

**LATERAL FACIES TRENDS IN DEEP-MARINE SLOPE AND BASIN FLOOR
MATRIX-RICH BEDS, NEOPROTEROZOIC WINDERMERE SUPERGROUP,
BRITISH COLUMBIA, CANADA**

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Abstract

This study investigates the lithological characteristics, and lateral and vertical facies trends of poorly understood, deep-marine matrix-rich sedimentary rocks. Two laterally extensive, well-exposed outcrops of slope and proximal basin floor deposits were investigated from the Neoproterozoic Windermere Supergroup. Significantly, matrix-rich beds have been found to undergo the same lateral trends (over ~200-650 m) in both outcrops. Initially, thicker, clayey sandstone transitions laterally to a bipartite bed with the development of an upper, planar-based, more matrix-rich unit. Further laterally, the basal unit progressively thins until it pinches out, and all that remains is the upper, more matrix-rich unit – a sandy claystone. It too thins and then pinches out. Draping the entire transect is a thin, matrix-poor structured unit overlain by a mudstone or claystone cap. These trends are interpreted to reflect a progressive but rapid lateral evolution of flow structure controlled primarily by particle settling, namely sand, from mud-rich avulsion-related flows.

Résumé

La présente étude examine la lithologie et les variations latérales et verticales de faciès mal connues dans les roches sédimentaires à matrice argileuse de milieu marin profond. Les dépôts de milieu marin de pente et de plancher proximal du bassin ont été étudiés dans le Supergroupe de Windermere d'âge Néoprotérozoïque le long de deux affleurements latéralement continus et bien exposés. Les roches sédimentaires à matrice argileuse montrent des tendances latérales (~200-650 m) similaires le long de ces deux affleurements. Les grès à matrice argileuse en lits épais passent latéralement à des grès (« bipartite ») dont la partie supérieure est un claystone sableux avec une base planaire et une composition plus argileuse. La partie inférieure progressivement s'amincit et disparaît ne laissant que le claystone sableux. On retrouve tout le long de ce transect une unité mince sans matrice argileuse surmontée à son tour par un mudstone or claystone. Ces tendances latérales sont contrôlées par l'apport et la déposition rapide de sable provenant d'écoulements par avulsion riche en boue.

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1 Thesis Introduction

1.1 Thesis Rationale

Science has always, most fundamentally, been driven by two things – basic human curiosity and societal needs. The motivation to understand and manipulate the Earth and its surroundings, from the atom to the contents of the solar system, has driven and illuminated passionate and curious minds since the dawn of man. Early academics paved the way for generations of like-minded, scientifically inclined individuals to continue the search for answers to life's questions – big and small. Basic human curiosity and societal needs, which have ignited and fueled the progress of science over human history, can be credited for the advances and breakthroughs that have led to the development of 'science and technology' that humans depend on today. For sedimentologists, the deep-marine realm of ocean basins, a vast and largely unobservable part of the Earth's surface, has held a certain fascination and sparked much curiosity. However, it is ultimately the economic potential of the deep-marine that has led many researchers to delve into the mysteries that lie on, but more importantly beneath the modern ocean floor.

Deep-water (>200 m) exploration for oil and gas started in the late 1970s, but grew rapidly in the 1980s (Shanmugam, 2000; Pettingill and Weimer, 2002). Today, the primary global conventional hydrocarbon reservoirs occur in deep-marine sedimentary rocks. Most of the exploration and development activity is located along passive margins, down-dip from producing Cenozoic deltaic systems (Pettingill and Weimer, 2002). In these areas, sand-rich turbidites, with their generally exceptional porosity and permeability, form the primary reservoir units and currently account for approximately 90% of all global deep-marine reservoirs (Pettingill and

Weimer, 2002). Despite their excellent reservoir quality characteristics, turbidite sandstone reservoirs are difficult to exploit where their stratigraphic architecture is poorly known.

Details of the internal stratigraphic geometries and architectures of passive margin turbidite systems remain poorly understood in part because of their paucity in the ancient sedimentary rock record. Even more fundamentally, the mode of deposition of an individual turbidite remains unclear due to the inaccessibility, rarity, and unpredictable occurrence of the sediment-gravity flows that emplace them (Piper and Normark, 2009). Thus, the behaviour of these flows is largely inferred from depositional products, and models based on theory and experimentation (Covault, 2011).

Today, modern turbidite systems are principally investigated using 2D and 3D seismic reflection data and drill cores. 2D and 3D seismic reflection data have proven invaluable in imaging sub-surface structures that otherwise would remain unknown. Nevertheless, despite the benefit, both 2D and 3D seismic have shortcomings. Most importantly, seismic surveys at the depth (>200 m) of deep-marine hydrocarbon exploration yield a minimum resolution of 10-15 m, which in many cases approximates the thickness of larger-scale architectural elements, but far exceeds the thickness of the individual stratal components that make them up (Arnott, 2010).

The Windermere Consortium, of which this thesis is part, is a government- and industry-funded research initiative that aims to improve the understanding of deep-marine turbidite deposits by studying an ancient, well-exposed, passive-margin turbidite system (termed the Windermere turbidite system) in the Neoproterozoic Windermere Supergroup (*Figure 1*). While the Windermere Supergroup has no economic potential, it does represent an important stratigraphic analogue for the analysis of other deep-marine turbidite systems (Ross and Arnott, 2007). The study area is located at Castle Creek in the Cariboo Mountains of the southern

Canadian Cordillera, in British Columbia, Canada (*Figure 1*). Here, the exceptional preservation, exposure, and accessibility of the rocks allows stratigraphic relationships, lithology, and lateral associations of the sedimentary complexes to be examined in detail.

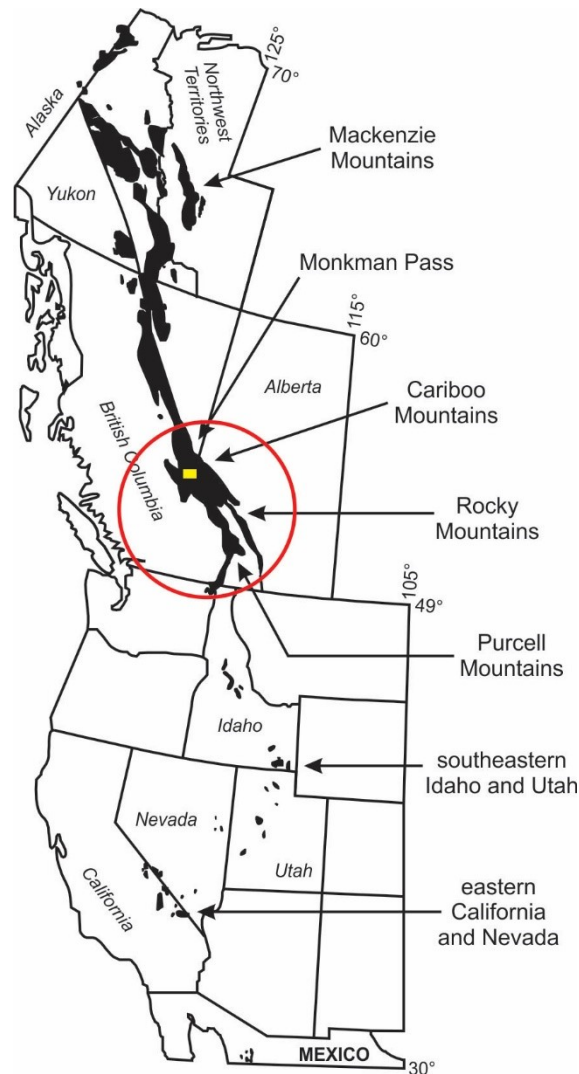


Figure 1. Exposure of the Windermere Supergroup (black) through Canada, the United States, and Mexico. Red circle indicates location of Windermere turbidite system, and yellow rectangle indicates location of Castle Creek study area (redrawn by Smith et al., 2014; after Ross, 1991).

The objective of this thesis is to describe the stratigraphic occurrence and origin of deep-marine matrix-rich sedimentary beds. Although increasingly recognized in the deep-marine

sedimentary record, little is known about their lateral facies changes, which in turn has hindered the development of a consistent depositional model. At Castle Creek, however, these strata crop out continuously over several hundreds of meters laterally in uncovered periglacial exposures. In addition, these strata occur in both slope and proximal basin floor deposits and therein provides the opportunity for comparison of matrix-rich beds in both depositional locations.

1.2 Windermere Supergroup

The Windermere Supergroup (WSG) is an unconformity-bounded, 2-9 km thick Neoproterozoic (740-569.9 Ma) succession of syn- and post-rift sedimentary rocks that are thought to be associated with the break-up (rifting) of Rodinia and subsequent formation of the proto-Pacific Ocean (McMechan, 2015). It extends more than 4000 km southward from the Yukon-Alaska border (Ross and Arnott, 2007). In the western Canadian Cordillera, it is largely confined in the western Foreland Belt and Omineca Belt (Ross et al., 1989; Ross and Arnott, 2007), whereas further south, in the western USA, it crops out in parts of the Sevier Belt, and Basin and Range Province, and finally culminates in northwestern Mexico (*Figure 1*) (Ross et al., 1989). Strata of the Windermere Supergroup consist primarily of coarse-grained feldspathic conglomerate, pebbly sandstone, and shale with lesser carbonate rock. In the southern Canadian Cordillera, these rocks have undergone low-grade (greenschist facies) to high-grade (amphibolite facies) metamorphism and tectonic deformation, and therefore are more appropriately termed metasedimentary to metamorphic rocks. Nevertheless, in the Castle Creek study area (Cariboo Mountains, British Columbia) (*Figure 1*), where rocks have undergone low-grade (greenschist facies) metamorphism, primary sedimentary structures are well preserved, even though primary textural characteristics have been modified (i.e. detrital grains have been recrystallized – see

sections 1.4 Methodology, 3.1 Facies Terminology, and 3.3.2 Facies 1. Microscopic Characteristics for additional information). For this reason, terminology for clastic sedimentary rocks is used herein.

1.2.1 Terminology

The name ‘Windermere System’ was first used by Walker (1926) to collectively describe sedimentary rocks that cropped out in the Windermere Valley of southern British Columbia. Today, the ‘Windermere Supergroup’ includes all Neoproterozoic rocks in western North America which, as noted above, form a discontinuous band that extends from the Yukon-Alaska border to northwestern Mexico (*Figure 1*). In the southern Canadian Cordillera, stratigraphic nomenclature for the Windermere Supergroup is largely dependent on geographical location: the Miette Group includes Windermere Supergroup rocks in the western Rocky Mountains, the Horsethief Creek Group includes those in the northern Purcell, Selkirk, and Monashee Mountains, the Kaza and Cariboo groups in the Cariboo Mountains, and finally the Mica Creek assemblage comprising high grade metamorphic rocks in the Monashee Complex (Arnott and Ross, 2007; Smith et al., 2014). For the purpose of this thesis, the strata in the Cariboo Mountains and, more specifically, at the Castle Creek study area will be described (*Figure 2*).

COLUMBIA MOUNTAINS					ROCKY MOUNTAINS						
CARIBOO MOUNTAINS ¹		SELKIRK MOUNTAINS ²		PURCELL MOUNTAINS ³	SELWYN AND MAIN RANGES ⁴		JASPER ⁵		LAKE LOUISE ⁶		
Cariboo Group	Yankee Belle Fm										
	Cunningham Fm										
	Isaac Formation										
Kaza Group	Upper	Horseshief Group	Upper Pelite Unit	Horseshief Group	Upper Clastic Div.	Miette Group	Upper Miette (East Twin Fm)	Miette Group	Upper Wynd Formation	Miette Group	Hector Formation
	'Marker'		Central Clastic Unit		Carbonate Div.		Middle Miette (McKale Fm)		Lower Wynd Formation		Taylor Lake/Mt. Temple mbrs
	Middle		Comedy Ck. Unit		Slate Division		MMM/OFP		Old Fort Point Fm		Corral Creek Formation
	Lower		Basal Pelite Division		Upper Grit Division	Middle Miette (McKale Fm)	Meadow Creek Formation				
	Middle Marble Unit		Mid-Marble Unit		Lower Grit Division	Lower Miette (Cushing Ck. Fm)					
	Semipelite Amphibolite Unit		Semipelite Amphibolite Unit		Basal Pelite Division & Carbonate Unit						
	Lower Pelite Unit		Pelite Unit		Irene Formation						

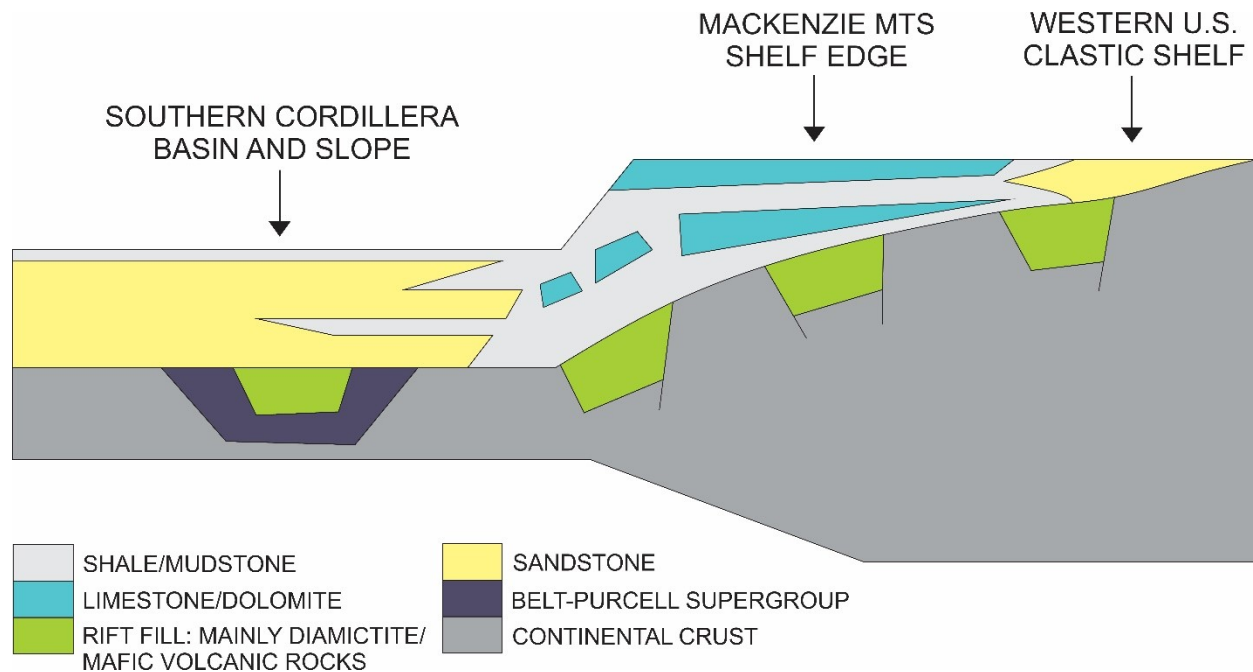


Figure 3. Interpretation of the western North American margin during the Neoproterozoic. Lower green deposits are syn-rift, whereas overlying deposits are post-rift (redrawn from Ross, 1991).

Glacial/glaciomarine deposits of the Toby Formation occur mostly at the base of the Windermere Supergroup where they are intercalated and/or overlain by volcanic rocks of the Irene Formation (Figure 4). Continued rifting eventually led to the development of the proto-Pacific Ocean and formation of a passive continental margin along the edge of Laurentia (Figure 3). These rocks are then overlain by a post-rift succession, which in the Cariboo Mountains of east central British Columbia comprises strata of the Kaza and Cariboo groups. This succession, which consists of basin floor deposits (Kaza Group) overlain progressively by slope and then shelf strata (Cariboo Group), is interpreted to form a continuous, several km-scale upward-shoaling succession related to the progradation of the continental margin of Laurentia in the proto-Pacific Miogeocline (Ross and Arnott, 2007).

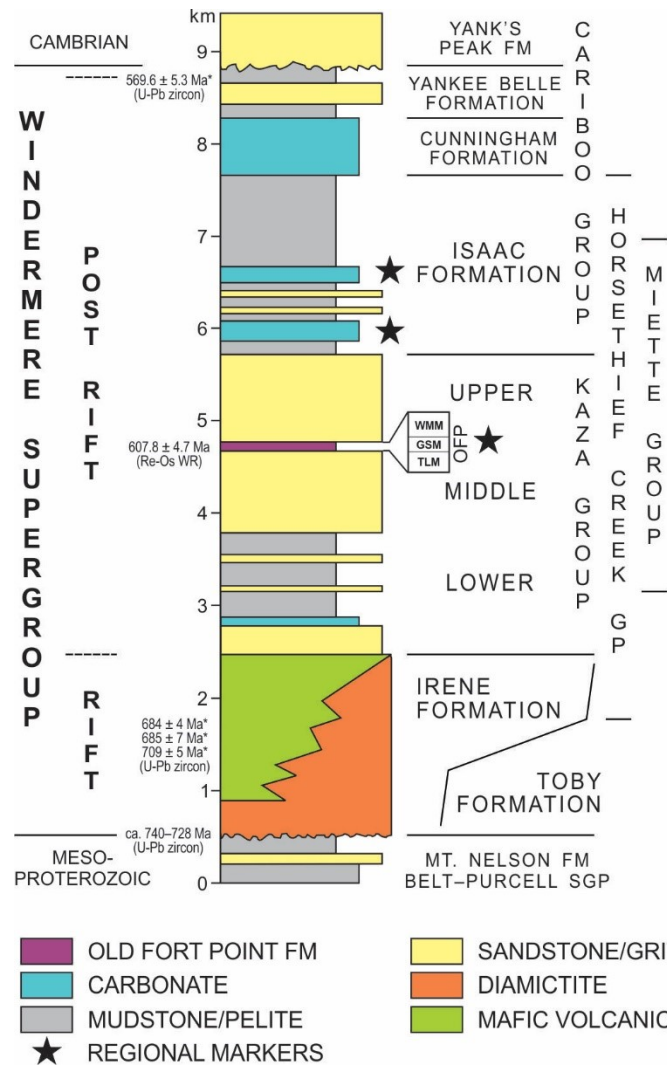


Figure 4. Stratigraphic log of the Windermere Supergroup showing syn-rift (rift) and post-rift successions (modified from Ross and Arnott, 2007).

In the southern Canadian Cordillera, a regional unconformity separates the top of the Windermere Supergroup from Lower Cambrian rocks (Aitken, 1969). Here evidence of rifting, uplift, and erosion of the Windermere Supergroup suggests that Lower Cambrian sedimentation was initiated by a second rifting event, which would have been necessary to produce the required heat flow to drive Paleozoic subsidence that led to the development of the Western Canada Sedimentary Basin (Armin and Mayer, 1983; Bond et al., 1983; Bond and Kominz, 1984; Roots and Parrish, 1988; Delvin and Bond, 1988, Lickorish and Simony, 1995).

1.2.3 Sediment Provenance and Paleocurrents

Although uncommon in sedimentary rocks of the Windermere Supergroup, detrital zircons show a bimodal distribution of Archean (>2.6 Ga) and Paleoproterozoic (1.9-1.75 Ga) ages, with a notable absence between 1.9-2.6 Ga (Ross and Bowring, 1990; Ross and Parrish, 1991a, b). In the general vicinity of the southern Canadian Cordillera, basement rocks of this age are confined to western North America – specifically cratonic rocks in southern Alberta and the northwestern United States (Ross and Arnott, 2007). Lemieux et al. (2007), however, state that zircons could also have been sourced from an outboard or eastern source. Regional paleocurrent data within the Windermere Supergroup suggest a general west to northwest direction of sediment transport, which is consistent with an eastern to southeastern sediment source (Ross and Arnott, 2007).

1.2.4 Geochronology

Geochronological control in the Windermere Supergroup is poor, largely because of its siliciclastic nature, lack of volcanic rocks, and paucity of fossils. U-Pb zircon dating of granitic basement rocks that non-conformably underlie the Windermere Supergroup has yielded a maximum age of $728 \pm 8-7$ to 740 ± 36 Ma (*Figure 4*) (Evenchick et al., 1984; Parrish and Scammell, 1988). Within the Windermere Supergroup, an age of approximately 685 Ma (U-Pb zircon dating) has been assigned to the lower rift-related volcanic rocks of the Irene Formation based on stratigraphic correlation with a rhyolite in Idaho (*Figure 4*) (Lund et al., 2003). Kendall et al. (2004) have dated the Old Fort Point Formation (see next) at 607.8 ± 4.7 Ma using Re-Os dating of an organic-rich mudstone (*Figure 4*) (Smith et al., 2014). Finally, a minimum U-Pb zircon age of 569.6 ± 5.3 Ma was obtained from volcanic rocks from near the base of the Hamill

Group that unconformably overlies the Windermere Supergroup (*Figure 4*) (Colpron et al., 2002).

1.2.5 Regional Markers

Three regional marker units have been used to correlate strata of the Windermere Supergroup throughout the southern Canadian Cordillera (Ross and Murphy, 1988; Ross et al., 1995). These include the Old Fort Point Formation and two deep-water limestone units. The Old Fort Point Formation can be correlated over 35,000 km² in the southern Canadian Cordillera and ranges in thickness from ~50-450 m, but most commonly is 60-125 m (Smith et al., 2014). It separates the middle and upper parts of the Kaza Group, and is stratigraphically lowest of the three regional markers (*Figure 4* – shown as ‘OFP’). Compositionally, the Old Fort Point Formation is characterised by three lithologically distinct units: siltstones, limestone-siltstone rhythmites, and mudstone-pelites (Smith et al., 2014). These three units are consistently bounded by coarse-grained Windermere Supergroup deposits. The depositional age of this marker unit, 607.8 ± 4.7 Ma, combined with its mineralogical composition, fine grain size, and geochemical characteristics has led to the suggestion that it was deposited during a sea-level highstand (Ross and Murphy, 1988) caused by melting of the Neoproterozoic Marinoan ice sheet (Ross et al., 1995). Further support for this hypothesis is provided by Ediacaran trace fossils and mega-fossils found in deep sea floor strata of the late Neoproterozoic which may indicate waning glaciations (Aitkin, 1989).

In addition to the Old Fort Point Formation, two deep-water limestone units are regional markers in the Isaac Formation (*Figure 4* – shown in blue). These units range from 10 to over 200 m thick, and are composed of thin-bedded carbonate turbidites, debris-flow deposits,

calcarene wackestones, and large carbonate slump blocks (Arnott and Ross, 2007). Like the Old Fort Point Formation, these units were deposited during highstand conditions, but this time with associated increased carbonate production on an adjacent continental shelf and reduced siliciclastic input into the deeper marine realm (Ross et al., 1989; Ross and Ferguson, 2003a, b; Ross and Arnott, 2007).

1.2.6 Basement

The Windermere Supergroup is underlain by either the Mesoproterozoic Belt-Purcell Supergroup metasedimentary rocks (1.5-1.4 Ma) in the Purcell and southern Selkirk Mountains (southeastern British Columbia, northeastern Washington, and northwestern Montana) (Evans et al., 2000), crystalline basement (2.2-0.73 Ma) (Crowley, 1996) of the Deserters Gneiss in the Deserters Range (Evenchick et al., 1984), or Malton Gneiss and associated gneisses south of Valemount, British Columbia (Murphy, 1990).

1.2.7 Toby Formation

The Toby Formation is stratigraphically equivalent to the Shedroof and Huckleberry conglomerates of Idaho and Washington (Aalto, 1971). It unconformably overlies the Upper Purcell Supergroup (Reesor, 1957), and is conformably overlain by either volcanic greenstones (in parts of Washington, Idaho, and British Columbia) (*Figure 4*) or clastic sedimentary rocks (conglomerate, sandstone, argillite, and carbonate) of the Horsethief Creek Group (Purcell Mountains) (Aalto, 1971).

The Toby Formation is composed predominantly of massive diamictite deposits (65%), with lesser amounts of argillite (20%), and conglomerate and sandstone (15%). Its thickness

ranges from 0 m (where absent) to 1864 m (Flint et al., 1960; Aalto, 1971), and changes dramatically over short geographical distances, which has been attributed to filling of laterally discontinuous rift-related sub-basins during glaciations (Stewart, 1972). Stratigraphically upward, the Toby Formation shows only subtle change in lithology. Diamictites are more prevalent at the stratigraphic base of the formation, and they gradually transition upward to argillaceous and fine sandstones (Aalto, 1971). Rare striated clasts, dropstones, and “exotic” granitic clasts support the interpretation that the Toby Formation is glacial or glaciomarine in origin (Aalto, 1971); an interpretation that is consistent with the high percentage of diamictite, and also the timing of the late-Precambrian Sturtian glaciation (approximately 750-700 Ma) (Ross et al., 1995).

1.2.8 Irene Formation

Conformably overlying and interstratified with glaciogenic strata of the Toby Formation is the Irene Formation (*Figure 4*). This formation is stratigraphically equivalent to the Huckleberry Greenstone and Leola Volcanic Unit in northeastern Washington (Aalto, 1971). It is made up of foliated and altered andesitic greenstone, tuff, breccia, agglomerate, and volcanic conglomerate; vesicle and pillow structures are present (Aalto, 1971). Beds alternate between green vitric-lithic tuff and chloritized ash-rich mud; these are mixed with other volcanic lithologies (Aalto, 1971). The Irene Formation is interpreted to be a syn-rift volcanic unit that was deposited both during and after Toby Formation deposition.

1.2.9 Kaza Group

In the Cariboo Mountains, the Kaza Group is 2-4 km thick and is interpreted to conformably overlie the Toby and Irene formations (*Figure 4*). It is subdivided into lower, middle, and upper parts, and also contains the Old Fort Point Formation (a regional marker – see section 1.2.5 Regional Markers), which separates the middle and upper Kaza. This group is composed predominantly of feldspathic, carbonate-cemented sandstone interbedded with mudstone. The sandstone to mudstone ratio is approximately 75:25. Sandstones are typically sheet-like and grade laterally and intercalate vertically with thin-bedded turbidites. These expansive sandstone bodies have been interpreted as depositional lobe elements that were deposited on the Windermere basin floor by relatively unconfined flows (Meyer and Ross, 2007; Terlaky et al., 2015). Moreover, uncommon mass movement deposits in the upper Kaza Group (compared to the overlying Isaac Formation) suggest negligible gravitational stability, and hence an area located away from the continental slope.

1.2.10 Isaac Formation

The Isaac Formation is 1.4 km thick, contains the two deep-water limestone regional markers, and conformably overlies the Kaza Group (*Figure 4*) (Arnott et al., 2011). Unlike the Kaza Group, it is composed predominantly of mudstone (sandstone to mudstone ratio of 25:75). Isolated coarse-grained to conglomerate sandstone bodies within the Isaac Formation are interpreted to be channel fill complexes, with adjacent sandstone and mudstones interpreted to be levee or mass transport deposits. Channel complexes are generally 20-100 m thick and are composed of multiple individual channel fills (Davis, 2011). At Castle Creek, six channel complexes have been identified in the Isaac Formation. Channels are mostly aggradationally

filled with medium- and coarse-grained sandstone, although less common laterally accreting channel fills are observed. Levee deposits are composed of thin-bedded turbidites, interbedded with splay deposits. Mass movement deposits including slump, slide, and debris flow deposits are common and suggest gravitational instability. These strata, in addition to the leveed-channel fills, suggests deposition on the continental slope.

1.2.11 Cunningham and Yankee Belle Formations

The Isaac Formation is then conformably overlain by the Cunningham, and, in turn, Yankee Belle formations (*Figure 4*). The Cunningham Formation is an oolitic intraclastic limestone that is up to 550 m thick, and the Yankee Belle Formation is composed of interbedded siliciclastic and carbonate rocks that are up to 900 m thick (Gabrielse and Campbell, 1992). Ross et al. (1995) have interpreted both formations as shallow-marine, high energy deposits.

Collectively, the Kaza Group to Yankee Belle Formation form a several km-thick, upward-shallowing succession of basin floor (Kaza Group) to continental shelf (Yankee Belle Formation) deposits, which Ross and Arnott (2007) attribute to progradation of the Laurentian continental margin into the Pacific miogeocline.

1.2.12 Metamorphism and Structure: The Canadian Cordillera

As a result of the Mesozoic deformation event that produced the North American Cordillera, the present-day position, geometry, and thickness of the Windermere Supergroup have been post-depositionally altered. More specifically, Late Jurassic to Early Tertiary allochthonous terrane accretion and orogenesis on the western margin of Laurentia caused severe shortening and crustal thickening (via thrust faulting and folding) of autochthonous western

Laurentian margin Proterozoic and Paleozoic sedimentary rocks. Today, Windermere Supergroup sediments are located primarily in the Omineca and Foreland belts; two of five belts that make up the North American Cordillera.

The western Canadian Cordillera is made up of five geologically distinct (characteristic lithologies, structures, and morphologies) belts termed the Insular Belt, Coast Belt, Intermontane Belt, Omineca Belt, and Foreland Belt (from west to east) (*Figure 5*). This orogen evolved in three main stages. The first stage, as discussed above, involved the formation of the Windermere Supergroup and the overlying Paleozoic succession. Windermere Supergroup sedimentation was initiated with Neoproterozoic rifting of Rodinia and continued as a passive margin formed on the western continental margin of Laurentia. These events were followed by Paleozoic passive margin sedimentation. The second stage involved terrane accretion and orogenesis on the western margin of the Laurentian craton during the Paleozoic, and Mesozoic to Eocene. As a result of terrane accretion, the western edge of North America was extended by approximately 500 km (Monger et al., 1982). The third stage involved tectonism (strike-slip faults and regional extension) in the North American Cordillera from the late Cretaceous to early Eocene. Today, these three events are expressed in the five belts of the North American Cordillera.

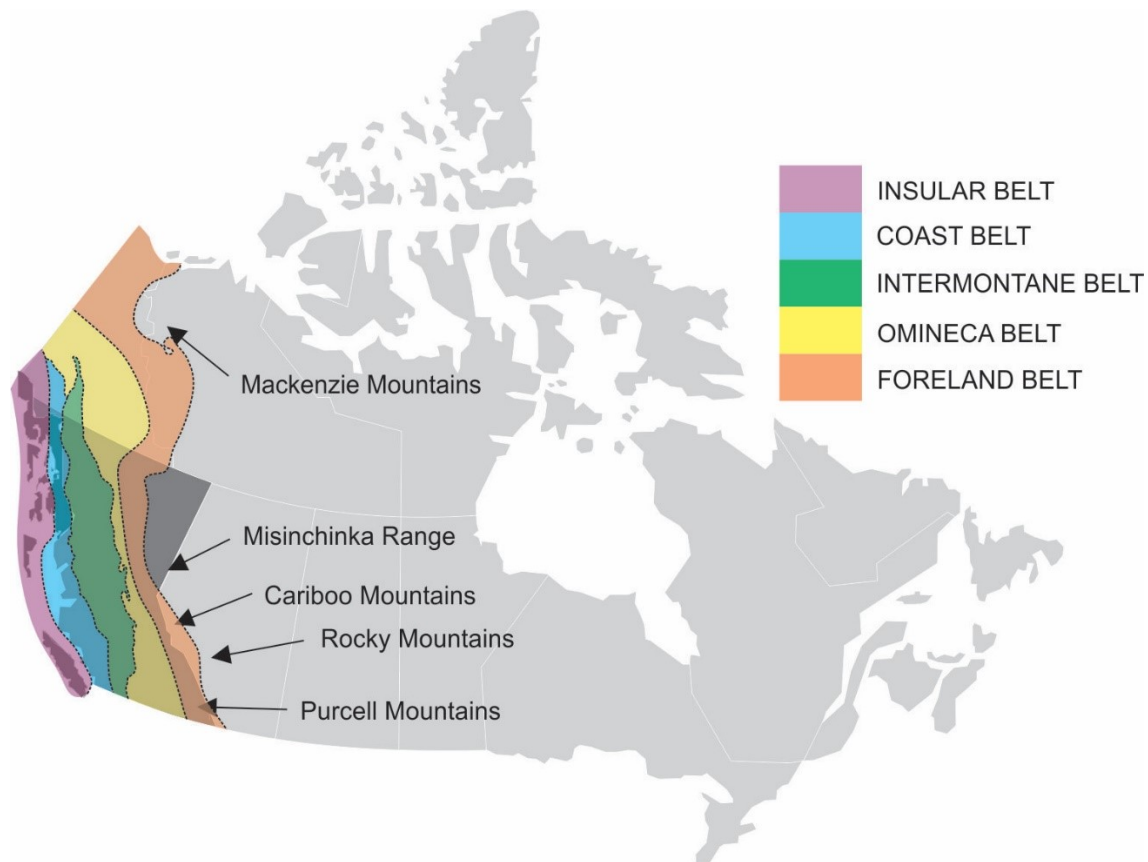


Figure 5. The western Canadian Cordillera is made up of five geologically distinct belts which, from west to east, include the Insular Belt, Coast Belt, Intermontane Belt, Omineca Belt, and Foreland Belt (modified from Davis, 2011).

The Foreland Belt, the eastern-most belt, overlies Precambrian basement gneisses and is composed primarily of slightly metamorphosed Proterozoic to Mesozoic sedimentary rocks. The oldest sedimentary sequences, which are Proterozoic in age, include the Belt Purcell and Windermere supergroups. Subsequent deposition of passive margin carbonates occurred in the Paleozoic, and was followed by Middle to Upper Paleozoic continental margin sedimentation. The youngest units were deposited in a foreland basin during Mesozoic terrane accretion and orogenesis.

The Omineca Belt is a mountainous region that includes metamorphosed sedimentary autochthonous rocks, and pericratonic terranes (allochthonous) that were accreted during the Mesozoic. It is composed primarily of granitic and metamorphic rocks. During the Mesozoic, the Omineca Belt experienced three major tectonic pulses (middle Jurassic, early Cretaceous, and late Cretaceous to Eocene) associated with the obduction of three suspect terranes onto the western edge of North America, which resulted in east-verging brittle and ductile deformation. It was these pulses that exposed the Windermere Supergroup, and moved it to its present-day position inboard of the western margin of ancestral North America. During deformation, the stratigraphic base of the Windermere underwent ductile deformation, whereas the overlying shallower rocks underwent both ductile and brittle deformation (Campbell, 1970). Most of the rocks in the Omineca belt have undergone high grade (amphibolite facies), Barrovian-style metamorphism, which is common in orogenic belts. However, expansive but localized areas of low grade (greenschist facies) rocks occur in the eastern Purcell and northern Cariboo Mountains, and it is here that sedimentological investigations are possible (Ross and Arnott, 2007).

The Intermontane Belt, composed of three accreted terranes – Stikinia, Cache Creek, and Quesnellia, was the first superterrane to dock along the western margin of North America during the late Triassic, forming a small, cold accretionary orogen. This belt is topographically lower than the surrounding Omineca and Coast belts, and low-grade metamorphic rocks that comprise it are intruded by Jurassic- to Tertiary-aged plutons.

The Coast Belt is a high-relief region composed primarily of high grade metamorphic rocks with Late Jurassic to Early Tertiary plutonic rocks. Similar to other belts within the

Cordillera, most of this belt is composed of terranes that had been accreted to the western margin of the North American craton.

The Insular Belt, like the Intermontane Belt, is a superterrane that is composed of the Wrangellia and Alexander terranes. It docked during the middle Cretaceous and formed a large, doubly vergent orogeny. The modern Pacific margin is located along the western edge of this belt.

1.2.13 Castle Creek Study Area

The Castle Creek study area is located in the northern Cariboo Mountains of the southern Canadian Cordillera (*Figure 1*). Here deep-water rocks of the Windermere turbidite system are exposed on one limb of a major anticline on the eastern side of the Isaac Synclorium (Arnott and Ross, 2007) (*Figure 6*). The exposure is 2.5 km perpendicular to bedding, and 7 km parallel to bedding and exposes approximately 800 m of the upper Kaza Group overlain by the Isaac Formation of the Cariboo Group. Beds are near-vertically dipping ($\sim 89^\circ$) and are superbly exposed. Recent glacial retreat has removed most surface debris and has polished the outcrop surface. As a consequence, strata can be traced laterally for 100's of meters to over 1000 meters.

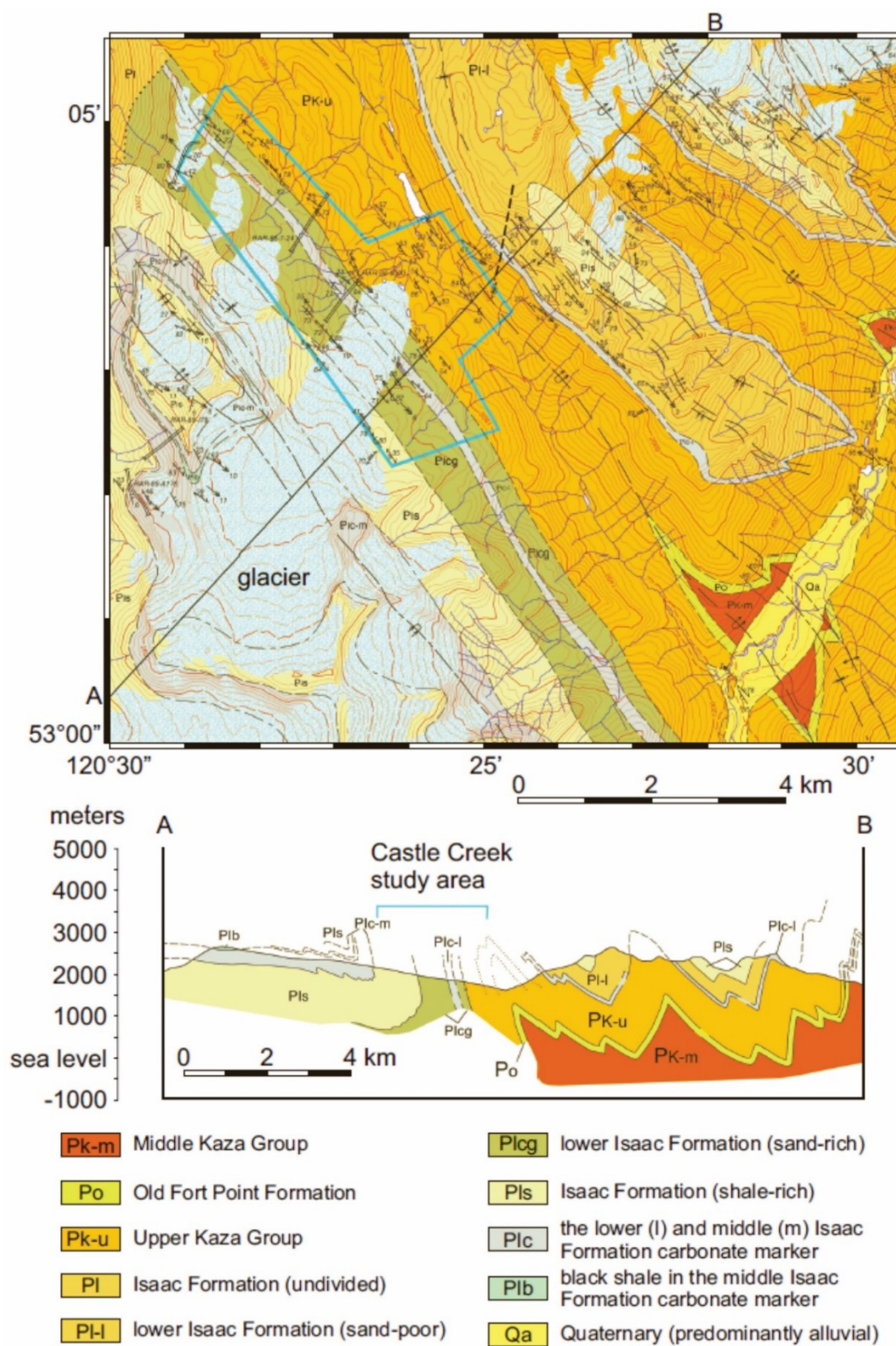


Figure 6. Regional geology of the Castle Creek study area (outlined in blue) (excerpt from the Eddy 1:50,000 map area; Ross and Ferguson, 2003a).

Within the Kaza Group of the Windermere Supergroup, micro- and macro-structures have been analysed, particularly their superimposition and cross-cutting relationships in addition to mid-Jurassic mineral growth (Murphy, 1987b). Findings suggest that the Castle Creek study area has undergone four deformation events (D1, D2, D3, and D4) and at least two episodes of low-grade metamorphism (M1 and M2) (Murphy and Rees, 1983; Murphy, 1987a, b). These deformation events occurred before, during, and after a mid-Jurassic peak in regional metamorphism. Metamorphism is thought to have occurred at temperatures ranging from 350 to 450°C, and pressures ranging from 4 to 9 kbar (Greenwood et al., 1991). In general, metamorphic grade decreases from the base of the Kaza Group to the top of the Isaac Formation. Higher grade regions are dominated by garnet-rich tight to isoclinal, southwest verging folds with gently dipping axial surfaces. Lower grade regions, on the other hand, are dominated by syn-kinematic muscovite- and chlorite-rich, open to tight, south-east verging folds with steeply dipping axial surfaces (Murphy, 1987b). The Isaac Formation, upper Kaza, and part of the middle Kaza are more brittlely deformed compared to the lower part of the middle Kaza and the lower Kaza which underwent more ductile deformation. Both deformation and metamorphism are interpreted to be associated with collision and accretion of terranes along the western margin of North America from the mid-Jurassic to Tertiary (Murphy, 1987b).

A pre- to early-metamorphic D1 event produced north west-trending, north east-verging, recumbent, isoclinal local folds with cleavage that is bedding parallel or locally inclined to bedding (Murphy, 1987b; Ross and Arnott, 2007). Associated M1 low-grade metamorphism formed a low-grade mineral assemblage of muscovite, biotite, and chlorite, and resulted in minor structural thickening (Murphy, 1987b). This deformation event is thought to be associated with the obduction of the first suspect terrane onto the western North American margin during the

middle Jurassic (Murphy, 1987b). A syn-metamorphic D2 event produced northwest trending, south west verging, overturned folds and reverse faults, which are the predominant structures that affect the Castle Creek study area. This event is associated with significant structural thickening, and is also thought to have initiated M2. This second metamorphic event peaked after D2 as evidenced by metamorphic minerals that overgrow D2 structures (Murphy, 1987b). The D3 event is late- to post-metamorphic and refolds D1 and D2 structures into northwest trending, north east verging, open, upright folds with associated reverse faults. This event largely impacted the stratigraphic base of the sequence (i.e. lower middle Kaza and lower Kaza), but is less prominent in younger strata. The final D4 event is post-metamorphic and refolds D1, D2, and D3 structures into meter to decimeter north east trending, steeply-inclined to upright folds with open to tight inter-limb angles and chevron hinge zones; strike-slip faults are associated (Campbell, 1970; Campbell et al., 1973; Ferguson, 1994; Reid et al., 1997).

Palinspastic restoration of the Windermere turbidite system in the southern Canadian Cordillera assumes 30% shortening related to these deformation events (e.g. Brown et al., 1986; McDonough and Simony, 1988; Price, 2000), thus a 35,000 km² unrestored system would have been approximately 80,000-100,000 km² prior to deformation (Arnott and Ross, 2007). This, then, suggests that the Windermere turbidite system is comparable in size to a number of modern turbidite fan systems like the Amazon and Mississippi fans.

1.3 Thesis Objectives

Although first described in the late 1960's, deep-marine mud-rich rocks have, over the past decade, been increasingly and now widely recognized in the ancient sedimentary record (e.g. Lowe and Guy (2000), Haughton et al. (2003), Sylvester and Lowe (2004), Talling et al.

(2004), Amy and Talling (2006), Davis et al. (2009), Haughton et al. (2009), Hodgson (2009), Pyles and Jennette (2009), Barker et al. (2008), Girard et al. (2012), and Terlaky and Arnott (2014)). Based on published data of bed characteristics and their associated depositional interpretations, these mud-rich rocks can be divided into two general sub-groups (see section 4.3 Comparison with Published Mud-Rich Beds for more detail). The first sub-group includes beds composed of genetically linked turbidite and debrite parts, and have been termed linked debrites, hybrid event beds, co-genetic turbidite-debrite beds, or transitional flow deposits. The second sub-group is less well-represented in the literature, and includes slurry beds (first described by Lowe and Guy, 2000). Mud-rich ('matrix-rich') rocks at Castle Creek share few similarities with the mud-rich sub-groups discussed above, suggesting that each bed type has a unique mode of deposition. The objective of the current study, then, is to describe the lithological, textural, and mineralogical characteristics of matrix-rich rocks at Castle Creek, and illustrate their lateral and vertical trends in order to ultimately understand the physical basis for their deposition.

1.4 Methodology

At the Castle Creek study area (*Figure 7*), three sections comprising matrix-rich beds were logged in detail – two sections from the Kaza Group (proximal basin floor deposits) and one from the Isaac Formation (slope deposits). A total of 63 stratigraphic columns (the vertical succession of beds in a particular location) were measured bed-by-bed. In the Isaac, the section is approximately 350 m wide. Here, 11 logs ranging from 3.5-5.5 m long were measured (*Appendix A and B*). In the stratigraphically highest Kaza section, which is approximately 650 m wide, 38 logs were measured (ranging from 0.1-8.5 m in length) (*Appendix C and D*). In the stratigraphically lowest Kaza section, which is approximately 230 m wide, 14 logs were

measured (ranging from 1-18.5 m in length) (*Appendix E*). The relatively poor lateral exposure of matrix-rich beds (a result of scouring) in this section resulted in less detailed lateral lithological trends. For this reason, this section is not discussed in this thesis. The location of each stratigraphic log was determined after “walking out” beds and placed where important lithological changes were noted or simply where outcrop conditions were best. In each log, bed thickness, colour, basal contact morphology, grain size, primary and secondary sedimentary structures, mud clast presence and features (size, shape, rounding, composition, orientation, abundance, etc.) were noted and described in detail. Grain size was determined using a standard grain size comparator and a 16X Ruper hand lens, and all thickness measurements were made with a standard measuring tape. Photographs were taken of each bed. Excellent lateral continuity of the outcrop allowed individual beds to be walked out and mapped on high-resolution (1:300) aerial photographs – this was especially important for lateral bed correlation and illustrating the dimensions of the observed lithological changes along depositional strike.

Hand samples of matrix-rich and matrix-poor strata were collected for petrographic analysis. From these samples, a total of 59 thin sections were prepared at the University of Ottawa and analysed using a Leitz Wetzlar (Germany) microscope. Thin sections were analyzed for grain size (minimum, maximum, and mode), structures, fabric, in addition to grain sorting, rounding, shape, and also grain contacts within each sample. Significantly, all thin sections were point counted to quantitatively determine the relative abundance of framework grains, matrix, altered grains, and cement. Point counts were made every 1.125 mm ($\frac{1}{4}$ cross-hair in low (40X) power) until approximately 500 points were counted. The following equations, adapted from Popović (2016), were used to classify thin section constituents into framework grains, matrix, altered grains, and cement:

$$Total = Framework\ Grains + Matrix + Altered\ Grains + Cement$$

$$Total = [Q_S + F_S + L_S] + [M_{QS} + M_{RXC} + O_S] + [RXC] + CC$$

Where Q_S is sand-sized (or greater) quartz grains; F_S is sand-sized (or greater) feldspar grains; L_S is sand-sized (or greater) lithic grains; M_{QS} is silt-sized quartz matrix; M_{RXC} is recrystallized matrix (composed of muscovite and chlorite); O_S is sand-sized (or greater) chlorite or muscovite crystals; RXC is recrystallized silt-sized (or greater) quartz crystals; and CC is post-depositional calcite and/or metamorphic dolomite cement (in thin section, calcite is pore-filling and displays characteristically high birefringence and rhombic cleavage, whereas dolomite occurs as individual rhombic crystals with high relief imparted by iron-rich rims). Since all rocks in the study area have undergone low-grade metamorphism, greenschist facies features are present in all thin sections. The most important changes that result from this metamorphism are (1) primary clay matrix has been recrystallized to muscovite and chlorite, and (2) bulging recrystallization and sub-grain rotation occur along some grain boundaries (see sections 3.1 Facies Terminology and 3.3.2 Facies 1. Microscopic Characteristics for further explanation). These changes were taken into account when point counting. Small, intergrown crystals of muscovite and chlorite, as well as silt-sized or larger crystals were counted as matrix. Framework grains that appeared to be the product of bulging recrystallization or sub-grain rotation were counted as recrystallized grains, not primary framework grains. The resulting counts were recorded in an Excel spreadsheet and percentage totals calculated for framework grains, matrix, altered grains, and cement.

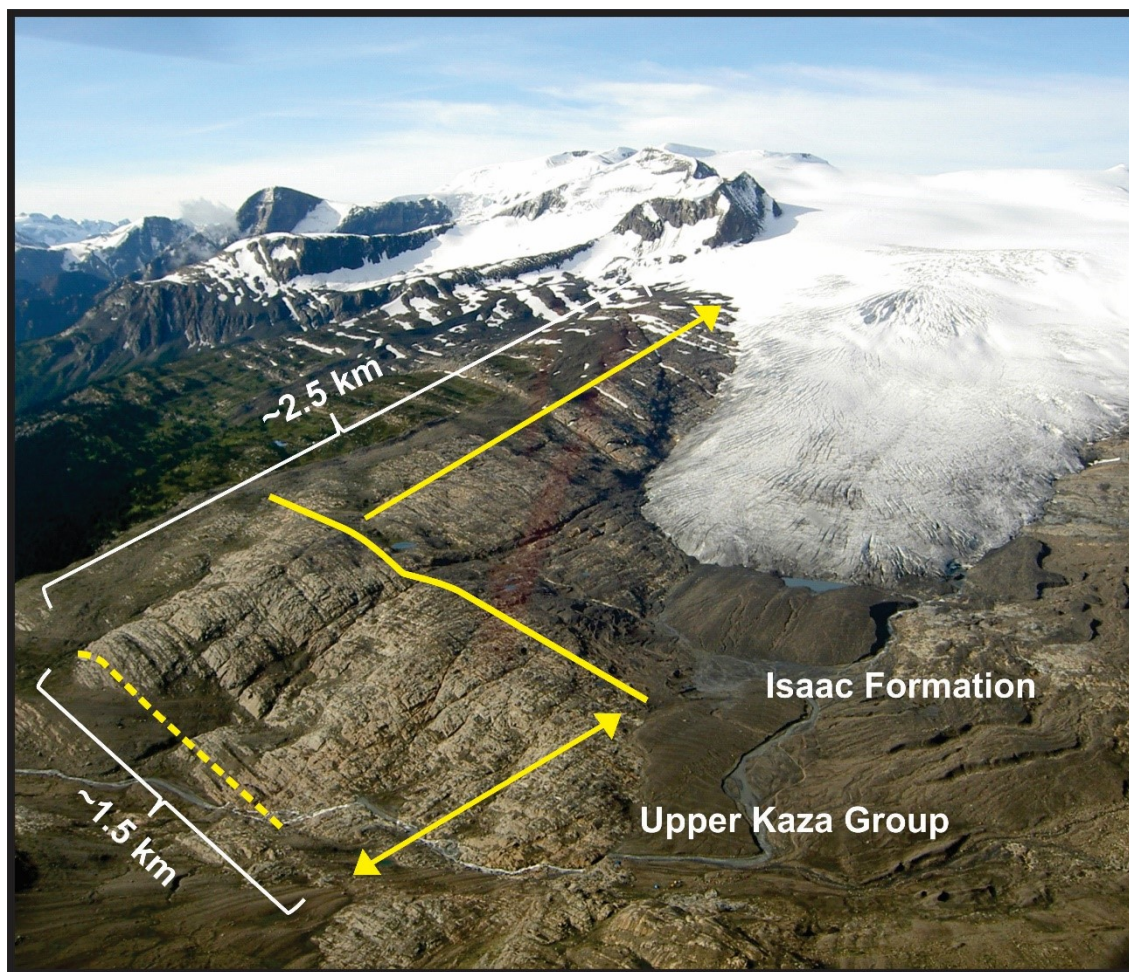


Figure 7. The Castle Creek study area is located in the Cariboo Mountains, east central British Columbia, Canada. Here, a 2.5 km thick section of vertically-dipping, recently deglaciated stratigraphy crops out and consists of proximal basin floor deposits of the upper Kaza Group (0.8 km thick) overlain by 1.7 km thick succession of leveed slope channels of the Isaac Formation.

1.5 Previous Work

In the Windermere Supergroup, Terlaky and Arnott (2014) have recognized matrix-rich beds in the upper Kaza (proximal basin floor), middle Kaza (medial to distal basin floor), and Isaac Formation (base of slope), where they make up approximately 6%, 4% and <1% of the total section, respectively. In their most proximal parts, these matrix-rich beds are typically laterally adjacent to coarser-grained, matrix-poor, scour-based beds that contain large, often

angular sandstone or mudstone clasts (several cm thick and long, up to about 15 cm thick and 1 m long; uncommonly several dm thick and several m long) near their bases. In their most distal parts, Terlaky and Arnott (2014) found that matrix-rich beds are continuous with thin-bedded turbidites. Additionally, matrix-rich beds in various depositional environments (proximal to distal basin floor, and base of slope) tend to be overlain by matrix-poor architectural elements of coarse sandstone (i.e. channels and splays). Together, these observations led to the interpretation that matrix-rich beds represent deposition during the initiation, or activation of the local sedimentary system. More specifically, they are interpreted to be deposited from avulsing flows that locally scoured a mud-rich seafloor, enriching the resulting unconfined flow with finer grained sediment (*Figure 8*). In the axial part of the avulsed flow, which in reality is a plane wall jet, coarser grains would have been deposited, forming clean (low matrix content), coarser-grained sandstone deposits. Fines would have become segregated to the tops and lateral margins of the flow, where they would eventually deposit matrix-rich beds. Further laterally, as the flow waned and became progressively more dilute, thin-bedded turbidites would have been deposited. Matrix-rich beds, then, would be succeeded vertically by typical deep-marine sedimentation – commonly terminal splays or distributary channels if the avulsion succeeded, and thin-bedded turbidites if it failed. Based on observed matrix-rich bed locations, Terlaky and Arnott (2014) suggest that this mode of matrix-rich sandstone deposition can occur anywhere on the slope to the distal basin floor, however, it is most common in proximal basin floor settings where avulsion is typically most frequent.

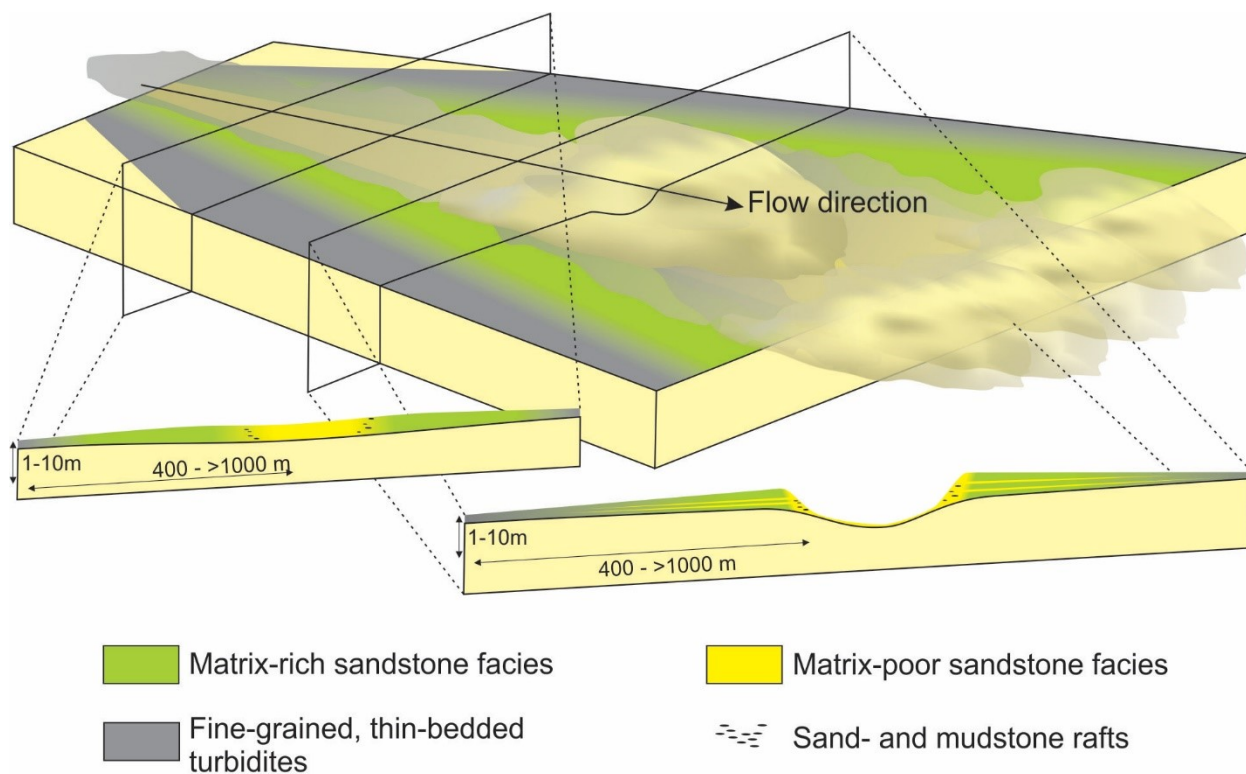


Figure 8. Depositional model for matrix-rich beds at Castle Creek (Terlaky and Arnott, 2014). An avulsing flow locally scours a mud-rich seafloor, enriching the resulting unconfined flow with finer grained sediment. Matrix-poor, coarser-grained sandstones are deposited near the axis of the flow, and matrix-rich beds further laterally. At the most distal extents, fine-grained, thin-bedded turbidites are deposited (from Terlaky and Arnott, 2014).

2 Deep-Water Physical Sedimentation Processes

Sediment-gravity flows, namely turbidity currents and debris flows, are the dominant agents for transporting and depositing sediment in the deep sea (>200 m water depth and seaward of the continental shelf break) (Shanmugam, 2000). As their name implies, they are mixtures of sediment and water that are driven down slope along the sea floor by gravity. Due to the local occurrence of suspended sediment, sediment-gravity flows are denser than surrounding sea water, resulting in (negative) buoyancy forces that cause the suspension to move. In the presence of a slope, these flows move in a single direction. Most often, sediment-gravity flows are initiated at the continental shelf break via slope failure, however, they can also be sourced from hyperpycnal river flows, or local storm suspensions near the shelf edge (Piper and Normark, 2009). From the site of initiation, sediment-gravity flows can transport sediment hundreds to thousands of kilometers across the sea floor. Longer run-out distances typically associate with fine-grained flows (especially where clay particles are present) due to reduced shear (and therefore mixing) at the base and tops of these flows, which is a consequence of a more plug-like density profile (Tilston, 2015). Although some flows may have longer run-outs, all sediment-gravity flows evolve (with distance and time) down-slope as a result of gradient change, fluid entrainment, erosion, deposition, and/or flow expansion (Mulder and Alexander, 2001; Piper and Normark, 2009). Thus, the final deposit type is generally a poor indicator of the initiation process of a sediment-gravity flow (Piper and Normark, 2009).

Slides and slumps, in contrast to turbidity currents and debris flows, are gravitationally-driven mass movements initiated by slope failure. A slide occurs when a consolidated block detaches from the sea floor along a plane of weakness, and moves down slope along a mostly planar surface. Deformation in the transported sediment mass is generally limited to its lower

part. Slumps, on the other hand, detach along a concave-up plane of weakness, but also show more extensive internal disruption, commonly in the form of overturned folds. Slides and slumps deposit en-masse when friction at the base of the moving mass exceeds the driving force of gravity. Most commonly, slides and slumps are deposited on the continental slope, but rarely do travel as far as the basin floor (as has been shown in the Windermere Supergroup) (Bill Arnott, 2015, personal communication). When travelling down high-gradient slopes, slides and slumps can evolve into debris flows and ultimately turbidity currents which, as noted above, deposit most of the sand-rich part of the deep-marine sediment record (Shanmugam, 2006). Due to their absence in the current study, slide and slump deposits will not be discussed further.

Turbidity currents are classified as dilute Newtonian fluids, which essentially means that they lack strength (*Figure 9*). Flows are considered fully turbulent when mixing by eddies occurs throughout the entire body of the flow. Within turbulent flows, grains are principally supported by the upward component of fluid turbulence, and are deposited via differential settling to produce classical, or Bouma, turbidites (*Figure 10*). Bouma turbidites are widely recognized in the deep-marine sedimentary record as deposits that exhibit a generally gradual upward change in grain size, usually fining, and an associated upward change in sedimentary structures (Bouma, 1962) – both features a consequence of waning flow conditions. A complete Bouma sequence comprises five divisions, labelled A through E. Although examples of full-sequence Bouma turbidites exist, most turbidites found in nature are missing at least one division. The typical changes in grain size and bedforms seen in a Bouma sequence are shown in *Figure 10*.

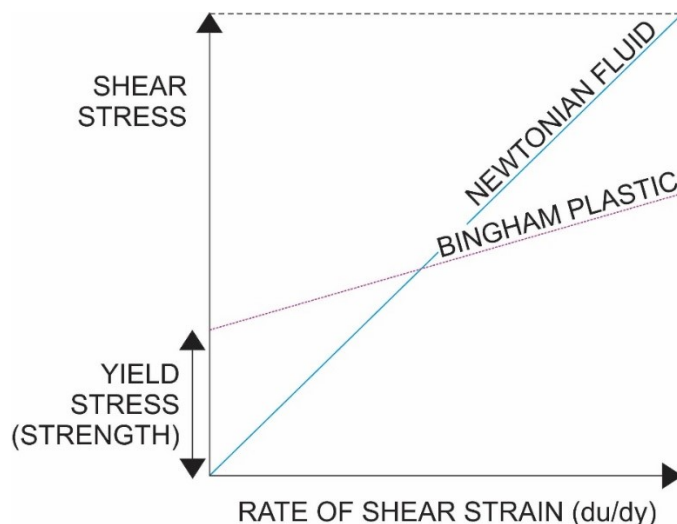


Figure 9. Rheology (stress-strain relationships) of Newtonian fluids (e.g. turbidity currents) and Bingham plastics (e.g. debris flows) (redrawn from Shanmugam, 2000).

Grain Size	Bouma Division	Description
Mud	Te	Massive mudstone (pelagic and hemipelagic sedimentation)
	Te	Massive or graded mudstone (turbidity current deposition)
Sand-silt	Td	Laminated siltstone
	Tc	Ripple cross-laminated fine-grained sandstone
	Tb	Parallel laminated fine- to medium-grained sandstone
Sand	Ta	Massive or normally graded fine- to coarse-grained sandstone (with granules at base) +/- pebbles, dewatering structures Planar or scoured base +/- rip-up clasts

Figure 10. Complete Bouma sequence showing sedimentary structures and grain-size typical of each Bouma division (redrawn from Bouma, 1962).

In contrast to turbidity currents, debris flows are Bingham plastics, which means that they possess measurable mechanical strength (*Figure 9*). In order to flow (strain), a threshold yield stress must be surpassed. Within debris flows, sand-grade and larger grains are maintained in suspension by a cohesive matrix made up of a mixture of interstitial fluid and fine (clay- and silt-sized) sediment. These flows deposit their sediment load en masse when the shear resistance of the flow (interplay of viscosity and friction) exceeds the driving force of flow movement (gravity). Turbidity currents and debris flows are considered end-member flow types – non-cohesive and cohesive flows, respectively. Although these two processes are important, they are not the only sediment-gravity flows responsible for sediment transport and deposition in the deep sea. Findings from ancient and modern deep sea deposits, experimental results, and theoretical models suggest that a variety of intermediate flow types exist between these two end-members. Summarizing this diversity, Mulder and Alexander (2001) proposed a simplified classification scheme for sediment-gravity flows based on (1) physical flow properties, (2) grain support mechanisms, and (3) characteristics of flow deposits. In the classification, sediment-gravity flows are separated into two groups: (1) cohesive flows (mud flows and debris flows), and (2) non-cohesive friction flows (turbulent flows, concentrated flows, and hyperconcentrated flows). For the purpose of this thesis, the sediment-gravity flow classification scheme of Mulder and Alexander (2001) is used.

Flows that contain clay are typically partially or fully cohesive due to the forces formed through the interaction of adjacent clay particles. Clay particles have weak (negative) electrostatic surface charges and, when in close proximity, form weak electrostatic bonds which change (strengthen) the rheology of the flow. When the clay content becomes sufficiently high, and an extensive network of bonded clay particles forms, the flow becomes increasingly viscous

and eventually can dampen and even suppress turbulence, at which point the flow is considered fully cohesive. The concentration of clay required to form a fully cohesive flow depends on flow conditions, sediment concentration, and clay mineralogy (Mulder and Alexander, 2001).

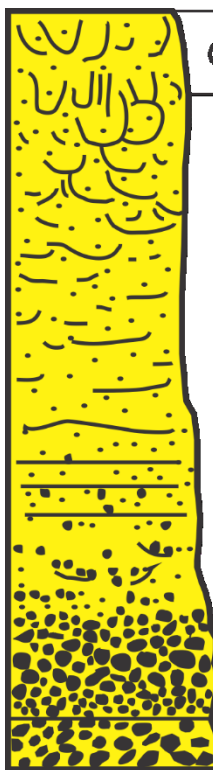
Cohesive flows, as described in Mulder and Alexander (2001), include mud flows and debris flows. Matrix-strength is the dominant grain-support mechanism in cohesive flows and can continuously suspend grains up to coarse-grained sand (Hampton, 1975). The presence of larger grains or clasts suggests additional supporting mechanisms, including local high pore pressure, particle-particle interactions, and/or buoyancy (Mulder and Alexander, 2001). Mud flows typically run-out less than 100 km from their point of origin and deposit mudstones with or without dispersed sand grains (<50% of the bed volume), granules (<5% of the bed volume), and/or clasts (Mulder and Alexander, 2001). Debris flows, although similar to mud flows, generally have longer run-out distances (>100 km), and their deposits are more poorly sorted with gravel (>5% of the bed volume), sand grains (variable percentages of the bed volume), and larger clasts (up to boulder-sized) (Mulder and Alexander, 2001). As they move, cohesive flows can transform along their margins and/or down slope into non-cohesive flows via flow transformation. In order for this to take place, turbulence must develop so that electrostatic bonds between clay particles can be broken. Such changes can be enhanced by a hydraulic jump or a steepening of the slope, which serve to increase the entrainment and mixing of ambient fluid thereby reducing the bulk density of the flow and promoting turbulence production (Mulder and Alexander, 2001).

Unlike cohesive flows, frictional (non-cohesive) flows are composed dominantly of discrete particles that do not interact electrostatically, although some contain clay particles that impart (partial) cohesive properties to the flow. Mulder and Alexander (2001) defined three

kinds of friction flows: hyperconcentrated flows, concentrated flows, and turbidity currents.

Hyperconcentrated flows are made up of >25% sediment by volume (with low clay concentrations) and have similar grain-to-liquid ratios as cohesive flows, but they are transitional between Bingham plastics (i.e. fully cohesive flows) and Newtonian fluids (i.e. fully turbulent flows). The lower concentration of clay or more energetic grain-to-grain interactions in hyperconcentrated flows have been suggested as possible reasons why the flow does not become fully cohesive in spite of a similar grain-to-liquid ratio (Mulder and Alexander, 2001). The principal particle support mechanism in hyperconcentrated flows is grain-to-grain support (energy exchange produced by suspended grain collisions) with lower amounts of local matrix support. Deposition occurs as a succession of small-scale flow surges that build up gravel or gravelly sand deposits with little or no normal grading (Mulder and Alexander, 2001).

Concentrated flows, which contain 9-25% sediment by volume, typically behave as Newtonian fluids, however, they too can be transitional if the clay content is sufficient to produce a supportive matrix (Mulder and Alexander, 2001). The main particle support mechanism in concentrated flows is grain-to-grain interaction, with lesser matrix support and, importantly, fluid turbulence. Deposition takes place progressively by differential grain settling, eventually building up a massive or normally graded sand to gravel Bouma "A" division (1962; *Figure 10*) or Lowe S3 sequence (1982; *Figure 11*) with a highly erosive base (Mulder and Alexander, 2001). Lowe S3 divisions are typically deposited under capacity-driven conditions. Higher concentrations, and therefore more frequent grain-to-grain interactions, result in hindered settling and finer particles are often driven downward by high volumes of settling coarser sand, producing poorly graded or ungraded deposits. The typical changes in grain size and bedforms seen in a Lowe sequence are shown in *Figure 11*.



Grain Size	Lowe Division	Description
Gravel and sand	S3	Suspension - Massive or normally graded fine- to coarse-grained sandstone
	S2	Traction carpet - Reverse-graded fine- to coarse-grained sandstone
	S1	Traction - Sandstone or gravel conglomerate with parallel- or cross-lamination
Gravel	R3	Suspension
	R2	Traction carpet
	R1	Traction

Figure 11. Complete Lowe sequence showing sedimentary structures and grain-size typical of each Lowe division (redrawn from Lowe, 1982).

Finally, turbidity currents, as mentioned above, are turbulent, fully Newtonian fluid flows that contain <9% sediment by volume. Turbulence is the primary grain support mechanism, with additional support provided by grain-to-grain interaction. Deposition takes place under competence-driven conditions marked by gradual and selective differential settling of grains that produce Bouma sequences (classical turbidites) in which grain size does not exceed coarse sand. Turbulent flows typically have longer run-out distances compared to concentrated flows, which in turn are longer than hyperconcentrated flows. This is because the main particle support mechanism in both concentrated and hyperconcentrated flows is grain-to-grain interaction, which is a significant energy-consuming process. All frictional flows can transform laterally or down slope into either a more or less cohesive flow: (1) higher gradients accelerate the flow which

increases ambient water entrainment into the flow, producing less cohesive flows, whereas lower gradients decelerate the flow, damp turbulence and promote clay particle aggregation, producing more cohesive flows; (2) entrainment of water dilutes the flow and promotes turbulence production; and/or (3) erosion of the sea floor concentrates the flow with sediment and promotes cohesion if the sea floor is clay-rich, producing more cohesive flows.

Flows that exhibit both non-cohesive and cohesive properties are termed transitional flows, and include all high-density frictional flows ($>9\%$ sediment concentration by volume). In these flows, variable clay content imparts a weak to strong cohesive component to the flow (as mentioned above, cohesive strength depends on flow conditions, sediment concentration, and clay mineralogy) (Mulder and Alexander, 2001), which in turn controls the velocity and turbulence structure of the flows (Baas and Best, 2002; Baas et al., 2009; Baas et al., 2011; Sumner et al., 2009; and Talling, 2013). Experiments by Baas et al. (2009), expanding on the earlier work of Baas and Best (2002), showed that flows pass from turbulent to transitional and finally to fully laminar as clay concentration increases. Generally, flows with low clay content are turbulent with logarithmic velocity profiles and sediment is supported by turbulence. As clay content progressively increases, flows develop partial cohesive properties (i.e. they are ‘transitional’ between fully non-cohesive and fully cohesive) as clay particles begin to interact and form electrostatic bonds. Eventually, in the outer part of the flow where turbulence is lowest and therefore most easily attenuated by ongoing clay network production, a laminar plug begins to form. With increasing clay content, this outer plug expands downward in the flow. Eventually, clay concentration becomes so high that turbulence is suppressed throughout the entire thickness of the flow, and therefore the plug is the entire flow (excluding a very thin laminar shear layer at the base of the flow). It is at this point where the flow is considered to be fully laminar (or

cohesive). Using these experimental results, a suite of hypothetical transitional flow deposits has been proposed (Baas et al., 2009; Sumner et al., 2009; Talling, 2013; *Figure 12*). Of particular interest in this thesis are the mud-rich parts that develop in these deposits.

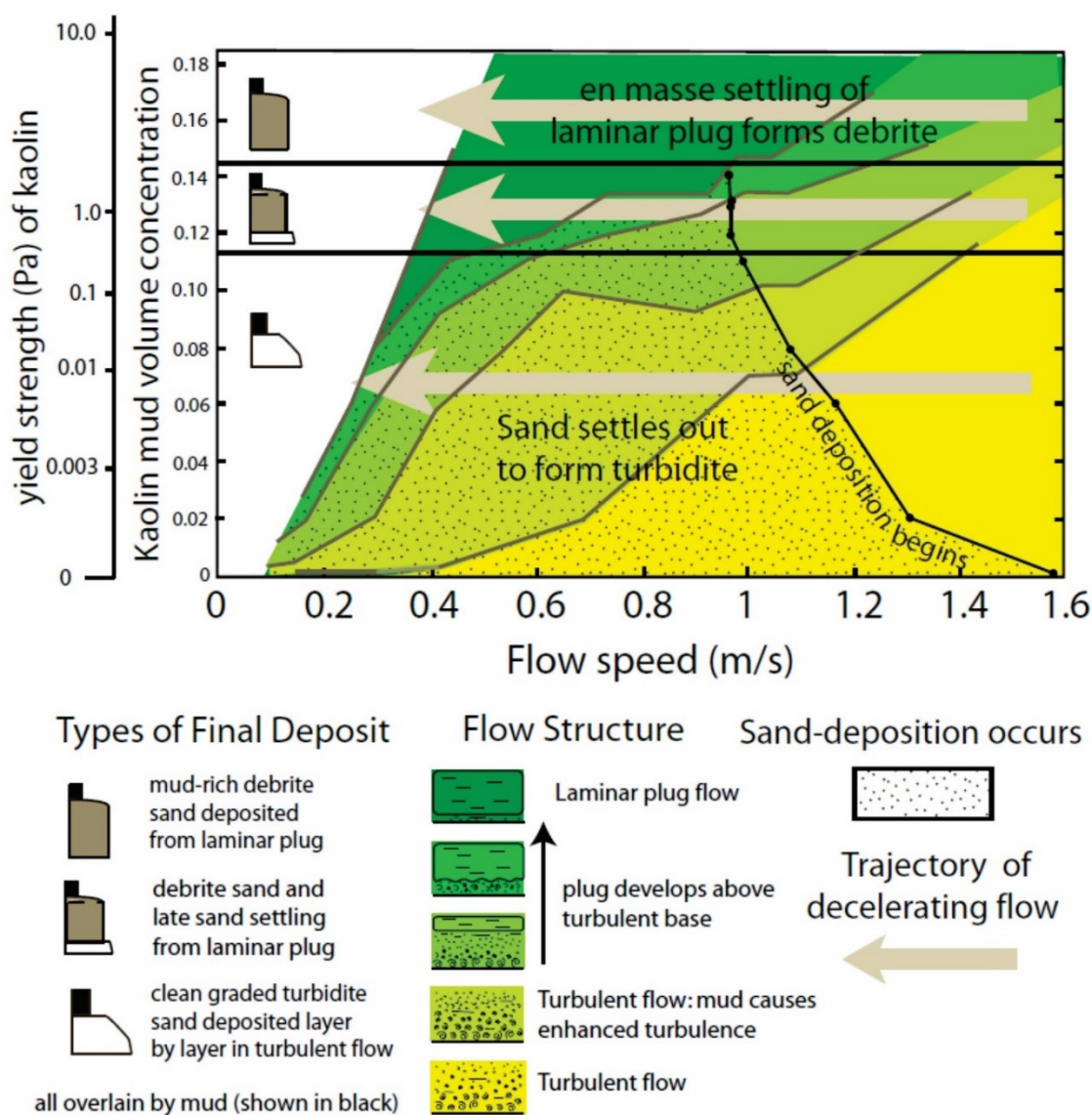


Figure 12. Phase diagram showing the relationships between flow speed, cohesive mud fraction, flow structure, and the type of final deposit. The figure combines results from recirculating flume experiments that constrained flow structure (Baas et al., 2009) and annular flume experiments that documented deposit types (Sumner et al., 2009). Both sets of experiments involved mixtures of kaolin and water, with sand added to the annular flume experiments (Sumner et al., 2009).

Decelerating flows follow trajectories from right to left (grey arrows), and are initially fully turbulent. Flow deceleration can lead to the development of a laminar plug. Mud content controls whether sand deposition occurs under turbulent or laminar plug flow conditions, and determines which of three different types of deposit are formed. At low mud content sand deposition occurs from turbulent flow and forms graded clean turbidite sand deposited in a grain by grain fashion. At high mud contents ($>14.25\%$ kaolin clay), en masse consolidation of a laminar plug forms ungraded mud-rich debrite sand. At intermediate concentrations late-stage sand settling occurs from the laminar plug forming a two-layer deposit (figure and caption from Talling, 2013).

3 Facies Descriptions and Interpretations

Macro- and micro-scale analysis of strata in both the upper Kaza Group and Isaac Formation have identified five lithofacies (*Table 1*):

Table 1. Summary of facies

Facies Number	Description	Depositional Interpretation
1	Coarse-tail graded or massive, fine- to medium-grained sandy claystone	Rapid suspension fallout from high-density, mud-rich turbidity currents
2	(2a) Coarse-tail graded or massive, fine- to medium-grained bipartite beds, and (2b) Coarse-tail graded or massive, fine- to coarse-grained clayey sandstone beds	(2a) Rapid suspension fallout from high-density, mixed sand/mud, stratified turbidity currents, and (2b) rapid suspension fallout from high-density, sand-rich turbidity currents
3	Coarse-tail graded or massive sandstone with finer-grained stratified tops	Rapid suspension fallout from high-density, sand-rich turbidity currents followed by a period of bypass and subsequent finer-grained deposition and traction transport
4	Coarse-tail graded or massive sandstone with traction structured tops	Rapid suspension fallout from high-density, sand-rich turbidity currents and in situ reworking of bed tops by the tailward part of the same depositional flow
5	Laminae and very thin- to thick-bedded sandstone, mudstone, and claystone	Differential grain settling from dilute turbidity currents – represent parts of the classic Bouma sequence

Each facies described below represents an end-member type of bed with its associated variations.

All facies, excluding Facies 3 and 4, are present in all of the studied stratigraphic sections. Facies 3 and 4 are observed only in the Kaza section (basin floor). Although this chapter describes all the lithofacies, Facies 1, 2, and the laminae and very thin- to thin-bedded turbidites of Facies 5 are described in particular detail.

3.1 Facies Terminology

Macroscopic and microscopic characteristics of each facies are outlined in this chapter. Macroscopic characteristics are observed at bed-scale (millimeter- to meter-scale), and include bed features that can be viewed with the naked eye and through a hand lens (>0.126 mm). Microscopic characteristics are observed at thin section scale, and include features as small as clay-size (0.0039 mm) grains or crystals. Additionally, field-based grain size measurements were confirmed in thin section.

Due to the low-grade (greenschist facies) metamorphic alteration of rocks in the study area, all primary detrital clay matrix has been recrystallized to muscovite and chlorite (*Figure 13 B*). Thus, the mineralogy and grain size of the primary clay matrix is unknown. In this thesis, all muscovite and chlorite crystals, regardless of their size, and quartz grains that are silt or finer (<0.0625 mm) are considered matrix. Grains between 0.0625 - 2 mm in their longest dimension are considered sand, and those >2 mm in length are considered granules. In this study, grain size is rarely coarser than granule and never more than pebble (~ 6 mm).

The classification scheme of Folk (1964) is used to classify bed types based on relative sand (0.0625 - 2 mm), silt (0.0039 - 0.0625 mm), and clay (<0.0039 mm) content (*Figure 14*). These contents were determined by point counting several (5 to 35) representative samples from each facies (as discussed in Chapter 1, Methodology). Resulting matrix (clay and silt) percentages were then used to group strata into three matrix categories: matrix-poor ($<10\%$ matrix), intermediate (10 - 50% matrix), and matrix-rich (50 - 90% matrix) (*Figure 13 C*). Matrix is typically composed of over 80% clay, therefore these three matrix categories correspond, respectively, with sandstone, clayey sandstone, and sandy claystone in the Folk (1964) classification scheme (*Figure 14*).

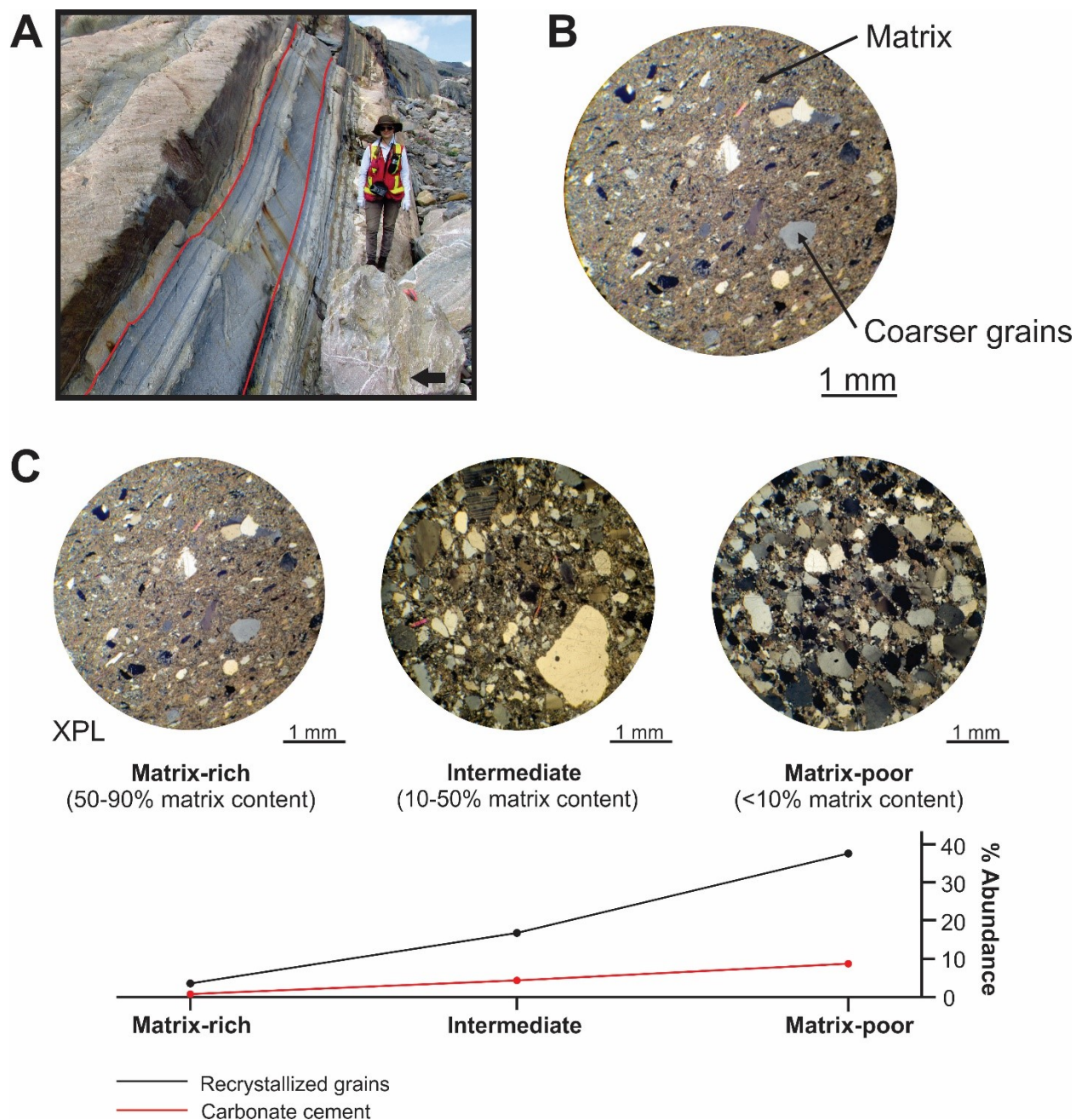


Figure 13. A) Example of a matrix-rich unit from the Isaac Formation (slope deposits); red lines indicate the base and top of the unit. B) Photomicrograph of a typical matrix-rich rock (~70% matrix content). Matrix is the finer-grained tan- or brown-coloured material in which coarser grains are dispersed. C) Photomicrographs of matrix-rich, intermediate, and matrix-poor rocks. Graph at bottom shows the percentage abundance of recrystallized grains and carbonate cement in each rock type; for recrystallized grains, the average percentage for each rock type was plotted, and for carbonate cement, the maximum percentage was plotted (see data below).

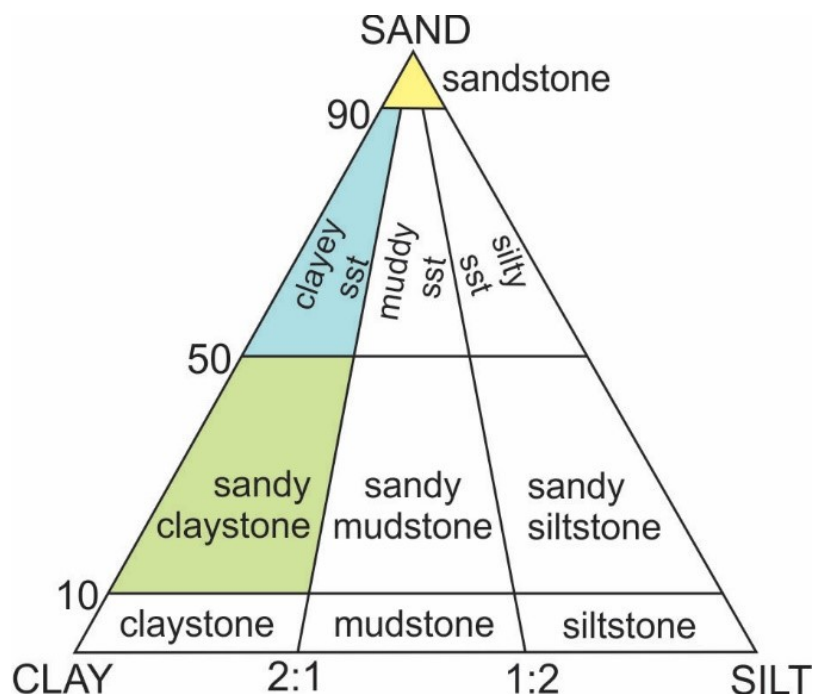


Figure 14. Folk (1964) classification scheme for siliciclastic sediments based on sand (>0.0625 mm), silt (0.0039 - 0.0625 mm), and clay (<0.0039 mm) content (redrawn from Tucker, 1981).

As a result of their low matrix content, and thus higher initial porosity, matrix-poor beds tend to contain more diagenetic carbonate cement (up to 9%) than intermediate or matrix-rich beds, which contain up to 4.5% and 0.5%, respectively (*Figure 13 C*). Additionally, recrystallized grain boundaries are more common in matrix-poor beds, typically making up 32-42% of the total thin section. In contrast, intermediate and matrix-rich beds contain 0.5-35% and 0-8.5%, respectively (*Figure 13 C*). When tectonized, the rigid, cemented framework of matrix-poor rocks caused stress to be relieved along contacts between adjacent grains, resulting in the formation of recrystallized grain boundaries. In matrix-rich beds, on the other hand, grain-grain contacts are rare or absent and tectonic stress was dissipated by mineralogical (clay mineral) slip.

3.2 Facies Field Recognition

In the field, matrix-rich beds are easily identified due to their darker colour (light to dark grey) and blue/green hue; characteristics imparted by high matrix and chlorite content, respectively (*Figure 13 A*). Beds with lower matrix contents, in contrast, tend to be tan coloured, with a red hue that is imparted by the iron-rich carbonate cement. Thus, where interstratified in the field, matrix-rich and matrix-poor beds can be easily differentiated. Where interstratified with laminae and/or very thin- to thin-bedded turbidites, matrix-rich beds become harder to distinguish due to their similar colour (light to dark grey) to mud-rich Td,e divisions. Yet, matrix-rich beds are characteristically sandy, whereas Td,e divisions lack sand.

3.3 Facies 1. Coarse-tail graded or massive, fine- to medium-grained sandy claystone

3.3.1 Facies 1. Macroscopic Characteristics

Facies 1 beds are composed of a basal medium to dark grey sandy claystone overlain abruptly by a matrix-poor structured unit and mudstone or claystone cap (*Figure 15*). Beds range from 2-25 cm, but most commonly are 5-15 cm thick (*Figure 16*). Basal contacts are usually sharp and planar, although uncommonly (8% of beds) they locally (15-30 cm laterally) undulate with 1-2 cm relief. The sandy basal part of Facies 1 beds makes up between 35-95% (1-15 cm) of the bed thickness (most commonly 65-90%), and typically consists of 50-80% matrix (most commonly 60-75%) with dispersed sand grains (*Figure 16*).

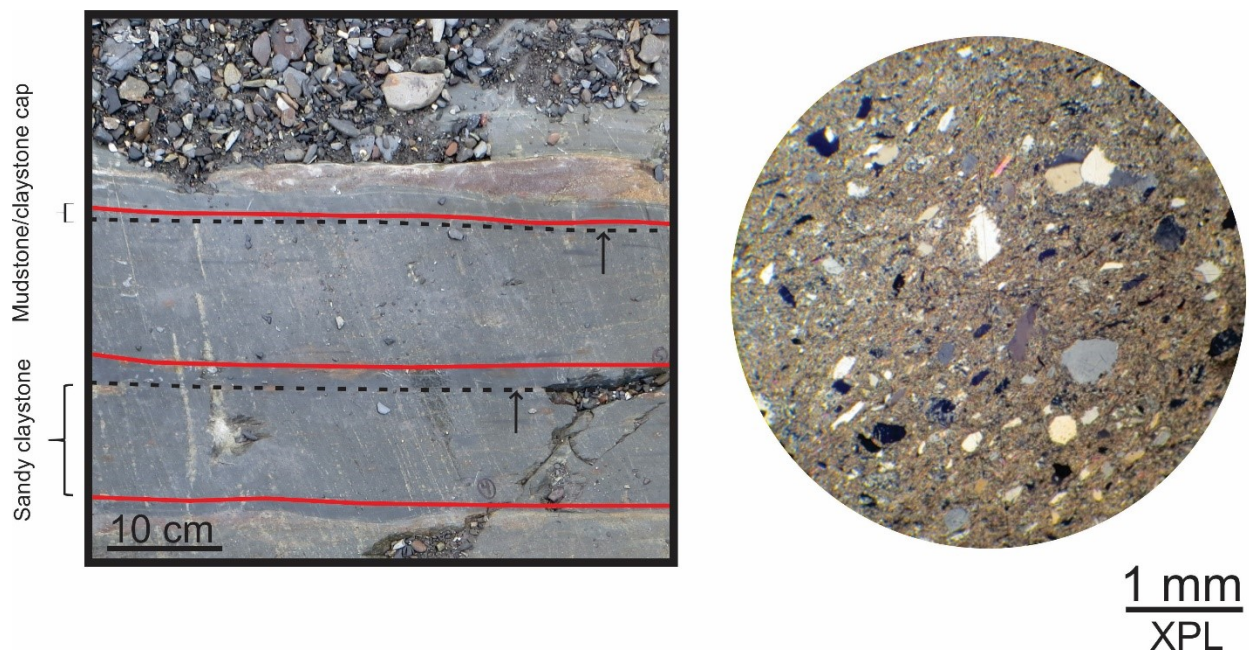


Figure 15. Photograph of two sandy claystone (Facies 1) beds (left). Note: the matrix-poor structured unit is present in both beds, however, is very thin (few mm) (indicated by arrows). Red lines indicate bed bases and black dashed lines indicate contacts between the sandy basal unit and fine-grained cap. Photomicrograph (in cross polarized light – XPL) of a matrix-rich rock (~70% matrix content) (right).

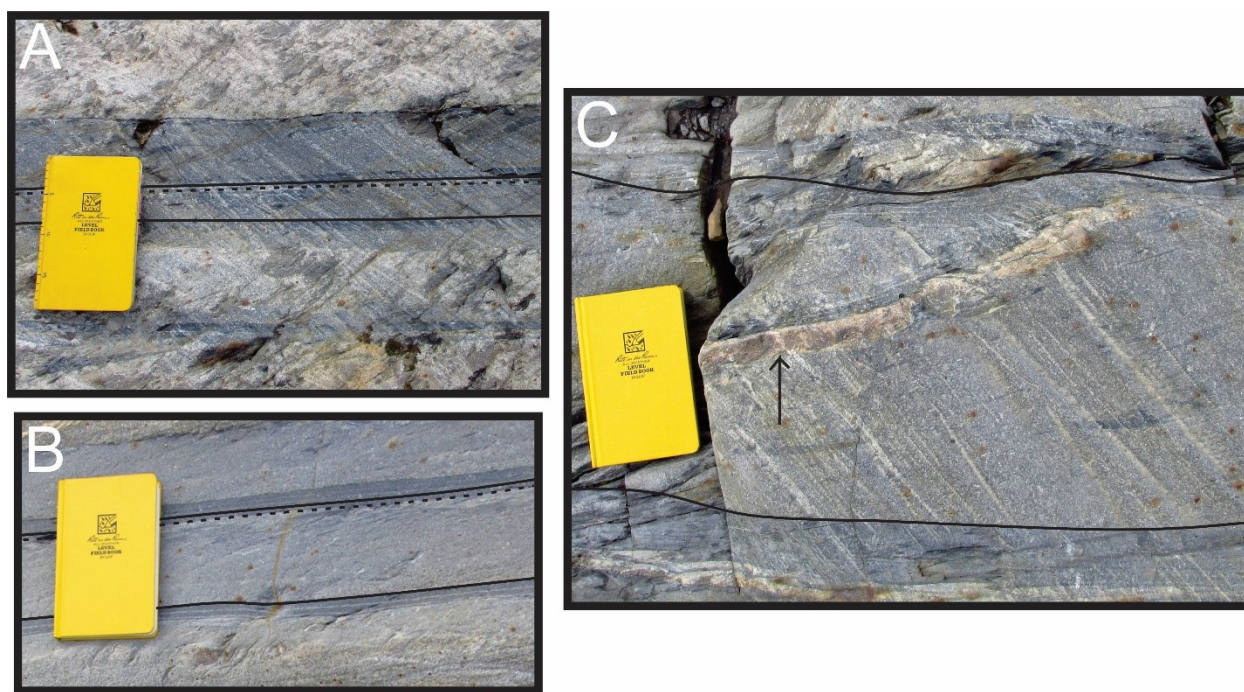


Figure 16. Thickness and matrix content variations in sandy claystone (Facies 1) beds. Beds with a higher matrix content are typically darker grey. Solid lines indicate bed bases and dashed lines

the contact between the sandy basal unit and fine-grained cap. Note: beds in A and B are overlain by a few very thin-bedded turbidites. In C the light-coloured sandstone clast (arrow) has foundered from the base of the overlying sandstone bed.

Based on the distribution of sand grains, approximately 55% of Facies 1 beds are massive. In these beds, sand grains consist mostly of lower fine to upper medium sand (with lesser amounts of lower to upper very fine, and lower coarse to upper very coarse sand). Less commonly, these grains are dominated by upper very fine sand. Coarse-tail graded beds make up the remaining 45% of Facies 1 beds. At the base of these beds, dispersed sand grains are typically lower fine to upper medium (with lesser amounts of lower to upper very fine, and lower coarse to upper very coarse sand). Beds fine slightly upward; typically, finer grains become minimally more abundant, but the grain size range remains consistent. At the top of these beds, grains are lower fine to lower medium sand (with lesser amounts of lower to upper very fine, and upper medium to lower very coarse sand). Rarely, in massive and coarse-tail graded strata, dispersed 2-4 mm granules are observed at bed bases. In addition, clasts, which are present in approximately 85% of Facies 1 beds, typically occur as elongate mudstone or claystone fragments although, uncommonly, clasts are planar laminated with silt, are deformed, or are composed of lower fine to upper medium sand (matrix-poor or matrix-rich). Clasts range from <1 cm thick and a few cm long to 3.5 cm thick and 20 cm long, and are oriented sub-parallel to bedding (*Figure 17 B*). Where present, clasts make up 5-30% of the bed volume and are either dispersed throughout the thickness of the sandy basal part of the bed or are preferentially concentrated in its upper half. Most commonly, beds that contain clasts show no relationship between clast size and their position in the bed, however, in a small number of beds, clasts are thicker and/or longer in the upper part of the sandy basal unit. In 95% of Facies 1 beds, a well-sorted, matrix-poor traction structured unit caps the sandy basal unit. The contact between the

sandy basal unit and structured unit is typically sharp and undulatory (0.1-1 cm locally). The sharp contact is the product of an abrupt change in sorting from the underlying matrix-rich basal unit. This unit makes up 1-25% of the bed thickness. In approximately 85% of these beds, the structured unit is 0.2-2.5 cm thick and is ripple cross-stratified (*Figure 17 C*). In the thinnest units (i.e. <0.5 cm), structures are more difficult to discern and therefore may be absent. In the remaining 15% of beds, the structured unit is 1-3 cm thick and is diffusely parallel- to wavy-laminated (*Figure 17 D*). Laminae are formed of thin (few mm) layers of highly concentrated matrix (approximately 80% matrix content) interstratified with thin layers of matrix-poor sand (<10% matrix content), which causes the strata to exhibit a distinctive dark-light striping. In both ripple cross-stratified and parallel- to wavy-laminated units, the modal grain size ranges from lower very fine to upper medium sand, which is typically equal to or slightly finer than the modal grain size at the top of the sandy basal unit, however, rarely the grain size of the structured unit is slightly coarser. Thinner traction structured units tend to be finer-grained than thicker units.

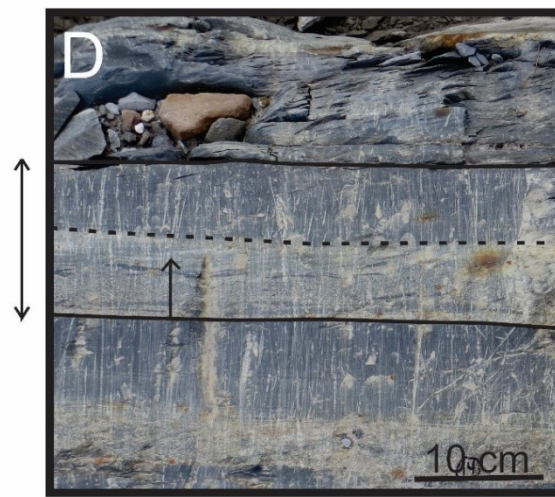
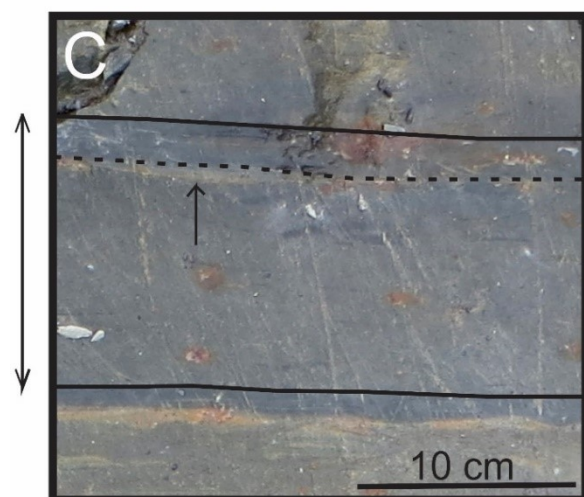
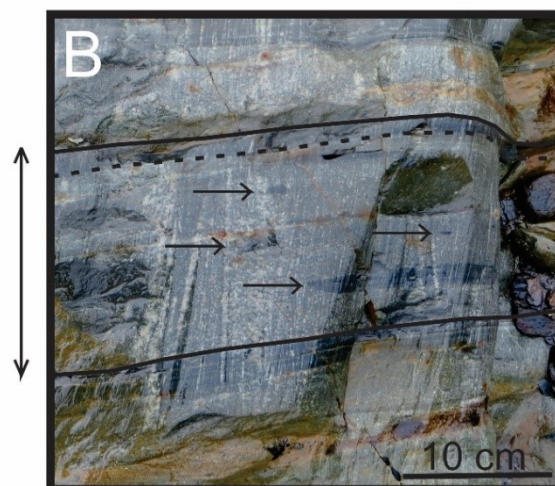
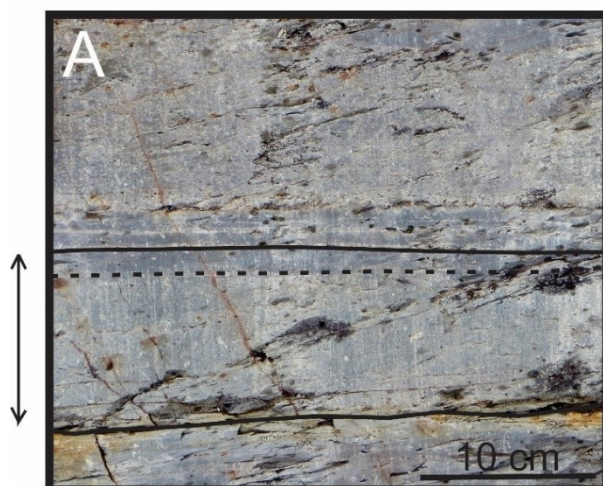
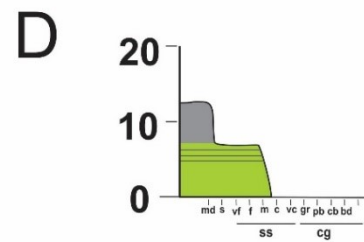
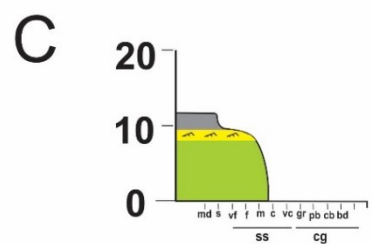
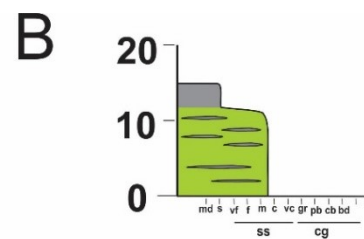
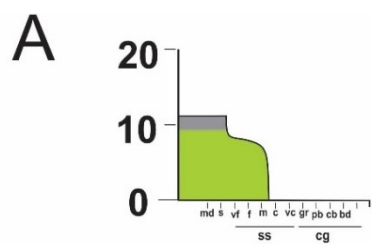


Figure 17. Illustrations of sandy claystone (Facies 1) beds (top) and representative photographs (bottom). In photographs, solid lines indicate bed bases and dashed lines the contact between the sandy basal unit and fine-grained cap. A) Bed lacking clasts (note that the cap is overlain by two very thin-bedded turbidites), B) Bed with elongate mudstone clasts (indicated by arrows), C) Sandy basal unit overlain abruptly by a well-sorted, discontinuous ripple form set (indicated by arrow), and D) Bed with diffuse planar laminated upper part (indicated by arrow) overlain abruptly by a mudstone cap. Note: the matrix-poor structured unit is present in all beds, but is very thin (few mm) in all but bed C.

All beds in Facies 1 have a medium to dark grey, fine-grained cap. Caps range from 0.2-5 cm thick. Typically, they abruptly overlie the sandy basal unit or traction structured unit. Where well-developed ripple cross-stratification occurs at the top of the bed, the cap fills the negative topography of the ripple form. Caps that gradationally overlie the basal unit occur rarely, and only above parallel- to wavy- laminated units. Most caps are composed of ungraded mudstone, however, some grade upward from mudstone to claystone. Rarely, caps of ungraded mudstone or claystone contain dispersed very fine sand grains.

3.3.2 Facies 1. Microscopic Characteristics

Medium to dark grey sandy claystone unit

Based on relative percentages of grains (quartz, feldspar, and lithic fragments) and matrix in the sandy basal unit, Facies 1 beds can be classified as sandy claystone (as in the sedimentary classification scheme of Folk (1964) - *Figure 14*). Beds in Facies 1 typically comprise 50-80% matrix (most commonly 60-75%), 20-40% dispersed sand (or coarser) grains, 0-10% recrystallized quartz grains, and 0-0.5% calcite cement (*Figure 18*). Beds with higher matrix contents tend to have fewer recrystallized quartz grains. Beds are poorly to very poorly sorted, and the dispersed sand grains are typically sub-angular to sub-round, equant to elongate, and sub-

spherical to non-spherical. At bed bases, sand grains are typically more closely spaced, and in some beds also shallowly (<1 mm) penetrate (i.e. load) into the underlying fine-grained cap.

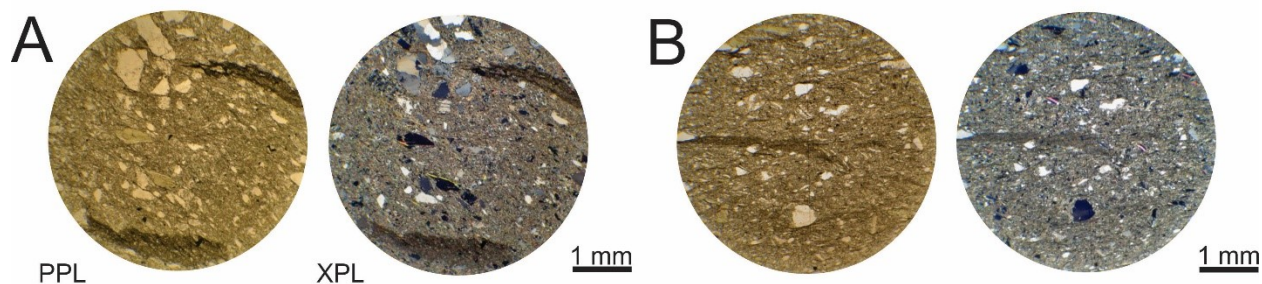


Figure 18. Photomicrographs in plane (PPL) and cross-polarized light (XPL) of sandy claystone (Facies 1) beds. A) 78% matrix content. B) 67% matrix content. Note the presence of mudstone clasts in both examples.

Grains are dominated by quartz (70-95%) that displays undulose extinction under cross-polarized light. Recrystallized quartz forms as <0.0625 mm grains along the boundaries of larger (>0.25 mm) quartz grains, and does not display undulose extinction under cross-polarized light. These grains are mostly the result of tectonic deformation and related bulging recrystallization. Bulging recrystallization is the process of quartz crystal regrowth that occurs under conditions of elevated stress and/or temperature, and reduces strain on quartz crystal lattices. It is typical of lower greenschist facies recrystallization, occurring at temperatures of 280-400°C (Stipp et al., 2002). During this process, the outer boundaries of quartz grains migrate and develop strain-free bulges that eventually separate from the primary grain (Stipp et al., 2002). The separated bulges appear as small quartz grains around the older, larger quartz grain (Stipp et al., 2002). Since these grains are a tectonic product, they are not considered primary grains. The other sand-sized components of Facies 1 beds include mudstone clasts (0-30%), feldspar (plagioclase) grains (0-10%), and euhedral to subhedral pyrite grains (0-7%). Mudstone and/or claystone clasts, in thin section, are typically 0.2-0.5 mm thick and 2-10 mm long, dark brown to black, ungraded,

undeformed, and are oriented parallel to bedding. Uncommonly, clasts contain trace amounts of quartz silt grains.

In Facies 1 beds, small (<0.0625 mm), inter-grown muscovite and chlorite crystals make up 80-97% of the matrix. Larger (>0.25 mm in diameter) muscovite and chlorite crystals make up 0-10% of the matrix. Where crystal size exceeds 0.25 mm, chlorite crystals often grow around muscovite crystals. At the macroscopic scale, the high abundance of chlorite imparts a distinctive green colour to the beds of this facies. Silt-sized (0.0039-0.0625 mm) quartz grains make up 3-20% (most commonly 15-20%) of the matrix.

Structured unit

Only thinner (<1 cm) structured units were sampled for thin section analysis (*Figure 19*). As noted above, thinner units tend to be finer grained, thus the modal grain sizes discussed here apply only to thinner (<1 cm) units. Structured units have the same mineralogical composition as the basal unit of a matrix-rich bed (see above – medium to dark grey sandy claystone unit). Typically, however, these structured units are better sorted (moderately to very well sorted) with $<10\%$ clay-sized (<0.0039 mm) muscovite and chlorite crystals. Based on relative percentages of grains and matrix, these units can be classified as sandstone (as in the sedimentary classification scheme of Folk, 1964). Approximately 80% of these units have a modal grain size of lower to upper very fine sand. The maximum grain size, which generally makes up less than 5% of the total number of grains, is typically lower medium to upper coarse sand or, less commonly, lower to upper fine sand or lower very coarse sand. The remaining 20% of structured units have a modal grain size of upper fine sand with a maximum grain size of upper coarse to lower very coarse sand.

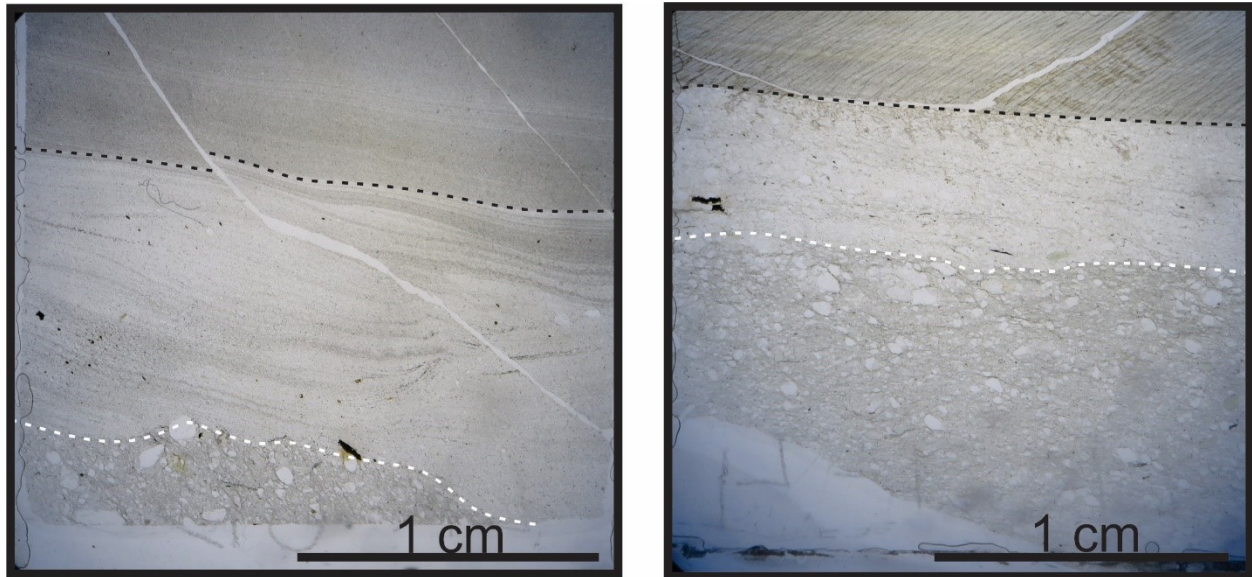


Figure 19. Thin section photographs showing the upper, matrix-poor structured unit. White dashed lines indicate the contact between the underlying sandy claystone unit and structured unit, and black dashed lines indicate the contact between the structured unit and overlying mudstone cap.

Cap

Within claystone caps of Facies 1 beds, minerals are aligned oblique (45°) to bedding, creating foliation. These foliations, otherwise known as crenulation cleavage, result from re-alignment of minerals during periods of metamorphic stress. This thesis, being focused on sedimentology, will not discuss these features further.

3.3.3 Facies 1. Interpretation

Facies 1 beds are interpreted to form via rapid suspension fallout from high-density, mud-rich turbidity currents. These beds are composed of 50-80% matrix (<0.0625 mm grains) with dispersed coarser grains that range up to upper very coarse sand (excluding the 2-4 mm granules

rarely present at bed bases). Most of these particle sizes can be suspended solely by fluid turbulence (i.e. turbidity current) and all by matrix-strength (i.e. debris flow). Intuitively, the high matrix content of Facies 1 beds suggests that they were deposited by debris flows, however, closer analysis of bed characteristics and lateral facies trends (see Chapter 4 Lateral and Vertical Facies Trends) indicates that these beds are deposited from flows that were at least partly, if not fully, turbulent and therefore are similar, but not identical, to slurry (mud-rich) beds described by Lowe and Guy (2000) and mud-rich sands of Sylvester and Lowe (2002).

Debrites have been variously described, and according to Shanmugam (1996) exhibit a sharp upper contact, a basal shear zone, rafted clasts (oriented either parallel or randomly to bedding), floating granules in a fine-grained matrix, inverse grading, and plastic deformation. Alternatively, Haughton et al. (2009) recognized debrites as muddy sand deposits with mudstone clasts, sand patches, injections, outsized granules, and shear fabrics. In spite of these differences, all debrites are interpreted to be deposited *en masse*, and therefore thought to reflect the lateral and vertical structure of a debris flow at the time of deposition. Although post-depositional grain settling can occur in less cohesive debrites, it is typically limited when matrix strength (the primary sediment support mechanism of debris flows; produced by networks of clay particles) is high (Mulder and Alexander, 2001), especially in true “quasi-laminar plug flows” (Baas et al., 2009) (see Chapter 2 Deep-Water Physical Sedimentation Processes). True debris flows are laminar (i.e. turbulence is absent), and therefore incapable of forming normally or coarse-tail graded beds; both consequences of differential grain settling in waning, and at least partially turbulent flows (Leclair and Arnott, 2003). It should be noted that, although not studied in this thesis, debrites have been recognized throughout the Castle Creek study area. Here, they are typically a few to in excess of a few decameters thick, ungraded, poorly sorted, and composed of

a chaotic, heterogeneous silty mudstone or sandstone matrix with deformed/sheared or undeformed clasts of various lithologies (quartz sand- to pebble-sized grains, mudstone, sandstone, and carbonate) and sizes (up to 54 m long and 4 m thick) (Arnott et al., 2011). Castle Creek debrites can be correlated laterally up to a few km along strike, whereas Facies 1 beds are laterally continuous over only a few hundred meters (maximum) before they transition into a different facies (see Chapter 4 Lateral and Vertical Facies Trends).

In Facies 1 beds, uncommon locally undulating bed bases (1-2 cm relief over 15-30 cm laterally) reflect local sea-floor scour by turbulent near-bed eddies or, alternatively, gravitational loading of an under-consolidated sea floor, which most probably is a consequence of high sedimentation rates. Planar bases indicate turbulence suppression near the bed. Facies 1 beds are either coarse-tail graded or massive, suggesting that they were deposited by rapid suspension fallout from high-density turbidity currents (with >20% sediment concentration, according to Sumner et al. (2008); see also Leclair and Arnott, 2003; Sumner et al., 2012). When bed aggradation rates and the near-bed concentration of sediment are both high, particles of different sizes cannot segregate due to hindered settling and grain hyper-concentration effects, resulting in the deposition of poorly sorted beds (Banerjee, 1977; Kneller and Branney, 1995; Amy et al., 2006; Sumner et al., 2008; Dorrell et al., 2011). In coarse-tail graded beds, the progressive improvement in sediment sorting reflects a temporal decrease in sedimentation rates (Arnott et al., 2011). Massive beds, therefore, were likely deposited from flows that were either more concentrated with suspended sediment (preventing differential grain settling) and/or decelerated more rapidly resulting in higher rates of sedimentation. In addition to its effect on the textural character of the deposit, rapid depositional rates also suppressed bedform development, hence

the structureless nature of Facies 1 beds (Arnott and Hand, 1989; Leclair and Arnott, 2005; Sumner et al., 2008).

Clasts in Facies 1 beds range up to a few cm thick and 20 cm long, and are typically composed of either mudstone or claystone, some with interlaminated silt. The fact that some clasts preserve their original interlaminated fabric implies that flows were partly cohesive (i.e. turbulence was at least slightly dampened), and that erosion was local (i.e. negligible transport), thus limiting clast disintegration (Smith, 1972). Sandstone (matrix-poor and matrix-rich) clasts, although uncommon, were probably sourced from underlying sand-rich strata or were foundered from overlying strata. Where clasts are dispersed evenly throughout the thickness of Facies 1 beds, depositing flows were probably not longitudinally or vertically stratified. However, beds where clasts are preferentially concentrated in the upper half indicate that the flows were either longitudinally or vertically stratified. If longitudinally stratified, clasts would have been carried in the middle or rearward part of the turbidity current and deposited on clast-free sediment that had been deposited by the front of the current (Kneller and Branney, 1995). Alternatively, in vertically stratified flows, positively buoyant clasts were carried in the upper, more dilute part of the flow that was separated from the lower, more concentrated part by a rheological interface (Postma et al., 1988).

The thin, matrix-poor, ripple cross-stratified units of Facies 1 beds often have bases that scour the underlying matrix-rich part of the bed, suggesting that they formed via reworking of previously deposited sediment by overriding, unidirectional, turbulent flows. The grain size of the ripple cross-stratification is either the same or slightly finer than the underlying matrix-rich portion of the bed, which is consistent with in situ reworking that winnowed fine sediment and formed a sand-rich near-bed transport layer which, in the presence of bed defects, formed current

ripples. The lateral discontinuity of Facies 1 ripple cross-stratified units, however, is likely the result of post-depositional erosion.

Similar to the formation of ripple cross-stratified bed tops, the sharp-based, cleaner, parallel- or wavy-laminated tops of Facies 1 beds, which are typically similar in grain size to the underlying matrix-rich part of the bed, reflect in situ reworking by an overriding turbulent flow. These structures are interpreted to be formed in the tractional near-bed boundary zone of low-density turbidity currents. In these flows, laminations form by the migration and burial of long-wavelength, low-amplitude bed waves (Best and Bridge, 1992; Sumner et al., 2008). The diffuse nature of the parallel- or wavy-lamination in Facies 1 beds may result from the interaction/alternation of deposition of sediment carried at or near the bed and sediment being deposited from suspension (Leclair and Arnott, 2003).

Massive or normally graded claystone and mudstone caps are interpreted to have been formed via settling and compaction of clay and silt from a dilute upper flow, which led to the formation of a fluid mud layer. Eventually the fluid mud layer froze and deposited. Alternatively, these caps may have formed during periods of quiescence by suspension fallout from the low energy, dilute, turbulent tails of flows. Hemipelagic fallout would have also likely played a role in forming these caps.

3.4 Facies 2a. Coarse-tail graded or massive, fine- to medium-grained bipartite beds, and Facies 2b. Coarse-tail graded or massive, fine- to coarse-grained clayey sandstone beds

3.4.1 Facies 2a. Macroscopic Characteristics

Facies 2a beds are composed of a basal light to medium grey clayey sandstone unit abruptly overlain by a medium to dark grey clayey sandstone or sandy claystone (herein referred to as sandy claystone) capped by a matrix-poor structured unit and mudstone or claystone (*Figure 20*). Beds range from 3-45 cm thick, but typically are 10-25 cm. Basal contacts are sharp, and either undulate (60% of beds) with 1-5 cm relief over 15-150 cm laterally, or planar (30% of beds). Some beds (10%) have 1-4 cm high and 1-30 cm long flame structures at their bases.

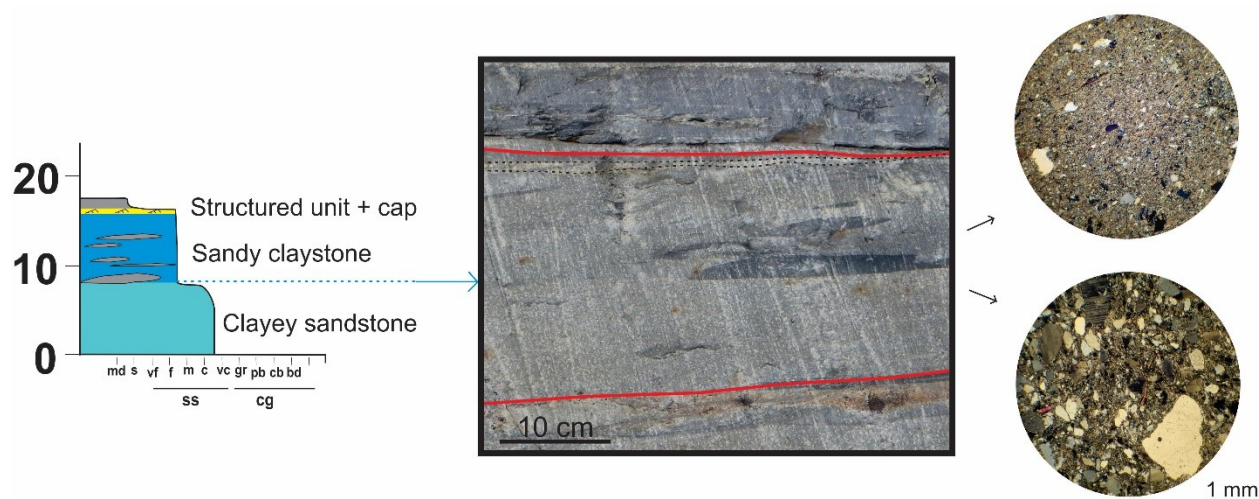


Figure 20. Illustration (left) and photograph (middle) of a bipartite (Facies 2a) bed, and cross-polarized photomicrographs of the basal and upper unit (right). Dashed blue line and arrow indicate the contact between the basal and upper unit. Red lines indicate bed bases, and black dashed lines indicate the base and top of the matrix-poor structured unit.

The basal light to medium grey clayey sandstone makes up between 5-95% (0.2-43 cm) of the total bed thickness. In beds that are more than 15 cm thick, it typically makes up 50-95%

(8-43 cm) whereas, in beds less than 15 cm thick, it typically makes up 5-50% (0.2-8 cm). Exceptions to this general trend are uncommon. These clayey sandstone units are generally composed of 10-45% matrix and coarser grains (>0.0625 mm). Based on the distribution of coarser grains, approximately 60% are coarse-tail graded, and the remaining 40% are massive. At the base of coarse-tail graded units, grains range from upper very fine to upper very coarse sand, but most commonly are lower fine to upper medium with or without dispersed lower coarse to upper very coarse sand and/or 2-4 mm granules. Beds fine slightly upward; typically, finer grains become minimally more abundant, but the grain size range remains consistent. At the top of these beds, grains are most commonly lower fine to lower medium with or without dispersed upper medium to upper very coarse sand and/or 2-3 mm granules. In massive units, grains are typically lower fine to upper medium with or without dispersed lower coarse to upper very coarse sand and/or 2-3 mm granules. In addition, clasts, which are present in approximately 20% of the basal light to medium grey units, typically occur as elongate, mudstone or claystone fragments that are oriented sub-parallel to bedding. Uncommonly, clasts are composed of lower fine to upper medium sandstone (matrix-poor or matrix-rich). Clasts range in size from 0.5-5 cm thick and 3-20 cm long, but are typically 0.5-2 cm thick and 3-10 cm long. Where present, they make up less than 5% of the total unit volume. Typically, they are concentrated mid-unit or at, or slightly below, the contact between the basal light grey unit and the overlying medium to dark grey unit. Few Facies 2a beds contain clasts that concentrate at the base of the basal unit. Beds that exceed 15 cm in thickness more commonly have clasts in the basal light grey unit compared to thinner beds.

In Facies 2a beds, abruptly overlying the basal light to medium grey clayey sandstone unit is a planar-based medium to dark grey sandy claystone unit. In 30% of beds, the contact

between the two units is marked by a zone of abundant mudstone or claystone clasts. This only occurs in beds that are thinner than 15 cm. Clasts range in size from <0.1-5 cm thick and 2-45 cm long, but are typically 0.1-2 cm thick and 2-10 cm long. Sandy claystone units make up between 2-90% (0.5-10 cm) of the total bed thickness. Where total bed thickness is more than 15 cm, these units make up between 2-35% (0.5-7 cm), compared to 35-90% (1-10 cm) in beds less than 15 cm. Again, exceptions to this general trend are uncommon. These units contain approximately 5-30% more matrix than the underlying light grey unit. Where the upper unit contains only slightly more matrix (i.e. 5%) than the basal unit, the two units are very similar in colour (*Figure 21*). Where the upper unit contains much more matrix (i.e. 30%) than the basal unit, the two units have a greater colour contrast, thus the contact between the two units is easier to identify. Typically, beds that are thinner than 15 cm, where the upper unit makes up a larger percentage of the total bed thickness, contain units more similar in colour. Upper units contain between 20-65% matrix, but most commonly are composed of 30-65% matrix with dispersed coarser grains (>0.0625 mm). The modal grain size at the base of the upper unit is either equal to or finer than the grain size at the top of the basal unit. Based on the distribution of coarser grains, approximately 60% of these medium to dark grey units are massive, and the remaining 40% are coarse-tail graded. Massive units are typically composed of lower fine to upper medium sand (with lesser amounts of lower to upper very fine sand) with or without dispersed lower coarse to upper very coarse sand. Where units are coarse-tail graded their bases generally contain lower fine to upper medium sand (with lesser amounts of lower to upper very fine, and lower to upper coarse sand) with or without dispersed lower to upper very coarse sand and rare 2-3 mm granules. Beds fine slightly upward; typically, finer grains become minimally more abundant, but the grain size range remains consistent. At the top of these beds, grains are lower fine to

lower medium sand (with lesser amounts of lower to upper very fine, and upper medium to lower coarse sand) with or without dispersed upper coarse and lower to upper very coarse sand and rare 2 mm granules. Typically, coarse-tail graded basal units are overlain by either massive or coarse-tail graded upper units, however, where the basal unit is massive the upper unit is also typically massive. In addition, clasts are present in approximately 40% of medium to dark grey upper units, and are most common within thicker units (i.e. in beds less than 15 cm thick) (*Figure 22 A*). Typically, clasts occur as elongate, mudstone or claystone fragments that are oriented parallel to bedding. Uncommonly, clasts are composed of upper fine to lower coarse sandstone. Rarely clasts are deformed. Clasts range in size from <0.1-5 cm thick and 0.5-47 cm long, but most commonly are 0.1-1.5 cm thick and 5-10 cm long. Where present, they make up 5-15% of the unit volume and are usually dispersed evenly throughout the unit. A matrix-poor traction structured unit caps the upper sandy claystone unit in 75% of beds. The contact between the sandy claystone unit and structured unit is typically sharp and undulatory (0.1-1 cm locally). The sharp contact is the product of an abrupt change in sorting from the underlying matrix-rich unit. The structured unit ranges from 0.2-5 cm thick, and makes up 1-50% of the total bed thickness. Most commonly, it is 0.2-2 cm thick, and makes up 5-15% of the total bed thickness. Approximately 95% of these structured units are ripple cross-stratified, lower very fine to upper medium sandstone (*Figure 22 B*). Uncommonly, these units contain lower coarse to upper very coarse sand grains at their bases. Generally, one or two ripple sets will be visible; thinner ripples are generally finer grained. In the thinnest units (i.e. <0.5 cm), structures are more difficult to discern and therefore may be absent. The remaining 5% of structured units are diffusely parallel-to wavy-laminated, upper very fine to lower fine sandstone with uncommon dispersed lower coarse to upper very coarse sand at their bases. Laminae are similar to those described in Facies

1. Where present, traction structures have a similar grain size to that in the underlying unit, or are slightly finer grained.

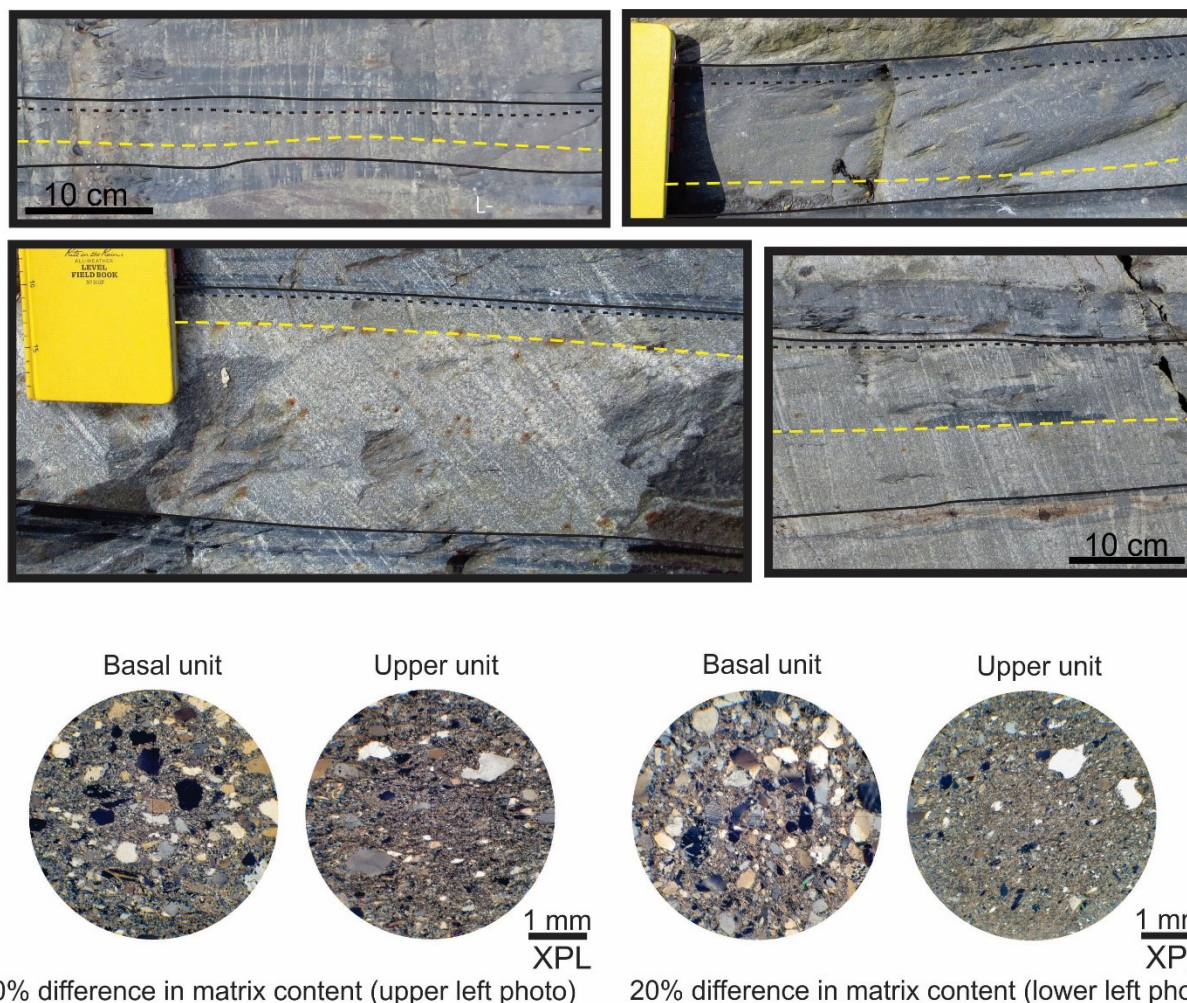


Figure 21. Thickness and matrix content variations in bipartite (Facies 2a) beds. Note that the basal and upper units of bipartite beds make up varying percentages of total bed thickness. In addition, as noted above, matrix content increases by 5-30% from the basal to upper unit. A larger increase in matrix content results in greater colour contrast between the two units (best illustrated in the lower left photograph). This bed-scale observation has been verified through thin section analysis.

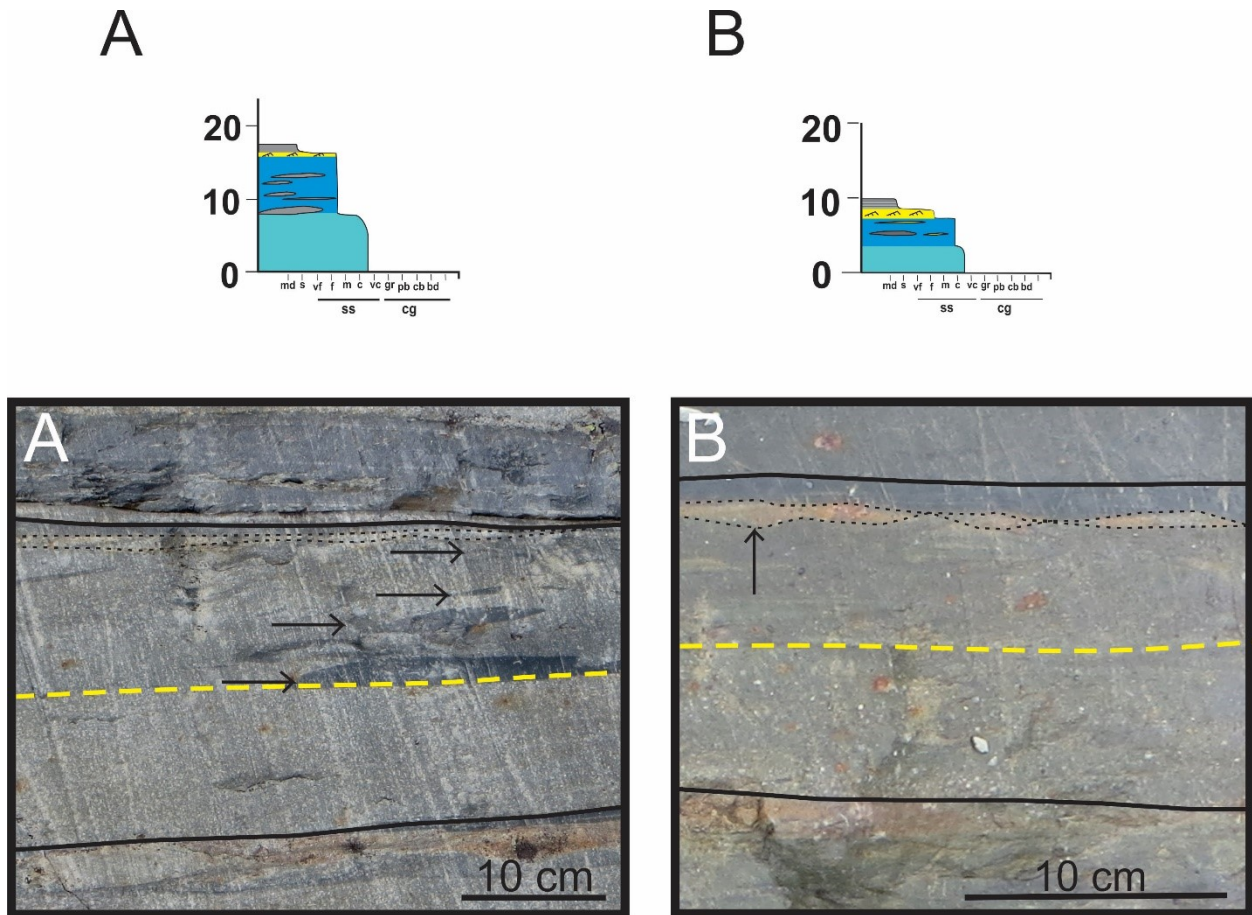


Figure 22. Illustrations of bipartite (Facies 2a) beds (top) and representative photographs (bottom). In photographs, solid black lines indicate bed bases, dashed yellow lines indicate the contact between the basal clayey sandstone unit and upper sandy claystone unit, and the dashed black lines indicate the base and top of the structured unit. A) Bed with elongate mudstone clasts (indicated by arrows) concentrated in the upper sandy claystone part of the bed, B) Sandy claystone unit overlain by a well-sorted, discontinuous ripple form set (indicated by arrow).

The mudstone or claystone caps of Facies 2a beds make up 2-55% of the total bed thickness, but generally only represent approximately 5-20%. Their thickness ranges from 0.1-6 cm, but usually is 0.5-3 cm. Typically, they abruptly overlie the upper sandy claystone unit or traction structured unit. Where overlying well-developed ripple cross-stratification, the cap fills the negative topography of the ripple form. Caps that gradationally overlie the basal unit occur

rarely, and only occur above parallel- to wavy- laminated units. Most caps are composed of ungraded mudstone, however, some grade upward from mudstone to claystone.

3.4.2 Facies 2a. Microscopic Characteristics

Within Facies 2a beds, the basal light to medium grey clayey sandstone and the overlying medium to dark grey sandy claystone differ by 5-30% in matrix content, and therefore are described separately in this section.

Light to medium grey clayey sandstone unit

Light to medium grey basal units of Facies 2a beds typically comprise 45-65% dispersed sand grains, 10-45% matrix, 0.5-35% recrystallized quartz grains, and 0-4.5% calcite and/or dolomite cement (dolomite cement is much less common), and according to the classification of Folk (1974), are classified as clayey sandstone. Beds are poorly to very poorly sorted and dispersed sand grains are typically sub-angular to sub-round, equant to elongate, and sub-spherical to non-spherical. Beds with lower matrix percentages typically contain higher percentages of recrystallized quartz grains and calcite and/or dolomite cement.

Grains are dominated by quartz (88-100%) with undulose extinction. Unstrained, recrystallized quartz, which is the result of bulging recrystallization (see section 3.3.2 Facies 1. Microscopic Characteristics), occurs as <0.0625 mm grains along the boundaries of larger (>0.25 mm) quartz grains. Other sand-sized or coarser components include feldspar (plagioclase) (0-8%), euhedral to subhedral pyrite (0-8%), and mudstone or claystone clasts (0-0.5%).

In the light to medium grey basal unit of Facies 2a beds, small (<0.0625 mm), inter-grown muscovite and chlorite crystals make up 80-100% of the matrix, and larger (>0.25 mm in

diameter) muscovite and chlorite crystals make up 0-20%. Finally, 0-17% of the matrix is composed quartz silt grains.

Medium to dark grey clayey sandstone or sandy claystone unit

Based on percentages of grains (quartz, feldspar, and lithic fragments) and matrix, the middle part of Facies 2a beds can be classified as either clayey sandstone (where matrix 10-50%) or sandy claystone (where matrix 50-90%). This unit contains between 20-65% matrix, but most commonly is composed of 30-65% matrix, 28-65% dispersed sand grains, 0-25% recrystallized quartz grains, and 0-3% calcite and/or dolomite cement (dolomite cement is much less common). Recrystallized grains are mostly the result of bulging recrystallization (see section 3.3.2 Facies 1. Microscopic Characteristics). Beds are poorly to very poorly sorted and dispersed sand grains are typically sub-angular to sub-round, equant to elongate, and sub-spherical to non-spherical. Although much the same as the basal clayey sandstone units of Facies 2a beds, the dispersed sand (or coarser) grains of sandy claystone units are 70-100% quartz, 0-27% mudstone or claystone clasts, 0-7% euhedral to subhedral pyrite, and 0-6% feldspar (plagioclase). Beds with lower matrix percentages typically contain higher percentages of recrystallized quartz grains and calcite and/or dolomite cement.

Matrix is composed of 75-100% small (<0.0625 mm), inter-grown muscovite and chlorite crystals, and 1-11% larger (>0.25 mm in diameter) muscovite and chlorite crystals. Finally, 0-20% of the matrix is composed of silt-sized (<0.0625 mm) quartz grains.

Structured unit

Thinner (<1 cm) structured units are very similar between Facies 1 and Facies 2a; please see Facies 1. Coarse-tail graded or massive, fine- to medium-grained sandy claystone (structured unit) for information.

Cap

In some claystone caps, muscovite and chlorite crystals are oriented oblique (45°) to bedding, creating foliation (see section 3.3.2 Facies 1. Microscopic Characteristics for discussion).

3.4.3 Facies 2b. Macroscopic Characteristics

Facies 2b beds are lithologically similar to Facies 2a beds, however, they lack the upper medium to dark grey, more matrix-rich clayey sandstone or sandy claystone unit (*Figure 23*). Beds range from 25-55 cm thick. Basal contacts are sharp, and either undulate (70% of beds) with 1-7 cm relief over 30-200 cm laterally, or planar (30% of beds).

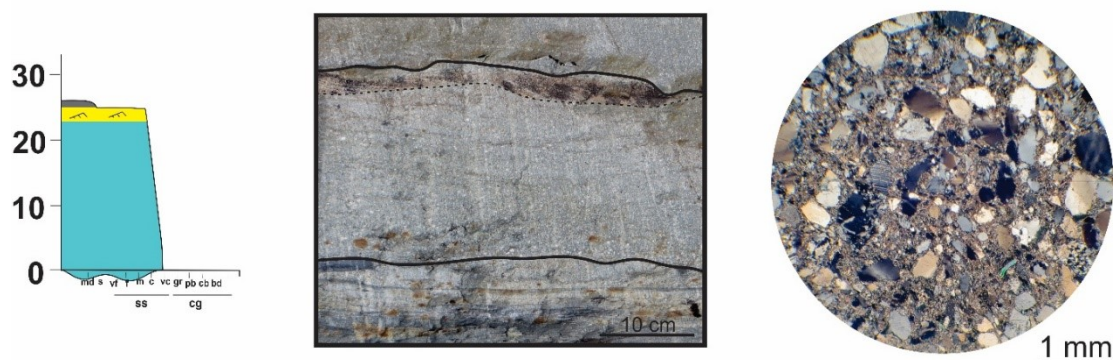


Figure 23. Illustration (left), photograph (middle), and photomicrograph of a clayey sandstone (Facies 2b) bed. Black lines indicate bed bases, and black dashed lines indicate the base of the matrix-poor structured unit. Note that the cap has been eroded.

The basal part of the bed, excluding the upper structured unit and mudstone or claystone cap, contains 10-45% matrix and is typically 16-52 cm thick. Based on the distribution of sand-grade or coarser grains, approximately 85% of Facies 2b beds are coarse-tail graded. At the base of these beds, grain size is lower fine to upper coarse sand with dispersed lower to upper very coarse sand and 2-4 mm granules. Less commonly, 2-4 mm granules are abundant in the basal 5-10 cm of the bed, forming granule conglomerate. Upward, grain size fines slightly to lower fine to upper medium sand with or without dispersed lower coarse to upper very coarse sand. The remaining 15% of Facies 2b beds are massive and contain lower fine to upper coarse sand with dispersed lower to upper very coarse sand and, in some beds, 2-4 mm granules. Clasts are present in 8% of Facies 2b beds, and typically occur as elongate, mudstone or claystone fragments that are oriented sub-parallel to bedding. Uncommonly, clasts are composed of lower fine to upper medium sandstone (matrix-poor or matrix-rich). Clasts range from 1-5 cm thick and 3-20 cm long and typically occur mid-bed or at the base of the bed. Where present, clasts make up <1% of the total bed volume. Where Facies 2b bed tops are not scoured by the overlying bed (i.e. in 65% of Facies 2b beds), a well-sorted, matrix-poor traction structured unit caps the bed. This unit makes up 2-25% of the bed thickness, and ranges in thickness from 1-8 cm. Most commonly, it is ripple cross-stratified and composed of upper very fine to upper medium sand, with less common lower coarse and upper coarse sand. Dispersed at the bases of some units are lower to upper very coarse sand and rare 2 mm granules. Generally, the unit consists of one or two ripple sets, which are generally thinner in finer grained strata. The grain size of the structured unit is typically the same or slightly finer than the underlying structureless unit (described above).

The upper part of Facies 2b beds is commonly scoured by the base of the overlying bed. Therefore, Facies 2b beds are rarely capped by a medium to dark grey, fine-grained unit composed of ungraded mudstone. Where preserved caps are typically very thin (<1 cm) and abruptly overlie the sandy basal unit or traction structured unit.

3.4.4 Facies 2b. Microscopic Characteristics

Facies 2b beds typically comprise 45-65% dispersed sand grains, 10-45% matrix, 0.5-35% recrystallized quartz grains, and 0-4.5% calcite and/or dolomite cement (dolomite cement is much less common), and according to the classification of Folk (1974), are classified as clayey sandstone. Beds are poorly to very poorly sorted and dispersed sand grains are typically sub-angular to sub-round, equant to elongate, and sub-spherical to non-spherical. Beds with lower matrix percentages typically contain higher percentages of recrystallized quartz grains and calcite and/or dolomite cement.

Grains are dominated by quartz (90-97%) with undulose extinction. Unstrained, recrystallized quartz, which is the result of bulging recrystallization (see section 3.3.2 Facies 1. Microscopic Characteristics), occurs as <0.0625 mm grains along the boundaries of larger (>0.25 mm) quartz grains. Other sand-sized or coarser components include feldspar (plagioclase) (2-10%), euhedral to subhedral pyrite (0-4%), and mudstone or claystone clasts (0-0.5%).

Small (<0.0625 mm), inter-grown muscovite and chlorite crystals make up 83-92% of the matrix. Larger (>0.25 mm in diameter) muscovite and chlorite crystals make up 3-12% of the matrix. Finally, 4-8% of the matrix is composed quartz silt grains.

3.4.5 Facies 2a. Interpretation

Facies 2a beds are interpreted to form via rapid suspension fallout from high-density, mixed sand/mud, stratified turbidity currents. The generally undulatory contact of Facies 2a beds indicates that the flow was at least partially turbulent at its base. Energetic, down-sweeping turbulent eddies would have interacted with and scoured the sea floor by 1-5 cm over 15-150 cm laterally. Where bed bases are sharp and planar, the flow was at least partially cohesive, suppressing turbulent eddies from interacting with the bed. The occurrence of uncommon 1-4 cm high and 1-30 cm long flame structures at the base of some beds indicates gravitational loading of an under-consolidated sea floor, which most probably is a consequence of high sedimentation rates.

Excluding the matrix-poor traction structured unit and mudstone/claystone cap, beds of Facies 2a comprise two sharply bounded parts (bipartite) marked by an abrupt upward increase (5-20%) in matrix content, which in the field manifests itself by darker coloured strata. Although more matrix-rich, the upper unit has a similar range of sand-grade particles compared to the basal unit. In addition, the coarsest sand grains in the upper unit are similar or finer than those in the associated basal unit, but only less abundant. Facies 2a basal units are typically coarse-tail graded, and upper units are either massive or coarse-tail graded. These features, plus the sharp, planar contact between the lower and upper parts suggest that both units were deposited by the same flow. Coarse-tail graded (60% of Facies 2a beds) and massive (40% of Facies 2a beds) basal units are interpreted to form in a similar manner to Facies 1 coarse-tail graded and massive beds, but in this case the depositing flows were much more concentrated with sand-grade particles, and as a consequence were denser and deposited more rapidly. The upper part of Facies 2a beds is lithologically analogous to Facies 1 beds, and therefore is similarly interpreted (see

section 3.3.3 Facies 1. Interpretation). This also includes the upper, cleaner, ripple cross-stratified or planar- to wavy-laminated tops of Facies 2a beds, and the mudstone/claystone cap.

The distinctive bipartite make up of Facies 2a beds may reflect deposition from vertically or longitudinally stratified flows (see below), which are known to deposit beds composed of two vertically stacked parts marked by an abrupt grain size change across the contact (Gladstone and Sparks, 2002; Kneller and McCaffrey, 2003). According to Gladstone et al. (2004), sediment laden flows become naturally stratified due to (1) variable settling velocities of grains, (2) differences in turbulence through the flow thickness, (3) entrainment of ambient water at the head and margins of the flow, and (4) entrainment of sediment from the bed. Stratified flows can also be triggered by changes in sea floor topography, hydraulic jumps, and obstacles to flow (Kneller and Buckee, 2000).

Almost all low-concentration turbulent flows are continuously vertically stratified with particle concentration, and hence density, increasing toward the bed (i.e. denser and/or larger particles preferentially settle). Fine grains, on the other hand, remain more evenly distributed through the flow (Garcia, 1994). High concentration flows are more strongly stratified, with a basal high concentration layer overlain by a more dilute and relatively more turbulent upper layer (Gladstone and Sparks, 2002; Amy et al., 2005). The contact between the two layers becomes more distinct as the basal layer becomes increasingly concentrated with sediment, and as a consequence reduces mixing between the layers (Amy et al., 2005). Facies 2a beds with larger increases (20%) in matrix content from the basal to upper unit are likely deposited from highly stratified flows, with a basal layer highly enriched in coarse grains and a comparatively depleted upper layer. Beds with a lower increase (5%) in matrix content across the basal to upper unit contact suggest deposition from more weakly stratified flows due to either an initial lower

sediment concentration within the flow, and hence reduced particle settling, or a less evolved flow.

Similar to vertically stratified flows, almost all natural turbidity currents are longitudinally stratified. Coarse particles are preferentially segregated into the head and body of the flow, and fines into the turbulent wake (Gladstone and Sparks, 2002). In Facies 2a beds, deposition of the coarser-grained lower unit would represent capacity-driven sedimentation from the forward part of the flow, and the finer-grained upper unit competence-driven sedimentation from the more rearward parts of the flow (Kneller and McCaffrey, 2003).

Where present, clasts in the basal part of Facies 2a beds are concentrated in the basal, middle, or top part of the unit and typically make up less than 5% of the total bed volume. In the upper matrix-rich part, clasts are concentrated at or near the contact between the two units, or are evenly dispersed throughout the upper unit and make up 5-15% of the total unit volume. The consistent (i.e. non-random) position of clasts in beds of Facies 2a also suggests the effects of vertical or longitudinal flow stratification. Sumner et al. (2012) suggested that, if not deposited en masse, clasts positioned along specific horizons were either preferentially transported along a rheological interface within the flow, or deposited in an aggrading bed based on their longitudinal position within a flow (i.e. front, middle, or back). In this study, the preferential position of the clasts in Facies 2a beds is alternatively interpreted to be the result of the upward movement of low-density, positively buoyant clasts within a high-density flow (see section 4.1.9 Clasts for the lateral correlation of clasts and their vertical positions within a bed). Thus, clasts were transported and then deposited on specific horizons based on their ability to be segregated upward. This, in large part, is controlled by the yield strength of the flow, which would have been sufficiently low to allow for the upward movement of clasts (Talling, 2013). Since clast

density is principally controlled by (pre-erosion) depth of burial and, in turn, level of compaction, the occurrence of low-density clasts suggests sourcing from a shallowly-buried (sea floor) substrate composed mostly of thin-bedded turbidites.

3.4.6 Facies 2b. Interpretation

Facies 2b beds are interpreted to form via rapid suspension fallout from high-density, sand-rich turbidity currents. These beds are lithologically similar to Facies 2a beds, except they lack the upper (more matrix-rich) sandy claystone unit. Thus, the basal unit in Facies 2b beds is interpreted to have been deposited in the same way as the basal clayey sandstone unit in Facies 2a beds. Facies 2b beds, however, were deposited from more concentrated (specifically, more sand-rich) and more rapidly deposited flows, which prevented flow stratification and, therefore, the upper, more matrix-rich part from forming (see section 5.1.4 Depositional Interpretation for further explanation). The upper, cleaner, ripple cross-stratified or planar- to wavy-laminated tops of Facies 2b beds are interpreted like those in Facies 1 and Facies 2a beds.

3.5 Facies 3. Coarse-tail graded or massive sandstone with finer-grained, normally graded, stratified tops

3.5.1 Facies 3. Macroscopic Characteristics

Facies 3 beds are composed of a basal tan-coloured sandstone overlain abruptly or gradationally by a medium to dark grey, finer-grained, stratified unit and a mudstone cap (*Figure 24*). Beds range from 5-225 cm thick, but most commonly are 40-60 cm. Basal contacts are usually sharp and undulate with <1-27 cm relief over a few meters laterally, however, in 44% of

beds, basal contacts are sharp and planar. Where bed bases are undulatory, they are typically associated with flame structures that are approximately 3-20 cm long and 2-80 cm wide.

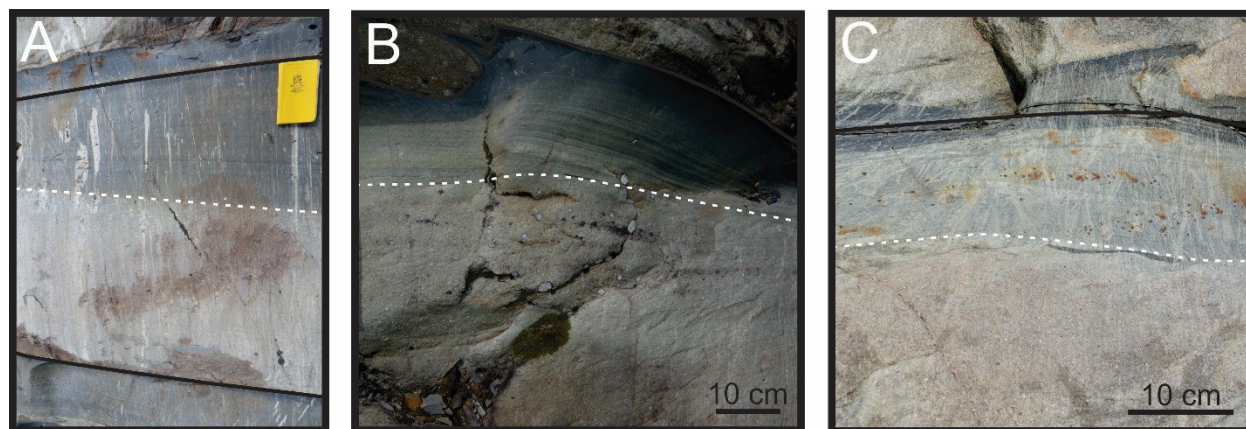


Figure 24. Photographs of Facies 3 beds. Solid black lines indicate bed bases, and dashed white lines indicate the contact between the lower coarser sandstone unit and upper finer-grained stratified unit.

The thickness of the basal tan-coloured sandstone unit ranges from 2-194 cm, but most commonly is 20-45 cm. This unit is matrix-poor or intermediate, typically containing <10-15% matrix content. Based on the distribution of grains, approximately 75% of basal sandstone units are coarse-tail graded. At the bases of these beds, grains are typically lower medium to upper very coarse sand with or without dispersed 3-4 mm granules and/or 4-6 mm pebbles and then fine upward to lower medium to upper coarse (with lesser upper fine and lower very coarse sand grains). In addition to coarse-tail graded bases, 15% of basal units are massive with a grain size of lower medium to upper medium sand with lesser upper very coarse sand and dispersed 2-4 mm granules. The remaining 10% of basal sandstone units are normally graded. Grain size at the bases of these units is upper medium to upper very coarse sand and then fines upward to upper fine to upper medium sand. Clasts, which are present in 65% of basal sandstone units, are

typically composed of mudstone (massive, laminated with silt, or internally stratified with sandstone) or medium to coarse sandstone (matrix-poor or matrix-rich). Less commonly, they are composed of matrix-rich fine to medium sandstone. Clasts are typically elongate and oriented parallel to bedding, however, less commonly they appear sheared, deformed, or rounded, and are oriented oblique to bedding. Clasts are either dispersed evenly throughout the thickness of the unit (15% of beds with clasts), or are concentrated at the base (6% of beds with clasts), middle (12% of beds with clasts), or top (32% of beds with clasts). Typically, clasts are 0.1-20 cm thick and 3-260 cm long. Uncommonly, the basal sandstone unit is faintly to distinctly parallel-laminated throughout its thickness, or in the top few centimeters. Rarely, basal sandstone units have a 1-4 cm thick ripple formset at their top, or contain dewatering pipes that are oriented perpendicular to bedding.

A medium to dark grey, finer-grained, stratified unit overlies the basal sandstone unit with either a sharp, planar contact (95% of beds) or gradationally (5% of beds). This medium to dark grey unit ranges from 3-45 cm thick, but most commonly is 15-40 cm. It typically makes up approximately 20-60% of the total bed thickness but ranges from 5-85%. Generally, thinner beds have thicker upper units than thicker beds. Based on the distribution of grains, approximately 85% of these units are normally graded, 10% are massive, and 5% are coarse-tail graded. In normally graded units, upper fine to lower fine sand with dispersed lower medium sand fines upward to silt and clay or lower medium to upper medium sand with rare lower coarse to upper coarse sand fines upward to upper very fine to lower fine sand and finally to silt and clay. Massive units are generally composed of lower medium to upper medium sand, and coarse-tail graded units typically fine upward from lower fine to upper medium sand to very fine to upper fine sand. Clasts occur in 5% of upper units, and only where the contact between the basal

sandstone unit and upper unit is gradational. Clasts are composed of mudstone (massive or laminated with silt), typically 0.3-3 cm thick and 5-23 cm long and oriented parallel to bedding. All upper units are parallel-laminated and/or wavy-laminated and ripple cross-stratified; laminated and ripple cross-stratified units are 0.2-3 cm and 1.5-3 cm thick, respectively. Where present, the parallel-lamination is commonly discontinuous through the thickness of the unit, and may be either subtle or distinct. Wavy-laminated and ripple cross-stratified zones typically occur interstratified with parallel-laminated zones or are present at the top of the unit.

A medium to dark grey, fine-grained mudstone cap overlies the upper stratified unit in all Facies 3 beds, with either a sharp, planar contact (85% of beds) or a gradational contact (15% of beds). Mudstone caps range in thickness from 2-11 cm, but most commonly are 2-3 cm. Mudstone caps are typically massive, and uncommonly parallel-laminated with silt.

3.5.2 Facies 3. Microscopic Characteristics

Although Facies 3 is not part of the lateral transect described in this thesis (Chapter 4), a few samples were collected from the basal unit. Based on relative percentages of grains (quartz and feldspar) and matrix, the basal sandstone unit of Facies 3 beds can be classified according to the classification scheme of Folk (1964) as sandstone or clayey sandstone. These units typically comprise 47-78% framework grains, 8-39% recrystallized quartz grains, <10-15% matrix, and 0-4% carbonate cement (dominantly calcite with lesser dolomite). They are moderately sorted and therefore have many more grain-to-grain contacts than matrix-rich beds of Facies 1 and 2. Framework grains are typically sub-angular to well-rounded, equant to elongate, and spherical to non-spherical. Beds with higher framework grain percentages typically have more recrystallized quartz grains.

Framework grains are dominated by quartz (96-98%) that display undulose extinction under cross-polarized light. Recrystallized, non-undulose, <0.0625 mm quartz grains form along the boundaries of larger (>0.25 mm) quartz grains (see section 3.3.2 Facies 1. Microscopic Characteristics for the formation of recrystallized grains). The other framework component is feldspar (plagioclase) which makes up 2-4% of the total framework grains.

In the basal sandstone unit of Facies 3 beds, small (<0.0625 mm), intergrown muscovite and chlorite crystals make up 70-88% of the matrix. Larger (>0.25 mm in diameter) muscovite and chlorite crystals make up 0-10% of the matrix. Finally, 13-20% of the matrix is composed of silt-sized (0.0039-0.0625 mm) quartz grains.

The upper medium to dark grey, finer-grained, stratified unit was not sampled in this thesis, however, beds of the same facies are the focus of work by Popović (2016) and were therefore sampled and described in detail (please refer to Popović (2016) for microscopic analyses of this upper unit).

3.5.3 Facies 3. Interpretation

Facies 3 beds are interpreted to have formed from flows where the upward component of turbulence was the primary grain-support mechanism. In these beds, scoured basal contacts with <1-27 cm relief over a few meters laterally are interpreted to be the result of energetic, down-sweeping turbulent eddies that eroded the sea floor. Planar-based beds, on the other hand, are interpreted to indicate local near-bed turbulence suppression. Flame structures along some bed bases indicate gravitational loading into an under-consolidated substrate, which in turn suggests a short recurrence interval between successive sedimentation events. The basal coarse-tail graded or massive part of Facies 3 beds are interpreted to be Bouma Ta divisions, and therefore

deposited under similar conditions to coarse-tail graded and massive beds of Facies 1 and 2 (see above) – specifically, deposition from highly concentrated, rapidly waning turbulent flows. Normally graded beds were deposited from flows in which turbulence and velocity decreased gradually and the effect of hindered settling was limited. Accordingly, they indicate deposition from lower concentration flows than those that formed coarse-tail graded or massive beds. In addition, normal grading indicates deposition from longitudinally stratified flows where coarser grains were concentrated in the head of the flow and finer grains preferentially displaced toward the tail. The preferential location of clasts along discrete horizons in some Facies 3 beds indicates their ability to be segregated upward during transport, which results from their comparatively lower (clast) density and an overall lower particle concentration in the flow. The occurrence of parallel-lamination throughout the thickness of some Ta beds suggests that rates of deposition from suspension, and related bed aggradation, were lower than in flows depositing structureless beds. This, then, permitted more prolonged transport of bed load sediment, and in turn the development of diffuse planar lamination (Arnott and Hand, 1989; Leclair and Arnott, 2005). Ripple cross-stratification at the tops of Facies 3 basal units reflects waning of the flow, and bed-load transport. Angular bedforms, like current ripples, developed in the presence of bed defects, indicating that rates of sediment fallout were low enough and the near-bed density profile steep enough to allow bed defect formation and amplification (Tilston et al., 2015). The lack of climbing ripples indicates that sediment fallout was low. Dewatering pipes, although uncommon in strata of Facies 3, but also generally in deep-water Windermere strata, indicate post-depositional liquefaction of a water saturated sediment related to rapid deposition from high-concentration flows.

The rapid grain size change between the basal and upper units of Facies 3 beds suggests a period of sediment bypass between deposition of the basal Ta and the upper finer-grained part of the bed. As mentioned above, the depositing flows were longitudinally stratified with coarser-grained sediment in the head of the flow, which deposited the lower part of the bed, followed some time thereafter by fine grained deposition from the tail of the flow. The small number of beds (~5%) with gradational contacts between basal and upper units suggest similarly longitudinally stratified flows, but more of a continuous history of sedimentation. Coarse-tail graded, massive, and normally graded upper units are interpreted to form in the same way as the basal Ta part of this facies. Parallel- or wavy-laminated upper units (Bouma Tb divisions) reflect traction transport at the bed. The diffuse nature of these structures may be the result of interaction between bed load sediment and sediment being deposited rapidly from suspension (Leclair and Arnott, 2003). Ripple cross-stratification (Bouma Tc division), either interstratified with parallel- or wavy-lamination, or present at the upper unit tops, forms via bed-load transport in the presence of bed defects, and in the absence of high rates of sediment fallout (Khan and Arnott, 2010).

Mudstone caps, regardless of having either a sharp, planar contact or a gradational contact with the underlying sandier part of the bed, are interpreted to form via suspension fallout from low density turbidity currents, likely the dilute tail of the flow that deposited the underlying part of the bed. Additionally, hemipelagic fallout likely played a role in forming these caps after all other sediment input ceased.

3.6 Facies 4. Coarse-tail graded or massive sandstone with traction structured tops

3.6.1 Facies 4. Macroscopic Characteristics

Facies 4 beds are composed of a basal tan-coloured sandstone overlain abruptly by a coarser-grained, red carbonate-cemented traction structured unit and a medium to dark grey mudstone cap (*Figure 25*). Beds range from 13-158 cm thick, but most commonly are 70-150 cm. Basal contacts are usually sharp and planar, although in 40% of beds they undulate slightly with 1-2 cm relief over a few meters laterally. Rare flame structures at the base of beds are a few cm long and 5-6 cm wide.

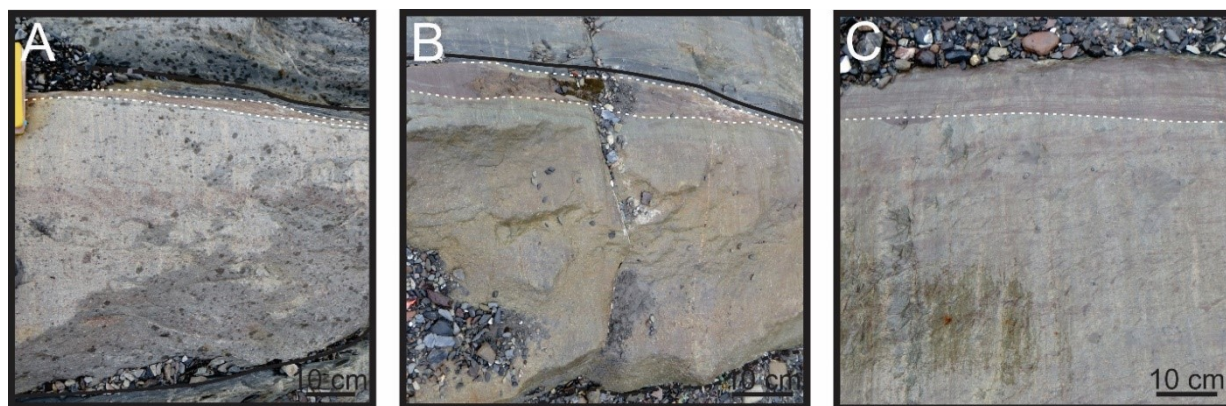


Figure 25. Photographs of Facies 4 beds. A) Single ripple formset (bounded by dashed white lines) at top of bed, B) Multiple ripple sets at bed top, and C) Parallel lamination at bed top. Solid black lines indicate bed bases, dashed black lines indicate uncertainty in contact location, and dashed white lines indicate the base and top of the upper structured unit.

The thickness of the basal tan-coloured sandstone ranges from 5-145 cm, but most commonly is 60-100 cm. Basal units are matrix-poor or intermediate, typically containing <10-15% matrix content. Based on the distribution of grains, approximately 80% of basal sandstone units are coarse-tail graded. At the base of these beds, grains are typically lower medium to

upper very coarse sand (coarse sand is dominant) with dispersed 2-3 mm granules and then fine upward to upper fine to upper coarse sand with dispersed lower and upper very coarse sand (medium sand is dominant). Rarely, the basal 10-30 cm of a bed is composed of granule conglomerate and gradually fines upward. Uncommonly, coarse-tail graded basal sandstone units gradationally become more mud-rich (about 20-30% matrix content) in their top 7-30 cm. The remaining 20% of basal sandstone units are massive. In these beds, grains consist mostly of lower medium to upper coarse sand. Clasts, which are present in approximately 45% of basal sandstone units, are oriented sub-parallel to bedding and typically occur as elongate mudstone fragments. Less commonly, clasts are composed of lower medium to upper coarse sandstone (matrix-poor or matrix-rich). Clasts are typically 0.1-5 cm thick and 2-15 cm long, but larger clasts (up to 19 cm thick and 56 cm long) are observed. In most beds, clasts are dispersed throughout the thickness of the basal sandstone and make up <5% of the bed volume. Where basal sandstone units grade upward to more mud-rich tops, clasts are preferentially concentrated at or near the top of the bed. Here, clasts are up to 93 cm long and some are deformed. In addition, rare dewatering pipes oriented perpendicular to bedding are observed. In some locations, basal units are amalgamated (i.e. they lack the upper traction structured unit and mudstone cap) and form units 50-200 cm thick.

Abruptly overlying the basal sandstone is a red carbonate-cemented traction structured unit that ranges from 4-34 cm thick. Locally, the base of the unit undulates and penetrates about 2-7 cm into the basal sandstone. Most commonly, it comprises parallel-laminated (40% of beds), ripple (45% of beds) or dune cross-stratified (15% of beds) sandstone. Some traction structured units contain parallel-laminated bases and ripple or dune cross-stratified tops. Parallel-laminated sandstone, which typically is 3-26 cm thick, is composed of upper very fine to upper coarse sand

with uncommon dispersed upper very coarse sand grains near their bases. Most parallel-laminated units are ungraded but some grade slightly upward. Ripple cross-stratified sandstone is typically 2-8 cm thick, and comprise 1-4 ripple sets that individually are 2-4 cm thick. Ripple cross-stratified units typically contain lower medium to upper coarse sand at their bases and fine upward to upper fine to lower medium sand. Finally, dune cross-stratified sand is thickest at the dune crest (30-35 cm) and thins to a few cm on either side. Laterally, dunes are continuous with single ripple sets or parallel-laminated sandstone. These units are composed of lower medium to upper very coarse sand. Where ripple cross-stratified sandstone overlies parallel-laminated sandstone, ripples are finer-grained than the underlying parallel-laminated unit. When dune cross-stratified sandstone overlies parallel-laminated sandstone, the dune is typically the same grain size as the underlying parallel-laminated unit. Also, at the base of some dunes, rare deformed mudstone clasts that are a few centimeters thick and 15-30 cm long are present.

Medium to dark grey, fine-grained mudstone overlies traction structured units in 85% of beds. In the 15% of beds that contain dune cross-stratified tops, mudstone caps are absent. Caps range from 2-23 cm thick, but most commonly are 2-12 cm. Mudstone overlies parallel-laminated units with sharp, planar contacts, however, where it overlies ripple cross-stratified units, it undulates with 1-2 cm relief as it conforms to the shape of the underlying ripple. Most caps are composed of ungraded mudstone, however, 15% of caps contain dispersed upper very fine and fine sand at their base that then grades upward to claystone. Approximately 30% of mudstone caps are diffusely parallel-laminated.

3.6.2 Facies 4. Microscopic Characteristics

The basal and upper units of Facies 4 were not sampled since Facies 4 is not part of the lateral transect described in this thesis. However, macroscopically, the basal unit of Facies 4 is analogous to that of Facies 3, and therefore is likely also similar microscopically (please refer to Facies 3. Microscopic Characteristics).

3.6.3 Facies 4. Interpretation

Facies 4 beds are interpreted to form via rapid suspension fallout from high-density, sand-rich turbidity currents. Although lithologically similar to the basal part of Facies 3 beds, the basal part of Facies 4 beds is not normally graded (i.e. it is either coarse-tail graded or massive), suggesting that the depositional flows were either more (sediment) concentrated, less longitudinally stratified, or deposited sediment more rapidly. Additionally, scours are shallower (1-2 cm over a few meters laterally), which then suggests reduced turbulence, quite possibly related to higher sediment concentration.

Unlike Facies 3 beds, Facies 4 beds are erosively overlain by traction transport deposits. In outcrop these strata appear coarser grained compared to the underlying Ta part of the bed. Upon detailed inspection, however, it can be shown that both parts of the bed consist of similar grain sizes, but that the sand in the upper, structured part of the bed is better sorted (i.e. more unimodal). As a consequence, this, in addition to the scoured basal contact between the lower and upper parts of the bed, suggests that sediment in the upper part was sourced from the in-situ reworking of the underlying Ta unit by the tail-ward part of the same depositional flow. Parallel-lamination typically occurs at the base of the traction structured units and likely represents reworking by flows that were either faster than those that formed ripple or dune cross-

stratification and/or had a near-bed sediment concentration profile that inhibited the initiation of angular bed forms. In contrast, flows that produced ripple or dune cross-stratification were either slower and/or had undergone sufficient sediment settling for the near-bed sediment concentration profile to become exponential, which then allowed for the development of bed defects, and ultimately angular bed forms to grow (Tilston et al. 2015).

Mudstones that cap Facies 4 beds have a similar depositional origin to those that overlie beds of Facies 3.

3.7 Facies 5. Laminae and very thin- to thick-bedded sandstone, mudstone, and claystone

3.7.1 Facies 5. Macroscopic Characteristics

Beds of Facies 5 beds are mostly thin-bedded (3-10 cm) and composed of sandstone, mudstone, and/or claystone, although some are thin laminae (0.1-0.3 cm), thick laminae (0.3-1 cm), very thin- (1-3 cm), medium- (10-30 cm), or thick-bedded (30-70 cm) and composed of sandstone overlain by mudstone and/or claystone (*Figure 26*). Typically, beds form bed-sets 10-600 cm thick that extend laterally over several hundreds of meters. Successions consisting mostly of laminae, very thin- and thin-bedded sandstone, mudstone, and claystone are medium to dark grey in colour, with thinner light grey- or tan-coloured stripes where sandstone is present. Successions consisting mostly of medium- or thick-bedded sandstone overlain by mudstone and/or claystone are dominantly tan-coloured, with stripes of medium to dark grey where mudstone and claystone are present. Bed bases are usually sharp and planar, although rarely and locally in medium to thick beds they shallowly undulate with <1 cm relief.

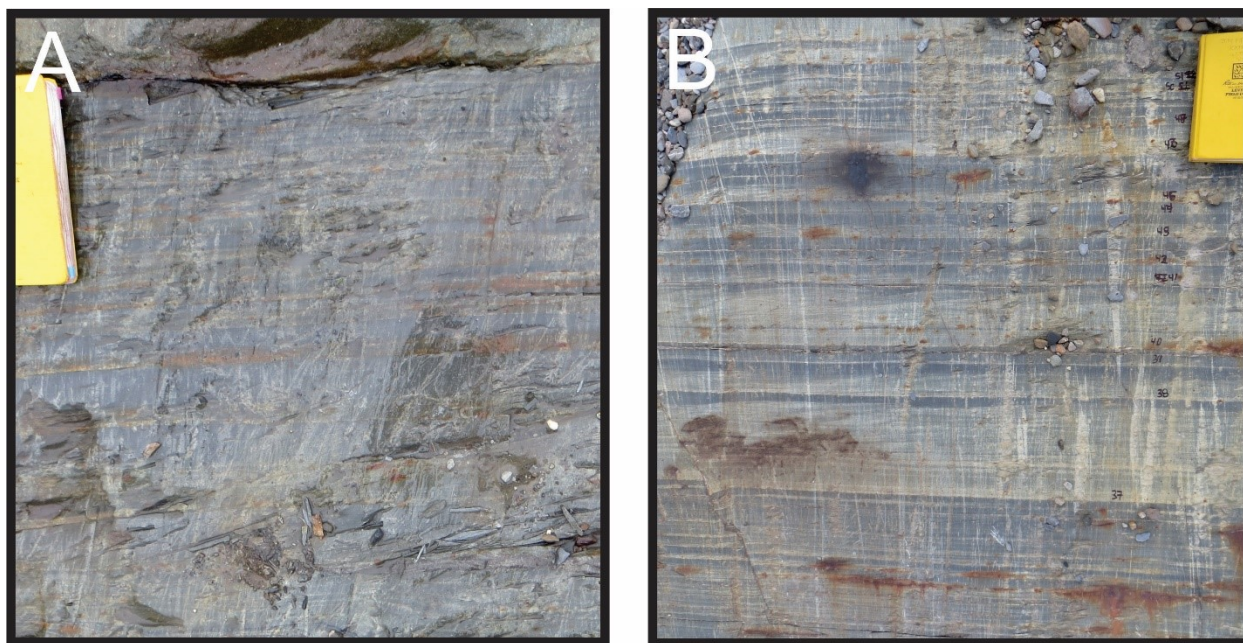


Figure 26. Photographs of Facies 5 beds. A) Very thin- to thin-bedded Bouma Tcd turbidites. B) Thin- to medium-bedded Bouma Tbcde turbidites.

For ease of explanation, thin and thick laminae, as well as very thin-bedded turbidites will be described together and referred to as ‘beds’. These beds range in thickness from 0.1-3 cm, but usually are 0.2-2 cm. The basal part of each bed has a modal grain size of silt or sand (see section 3.7.2 Facies 5. Microscopic Characteristics for a more detailed grain size analysis) and ranges in thickness from 0.001-2 cm, but is typically 0.1-0.5 cm; typically, too thin for structures to be discerned, however, where possible, these units are either discontinuously ripple cross-stratified or parallel-laminated. The associated mudstone or claystone cap typically abruptly overlies the basal silty or sandy unit with a sharp, planar contact, however, in 5% of turbidites, where the cap fills negative ripple topography, this contact undulates. In 15% of turbidites, the underlying light grey base grades into the overlying medium to dark grey cap. Medium to dark grey caps themselves are either ungraded or graded. Ungraded caps are composed of mudstone (83% of beds) or claystone (17% of beds). Generally, when graded, cap bases contain silt- and clay-sized

grains and cap tops are dominated by clay-sized grains with few to no silt-sized grains. Caps range in thickness from 0.03-1.5 cm, but are typically 0.1-1.5 cm.

In order of relative abundance, thin beds are made up of either (1) sandstone overlain by mudstone and/or claystone, (2) mudstone overlain by claystone, (3) mudstone, or (4) claystone. Where beds are composed of sandstone overlain by mudstone and/or claystone, the sandstone part of the bed ranges from 0.1-8 cm thick, but most commonly is 0.1-2 cm. The overlying mudstone is typically between 0.5-9 cm thick, but on average is 1-4 cm. Where claystone is the primary cap on a bed (i.e. mudstone is absent) it is 3-8 cm thick, however, if it overlies mudstone it is between 0.1-1 cm thick. Sandstone units are typically ripple cross-stratified, although some are parallel-laminated and/or structureless. Ripple cross-stratified sandstone units typically comprise 1-3 upper very fine to upper medium ripple sandstone sets (with rare dispersed upper coarse sand grains), with each set about 1-3 cm thick. Thinner (<1 cm) ripple cross-stratified sandstone units are finer-grained and ungraded, whereas thicker (>5 cm) units generally grade upward from upper medium (with dispersed upper coarse sand grains) to lower medium sandstone. Parallel laminated sandstone is typically ungraded and composed of lower fine to upper medium sand. Structureless sandstone typically grades upward from lower or upper medium sand (with dispersed upper coarse sand grains and rare 2-3 mm granules) to lower fine or lower medium sand, respectively, however, are also commonly ungraded. In thin (<1 cm) structureless sandstone units, grain size is upper very fine to lower fine sand. Sandstone units that have parallel-laminated or structureless bases and ripple cross-stratified tops, or structureless bases and parallel-laminated tops are 3-8 cm thick and normally graded. Typically, the change in structure, for example from parallel-laminated to ripple cross-stratified sandstone, occurs mid-bed. Mudstone and claystone units either occur as caps above a basal sandstone unit (as

discussed above), or as separate beds. Where they cap a parallel-laminated or structureless sandstone the mudstone or claystone contact is planar, but where they overlie a ripple cross-stratified sandstone unit their contact conforms to the shape of the ripple formset. Where they form a discrete bed, mudstone is 1-5 cm thick and claystone is 2-8 cm. Both have sharp, planar bases. Mudstone is either massive or graded (i.e. decreasing abundance of silt grains upward) and, in a few beds, is diffusely to distinctively parallel-laminated. In some beds, mudstone grades upward to claystone. Where claystone is present it is generally massive, although some beds are graded or are subtly parallel-laminated.

Medium- and thick-bedded sandstone overlain by siltstone and/or mudstone differs from thin beds due to the much thicker basal sandstone unit. Beds typically range from 10-70 cm and consist mostly of structureless, parallel-laminated, and/or ripple cross-stratified sandstone. Mudstone and claystone caps range from 1-10 cm thick, but most commonly are 1-3 cm. Parallel-laminated sandstone is most common and forms 10-70 cm thick sandstone units. These units are typically ungraded and contain lower fine to upper coarse sand. Ripple cross-stratified sandstone units are typically 8-15 cm, and their bases contain lower fine to upper medium sand (with dispersed upper coarse sand grains) that grades upward to upper very fine to lower medium sand. These units contain between 3-9 sets of vertically stacked 1-3 cm thick ripple sets. Structureless sandstone is typically 10-12 cm thick, and grades from lower medium to upper coarse sand (with dispersed upper very coarse sand grains and rare 2-3 mm granules) to upper fine to lower medium sand. Sandstone units that have parallel-laminated or structureless bases and ripple cross-stratified tops, or structureless bases and parallel-laminated tops are 12-40 cm thick, as are units that contain lower parallel-laminated sandstone and upper ripple cross-stratified sandstone. These beds all fine upward. In beds with parallel-laminated bases and ripple

cross-stratified tops, the top 5-50% (most commonly 5-15%) of the bed is ripple cross-stratified. In beds with structureless bases and ripple cross-stratified or parallel-laminated tops, the top 50-70% of the bed is ripple cross-stratified or parallel-laminated. Few beds contain structureless bases overlain by parallel-lamination overlain by ripple cross-stratification. In these beds, the basal half of the bed is structureless, and the upper half is parallel- overlain by ripple-laminated sandstone. Rarely, medium and thick beds contain mudstone clasts that, on average, are 1 cm thick and 12 cm long, undeformed, and oriented parallel to bedding. Mudstone and claystone caps overlie most basal sandstone units. Most caps are massive, however, some mudstone caps are graded (i.e. upward decrease in the abundance of silt) or diffusely to strongly parallel-laminated. Rarely, mudstone forms individual beds that are 10-20 cm thick, and diffusely parallel-laminated or graded (i.e. upward decrease in the abundance of silt).

3.7.2 Facies 5. Microscopic Characteristics

For the purpose of this thesis, only laminae and very thin- to thin-bedded turbidites will be discussed in this section. These turbidites are of particular importance because they make up part of the lateral transect described in Chapter 4 and interpreted in Chapter 5. For this reason, they were sampled much more extensively than thicker turbidites (discussed above), which are not part of the lateral transect. The light grey basal part of each turbidite has the same mineralogical composition as the sandy basal unit of a matrix-rich bed (see section 3.3.2 Facies 1. Microscopic Characteristics). Typically, however, turbidites are better sorted (moderately to very well sorted) with only 10-30% clay-sized (<0.0039 mm) muscovite and chlorite crystals. Turbidites are typically ungraded, however, approximately 25% are poorly- to well-graded. Turbidites tend to be graded where their modal grain size is silt to very fine sand. Approximately

55% of turbidites have a modal grain size of silt. In these turbidites, the maximum grain size, which generally makes up less than 5% of the total number of grains, ranges from lower very fine to upper fine sand, although rare grains of lower medium to upper coarse sand are observed. Importantly, it should be noted that these turbidites, as defined earlier, should be considered to be composed of matrix. However, they differ significantly from the matrix-rich units in Facies 1 and 2 due to their comparatively lower amount of clay-sized muscovite and chlorite crystals. In addition to silt-dominant turbidites (55%), approximately 40% of turbidites have a modal grain size of lower to upper very fine sand. The maximum grain size, which generally makes up less than 5% of the total number of grains, is typically lower medium to upper coarse sand or, less commonly, lower to upper fine sand or lower very coarse sand. The remaining 5% of turbidites have a modal grain size of upper fine sand with a maximum grain size of upper coarse to lower very coarse sand.

3.7.3 Facies 5. Interpretation

Facies 5 beds are interpreted to be deposited from fully turbulent flows, and represent parts of the classic Bouma sequence (Ta-e; Bouma, 1962). Thinner beds are interpreted to form further laterally from the axis of the flow where flow velocity and sediment concentration are lower. Structureless units are interpreted to be Bouma Ta deposits, and represent direct suspension deposition from high-concentration, rapidly depositing flows (see discussion on coarse-tail grading in Facies 1, 2, 3, and 4). Parallel-laminated units are interpreted as Bouma Tb deposits, and represent deposition from a tractional flow boundary zone and form when bed aggradation rates are high, and suppresses the formation of 3D bedforms (Leclair and Arnott, 2005). Ripple cross-stratified units are interpreted as Bouma Tc deposits, and represent

deposition from flows that were sufficiently slow and had a suitably steep near-bed sediment concentration to allow bed defects to form, become amplified, and ultimately develop current ripples.

Mudstone and claystone caps (Bouma Td and Te deposits) represent suspension fallout from the dilute, fine-grained, low-energy tails of flows. Additionally, inter-event hemipelagic fallout likely contributed sediment too.

4 Lateral and Vertical Facies Trends

This chapter will describe the lateral and vertical trends within matrix-rich beds at Castle Creek. Specifically, it will detail general facies transitions, overall bed thickness, individual unit thickness (i.e. parts of the bed), basal contact morphology, grain size, matrix percentage, clast data (i.e. composition, size, location), and bedform structures. Chapter 5 will then summarize the various trends described in this chapter and provide a conceptual depositional model to explain the origin of matrix-rich beds at Castle Creek.

The most laterally extensive stratigraphic section studied in this thesis is 650 m wide and located in the upper Kaza Group (proximal basin floor deposits). It is the only studied section in which all of the lithological variants of matrix-rich rocks are exposed along a single transect, and therefore is considered the “type section”. Moreover, this section is thought to most closely parallel the direction of local sediment transport. For this reason, it was logged and mapped in the greatest detail. Nevertheless, matrix-rich beds in all measured sections exhibit the same lateral trends, the difference being that in narrow outcrops the full lateral succession of facies is only partly exposed.

Data for this chapter have been compiled from the detailed description of 12 matrix-rich beds with lateral extents ranging from approximately 350 to 650 m; 8 of the beds were deposited on the slope, and the remaining 4 on the proximal basin floor. Laterally, from the proximal to distal part of the lateral transect, each bed shows a change from sand-rich to matrix-rich, respectively.

4.1 Lateral Trends

4.1.1 General Facies Trends

Amalgamated clayey sandstone beds (Facies 2b) transition laterally to bipartite (Facies 2a) beds with the development of an upper more matrix-rich (clayey sandstone or sandy claystone) unit. Laterally, the basal unit progressively thins whereas the upper unit thickens slightly (see section 4.1.5 Thickness of Bed Parts). Eventually, the basal unit pinches out and the bed consists entirely of sandy claystone (Facies 1), which then fines and thins laterally until it too pinches out. Across this entire lateral transect, a thin, matrix-poor cross-stratified or parallel-to wavy-laminated unit directly overlies the upper more matrix-rich (clayey sandstone or sandy claystone) layer, and in turn is overlain by a mudstone or claystone cap. Notably, this cleaner upper layer extends beyond the pinch-out of the sandy claystone (Facies 1) bed, and correlates with a very thin- to thin-bedded turbidite in the distal part of the lateral transect (see *Appendix F* for lateral facies descriptions). Associated with lateral changes in lithofacies are changes in matrix content, increasing gradually from 10-30% in clayey sandstone (Facies 2b) beds to a maximum of approximately 65-80% in sandy claystone (Facies 1) beds, and then decreasing to <10% in the silt- or sand-rich very thin- to thin-bedded turbidites (Facies 5). In addition, clasts are generally absent in the most proximal part of the lateral transect (i.e. clayey sandstone beds; Facies 2b), but are present in bipartite (Facies 2a) and sandy claystone (Facies 1) beds. Clasts are initially present near the bases of beds, but laterally are located progressively higher in the bed. *Figure 27* (below) illustrates these general facies trends.

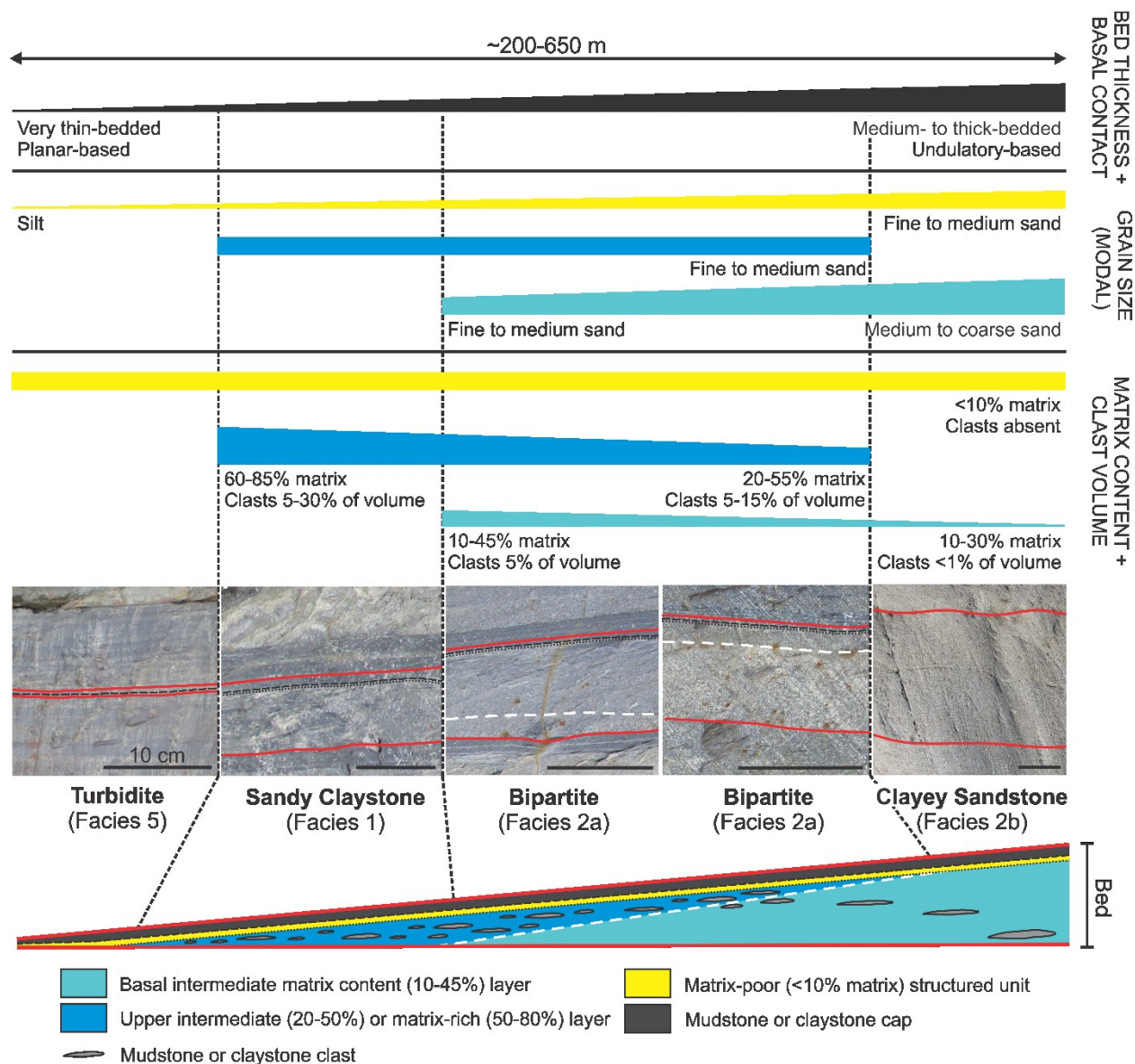


Figure 27. Right to left – Lateral trends in a matrix-rich bed: initially, a clayey sandstone (Facies 2b) transitions to a bipartite (Facies 2a) bed, in turn to a sandy claystone (Facies 1), and ultimately a very thin- to thin-bedded turbidite (Facies 5). In photographs (middle) and the schematic of an idealized lateral transect (bottom), red lines indicate bed bases, white dashed lines indicate the contact between the basal and upper parts of a bipartite (Facies 2a) bed, lower black dashed lines indicate the base of the matrix-poor structured unit, and upper black dashed lines indicate the base of the mudstone or claystone cap. Note that the matrix-poor turbidite is very thin in the photographs (few mm).

4.1.2 Trend Dimensions

For the purpose of this thesis the lateral distance over which the various facies trends occur will not be discussed in detail, since these dimensions vary based on the orientation of the outcrop in comparison to the direction of transport. Additionally, distances would be expected to change in each example due to differences in sea floor topography and gradient, flow size and velocity, and grain size. Notwithstanding, the lateral transect of a matrix-rich bed, namely from a clayey sandstone (Facies 2b) bed to a very thin- to thin-bedded turbidite (Facies 5), occurs over approximately 200 to 650 m. Since the most sand-rich part of the lateral transect is covered in each example, it is unknown how clayey sandstone (Facies 2b) beds change in the direction opposite to the progressively more matrix-rich trend. For this reason, it is not possible to accurately determine a total width of this part of the full lateral transect. Based on observations within the studied sections, the bipartite (Facies 2a) zone typically ranges from 50 to 200 m wide (from the initial formation of the upper unit to the pinch-out point of the basal unit), and the sandy claystone (Facies 1) zone typically ranges from 50 to 280 m wide (from the pinch-out point of the basal bipartite unit to the pinch-out point of the sandy claystone bed). The lateral extent of a very thin- to thin-bedded turbidite bed, beyond the pinch-out point of a sandy claystone (Facies 1) bed, was not measured.

4.1.3 Bed Thickness

Beds generally thin away from the sand-rich part of the lateral transect toward the matrix-rich part. Beds within the clayey sandstone (Facies 2b) zone of the lateral transect are typically medium- to thick-bedded (approximately 25-55 cm). The laterally adjacent bipartite (Facies 2a) beds initially range from medium- to thick-bedded (approximately 25-45 cm) but then gradually

diminish to thin- to medium-bedded (approximately 3-25 cm) laterally. These then transition to sandy claystone (Facies 1) beds, which initially are thin- to medium-bedded (3-25 cm) and then they too progressively thin to very thin- to thin-bedded (2-5 cm). These are then succeeded by thin- to thick-laminae or very thin- to thin-bedded (0.1-5 cm) turbidites at the end of the lateral transect. *Figure 28* and *Figure 29* illustrate the lateral thickness trends observed in proximal basin floor deposits of the upper Kaza Group and slope deposits of the Isaac Formation, respectively. A more detailed analysis of bed thickness is presented later (section 4.1.5 Thickness of Bed Parts).

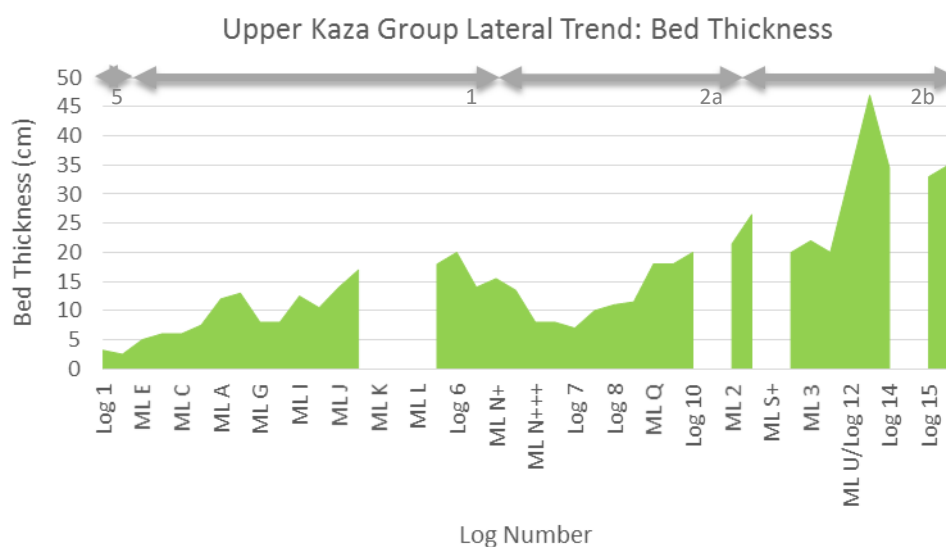


Figure 28. Lateral thickness trend of a single bed in the upper Kaza Group (proximal basin floor) type section. Arrows and associated numbers indicate facies zones. Spaces indicate locations where the bed was covered.

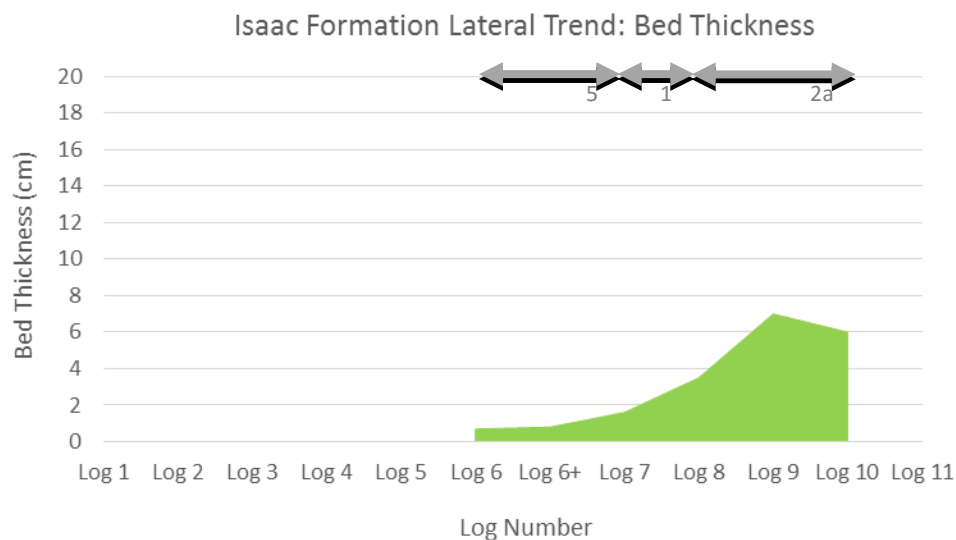


Figure 29. Lateral thickness trend of a single bed in the Isaac Formation (base of slope). Arrows and associated numbers indicate facies zones. Spaces indicate locations where the bed was covered.

4.1.4 Basal Contacts

Over the lateral transect, basal contacts also show a consistent evolution from sharp and scoured (or, less commonly, loaded) in more sand-rich strata (Facies 2b) to sharp and planar in more matrix-rich strata (Facies 1). The depth of scour (or loading) decreases progressively over the lateral transect. As a result, sand-rich strata tend to be amalgamated, whereas beds become well-bedded laterally. The transition from sharp and scoured to sharp and planar bases occurs in the bipartite (Facies 2a) zone of the lateral transect, and typically where the upper unit of the bipartite bed becomes equal to or thicker than the basal unit.

Due to greater scour in the sand-rich part of the lateral transect, the upper matrix-rich part of a bipartite (Facies 2a) bed may be completely eroded. As a consequence, it becomes difficult to differentiate a bipartite (Facies 2a) bed from a clayey sandstone (Facies 2b) bed that lacks the upper matrix-rich unit. Observed also, but rarely, is that sharp, flat bases of sandy claystone

(Facies 1) beds become scoured over distances of a few meters laterally. Interestingly, these scours are filled with bipartite (Facies 2a) strata, but further laterally the bed base returns to being sharp and planar and then is overlain by sandy claystone (Facies 1) strata.

4.1.5 Thickness of Bed Parts

Although there are four main facies (Facies 1 – sandy claystone beds; Facies 2a – bipartite beds; Facies 2b – clayey sandstone beds; and Facies 5 – very thin- to thin-bedded turbidites) that have been described in the lateral transect of a matrix-rich bed, there are only truly four main bed ‘parts’ that persist across this lateral transect. These include: (1) an intermediate matrix content layer (i.e. a clayey sandstone bed – Facies 2b, or the basal part of a bipartite bed – Facies 2a), (2) an upper more matrix-rich layer (i.e. the upper part of a bipartite bed – Facies 2a, or a sandy claystone bed – Facies 1), (3) a matrix-poor, structured unit (i.e. the upper turbidite that caps all deposits except where the top has been scoured), and (4) a mudstone or claystone cap (that overlies all deposits except where eroded) (*Figure 30*).

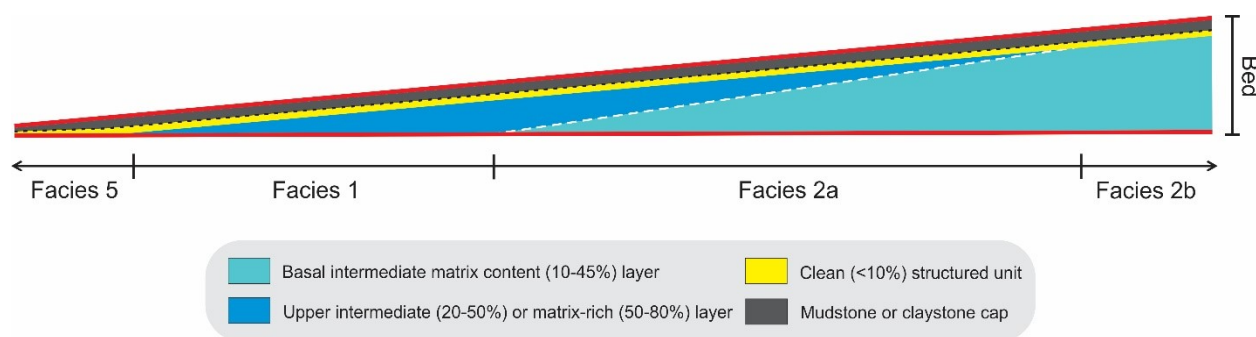


Figure 30. Schematic of the bed parts within a lateral transect of a matrix-rich bed.

The basal intermediate matrix content layer generally thins progressively laterally from 35-55 cm thick in more proximal Facies 2b strata to 25-50 cm in more distal Facies 2b strata. In

the bipartite (Facies 2a) zone of the lateral transect, the basal intermediate matrix content layer is initially 25-43 cm thick and then thins progressively to 0 cm at its pinch out. Interestingly, within this zone, the basal intermediate matrix content layer thins faster than the upper more matrix-rich layer thickens (*Figure 31* and *Figure 32*). In fact, the upper more matrix-rich layer, once developed, tends to maintain a similar thickness (0.5-10 cm) across the bipartite (Facies 2a) zone, with only minor thickening laterally until the basal intermediate matrix content layer pinches out. As a consequence, the upper layer becomes a greater portion of the total bed thickness laterally, despite a negligible increase in thickness (*Figure 31* and *Figure 32*). Beyond the pinch-out of the basal layer, the upper more matrix-rich layer, now a sandy claystone (Facies 1) that occupies the whole bed, continues to increase in thickness laterally, except at a faster rate than in the bipartite (Facies 2a) zone of the transect. It thickens over 40-100 m until it reaches a maximum of approximately 10-15 cm, and then thins progressively to its pinch-out. In the upper Kaza Group (proximal basin floor deposits), the maximum thickness of a sandy claystone (Facies 1) bed commonly exceeds the thickness of the laterally adjacent, most distal bipartite (Facies 2a) bed. In contrast, a sandy claystone (Facies 1) bed deposited at the base of slope (Isaac Formation) is typically equally thick, or only slightly thicker than the laterally adjacent, most distal bipartite bed.

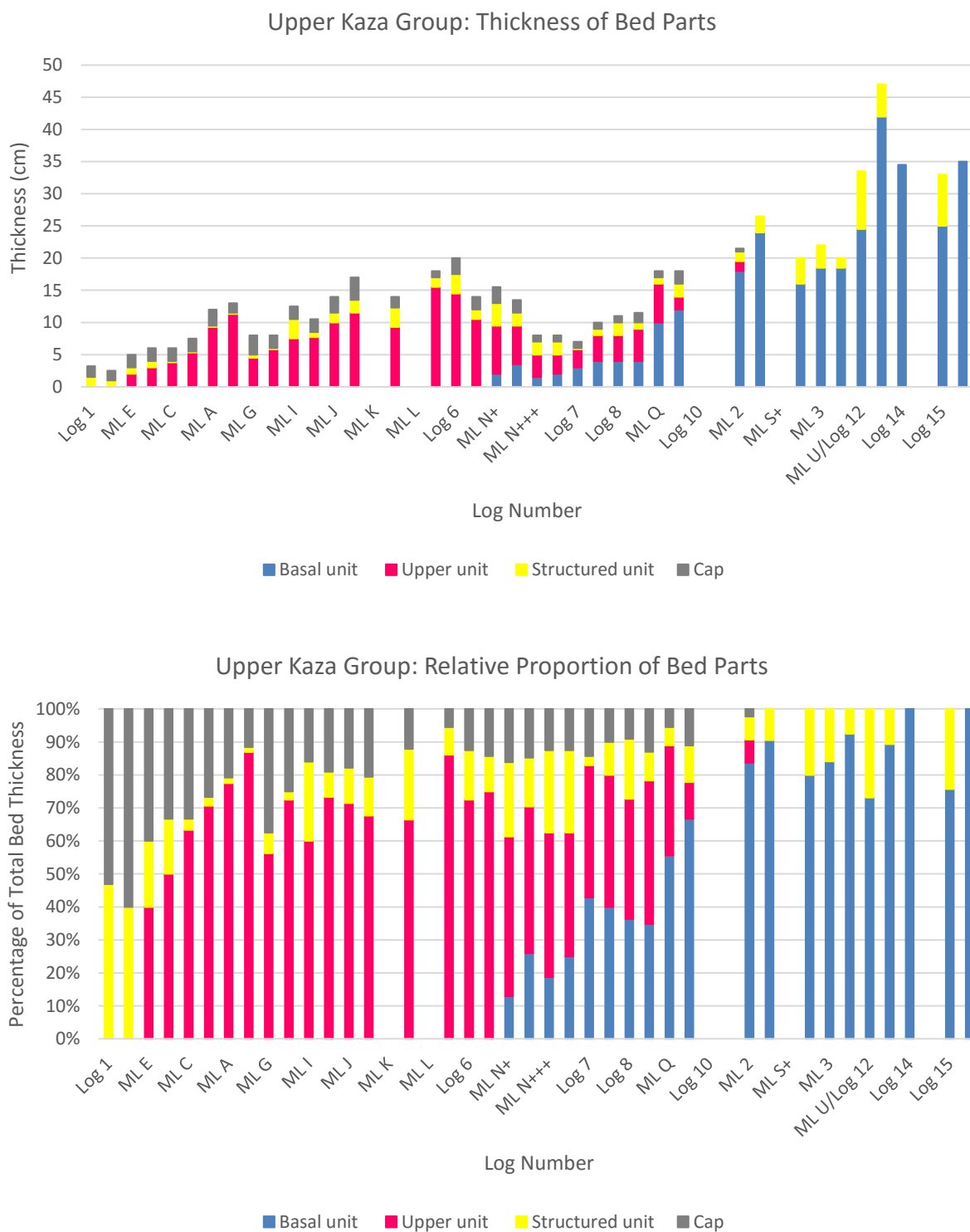


Figure 31. Thickness (Top) and Relative Proportion (Bottom) of individual bed parts from a single bed in the upper Kaza Group. Spaces indicate locations where the bed was covered.

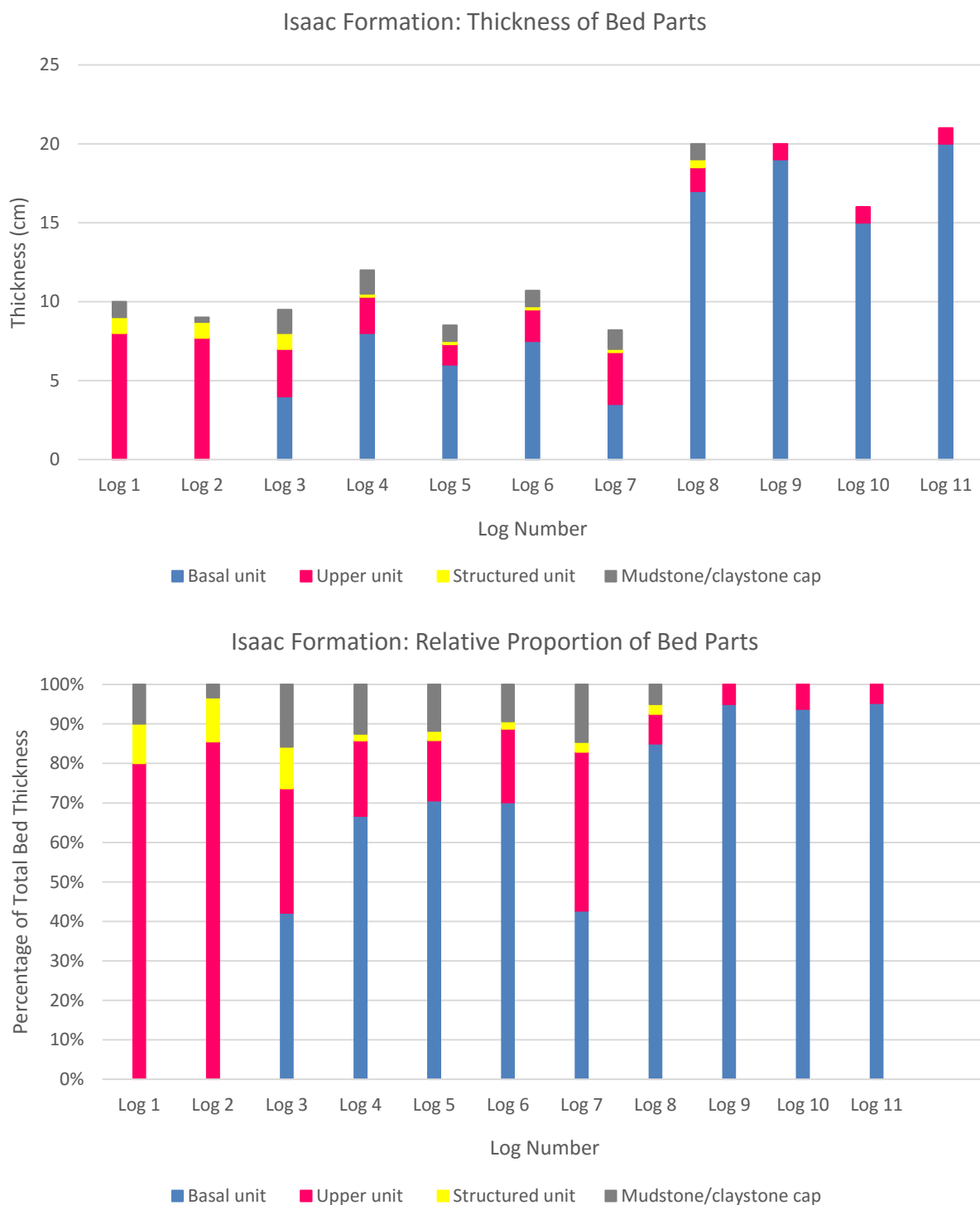


Figure 32. Thickness (Top) and Relative Proportion (Bottom) of individual bed parts from a single bed in the Isaac Formation.

In contrast to the laterally varying thickness of the basal intermediate matrix content and upper more matrix-rich layers, the matrix-poor structured unit changes little in thickness across the entire lateral transect, and is typically <1 to a few centimeters thick. Local increases and decreases in thickness are observed, but no apparent overall trend can be discerned. Typically, where this unit thickens slightly, the underlying matrix-rich unit antithetically thins. In the proximal part of the lateral transect, where some Facies 2b (clayey sandstone) bed tops have not been scoured by later events, structured units tend to be thicker (few cm to 10 cm) compared to elsewhere along the lateral transect. Mudstone or claystone caps are typically <1 to few centimeters thick and change little in thickness across the entire lateral transect. However, like the structured unit, the cap is locally thinner in the proximal part of the lateral transect due to scour by sand-rich Facies 2b beds.

4.1.6 Structured Unit

The structured unit that underlies the mudstone or claystone cap of each bed is typically ripple cross-stratified (single set) throughout the entire lateral transect. Locally, ripple cross-stratification is laterally discontinuous, or replaced by parallel- to wavy- lamination.

4.1.7 Grain Size

Like dimensional trends discussed above, only general grain size trends are described here. In order to more accurately capture these trends, a more systematic sampling of matrix-rich beds would be required. For example, in a single bed, samples could be obtained from the base, middle, and top at fixed lateral distances. From there, point counting of each grain size would be required. Also, to build a robust, statistically significant dataset, many beds would require

sampling. In this thesis, systematic sampling was not carried out, but would be an interesting topic for future research. Instead, grain size measurements were made using a 16X Ruper hand lens and a standard American/Canadian grain size comparator to assess grain size at bed bases and tops, as well as at any intervening interfaces, specifically below and above the bipartite interface. These data show a general fining along the lateral transect from proximal to distal, although local, irregularly-spaced increases or decreases in a single bed are observed. The basal intermediate matrix content layer is typically medium- to coarse-grained (modal grain size of the sand-grade or coarser fraction) in the most proximal parts of the lateral transect and becomes fine- to medium-grained in more distal strata. Within this layer, the range of grain sizes remains approximately constant across the lateral transect, specifically from upper very fine to 2-4 mm granules, however, the proportion of coarser grain sizes decreases progressively laterally. The upper more matrix-rich layer is typically fine- to medium-grained (modal grain size of the sand-grade or coarser fraction) in both proximal and distal strata. Similar to the basal unit, the range of grain sizes remains approximately constant across the lateral transect, specifically from lower very fine to 2-4 mm granules, however grains coarser than upper medium sand are typically dispersed, and progressively decrease in abundance laterally until they are rare or absent. The grain size at the base of the upper unit is typically equal to or finer than the grain size at the top of the basal unit. Grain size in the upper structured unit varies only slightly across the lateral transect, specifically from fine- to medium-grained sand with dispersed upper medium sand grains to fine-grained sand with dispersed lower medium sand grains. Finally, beds are capped by a uniform mixture of silt and clay (mudstone) or clay (claystone). Most beds, however, are capped by mudstone.

The basal layer is typically coarse-tail graded across the entire lateral transect (i.e. in a Facies 2b clayey sandstone bed and the laterally adjacent basal part of a Facies 2a bipartite bed). Locally, the basal layer is massive. Approximately 50% of beds contain an upper layer that is massive across the entire lateral transect (i.e. in the upper part of a Facies 2a bipartite bed and the laterally adjacent Facies 1 sandy claystone bed). The remaining 50% of beds contain upper units that laterally alternate between massive and coarse-tail graded, with no apparent pattern. Locally, where the basal layer is massive, the upper layer is also typically massive.

4.1.8 Matrix Content

A total of 59 samples from 7 beds were point counted for matrix content. These quantitative data were then used to calibrate qualitative field-based observations of bed shade and colour (i.e. light grey, medium grey, and dark grey). Significantly, it was shown that bed shade did in fact provide a reasonable approximation of matrix content with darker beds, as shown with point counting, containing more matrix.

Matrix content is lowest in proximal parts of the basal intermediate matrix content layer (i.e. Facies 2b beds). Here, matrix content is typically 10-30%, and bed shade is light grey. Laterally, matrix content either remains constant or increases slightly by approximately 10-15% to 20-45% and bed shade darkens slightly to light to medium grey. In contrast, the upper unit, when first developed (in Facies 2a – bipartite beds), contains between 20-55% matrix content, which actually places it in the intermediate matrix content and matrix-rich category. Here, bed shade is medium grey. Laterally, matrix content in the upper unit either remains constant, or increases progressively by about 10-20% to 30-65%. Increased matrix content causes the upper unit to be darker, typically medium to dark grey. Where the basal unit of a bipartite bed pinches

out, the colour of the basal and upper units is typically similar, making the contact between the two difficult to identify. Beyond the pinch-out of the basal layer, matrix content in the upper layer increases to a maximum of approximately 65-80% (now a Facies 1 sandy claystone bed), and changes little up to the pinch-out of this layer. Beds are characteristically dark grey. In terms of composition, the matrix in both the basal and upper layers tend to have low silt content (approximately 0-20% of the total matrix percentage; the remaining percentage being composed of clay), but, in the same bed, the upper unit tends to contain between 5-20% more silt than the basal layer. Silt content either remains roughly constant from proximal to distal deposits, or decreases slightly. The overlying structured unit typically contains <10% matrix content and is tan with a red hue. This remains constant over the entire lateral transect, although local increases or decreases in matrix content, and hence colour, are observed.

4.1.9 Clasts

Although mudstone and claystone clasts dominate, uncommon sandstone clasts (both matrix-poor and matrix-rich) are observed. Clasts are typically uncommon in clayey sandstone (Facies 2b) beds (occur in 8% of beds), but laterally become progressively more abundant in bipartite (Facies 2a) beds (occur in 20% of basal units and 40% of upper units) and sandy claystone (Facies 1) beds (occur in 85% of beds) (*Figure 27*). Clasts are absent in very thin- to thin-bedded turbidites. Where present in clayey sandstone (Facies 2b) beds, clasts are typically located near the base or middle part of the bed. In bipartite beds, clasts are initially located in the basal unit (typically mid-bed), but further laterally rise upward and begin to concentrate below the bipartite interface. Laterally, as the upper more matrix-rich unit makes up more than approximately 30% of the bed thickness, clasts begin to concentrate at the bipartite interface. At

this point, clasts are also present in the upper unit, and are evenly dispersed throughout its thickness. Here, only a few clasts are present in the basal unit, and those that are present are typically concentrated near its top. After the basal unit of the bipartite (Facies 2a) bed has pinched-out, the resulting sandy claystone (Facies 1) bed typically contains clasts that are evenly dispersed throughout its thickness. Across the lateral transect, the clast size range remains approximately constant (<1-few cm wide and a few-20 cm long), although mud chips (<0.1-0.5 cm wide and <0.5 to a few cm long) are much more abundant in sandy claystone (Facies 1) beds.

4.1.10 Bipartite vs. Amalgamated Beds

In order to avoid confusing a bipartite (Facies 2a) bed with two amalgamated beds (i.e. a lower, cleaner (Facies 2b) bed and an upper matrix-rich (Facies 1) bed), all beds must be traced out laterally. Field observations suggest that certain features are unique to (or exhibited by most) bipartite beds. For example, the interface that separates the basal and upper units in bipartite beds is sharp and planar, whereas the contact between amalgamated beds is sharp but undulatory and/or irregular. Additionally, in thicker (>15 cm) bipartite beds, the basal unit represents a larger overall proportion of the total bed thickness (see section 3.4.1 Facies 2a. Macroscopic Characteristics), and clasts are preferentially concentrated in the basal unit. In thinner (<15 cm) bipartite beds, on the other hand, the upper unit makes up more of the total bed thickness (see section 3.4.1 Facies 2a. Macroscopic Characteristics) and clasts are preferentially concentrated at the bipartite interface or dispersed in the upper unit. Where clasts are only present at the interface between two units, but are absent elsewhere, they likely represent an (amalgamated) scour contact and the remnants of the mudstone cap of the underlying unit.

4.2 Stacking Patterns

Where matrix-rich beds are present, they tend to stack on top of one another in vertical successions that range from 20-150 cm thick and contain between 2-15 beds. It is much less common to observe them intercalated with matrix-poor beds. Matrix-rich successions are made up of one or more vertically stacked packages, and each package contains a basal and upper part (*Figure 33*). The basal part is made up of one laterally extensive matrix-rich bed that sharply overlies either a matrix-poor or matrix-rich bed. The upper part, which abruptly overlies the basal bed, is made up of one or several less laterally extensive matrix-rich beds, which all have similar lateral extents. These less laterally extensive beds are thinner and finer-grained than the basal bed, but are similar in thickness and grain size to one another. The lateral fining and thinning trend in each matrix-rich bed, combined with the vertical stacking pattern of beds in each package, produces an overall wedge-shaped geometry, which at least in two dimensions resembles the gull-wing geometry commonly attributed to channel bounding levees (*Figure 34*).

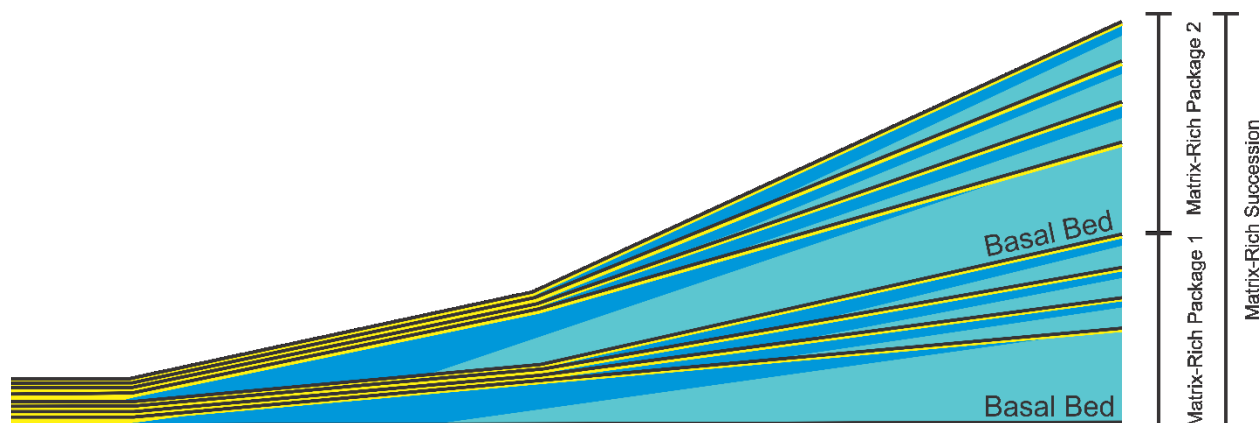


Figure 33. Schematic of the general vertical stacking pattern of a succession of matrix-rich beds. A single vertical succession may comprise several matrix-rich packages, each consisting of a basal laterally extensive bed overlain by less extensive beds. Note that this schematic is vertically exaggerated in order to display the stacking pattern more easily.

Within a single matrix-rich succession the basal bed in each package tends to have similar lateral dimensions; the same is true for the upper beds. As a consequence of having similar lateral dimensions, all basal beds within a matrix-rich succession experience facies changes (i.e. Facies 2b → Facies 2a → Facies 1 → Facies 5) at similar positions along the lateral transect (typically within 10 m). In addition, they have similar thicknesses, grain size, and matrix content at any given point in the lateral transect. The same is true for the less laterally extensive upper beds. As a result of the stacking patterns within one package, more distal lithofacies exhibited by the upper beds overlie more proximal lithofacies of the basal bed. Where the upper, less extensive beds have laterally transitioned to very thin- to thin-bedded turbidites (Facies 5), they produce a stacking pattern that, locally, appears to be a banded cap on top of the underlying bed (*Figure 35*, *Figure 36*, and *Figure 37*). The banding is produced by the alternation of lighter, coarser-grained turbidite bases and their darker, finer-grained caps. In poorly exposed outcrops this banding may be interpreted to be the interlaminated Td division of a single turbidite (i.e. single depositional event). However, at Castle Creek it can be clearly shown that, further laterally, each very thin- to thin-bedded turbidite in fact overlies a separate matrix-rich unit (that had pinched out), and therefore represents a separate depositional event. An example of this stacking pattern is shown in *Figure 38* (A) and (B). At a microscopic scale, it can be seen that the coarsest grains within a stacked succession of turbidites are contained within the lowermost turbidite base, and grain size decreases upward in successive bases.

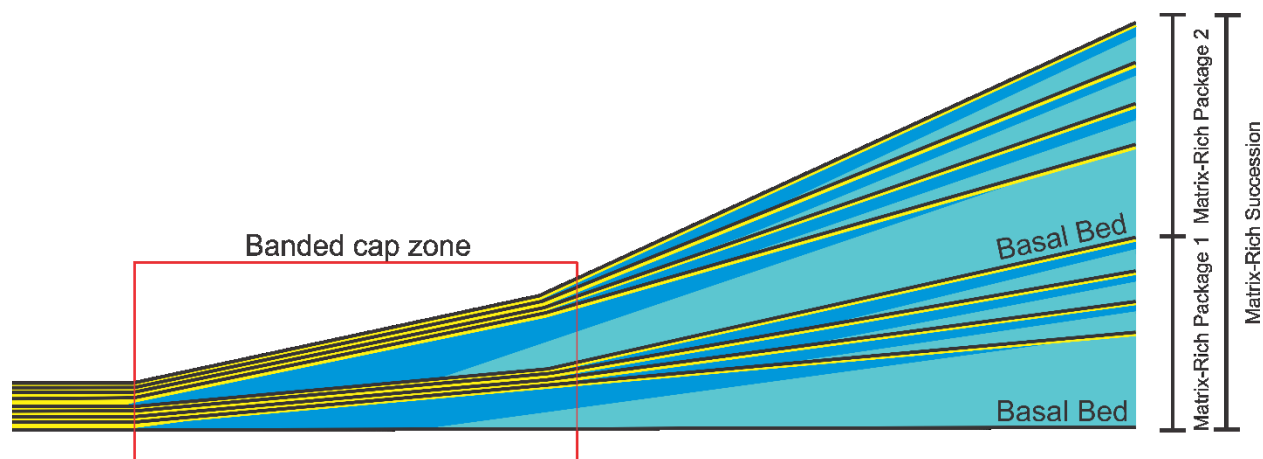


Figure 35. Example of how laminae and/or very thin- to thin-bedded turbidites may stack to form a misinterpreted single depositional event (banded cap).

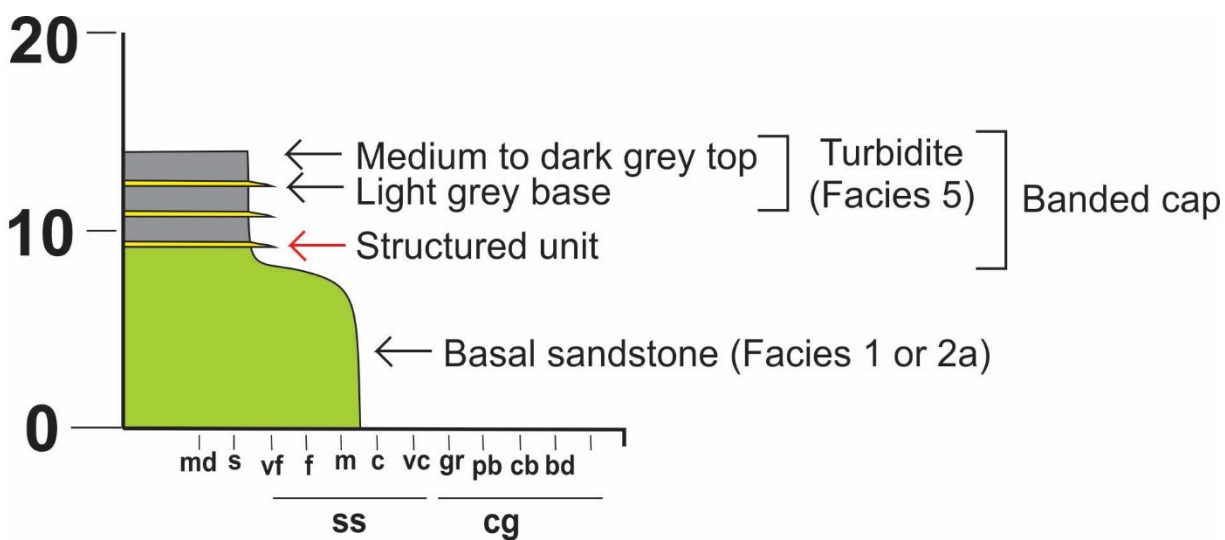


Figure 36. Example of very thin- to thin-bedded turbidites stacked on top of a mudstone or claystone cap of the underlying bed. Note that the basal-most turbidite (indicated by the red arrow) is the 'structured unit' described above.

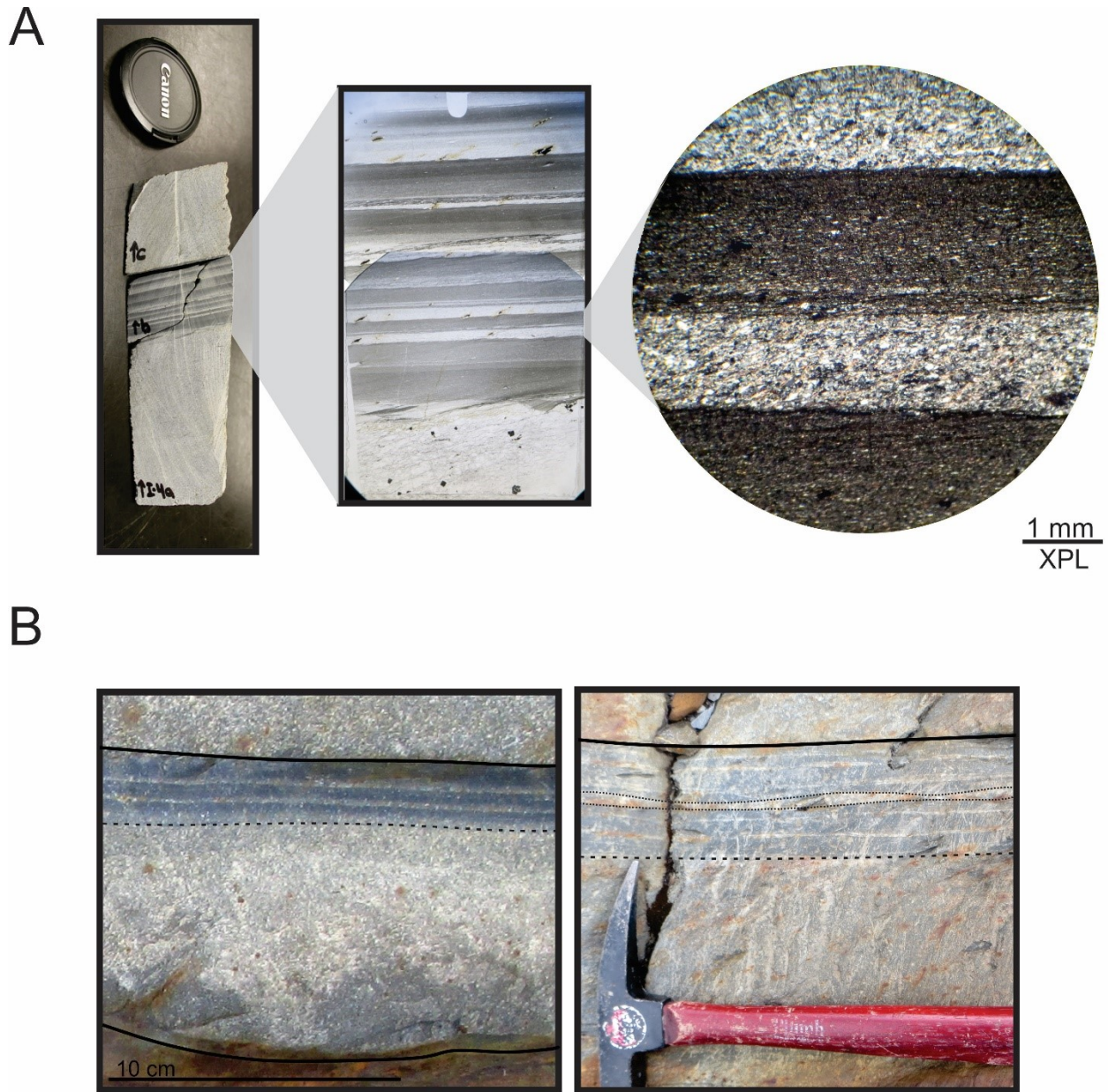
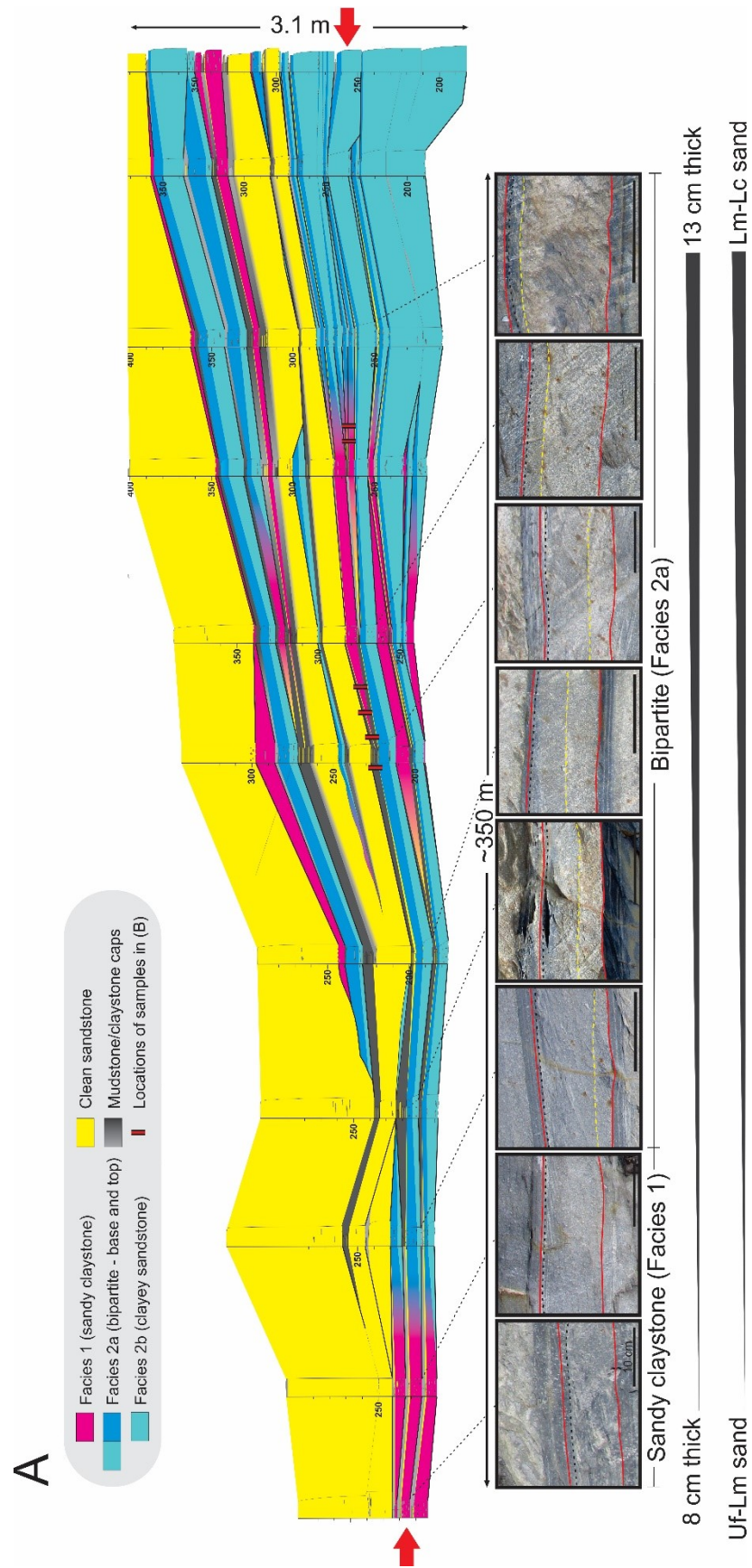


Figure 37. A) Left – Sample of a banded cap. Middle – Photograph of a thin section through the banded cap. Right – photomicrograph of one lamina. Note the better sorting in the light grey base compared to the basal part of a matrix-rich bed (i.e. *Figure 18*). B) Examples of beds with banded caps. Right – some laminae or very thin- to thin-bedded turbidites transition laterally into discontinuous ripple sets (1-3 cm thick). Solid black lines indicate bed bases, and dashed black lines indicate the base of the cap.



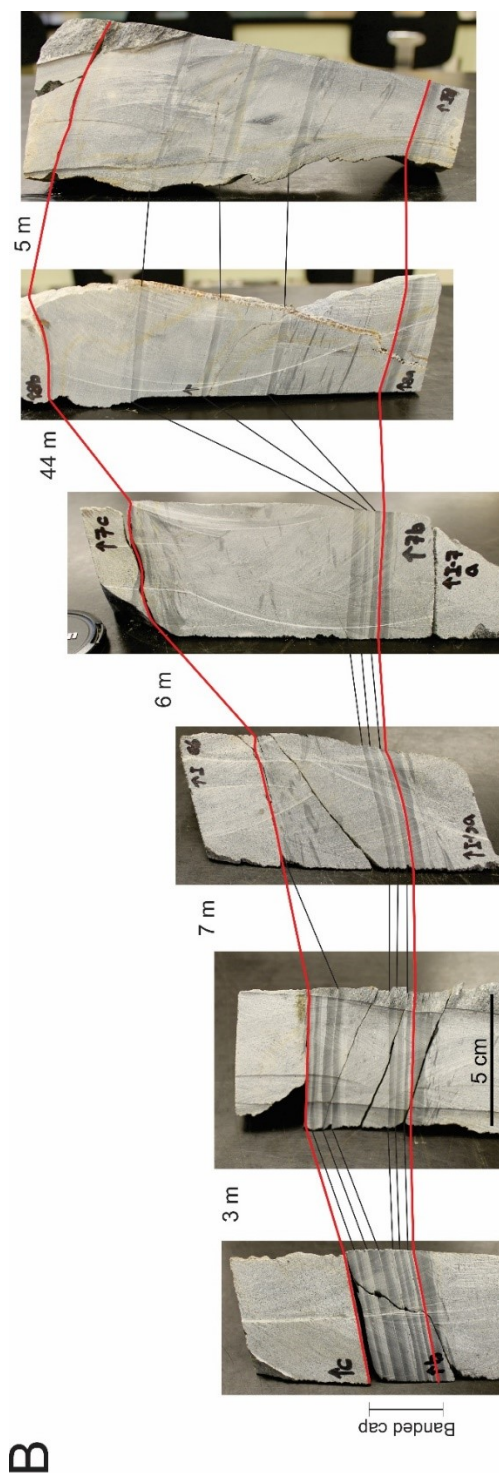


Figure 38. A) Isaac Formation correlation panel. Photographs show lateral trends along one bed horizon. In correlation panel, bed is indicated with red arrows. B) Representative field samples and their lateral correlations (sample locations are shown in (A)). Note the stacking of very thin-bedded turbidites on the far left, which form what appears as a banded cap on the underlying bed, and how each turbidite rapidly expands into an easily recognizable bed toward the right.

4.3 Comparison with Published Mud-Rich Beds

4.3.1 Linked Debrites vs. Bipartite Beds

Linked debrites (mentioned briefly in Chapter 2) (*Figure 39*), which were first described by Haughton et al. (2003), are of particular interest in this study because they superficially resemble bipartite (Facies 2a) beds. Both bed types contain two vertically stacked units, the upper unit being more mud-rich than the basal (*Figure 40*).



Figure 39. Schematic of a linked debrite (redrawn from Haughton et al., 2003) and representative photograph from the Ordovician Cloridorme Formation (Quebec, Canada); red lines indicate bed bases and dashed yellow line indicates the contact between the lower turbidite and upper debrite.

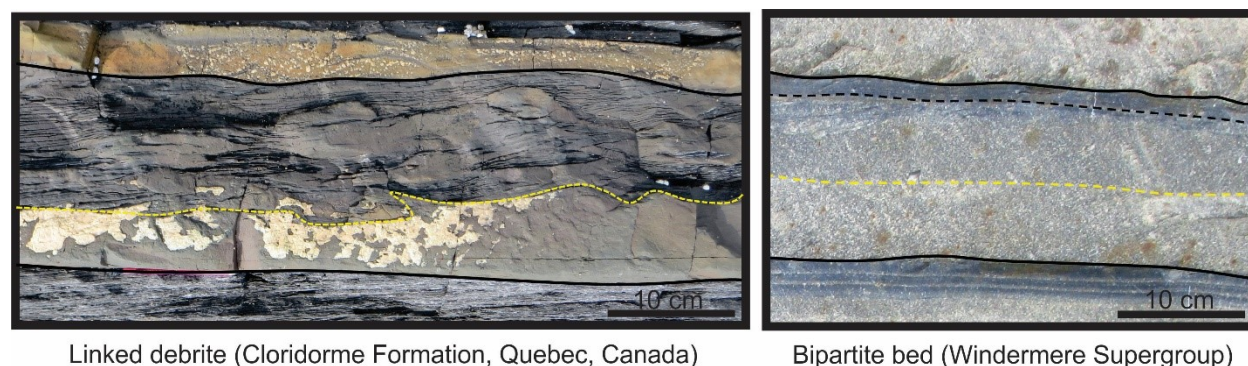


Figure 40. Comparison of a linked debrite and bipartite (Facies 2a) bed. Of particular interest is the difference in contact morphology between the basal and upper unit (indicated by the yellow dashed line), and also the heterogeneous (linked debrite) versus homogeneous (bipartite bed) matrix of the upper unit in each bed. Solid black lines indicate bed bases, and the dashed black

line indicates the contact between the upper sandy claystone unit and cap. Note: the white material in the linked debrite photograph is a superficial calcite precipitate.

Linked debrites are composed of a basal clean turbidite overlain by an upper matrix-rich (~20%) debrite, and therein are beds composed of two parts, or 'bipartite' beds. However, beyond this similarity, they differ in almost all other aspects to Facies 2a beds. Linked debrites are typically several decimeters thick (average 67 cm) and extend for kilometers to hundreds of kilometers longitudinally (Haughton et al., 2003; Talling et al., 2004). The basal turbidite is typically a sharp-based, massive or graded, dewatered, clean sandstone, which then is sharply overlain by the debrite. The contact between the units is typically highly irregular, with up to several decimeter relief and evidence of liquefaction and upward sand injection from the basal turbidite into the upper debrite. The debrite commonly has a heterogeneous matrix (patchy with areas of higher sand content) with abundant clasts of various sizes and lithologies, including exotic lithologies. Importantly, carbonaceous clasts often occur in the upper unit, suggesting that the depositing flows were far-travelled. Clasts are typically oriented oblique to bedding and are highly deformed or sheared. Additionally, the coarsest grain size in the debrite is commonly coarser than the coarsest grain size in the underlying turbidite. Overlying the debrite is a thin, laminated, clean sandstone, and then siltstone and mudstone. Linked debrites are interpreted to form from flows that exhibit a systematic longitudinal change in flow conditions, specifically a fully turbulent headward region followed by a fully laminar body, in turn followed by a fully turbulent tail region.

In contrast, bipartite (Facies 2a) beds are, on average, 25 cm thick and are laterally continuous for only 50-200 m. The basal unit is sharp-based, coarse-tail graded (or, less commonly, massive), contains 10-45% matrix (i.e. it is not clean), is devoid of dewatering

structures, and may contain clasts. An upper, more matrix-rich unit overlies the basal unit with a sharp, planar contact. The upper unit has a homogenous matrix (i.e. no sandy patches), is massive or coarse-tail graded, and contains undeformed, typically similar-sized mudstone or claystone clasts oriented parallel to bedding. The lack of carbonaceous clasts and preservation of mudstone and claystone clasts in both the basal and upper units suggest that they were sourced by local scour of the seabed. Finally, the coarsest grain size in the upper unit is either the same or finer than that in the basal unit, and the matrix content ranges from 20-65%, but is typically >50%. *Table 2* summarizes the differences between linked debrites and bipartite (Facies 2a) beds.

Table 2. Comparison of linked debrites and bipartite beds

Variable	Linked debrite	Bipartite bed
Average bed thickness (cm)	67	25
Lateral extent	Few km's to 100's of km	50-200 m
Primary components	Basal turbidite, upper debrite	Basal clayey sandstone, upper sandy claystone
Dewatering structures?	Present – dish and pipe structures in basal unit	Absent
Clasts?	Present – various sizes, lithologies, orientations in upper debrite; highly deformed/sheared	Present – few cm thick and long, elongate mudstone/claystone in both lower and upper units
Basal to upper unit contact	Sharp, highly irregular with evidence of liquefaction and upward sand injection from basal turbidite, or banded unit (see Haughton et al., 2009)	Sharp, planar; no banded unit
Basal unit matrix content	"Clean" (<20%)	Intermediate (20-50%)
Upper unit matrix content	20%	20-65%
Upper unit matrix characteristics	Heterogeneous (patchy with areas of higher sand concentration)	Homogeneous
Upper unit max. grain size	Often coarser than max. grain size in basal turbidite	Finer or equivalent to max. grain size in basal unit
Overlying turbidite and massive mudstone cap?	Present	Present

4.3.2 Slurry Beds vs. Clayey Sandstone Beds

The second sub-group of matrix-rich beds is less well-represented in the literature and includes slurry beds (*Figure 41*) (first described by Lowe and Guy, 2000). Beds average 4-15 m in thickness, and contain 10-35% matrix and a variety of sedimentary structures, including ubiquitous dewatering pipes and dishes, banding, and soft-sediment deformation features. They are interpreted to be deposited from a flow that is transitional between fully turbulent and fully laminar. These beds are most similar to clayey sandstone (Facies 2b) beds in this thesis, however, many dissimilarities suggest their modes of deposition differ. For example, Facies 2b beds are 40 cm thick on average and lack dewatering structures, banding or soft-sediment deformation features. Most notably, beds of Facies 2b can be shown to extend laterally for less than 100 m before transitioning into another lithofacies (see section 4.1.2 Trend Dimensions). *Table 3* summarizes the differences between slurry beds and clayey sandstone (Facies 2b) beds.

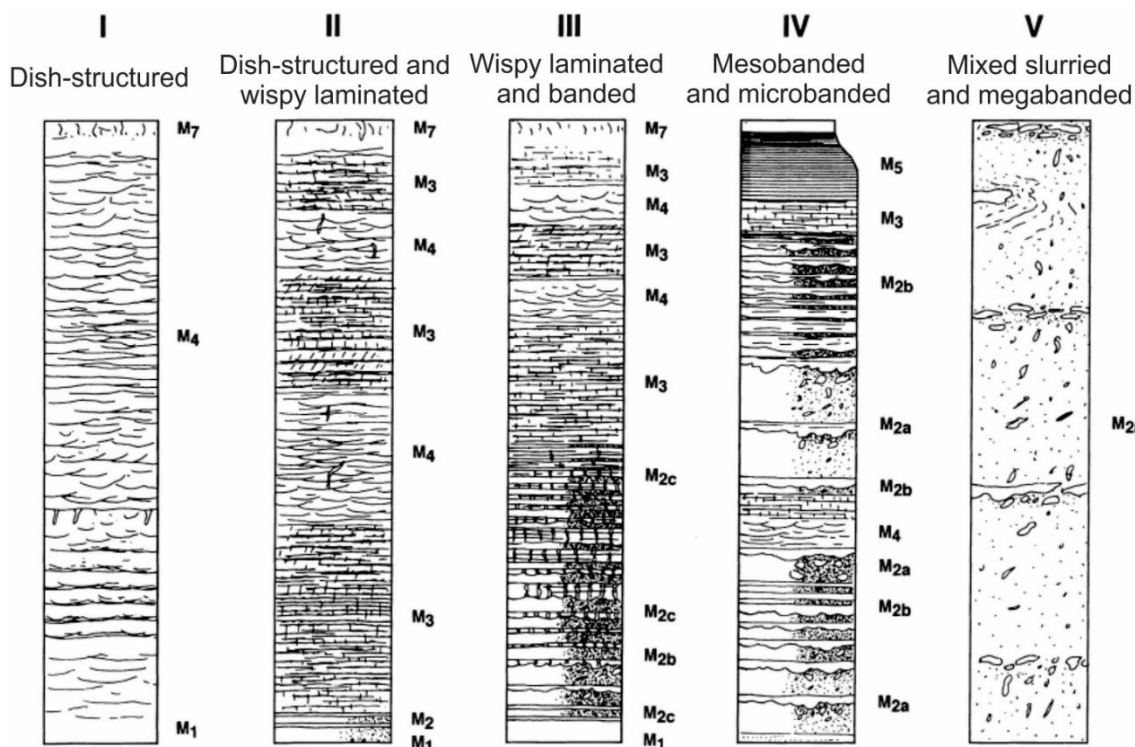


Figure 41. Schematic of different types of slurry beds (from Lowe and Guy, 2000).

Table 3. Comparison of slurry beds and clayey sandstone beds

Variable	Slurry bed	Clayey sandstone bed
Average bed thickness	~1 m to >50 m; most are 4-15 m	40 cm
Matrix content	10-35%	10-45%
Dewatering structures?	Present – dish and pipe structures	Absent
Clasts?	Absent – unless foundered	Present – not common; few cm thick and long, elongate mudstone/claystone
Matrix-rich/matrix-poor banding?	Present	Absent
Soft-sediment deformation?	Present	Absent

4.3.3 Significance of Comparisons

Matrix-rich rocks at Castle Creek share few similarities with the mud-rich sub-groups discussed above (linked debrites and slurry beds), which suggests that they represent a third sub-group of deep-marine matrix-rich bed with its own unique mode of deposition. Slurry beds differ in almost every aspect, except matrix content, from clayey sandstone beds. Significantly, they are much thicker which may suggest deposition in a more confined basin (Bill Arnott, 2015, personal communication). Most importantly, however, is the comparison between linked debrites and bipartite beds. Castle Creek bipartite beds have not yet been documented in the literature which means that, in modern core samples or ancient deep-marine outcrops, two-part beds with more matrix-rich upper parts would likely be interpreted as linked debrites. This is especially true for core samples, where limited lateral continuity hinders the ability to observe local lateral changes. Distinguishing between the two bed types has important implications in the petroleum industry. Linked debrites are much more extensive beds (they can be correlated longitudinally for kilometers to hundreds of kilometers), and are interpreted to form primarily on the most distal margins of sub-marine fans (Talling, 2013). In contrast, bipartite beds are laterally continuous for only 50-200 m before they transition to another facies, and are interpreted to form on both the

slope and proximal basin floor. Misinterpreting a bipartite bed as a linked debrite would not only confine the depositional location, but would also mistake a baffle (bipartite bed) as a barrier (linked debrite) in the petroleum system.

5 Discussion

5.1 Lateral Transect: Interpretation of a Single Bed

5.1.1 Summary of Observations

The following points will summarize important information presented in previous chapters. For a summary table of the results presented in Chapter 4, see *Appendix F*.

- A. Along a lateral transect, beds transition from clayey sandstone (Facies 2b) to bipartite (Facies 2a) with the development of a sharp, planar-based upper more matrix-rich (clayey sandstone or sandy claystone) unit. Laterally, within the bipartite zone of the transition, the basal unit progressively thins whereas the upper unit thickens slightly or remains a constant thickness. Eventually, the basal unit pinches out and the bed consists entirely of sandy claystone (Facies 1), which then fines and thins laterally until it too pinches out. Across this entire lateral transect, a thin, matrix-poor cross-stratified or parallel- to wavy-laminated unit directly overlies the upper more matrix-rich (clayey sandstone or sandy claystone) layer, and in turn is overlain by a mudstone or claystone cap.
- B. The lateral transect of matrix-rich beds, namely from a clayey sandstone (Facies 2b) bed to a very thin- to thin-bedded turbidite (Facies 5), occurs over approximately 200-650 m. The bipartite (Facies 2a) part of the lateral transect is 50-200 m wide, and the sandy claystone (Facies 1) part is 50-280 m. Laterally, beds thin progressively, giving them a wedge-shaped geometry.
- C. Basal contacts show a consistent evolution from sharp and scoured in more sand-rich strata (Facies 2b) to sharp and planar in more matrix-rich strata (Facies 1); the depth of scour decreases progressively over the length of the lateral transect.

- D. The modal grain size of the basal layer (*Figure 30*) fines slightly laterally (from medium-/coarse-grained to fine-/medium-grained), however, the range of grain sizes remains approximately constant across the lateral transect. The basal layer is wedge-shaped (35-55 cm in proximal parts and decreases to 0 cm laterally).
- E. The modal grain size of the upper layer (*Figure 30*) is fine to medium sand, and remains constant laterally. Once developed, the upper layer maintains a thickness of a few to 10 cm, but increases in thickness to a maximum of 10-15 cm (over 40-100 m laterally) after the pinch-out of the basal layer, which is subsequently followed by progressive thinning until it too pinches out.
- F. The basal layer is typically coarse-tail graded, and the upper layer is either massive or coarse-tail graded. The grain size at the base of the upper layer is either equal to or slightly finer than that at the top of the basal layer.
- G. In proximal parts of the basal layer, matrix content is 10-30%, and laterally either remains constant or increases to 20-45%. In proximal parts of the upper layer, matrix content is 20-55%, and laterally either remains constant, or increases to 30-65%. Beyond the pinch-out of the basal layer, matrix content of the upper layer increases to a maximum of approximately 65-80%, and changes little up to the pinch-out of the bed.
- H. Clasts are generally absent in the most proximal part of the lateral transect (i.e. clayey sandstone beds; Facies 2b), but are present in bipartite (Facies 2a) and sandy claystone (Facies 1) beds. Clasts initially crop out near the base of beds, but laterally become located progressively higher in the bed.

- I. The clean (<10% matrix content) structured unit and mudstone or claystone cap maintain a consistent grain size, fine- to medium-grained, and silt and clay or clay, respectively, and consistent thickness, both <1 to a few cm across the full lateral transect.
- J. In the study area, silt and coarser grains are composed predominantly of quartz ($\rho = 2.65 \text{ g/cm}^3$), and minor (0-10%, but most commonly 0-5%) feldspar, specifically plagioclase ($\rho = 2.67 \text{ g/cm}^3$). Due to the typically limited abundance of plagioclase, differences in grain density in silt-size and coarser grains are not considered significant when modelling particle deposition. However, when modeling matrix-rich bed deposition, the density differences between the clay mineral portion of the matrix and other particles must be considered. As discussed in Chapter 3, the low-grade metamorphic alteration of rocks in the study area has caused all primary detrital clay matrix to be recrystallized to muscovite and chlorite. Thus, the mineralogy, grain size, and density of the detrital clay are unknown. Although the absolute density of the clay particles cannot be determined, a general density range can be inferred from common clay mineral densities (2.08-2.77 g/cm^3), and also from the location of clay minerals in bipartite (Facies 2a) beds. The segregation of particles in these beds, namely the higher percentage of matrix (and therefore clay mineral particles, which make up 80-100% of the total matrix percentage) in the upper unit, suggests that clay mineral particles within the flow were, indeed, less dense ($\rho < 2.65 \text{ g/cm}^3$) and/or smaller (<0.0625 mm) than the sand-grade (or coarser) quartz particles. Additionally, differences in shape between platy clay particles and spherical quartz grains may also have enhanced segregation effects (see more detailed discussion below).

K. Facies 1 (sandy claystone) and 2a (bipartite) matrix-rich beds are unlike Castle Creek debrites, which have a very different composition, texture, and overall architecture (discussed in Facies 1. Interpretation). Also they are dissimilar to other matrix-rich beds discussed in the literature, which generally fall into two sub-groups: (1) linked turbidite-debrite deposits, and (2) transitional or slurry flow deposits (see section 4.3 Comparison with Published Mud-Rich Beds).

5.1.2 Settling Behaviour in Suspensions

The lateral transitions studied in this thesis are interpreted to have been controlled primarily by particle settling behaviour. Thus, published works that have studied particle settling will be first briefly described, and later applied to the observed facies transitions.

In low concentration suspensions, interaction between adjacent particles is negligible, and particles settle independently according to their size and density. In polydisperse suspensions, the largest and densest particles, which have the highest settling velocities, tend to settle first, followed by smaller and/or less dense particles. Graded and well-sorted deposits are typically the product of progressive settling from polydisperse suspensions. In monodisperse suspensions, all particles fall to the bed at the same rate, and accordingly emplace an ungraded deposit.

As concentration increases adjacent particles begin to increasingly interact. They continue to settle according to their size and/or density but, as a result of increased concentration, secondary modes of particle segregation become increasingly important. These effects are especially noticeable in suspensions with a wide range of grain sizes or densities, and even more profoundly in suspensions with bimodal distributions of grain size or density:

- (1) Each settling particle displaces upward an equal volume of fluid, known as the return flow (Davies, 1968). If the return flow is sufficiently strong and particles are not too far- or near-spaced, smaller or lighter particles, along with the fluid, will be moved upward (i.e. elutriated) through the spaces between settling larger or denser particles, and will be concentrated at the top of the suspension (Davies, 1968). In this way, the return flow creates shear as it moves past the particle, and as a consequence “hinders” individual particles from settling at their terminal fall velocities (i.e. reduced particle settling; what is termed ‘hindered settling’).
- (2) Each particle in a suspension has an upward buoyant force ($F_B = \rho_f v_f g$; where F_B is the buoyant force, ρ_f the density of the displaced fluid, v_f the volume of the displaced fluid, and g is acceleration due to gravity, 9.8 m/s^2) acting on it, which is equal to the weight of the fluid being displaced by the particle. If the weight of a submerged particle equals the weight of the displaced fluid (the buoyant force), the particle will become neutrally buoyant with no tendency to rise or sink. However, if the weight of a particle is greater or less than the buoyant force, the particle will sink or rise within the fluid, respectively.

A suspension containing particles of uniform size and density would not be expected to produce a suspension with distinct settling boundaries. Similarly, a suspension with a continuous spectrum of grain size and density would also prevent obvious grain settling boundaries from forming. In contrast, a suspension containing a bimodal distribution of particle size or density, would be expected to produce a stratified suspension with a sharp, horizontal boundary separating a basal zone of larger and/or denser particles from an upper zone composed of smaller and/or lighter particles (Richardson and Meikle, 1961).

The ability to segregate particles of different sizes and densities, and therefore produce a graded or segregated deposit, decreases as the absolute concentration of the suspension increases (Middleton and Southard, 1977). At a particular (high) concentration, also known as the critical concentration, segregation is completely suppressed due to particle interlocking (Davies, 1968) or hydrodynamic effects (Lockett and Al-Habbooby, 1974). Interlocking occurs where particles are hindered from moving past one another and the upward elutriation of the pore fluid-particle mixture by the return flow is suppressed due to the reduced space between particles (set up by the higher concentration). Under these conditions all particles settle together and form poorly sorted, ungraded beds. Interestingly, Davies (1968) found that critical concentration decreases as the mean size or density of the suspension fractions become more similar, and also in suspensions with more irregularly-shaped particles (Davies, 1968). Alternatively, the critical concentration can be explained using hydrodynamic principles. Consider a segregated suspension where larger particles make up the lower zone and smaller particles the upper zone, and both particles are of equal density. The lower boundary will fall faster than the upper since it contains grains with higher settling velocities. It is possible, however, to tweak the suspension so that both boundaries fall at the same rate, and thus the entire suspension settles as one. According to equation (1):

$$\text{relative fall velocity} = \text{terminal fall velocity} \times (1 - \text{concentration})^{n-1} \quad (1)$$

where n is 2.39 under turbulent settling conditions (Richardson and Zaki (1954) – the relative fall velocity of a particle decreases as concentration is increased (Lockett and Al-Habbooby, 1974), hence particles with higher terminal fall velocities require greater concentrations if they

are to fall at the same speed as particles with lower terminal fall velocities. Thus, if an appropriate concentration difference between two particle fractions in a suspension is met, the relative fall velocity of both particle types can approach one another and the suspension settles as if made up of equal size and density particles. In this model, the critical concentration depends on the relative proportion of each fraction in a mixed suspension. Based on all of the above, it seems that the critical concentration is dependent on the mean size and density, concentration, and shape of particles in each fraction of a suspension.

5.1.3 Cohesive Particles in Suspensions

All phenomena described above explain the different types of particle settling in non-cohesive particle suspensions. However, matrix-rich beds contain non-cohesive particles (quartz and feldspar), and a matrix that typically is 80-100% clay minerals (i.e. cohesive particles). Amy et al. (2006) have noted that, regardless of particle type (cohesive or non-cohesive), polydisperse suspensions typically produce well-graded and sorted beds if the concentration is below a limiting value. Grains settle according to their terminal fall velocities and experience the same modes of segregation, regardless of kind (i.e. non-cohesive or cohesive). Nevertheless, the presence of clay has important implications on the settling behaviour of suspensions. Fines, especially clay mineral particles, increase the density and viscosity of a suspension, which in turn decreases the settling velocity of coarser particles (Middleton and Southard, 1977). Additionally, as mentioned in Chapter 2, electrostatic surface charges allow clay particles to form aggregates that settle more quickly than their individual constituent clay particles. Aggregation becomes especially important in turbulent flows as particle concentrations increase. Turbulence facilitates particle collisions and increased concentrations bring cohesive clay particles closer together,

allowing for more frequent interaction, and ultimately aggregate formation (Winterwerp, 2002; Amy et al., 2006). However, if turbulence becomes too intense, delicately bound particle aggregates become disaggregated.

At higher concentrations, aggregates join and form an extensive interconnected network throughout the suspension and imparts to it (matrix) strength. This concentration is known as the gelling point (Winterwerp, 2002), after which the settling of dispersed coarser particles becomes reduced or, depending on particle size, density, and shape, may become completely suppressed (Amy et al., 2006).

5.1.4 Depositional Interpretation

In the lateral transect described in Chapter 4, proximal sand-rich, typically scour-based (Facies 2b) beds, which are thicker, coarser grained, and less matrix rich than laterally adjacent matrix-rich beds, are interpreted to be formed closest to the high-energy axis of the depositing flow. These strata are succeeded laterally by bipartite (Facies 2a) and then sandy claystone (Facies 1) beds, and finally very thin- to thin-bedded turbidites (Facies 5) on the most distal fringe. It should be noted that, because of limited exposure, beds only thicken in a single direction. However, it is assumed that flows are more or less symmetrical, causing the axis-to-margin lateral trend to be mirrored on the opposite (covered) side of the flow axis. In other locations in the Castle Creek study area, matrix-rich beds have been shown to be continuous with matrix-poor coarse-grained (or coarser) sandstone that commonly terminate along several-meter-deep matrix-poor sandstone scour fills (Terlaky, 2014; Terlaky and Arnott, 2014; Nataša Popović, 2015, personal communication).

Due to their common occurrence below large-scale stratal elements (such as terminal splays, distributary channels, and slope channels), matrix-rich successions at Castle Creek are thought to indicate activation of the local sedimentary system by upflow channel avulsion (Terlaky and Arnott, 2014). Escape of the high-energy avulsing jet flow scoured the mud-rich seafloor and charged the already sand-rich flow with mudstone clasts and fine-grained (clay and silt) sediment (*Figure 42 – 1* and *2*). The jet flow continued to flow longitudinally away from the avulsion node and shear between the faster, denser suspension and surrounding ambient seawater led to the formation of Kelvin-Helmholtz instabilities along the top and margins of the flow (in fact, a vortex sheet between two immiscible fluids moving with relative motion) (*Figure 42 – 3*). With time the sheet “rolls” up into a discrete layer of individual rotating, semi-spherical fluid mass. Uncommonly, a rotating volume of sediment-laden fluid became detached from the margins of the main jet flow, forming a mini isolated plume that moved at a high angle away from the main flow (*Figure 42 – 3*).

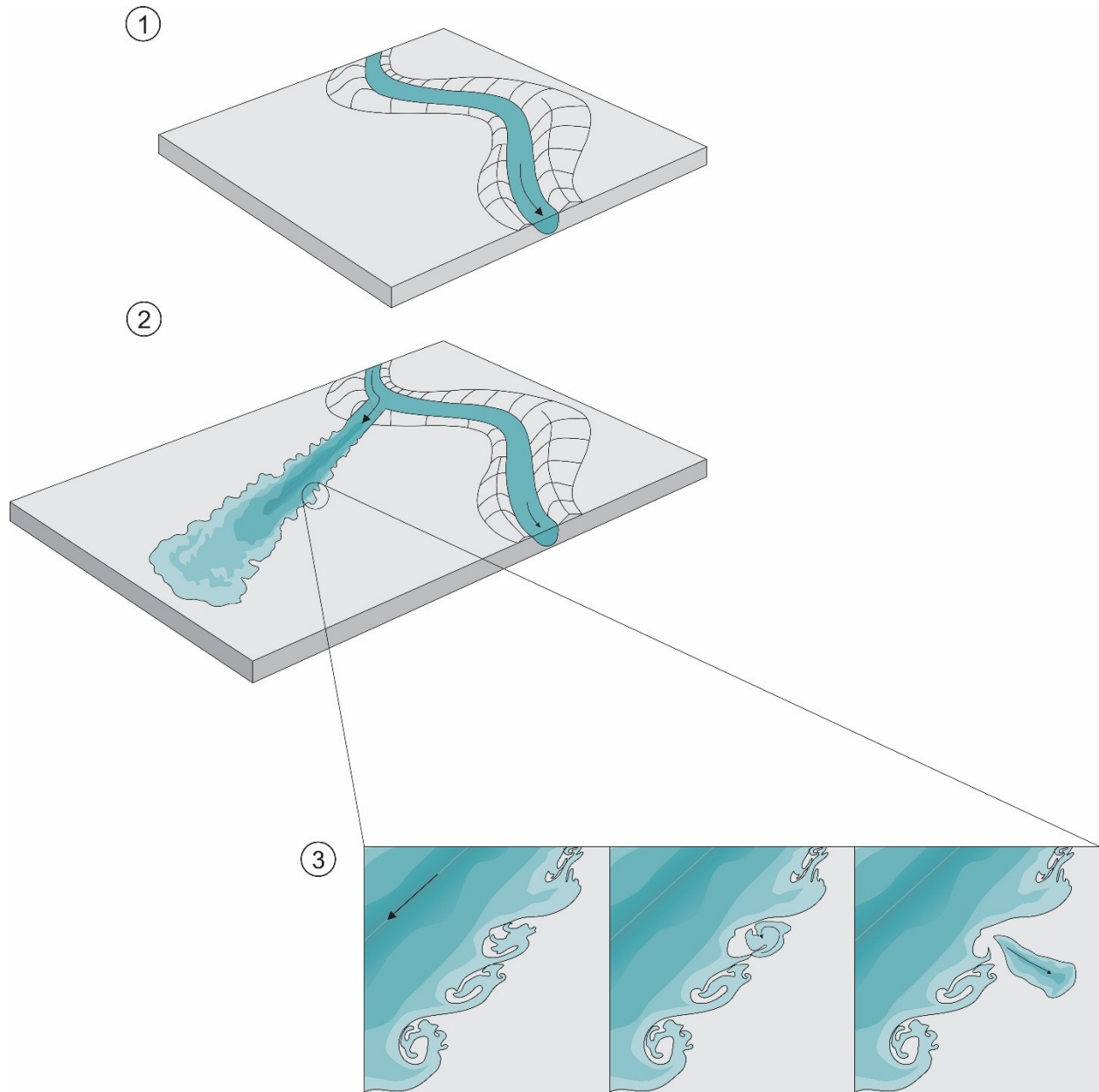


Figure 42. Schematic illustrating the detachment of a sediment-laden plume from the lateral margin of the main jet flow. Diagram is not to scale. Black arrows indicate flow direction. (1) Plan view of a subaqueous leveed-channel (adapted from Davis, 2011). (2) Upflow breach of the levee leads to flow re-direction. Sediment-laden fluid enters the interchannel area as a high-energy jet flow, which retains its initial velocity longitudinally for up to 4 to 6 outlet widths (Rowland et al., 2009). (3) Kelvin-Helmholtz instabilities form along the margins of the jet and, as the jet continues to flow longitudinally, instabilities roll up into a continuous layer of rotating, semi-spherical fluid mass. Uncommonly, a rotating volume of fluid becomes detached and forms an isolated, momentum-driven sediment-laden plume that moves away from the edge of the jet flow. Shear along the base and margins of the plume caused it wane and particles to settle progressively and systematically from suspension.

Once detached, this now completely momentum-driven flow began to rapidly wane, and soon thereafter particles began to differentially settle, which in turn led to the progressive deposition of the lateral transect described in Chapter 4. Close to the site of detachment, high energy and intense internal shearing formed the scours that are observed at the base of clayey sandstone (Facies 2b) beds (Detachment zone in *Figure 43*). Additionally, the high sediment concentration and internal shearing tended to hinder differential particle settling causing fine sediment to be distributed more or less uniformly throughout the thickness of the flow. The retention of fine sediment in the near-bed region was probably also enhanced by high fallout rates and elevated concentrations of sand-grade or coarser particles that further prevented fines from being preferentially displaced upward, and ultimately resulted in the deposition of coarse-tail graded units typically with 10-30% matrix (Kuenen, 1966; Banerjee, 1977; Amy et al., 2006). Beds with lower matrix content were probably formed from flows with lower initial concentrations of clay and silt. Further laterally, lower total sediment concentration (due to earlier fallout and flow expansion), in addition to waning flow conditions and reduced internal shearing, produced more shallowly scoured bed bases and allowed sand-grade or coarser grains to settle preferentially toward the bed (Differential settling zone in *Figure 43*). As coarser grains fell toward the bed, finer grains were displaced upward, eventually forming a sharp, planar-based upper unit with higher matrix content. Sand-grade or coarser grains that did not settle into the basal coarse-tail graded unit became “stranded” in the upper part of the suspension – in fact some of the coarser grains likely became neutrally buoyant as the density and viscosity of the fine-grained upper part of the suspension increased. The slight lateral increase in matrix content in both the basal and upper unit probably reflects the on-going preferential loss of coarser grains

laterally. As coarser grains became less abundant, fewer settled, therefore fewer fines were segregated upward.

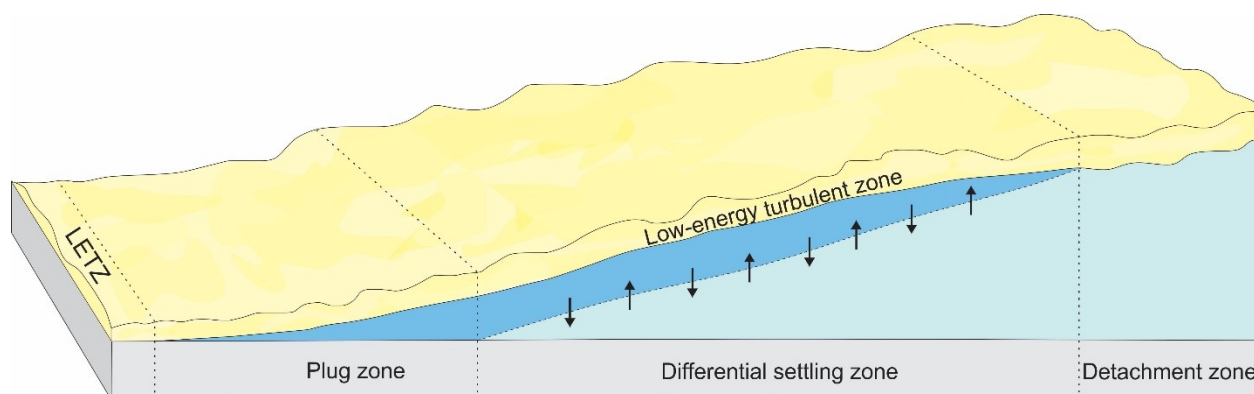


Figure 43. Schematic illustrating the main depositional zones that form the lateral lithological trends discussed in Chapter 4. Clayey sandstone (Facies 2b) beds are deposited in the detachment zone, bipartite (Facies 2a) beds within the differential settling zone, sandy claystone (Facies 1) beds within the plug zone, and matrix-poor structured units in the low-energy turbulent zone (LETZ). The dashed line within the differential settling zone indicates the segregation boundary that separates a basal zone of mostly coarser grains and an upper zone of mostly finer grains. Arrows indicate differential movement of particles within the suspension; finer grains are elutriated upward as coarser grain settle.

Eventually the flow becomes depleted of most of its coarser-grained particles. The basal unit pinches out as the flux of settling sand becomes negligible and the upper, matrix-rich part becomes the entire flow. At this point in the lateral transition, the density profile of the flow is probably plug-like, which results in reduced interfacial mixing with the overlying fluid (i.e. seawater) and reduces shearing in the lower part of the flow. Together these changes serve to reduce energy loss and allow the flow to typically run-out for up to 50-150 m beyond the pinch-out of the basal unit. Reduced basal shear also reduces turbulence in the lower part of the flow, which becomes manifest also by flat basal contacts. Progressive deceleration over the plug zone (sandy claystone; Facies 1; see *Figure 43*) allows increased interaction between adjacent clay minerals, and the production and preservation of cohesive networks between clay particles,

which would further reduce the settling velocities of the remaining coarser grains. The progressive deposition of sediment from the base of the flow upward results in the gradual thinning, and ultimately pinch-out of a sandy claystone (Facies 1) bed.

At all times, the detached sediment-laden plume is overlain by a low-concentration, fine-grained suspension. Upon the termination of the detached plume, the overriding suspension continues to move away from the axial jet. This dilute flow reworks the top of all previously deposited sediment and forms a very thin- to thin-bedded turbidite, which because of high flow mobility, extends beyond the terminus of the sandy claystone (Facies 1) zone in the lateral transect (low-energy turbulent zone in *Figure 43*). The final depositional episode is marked by the deposition of a “clean” (absence of sediment coarser than silt) mudstone or claystone, which probably is the manifestation of a coagulated fluid mud layer.

5.1.5 Clast Location

Clasts are introduced into the main jet flow via seafloor scour of previously deposited mudstone, claystone, and sandstone (matrix-poor or matrix-rich). As reported by Terlaky (2014), large, often angular sandstone or mudstone clasts (several cm thick and long, up to about 15 cm thick and 1 m long; uncommonly several dm thick and several m long) are deposited near the site of scour and occur near the base of beds (Terlaky, 2014). Smaller mudstone or claystone clasts are generally absent. These are probably formed later by the disintegration of the large clasts and then transported laterally and away from the site of erosion.

Initially, sediment concentrations were sufficiently high in the detached plume that low density clasts were hindered from rising upward in the suspension. Further laterally, reduced sediment concentration and the onset of significant differential settling, allowed small, low

density clasts to become progressively displaced upward into the middle and then top parts of the basal layer, and ultimately the mud-rich upper layer.

5.1.6 Interpretation of Stacking Patterns

As discussed in Chapter 4, matrix-rich beds commonly form vertical successions that exhibit a systematic stacking of matrix-rich lithologies. Specifically, the basal bed in each succession is the most laterally extensive. Successively upward beds show a systematic narrowing, suggesting deposition from progressively lower energy flow events. Collectively, this trend is interpreted to indicate deposition from discrete detached sediment-laden plumes, but which all are associated with a single but progressively waning discharge event. Although there are a few vertical trends outlined in Chapter 4, future work is needed to describe additional examples of each, and then to compare and contrast them. Specifically, a thicker stratigraphic section must be measured to determine if these trends in fact represent a recurring theme.

5.1.7 Contributions

Although bipartite beds had been identified at Castle Creek during past field work and studies (Bill Arnott, 2014, personal communication; Terlaky, 2014), no associated lateral trend was observed due to poor lateral exposure within the studied outcrops. This made it difficult to determine if these beds were deposited from one (i.e. the basal and upper unit were deposited by the same flow) or two event(s) (i.e. the basal and upper unit were deposited by separate flows). The lateral lithological changes documented in this thesis undoubtedly indicate deposition from a single event.

This study and others carried out at Castle Creek (i.e. Terlaky, 2014; Popović, 2016) suggest that there is a spectrum (or possibly several unrelated sub-groups) of matrix-rich bed types within the Windermere turbidite system. In general, the beds in this thesis contain more matrix than those in Terlaky (2014) and Popović (2016). Significantly, the less matrix-rich beds of other studies do not undergo the lateral trends described here. Specifically, bipartite beds do not form. This, then, suggests that each bed type reflects its own unique depositional mechanism. Future work will determine how matrix-rich beds at Castle Creek are related to one another, and to those formed in other deep-marine systems.

6 Conclusions

Although mud-rich strata have become widely recognized in both the modern and ancient deep-marine sedimentary records, it is becoming increasingly evident that much work is needed to differentiate the various different kinds of mud-rich sedimentary rocks and link them to their unique modes of deposition. The purpose of this thesis, therefore, was to investigate and describe matrix-rich rocks at Castle Creek. More specifically, it was to identify any lateral changes in their lithological make up in order to gain a better understanding of how they were deposited, and how the related depositional flows evolved. Detailed micro- and macro-scale analysis of many matrix-rich beds at Castle Creek has elucidated a consistent lateral trend that is thought to reflect deposition on the margins of an avulsed high-energy jet flow. In the most proximal part of the lateral transect, clayey sandstone (Facies 2b) beds are typically amalgamated, coarse-tail graded, medium- to coarse-grained, medium- to thick-bedded, scour-based, and contain 10-30% matrix. Laterally, these beds progressively transition to bipartite (Facies 2a) beds with the development of a sharp, planar-based upper more matrix-rich (clayey sandstone or sandy claystone) unit. The most proximal bipartite (Facies 2a) beds are typically coarse-tail graded or massive, fine- to medium-grained, medium- to thick-bedded, scour-based, and contain a basal unit with 10-30% matrix, and an upper unit with 20-55% matrix. Laterally, the basal unit progressively thins whereas the upper unit either thickens slightly or remains constant. Matrix content in the basal and upper units either remains constant or increases slightly. The most distal bipartite (Facies 2a) beds are coarse-tail graded or massive, fine- to medium-grained, thin- to medium-bedded, more planar based, and contain a basal unit with 10-45% matrix, and an upper unit with 20-65% matrix. Eventually, the basal unit pinches out and the bed consists entirely of sandy claystone (Facies 1). These sandy claystone (Facies 1) beds are initially massive or coarse-

tail graded, fine- to medium-grained, thin- to medium- bedded, planar-based, and typically contain >50% matrix content. Laterally, these beds fine and thin until they too pinch out. Across this entire lateral transect, a thin, matrix-poor cross-stratified or parallel- to wavy-laminated unit directly overlies the upper more matrix-rich (clayey sandstone or sandy claystone) unit, and in turn is overlain by a mudstone or claystone cap. Notably, this cleaner upper layer and cap extend beyond the pinch-out of the sandy claystone (Facies 1) bed, and correlate with a very thin- to thin-bedded turbidite in the distal part of the lateral transect.

This lateral continuum of lithofacies reflects an evolution of flow structure that is interpreted to be controlled primarily by particle deposition, namely sand, from a detached volume of sediment-laden fluid (originating from a Kelvin-Helmholtz instability) formed on the lateral margins of an avulsed jet flow. Soon after detachment, the momentum-driven plume begins to wane resulting in rapid suspension fallout of mostly sand and the deposition of medium- to thick-bedded, coarse-tail graded clayey sandstone (Facies 2b) beds. Progressive lateral loss of sand-grade or coarser grains (and lesser finer grains) from particle fallout led to the lateral thinning and slight fining of strata, in addition to a progressive enrichment in clay and silt. The differential settling of coarser grains caused fines to become preferentially segregated into the top of the flow, forming an upper matrix-rich part that thickened slightly or remained constant laterally. Accordingly, bipartite (Facies 2a) beds were deposited in this zone of differential particle settling. Progressively laterally, the more sand-rich basal part of the flow thinned and eventually pinched out as the flux of settling sand became negligible and the upper, matrix-rich part became the entire flow. Here, sandy claystone (Facies 1) beds were deposited in the plug zone. At all times, the detached sediment-laden plume was overlain by a low-concentration, fine-grained suspension which, upon termination of the detached plume, reworked

the top of all previously deposited sediment and deposited a very thin- to thin-bedded turbidite. Finally, a mudstone or claystone cap was deposited via the coagulation of a fluid mud layer.

Where matrix-rich beds are present, they tend to stack on top of one another in vertical successions that range from 20-150 cm thick and contain between 2-15 beds. Within a succession, beds typically show a systematic change in lateral extent upward. Packages, contained within matrix-rich successions, comprise a basal, more laterally extensive matrix-rich bed overlain by beds with progressively reduced lateral extent. This stacking pattern suggests multiple detachment events, but all associated with the gradual waning of a single discharge event. Significantly, matrix-rich strata in both slope (Isaac Formation) and proximal basin floor (upper Kaza Group) strata exhibit a similar lateral succession of matrix-poor to matrix-rich lithologies, as well as a similar vertical stacking pattern, suggesting that their formation is not paleogeographically, but rather mechanistically controlled.

6.1 Areas for Future Work

- A. Continue to document the detailed characteristics of matrix-rich strata at Castle Creek and compare them to examples elsewhere in the Windermere, in geological units elsewhere (e.g. Cloridorme Formation, Quebec), and from the literature.
- B. Only general lateral grain size trends are described in this thesis. In order to more accurately capture the trend, a more systematic sampling of matrix-rich beds is needed. For example, in a single bed, samples could be obtained from the base, middle, and top at fixed lateral distances. From there, point counting of each grain size would be required. Also, to build a robust, statistically significant dataset, many beds would require sampling. This data would provide a much better idea of how grain size percentages

change laterally, which would provide further insight into how flow characteristics, as gleaned from the depositional record, evolved in both time and space.

- C. Time-intensive manual point counting for matrix content percentages was performed in this thesis. Future work should create an automatic point counting system in order to ensure unbiased data capture and also minimize time expenditure.
- D. Post-depositional diagenetic and metamorphic processes have affected beds at Castle Creek but were not analysed in detail in this thesis. A comparative study of their effects on the textural characteristics of matrix-rich and matrix-poor strata may provide a qualitative tool for estimating the percent matrix content in Windermere strata.
- E. Finally, in order to develop a more accurate depositional model for matrix-rich beds at Castle Creek, specifically of the type described in this thesis, it will be necessary to log a thicker stratigraphic section containing multiple matrix-rich successions. These data would then better elucidate the relationship between superimposed matrix-rich beds in a multiple-bed bedset. Specifically, are superimposed beds genetically related and arranged systematically upward, and if so, are they deposited by a single or multiple flows?

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Appendix A: Map of the Isaac Formation section

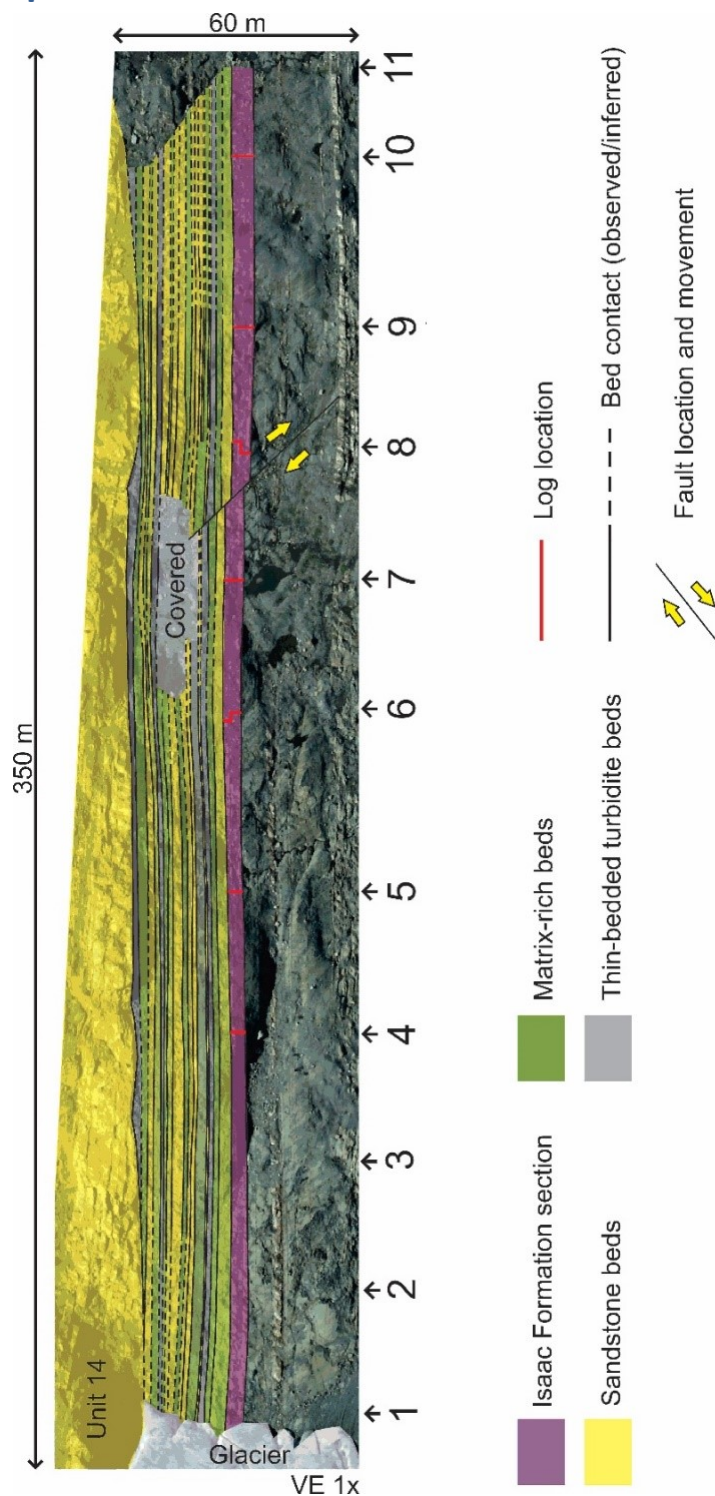
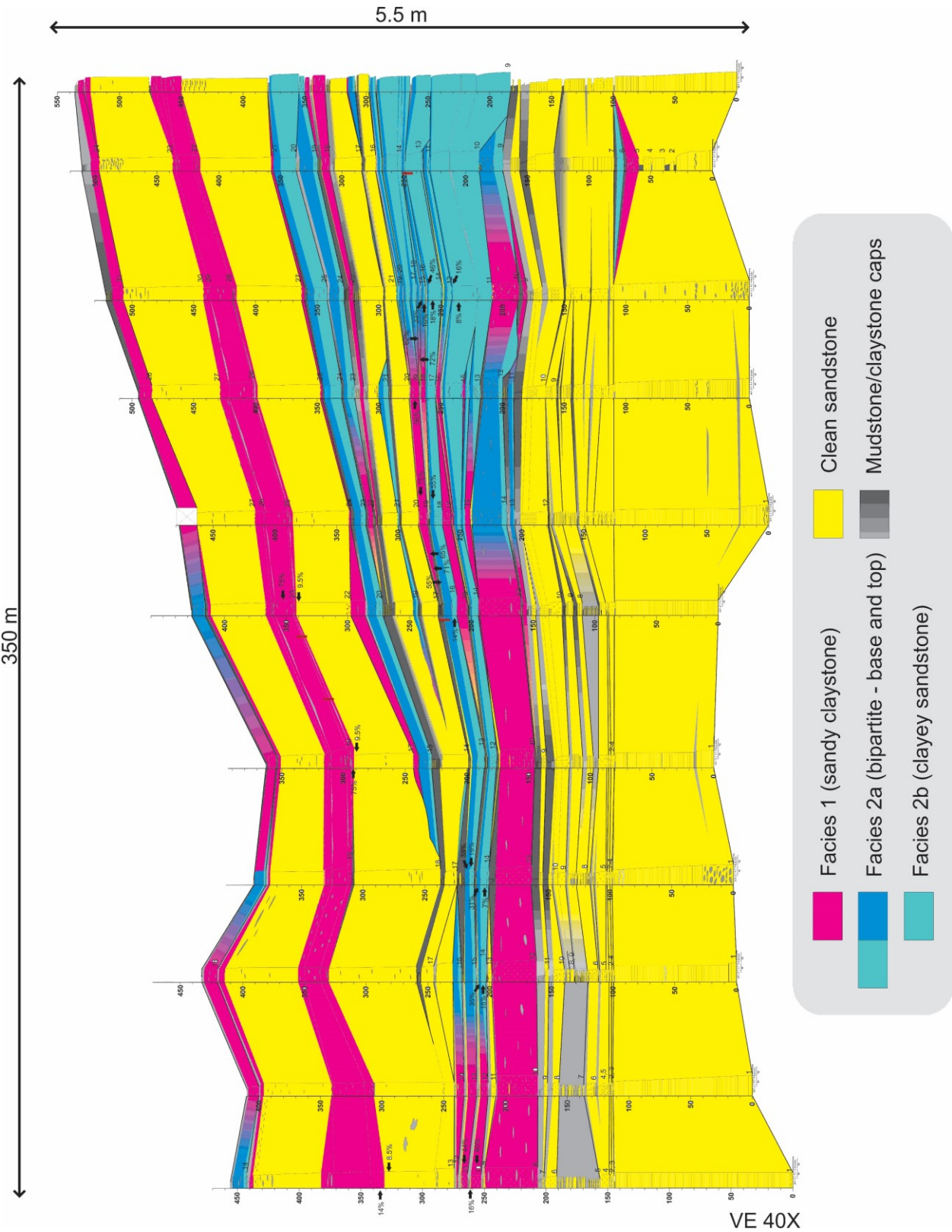
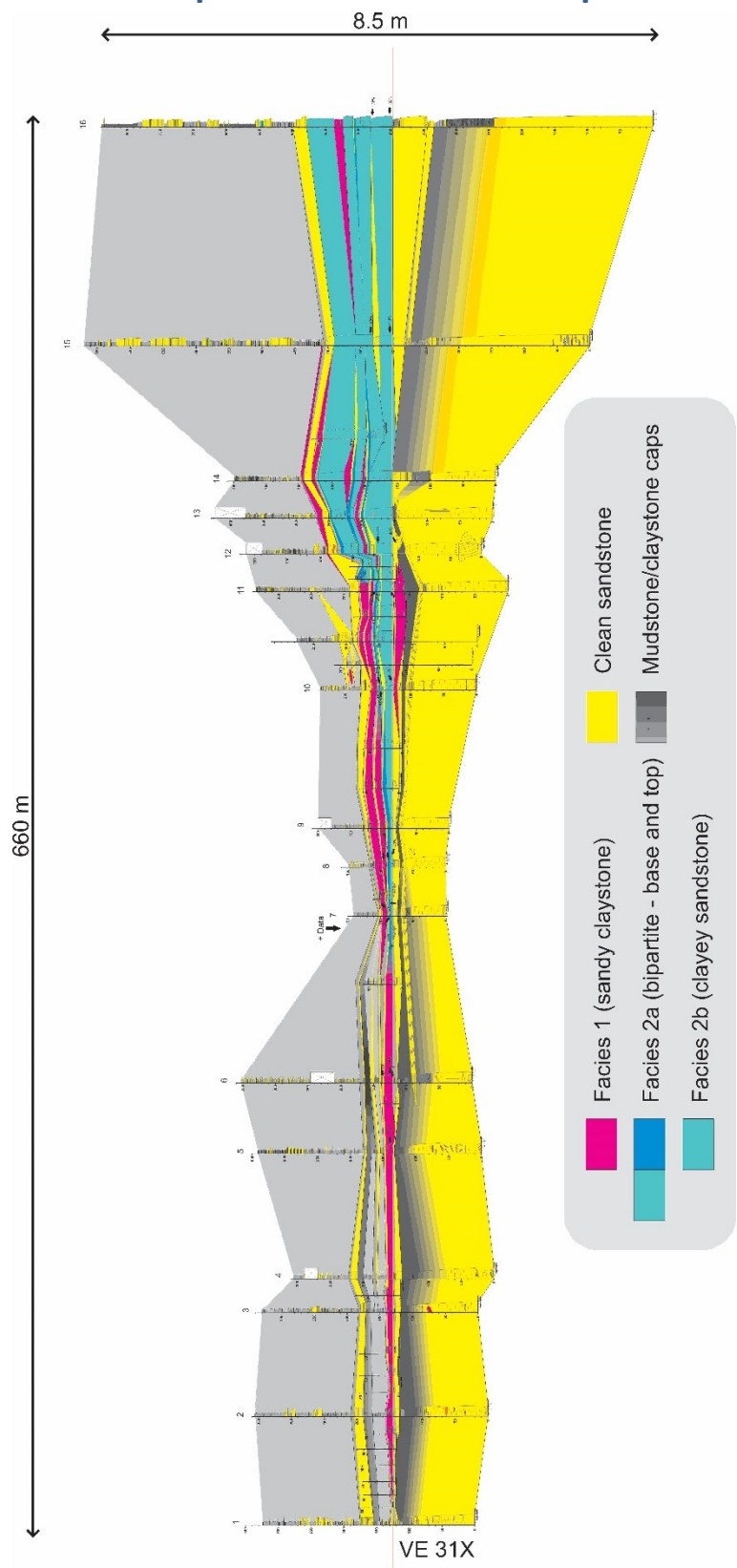


Figure 44. Map showing the location of the Isaac Formation study section (purple) and overlying lithologies (mapped by Viktor Terlaky). Numbers and associated arrows indicate log locations. Unit 14 is a channel (studied by Dumouchel (2015)). VE is vertical exaggeration.

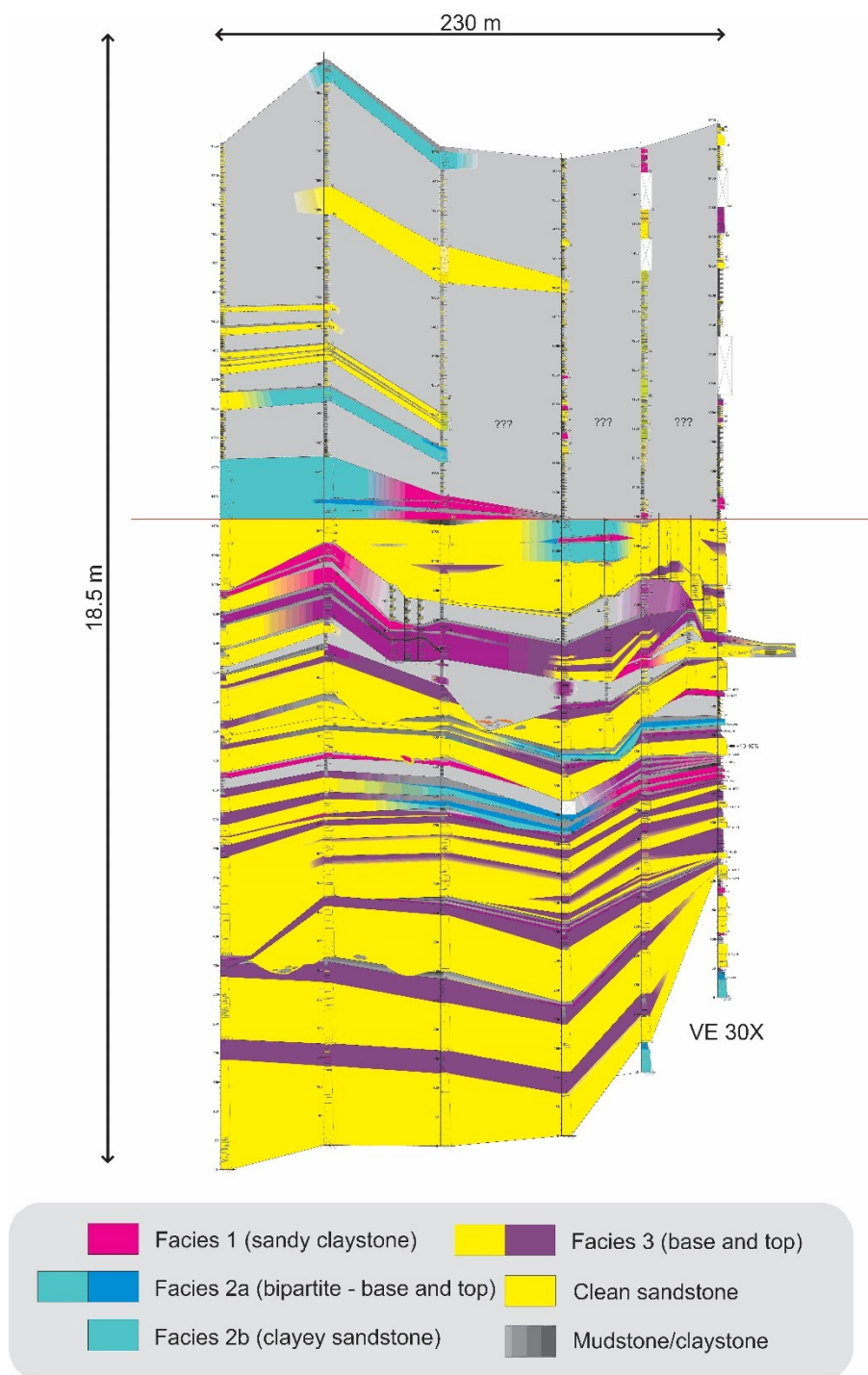
Appendix B: Correlation panel of the Isaac Formation section



Appendix D: Correlation panel of the Kaza Group section



Appendix E: Preliminary map and correlation panel of the stratigraphically lower Kaza Group section (not discussed in this thesis)



Appendix F: Summary of lithological data discussed in Chapter 4

VARIABLE	FACIES 5	FACIES 1	FACIES 2a	FACIES 2b
Bed type	Turbidite	Sandy claystone	Bipartite	Clayey sandstone
Location in lateral transect	Distal	Distal intermediate	Proximal intermediate	Proximal
Width of facies zone	N/A	50-280	50-200	Covered
Overall bed thickness range	0.1-5 cm	2-25 cm	3-45 cm	25-55 cm
Basal contact	Sharp, planar	Sharp, planar or undulatory	Sharp, undulatory or planar	Sharp, undulatory or planar
BASAL LAYER				
Basal layer thickness			0.2-43 cm	16-55 cm
Overall grain size range (excluding matrix)			Upper very fine to upper very coarse sand +/- 2-4 mm granules	Lower fine to upper very coarse +/- 2-4 mm granules
Modal grain size range (excluding matrix)			Lower fine to upper medium sand	Lower medium to upper coarse sand
Texture			CTG or massive	CTG or massive
Matrix content			10-45%	10-45%
Matrix components			80-100% clay and 0-20% silt	92-96% clay and 4-8% silt
Colour			Light to medium grey	Light to medium grey
Clast presence			20% of basal units	8% of basal units
Clast composition			Mst, cst, uncommon sst	Mst, cst, uncommon sst
Clast size range			<1-few cm x 3-20 cm;	1-5 cm x 3-20 cm;

and shape		elongate	elongate
Clast orientation		Sub-parallel to bedding	Sub-parallel to bedding
Clast volume		5% of unit volume	<1% of unit volume
Clast location		Mid-unit or at/slightly below bipartite interface	Base of unit or mid-unit
UPPER LAYER			
Upper layer thickness	1-15 cm	0.5-10 cm	
Contact with underlying unit	N/A	Sharp, planar	
Overall grain size range (excluding matrix)	Lower very fine to upper very coarse sand; rare 2-4 mm granules	Lower very fine to upper very coarse sand +/- 2-4 mm granules	
Modal grain size range (excluding matrix)	Lower fine to upper medium sand	Lower fine to upper medium	
Texture	Massive or CTG	Massive or CTG	
Matrix content	50-80%	20-65%	
Matrix components	80-97% clay and 3-20% silt	75-100% clay and 0-25% silt	
Colour	Medium to dark grey	Medium to dark grey	
Clast presence	85% of upper units	40% of upper units	
Clast composition	Mst, cst, uncommon sst	Mst, cst, uncommon sst	
Clast size range and shape	<1-few cm x <1-20 cm; elongate	<1-few cm x 0.5-47 cm; elongate	
Clast orientation	Sub-parallel to bedding	Sub-parallel to bedding	

Clast volume	5-30% of unit volume		5-15% of unit volume	
Clast location	Evenly dispersed through unit		Evenly dispersed through unit	
TURBIDITE				
Turbidite thickness	0.1-3 cm	0.2-3 cm	0.2-5 cm	1-8 cm
Contact with underlying unit	Sharp, planar	Sharp, undulatory (0.1-1 cm)	Sharp, undulatory (0.1-1 cm)	Sharp, undulatory (0.1-1 cm)
Structure type	Massive, ripple cross-stratified or planar-laminated	Ripple cross-stratified or planar- to wavy-laminated	Ripple cross-stratified or planar- to wavy-laminated	Ripple cross-stratified or planar- to wavy-laminated
Overall grain size range (excluding matrix)	Lower very fine to lower very coarse sand	Lower very fine to lower very coarse sand	Lower very fine to upper very coarse sand	Lower very fine to upper very coarse sand; rare 2-4 mm granules
Modal grain size range (excluding matrix)	Silt to upper fine sand	Lower very fine to upper medium sand	Lower very fine to upper medium sand	Lower very fine to upper medium sand
Matrix content	<10%	<10%	<10%	<10%
CAP				
Cap thickness	0.03-1.5 cm	0.2-5 cm	0.1-6 cm	<1 cm; typically scoured
Contact with underlying unit	Sharp, planar or fills ripple topography	Sharp, planar or fills ripple topography	Sharp, planar or fills ripple topography	Sharp, planar or fills ripple topography
Modal grain size range	Mst or cst; ungraded	Mst or cst; ungraded	Mst or cst; ungraded	Mst; ungraded

Key: CTG = coarse-tail graded; mst = mudstone; cst = claystone; sst = sandstone

Note: 'Basal' and 'upper' layer refer to those shown in *Figure 30*.