

Vertical and Lateral Facies Architecture of Levees and Their
Genetically-Related Channels, Isaac Formation, Neoproterozoic
Windermere Supergroup, Cariboo Mountains, B.C.

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Abstract

At the Castle Creek study area, levee deposits are well-exposed over an area of ~2.6 km wide and ~90 m thick. This provides an opportunity to describe their lateral and vertical lithological changes, and accordingly details about their reservoir geometry and stratal continuity. Here, levee deposits are divided vertically into packages, each consisting of a sand-rich lower part overlain sharply by a mud-rich upper part. Each lower part displays a consistent thickening then thinning trend laterally away from its genetically related channel. The characteristics of these packages suggest that they were controlled by recurring changes in the structure of channelized flows, which in turn was controlled by grain size and grain sorting. This ultimately was controlled by short-term changes in relative sea level. Moreover, some mud- and sand-rich strata are rich in residual carbon suggesting that mid-fan levees can serve as source rocks for hydrocarbon generation, and also reservoirs.

Résumé

Les dépôts de levées sont bien exposés sur une région de $\sim 2,6$ km de largeur et de 90 m d'épaisseur à la région d'étude Castle Creek, ce qui permet de décrire les changements lithologiques latéraux et verticaux et, ainsi, de bien décrire leur géométrie de réservoir et leurs continuités stratales. Ici, les dépôts de levées sont divisés verticalement en paquets, chacun composé d'une partie inférieure riche en sable, recouverte abruptement par une partie supérieure riche en boue. Chaque partie inférieure démontre un épaississement suivi de l'amincissement constant en s'éloignant latéralement de son canal sous-marin associé. Les paquets verticaux sont interprétés comme ayant été formés en raison des changements périodiques dans la structure de densité verticale des flux canalisés (contrôlés principalement par la taille et le tri des grains), ce qui a été finalement contrôlé par des changements dans le niveau relatif de la mer. De plus, certaines couches en boue, et en sable, sont riches en carbone résiduel, ce qui suggère que les dépôts de levées peuvent servir comme roches-mères et réservoirs aux hydrocarbures.

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Table of Contents

Abstract	ii
Résumé.....	iii
Acknowledgements.....	iv
List of Figures	viii
List of Tables	xx
1 Introduction.....	1
1.1 Thesis Rationale	1
1.2 Geological Setting and Regional Stratigraphy	1
1.3 Castle Creek Study Area, Cariboo Mountains	8
1.4 Previous Work.....	14
1.5 Methodology	15
1.5.1 Field Work	15
1.5.2 Petrography	16
1.5.3 Analytical Techniques: Total Organic Carbon Content and Stable Carbon (^{13}C) Isotopes	16
2 Deep-water sedimentation processes and their associated deposits	18
2.1 Mass-movement Deposits	18
2.2 Sediment-Gravity Flows	19
2.2.1 Cohesive Flows.....	21
2.2.2 Frictional Flows	22
2.2.2.1 Hyperconcentrated Density Flows.....	23
2.2.2.2 Concentrated Density Flows.....	23
2.2.2.3 Turbidity Currents	24
3 Submarine Fans.....	30

3.1	Overview of Deepwater Systems	30
3.2	Architectural Elements	32
3.3	Channels	33
3.4	Levees.....	38
4	Facies Descriptions and Interpretations	44
4.1	Facies 1 (F1): Structureless, Normally Graded Sandstone	44
4.1.1	Interpretation.....	45
4.2	Facies 2 (F2): Traction Structured Sandstone	46
4.2.1	Interpretation.....	49
4.3	Facies 3 (F3): Medium-Scale Cross-Stratified Sandstone	55
4.3.1	Interpretation.....	57
4.4	Facies 4 (F4): Chaotic Mud-Rich Deposits.....	58
4.4.1	F4a: Poorly-sorted mudstone-rich deposits	58
4.4.2	Interpretation.....	65
4.4.3	F4b: Deformed Bedded Deposits.....	66
4.4.4	Interpretation.....	70
4.5	Petrography	73
4.6	Carbonate Cements	77
4.7	Total Organic Carbon Content	79
4.8	Stable Carbon (^{13}C) Isotopes of Organic Material	82
5	Channel and Levee Descriptions and Interpretations	84
5.1	Channels.....	85
5.1.1	Channel Unit 1 (CH1).....	86
5.1.2	Channel Unit 2 (CH2).....	87
5.1.3	Channel Unit 3 (CH3).....	88

5.1.4	Channel Unit 4 (CH4).....	89
5.1.5	Channel Unit 5 (CH5).....	89
5.1.6	Channel Complex Interpretation.....	92
5.2	Levees.....	93
5.2.1	Type 1 Levee Deposits: Description of Vertical Trends	95
5.2.2	Type 1 Levee Deposits: Description of Lateral Trends	99
	Levee Packages 1 and 2: southeast drainage and southern outcrops	100
	Levee Packages 1 and 2: northern outcrop	105
	Levee Packages 3 and 4: southern outcrop	110
	Levee Packages 3 and 4: northern outcrop	112
5.2.3	Interpretation of Vertical Packaging in Type 1 Levee Deposits.....	114
5.2.4	Interpretation of Lateral Trends in Type 1 Levee Deposits.....	123
5.2.5	Type 2 Levee Deposits: Description.....	130
5.2.6	Type 2 Levee Deposits: Interpretation.....	132
5.2.7	Type 3 Levee Deposits: Description.....	135
5.2.8	Type 3 Levee Deposits: Interpretation.....	136
6	Discussion and Conclusions	139
6.1	Temporal and Spatial Relationships between Channel and Levee Sedimentation	139
6.2	Deposition of the leveed-channel complex	150
6.2.1	Time-Stratigraphic Development of Channel-Levee Packages 1 to 4.....	150
6.3	Implications for hydrocarbon exploration.....	155
6.4	Areas for future research	156
7	References.....	157
8	Appendix.....	169

List of Figures

Figure 1.1: Distribution of WSG outcrops (yellow) in western North America..	3
Figure 1.2: Global palaeogeographic maps for the 780-580 Ma interval.	4
Figure 1.3: Collision and accretion of allochthonous terranes along the western coast of Laurentia during the Mid-Jurassic to Early Cenozoic formed the Canadian Cordillera which can be subdivided into five orogen-parallel, morphogeological belts.	5
Figure 1.4: Schematic cross-section showing the interpreted western North American margin during the Neoproterozoic (redrawn by Khan, 2012 from Ross, 1991).	5
Figure 1.5: Stratigraphy of the WSG in the Cariboo Mountains of the southern Canadian Cordillera (from Khan, 2012 modified from Ross et al., 1995).	6
Figure 1.6: Geologic map of the WSG in the southern Canadian Cordillera, including location of the Castle Creek study area (green circle). Modified from Young et al., 1973; Murphy and Ross, 1988; Ross and Arnott, 2007).	9
Figure 1.7: Regional geology of the Castle Creek study area is outline in blue (from the Eddy 1:50,000 map area; Ross and Ferguson, 2003a).	10
Figure 1.8: Palinspatic restoration of the WSG makes it comparable in size to the modern Amazon and Mississippi fans. For comparison, other modern fans (black text) and ancient fans (red text) are shown (redrawn by Khan, 2012; inset map of the WSG from Ross, 2001, modified from Barnes and Normark, 1985).	10
Figure 1.9: A) Photomosaic of the Castle Creek study area. This study documents the stratigraphy of levee deposits associated with base of slope channels in the Isaac Formation. The three outcrops of this study are outlined in red boxes. B) Interpreted geologic map of the Castle Creek study area (after Ross and Arnott, 2007). Isaac channel complexes 1-6 (ICC1-6) are numbered on the left. The channel-levee complex of this study crops out between ICC4 and ICC5 on the lower slope and is outlined by the red boxes.	13
Figure 1.10: General stratigraphic nomenclature of strata exposed at Castle Creek. Levee deposits of this study are located between Isaac Channels 4 and 5 (modified from Ross and Arnott, 2007).	14
Figure 2.1: Initially, gravity-driven sediment failure at the shelf edge forms coherent slides and slumps. Over time, these deposits can evolve into debris flows and turbidity currents.	19

Figure 2.2: Schematic of the simplified classification scheme of Mulder and Alexander (2001). They classified sediment-gravity flows into two types: cohesive flows and frictional flows, where frictional flows are further subdivided into hyperconcentrated density flows, concentrated density flows and turbidity currents.....	20
Figure 2.3: The structure of a turbidity current consists of a well-defined head, body and tail. ..	26
Figure 2.4: Most turbidity currents are vertically density-stratified, with grain concentration highest near the base of the flow decreasing exponentially upwards.	26
Figure 2.5: Idealized Bouma sequence for low-density turbidity currents showing the typical grain size profile and sedimentary structures (redrawn by Angus, 2016 from Bouma, 1962).	28
Figure 2.6: Idealized Lowe sequence for high-density turbidity currents showing the typical grain size profile and sedimentary structures (redrawn by Angus, 2016 from Lowe, 1982).....	29
Figure 3.1: Sediment-gravity flows transport large volumes of clastic material from the shelf down the slope to the ocean floor where it is deposited in large-scale depositional systems called submarine fans (from Meiburg and Kneller, 2010).	31
Figure 3.2: Idealized depositional model for a point-source, mixed sand-mud submarine fan. The Windermere Supergroup deep-water turbidite system most likely represents a similar model (from Reading and Richards, 1994).....	32
Figure 3.3: Submarine fans are composed of an assemblage of geomorphic features termed architectural elements. The main architectural elements that make up deep-water environments are shown on this schematic.	33
Figure 3.4: Horizon 3D seismic slice of the top of the of the top of the Green Channel Complex, offshore Angola (Wynn et al., 2007 modified from Abreu et al., 2003).	34
Figure 3.5: Adjustment of a channel profile to maintain equilibrium transport conditions. (A) A channel profile that lies above the graded profile must erode to flatten its profile (Kneller, 2003). (B) A channel profile that lies below the graded profile must aggrade to steepen its profile (Kneller, 2003). (C) When the channel is at grade and the majority of the sediment is bypassed to areas farther downflow (Kneller, 2003; Arnott, 2010). (D) Mixed channels (erosional/depositional) alternate between channel-axis erosion and levee deposition at different locations downflow in order to maintain equilibrium transport conditions (Arnott, 2010). Figure A-C redrawn by Khan, 2012 from Kneller, 2003. Figure D redrawn by Khan, 2012 from Kneller, 2003 after Samuel et al., 2003.	36

Figure 3.6: The combined channel classification scheme proposed by Navarro (2005) to describe the components that make up submarine channels.	38
Figure 3.7: Three-dimensional view of the Joshua channel belt, Gulf of Mexico. Note the well-developed levees formed along both sides of the channel.	40
Figure 3.8: Seismic profile across a channel-levee complex in the Amazon fan. Channel deposits are between the white lines. Note how the levee deposits on either side of the channel exhibit a gull wing-shaped geometry away from the channel margins (from Piper and Normark, 2001). .	40
Figure 3.9: Process of flow stripping. Flow momentum causes the upper, more dilute, finer-grained portion of the flow to become detached from the lower, higher-concentrated, coarser-grained part of the flow and to overspill (Fildani et al., 2006; Arnott, 2010). Redrawn by Khan, 2012 from Peakall et al., 2000.	41
Figure 3.10: Process of inertial overspill. The entire flow, including the lower coarser-grained portion, may overspill the outer channel bends due to particle inertia (Khan, 2012). Khan, 2012 after Kane et al., 2007.	42
Figure 3.11: Process of continuous overspill. Where throughgoing flows are thicker than the depth of the channel the portion of the flows that are thicker than channel margins overspill and build up levees on both sides of the channel (Arnott, 2010). Khan, 2012 after Peakall et al., 2000; Kane et al., 2007.	42
Figure 3.12: Walker (1985) reported that levee strata commonly exhibit three attributes: climbing ripples, convolute lamination and mud rip-up clasts, terming them CCC-turbidites (Posamentier and Walker, 2006).....	43
Figure 3.13: Thin-bedded turbidites in the Chatsworth Sandstone, California with two of the three “Cs”. Orange arrows indicate the base of each bed, yellow arrows indicate climbing ripple sets and convolute lamination is highlighted in blue. Note the absence of mud clasts (Posamentier and Walker, 2006).....	43
Figure 4.1: Photos of very thick-bedded, structureless sandstones grading upward from lower coarse- to lower medium-grained sand.	45
Figure 4.2: Photos of F2 strata.....	48
Figure 4.3: Hypothetical bedform stability diagram for low concentration turbidity currents (redrawn from Davis, 2011 after Southard and Boguchwal, 1990). As flow velocity decreases for	

bedload sediment coarser than middle fine sand, the Tb division (upper plane bed field) should be succeeded by dune cross-stratification before entering the ripple stability field.	50
Figure 4.4: Turbulence model of Best (1992) focused on larger aggregate vortex structures to build up initial bed defects of sufficient height that they could influence local flow conditions and lead to the development of angular bedforms.	51
Figure 4.5: Hydrodynamic instability model of Venditti et al. (2006) for the initiation of bed defects that lead to the development of angular bedforms.	52
Figure 4.6: Top: High-speed, time-lapse photography showing how two turbidity currents of similar concentration and velocity but different grain sizes evolve at a point over time (modified from Tilston's experiments, 2014). The flow in the upper panel is composed of well-sorted medium sand (0.33 mm sand) compared with lower fine sand (0.15 mm) in the lower panel. Bottom: Graph showing the vertical density structure of the coarse- and fine-grained flows shown above (modified from Tilston, 2016). Tilston et al. (2015) showed that a steep near-bed density gradient (i.e., coarse-grained flow) is a requisite condition to form bed defects, which then quickly evolve into angular bed forms like ripples, and presumably also dunes.....	54
Figure 4.7: Photos of F3 strata.....	57
Figure 4.8: Photos of D1.....	60
Figure 4.9: Photos of D2.....	62
Figure 4.10: Photos of D3.....	64
Figure 4.11: Photos of D4.....	65
Figure 4.12: Photos of deformed bedded deposits showing the slide scar and deformed strata...	67
Figure 4.13: Photos of deformed bedded deposits showing the subtly deformed nature of beds.	67
Figure 4.14: Various photos of deformed bedded deposits	69
Figure 4.15: CHIRP profile across a straight segment of the modern Villafranca channel, southeastern Tyrrhenian Sea (from Gamberi and Rovere, 2011). Failure along the western steep inner margin of the channel triggered mass transport deposits, which on this seismic image are shown as thick, transparent layers with a rough surface. Rotated slide blocks along the western margin are bounded by extensional faults (black dashed lines).	71
Figure 4.16: Slump deposit in a channel-proximal location within the Upper Cretaceous Rosario Formation, Baja California, Mexico (modified from Kane et al., 2007).	71

Figure 4.17: (A) Aerial photograph showing the slide deposit (contact in red) that crops out in the northern outcrop of this study underlain by the channel fill deposit (contact in yellow) identified by Marion (2005). (B) Photograph taken by Marion (2005) in the field of the area marked by the white rectangle in (A). Slide contact outlined in red, channel in yellow and debrite D1 in blue.....	72
Figure 4.18: Photomicrographs of facies 2 (top) and facies 3 (bottom) strata under cross-polarized light. Quartz consists of the main framework grain.....	74
Figure 4.19: Photomicrographs of strata showing black and opaque wavy graphite laminae under plane-polarized (top) and cross-polarized light (bottom). Note how quartz framework grains are dispersed in the graphite. Matrix consists of metamorphic muscovite and chlorite.	75
Figure 4.20: Photomicrograph of facies 4a under cross-polarized light. This facies is typically poorly sorted and composed primarily of matrix.....	77
Figure 4.21: Carbonate-cemented strata. (A) Carbonate-cemented medium-scale cross-stratified sandstone (F3). (B) Carbonate-cemented Tbcde turbidite. Note the distinctive rusty-orange colour of these strata.	78
Figure 4.22: (A) Red arrows point to clasts of carbonate-cemented convoluted planar-laminated sandstone (F2) within a debrite (F4a). (B) Metre-scale carbonate-cemented dune cross-stratified sandstone clast (F3) within a debrite (F4a).....	79
Figure 4.23: (A) Carbonate-cemented Tbcde turbidite with elevated TOC. Dark-grey laminations are composed of graphite. (B) Photomicrograph under plane polarized light. Quartz framework grains are isolated within wavy black and opaque graphite laminations that are oriented parallel to bedding.....	79
Figure 5.1: Channels and levees were identified in the study areas (red boxes). The channel complex crops out in the southeast drainage outcrop (A). Levee deposits crop out in the southern (B) and northern (C) outcrops.....	84
Figure 5.2: Aerial photograph of the southeast drainage study area (marked A in figure 5.1) where the channel complex crops out. Log locations shown in red. From Terlaky, 2016.	85
Figure 5.3: Aerial photograph of the southern study area (marked B in figure 5.1) where levee deposits crop out. Log locations shown in red.....	85
Figure 5.4: Aerial photograph of the northern study area (marked C in figure 5.1) where levee deposits crop out. Log locations shown in red.....	85

Figure 5.5: The fill of CH1 is composed of coarse- to medium-grained sandstone with abundant granules and pebbles, mudstone, carbonate cemented sandstone and limestone clasts. Photo from Terlaky, 2016.	86
Figure 5.6: (A) Overlying CH2 is a poorly-sorted, mudstone-rich “starry night” debrite (D1-F4a). (B) The “starry night” debrite is overlain locally by a poorly-sorted, carbonate-clast-rich debrite (D2-F4a). Photo from Terlaky, 2016.	87
Figure 5.7: Photo of CH4. Thick-bedded turbidites of the lower fill (left) are sharply overlain by medium- to thin-bedded turbidites of the upper fill. Photo from Terlaky, 2016.	88
Figure 5.8: On the north side of the southeast drainage outcrop, debrite D3 (F4a) plugs CH5 except at the top of the channel unit, where the channel fill onlaps the debrite. Photo from Terlaky, 2016.	90
Figure 5.9: Strata in the southeast drainage outcrop. Top is an aerial photograph showing the contacts between the channel units, levee deposits and mass transport deposits identified in the study area. Bottom is a correlation panel of the strata showing the detailed architecture of the channel complex. Modified from Terlaky, 2016.	91
Figure 5.10: Correlation panel of the strata in all three outcrops of the study showing the detailed architecture of the channel complex. Although separated by moraine (~190 m wide) and the Castle Creek Glacier (~819 m wide), intercalated mass transport deposits, or MTDs, allow strata to be confidently correlated across the full width of the study area -- a distance of 2.6 km. The channel complex crops out in the southeast drainage outcrop (see section 5.1) and the genetically-related levees crop out on the northernmost part of the southeast drainage outcrop, across the southern and northern outcrops.	94
Figure 5.11: (Top) The quality of the northern outcrop is excellent with almost 100% exposure. Correlation panel of extra logs spaced 5-20 m apart to document small-scale lateral changes in individual beds in the lower part of a package. (Bottom) Correlation panel of extra logs spaced 5-20 m in the northern outcrop to document small-scale lateral changes in individual beds in the upper part of a package.	96
Figure 5.12: Type 1 levee deposits are divided vertically into four correlatable stratal packages, two of which are shown here. Packages can be easily subdivided into lower and upper parts, the transition being sharp and marked by a dramatic decrease in bed thickness, grain size and sandstone content.	95

- Figure 5.13: (A) The lower part of packages consists mostly of medium- to thick-bedded, upper medium- to coarse-grained, lower division turbidites intercalated with very thin- to thin-bedded, fine-grained, upper division turbidites. (B) Close up of a thick-bedded, upper medium-grained Tbcd/e turbidite in the lower part of a package. 97
- Figure 5.14: (A) The upper part of packages consists mostly of very thin- to thin-bedded, fine-grained, upper division turbidites intercalated with medium- to thick-bedded, upper medium-grained, lower division turbidites. Yellow arrow points to a medium-bedded, upper medium-grained Tbcd/e turbidite. (B) Close up of thin-bedded, fine-grained Tcd/e turbidites in the upper part of a package. 97
- Figure 5.15: Thickness data of the 295 beds that make up packages 1-4 in the northern outcrop shows how thickness varies upwards. Data is from 1 log and these plots use a 3-bed moving average. Note the dramatic decrease in bed thickness at the contact between the lower and upper parts of a package..... 98
- Figure 5.16: Lateral thickness (black lines) changes in the lower part of packages (LP) that extend from the southeast drainage outcrop (left) across the moraine to the southern outcrop (total lateral distance of 872 m). 100
- Figure 5.17: Lateral change in net-to-gross of the lower parts of packages (LP1 to LP4) extending from the southeast drainage outcrop (far left) across the moraine to the southern outcrop (total lateral distance of 872 m). 101
- Figure 5.18: Lateral thickness (black lines) changes in the upper part of packages (UP) that extend from the southeast drainage outcrop (left) across the moraine to the southern outcrop (total lateral distance of 872 m). 102
- Figure 5.19: Lateral changes in thickness of the lower (black lines) and upper (grey lines) parts of packages extending from the southeast drainage outcrop (left) across the moraine to the southern outcrop (a total lateral distance of 872 m). Note the antithetic thickening and thinning patterns of the lower and upper parts of each package. 103
- Figure 5.20: Lateral change in net-to-gross of the upper parts of packages (UP1 to UP4) extending from the southeast drainage outcrop (UP1 and UP2) (left) across the moraine to the southern outcrop (total lateral distance of 872 m). 104
- Figure 5.21: Lateral changes in thickness of the lower (orange) and upper (light blue) parts of packages extending across the southern outcrop for packages 1 to 4. Note the antithetic

thickening and thinning patterns of the lower and upper parts of each package. Cumulative thickness of Tb turbidites (yellow), individual multiple set Tc turbidites (dark blue), individual multiple set Tc turbidites plus multiple set Tc turbidites overlying Tb turbidites (grey), and Td/e turbidites (green). Note that it is the Td/e turbidites, and, to a lesser extent, the multiple set Tc turbidites that control the antithetic thickening and thinning of the upper parts. 105

Figure 5.22: Lateral thickness (black lines) changes in the lower parts of packages (LP) that extend from the glacier (left) across the northern outcrop (total lateral distance of 194 m).. 106

Figure 5.23: Thickness of select lower division turbidites from LP 1 to 4 in the northern outcrop. Graph on left shows how their thickness changes towards the glacier. Graph on right spaces them out in order to better show the thickness trends. About half of the beds thin towards the glacier while the other half either thicken or remain similar in thickness. Note that in general, beds do not thicken or thin linearly (smoothly). 107

Figure 5.24: Lateral change in net-to-gross in the lower part of packages (LP1 to LP4) that extend from the glacier (far left) across the northern outcrop (total lateral distance of 194 m).. 107

Figure 5.25: Lateral thickness (black lines) changes in the upper parts of packages (UP) that extend from the glacier (left) across the northern outcrop (total lateral distance of 194 m).. 108

Figure 5.26: Lateral change in thickness of the lower (black lines) and upper (grey lines) parts of packages that extend from the glacier (left) across the northern outcrop (total lateral distance of 194 m). Note the antithetic thickening and thinning of LP2 and UP2. Other packages only show subtle antithetic thickening and thinning trends. 109

Figure 5.27: Lateral change in net-to-gross in the upper part of packages (UP1 to UP4) that extend from the glacier (left) across the northern outcrop (total lateral distance of 194 m). Note that UP2 extends laterally for 59 m then becomes truncated by a slide (see fig. 5.10). 109

Figure 5.28: Lateral changes in thickness of the lower (orange) and upper (light blue) parts of packages extending across the northern outcrop for packages 1 to 4. Note the antithetic thickening and thinning patterns of the lower and upper parts of each package. Cumulative thickness of Tb turbidites (yellow), individual multiple set Tc turbidites (dark blue), individual multiple set Tc turbidites and multiple set Tc turbidites overlying Tb turbidites (grey), and Td/e turbidites (green). Note that it is the Td/e turbidites, and, to a lesser extent, the multiple set Tc turbidites that control the antithetic thickening and thinning of the upper parts. 110

Figure 5.29: Thickness of upper division turbidites from UP4 in the northern outcrop. Graph on left shows how their thickness changes towards the glacier (towards the left). Graph on right spaces them out in order to better show the thickness trends. About half of the thin-bedded, upper division turbidites remain similar in thickness, only varying by 1-2 cm in thickness. The other half of upper division turbidites rhythmically thicken and thin by as much as 10 cm.

Interstratified medium-bedded, upper division turbidites also rhythmically thicken and thin. Note that in general, beds do not thicken or thin linearly (smoothly). 114

Figure 5.30: (A) Digital elevation models of the deposits formed by the experimental turbidity currents of deLeeuw et al. (2016). This depositional pattern occurs due to spatial differences in sediment transport/deposition, specifically lower rates of deposition within the axis compared to the margins of the flow. (B) Cumulative digital elevation model of the deposits. Cross-sections show the amount of erosion/deposition on the lower, middle and upper slope. From deLeeuw et al., 2016..... 116

Figure 5.31: The sharp contact at the base of each package, overlain then by coarse-grained turbidites, is interpreted to mark the base of the incipient levees (initial ridges) that formed during channel inception and helped to confine the axial parts of subsequent flows that travelled through the now better-defined channel. White arrows in the photograph point to this sharp contact (highlighted by red lines) at the base of levee packages. 117

Figure 5.32: The velocity maximum is controlled by the vertical density profile of the flow. Experimental work by Tilston et al. (2015) suggests that turbidity currents consist of two end-member vertical density profiles controlled by grain size: 1) a plug-like structure where density changes little upward but then decreases sharply, and 2) an exponential structure with density decreasing continuously upward..... 119

Figure 5.33: The almost exclusively coarse-grained make-up of beds in the lower parts of packages suggest that channelized flows were similarly coarse grained and well sorted. These flows would have had a plug-like density structure with negligible vertical stratification and high flow efficiency. Also, the height of the velocity maximum would have been at roughly half the depth of the flow... 120

Figure 5.34: With time the levee continued to aggrade and conditions in the channel continued to be dominated by flow bypass with negligible net sedimentation. Relief between the channel floor to levee crest progressively increased. Eventually relief between the channel floor to levee crest

- exceeded the height of the velocity maximum of the average channelized flows after which only the upper, fine-grained parts of currents typically overspilled, building up the upper parts, while the lower, coarse-grained parts remained confined to the channel. White arrows point to the sharp contact between the lower and upper parts of the packages marks this condition (highlighted by yellow lines). 121
- Figure 5.35: At some point during deposition of the upper part of the package (white arrows point to upper parts in photograph), the make-up of the sediment supply changed, specifically an increase in the sand fraction. The result was the development of an exponential density structure in throughgoing currents with a velocity maximum height significantly depressed to an elevation well below the height of the levee crest. The exponential density profile promoted interfacial mixing and rapid energy loss, and accordingly deposition within the channel. 122
- Figure 5.36: Schematic showing the initial thickening of a lower part of a package away from the channel margin across the southeast drainage and southern outcrops. 125
- Figure 5.37: In modern systems, the inner walls of sinuous channels are much steeper than the slope on the levee backside, and as a result are more prone to gravitational failure. Top: Slump deposit in a channel-proximal location within the Upper Cretaceous Rosario Formation, Baja California, Mexico (modified from Kane et al., 2007). Middle: Aerial photograph showing the slide deposit (contact in red) that crops out in the northern outcrop of this study underlain by the channel fill deposit (contact in yellow) identified by Marion (2005). Bottom: Photograph taken by Marion (2005) in the field of the area marked by the white rectangle in the aerial photograph. Slide contact outlined in red, channel in yellow and debrite D1 in blue. 126
- Figure 5.38: (A) Very thick-bedded, lower-division, very coarse-grained sandstone turbidite with dispersed quartz and carbonate pebble clasts thins and fines to a medium-bedded, upper medium-grained sandstone over a distance of 300 m toward the glacier. (B) Photo of bed thinning towards the glacier ~190 m from the glacier. (C) Photos of this bed taken over the last 135 m towards the glacier show how it continues to thin and fine. 128
- Figure 5.39: Lateral changes in thickness of the type 2 levee deposit (blue) and the underlying debrite D1 (orange) extending across the southern outcrop. Note the antithetic thickening and thinning patterns of the levee and debrite. Cumulative thickness of Ta turbidites (purple), Tb turbidites (grey), Tc turbidites (yellow), Td/e turbidites (black) and dunes (green) that make up the levee are shown. Note that it is the Tb and Td/e turbidites, and, to a lesser extent, the Ta

turbidites that control the antithetic thickening and thinning of the levee. Tc turbidites and dunes change little in thickness.	131
Figure 5.40: Lateral change in net-to-gross of the type 2 levee deposit extending across the southern outcrop.....	131
Figure 5.41: Lateral change in net-to-gross of the type 2 levee deposit extending across the northern outcrop.....	132
Figure 5.42: In the southern outcrop, dune cross-stratified sandstone is more common in type 2 levee deposits compared to type 1.	133
Figure 5.43: A thick-bedded, upper medium-grained Tbcd/e turbidite marks the base of the type 3 levee deposit (yellow line). This deposit consists mainly of very thin- and thin-bedded, fine-grained Tc turbidites.	135
Figure 5.44: Graph showing how the thicknesses of four thick-bedded, upper medium-grained Tbc turbidites either change negligibly or thicken/thin across the width of the outcrop. Lower division turbidites that do thicken/thin do not do so in any preferred direction.	136
Figure 5.45: Lateral change in net-to-gross in the type 3 levee deposit extending from the glacier (left) across the northern outcrop (total lateral distance of 194 m).....	136
Figure 5.46: Deposition over the distal levee was dominated by finer, more slowly settling particles in which particle settling and subsequent bed-load transport was more or less spatially uniform (Khan and Arnott, 2011). Yellow arrows point to discontinuous ripple trains which suggest conditions of limited sediment availability and low energy (Kane et al., 2007), consistent with a distal levee setting.	137
Figure 5.47: Schematic diagram showing coarse-grained overspilling flows that accelerate down the steep slope of the proximal levee and deposit thick-bedded, coarse-grained, lower division turbidites (yellow) over the more shallowly inclined distal levee. Such deposits were termed overbank splays by Khan and Arnott (2011). Figure modified from Davis, 2011.	138
Figure 6.1: Block diagram (left) and seismic section (right) to show the development of channels in the Nile Delta, Egypt. Note how the channels in the block diagram (red arrow) stack vertically and are laterally offset. Modified from Samuel et al., 2003.	144
Figure 6.2: A three dimensional reconstruction of the two lowermost channel-levee packages in the Windermere channel-levee complex developed from a two dimensional dataset (face of the	

exposed outcrop). Black boxes indicate one possible face of the outcrop-- recall that the outcrop exposure is interpreted to be oriented oblique to the channel margins.	144
Figure 6.3: Depositional model for the channel-levee packages observed in this study.	149
Figure 6.4: Thin-bedded, fine-grained upper division turbidites in the upper parts of packages 3 and 4 are a distinctive red in the field.	152
Figure 6.5: Photomicrograph of a thin-bedded, fine-grained upper division turbidites from the upper part of package 3 in cross-polarized light. Iron-rich carbonate cement makes up approximately 60% of the total stratal volume.	153
Figure 6.6: Aerial photographs showing the interpreted channel, levee, slide and debrite deposits in the study areas (top: southeast drainage outcrop (from Terlaky, 2016), middle: southern outcrop, bottom: northern outcrop). Log locations are shown in red.	154
Figure 6.7: Model showing the time-stratigraphic development of the entire channel-levee complex. Short-term changes in relative sea level were shown to influence the development of the channel-levee packages whereas long-term changes controlled the overall mineralogy of the strata and the occurrence of mass-wasting deposits..	155
Figure 8.1: The five stages of organic-matter oxidation in anoxic sediments in ventilated and stagnant basins (Hesse and Schacht, 2011).....	170

List of Tables

Table 1: Total organic carbon (TOC) and stable carbon ($\delta^{13}\text{C}$) isotope data. The thickness, grain size and structure of each bed sampled are also listed.....	81
Table 2: Approximate $\delta^{13}\text{C}$ values for carbonates precipitated from normal Neoproterozoic seawater obtained from the literature. Discretion should be used when referencing these ages due to poor age constraints and discrepancies in isotopic values in the Neoproterozoic (Meyer, 2004).	169

1 Introduction

1.1 Thesis Rationale

Channel-levee complexes are a principal component of deep-marine slope systems and have been shown to form good hydrocarbon reservoirs (Navarro et al., 2007). Academic work on these channel-levee complexes has tended to focus on the channels rather than the levees, mostly because of the generally finer grained and more mudstone-rich nature of levees, which causes them to be more poorly exposed in outcrop and their less than decimetre-thick characteristics are below the resolution of even high resolution industry seismic surveys.

At the Castle Creek study area, channels, along with their levee deposits, are 100% exposed in recently-deglaciated, vegetation-free, vertically-dipping strata. This thesis will describe the lithological details of levee deposits. It will also delineate the lateral and vertical trends in bed thickness, structure and grain-size distribution that are crucial in providing details on the spatial characteristics of reservoir quality and continuity. Lastly, it will illustrate the temporal and spatial relationships between channel and levee sedimentation. The centimetre-scale data obtained from the outcrop at the Castle Creek study area will fill in the details which cannot be obtained from seismic surveys of modern submarine fans and will contribute to our knowledge of deep marine physical processes.

1.2 Geological Setting and Regional Stratigraphy

The Windermere Supergroup (WSG) consists of a succession of Neoproterozoic metasedimentary rocks exposed in an outcrop belt that stretches from northwestern Mexico northward through the Canadian Cordillera and up to the Yukon-Alaska border (Fig. 1.1) (Ross and Arnott, 2007). In the southern Canadian Cordillera, rocks of the WSG crop out mainly in the Foreland Fold and Thrust and Omineca belts. Here they form a sediment pile of mostly siliciclastic

deep-marine sedimentary rocks that is 6-9 km thick and extends over an area of at least 35 000 km² (Ross and Arnott, 2007; Ross et al., 1995).

A number of major global events took place during the Neoproterozoic (1000-544 Ma), including the break-up of the supercontinent Rodinia between 720 Ma and 650 Ma (Fig. 1.2) (Li et al., 2013). Associated with the break-up was the rifting of a landmass away from the present-day western coast of ancestral North America, Laurentia (Ross et al., 1995). This created an extensive passive continental margin (Ross, 1991) and expanded the proto-Pacific Ocean into which mostly continental- and also basinal-derived sediment was shed (Ross et al., 1995; Price, 2000). The conjugate counterpart that separated from Laurentia is a source of ongoing debate; Antarctica, Australia and Siberia have all been suggested as possible candidates (Ross and Arnott, 2007).

In southwestern Canada, following deposition of syn-rift strata into proto-Pacific Ocean during active continental drifting, post-rift deposition in a rapidly thermally subsiding basin formed a thick (> 6 km) accumulation of deep-water sedimentary strata that changes stratigraphically-upward to continental shelf strata (Ross et al., 1995). This accumulation of rocks was interpreted by Ross (1991) to represent the progradation of the passive continental margin of Laurentia (present-day western North America) into the proto-Pacific miogeocline as a response to decreased rates of thermally driven subsidence (Fig. 1.4).

A second rift and subsidence event occurred in the Early Paleozoic (Ross, 1991) and was followed by a protracted period of mostly shallow marine carbonate deposition (Monger and Price, 1979). In the Mesozoic, however, the passive margin along the western coast of Laurentia was transformed into a tectonically-active continental arc system. Collision and accretion of allochthonous terranes along the western coast of Laurentia during the Mid-Jurassic to Early

Cenozoic formed the Canadian Cordillera (Price, 2000), which can be subdivided into five orogen-parallel, morphogeological belts, each distinguished by different rock types, metamorphic grade and structure (Fig. 1.3) (Price, 2000; Reid et al., 2002). They are, from east to west, the Foreland Fold and Thrust, Omineca, Intermontane, Coast and Insular belts (Reid et al., 2002). This present-day configuration of the Canadian Cordillera was complete by Early Cretaceous time, however uplift and deformation continued into the Early Tertiary (Monger and Price, 1979).

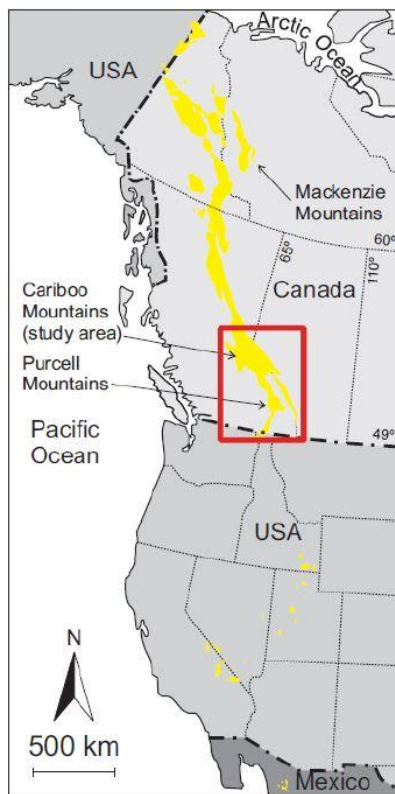


Figure 1.1: Distribution of WSG outcrops (yellow) in western North America. Well-exposed deep-water rocks in the southern Canadian Cordillera crop out in the red rectangle (Popovic, 2016 modified from Ross, 1991).

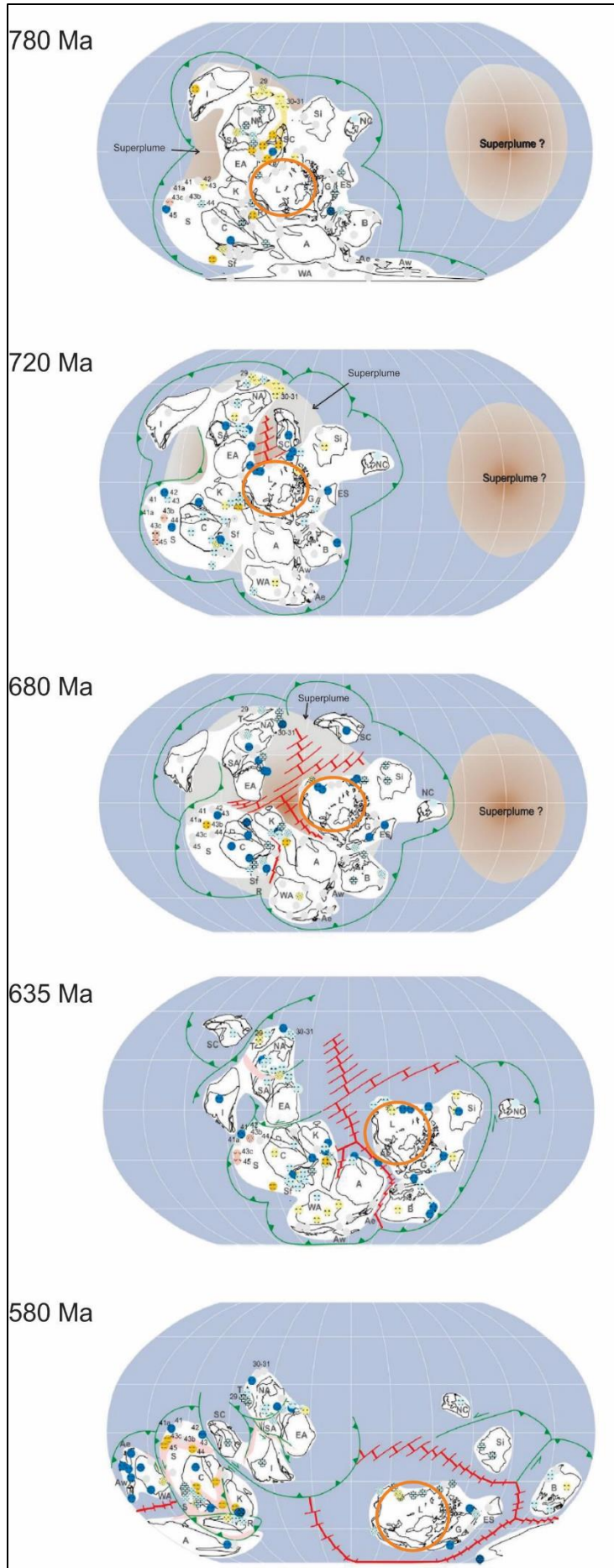


Figure 1.2: Global palaeogeographic maps for the 780-580 Ma interval. At 780 Ma Rodinia was intact. Laurentia is highlighted in the orange circle. At 720 Ma Rodinia begins to break up. The spreading ridge is shown by the red lines. At 680 Ma Rodinia is still breaking up and sea-level rises. At 635 Ma Rodinia is almost completely broken up. At 580 Ma the breakup of Rodinia is complete. Figure modified from Li et al., 2013.



Figure 1.3: Collision and accretion of allochthonous terranes along the western coast of Laurentia during the Mid-Jurassic to Early Cenozoic formed the Canadian Cordillera which can be subdivided into five orogen-parallel, morphogeological belts. (Price, 2000; Reid et al., 2002). They are, from east to west: the Foreland Fold and Thrust, Omineca, Intermontane, Coast and Insular belts (Reid et al., 2002). Map from Popovic, 2016; modified from Wheeler and McFeely, 1991.

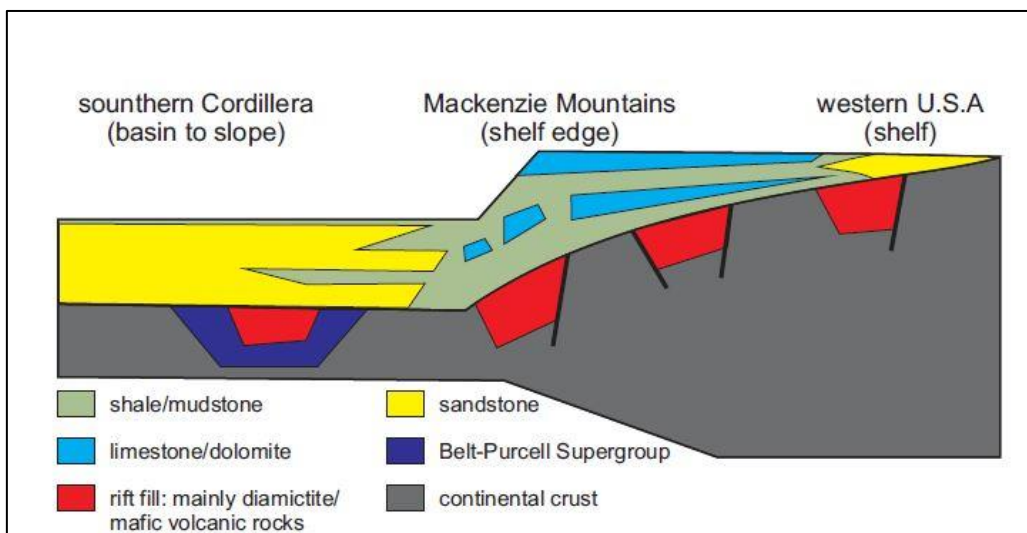


Figure 1.4: Schematic cross-section showing the interpreted western North American margin during the Neoproterozoic (redrawn by Khan, 2012 from Ross, 1991). Syn-rift deposits consisting mainly of mafic volcanic rocks are overlain by post-rift deposits consisting of sandstone, limestone/dolomite and shale/mudstone.

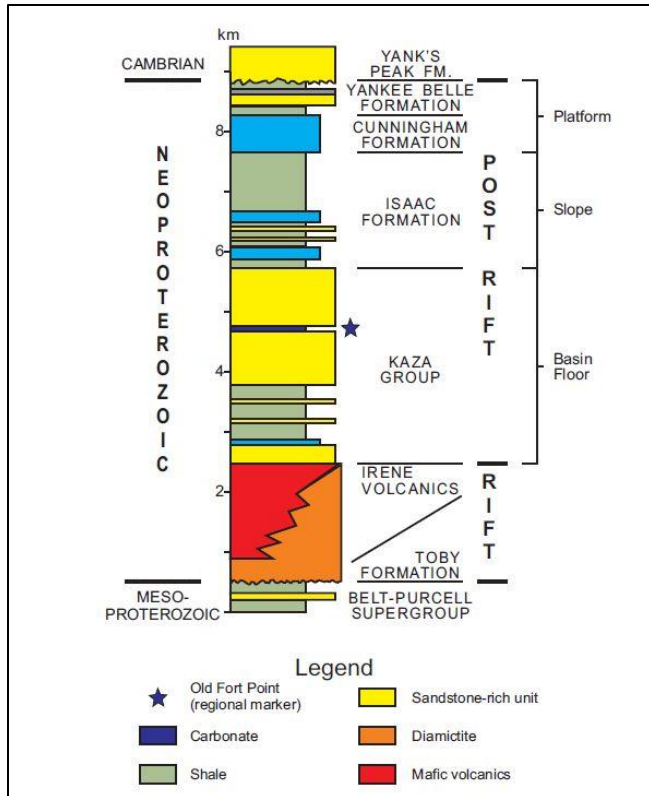


Figure 1.5: Stratigraphy of the WSG in the Cariboo Mountains of the southern Canadian Cordillera (from Khan, 2012 modified from Ross et al., 1995). Depositional environments are indicated on the right.

In the southern Canadian Cordillera, the WSG unconformably overlies sedimentary rocks of the Mesoproterozoic Belt-Purcell Supergroup (Evans et al., 2000). Here the WSG has been subdivided into 2 main stratigraphic units (Ross and Arnott, 2007). In southeastern B.C. the lower unit consists of syn-rift glaciogenic diamictites of the Toby Formation and mafic volcanics of the Irene Formation (Fig. 1.4). The upper unit, on the other hand, consists of a post-rift succession of mostly siliciclastic rocks that forms a shallowing-up sequence from basin floor to shelf strata deposited on a passive continental margin (Fig. 1.4). Post-rift strata are termed the Miette Group in the western Rockies, the Horsethief Creek Group in the Purcell Mountains and the Kaza Group and Isaac Formation (Cariboo Group) in the Cariboo Mountains (Ross and Arnott, 2007). In the Cariboo Mountains these rocks are then overlain successively by the Cunningham Formation composed mainly of oolitic-intraclastic limestone, and the Yankee Belle Formation composed mainly of a mixed siliciclastic-carbonate assemblage (Ross et al., 1995). In most parts of the

southern Canadian Cordillera this sedimentary pile is capped by the sub-Cambrian unconformity (Fig. 1.5) (Ross, 1991; Ross and Arnott, 2007).

The age of deposition of the WSG in the southern Canadian Cordillera is poorly constrained due to a paucity of datable markers and age-diagnostic fossils. A number of studies have dated basement rocks that unconformably underlie the WSG. Uranium-lead (U-Pb) dating of zircons from gneissic granites in the Sifton and Deserters Ranges in British Columbia, which presumptively predate deposition of the older Belt-Purcell Supergroup, yield an age of 728 Ma (Evenchick et al., 1984). More recently, McDonough and Parrish (1991) used U-Pb dating on gneisses from the Malton Complex and yielded a potential maximum age of 736 Ma for WSG sedimentation. Volcanics of the Huckleberry Formation, the equivalent of the Irene Formation in Washington State, have been dated at 762 ± 44 Ma by Sm-Nd dating (Devlin et al., 1988; Ross et al., 1995). The only direct age of WSG sedimentation is reported from the Old Fort Point Formation (OFP). Here black shales have been dated using Re-Os isochron techniques and yield a depositional age of 607.8 ± 4.7 Ma (Kendall et al., 2004). Lastly, syn-rift volcanic rocks in the Hamill-Gog Group, which overlies the WSG, have yielded an age of 569.6 ± 5.3 Ma and therefore provide an approximate minimum age constraint for WSG sedimentation (Colpron et al., 2002).

Within the WSG, there are three distinctive lithostratigraphic markers that are used for regional correlations across the southern Canadian Cordillera. The first, and oldest, is the OFP, which ranges from approximately 50 to 450 m in thickness and occurs between the Middle and Upper Kaza groups (Smith et al., 2014; Ross and Arnott, 2007). This unit consists mostly of thin-bedded siliciclastic and carbonate turbidites intercalated locally with distinctive purple and green slates and a black shale horizon (Ross and Arnott, 2007). The OFP is interpreted to represent deposition during a major rise in sea level that terminated siliciclastic sediment input into the

Windermere turbidite system (Kendall et al., 2004; Ross et al., 1995). This event eventually culminated in deposition of an anoxic black shale layer across most of the deep-water WSG basin (Smith et al., 2014). The other two regional markers occur stratigraphically above the OFP and range from 10 to 250 m in thickness. They consist of an intercalated succession of calcareous and siliciclastic turbidites, interbedded with mudstones, in addition to less common conglomerates and breccias containing shallow-water carbonate clasts (e.g., a stromatolite, oolite fragments). Both carbonate markers are interpreted to coincide with periods of high sea level that flooded the shelf and promoted shallow-water carbonate production, which then was resedimented into the deep-water part of the WSG basin (Ross et al, 1995; Ross and Arnott, 2007).

The source of siliciclastic sediment that builds up most of the WSG in the southern Canadian Cordillera has been determined using U-Pb dating on detrital zircons. Virtually all of these detrital zircon grains are characterized by a distinctive bimodal distribution of zircon ages: 1.65- 2.16 Ga (Paleoproterozoic) or >2.5 Ga (Archean) suggesting that siliciclastic strata of the WSG were derived from fluvial drainage of the southern Canadian Shield and northwestern United States (Ross and Parrish, 1991). Limited paleocurrent data based on current-ripple and dune cross-stratification, suggests that these sediments were then transported by powerful turbidity currents in a generally west-northwest direction (Ross and Arnott, 2007).

1.3 Castle Creek Study Area, Cariboo Mountains

Strata of the WSG are well exposed at the Castle Creek study area. This study area is located in the northern Cariboo Mountains of east-central British Columbia and forms part of the Omineca Belt in the southern Canadian Cordillera (Fig. 1.6, 1.7) (Ross and Arnott, 2007; Reid et al., 2002). This area has been affected by four phases of deformation. The first two phases formed northwesterly trending and plunging folds (D1 and D2) whereas the latter two phases consist of

upright folds, with D3 having a weak northwest trending axial planar fabric and D4 having a strong northeast trending fabric. Regional metamorphism started shortly after D1 and peaked between D2 and D3 during which time strata were metamorphosed to greenschist facies (Murphy, 1987; Ross and Arnott, 2007). Deformation and metamorphism occurred mainly in the Jurassic and Cretaceous, with the first two phases of deformation having formed between the Early and Middle Jurassic, when an oceanic terrane accreted to the North American continental margin (Brown et al., 1986; Murphy, 1987). When palinspatically restored (assuming ~30% shortening from the deformation events), the Windermere turbidite system would have had a minimum extent of 80,000 km² and would, therefore, be comparable in size to the modern Amazon and Mississippi turbidite fan systems (Fig. 1.8) (Ross and Arnott, 2007).

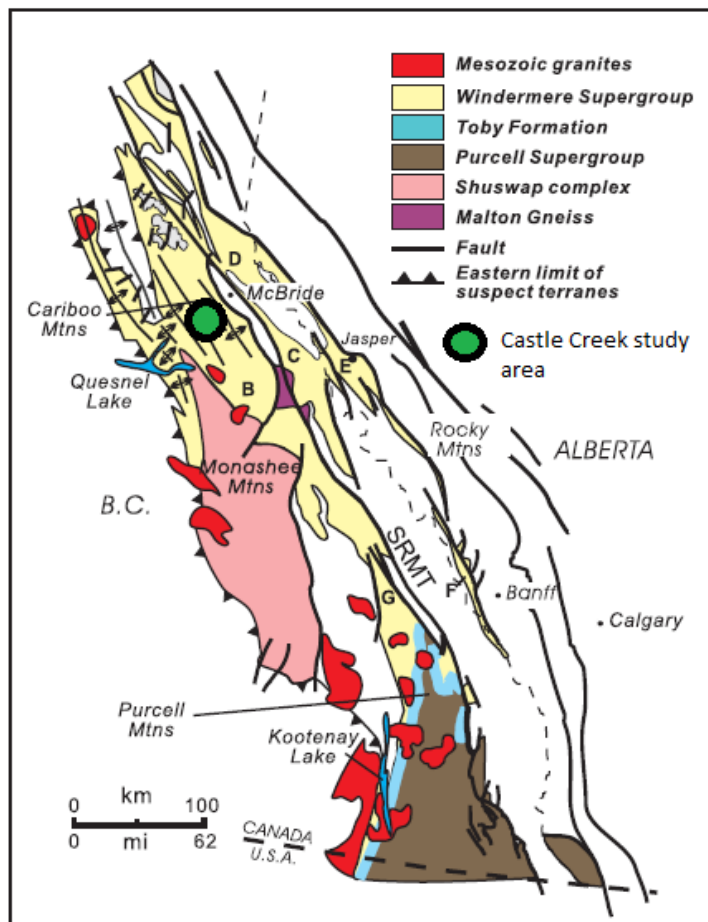


Figure 1.6: Geologic map of the WSG in the southern Canadian Cordillera, including location of the Castle Creek study area (green circle). Modified from Ross and Arnott, 2007; modified from Young et al., 1973 and Ross and Murphy, 1988).

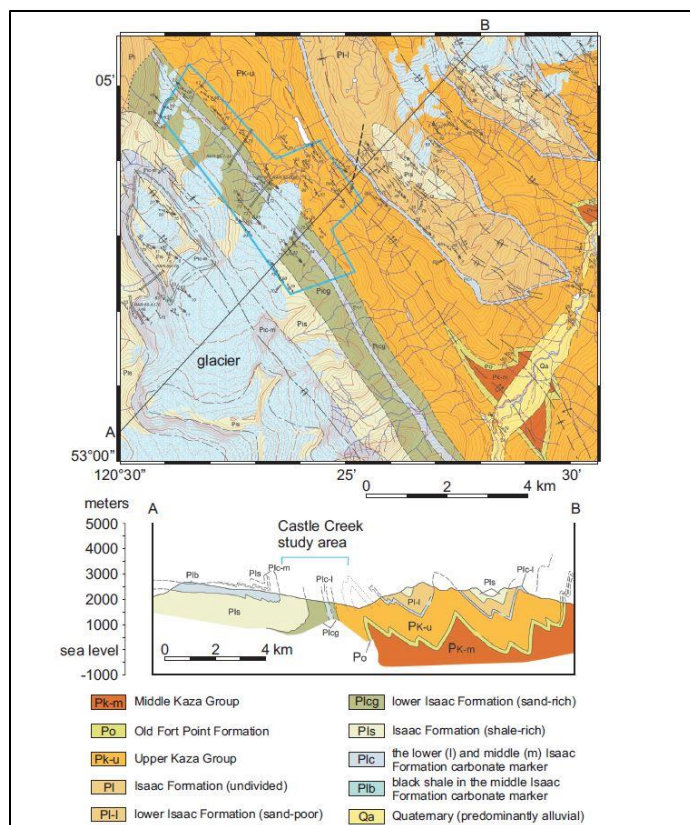


Figure 1.7: Regional geology of the Castle Creek study area is outline in blue (from the Eddy 1:50,000 map area; Ross and Ferguson, 2003a).

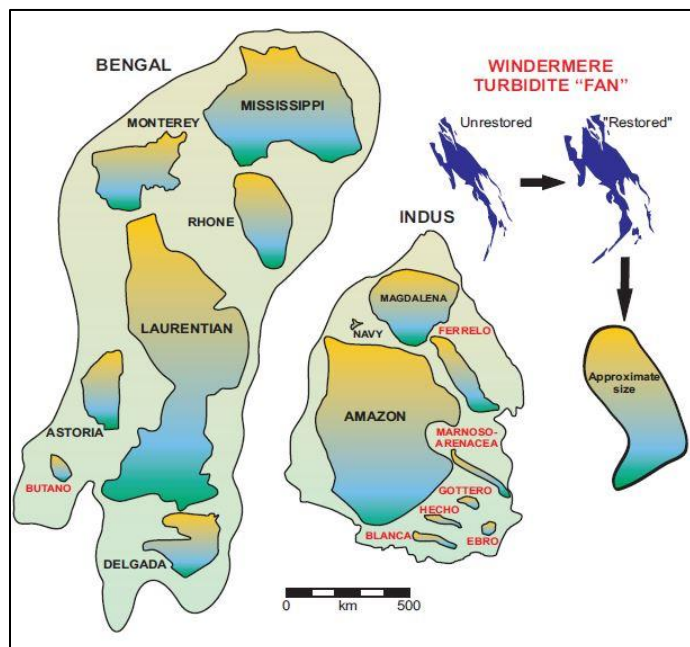


Figure 1.8: Palinspastic restoration of the WSG makes it comparable in size to the modern Amazon and Mississippi fans. For comparison, other modern fans (black text) and ancient fans (red text) are shown (redrawn by Khan, 2012; inset map of the WSG from Ross, 2001, modified from Barnes and Normark, 1985).

In the Castle Creek study area, the only significant regional structures that are preserved were formed during D2 (Murphy and Rees, 1983). Here, strata are superbly exposed on a nearly vertically-dipping limb of a regionally extensive southwest-verging D2 anticline on the eastern side of the Isaac synclinorium, a structural depression bounded on its margins by dextral strike-slip faults (Ross and Arnott, 2007; Reid et al., 2002). At Castle Creek strata crop out in an area up to 7 km parallel to regional bedding and 2.5 km perpendicular to bedding (Fig. 1.9A). Even though strata have been metamorphosed to lower greenschist, primary sedimentary structures are well preserved. As a result, sedimentary terminology, instead of metamorphic, is used to describe these strata. In addition the rapidly retreating Castle Creek Glacier divides the study area into south and north regions (Fig. 1.9A). Exposed strata have been polished smooth, are vegetation-free and are nearly 100% exposed with only local cover by glacial debris and moraine. A third region, informally called the hill section, is located farther to the northwest.

Strata of the upper Kaza Group make up the lower 800 m of exposed section at Castle Creek (Fig. 1.9B). Strata consist of massive to normally graded, medium- to very thick-bedded coarse-grained to pebble conglomerate turbidites interbedded with up to 40 m thick intervals of very thin- to thin-bedded fine-grained turbidites. These rocks have been interpreted to represent channelized-lobe elements deposited by poorly confined flows within a basin-floor setting. Moreover, the sandstone-shale ratio in the upper Kaza Group is high, 75:25 (Khan and Arnott, 2011; Meyer and Ross, 2007; Ross et al., 1995).

The Isaac Formation conformably overlies the upper Kaza Group and comprises the upper 1.6 km of the exposed section. Coarse-grained sandstone and conglomerates form laterally discontinuous stratal units up to 100 m thick. They are interpreted to be the fills of continental slope channels that built up units informally termed Isaac channel complexes 1-6 (Fig. 1.9B).

These channel complexes are bounded on all sides by thin-bedded turbidites interbedded with thicker, amalgamated sandstone units that are interpreted to represent levee and other extra-channel deposits. Thick debris flow, slump and slide units are common in the Isaac Formation and indicate gravitational instability likely on the continental slope. Strata are more mud-rich compared to the upper Kaza Group, with a sandstone-shale ratio of 25:75 (Khan and Arnott, 2011; Ross and Arnott, 2007).

Collectively, strata exposed at the Castle Creek study area record a km-scale upward-shoaling succession from proximal basin floor deposits of the upper Kaza Group to base of slope deposits of the Isaac Formation. As described earlier, this trend is a consequence of the progradation of the passive continental margin of Laurentia into the proto-Pacific Ocean miogeocline (Khan and Arnott, 2011).

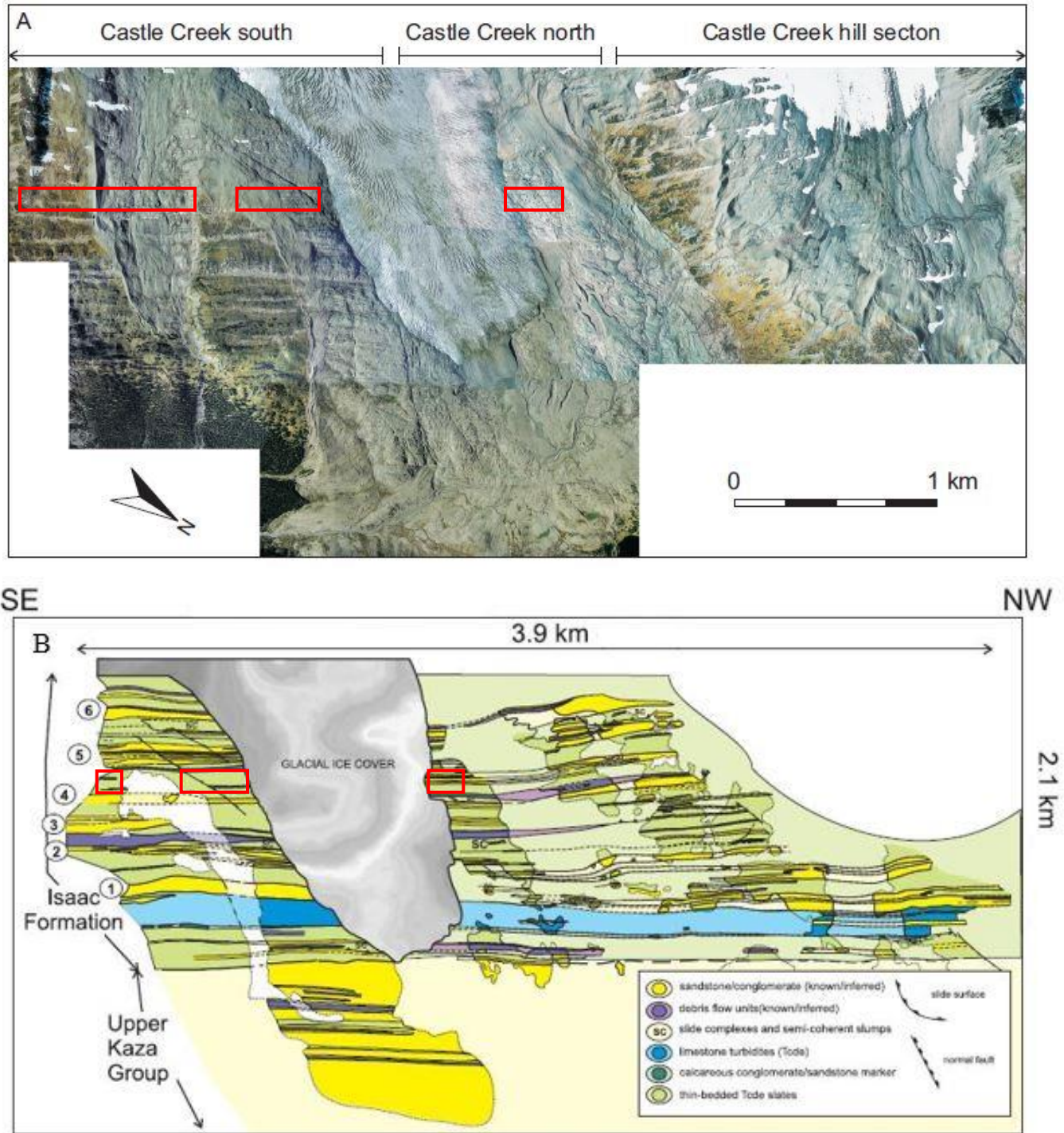


Figure 1.9: A) Photomosaic of the Castle Creek study area. This study documents the stratigraphy of levee deposits associated with base of slope channels in the Isaac Formation. The three outcrops of this study are outlined in red boxes. B) Interpreted geologic map of the Castle Creek study area (after Ross and Arnott, 2007). Isaac channel complexes 1-6 (ICC1-6) are numbered on the left. The channel-levee complex of this study crops out between ICC4 and ICC5 on the lower slope and is outlined by the red boxes.

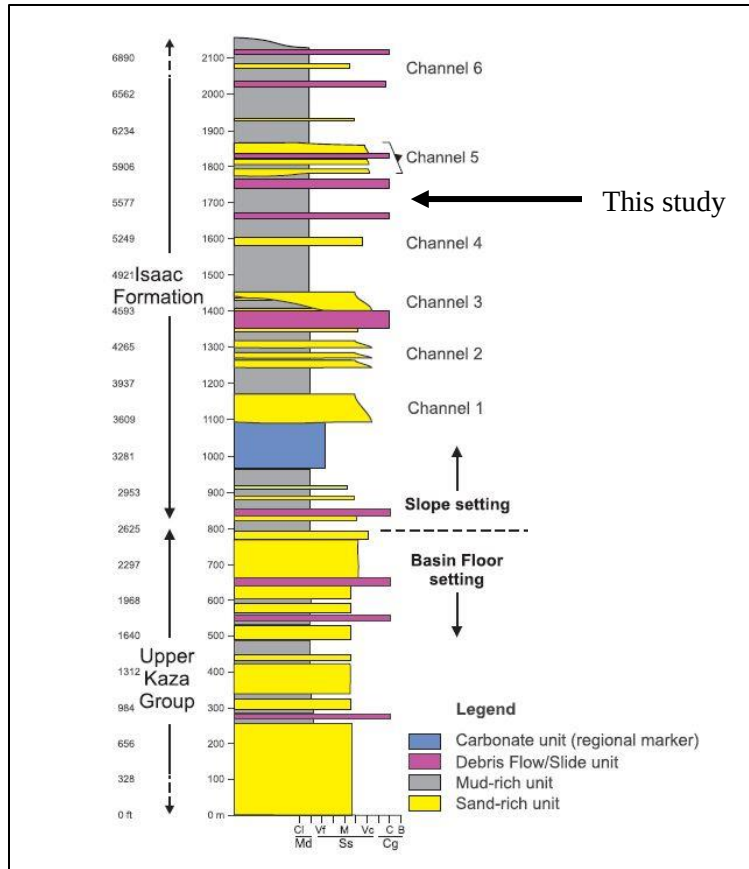


Figure 1.10: General stratigraphic nomenclature of strata exposed at Castle Creek. Levee deposits of this study are located between Isaac Channels 4 and 5 (modified from Ross and Arnott, 2007).

1.4 Previous Work

Campbell et al. (1973) were the first to map and describe the regional geology, deformation and metamorphism of the Castle Creek study area, in addition to producing a regional (1:250 000) map. Later, Ross and Ferguson (2003) produced a 1:50 000 map which included the Castle Creek study area (Fig. 1.6). Since then many others have also studied the structural and regional geology of the area, including Murphy and Rees (1983), Murphy (1987), Ross (1991), Ross et al. (1995) and Reid et al. (2002).

More recently detailed sedimentological, stratigraphic and architectural analyses of the deep-marine deposits in the Castle Creek study area have been conducted by the industry-government funded Windermere Consortium research group from the University of Ottawa. In this

work levee deposits have been identified (e.g., Navarro, 2006; Navarro et al., 2007; Arnott, 2007; Schwarz and Arnott, 2007; Davis, 2011) but to date only Khan (2012) has described them in detail. The goal of this thesis, therefore, is to expand upon this work by illustrating the make-up and stratal geometry of a second leveed-channel succession (Fig. 1.10) and to compare and contrast it with the previously described channel-levee complex. In particular, the improved exposure of strata in this study compared to Khan (2012) will better elucidate the lateral and vertical facies changes in these strata.

1.5 Methodology

1.5.1 Field Work

This study documents the stratigraphy of levee deposits associated with base-of-slope channels in the Isaac Formation (Fig. 1.10). Fieldwork was conducted over two summer seasons, July-August 2014 and July-August 2015, in two major field areas separated by the Castle Creek Glacier (Fig. 1.9). In the Castle Creek south study area, bed-by-bed measurements and detailed sedimentological descriptions of nine vertical stratigraphic sections, each ~130 m thick and about 400 m in strike length were made. Descriptions document strata thickness, bedding contacts, grain size, sedimentary structures and lithology. In the Castle Creek north study area, 5 vertical stratigraphic sections ranging from ~60 m to ~90 m thick and about 250 m in strike length were made. In addition, 32 short stratigraphic sections ranging from 2 m to 5.5 m in thickness were measured within these areas in order to capture lateral lithofacies changes in more detail.

Correlation of beds was accomplished by walking out bounding surfaces in the field. These surfaces were then traced onto high-resolution aerial photographs (1:300-1:500). Stratigraphic sections were digitized, correlated and used to identify and interpret lateral and vertical trends.

1.5.2 Petrography

A total of 37 samples were collected for detailed petrographic description of grain size, sorting, mineralogy and microstructure using an Olympus BX-41 plane and polarized light microscope. The approximate modal percentage of minerals in each thin section are based on visual estimates.

1.5.3 Analytical Techniques: Total Organic Carbon Content and Stable Carbon (^{13}C) Isotopes

Geochemical analyses were done on 9 samples to determine the weight percent of total organic carbon (TOC) and the stable carbon (^{13}C) isotope signature of organic material. Whole rock samples were first crushed with a sledge hammer in order to make smaller rock chips. Rock chips were then pulverized into a powder using a vibratory pulverizer and powdered samples were placed into containers.

First, to determine the TOC, an acid wash was performed on all samples to remove any inorganic carbonate (which occurs mainly in the carbonate cements). Approximately 500 mg of each powdered sample was weighed and placed in 10 mL beakers. Approximately 7 mL of 10% HCl was added to each beaker and the solution vigorously stirred to ensure that all of the powdered material was thoroughly soaked in acid. The solutions were then left to soak for 24 hrs. Next, samples were decanted and mixed into distilled water, stirring the powdered solution and allowing the powder to soak and rest for 1.5 hours between each wash. This procedure was repeated a total of three times. Samples were decanted a final time and the wet powders were placed in an oven and heated at 70°C for 24 hours.

Approximately 10 mg of each powdered sample were weighed into tin capsules. The capsules were loaded with standards and tungsten oxide, which acts as a combustion catalyst, into a Micro Cube elemental analyser to measure the percent carbon content (% C) in each sample. In

the elemental analyser, samples were flash combusted with oxygen at about 1800°C and then carried by helium through columns of reducing/oxidizing chemicals to produce CO₂ gas. Next, the gas was trapped within a single "trap and purge" adsorption column, and released separately so that the thermal conductivity detector (TCD) could detect and measure the gas separately as it was released. The routine analytical precision (2sigma) for the analyses is +/-0.1%. (Paul Middlestead (G.G. Hatch Stable Isotope Laboratory) personal communication, 2016). The TOC of the samples is measured from this procedure.

Next, to measure the various carbon isotopes, an Isotope Ratio Mass Spectrometer (IRMS) was used. The IRMS measures a ratio and requires the amount of carbon in each sample to be between 0.02 and 0.2 mg (Davis, 2011). Using the percent carbon previously measured by the elemental analyser for each sample, the weight of powdered sample required to measure carbon isotopes was calculated and determined to be approximately 50 mg. The samples and standards were weighed into tin capsules and loaded into an elemental analyser connected to an IRMS to measure the various carbon isotopes. In the elemental analyser the samples were flash combusted at about 1800°C (Dumas combustion) and the exsolved gas products carried by helium through columns of oxidizing/reducing chemicals optimised for CO₂. The gas was separated by a "purge and trap" adsorption column, sent to the IRMS interface and then lastly sent to the IRMS, where the relative abundance of each carbon isotope was determined. The routine analytical precision for the analyses is +/-0.20‰ (Paul Middlestead (G.G. Hatch Stable Isotope Laboratory) personal communication, 2016).

2 Deep-water sedimentation processes and their associated deposits

The processes by which large volumes of sediment are transported downslope into deep-marine environments were poorly understood up until the 1950s as they could not be directly observed (Arnott, 2010). In 1950, Kuenen and Migliorini began experimenting by releasing sediment into a basin of still-standing water. They observed a characteristic upward-fining trend in the deposits of their experiment and were the first to propose that this motif was the result of deposition from turbidity currents. In 1962, Bouma observed both a fining-upward trend and a characteristic sequence of sedimentary structures in deep-marine rocks. Today, it has been established that sediment is supplied to the deep-marine environment by gravity-driven processes in the form of mass-movement processes and sediment gravity flows (Arnott, 2010; Shanmugam, 2006).

2.1 Mass-movement Deposits

Mass-movement is a term used to describe the failure, dislodgment and downslope movement of coherent to semi-coherent masses of sediment under the influence of gravity (Fig. 2.1) (Arnott, 2010; Shanmugam, 2006). There are many different mechanisms that initiate shelf-edge sediment failures, which in turn generate mass-movement. Some examples include: eustatic changes in sea level, earthquakes, oversteepening of the submarine slope or high sedimentation rates (Shanmugam, 2006). Movement of these deposits occurs when the force of gravity exceeds the tensile strength of the sediment mass. Movement continues until the resisting forces, mainly friction along the base, exceed the gravitational forces and en masse deposition occurs. Mass-movement deposits are subdivided into two end-members: slides and slumps, based on the degree of internal deformation (Arnott, 2010). According to Shanmugam (2006), a slide is a coherent mass of sediment that moves along a planar glide plane and shows negligible internal deformation

whereas a slump moves along a concave-up glide plane and undergoes rotational movements and more extensive internal deformation. Mass-movements can travel hundreds of kilometres in distance and can form deposits up to hundreds of metres thick.

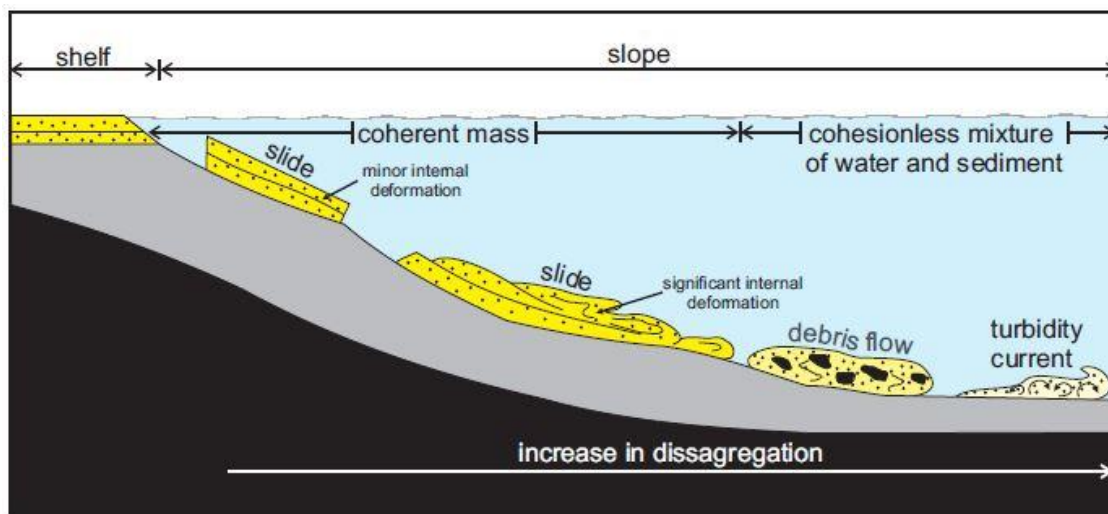


Figure 2.1: Initially, gravity-driven sediment failure at the shelf edge forms coherent slides and slumps. Over time, these deposits can evolve into debris flows and turbidity currents. Importantly, this evolution does not always occur and any one of these gravity-driven mass movements or sediment flows may form directly at the shelf edge, on the slope or in the distal basin (Khan, 2012) (redrawn by Khan, 2012 from Shanmugam et al., 1994).

2.2 Sediment-Gravity Flows

Gravity flows are generated when a denser fluid moves through a less-dense fluid and displaces it. In deep-marine environments, suspended sediment makes a local fluid-sediment volume denser than the surrounding water, and gravity displaces the denser fluid-sediment volume downslope (Arnott, 2010). The early classification scheme of Middleton and Hampton (1976) classified sediment-gravity flows into four types based on their respective sediment-support mechanisms. They are: debris flows (matrix strength), grain flows (grain-grain interaction), fluidized sediment flows (upward escaping intergranular flow) and turbidity currents (fluid turbulence). This thesis uses the simplified classification scheme of Mulder and Alexander (2001)

based on physical flow properties and grain-support mechanisms. In it, they classified sediment-gravity flows into two types: cohesive flows and frictional flows, where frictional flows are further subdivided into hyperconcentrated density flows, concentrated density flows and turbidity currents (Fig. 2.2).

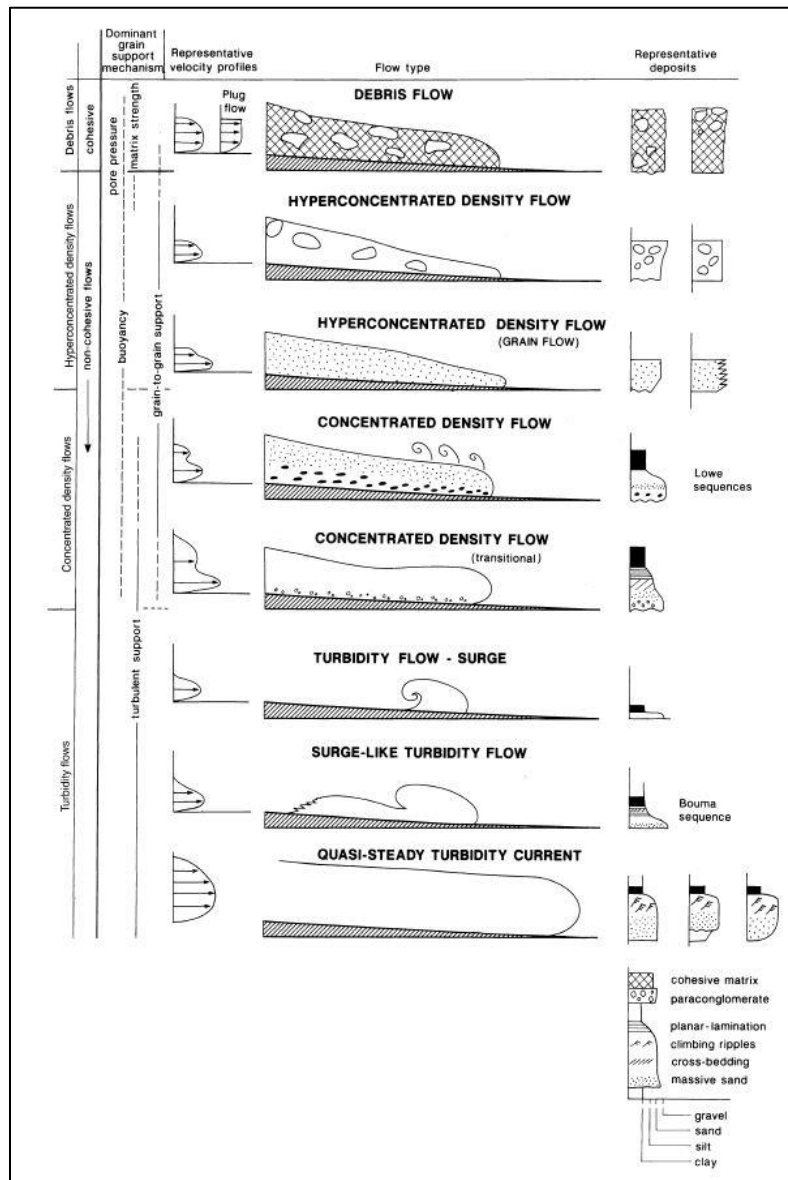


Figure 2.2: Schematic of the simplified classification scheme of Mulder and Alexander (2001). They classified sediment-gravity flows into two types: cohesive flows and frictional flows, where frictional flows are further subdivided into hyperconcentrated density flows, concentrated density flows and turbidity currents.

2.2.1 Cohesive Flows

Cohesive flows, also termed debris flows and mud flows, are sediment-gravity flows with a pseudoplastic rheology and laminar state from which deposition occurs through en masse freezing (Mulder and Alexander, 2001; Shanmugam, 2006). Cohesion provided by the matrix of fluid and fine-grained sediments is the main grain-support mechanism, however buoyancy, particle-particle interaction, hindered settling, elevated fluid turbulence and pore pressure are all other important factors which may assist in suspending particles (Arnott, 2010; Mulder and Alexander, 2001). Hampton's (1975) experiments showed that the matrix strength of cohesive flows was capable of supporting up to coarse-sand sized particles. However, there have been many examples where debris flows have supported decametre-scale sized clasts, which are further supported by high local pore pressures, buoyancy and clast-to-clast interaction (Mulder and Alexander, 2001).

Cohesive density flows are divided into mud flows and debris flows based on sediment size sorting. Mud flows are defined as having less than 5% gravel by volume, a mud to sand ratio of more than 1:1 and transport little coarse sediment, although they may transport isolated blocks. In contrast, debris flows have more than 5% gravel by volume (and therefore are more poorly sorted) and can transport boulder-sized clasts and decametre-scale sized blocks (Mulder and Alexander, 2001).

Debris flows travel over long distances of up to several hundred kilometres and commonly have planar, non-erosional basal contacts (Arnott, 2010). Although turbulence damping by the cohesive matrix may partly explain the long runout distances and non-erosional contacts, hydroplaning also has an effect. Hydroplaning occurs when a layer of water becomes trapped

between the flow snout and the underlying bed, reducing friction, and therefore erosion, allowing the flow to glide over long distances (Mulder and Alexander, 2001).

Over the course of the flow, debris flows may transform into sediment-driven, non-cohesive flows, however such transformations are poorly understood. Flow transformation may occur through mixing and dilution at the flow head (Hampton, 1972), fluid entrainment along the upper boundary of the body or by passing through a hydraulic jump (Fisher, 1983; Weirich, 1988). Such transformations are more likely to occur in debris flows which have high proportions of medium silt-sized particles or coarser (Hampton, 1972).

En masse deposition of debris flows occurs when the forces resisting shear (mainly viscosity and friction) become equal to the main driving force (force due to gravity) (Alexander and Mulder, 2001). This causes the flow to freeze from areas of lower shear near the top downwards to areas of higher shear at the base. The deposits, called debrites, are lobe-shaped masses that have steep margins and can be up to 10s of km wide and over 100 m thick. Internally, they range from being mud to sand rich and are characteristically disorganized and poorly sorted, with clasts ranging from sand grains to large blocks supported in a fine-grained matrix (Arnott, 2010).

2.2.2 Frictional Flows

Frictional flows include a sequence of different flows, ranging from sediment slides to turbidity currents. Subdivision of these flows is based on the dominant particle-support mechanisms including buoyancy, pore pressure, grain-to-grain interaction, Reynolds stresses and bed support. Moreover, the dominant particle-support mechanism depends on flow conditions, particle concentration, grain-size distribution and particle type (Mulder and Alexander, 2001).

2.2.2.1 Hyperconcentrated Density Flows

Although hyperconcentrated density flows have similar sediment:water proportions as cohesive flows, they behave as non-Newtonian fluids due to a lesser amount of mud-sized particle content. However, the ratio of cohesive to non-cohesive particles that distinguish cohesive flows from frictional flows is not well defined. The main difference between hyperconcentrated density flows and cohesive flows is that hyperconcentrated density flows are friction dominated. In hyperconcentrated density flows, high sediment concentrations (over 25% volume) make grain-to-grain collisions an important mechanism for suspending sediment. Gravity is the driving force of these flows and frictional freezing resulting from grain-to-grain interactions is the main depositional mechanism. Deposits are commonly massive coarse silt, sand or gravel with localized inverse grading (Mulder and Alexander, 2001).

2.2.2.2 Concentrated Density Flows

Concentrated density flows have lower sediment concentration and fewer grain-to-grain interactions than hyperconcentrated density flows. They are more dilute and as a result behave as Newtonian fluids. The boundary between hyperconcentrated and concentrated density flows is marked by a change in particle behaviour where the larger grains are no longer supported by grain-to-grain interactions in the latter, causing coarse-tail grading to occur. The main driving force of concentrated density flows is gravity and sediment is mainly supplied by erosional processes. Grain-to-grain interaction is the main particle-support mechanism in these flows, however fluid turbulence acts as the dominant support mechanism near the top of the flow. Coarse sediment (gravel and coarse sand) are transported as bed load and commonly form traction transport structures. Deposits commonly have erosional bases, consist of massive sand or gravel and are coarse-tail graded (Mulder and Alexander, 2001).

2.2.2.3 Turbidity Currents

Turbidity currents are the most common type of frictional flow and are thought to be the principal transport agent of sediment into the deep marine (Arnott, 2010; Khan, 2012). Alexander and Mulder (2001) define turbidity currents as sediment-gravity flows in which sediment is mainly supported by the upward component of fluid turbulence, with other mechanisms providing additional sediment support to varying degrees. The effect sediment concentration has on these types of flows is poorly understood (Arnott, 2010). Lowe (1982) recognized two end-member types of turbidity currents: low-density and high-density. In low-density turbidity currents, fluid turbulence is the only particle support mechanism and flows cannot exceed 9% sediment concentration by volume (Bagnold, 1954). Once sediment concentration exceeds this limit, fluid turbulence is no longer the only operating particle support mechanism, and such flows are termed high-density turbidity currents. Grain-to-grain interactions damp fluid turbulence and as a result, dispersive pressure, hindered settling and buoyancy are needed to support sediment, especially the coarser particles (Arnott, 2010).

Turbidity currents can be initiated from the discharge of fluvial systems at the shelf edge or may form from the transformation of slumps, slides and debris flows. The structure of a turbidity current consists of a well-defined head, body and tail (Fig. 2.3). The head has an overhanging nose, which forms as a result of the frictional forces being exerted along the upper and lower boundaries of the flow (Kneller and Buckee, 2000). The head is the sediment-rich, most erosive part of the flow and where most of the mixing with the ambient fluid occurs (Arnott, 2010; Kneller and Buckee, 2000). The body of the flow has a velocity 30-40% faster than the velocity of the head and continuously supplies sediment to the head, sustaining the current (Kneller and Buckee, 2000). Over time, the flow becomes longitudinally stratified in terms of grain size -- coarser sediments

accumulating in the lower part of the head whereas finer sediment becomes concentrated into the body and tail of the flow. Also at the tail of the flow, sediment concentration is low and flow velocity is slower than that of the body (Arnott, 2010).

Most turbidity currents are also vertically density-stratified, with grain concentration highest near the base of the flow decreasing exponentially upwards and velocity highest at approximately 20-30% flow depth and decreasing upwards (Fig. 2.4) (Alexander and Mulder, 2001). Deposition of turbidity currents occurs when the flow moves down a shear velocity gradient, which is equivalent to a reduction in flow velocity. This can either occur when the flow velocity decreases spatially, for example due to a decrease in slope, flow expansion or reduced sediment load, or temporally, for example due to fluctuations in the rate of sediment supply. Sediment may deposit in a few different ways. First, sediment may settle from a static suspension if the flow comes to rest or slows to the point that fluid turbulence is no longer capable of maintaining the grains in suspension. This is the process by which mud is deposited from turbidity currents because the bed shear stress required to maintain mud-sized particles in suspension is small, and only when the current comes to rest will the particles fall from suspension. Second, sediment may fall from suspension and be moved as bed load along the bed, resulting in the formation of unidirectional sedimentary structures before coming to rest. Third, where the flux of sediments from suspension to the bed is very high, sediment may be deposited directly from suspension with negligible bed-surface traction transport, even when conditions are conducive to traction transport (Kneller and McCaffrey, 2003).

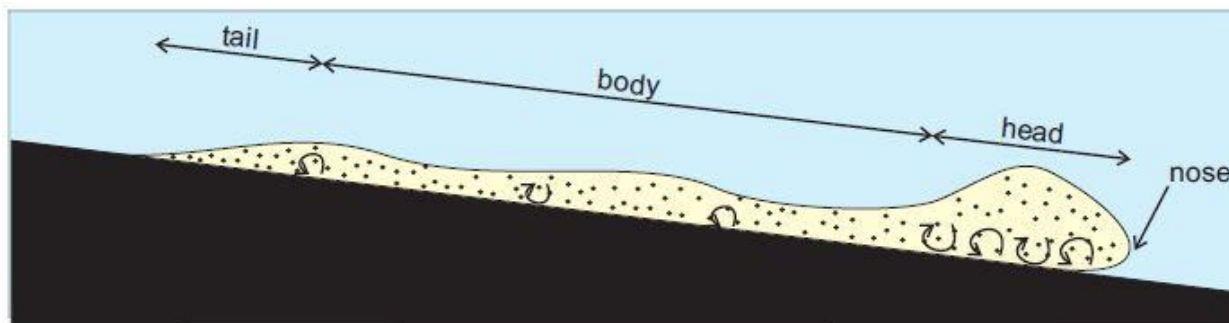


Figure 2.3: The structure of a turbidity current consists of a well-defined head, body and tail. The head has an overhanging nose that forms as a result of the frictional forces being exerted along the upper and lower boundaries of the flow (Kneller and Buckee, 2000). Illustration from Khan, 2012.

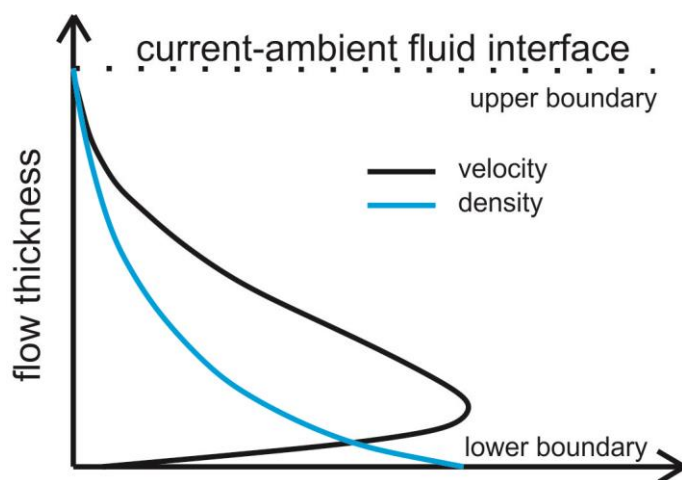


Figure 2.4: Most turbidity currents are vertically density-stratified, with grain concentration highest near the base of the flow decreasing exponentially upwards. The velocity is highest at approximately 20-30% of the flow depth and decreases upwards (Alexander and Mulder, 2001). Modified from Khan, 2012.

Deposits from turbidity currents are called turbidites. They are composed of mud- to gravel-sized sediment and can range from a few millimetres to several metres thick (Khan, 2012). Bouma (1962) was the first to describe classical turbidites, which are interpreted to be deposited by decelerating, low-density turbidity currents (<9% sediment concentration by volume). Bouma's idealized turbidite sequence consists of five divisions, which, from base to top, are termed the Ta-Te divisions (Fig. 2.5). The Ta division consists of normally graded, structureless sandstone or conglomerate deposited directly from suspension. This division is interpreted to form in the upper

flow regime. Ta turbidites are interpreted to be deposited rapidly from high-density turbidity currents. The structureless nature of the bed is suggestive of high suspension fallout rates that inhibit the development of tractional sedimentary structures like planar lamination or cross-stratification (Arnott and Hand, 1989). The Tb division consists of fine- to medium-grained, planar laminated sandstone and is commonly interpreted to form in the upper flow regime plane bed in flows with lower sediment concentrations than the Ta division. However, recent work by Tilston and Arnott (2015) showed that planar lamination is formed by traction transport on a surface that lacks the defects required to initiate the formation of angular bed forms such as ripples or dunes. This, then, is overlain by the Tc division consisting of fine- to very fine-grained small-scale cross-stratified sandstone interpreted to form from current ripples in the lower flow regime (Khan, 2012). Interlaminated sandstone/siltstone and mudstone make up the Td division, which forms by mixed traction and suspension sedimentation processes (Lowe, 1982). Lastly, the Td division is capped by the mudstone Te division deposited by suspension sedimentation processes (Lowe, 1982). Few turbidites show the complete Bouma sequence, with most only preserving parts of the Tabcde sequence.

On the other hand, high-density turbidity currents (> 9% sediment concentration by volume) do not deposit classic Bouma turbidites. Lowe (1982) proposed a classification scheme for these deposits based on the grain size and concentration of the flows that formed them (Fig. 2.6). In it, R represents gravel while S represents sand. In this classification, the lowest unit (R2) consists of inversely-graded gravel deposited by traction carpet deposition. This type of deposition occurs when grain collisions become the dominant support mechanism of the flow. This happens due to a high concentration of coarse particles in the bed load layer that suppresses turbulence (Khan, 2012). This unit is then overlain by R3, which consists of normally-graded gravel deposited by

suspension sedimentation. As the flows progressively lose their coarse-grain fraction they become sandy high-density turbidity currents and are subdivided into 3 units (S1-S3). S1 is interpreted to form by bed load transport of sand and can consist of planar laminated or dune cross-stratified sandstone. As sedimentation rates increase, S2 is deposited by traction carpet sedimentation and lastly S3 is deposited by suspension sedimentation of sand after traction carpet sedimentation stops. S3 is equivalent to the Ta division of the Bouma sequence.

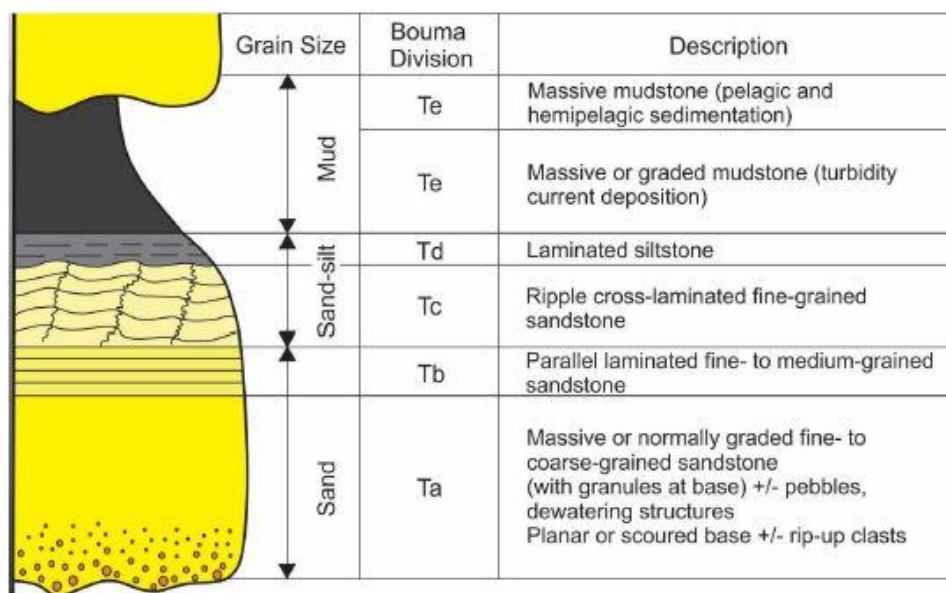


Figure 2.5: Idealized Bouma sequence for low-density turbidity currents showing the typical grain size profile and sedimentary structures (redrawn by Angus, 2016 from Bouma, 1962). Few turbidites show this complete Bouma sequence, with most only preserving parts of the Tabcde sequence.

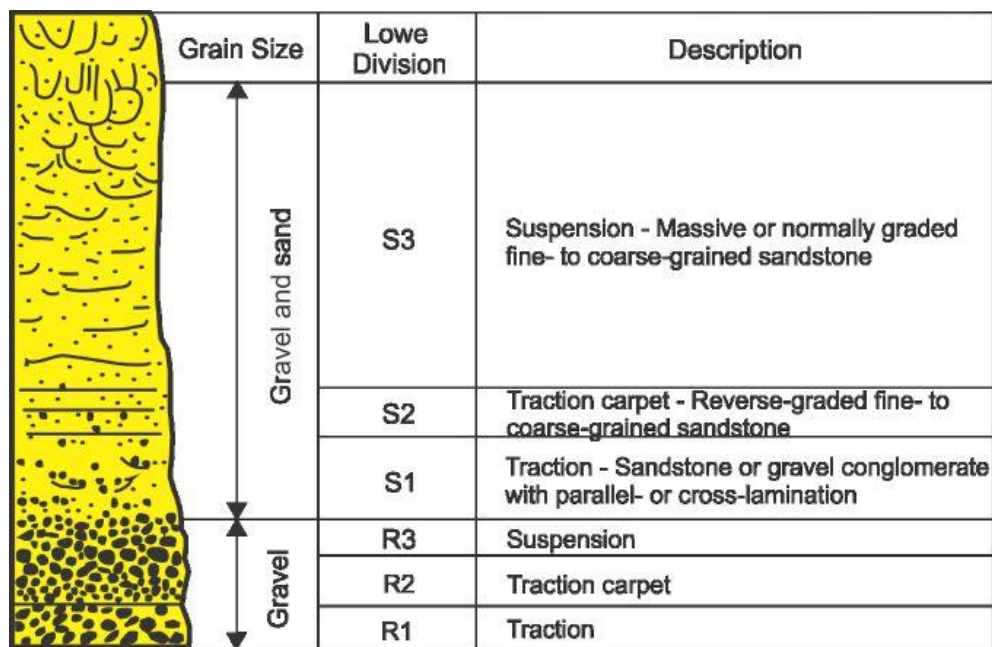


Figure 2.6: Idealized Lowe sequence for high-density turbidity currents showing the typical grain size profile and sedimentary structures (redrawn by Angus, 2016 from Lowe, 1982). The S3 division is equivalent to the Ta division of a Bouma sequence.

3 Submarine Fans

3.1 Overview of Deepwater Systems

Sediment-gravity flows transport large volumes of clastic material from the shelf down the slope to the ocean floor where it is deposited in large-scale depositional systems called submarine fans (Fig. 3.1) (Khan, 2012). Both the origin and characteristics of submarine fans depend on a number of parameters including changes in sea level, tectonics, the nature of the sediment supply system, basin salinity, sediment composition and organic supply (Arnott, 2010; Reading, 1996). Following the discovery that graded bedding is formed by turbidity currents (Kuenen and Migliorini, 1950), numerous submarine fan models have been proposed since the pioneering works of Menard (1960), Normark (1970) and Haner (1971). Reading and Richards (1994) classified deep-marine depositional systems based on the nature of the supply system, recognizing point-source submarine fans, multiple-source submarine ramps and linear-source slope aprons. They then further subdivided these systems based on grain size of the available sediment, distinguishing between gravel-rich, sand-rich, mud-rich and mixed sand-mud depositional systems. Based on these criteria Reading and Richards (1994) recognized 12 classes, ranging from small (radius of ~1-50 km) gravel-rich systems associated with steep slope gradients and limited source areas, like a fan delta, to large (radius of ~100-3000 km) fine-grained systems with shallow slope gradients and very large source areas, like a delta.

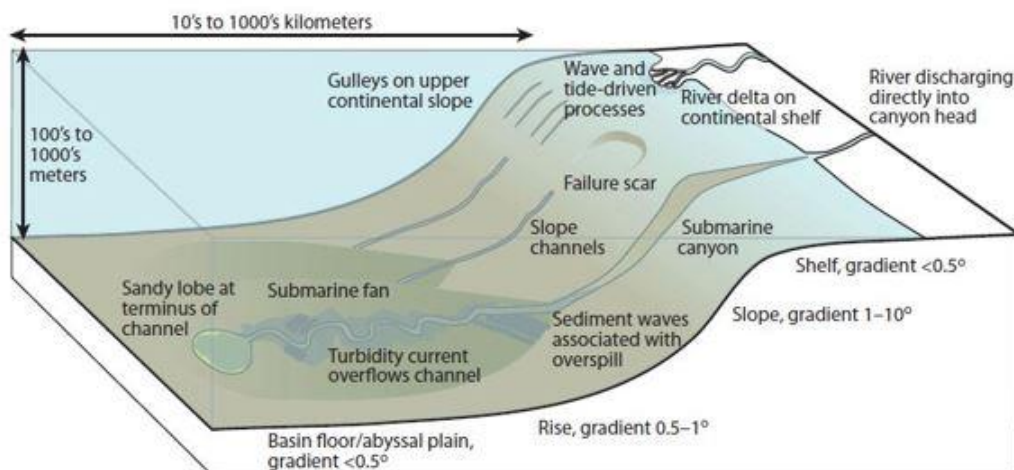


Figure 3.1: Sediment-gravity flows transport large volumes of clastic material from the shelf down the slope to the ocean floor where it is deposited in large-scale depositional systems called submarine fans (from Meiburg and Kneller, 2010).

The Windermere Supergroup deep-water turbidite system most likely represents a mixed sand-mud submarine fan sensu Reading and Richards (1994), and therefore forms an expansive (32 000 km² to 3 000 000 km²), low gradient (18 to 0.74 m/km) system fed by efficient, large volume, far traveling, low-density turbidity currents generated by major slump events. The upper fan consists of canyons and farther downslope is the middle fan which is composed of large, meandering channels confined by well-developed levees. At the base of the fan, the sand:shale ratio increases dramatically as the turbidity currents lose confinement and spread out over the basin floor. The thick channel-fills, overbank deposits and low sandstone-to-mudstone ratio (25:75) of the Isaac Formation is interpreted to be a mid-fan setting whereas the shallow distributary channels, depositional lobes and high sandstone-to-mudstone ratio (75:25) of the Kaza Group likely reflects a lower fan setting (Khan, 2012; Ross and Arnott, 2007). Examples of modern submarine-fan systems comparable in size to the ancient Windermere turbidite system include the Amazon (330 000 km²) and Mississippi (300 000 km²) fans (Reading and Richards, 1994). However, there is a marked difference in the grain size of channel fills, levee and lobe deposits in

the Windermere turbidite system compared to these modern systems, with most of these modern systems being mud-rich. At Castle Creek, the abundance of coarser grain sizes of these architectural elements is most comparable to the mixed sand-mud model (Fig. 3.2).

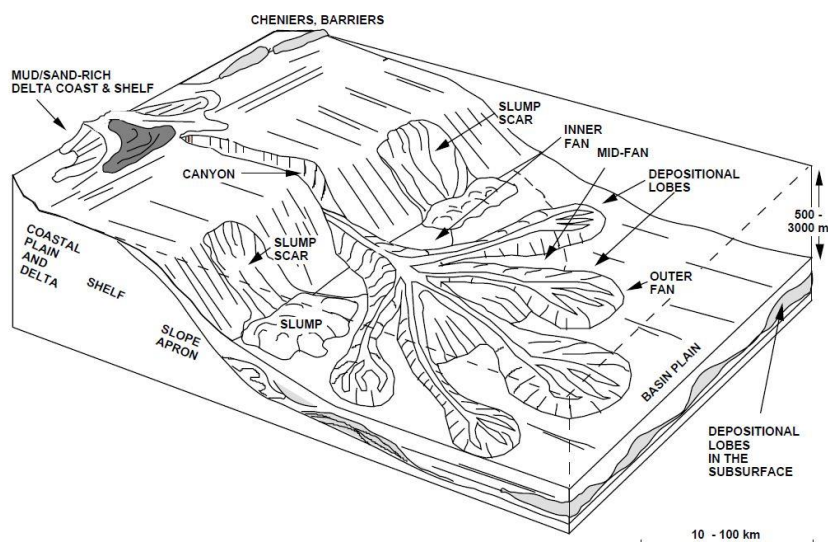


Figure 3.2: Idealized depositional model for a point-source, mixed sand-mud submarine fan. The Windermere Supergroup deep-water turbidite system most likely represents a similar model (from Reading and Richards, 1994).

3.2 Architectural Elements

Submarine fans are composed of an assemblage of geomorphic features termed architectural elements (Fig. 3.3) (Posamentier and Kolla, 2003). Architectural elements are the basic mappable geomorphic elements of modern and ancient turbidite systems that are recognised in seismic and outcrop studies, and, which individually consist of a distinctive assemblage of facies and facies associations (Mutti and Normark, 1991). Architectural elements commonly identified in deep-marine turbidite systems include channels, levees, overbank splays, crevasses splays and depositional lobes (Posamentier and Kolla, 2003). This study focuses on a ~150 m thick interval of strata deposited on the base of slope. Based on distinctive assemblages of structureless, normally graded sandstone, structured sandstone and medium-scale cross-stratified sandstone (facies 1-3)

(see Chapter 4), both channels and levees were identified. In addition, these channels and levees are intercalated with slides and debrites (facies 4). An overview of channels and levees, the main focus of this thesis, is described next.

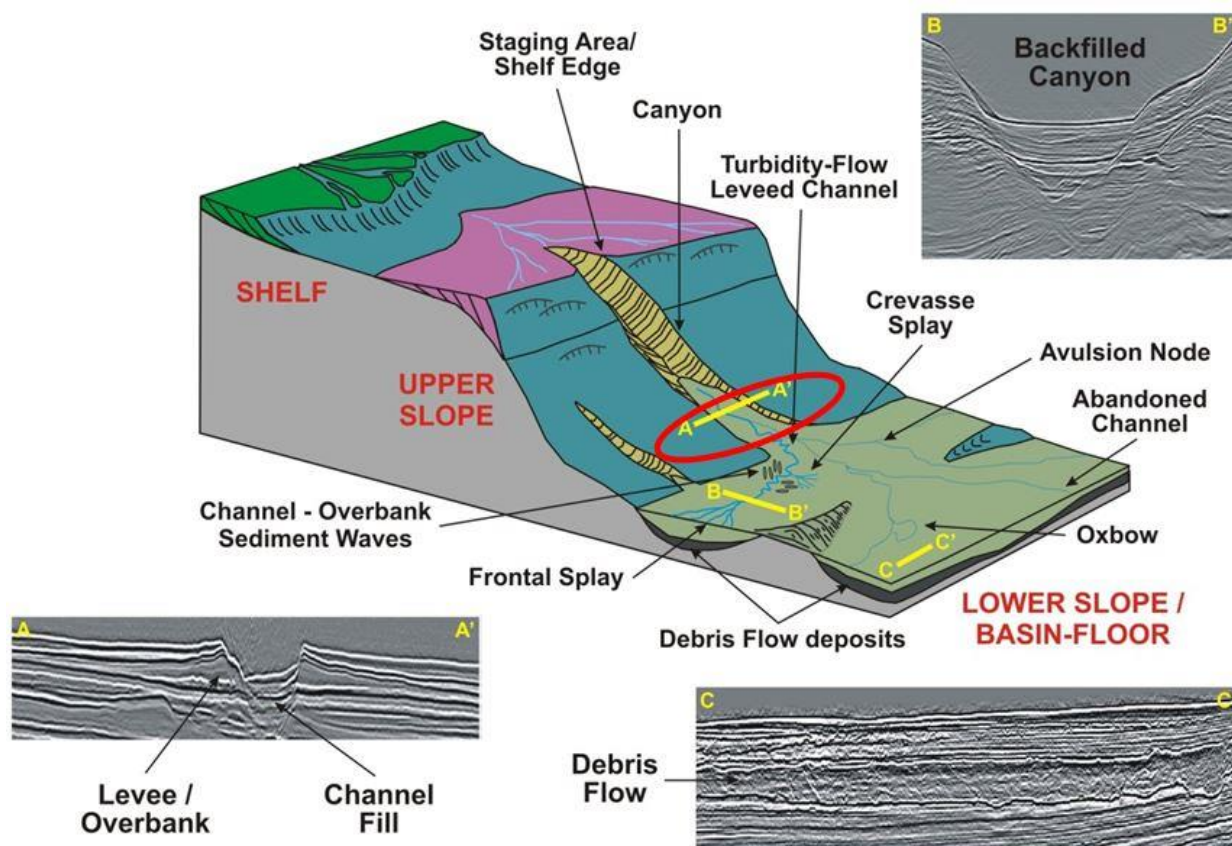


Figure 3.3: Submarine fans are composed of an assemblage of geomorphic features termed architectural elements. The main architectural elements that make up deep-water environments are shown on this schematic. Red circle highlights the base of slope- the focus of this study (from Posamentier and Kolla, 2003).

3.3 Channels

Channels are a common element in deep-marine systems, are excellent hydrocarbon reservoirs and are typically well exposed in outcrop because their sandy fills are resistant to weathering. Accordingly, channels are well documented in the literature (for example, Mutti and Normak, 1987; Pickering et al., 1995; Bain and Hubbard, 2016). A submarine channel is a negative topographical element that acts as a long-term conduit for sediment transported basinward by

turbidity currents and mass movements (Arnott, 2007; Covault et al., 2014). Channels are large features that range up to several thousand kilometers long, hundreds of meters deep and several kilometers wide (Fig. 3.4) (Clark and Pickering, 1996, Khan, 2012).

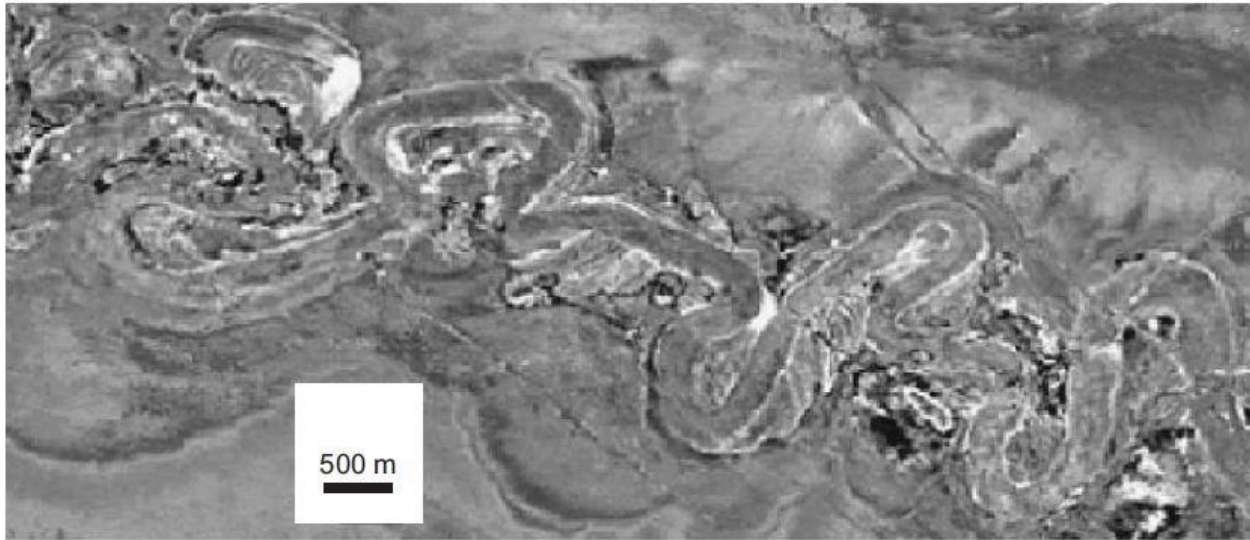


Figure 3.4: Horizon 3D seismic slice of the top of the top of the Green Channel Complex, offshore Angola (Wynn et al., 2007 modified from Abreu et al., 2003).

For a given set of average flow conditions, an equilibrium gradient will exist in order to maximize basinward sediment transport (Fig. 3.5). When out of grade, submarine channels, like fluvial channels, either erode or aggrade to achieve their equilibrium profile. In the case of submarine channels, base level is determined by gravity base, that is, the lowest point that a turbidity current can flow on a particular pathway (Kneller, 2003). Where the gradient of the channel and sediment-transporting flows are in equilibrium, the channel is at grade and the majority of the sediment is bypassed to areas farther downflow (Fig. 3.5 C) (Kneller, 2003; Arnott, 2010). This condition likely only describes average flows passing through a channel, as deposition will occur in smaller than average flows and erosion in larger than average flows (Khan, 2012). Normark (1970) subdivided channels into three types: erosional, depositional and mixed (erosional/depositional) based on observations in both ancient and modern systems (Clark and

Pickering, 1996). Erosional channels form where flow parameters change in a way that causes the channel profile to lie above the graded profile, and as a consequence the channel erodes to flatten its profile (Fig. 3.5 A) (Kneller, 2003). Erosional channels are commonly associated with areas of steep slope, low sinuosity planforms, and coarse grained, high concentration turbidity currents (Clark and Pickering, 1996). Channels form a topographic low erosionally bounded by older strata (Arnott, 2010).

In contrast to erosional channels, depositional (aggradational) channels form where flow parameters change in a way that the channel profile lies below the graded profile (Fig. 3.5 B). In this case, channels aggrade to steepen their profile. However, it is important to note that aggradation must be balanced between the channel axis and overbank areas in order to maintain channel confinement (Kneller, 2003). Depositional channels are associated with high channel sinuosity, low slopes and fine grained turbidity currents (Clark and Pickering, 1996). These channels are bounded by channel-margin levees that become increasingly elevated above the seafloor with time (Arnott, 2010). Mixed channels (erosional/depositional) alternate between channel-axis erosion and levee deposition, and, may lie above or below the regional seafloor surface (Fig. 3.5 D) (Arnott, 2010).

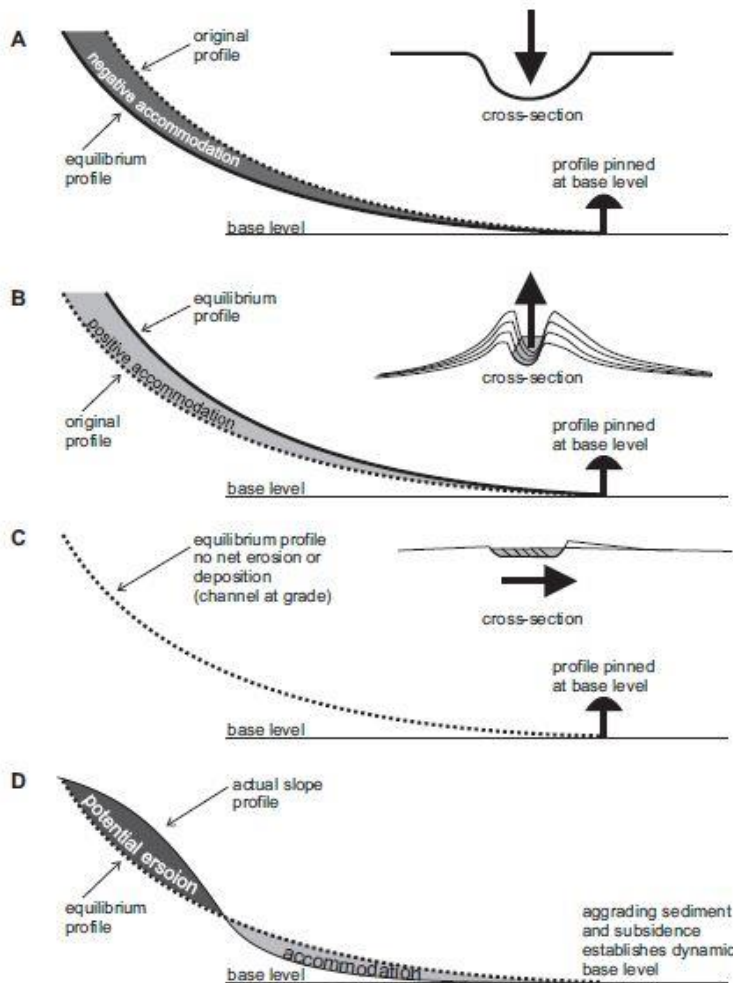


Figure 3.5: Adjustment of a channel profile to maintain equilibrium transport conditions. (A) A channel profile that lies above the graded profile must erode to flatten its profile (Kneller, 2003). (B) A channel profile that lies below the graded profile must aggrade to steepen its profile (Kneller, 2003). (C) When the channel is at grade and the majority of the sediment is bypassed to areas farther downflow (Kneller, 2003; Arnott, 2010). (D) Mixed channels (erosional/depositional) alternate between channel-axis erosion and levee deposition at different locations downflow in order to maintain equilibrium transport conditions (Arnott, 2010). Figure A-C redrawn by Khan, 2012 from Kneller, 2003. Figure D redrawn by Khan, 2012 from Kneller, 2003 after Samuel et al., 2003.

Channels are also classified based on the degree of confinement. Highly confined channels are those in which throughgoing flows are contained mostly within the channel and typically occur in the upflow parts of a turbidite system. Further downflow confinement decreases and forms leveed-channel systems. Here the upper part of the flow commonly overflows the sides of the channel, building channel-bounding depositional levees. Farther downflow the system evolves into

a complex of poorly confined channels associated with a depositional lobe at the terminus of the channel (Arnott, 2010).

Channels commonly experience a systematic evolution including channel inception, bypass, fill and finally abandonment. Channel inception consists of a period of net erosion, where multiple flows with high transport efficiency scour a sharp-based topographical low that acts principally as a conduit for flows transporting sediment basinward (Arnott, 2010). This stage is followed by the channel-bypass stage, where the majority of throughgoing flows neither erode nor deposit sediment (complete bypass). Channel lag deposits are commonly formed as throughgoing flows rework any previously deposited sediment (Clark and Pickering, 1996). Incomplete bypass can also occur where flows leave behind a small part of their sediment load (Navarro, 2005). During bypass, channel overflow and flow stripping (see section 4.4) cause levee deposits to build up along the channel margins (Clark and Pickering, 1996). Channel-bypass is then succeeded by channel fill during which the transport efficiency of the flows decreases, and sediment is deposited within the channel, partly to completely filling it (Arnott, 2010). Strata typically consist of thick, coarse-grained, amalgamated or layered deposits (Clark and Pickering, 1996; Navarro, 2005). Finally, and as a result of autogenic controls, including upflow channel avulsion, or allogenic controls, like a relative sea-level rise, the channel becomes gradually or abruptly abandoned resulting in the deposition of a fine-grained hemipelagic drape (Clark and Pickering, 1996; Arnott, 2010).

Numerous classification schemes have been proposed to describe the architectural elements that make up submarine channels (for example: Mutti and Normark, 1987; Pickering et al., 1995; Moraes et al., 2004). The combined channel classification scheme proposed by Navarro (2005) will be used in this study (Fig. 3.6). The basic element is a discrete channel fill, which is a

negative topographic feature that confined turbidity currents (Mutti et al., 2000). Channel fills are less than 10 m thick and 200 m wide. A channel unit, in turn, comprises two or more genetically related channel fills, and a channel complex comprises two or more stacked channel units. A channel complex set consists of two or more channel complexes marked by a facies change or an abandonment surface at its base or top (Abreu et al., 2003). The highest order of submarine channel elements is the channel system, which comprises two or more genetically related channel complex sets.

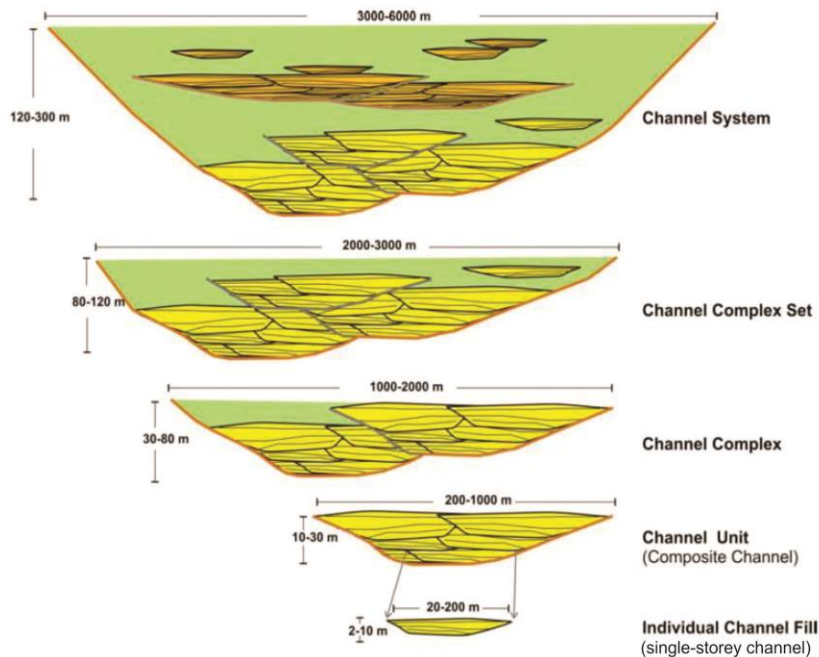


Figure 3.6: The combined channel classification scheme proposed by Navarro (2005) to describe the components that make up submarine channels.

3.4 Levees

Sinuuous submarine channels in slope environments commonly have levees of varying development along their margins (Fig. 3.7) (Posamentier and Kolla, 2003). In cross-section, they exhibit a characteristic “gull wing” or wedge-shaped geometry caused by the lateral thinning and fining of beds away from the channel (Fig. 3.8) (Skene et al., 2002; Beauboeuf, 2004; Arnott, 2010, Kane et al., 2011; Morris et al., 2014). Levees form as flows overspill the channel margins

and deposit sediment, forming topographical highs that act to partially confine flow to the channel (Arnott, 2010; Khan and Arnott, 2010; Kane et al., 2011; Morris et al., 2014). Compared to fluvial systems, the density contrast between the overspilling flow and the ambient fluid in submarine systems is considerably reduced (Kane et al., 2009), and as a consequence of its more buoyant nature allows overspilling flows to travel for up to several tens of kilometers laterally, forming significant levees that commonly are an order of magnitude or more wider than their associated channels (Posamentier and Kolla, 2003). In the modern submarine Mississippi fan, for example, the Mississippi Channel is 2-4 km wide and 25-45 m deep but its levees are 50 km wide and up to 200 m high (Wynn et al., 2007). Thus, by volume, levees make up a significant portion of deep-water turbidite systems (Khan and Arnott, 2011; Straub et al., 2011). Additionally, they have been proven to form productive hydrocarbon reservoirs, such as the levee-overbank turbidite sands in the Ram/Powell Field, deepwater Gulf of Mexico (Clemenceau et al., 2000; Skene, 1998). In spite of this, levees remain one of the most poorly understood deep-marine sedimentary systems (Kane et al., 2009; Sylvester et al., 2011). In modern systems their large size and thickness makes capturing their lithological and architectural details difficult. In the ancient sedimentary record, on the other hand, the generally fine-grained nature of levee deposits causes them to be typically extensively weathered and poorly exposed (Khan and Arnott, 2011). Therefore, many channel-levee models are typically based on high-resolution seismic images of modern systems. Although seismic images are useful for determining large-scale channel-levee geometries, outcrop studies are needed to acquire details about the internal architecture of levee deposits that cannot be resolved from even high-resolution seismic (Khan, 2012; Davis, 2011).

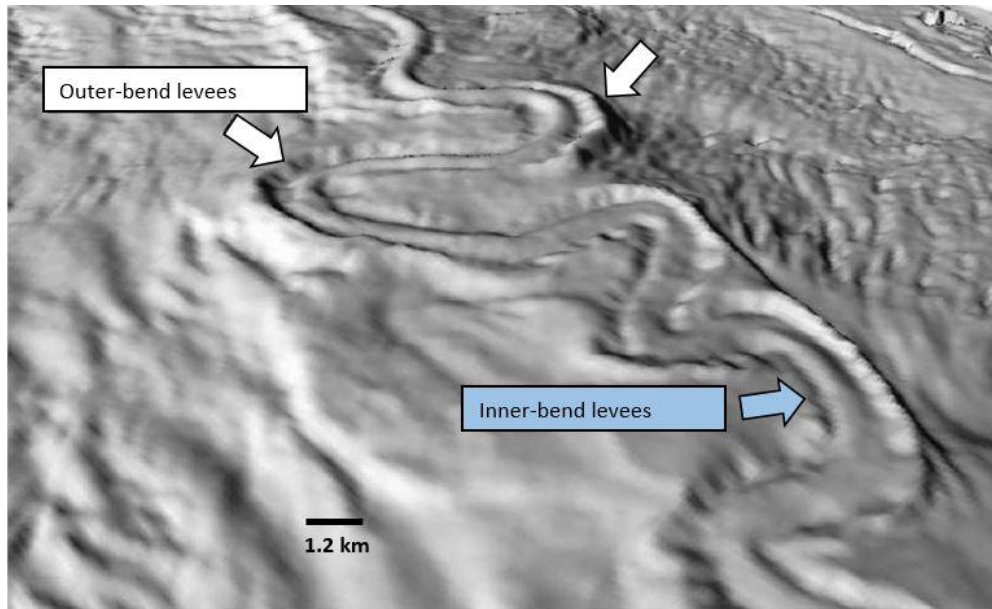


Figure 3.7: Three-dimensional view of the Joshua channel belt, Gulf of Mexico. Note the well-developed levees formed along both sides of the channel. Outer-bend levees (white) tend to be thicker and better developed compared to their inner-bend (blue) counterparts. (Modified from Posamentier, 2003).

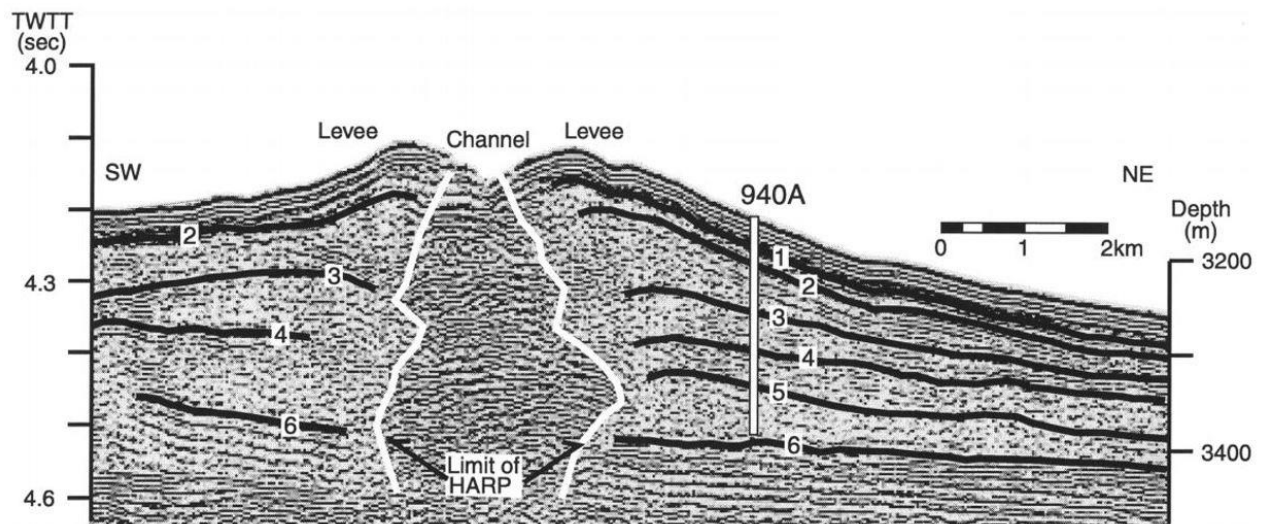


Figure 3.8: Seismic profile across a channel-levee complex in the Amazon fan. Channel deposits are between the white lines. Note how the levee deposits on either side of the channel exhibit a gull wing-shaped geometry away from the channel margins (from Piper and Normark, 2001).

Three main types of overspilling processes build up channel-margin levees: flow stripping, inertial overspill and continuous overspill. Flow stripping occurs in channel bends where flow momentum causes the upper, more dilute, finer-grained portion of the flow to become detached

from the lower, higher-concentrated, coarser-grained part of the flow confined to the channel (Fig. 3.9). The upper part then over spills the channel banks and spreads out over the levee (Fildani et al., 2006; Arnott, 2010). Superelevation created in channel bends acts to augment the effects of flow stripping by increasing the thickness of the flow along the outer bends, causing the surface of the flow to slope towards the inner bends (Khan, 2012). During inertial overspill the entire flow, including the lower coarser-grained portion, may spill over the outer channel bends due to particle inertia (Fig. 3.10) (Khan, 2012). Continuous overspill occurs where the throughgoing flow is thicker than the depth of the channel (Fig. 3.11). As a result, the portion of the flow above the channel margins continuously over spills and builds up levees on both sides of the channel. Levees along the inner bends and the straight portions of channels form principally by continuous overspill (Arnott, 2010).

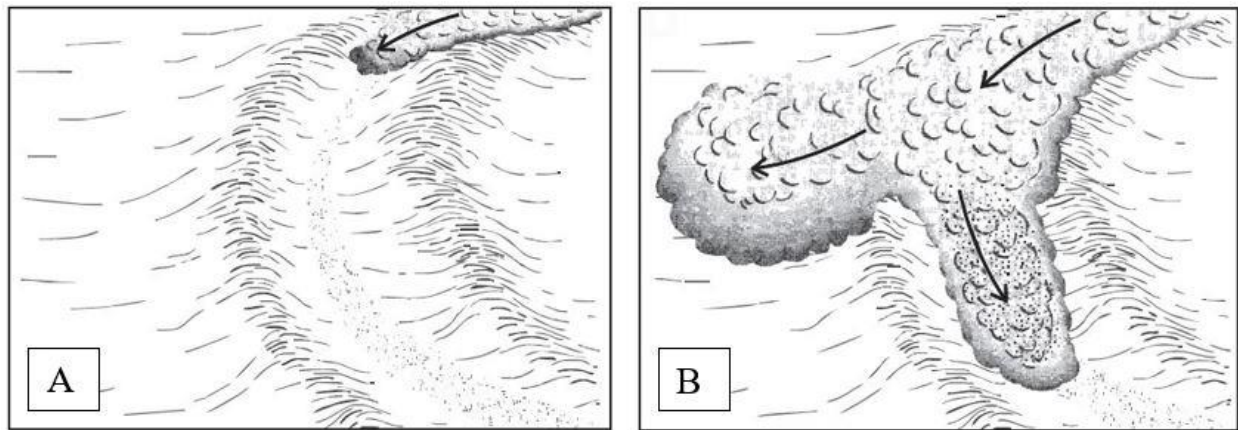


Figure 3.9: Process of flow stripping. Flow momentum causes the upper, more dilute, finer-grained portion of the flow to become detached from the lower, higher-concentrated, coarser-grained part of the flow and to overspill (Fildani et al., 2006; Arnott, 2010). (Redrawn by Khan, 2012 from Peakall et al., 2000.)

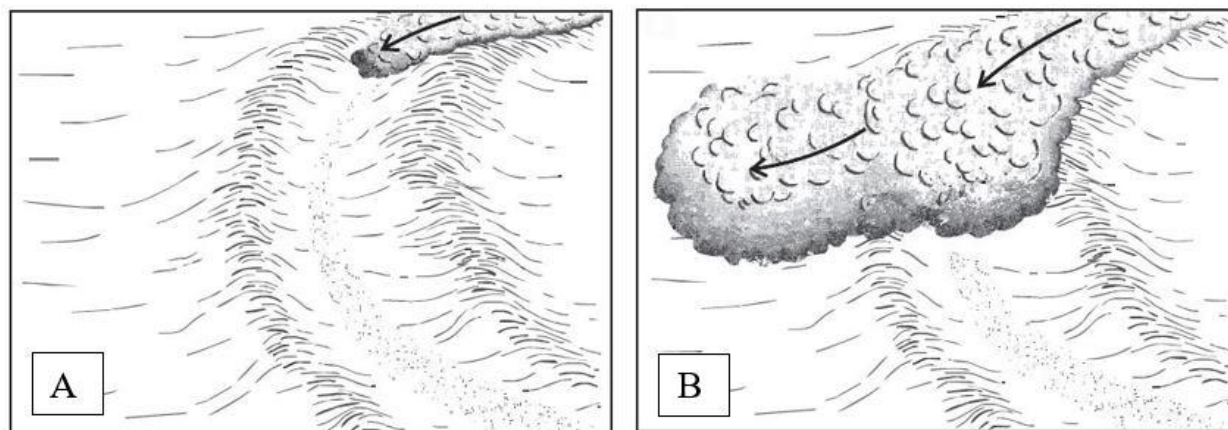


Figure 3.10: Process of inertial overspill. The entire flow, including the lower coarser-grained portion, may overspill the outer channel bends due to particle inertia (Khan, 2012). (Khan, 2012 after Kane et al., 2007.)

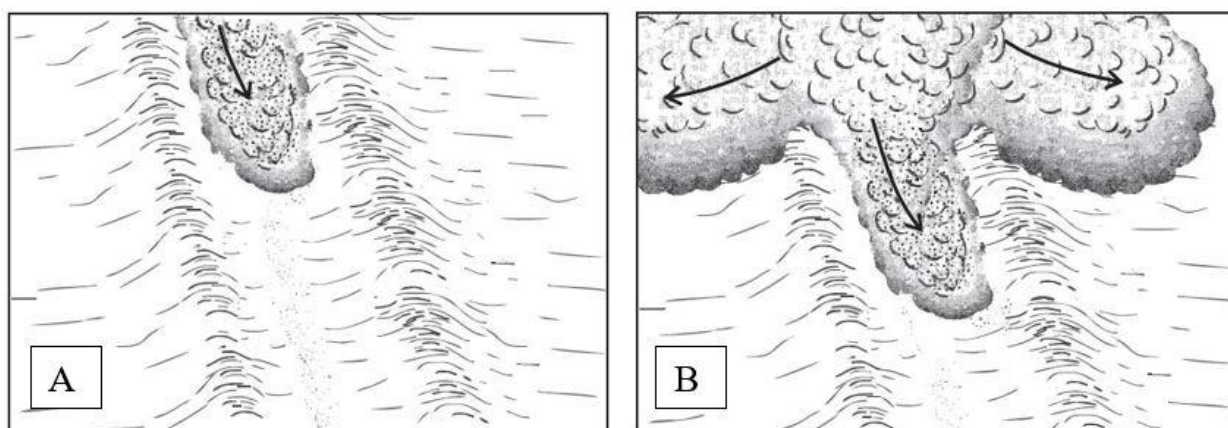


Figure 3.11: Process of continuous overspill. Where throughgoing flows are thicker than the depth of the channel the portion of the flows that are thicker than channel margins overspill and build up levees on both sides of the channel (Arnott, 2010). (Khan, 2012 after Peakall et al., 2000; Kane et al., 2007.)

Walker (1985) reported that levee strata commonly exhibit three attributes: climbing ripples, convolute lamination and mud rip-up clasts, terming them CCC-turbidites (Fig. 3.12). Climbing ripples have been observed in a number of modern and ancient proximal-levee deposits and indicate rapid sediment fallout from suspension as the flow expands, although not so rapid as to completely suppress bedload transport (Fig. 3.13) (Arnott, 2010; Kane et al., 2007). Convolute lamination is also suggestive of high rates of sediment deposition, which results in low-sediment yield strength and therefore deformation by shearing from the same or later turbidity currents (Fig.

3.13). Although CCC-turbidites have been observed in many submarine levee deposits (for example: King et al., 1994; Coleman, 2000; Hickson and Lowe, 2002; Kane et al., 2007) they are uncommon in the Castle Creek study area. Here ripples typically show negligible or low angles of climb, and convolute lamination is rarely observed, which might be related to coarser grain size and/or poorer sorting, or lower rates of sediment fallout (Arnott, 2010; Khan, 2012). In addition, mud rip-up clasts are generally uncommon.

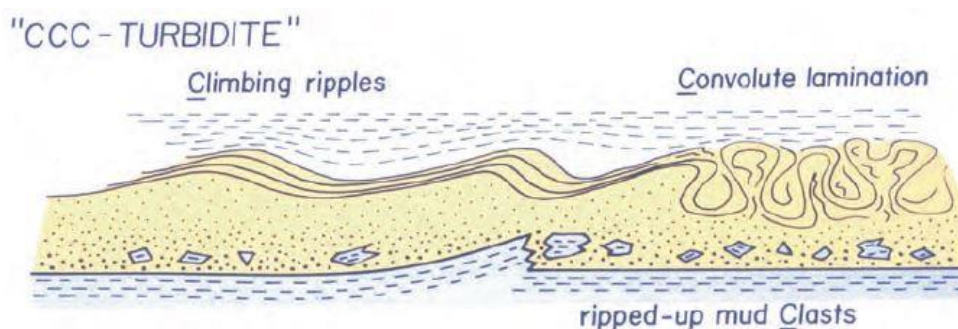


Figure 3.12: Walker (1985) reported that levee strata commonly exhibit three attributes: climbing ripples, convolute lamination and mud rip-up clasts, terming them CCC-turbidites (Posamentier and Walker, 2006).

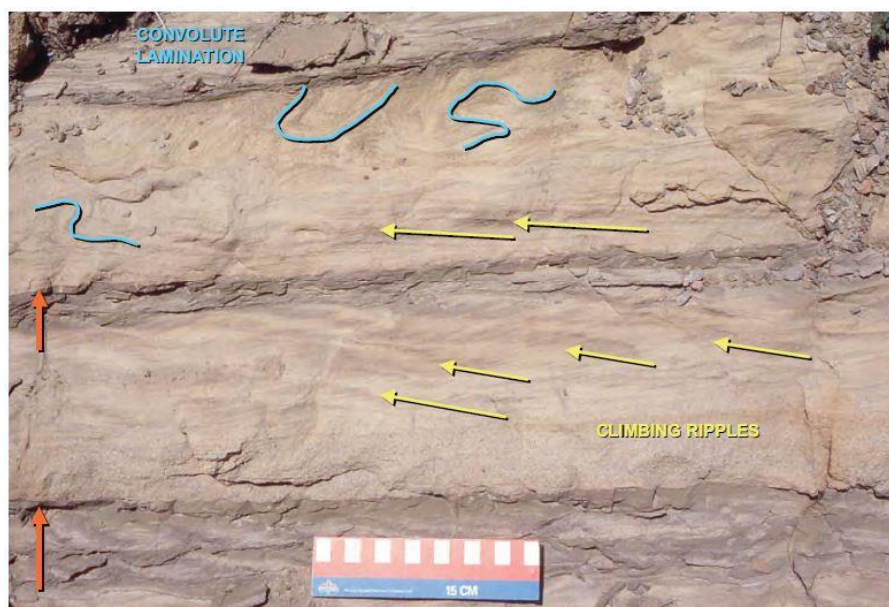


Figure 3.13: Thin-bedded turbidites in the Chatsworth Sandstone, California with two of the three "Cs". Orange arrows indicate the base of each bed, yellow arrows indicate climbing ripple sets and convolute lamination is highlighted in blue. Note the absence of mud clasts (Posamentier and Walker, 2006).

4 Facies Descriptions and Interpretations

Four facies are recognized in this study. Facies are subdivided based on bed thickness, sedimentary structures and grain size. Bed thickness follows the classification of Ingram (1954): very thin-bedded (1- 3 cm), thin-bedded (3- 10 cm), medium-bedded (10- 30 cm), thick-bedded (30- 100 cm), and very thick-bedded (> 100 cm). Grain size is described using the Wentworth grain size classification scale (Wentworth, 1922).

4.1 Facies 1 (F1): Structureless, Normally Graded Sandstone

Strata consist of well-sorted, structureless, normally graded, light pink sandstone (Fig. 4.1), with structureless indicating an absence of tractional sedimentary structures. Basal contacts are undulatory with common flame structures. Strata typically grade upward from lower coarse- or upper medium- to lower medium-grained sandstone. Beds range from 39-139 cm, and on average are 87 cm thick. Mudstone intraclasts typically occur locally and in the basal part of beds where they range from 114 cm x 3 cm (large) to 1.5 cm x 1cm (small), and are oriented with their long axis parallel to the base of the bed. These strata are then capped by an interlaminated sandstone/siltstone and mudstone unit that ranges from 3-12 cm thick, with the top ~1-2 cm composed of structureless mudstone.



Figure 4.1: Photos of very thick-bedded, structureless sandstones grading upward from lower coarse- to lower medium-grained sand.

4.1.1 Interpretation

Beds of F1 are interpreted to represent incomplete Bouma turbidites deposited from decelerating turbulent flows that contained clay to lower coarse sand. Specifically, strata of F1 represent the Tade divisions of a classical Bouma sequence. The sand rich Ta part of the bed was deposited rapidly from high-density turbidity currents (Lowe, 1982), or similarly concentrated or hyper-concentrated flows (Shanmugam, 1996, 1997; Mulder and Alexander, 2001). Particle support in the flows was a combination of particle-particle interaction, buoyancy and turbulence. The structureless nature of the bed is also suggestive of high rates of suspension fallout, which in this case muted the development of tractional sedimentary structures like planar lamination or cross-stratification (Arnott and Hand, 1989; Leclair and Arnott, 2006). Arnott and Hand's (1989) experiments showed that parallel lamination became more poorly developed as the rate of suspension fallout, and therefore bed aggradation, increased. Even if the flow velocity is within

the stability field for upper plane bed, dunes or ripples, incipient bedforms become buried before they've had sufficient time to develop. The rate of sediment fallout needed to suppress the formation of tractional structures is variable and depends on grain composition, size, shape and density (Arnott and Hand, 1989).

Common load and flame structures on the bases of F1 beds suggest substrate instability. This arises due to the insufficient strength of the undulating mud-rich layer and its inability to support the weight of the rapidly deposited sand-rich Ta part of the next flow event, causing the latter to founder into the former. Such conditions indicate a short recurrence interval between successive turbidity currents.

As flow velocity continues to decrease, silt and mud is added to the bed load layer. The planar interlaminated sandstone/siltstone and mudstone Td part near the top of the bed indicates alternating traction and suspension sedimentation processes. The ~1-2 cm thick, structureless fine-grained Te layer at the top of the bed formed during the final flow stages as the dilute tail of the flow deposited mud by full suspension settling (Lowe, 1982). With full flow cessation, hemipelagic fallout, interpreted to represent background, non-turbidite sedimentation, may have contributed to the formation of these mudstone caps (Angus, 2016; Meyer, 2004).

4.2 Facies 2 (F2): Traction Structured Sandstone

Strata consist of well-sorted, planar- or cross-stratified sandstone that are light pink or orange (Fig. 4.2). Basal contacts are usually sharp and planar (79%) but undulatory basal contacts with flame structures are not uncommon (21%). Mudstone intraclasts, although uncommon, occur locally in the basal part of a few (4%) F2 beds. They range from 98 cm x 8 cm (large) to 4 cm x 0.5 cm (small), and are oriented with their long axis parallel to the base of the bed. Beds range from 2-326 cm thick but generally (86%) are on the order of 3-30 cm thick. Very thin-bedded and

thin-bedded strata typically grade upward from lower medium- to fine-grained sandstone or upper fine- to lower fine-grained sandstone. Medium-bedded to thick-bedded strata typically grade upward from lower medium- to fine-grained sandstone or from upper medium-grained to fine-grained sandstone. Uncommon very thick beds consist of lower coarse-grained to upper coarse-grained sandstone in the basal part.

Most (66%) F2 beds consist of 3 to 7 non-climbing, small-scale cross-stratified sets that form cosets ranging from 3-27 cm thick, but on average are 9 cm thick. Sets range from 0.3-4 cm and average 1.7 cm in thickness. Typically they are interlaminated, especially in the uppermost part of the bed, with very fine-grained sandstone to siltstone. This unit is then overlain sharply by a thinly-interlaminated sandstone-siltstone and mudstone unit that ranges from 0.5-10 cm thick, with the top ~1 cm composed of structureless mudstone. Less commonly (18%) F2 beds consist of a single non-climbing cross-stratified set overlain by thinly-interlaminated sandstone-siltstone and mudstone unit that ranges from 1-5 cm thick with the top ~1 cm composed of structureless mudstone.

Some (16%) F2 beds consist of a basal planar laminated unit overlain sharply by a small-scale cross-stratified unit, and in turn a thin, structureless siltstone unit. The parallel-laminated part makes up approximately 9% to 98% of the total bed thickness and ranges from 2-304 cm thick. The overlying cross-stratified unit is typically composed of 2 to 3 non-climbing sets, although single sets are observed. The thinly planar laminated sandstone/siltstone and mudstone unit that overlies the cross-stratified part of the bed is usually about 8 cm thick with the top ~1 cm composed of structureless mudstone.

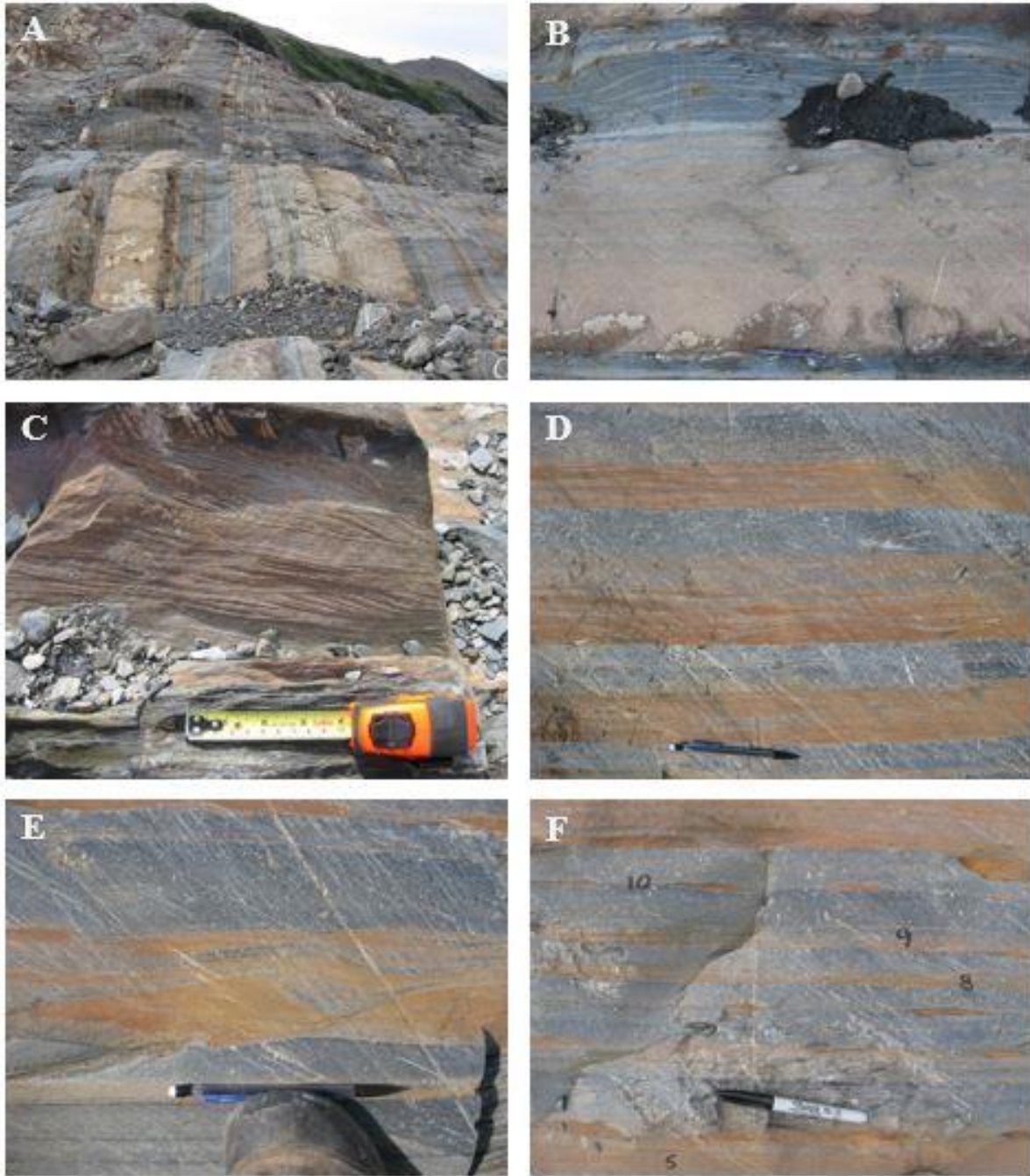


Figure 4.2: Photos of F2 strata. (A) Medium- and thin-bedded Tbcde/Tcde turbidites make up most of the stratigraphy in the study areas. (B) Thick-bedded, upper medium-grained Tbcde turbidite. (C) A ~15 cm thick, fine-grained, multiple set Tc turbidite. Individual ripple sets are 2-3 cm thick. Most Tc turbidites consist of 3 to 7 non-climbing, small-scale cross-stratified sets that form cosets that on average are 9 cm thick. (D) Three thin-bedded, fine-grained multiple set turbidites. (E) Thin-bedded, fine-grained, multiple set Tcde turbidite. Note how the fine-grained sand (orange) is interlaminated, especially in the uppermost part of the bed, with very fine-grained sandstone (light orange) to siltstone (grey). (F) Very thin-bedded, fine-grained, single set Tcde turbidites (beds are numbered 8-10).

4.2.1 Interpretation

F2 beds are interpreted to represent incomplete Bouma turbidites deposited from decelerating turbulent flows that contained clay- to upper coarse-grained sand particles suspended by flow turbulence (Lowe, 1982). Most basal contacts are sharp and planar indicating that flows generally became immediately depositional. Erosion is recorded in beds with undulating (scouring) basal contacts and rip-up mudstone intraclasts at their bases.

The basal planar-laminated part observed in some F2 beds is interpreted to represent the Tb division of a classical Bouma sequence, whereas the small-scale cross-stratified unit is interpreted to represent the Tc division and the thin, structureless siltstone unit is interpreted to represent the Td/e division. The Tb division is interpreted to have been deposited by flows with lower rates of fallout, and most likely also lower sediment concentration compared to the flows that deposited the Ta division of F1 strata (Arnott and Hand, 1989). Planar lamination forms as particles in the near-bed layer move and become sorted on the aggrading bed's surface by the overriding turbulent flow (Sumner, 2008). According to standard bedform stability diagrams (e.g. Southard and Boguchwal, 1990) as flow velocity decreases for bedload sediment coarser than middle fine sand, the Tb division should be succeeded by dune cross-stratification (Fig. 4.3). However they are everywhere overlain by the small-scale cross-stratified Tc or Td divisions, and not by dune cross-stratification. The absence of dune cross-stratification in the Bouma model and its general absence in the ancient and modern deep-marine sedimentary record is an ongoing source of debate. Previous explanations focused on the temporal and/ or hydraulic conditions needed for angular bedform development (Tilston et al., 2015), however recent attention has been directed toward the mechanisms by which angular bedforms become initiated (Arnott, 2012; Tilston et al., 2015).

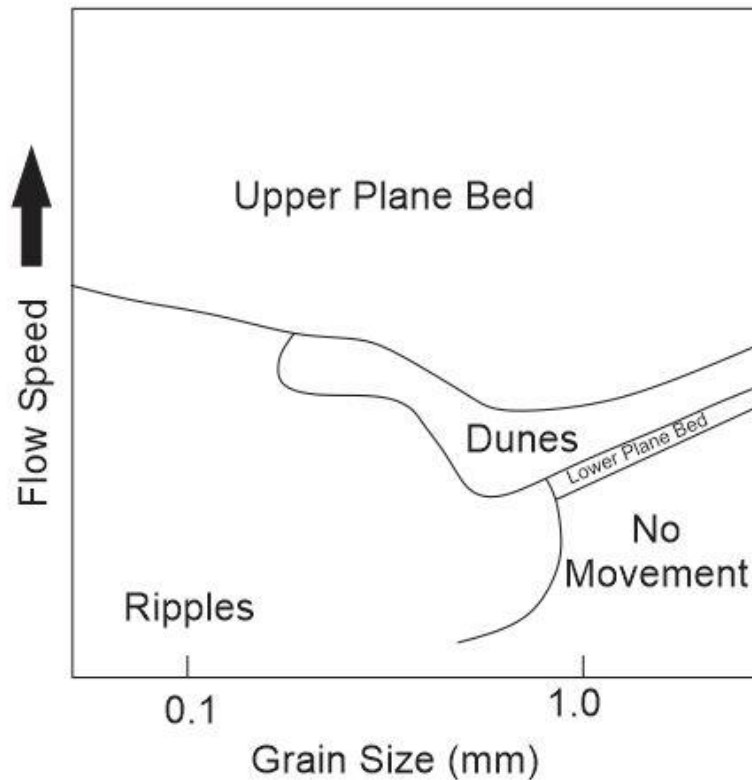


Figure 4.3: Hypothetical bedform stability diagram for low concentration turbidity currents (redrawn from Davis, 2011 after Southard and Boguchwal, 1990). As flow velocity decreases for bedload sediment coarser than middle fine sand, the Tb division (upper plane bed field) should be succeeded by dune cross-stratification before entering the ripple stability field.

Angular bedforms, like ripples and dunes, grow from incipient features called bed defects. Increased turbulence associated with the onset of near-bed flow separation over and/or around the defect causes the flow to scour the bed (Venditti et al., 2005), which then amplifies the height of the feature. Early models for the origin of bed defects focused on pre-existing defects, for example an isolated bed-surface obstacle like a stone (Raudkivi, 1963). Later, Williams and Kemp (1971) suggested that it was coherent (i.e., turbulence structures) high-speed, downward-directed fluid motions that caused brief episodes of sediment transport/deposition that formed the defects. This was later advanced by Best (1992) who focused on larger aggregate vortex structures to build up defects of sufficient height that they could influence local flow conditions (Fig. 4.4). However all these turbulence related models have been questioned based on the small size of sweep-induced

defects and that angular bedforms have been shown to form in laminar flows where turbulent motions are absent (Allen, 1982; Coleman and Eling, 2000).

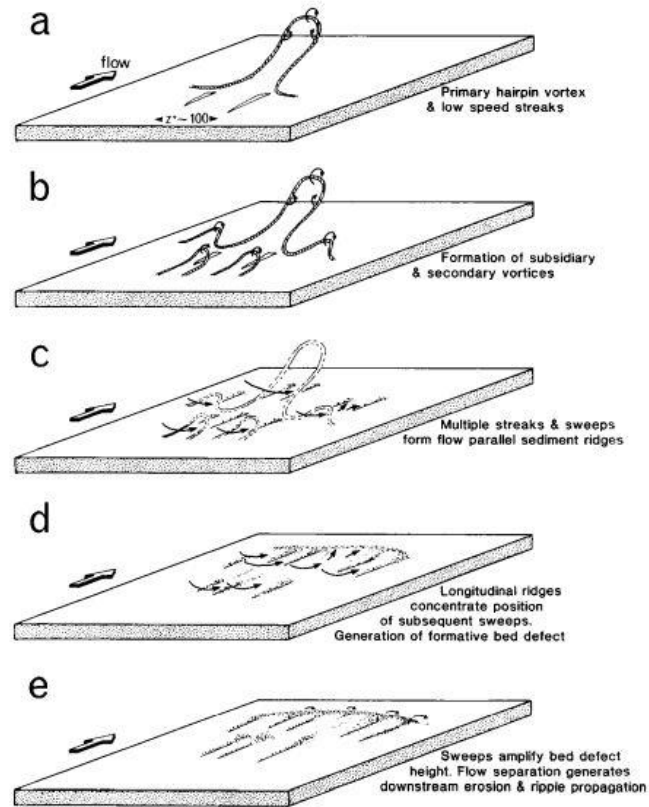


Figure 4.4: Turbulence model of Best (1992) focused on larger aggregate vortex structures to build up initial bed defects of sufficient height that they could influence local flow conditions and lead to the development of angular bedforms. As vortices (termed ejections) are lifted off the bed high-speed fluid rushes down towards the bed (termed sweeps). This inrush of high-speed fluid initiates localized sediment transport and subsequently the formation of sediment mounds that then propagate downstream.

More recently, Venditti et al. (2005) showed that bed defects can develop spontaneously over an entire plane bed when sediment transport is general and widespread, and do not have to propagate downflow from a single defect. The spontaneous initiation and growth of the defects was proposed to be related to a hydrodynamic instability of Kelvin-Helmholtz type that formed at the interface between the dense, slow-moving bed-load layer and the overlying, less dense but faster moving, near-bed fluid (Venditti et al., 2006). The regular waveform shape of the interface

caused a spatially regular pattern of sediment transport and accumulation on the bed that quickly and uniformly built up the bed defects (Fig. 4.5).

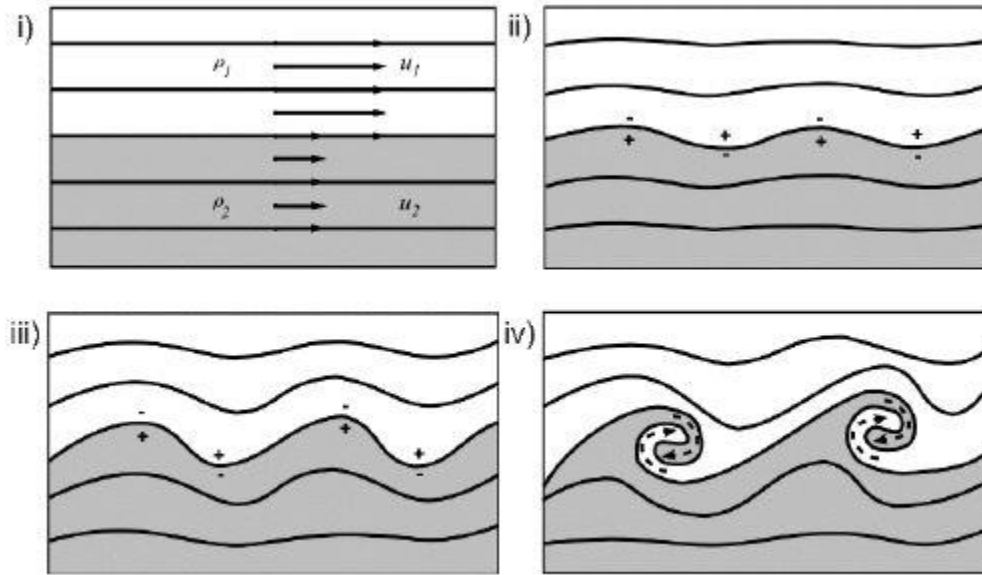


Figure 4.5: Hydrodynamic instability model of Venditti et al. (2006) for the initiation of bed defects that lead to the development of angular bedforms. In this model, the flow (top white layer labeled 1) and the bedload layer (bottom grey layer labeled 2) are treated as separate fluid layers with different densities (ρ_1 , ρ_2) and velocities (v_1 , v_2). A Kelvin-Helmholtz type hydrodynamic instability forms at the interface between the two fluids. The regular waveform shape of the interface causes a spatially regular pattern of sediment transport and accumulation on the bed that quickly and uniformly builds up bed defects, which then amplify and ultimately evolve into angular bedforms.

Arnott (2012) suggested that the lack of dune cross-stratification in turbidites could be related to the flow lacking a sufficiently well-developed near-bed hydrodynamic instability required for bed defect initiation and subsequent growth of angular bedforms. He suggested that the high suspended-sediment concentration near the base of most high energy turbidity currents inhibits the development of a sharply defined interface between the bed-load layer and the bottom of the current. His model proposed that a sufficiently well-developed density interface does not form until flow speed is generally well within the stability field for current ripples.

More recently, Tilston et al., (2015) conducted experiments that simultaneously measured the velocity and density structure of experimental turbidity currents and linked those characteristics to bed morphology. They found that sediment concentration and grain size are two parameters that alter the vertical density structure of the flow, which, in turn, controlled the final bed configuration. Specifically, a steep near-bed density gradient was required to form the hydrodynamic instability that imprints the bed and forms the bed defects from which ripples or dunes evolve. In their experiments with higher-concentration flows, or similarly finer-grained sediments, the sediment concentration in the near-bed region (i.e., flow density) showed little upward change (i.e., plug-like) (Fig. 4.6). As a result, the hydrodynamic instability was too weak or situated too high above the bed to affect sediment transport patterns on the bed (i.e., to become imprinted) and as a consequence planar lamination continued to form. The occurrence of ripple cross-stratification overlying planar lamination in F2 strata, then, is interpreted to indicate that the requisite steep near-bed density gradient needed to generate bed defects was not sufficiently developed until the flow speed was in the ripple stability field (Arnott, 2012; Tilston et al., 2015).

The Tc division is interpreted to have formed by traction transport and deposition under turbulent, unidirectional flows whose speed was within the ripple stability field. Moreover, high suspended-sediment concentration was limited to a thin near-bed that allowed the hydrodynamic instability to develop, the bed defects to form, and the ripples to grow (Fig. 4.6) (Venditti et al., 2006; Arnott, 2012; Tilston et al., 2015).

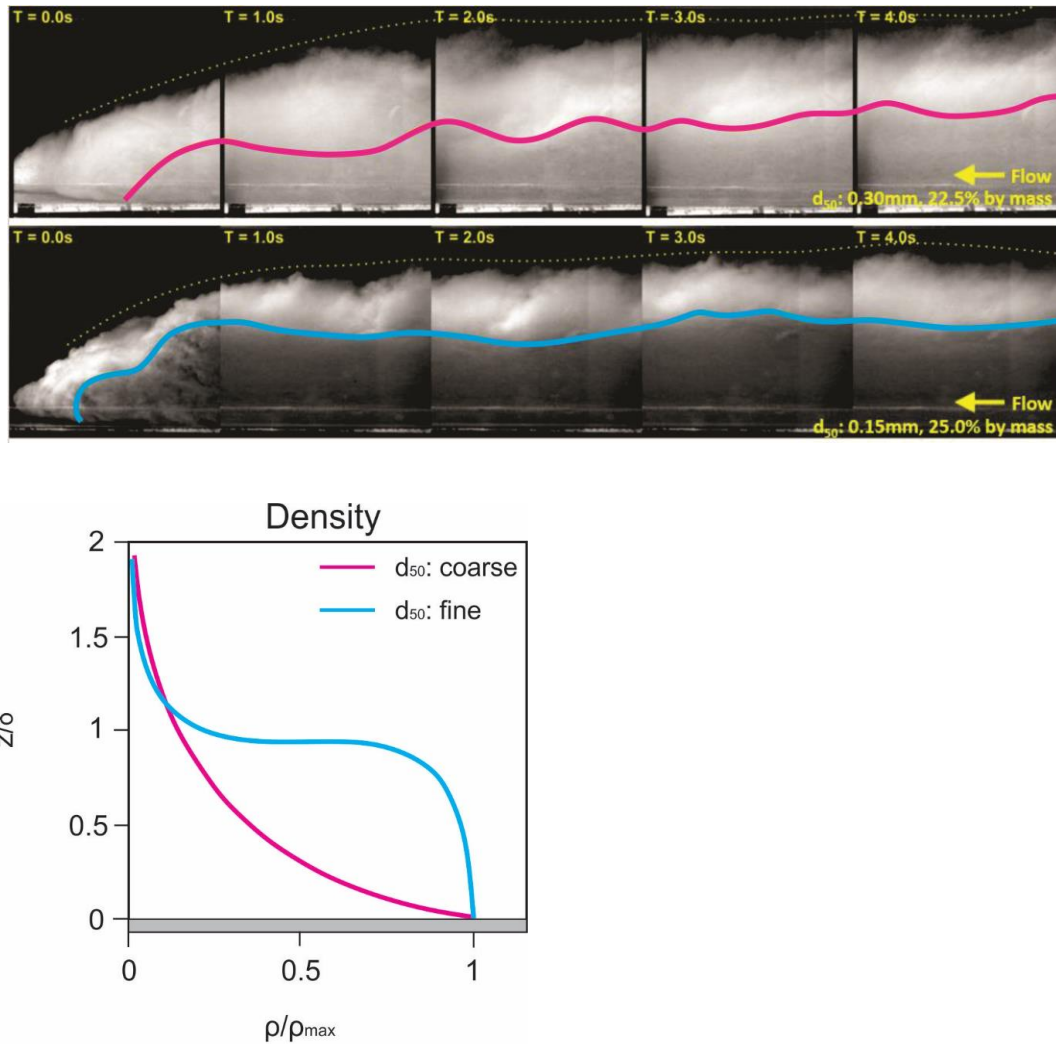


Figure 4.6: Top: High-speed, time-lapse photography showing how two turbidity currents of similar concentration and velocity but different grain sizes evolve at a point over time (modified from Tilston's experiments, 2014). The flow in the upper panel is composed of well-sorted medium sand (0.33 mm sand) compared with lower fine sand (0.15 mm) in the lower panel. The coloured traces delineate the Kelvin-Helmholtz instabilities formed in the upper part of the flow. Note the more pronounced irregularity and amplitude of the instability in the coarser flow (pink trace), which in turn leads to more intense mixing with the the ambient fluid. Bottom: Graph showing the vertical density structure of the coarse- and fine-grained flows shown above (modified from Tilston, 2016). Tilston et al. (2015) showed that a steep near-bed density gradient (i.e., coarse-grained flow) is a requisite condition to form bed defects, which then quickly evolve into angular bed forms like ripples, and presumably also dunes. In the finer-grained flow, sediment concentration in the near-bed region showed little upward change (i.e., plug-like) and as a result the hydrodynamic instability was too weak or situated too high to affect sediment transport patterns on the bed (i.e., to become imprinted) and as a consequence planar lamination continues to form.

As flow velocity continues to decrease, silt and mud is added to the bedload layer. The interlaminated sandstone/siltstone and mudstone unit that typically overlies the Tc division, but in some beds the Tb division, is interpreted to be the Td division of the Bouma sequence. The Td division is deposited by the rhythmic alternation of traction and suspension sedimentation. The ~1 cm thick structureless mudstone unit that overlies the Td division is interpreted to be the Te division formed during the final flow stages as the dilute tail of the flow deposited mud by full suspension settling (Lowe, 1982). Once the flow event terminated, hemipelagic fallout, interpreted to represent background, non-turbidite sedimentation, may have contributed to the formation of these mudstone caps (Angus, 2016; Meyer, 2004).

4.3 Facies 3 (F3): Medium-Scale Cross-Stratified Sandstone

F3 strata occur exclusively in the southern outcrop and consist of moderately well-sorted, medium-scale cross-stratified sandstone that ranges from dark pink to red (Fig. 4.7). Basal contacts are usually undulatory (68%) but sharp. Planar basal contacts are common. Beds range from 4-93 cm thick but most commonly (85%) are between 4-26 cm thick and consist of upper medium-grained (54%) or lower coarse-grained to lower very coarse-grained sandstone (30%). Upper coarse-grained to upper very coarse-grained sandstones make up a minority of the beds. In addition, approximately a quarter of all upper medium-grained, cross-stratified sandstone beds contain dispersed lower coarse sand grains to lower very coarse sand grains throughout the bed. Dispersed mudstone intraclasts occur locally within 59% of F3 beds and usually near the base of the set. Clasts typically parallel the cross-stratification, but where present in the basal part of the bed are subhorizontal. Clasts range from < 1 cm thick and 1-2 cm long to 3 cm thick and 29 cm long, although most are < 1 cm thick and a few centimetres long. Rare carbonate cemented sandstone clasts are observed and usually are 1 cm thick and a few centimetres long.

F3 beds commonly (63%) consist exclusively of cross-stratified sandstone overlain by a thin interlaminated siltstone/mudstone unit. Less commonly, the cross-stratified part of the bed abruptly overlies a massive (5%) or planar laminated unit (27%), or both (5%). The massive unit ranges from 4-90 cm thick and typically consists of lower coarse-grained sandstone. The planar laminated unit ranges from 2-62 cm thick and either consists of upper medium- or lower coarse-grained sandstone. Laterally, the cross-stratified part of these beds thickens and thins, whereas the basal part (massive or planar laminated) remains fairly consistent.

F3 commonly occur as single beds (69%) that pinch and swell along strike. For example, individual beds can thin by as much as 90% of their thickness over lateral distances of a few decimetres, where they then typically remain thin (~ few centimetres thick) over lateral distances of a few decimetres before thickening again by as much as 90%. Less commonly (31%), groups of 2 to 7 beds stack vertically where they display the same lateral thickening and thinning and then thickening trend. Beds typically (74%) overlie and are overlain by mudstone rich units; less commonly beds are interbedded with F2 strata.

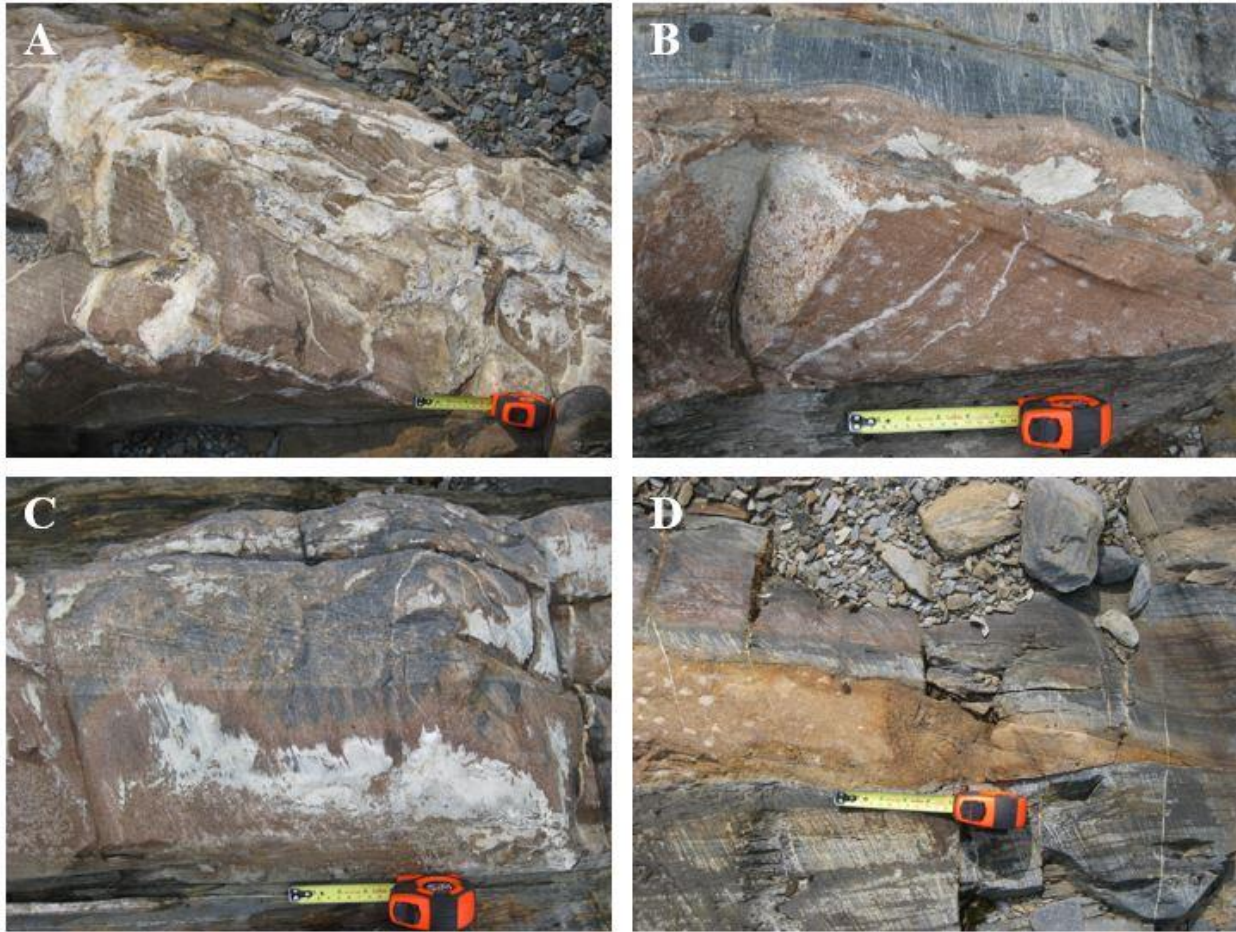


Figure 4.7: Photos of F3 strata. (A) Thick-bedded, upper coarse-grained, medium-scale cross-stratified sandstone. (B) Medium-bedded, upper medium-grained, medium-scale cross-stratified sandstone. (C) Less commonly, the cross-stratified part of the bed abruptly overlies a planar laminated unit. (D) F3 strata commonly occur as single beds that pinch and swell along strike. Here a 12 cm thick F3 bed pinches to a thickness of 1 cm over a lateral distance of 30 cm.

4.3.1 Interpretation

F3 is interpreted to have formed by traction transport and deposition under turbulent, unidirectional flows whose speed was within the dune stability field. Like ripples in strata of F2, suspended-sediment was concentration in the basal part of the flow and allowed for the development of the hydrodynamic instability needed to form bed defects, which because of higher flow velocity evolved instead into dunes (see interpretation of F2; Venditti et al., 2006; Arnott, 2012; Tilston et al., 2015).

In some F3 strata the cross-stratified part of the bed overlies a massive or planar laminated sandstone part of similar grain size. In these beds, the dunes are interpreted to be formed by tractional reworking of the underlying part of the bed and represent waning flow conditions with a change from higher energy direct suspension deposition graded sandstone (Ta) and/or planar stratified sandstone (Tb) (Kneller and McCaffrey, 2003). Where the grain size between the two parts differs, it is likely that the dune was formed by a later flow carrying a different grain size.

Like strata of F1 and F2, the thin interlaminated siltstone/mudstone unit at the top of F3 beds is interpreted to be the Td and Te divisions formed by alternating traction and suspension sedimentation processes and then eventually by full suspension settling during the final flow stages as the dilute tail of the flow deposited mud (Lowe, 1982).

4.4 Facies 4 (F4): Chaotic Mud-Rich Deposits

Facies 4 (F4) is divided into two subfacies: poorly-sorted mudstone-rich deposits (F4a) and deformed bedded deposits (F4b). Strata of F4 are important because they commonly form thick, easily identifiable units that in some cases extend across the width of the study area (> 2.5 km). They are used as stratigraphic markers to correlate between outcrops that are separated by moraine and/or the glacier.

4.4.1 F4a: Poorly-sorted mudstone-rich deposits

Facies 4a (F4a) consists of four sharply bounded, poorly-sorted mudstone-rich beds, which stratigraphically upward are termed D1- D4. Each of these is described next.

D1 is poorly-sorted and consists of pebble-size quartz, mudstone, carbonate-cemented sandstone and carbonate clasts dispersed in a silty mudstone matrix; imparting a distinctive “starry night” appearance to these strata (Fig. 4.8). D1 extends laterally across the entire study area -- a

distance of > 2.5 km. The basal contact is sharp and planar. D1 thins and thickens laterally, ranging from 0.5 m thick to 12.6 m thick but generally it is of the order of 4-9 m thick. Rounded quartz pebbles range from 0.4 cm x 0.4 cm to 3 cm x 1.5 cm but most commonly are 1.5 cm x 1 cm. Mudstone clasts range from 13 cm x 3 cm to 2.68 m x 36 m and typically are oriented with their long axis parallel to the base of the bed. Quartz pebble clasts and mudstone clasts are the most common clast types in D1. Carbonate-cemented sandstone clasts range from 4 cm x 2 cm to 7.7 m x 13 cm, and larger clasts, for example 2.5 m x 24 cm, 1.5 m x 48 cm, commonly overlie the basal contact and are oriented with their long axis parallel to the base of the bed. Clasts are composed of well-sorted, upper medium-grained sandstone that in some cases contain dispersed pebble-size quartz, mudstone and/or carbonate clasts. Carbonate clasts are rare in the southern outcrop but are more abundant in the northern outcrop where they range in size from 1 cm x 0.5 cm to 59.5 cm x 10 cm. Clasts are usually structureless but oolite and stromatolite fragments are observed locally.

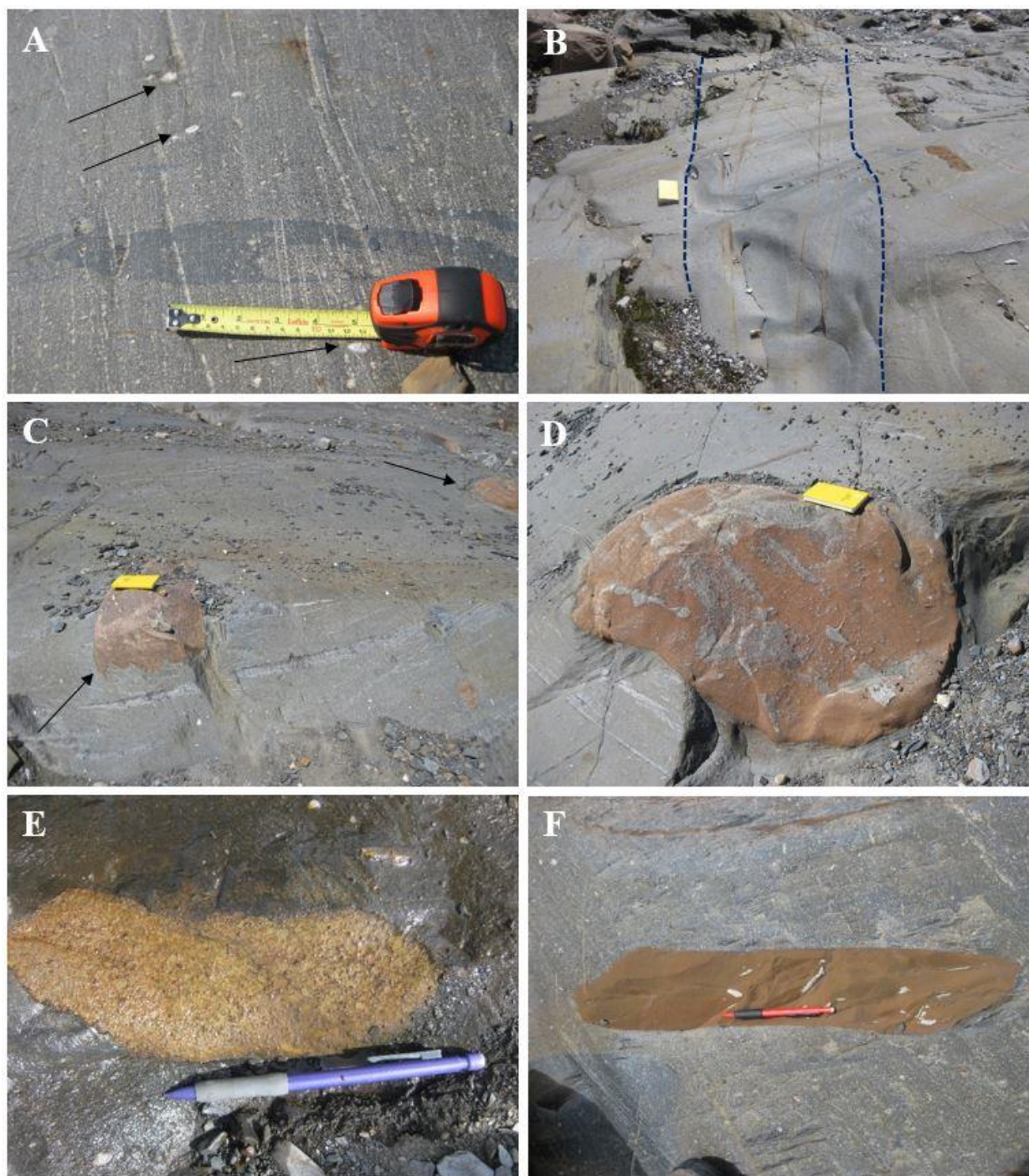


Figure 4.8: Photos of D1. (A) Pebble-size quartz clasts dispersed in a silty mudstone matrix impart a distinctive “starry night” appearance to D1. Black arrows point to quartz clasts. Small-sized mudstone clast crops out above tape measure. (B) Large clast of deformed thin-bedded, fine-grained, upper division turbidites is outlined by blue dashed lines. (C) Black arrows point to large, carbonate-cemented sandstone clasts. (D) Large carbonate cemented sandstone clast just above the basal contact. (E) Uncommon oolite clast. (F) Uncommon structureless carbonate clast.

D2 is poorly-sorted and consists of a variety of clasts in addition to lower very coarse grains and upper very coarse grains dispersed in a matrix of upper coarse-grained carbonate-cemented sandstone (Fig. 4.9). Mudstone clasts range from 3.5 cm x 2.5 cm to 28 cm x 3 cm, are located mainly at the base of D2, and are commonly oriented with their long axis parallel to the base of the bed. Carbonate clasts are abundant, range in size from 2 cm x 0.4 cm to 115 cm x 9.5 cm and are dispersed throughout. Most carbonate clasts are structureless but oolite and stromatolite fragments are observed. D2 is best developed in the southeast drainage study area, where it directly overlies D1 and crops out over a lateral distance of ~35 m, and ranges from 0.5 m to 6 m in thickness. In the southern outcrop, D2 crops out over a lateral distance of 15 m, is 42 cm thick and directly overlies D1. In the northern outcrop, D2 forms a laterally discontinuous layer that has foundered into D1 forming isolated rafts that are up to 3 m long and 1.2 m thick.

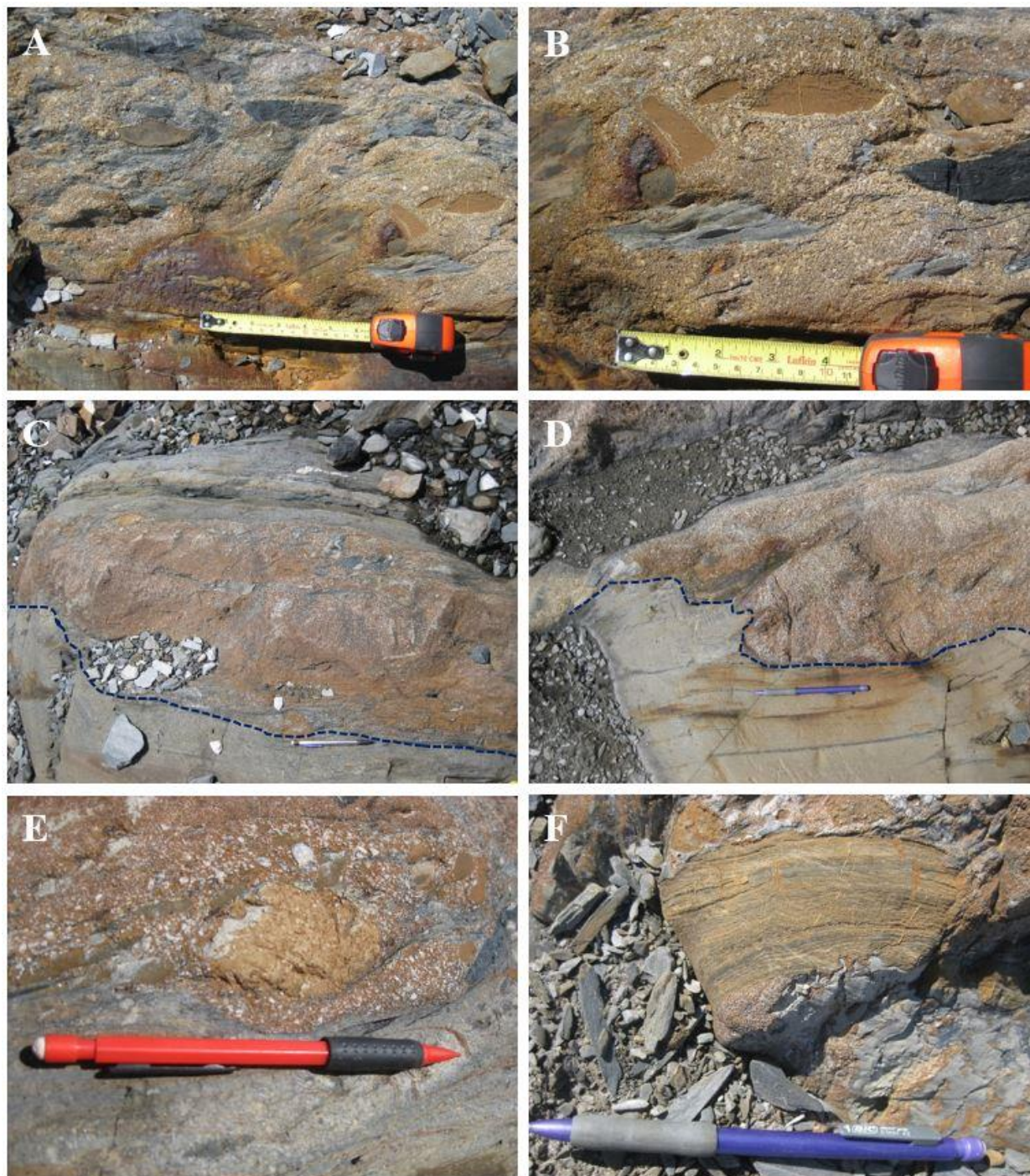


Figure 4.9: Photos of D2. (A) Photo of D2 in the southern outcrop where it overlies D1. Photo shows dispersed carbonate and mudstone clasts along with lower very coarse and upper very coarse grains dispersed in a matrix of upper coarse-grained carbonate-cemented sandstone that make up a poorly-sorted D2. (B) Close-up of photo A. (C, D) In the northern outcrop, D2 forms a laterally discontinuous layer that has foundered into D1 forming isolated rafts that are up to 3 m long and 1.2 m thick. Contact with rafts of D2 and underlying D1 is outlined in blue dashed line. (E) Oolite fragment in a raft of D2. (F) Stromatolite fragment in a raft of D2.

In the southeast drainage study area, D3 extends over a lateral distance of ~15 m, where it thickens towards the northwest reaching a maximum thickness of 11 m. In the southern study area, D3 thins and thickens laterally, ranging from 1.38 m thick to 9.01 m thick. D3 consists of very coarse and upper very coarse sandstone grains, mudstone, carbonate-cemented sandstone and uncommon structureless carbonate clasts dispersed in a silty mudstone matrix (Fig. 4.10). It is poorly sorted and the basal contact is sharp and planar. Carbonate-cemented sandstone clasts are the most abundant clast type, range in size from 9 cm x 3 cm to 9.7 m x 66 cm and are commonly isoclinally folded. Mudstone clasts are not uncommon and range in size from 4 cm x 0.5 cm to 77 cm x 10 cm. Notably, at the southernmost end of the outcrop, a 1.8 m thick raft composed of lower coarse-grained sandstone extends for 10 m laterally and crops out at the base of D3. Similarly, but at the northernmost end of the outcrop, a 2.9 m thick x 26 m long raft of lower coarse-grained sandstone occurs at the base of D3. Both clasts contain abundant mudstone intraclasts.

In the southern outcrop D4 directly overlies S1 (F4b, discussed later in this chapter) and thins from 18 m to 8.2 m towards the southeast. In the southeast drainage, it is 7.6 m thick on average. D4 has a sharp and planar basal contact, is poorly sorted and consists of upper very coarse sand grains, pebble-sized quartz, carbonate-cemented sandstone, mudstone and carbonate clasts dispersed in a silty mudstone matrix (Fig. 4.11). Carbonate-cemented clasts consist of lower medium-grained sandstone fragments that commonly are isoclinally folded and deformed. They range in size from 9 cm x 2 cm to 3.2 m x 62 cm. Both mudstone clasts (~7 cm x 2 cm) and carbonate clasts (~7 cm x 3 cm) are less common than carbonate-cemented sandstone clasts.

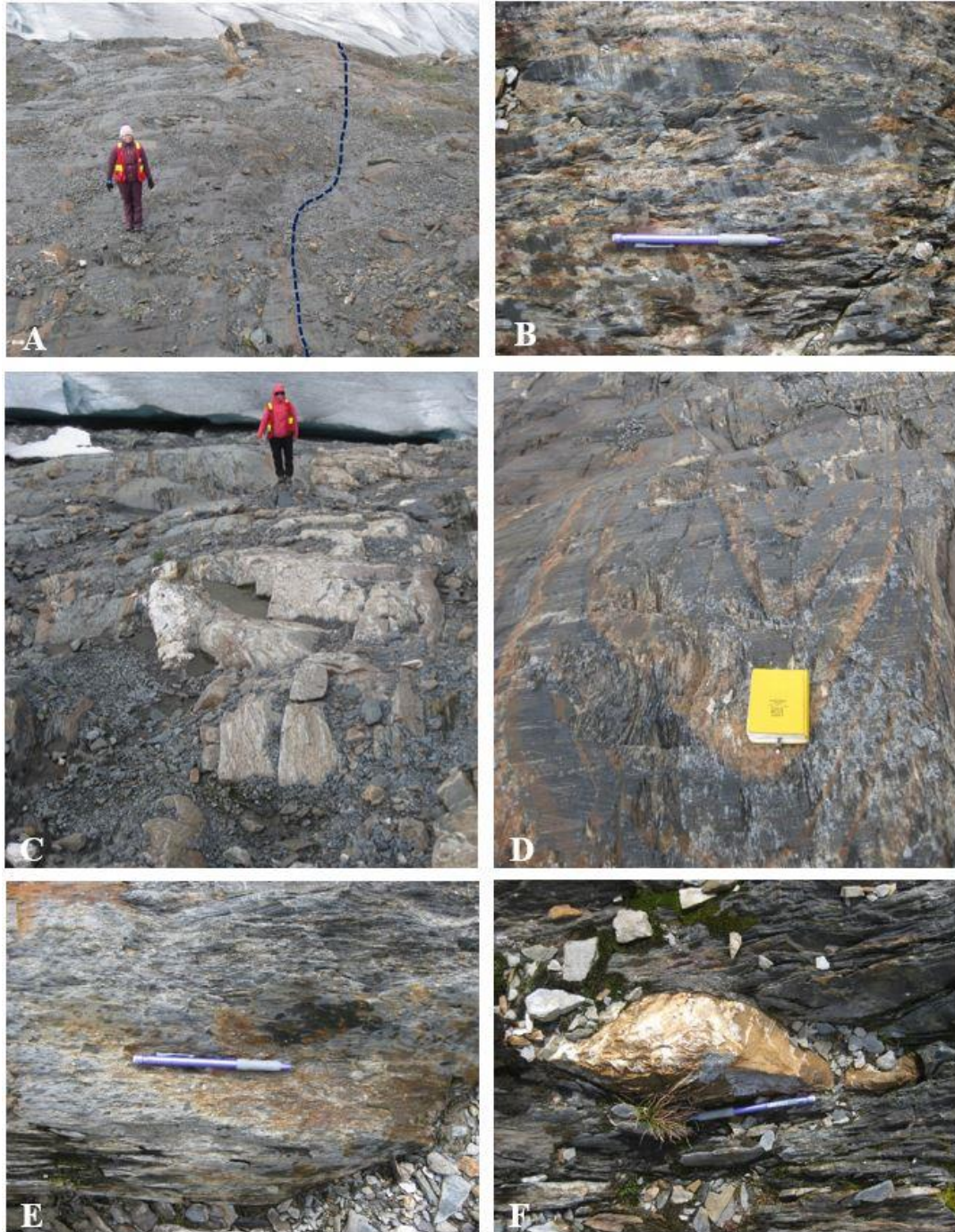


Figure 4.10: Photos of D3. (A) Sharp and planar basal contact of D3 is outlined in blue dashed line. It is overlain by deformed, thin-bedded, fine-grained upper division turbidites. (B) Carbonate-cemented sandstone clasts dispersed in a silty mudstone matrix. (C) At the northernmost end of the outcrop, this 2.9 m thick x 26 m long raft of lower coarse-grained sandstone crops out at the base of D3. (D) Isoclinally folded and deformed clast of thin-bedded, fine-grained upper division turbidites. (E) Very coarse and upper very coarse sandstone grains dispersed in the silty mudstone matrix. (F) Uncommon structureless carbonate clast.

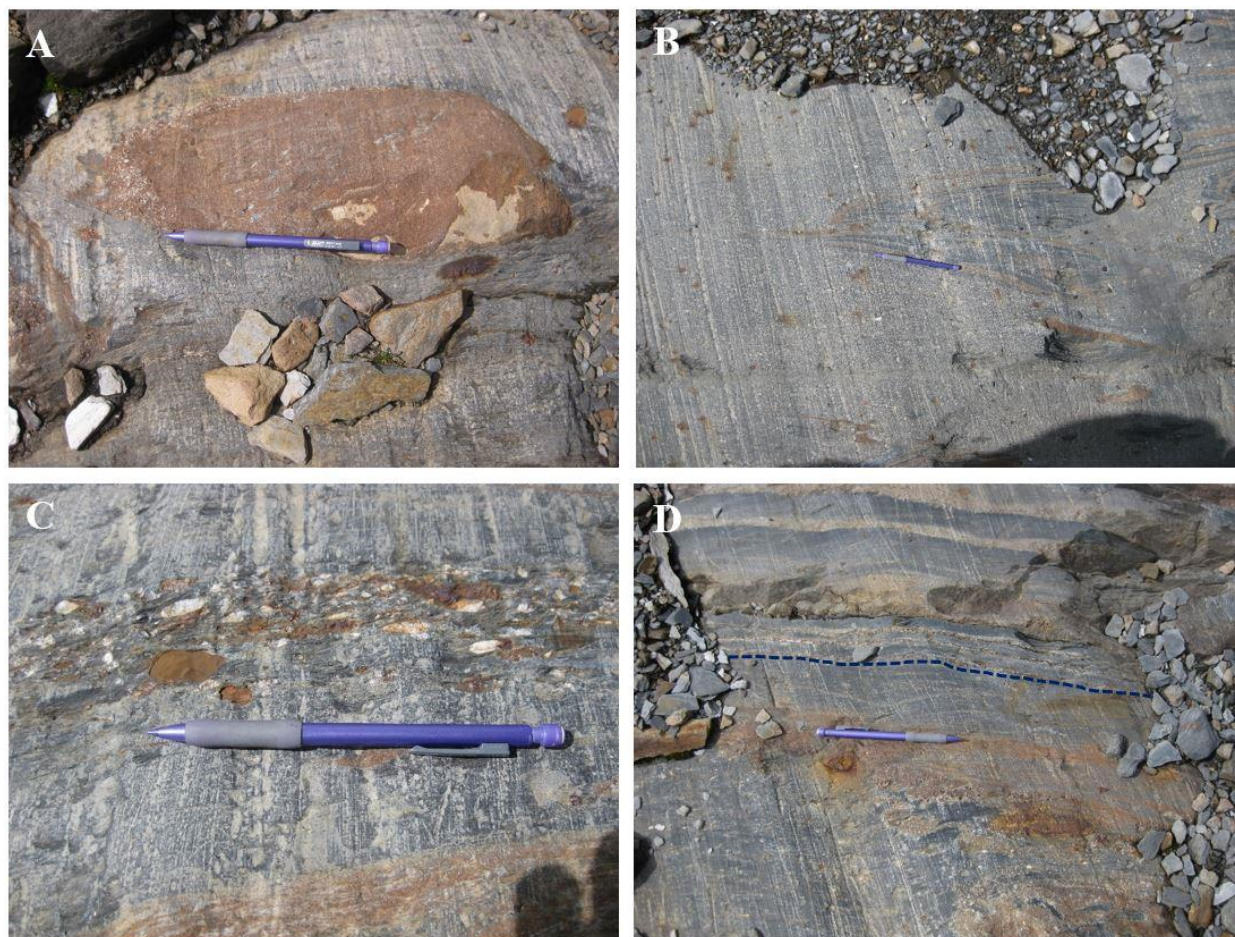


Figure 4.11: Photos of D4. (A) Carbonate-cemented sandstone clast. (B) Deformed, very thin-bedded turbidites in a silty mudstone matrix. (C) Pebble-size quartz and structureless carbonate clasts. (D) Upper contact of D4 is outlined by blue dashed line. It is overlain by thin-bedded, fine-grained turbidites.

4.4.2 Interpretation

Deposits D1- D4 consist of poorly-sorted, ungraded, matrix-rich units containing clasts of various size, composition and degree of deformation. The lack of bedding and presence of clasts within a poorly-sorted matrix suggests that these deposits were deposited by cohesive sediment-gravity flows termed debris flows (Mulder and Alexander, 2001). Large metre-scale clasts were supported mainly by matrix-strength provided by the ubiquitous fine-grained sediment that originally consisted mostly of clay minerals. The electrostatic attraction and atomic bonding of adjacent clay particles provides cohesive strength to the moving mass and, because of its low

permeability, helped reduce the influx of ambient fluid and thereby fluid dilution. Sediment support was also provided by a combination of elevated local pore pressure, buoyancy and clast-to-clast interaction (Mulder and Alexander, 2001). Debris flow deposition occurs mainly *en masse* when the forces resisting shear, mainly viscosity and friction, equal or exceed the gravitational force – thickness of the deposit being similar to the thickness of the parent flow (neglecting post-depositional compaction) (Iverson, 1997, Mulder and Alexander, 2001). *En masse* deposition, in addition to limited internal grain mobility, explains the poorly sorted and chaotic texture of these deposits, which most probably resembles the grain size distribution and sorting within the parent flow (Mulder and Alexander, 2001). Mudstone clasts are commonly isoclinally folded, which likely reflects shearing within the flow or shearing and/or compaction after *en masse* deposition (Mulder and Alexander, 2001).

4.4.3 F4b: Deformed Bedded Deposits

Facies 4b (F4b) crops out at the base of the northern outcrop where it is approximately 23 m thick. A second F4b deposit crops out near the top of the southern outcrop but cannot be correlated across the glacier to the northern outcrop. There it overlies D3, is approximately 35 m thick, and extends laterally across the moraine and into the southeast drainage outcrop where it again overlies D3 and is approximately 30 m thick.

In the northern outcrop, undeformed, flat-lying, lower-division turbidites are truncated by a steep and irregular surface towards the northwest (Fig. 4.12 A). Strata that crop out on the other side of this surface consist of deformed, upper division turbidites (Fig. 4.12) that, laterally away from this surface (towards the northwest), become progressively less deformed (Fig. 4.13). Strike of bedding within the unit diverge by about 5° from regional strike measurements (330°) (Marion, 2005).

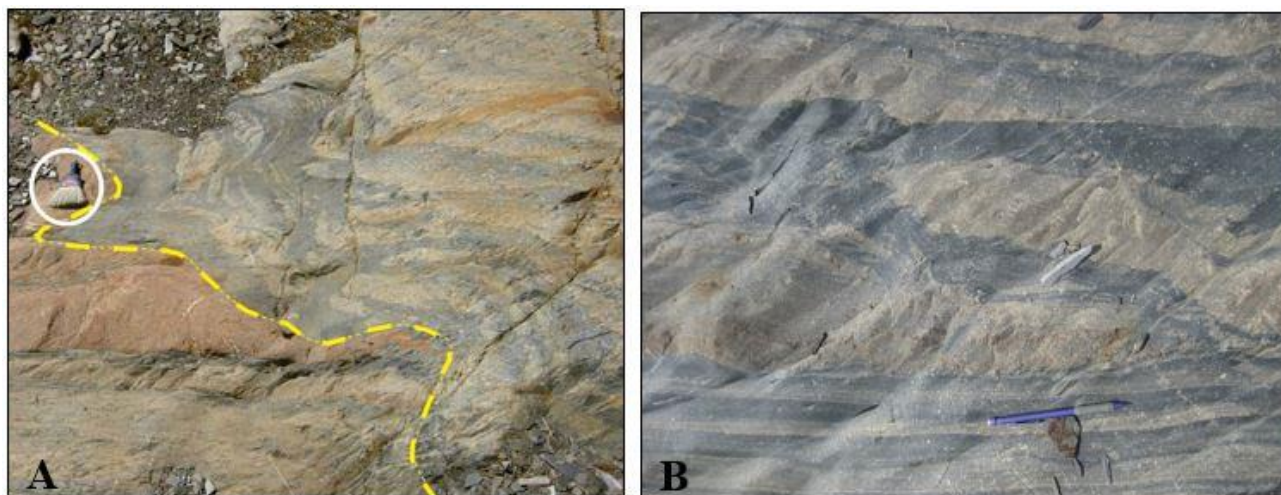


Figure 4.12: (A) Steep, sharp and irregular surface (yellow dashed line) that separates thick-bedded, lower division turbidites to the left from thin-bedded, upper division turbidites to the right. Note how strata to the left are truncated whereas strata to the right are deformed. Photo from Marion (2005). (B) Close up of deformed upper division turbidites ~5 m to the northwest (right) of the surface.

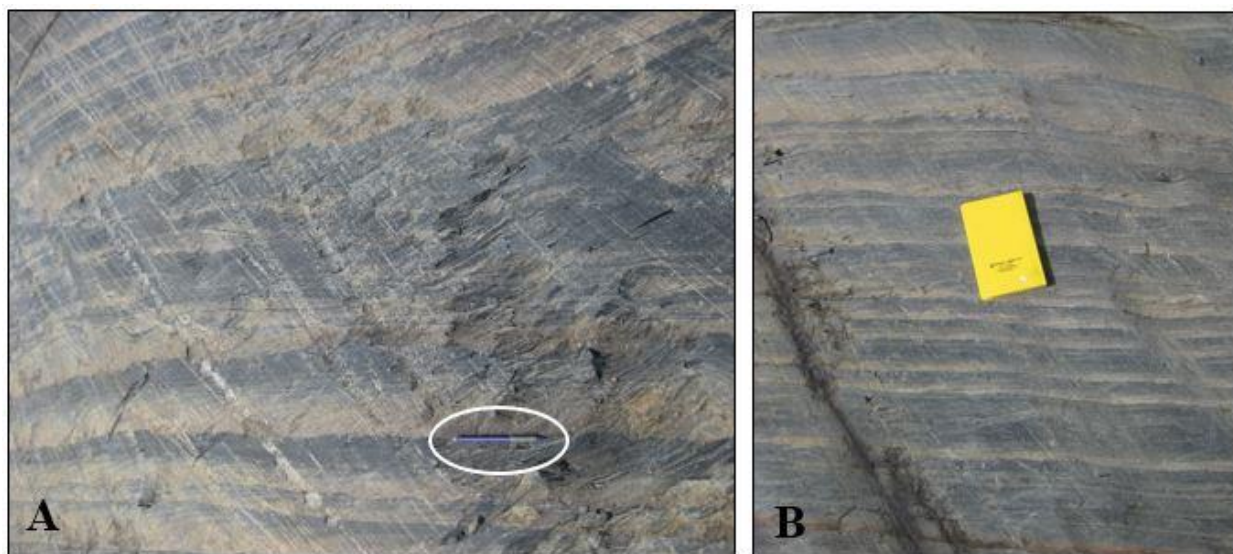


Figure 4.13: (A) Deformed upper division turbidites ~10 m to the northwest (right) of the steep, sharp and irregular surface. (B) At a lateral distance of ~25 m towards the northwest, strata only exhibit subtle soft sediment deformation structures.

In the southern outcrop, F4b is made up of stacked units that are each a few metres thick. Each unit has a sharp and planar basal surface (Fig. 4.14 A) and is composed of two parts. The lower part is a few decimetres thick and poorly sorted, consisting of coarse sand grains, quartz pebbles and mudstone clasts (typically on the order of 1.5 cm x 1 cm) dispersed in a silty mudstone

matrix (Fig. 4.14 A-B). The upper part gradationally overlies the lower part, is a few metres thick and is made up of deformed upper division turbidites that become progressively less deformed upward (Fig. 4.14 E-F), with the exception of strata at the very top, which are deformed. Dispersed carbonate-cemented sandstone clasts (typically on the order of 50 cm x 20 cm) crop out in mud-rich, thin-bedded Tcd/e intervals in the upper part (Fig. 4.14 C). The strike of these deformed strata (320°) differs slightly from the regional strike (330°) by about 10° on average (Fig. 4.14 A). Strata of F4b can be correlated laterally across the moraine and into the southeast drainage outcrop where deformed upper division turbidites are observed, but due to poor outcrop exposure discrete stacked units were not observed.

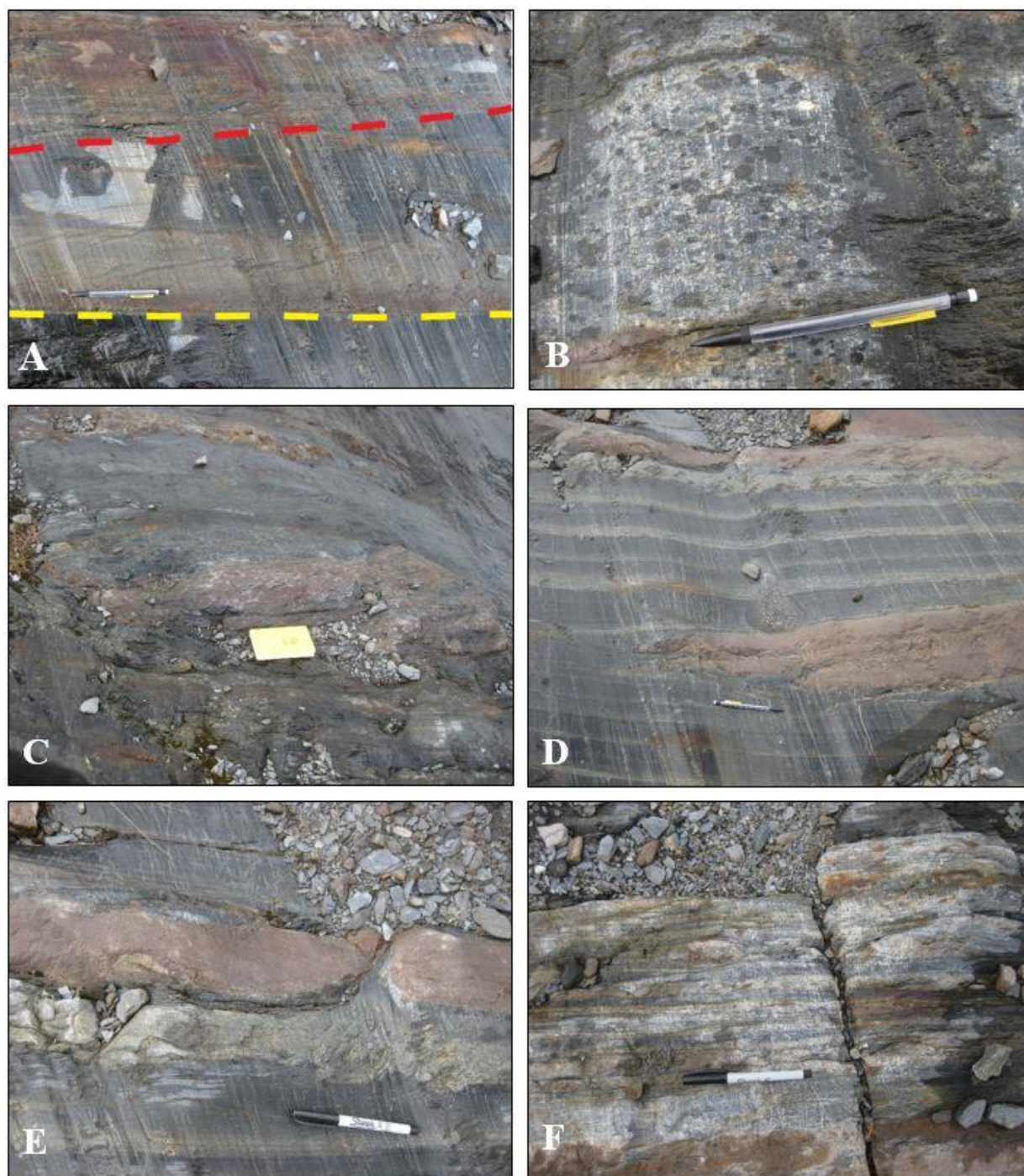


Figure 4.14: (A) Sharp and planar basal surface of F4b in the southern outcrop highlighted with red dashed line. A ~20 cm thick poorly-sorted layer with dispersed, abundant quartz pebbles surrounded by dark black silt directly overlies the base. Note the difference in strike between basal contact (red) and underlying strata (yellow). (B) Close up of poorly-sorted lower layer in A). (C) Carbonate-cemented sandstone clast in the upper part. (D) Planar-laminated sandstone bed that ends abruptly in the upper part. Note the undeformed appearance of the overlying strata. (E) Deformed sandstone bed in the upper part. (F) Slightly deformed sandstone bed located farther away from the basal surface within the upper part.

4.4.4 Interpretation

The steep and irregular surface that truncates undeformed, flat-lying, F4b lower division turbidites in the northern outcrop (Fig. 4.12 A) is interpreted to be a slide scar that formed when part of a levee succession became detached and moved some distance downslope and beyond the plane of the outcrop (Marion, 2005; Arnott, 2012). Later, the slide scar and associated seabed depression became plugged with a semi-coherent block of upper division turbidites that crop out on the northwest side of the scar. Failure of the levee deposits could have been triggered by a variety of factors including high sedimentation rates, sudden loading, earthquakes or sea level changes (Navarro, 2016); but most probably was simply the result of failure along the typically steep inner margin of a sinuous subaqueous channel – a feature commonly observed in both modern (Fig. 4.15) (e.g., Gamberi and Rovere, 2011) and ancient Upper Cretaceous (Fig. 4.16) (e.g., Kane et al., 2007) deep-marine turbidite systems. In this study a channel fill directly underlies the slide deposit in the northern outcrop (Fig. 4.17) (Marion, 2005). Therefore, it is likely that a portion of the levee facing into the channel failed and formed an irregular slide scar. This depression was then filled, possibly during the same failure event, with a large block of thin-bedded, fine-grained, upper division turbidites. The absence of coarse-grained turbidite interbeds suggests that the block represents a remobilized fragment of distal levee deposits (see section 5.2.7) unrelated to the levee truncated by the slide, but instead sourced from levees located further upslope.

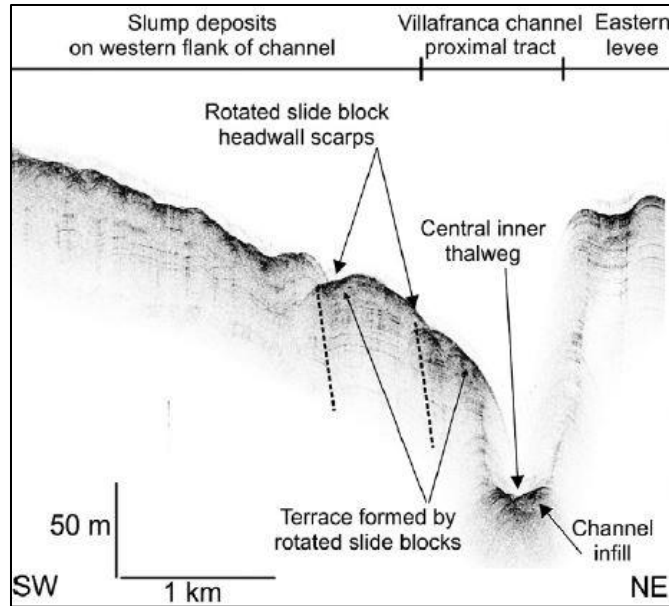


Figure 4.15: CHIRP profile across a straight segment of the modern Villafranca channel, southeastern Tyrrhenian Sea (from Gamberi and Rovere, 2011). Failure along the western steep inner margin of the channel triggered mass transport deposits, which on this seismic image are shown as thick, transparent layers with a rough surface. Rotated slide blocks along the western margin are bounded by extensional faults (black dashed lines).



Figure 4.16: Slump deposit in a channel-proximal location within the Upper Cretaceous Rosario Formation, Baja California, Mexico (modified from Kane et al., 2007). Some convoluted bedding contacts have been outlined in yellow. Here, deformation of channel-proximal strata includes local folding but is dominated by slide blocks up to 100 m in thickness.

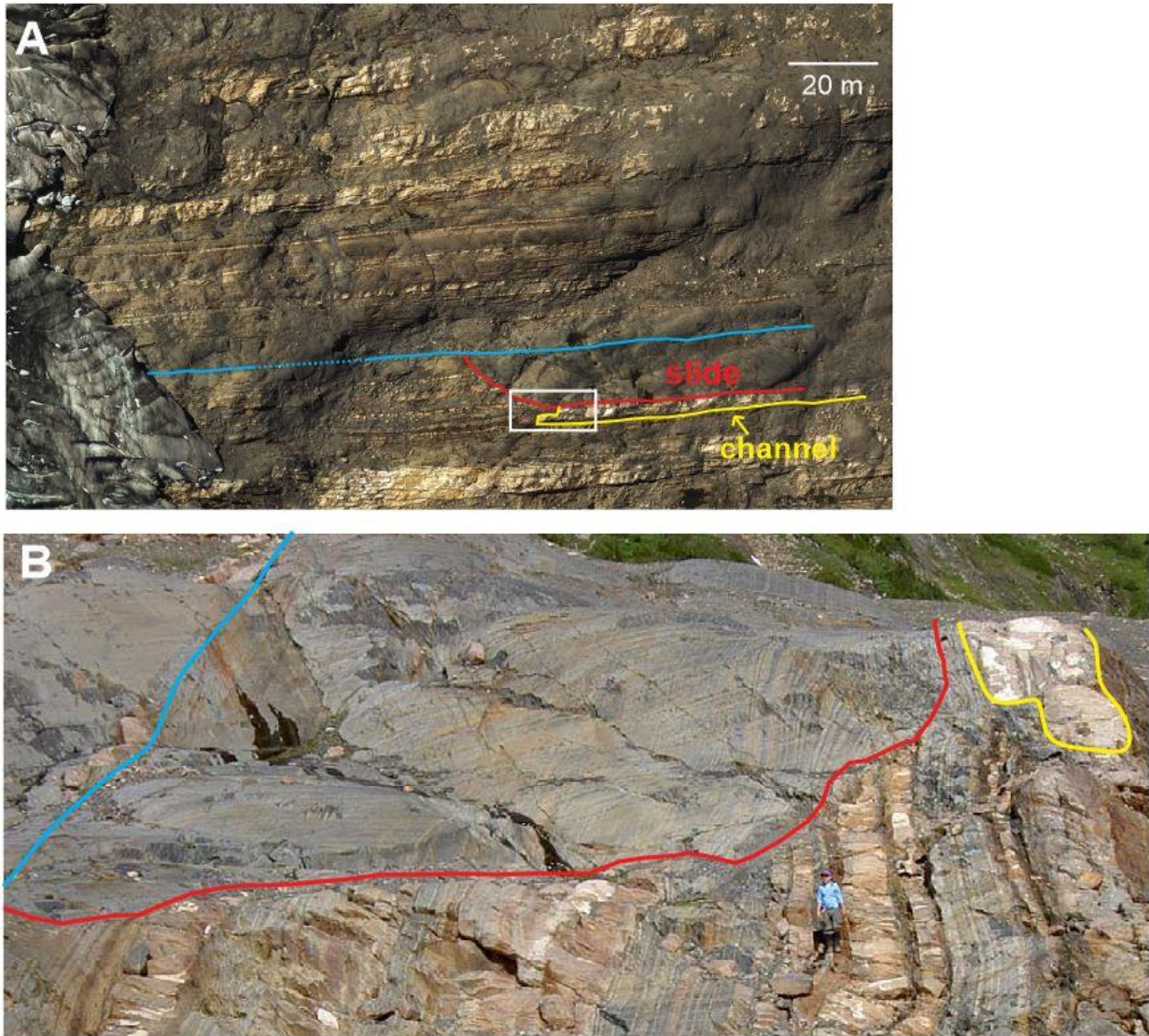


Figure 4.17: (A) Aerial photograph showing the slide deposit (contact in red) that crops out in the northern outcrop of this study underlain by the channel fill deposit (contact in yellow) identified by Marion (2005). Blue contact marks the base of debris 1 (D1). (B) Photograph taken by Marion (2005) in the field of the area marked by the white rectangle in (A). Slide contact outlined in red, channel in yellow and debris D1 in blue.

The slide deposit that crops out in the southeast drainage and southern outcrops most probably formed by gravity-driven downslope failure on a levee. Seismic profiles of modern channel-levee systems show that levee backslopes can be as steep as 5° (Pirmez and Imran, 2003). Consequently, levee deposits are intrinsically gravitationally unstable (Kane et al., 2007) and failure resulting in slides is common. Here, the slide deposit consists of thin-bedded, fine-grained,

upper division turbidites intercalated with thick-bedded, coarse-grained, lower division turbidites. These deposits resemble the lithological makeup in the distal levee (see section 5.2.7), thereby suggesting that this slide deposit is composed of remobilized distal levee deposits. Friction, and related shear deformation, is greatest along the basal slip plane of a mass movement deposit (Khan, 2012; Arnott, 2012). In addition, weaknesses within the sliding mass, for example a sand-rich layer within a pile of compacted mud, form internal surface slide surfaces. The lower, poorly-sorted layer with abundant dispersed quartz pebbles surrounded by dark black silt at the base of each slide block (unit) in the southern outcrop is interpreted to have formed by shearing along interspersed weak layers within a larger-scale sliding mass, resulting in an assemblage of stacked, discrete coherent blocks. Away from the shear surfaces strata become less deformed. However, strata at the top of each slide block are slightly deformed by shearing caused by the movement of the overlying slide block. Consistent strike orientations within each block, but variation between blocks, suggests that the blocks were independent parts of a single slide.

4.5 Petrography

Strata of F1, F2 and F3 are mineralogically similar (Fig. 4.18). Strata are moderately well-sorted and based on the Folk classification scheme (1974) are classified in terms of framework mineralogy to be subarkose. Subangular to subrounded framework grains consist predominantly of quartz (75% of the total stratal volume). However, in some F2 and F3 beds quartz content can be as low as 25% of the stratal volume because in these strata quartz grains are surrounded by a groundmass of recrystallized calcite. Anhedral quartz grains are commonly monocrystalline with undulose extinction. Plagioclase makes up a minor portion of the framework mineralogy (~3% of the stratal volume). Recrystallized quartz occurs as smaller (< 0.087 mm) monocrystalline grains along the boundaries of larger (~0.33 mm) framework quartz grains. These grains show no

undulose extinction. The matrix comprises ~10% of the total stratal volume, and consists of metamorphic muscovite and chlorite, with additional but minor (up to 2% of the stratal volume) euhedral to subhedral pyrite porphyroblasts. Carbonate cement, made up of both calcite and dolomite, makes up 10% to 60% of the stratal volume in F1, F2 and F3 beds.

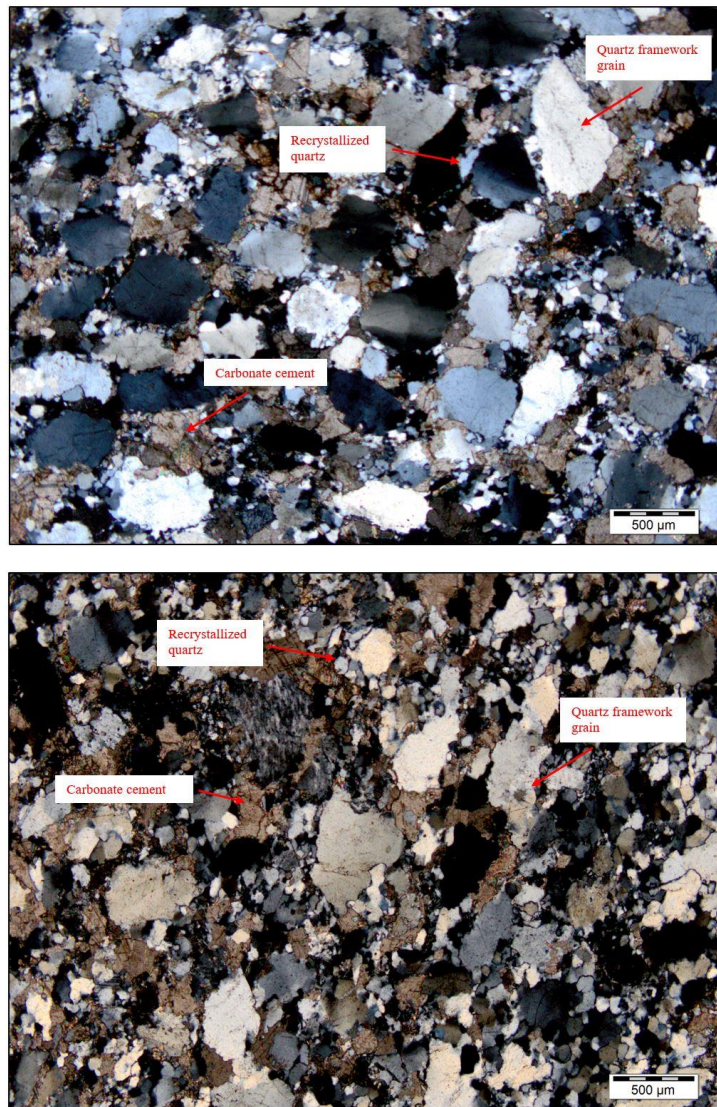


Figure 4.18: Photomicrographs of facies 2 (top) and facies 3 (bottom) strata under cross-polarized light. Quartz consists of the main framework grain. Note how the recrystallized quartz occurs as smaller monocrystalline grains along the boundaries of larger framework quartz grains. Carbonate cement in these strata is made up of both calcite and dolomite.

In some F2 and F3 beds, wavy graphite laminae that are black and opaque in thin section can make up to 50 % of the stratal volume of the bed (Fig. 4.19). In these samples, quartz framework grains are isolated within the laminations and make up ~20% of the total stratal volume. The matrix of metamorphic muscovite and chlorite makes up ~10% of the total stratal volume; carbonate cement and pyrite make up ~17% and ~3% of the total stratal volume, respectively. Graphite laminations are discussed in section 4.7.

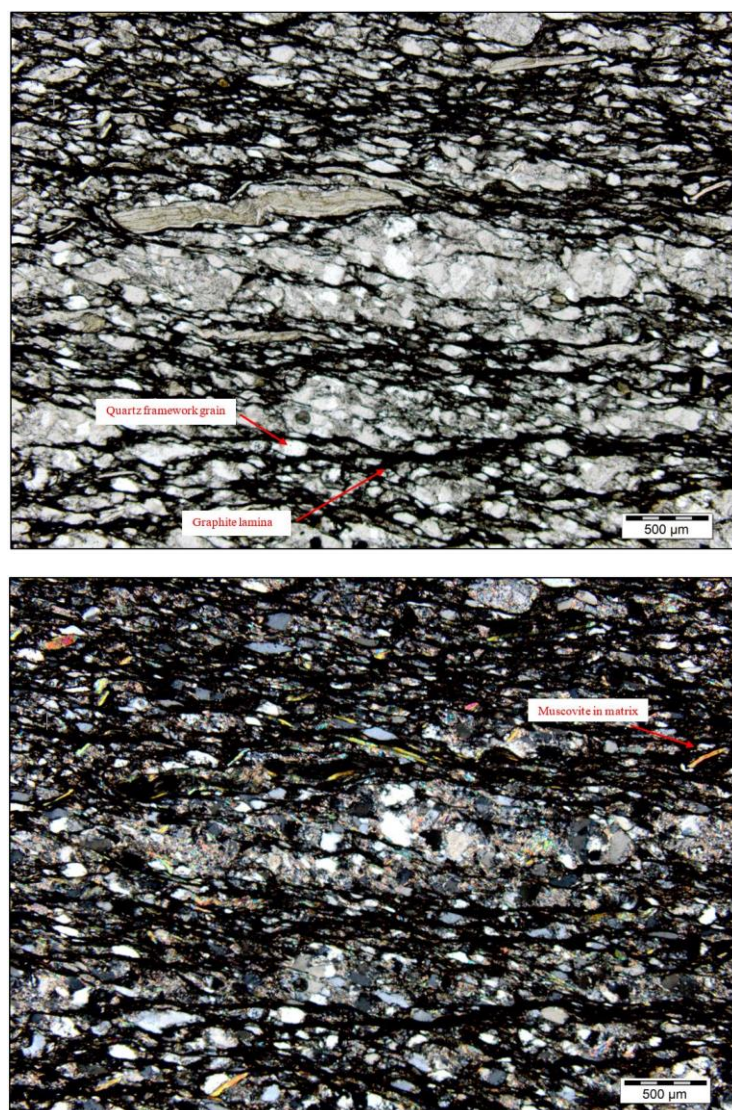


Figure 4.19: Photomicrographs of strata showing black and opaque wavy graphite laminae under plane-polarized (top) and cross-polarized light (bottom). Note how quartz framework grains are dispersed in the graphite. Matrix consists of metamorphic muscovite and chlorite.

The majority of the quartz framework grains show undulose extinction, which is a manifestation of elevated crystal lattice energy (Winter, 2010). The ubiquity of strained quartz grains indicates that temperatures remained too low (i.e., insufficient energy) for grains to recover and form new strain-free subgrains. In turn, deformation drives recovery and recrystallization. The sutured grain boundaries observed are a result of recovery, where high-strained material at grain boundaries is incorporated onto larger adjacent crystals by grain boundary migration (Winter, 2010). Recrystallized quartz grains are the result of bulging recrystallization. Bulging recrystallization reduces the strain on quartz crystal lattices when the rock is stressed or under elevated temperatures. This recrystallization process occurs at lower greenschist facies conditions (~280 and ~400°C) and affects only the outer boundaries of strained quartz grains (Stipp et al., 2002). During this process, grain boundaries migrate, or “bulge” into more highly strained areas. Eventually the bulges separate from the parent quartz grain resulting in many smaller quartz grains that come to envelop the parent quartz grain (Stipp et al., 2002). This is observed in F1, F2 and F3 beds where smaller, straight extinction quartz grains occur along the boundaries of larger framework quartz grains and are an indicator of strain relaxation.

Strata of F4a are typically distinctively poorly sorted and composed primarily of matrix that makes up to 80% of the stratal volume (Fig. 4.20). The matrix is composed of fine-grained metamorphic muscovite and chlorite, in addition to quartz grains that are silt-size or finer (<0.0625 mm). The chlorite and muscovite are well aligned and form the foliation. Rounded monocrystalline and polycrystalline quartz grains with undulose extinction account for 20% of the stratal volume. Quartz grains have been elongated parallel the foliation, which wraps around the individual grains. Additionally, quartz grains have overgrowths of quartz and quartz-phyllosilicate fringes (“beards”)

of fibrous grains that extend outward on opposite ends of the grain and parallel the foliation (e.g., Vernon, 2004). Euhedral to subhedral pyrite porphyroblasts can make up about 2% of F4 units.

Unlike matrix-poor, structureless, normally-graded sandstone (F1), structure sandstone (F2) and medium-scale cross-stratified sandstone (F3) that, when tectonized, relieved stress along framework grain contacts and formed recrystallized grain boundaries, the matrix-rich, poorly-sorted mudstone-rich deposits (F4a) have few grain-grain contacts. As a result, tectonic stress was accommodated principally by slip within (i.e., basal slip) detrital clay minerals that make up the matrix.

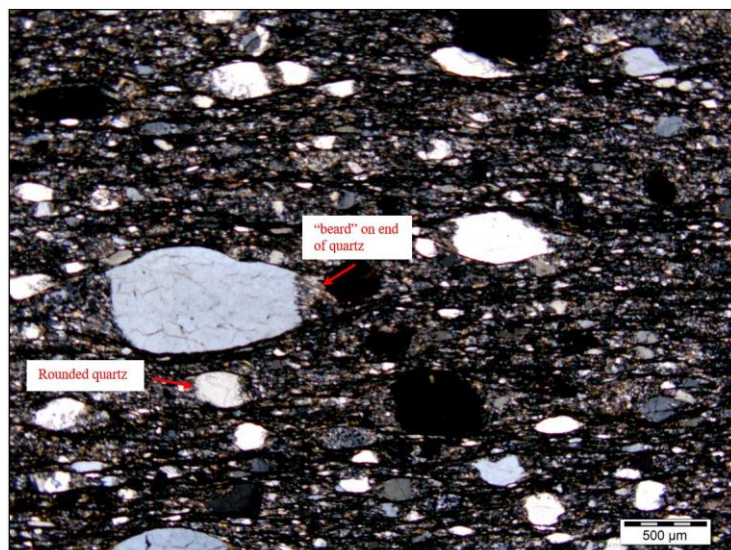


Figure 4.20: Photomicrograph of facies 4a under cross-polarized light. This facies is typically poorly sorted and composed primarily of matrix. Rounded quartz grains commonly have overgrowths of quartz and quartz-phylosilicate fringes (“beards”) of fibrous grains that extend outward on opposite ends of the grain and parallel the foliation.

4.6 Carbonate Cements

Carbonate cement, composed mostly of calcite with minor dolomite, makes up 10% to 60% of the stratal volume in structureless, normally graded sandstone (F1), structured sandstone (F2) and medium-scale cross-stratified sandstone (F3) beds (Fig. 4.21). Calcite occurs as pore-filling anhedral crystals with characteristic high birefringence and rhombic cleavage in thin section.

Dolomite cement is much less common (trace amounts) and in thin section occurs as high relief, dark brown, single rhombic crystals. Previous microscopic work on carbonate-rich strata at Castle Creek stained thin sections with Alizarin Red S and potassium ferrocyanide. It was determined that most of the calcite cement was iron-rich (Al-Mufti, 2013; Navarro, 2005), which upon oxidation imparts a distinctive rusty-orange colour to these strata.

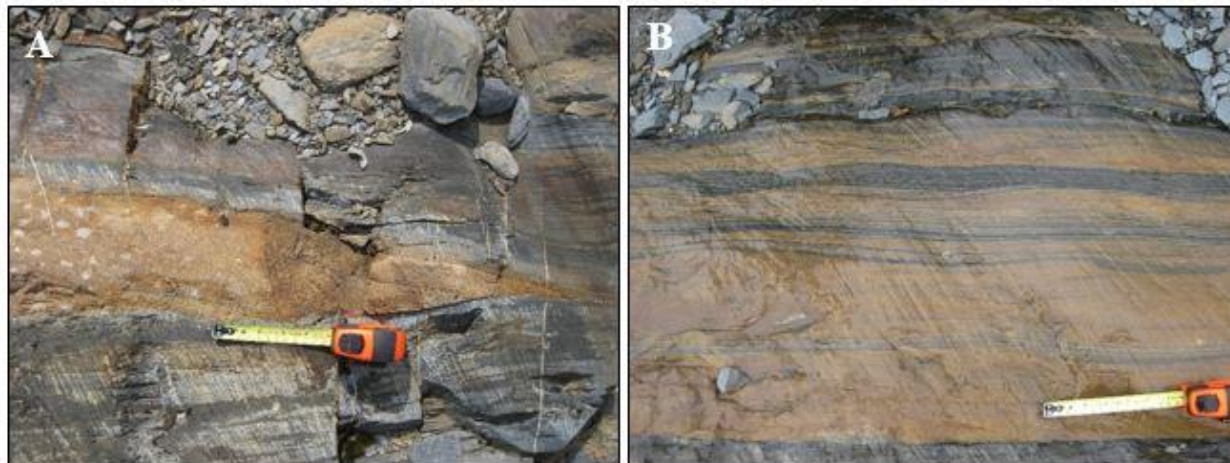


Figure 4.21: Carbonate-cemented strata. (A) Carbonate-cemented medium-scale cross-stratified sandstone (F3). (B) Carbonate-cemented Tbcde turbidite. Note the distinctive rusty-orange colour of these strata.

Strata with high carbonate cement content commonly consist of planar-laminated (F2), ripple cross-stratified (F2) or dune cross-stratified (F3) sandstone. High intergranular porosity suggests that pore-filling calcite cements formed soon after deposition (Al-Mufti, 2013; Meyer, 2004). Moreover, the occurrence of deformed planar-laminated and dune-cross stratified carbonate-cemented sandstone clasts within mass movement deposits (F4a and F4b; discussed in section 3.6) further supports early cementation (Fig. 4.22). It is likely that after deposition, these strata became partly cemented and soon after eroded by a mass movement flow.

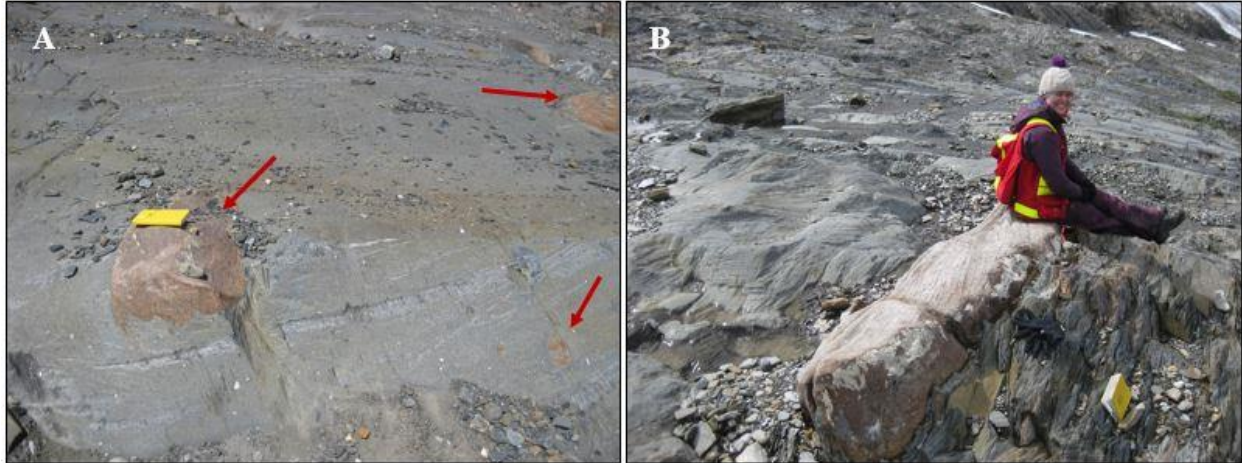


Figure 4.22: (A) Red arrows point to clasts of carbonate-cemented convoluted planar-laminated sandstone (F2) within a debrite (F4a). (B) Metre-scale carbonate-cemented dune cross-stratified sandstone clast (F3) within a debrite (F4a).

4.7 Total Organic Carbon Content

Sand-rich Tbcde turbidites commonly exhibit elevated total organic carbon (TOC) values compared to other strata in the study area (e.g., Ta turbidites) (Fig. 4.23). They typically weather a distinctive rusty-orange colour and contain dark-grey laminations composed mostly of graphite. They range from 70 cm to 140 cm thick and are composed of lower medium- grading upward to fine-grained sandstone.

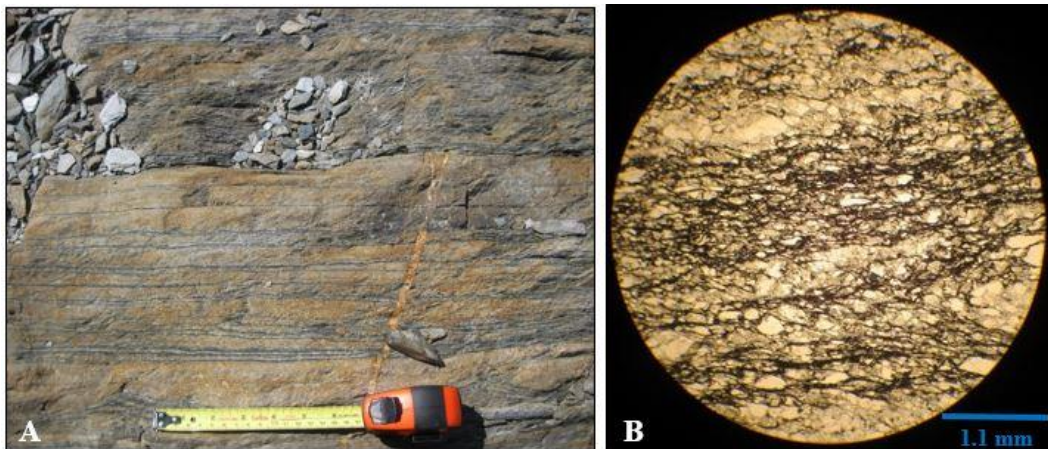


Figure 4.23: (A) Carbonate-cemented Tbcde turbidite with elevated TOC. Dark-grey laminations are composed of graphite. (B) Photomicrograph under plane polarized light. Quartz framework grains are isolated within wavy black and opaque graphite laminations that are oriented parallel to bedding.

Natural graphite can be formed in two ways: precipitation from C-O-H rich fluids (epigenetic graphite), or by the in situ metamorphism of organic matter (syngenetic graphite) through graphitization (Crespo et al., 2004). Epigenetic graphite occurs as veins (Luque et al., 1998) or as overgrowths on existing graphite grains (Ueno et al., 2002) whereas syngenetic graphite occurs as stratabound disseminated or massive concentrations (Crespo et al., 2004). The graphite in strata of this study is stratabound (syngenetic), and therefore is interpreted to have formed during Mesozoic regional metamorphism of detrital carbonaceous matter (Crespo et al., 2004). Metamorphism has completely transformed the organic material making its original size, shape and composition indeterminable. Nevertheless, organic matter was concentrated into discrete, dark-grey laminae in planar laminated (F2) and dune cross-stratified (F3), medium-grained sandstones, suggesting that the organic particles were hydraulically equivalent to medium quartz sand grains. The organic matter was most probably sourced from disaggregated bacterial/algal mats that grew on shallower parts of the slope or shelf before being resedimented downslope by turbidity currents where it was probably shape sorted into discrete laminae during bed-load transport.

Background TOC values of strata in the Isaac formation are generally 0.1% (Davis, 2011). Geochemical analyses of samples from the study area show that TOC ranges from 0.07- 0.87% (+/-0.1%) (Table 1). Note that these strata underwent low-grade metamorphism and therefore have lost (i.e., pyrolysis) about 50-75% of the original organic material (Smith, 2009; Hayes et al., 1983), indicating that depositional values ranged between 0.28- 3.48%. It is important to note that although some of the samples contain elevated TOC (0.87-0.21%), not all strata with graphite layers show such elevated contents (0.15-0.07 %).

Sample ID	Bed Thickness (cm)	Grain Size	Description	TOC (%)	$\delta^{13}\text{C}_{\text{org}}$ (‰)
S23	80	Fss	Top of planar laminated sandstone near mudstone cap	0.87	-24.32
S22	90	Silt	Mudstone cap	0.33	-23.34
S24	53	Lmss→Fss	Planar laminated sandstone	0.31	-22.70
S13	137	Lmss→Fss	Planar laminated sandstone	0.21	-21.82
S2	137	Lmss→Fss	Planar laminated sandstone	0.15	-22.63
S12	100	Umss	Planar laminated sandstone	0.11	-25.80
S11	117	Umss	Planar laminated sandstone	0.10	-24.34
IS4	137	Lmss→Fss	Planar laminated sandstone	0.09	-25.17
S16	85	Umss→Lmss	Planar laminated sandstone	0.07	-26.15

Table 1: Total organic carbon (TOC) and stable carbon ($\delta^{13}\text{C}$) isotope data. The thickness, grain size and structure of each bed sampled are also listed.

4.8 Stable Carbon (^{13}C) Isotopes of Organic Material

Stable carbon ($^{13}\text{C}_{\text{org}}$) isotopes are calculated using the following formula (Arthur et al., 1983; Meyer, 2004):

$$\delta^{13}\text{C}_x = \frac{(R_x - R_{\text{std}})}{(R_{\text{std}})} \times 1000 \text{ ‰} \quad [1]$$

where R_x is the $^{13}\text{C}/^{12}\text{C}$ ratio of the sample and R_{std} is the $^{13}\text{C}/^{12}\text{C}$ ratio of the standard. Typically, the $\delta^{13}\text{C}$ of organic matter reflects the $\delta^{13}\text{C}$ of the source of organic carbon, with diagenetic processes only changing the $\delta^{13}\text{C}_{\text{org}}$ by minor amounts (Arthur et al., 1983). Depleted $\delta^{13}\text{C}_{\text{org}}$ values, indicating an enrichment in ^{12}C relative to ^{13}C , are the result of isotopic fractionation: the mass-dependent separation of isotopes during physical and chemical processes (Clark, 2015; Meyer, 2004). Bacteria, archaea and plants maximize energy generation in reduction-oxidation (redox) reactions by preferentially using the light isotope because its bonds are easier to break (for example, phototrophs preferentially use $^{12}\text{C}-\text{O}$ rather than $^{13}\text{C}-\text{O}$ during photosynthesis) (Clark, 2015; Arthur et al., 1983). It is for this reason that biologically mediated redox reactions have high fractionation factors resulting in ^{12}C enrichment, and accordingly, depleted $\delta^{13}\text{C}_{\text{org}}$ values in equation (1) (Clark, 2015; Meyer, 2004; Arthur et al., 1983). Importantly, the amount of organic carbon isotope fractionation depends on the biological metabolic pathway (Arthur et al., 1983). For example, terrestrial plants using the C_3 (Calvin-Benson) photosynthetic pathway result in significant $\delta^{13}\text{C}$ fractionation, and $\delta^{13}\text{C}_{\text{org}}$ signatures that range from -23 to -33 ‰. Alternatively, plants using the C_4 photosynthetic pathway are less discriminating and exhibit $\delta^{13}\text{C}_{\text{org}}$ values that range between -9 to -16 ‰ (Arthur et al., 1983).

The measured $\delta^{13}\text{C}_{\text{org}}$ values in this study are presented in Table 1 (+/-0.20‰) and range from -26.15 to -21.82 ‰. Since vascular plants did not exist during the Neoproterozoic, potential

sources for this range of $\delta^{13}\text{C}_{\text{org}}$ values include algae and marine phytoplankton (Arthur et al., 1983). The $\delta^{13}\text{C}_{\text{org}}$ signature of algae ranges from about -3 to -23 ‰ whereas marine phytoplankton impart an even stronger carbon isotope fractionation and $\delta^{13}\text{C}_{\text{org}}$ signatures ranging between -10 to -31 ‰ (Arthur et al., 1983). Due to low-grade metamorphism fine scale morphological structure that would help differentiate these organic components have been obliterated, and hence assessing the relative contribution of either organic component is not possible.

5 Channel and Levee Descriptions and Interpretations

In this study both channels and levees were identified based on distinctive assemblages of structureless, normally graded sandstone, structured sandstone and medium-scale cross-stratified sandstone (facies 1-3) (see Chapter 4). Channels are on the order of 8-25 m thick and consist mainly of structureless, normally graded sandstone. They crop out in the southeast drainage study area (Fig. 5.1, 5.2). Levees, on the other hand, consist principally of thin-, medium- and thick-bedded, traction-structured sandstone and can be divided vertically into packages. They crop out on the northwestern side of the southeast drainage study area and across the southern and northern outcrops (Fig. 5.1, 5.3, 5.4). Moreover, channel and adjacent levee deposits are interpreted to be genetically-related.

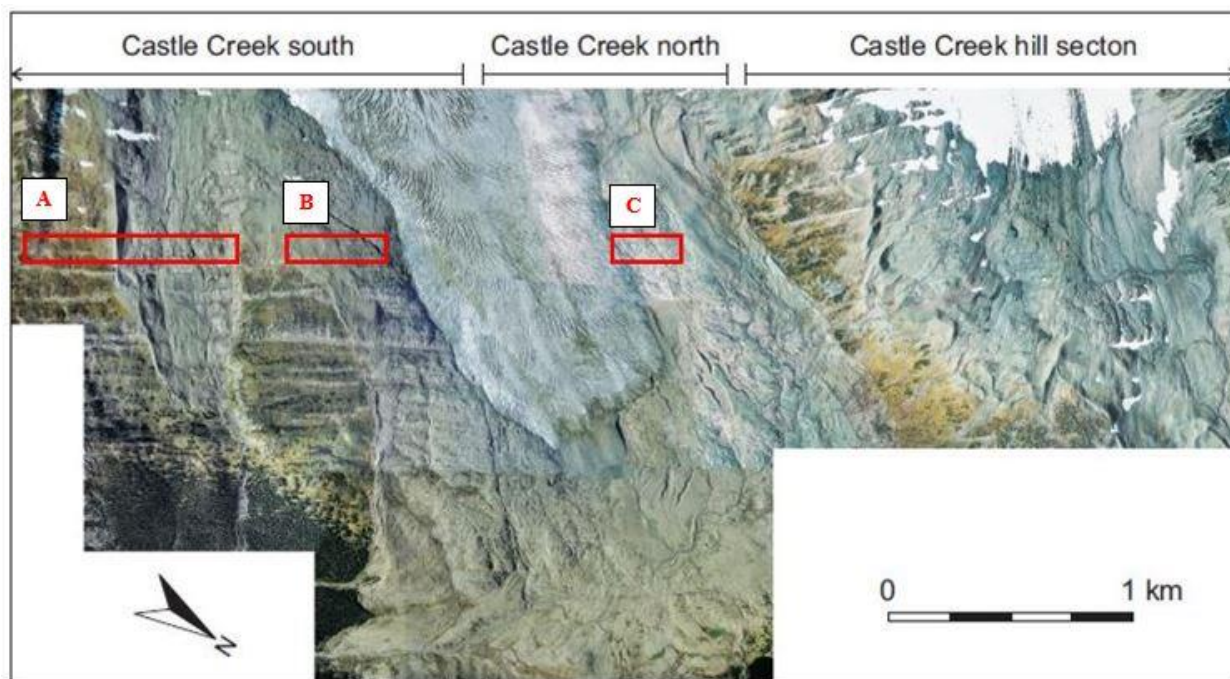


Figure 5.1: Channels and levees were identified in the study areas (red boxes). The channel complex crops out in the southeast drainage outcrop (A). Levee deposits crop out in the southern (B) and northern (C) outcrops.

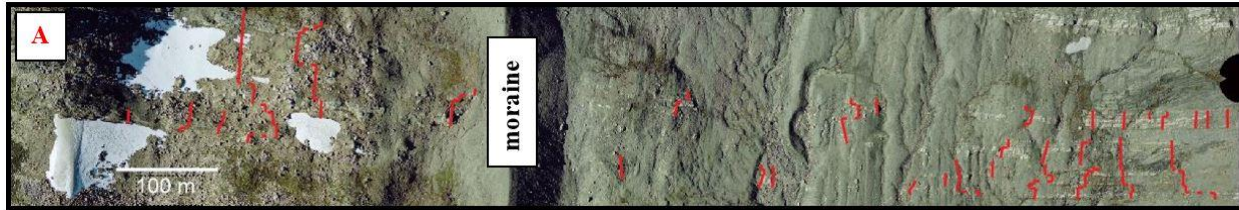


Figure 5.2: Aerial photograph of the southeast drainage study area (marked A in figure 5.1) where the channel complex crops out. Log locations shown in red. From Terlaky, 2016.

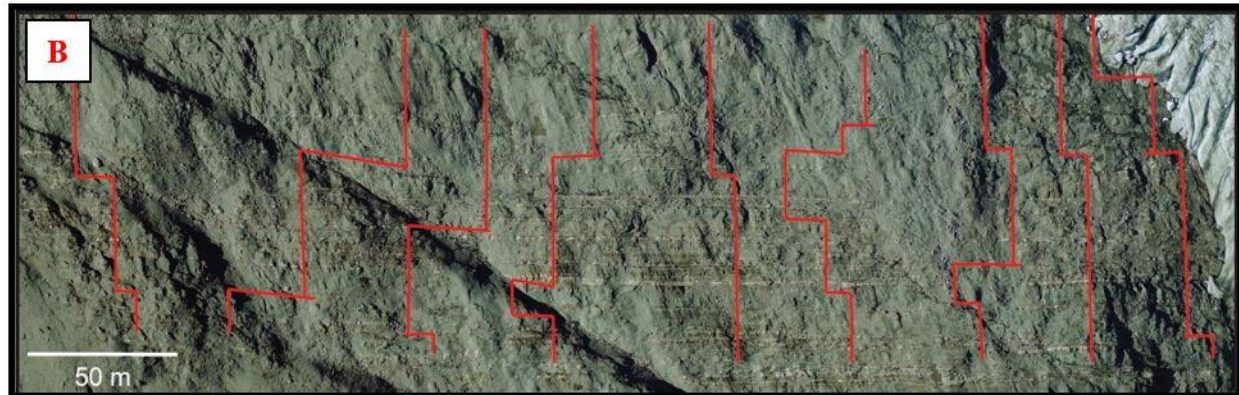


Figure 5.3: Aerial photograph of the southern study area (marked B in figure 5.1) where levee deposits crop out. Log locations shown in red.

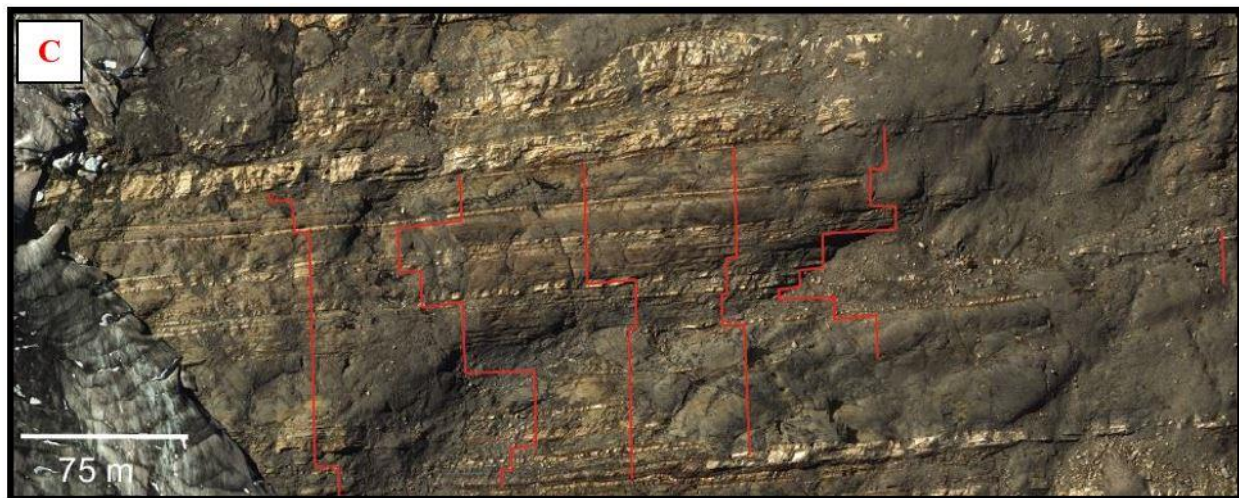


Figure 5.4: Aerial photograph of the northern study area (marked C in figure 5.1) where levee deposits crop out. Log locations shown in red.

5.1 Channels

The channel complex identified in this study crops out in the southeast drainage where it is ~130 m thick and 900 m wide (Fig. 5.1, 5.2). It consists of five channel units (CH1-5)

intercalated with mass transport deposits (Fig. 5.9). Viktor Terlaky mapped and measured stratigraphic sections in the southeast drainage over two field seasons (July-August 2014 and July-August 2015) and it is his work that is described next. At a larger scale, the five channel units exhibit a general upward-fining trend. Additionally, carbonate detritus is present in the lower two channel units (CH1-2) but absent in the upper three channel units (CH3-CH5).

5.1.1 Channel Unit 1 (CH1)

Channel unit 1 (CH1) has an erosional basal surface and overlies a ~10 m thick interval of fine-grained, thin-bedded, upper division turbidites (F2) (Fig. 5.9). CH1 ranges from 11 m to 13 m thick and is filled with amalgamated, thin- to very thick-bedded, ungraded and normally-graded, coarse- to medium-grained Ta turbidites with granules and pebbles (F1) (Fig. 5.5). At its base strata are composed of coarse-grained sandstone with locally abundant quartz and feldspar pebbles ranging from 1 cm to 4 cm in diameter. In addition, mudstone intraclasts, carbonate cemented sandstone clasts and limestone clasts are abundant and range up to several decimeters long. Clasts are rounded to semi-angular and are oriented with their long axis parallel to the base of the channel.

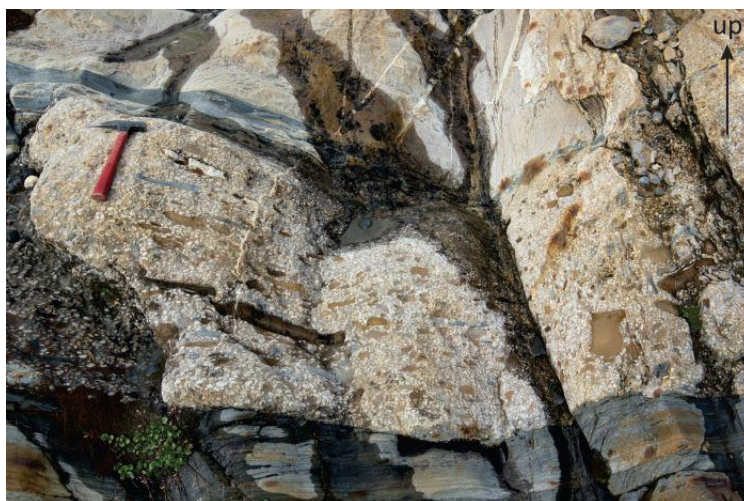


Figure 5.5: The fill of CH1 is composed of coarse- to medium-grained sandstone with abundant granules and pebbles, mudstone, carbonate cemented sandstone and limestone clasts. Photo from Terlaky, 2016.

5.1.2 Channel Unit 2 (CH2)

Channel unit 2 (CH2) directly overlies CH1 and ranges from 7 m to 11 m thick (Fig. 5.9). The basal contact of CH2 is sharp and scours into CH1. CH2 consists of amalgamated, normally-graded, coarse- and medium-grained Ta turbidites. The channel fill fines upwards and the top part of the fill fines laterally towards the north. Adjacent to CH1 and CH2 is a ~25 m thick succession of thin- to thick-bedded, traction-structured turbidites (F2), interpreted to be the genetically related levee deposits. These levee deposits extend laterally towards the north across the entire study area (a distance of ~2.5 km). Overlying CH2 is a poorly-sorted, mudstone-rich debrite (D1-F4a), which, in turn, is overlain locally by a poorly-sorted, carbonate-clast-rich debrite (D2-F4a) (Fig. 5.6) (Terlaky, 2016).

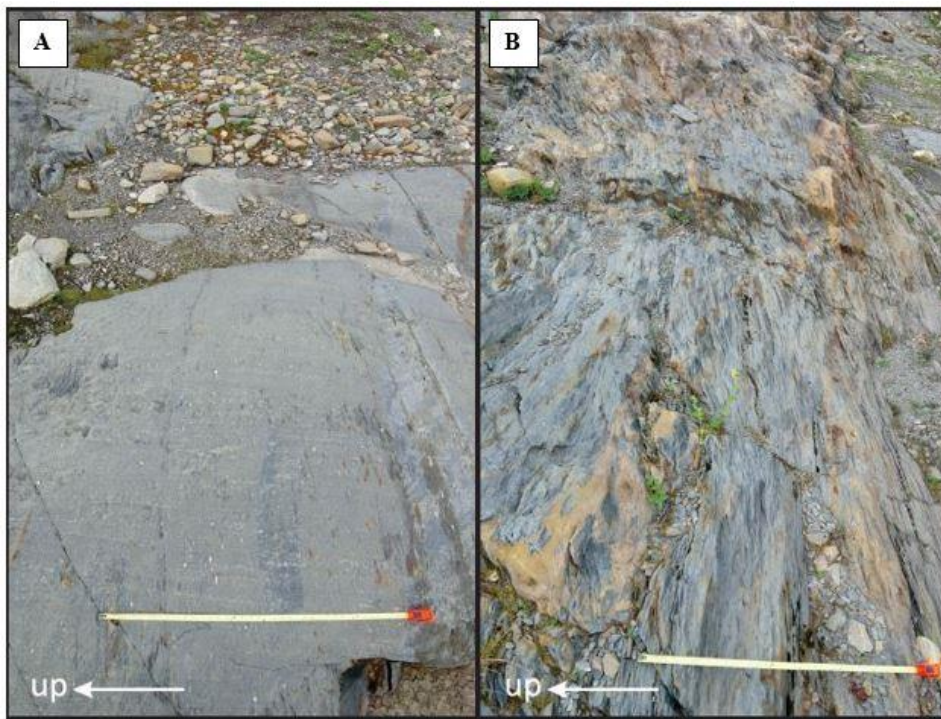


Figure 5.6: (A) Overlying CH2 is a poorly-sorted, mudstone-rich “starry night” debrite (D1-F4a). (B) The “starry night” debrite is overlain locally by a poorly-sorted, carbonate-clast-rich debrite (D2-F4a). Photo from Terlaky, 2016.

5.1.3 Channel Unit 3 (CH3)

Channel unit 3 (CH3) has an erosional basal surface and sharply overlies or onlaps debrite D1, and locally debrite D2 (F4a) (Fig. 5.9). CH3 is composed of two lithologically distinct channel fills (Fig. 5.7) and the total thickness of the channel unit ranges from 8 m to 12 m thick. The lower fill comprises amalgamated, normally-graded, upper coarse-grained Ta turbidites (F1) with local mud-clast breccia. Beds in the lower fill have undulatory basal contacts and are up to 2 m thick. Locally, strata grade upward to medium-grained sandstone and are overlain by thin mudstone caps (~1-2 cm thick) that have limited lateral extent. The upper fill consists of lower coarse- and/or upper medium-grained sandstone grading upward to fine-grained, thin-bedded, upper-division turbidites (F2) with sharp, planar basal contacts that locally are extensively ornamented with flame structures. Beds in the upper fill typically range from a few cms to a few 10s of cm thick. Like CH2, CH3 shows well-developed fining-upward and -lateral trends. The coarsest part of CH3 is on the south side of the channel unit and it fines laterally towards the north. Additionally, beds within CH3 also fine upwards (Terlaky, 2016). Towards the southeast, CH3 is overlain by a slide and towards the northwest it is overlain by CH4 (Fig. 5.9).

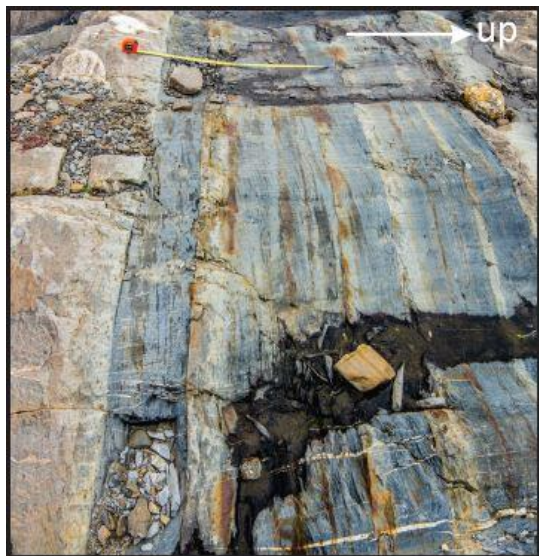


Figure 5.7: Thick-bedded turbidites of the lower fill (left) are sharply overlain by medium- to thin-bedded turbidites of the upper fill. Photo from Terlaky, 2016.

5.1.4 Channel Unit 4 (CH4)

Channel unit 4 (CH4) has an erosional basal surface and crops out on the north side of the southeast drainage outcrop where it is ~25 m thick (Fig. 5.9). Towards the southeast, this channel unit becomes eroded by the slide deposit (F4b). At its base it consists of ~3 m of amalgamated, normally-graded, coarse-grained Ta turbidites. The upper ~22 m of this channel unit is poorly exposed in the study area, and consists mainly of mudstone and siltstone.

5.1.5 Channel Unit 5 (CH5)

On the south side of the study area the base of channel unit 5 (CH5) is erosional and overlies a ~10 m thick slide (F4b) whereas on the north side it overlies CH4 (Fig. 5.9). On the north side debrite D3 (F4a) plugs CH5 except at the top of the channel unit, where the channel fill onlaps the debrite (Fig. 5.8). Like CH3, CH5 comprises two channel fills that together form a channel unit that ranges from 12 m to 20 m thick. Lithologically the lower and upper channel fills are similar to those in CH3 but they are slightly finer grained. Amalgamated, normally-graded Ta turbidites in the lower channel fill are composed of lower coarse- grading upward to medium-grained sandstone. Thin-bedded, upper division turbidites in the upper fill are composed of very fine-grained sandstone. Additionally, strata in the lower channel fill exhibit extensive soft sediment deformation where it is truncated by debrite D3. Significantly, the lower part of the lower fill is laterally continuous under debrite D3 (Fig. 5.8). Similar to CH3, CH5 shows well-developed upward and lateral fining. The coarsest part of CH5 is on the south side of the channel fill and sandstone beds fine stratigraphically upwards. Also, grain size decreases laterally to the north across the channel unit. CH5 is overlain by a ~30 m thick slide (F4b) that extends continuously from the southeast drainage to the glacier. This is then overlain by debrite D4 that extends continuously across the southern and northern parts of the study area (Fig. 5.9) (Terlaky, 2016).

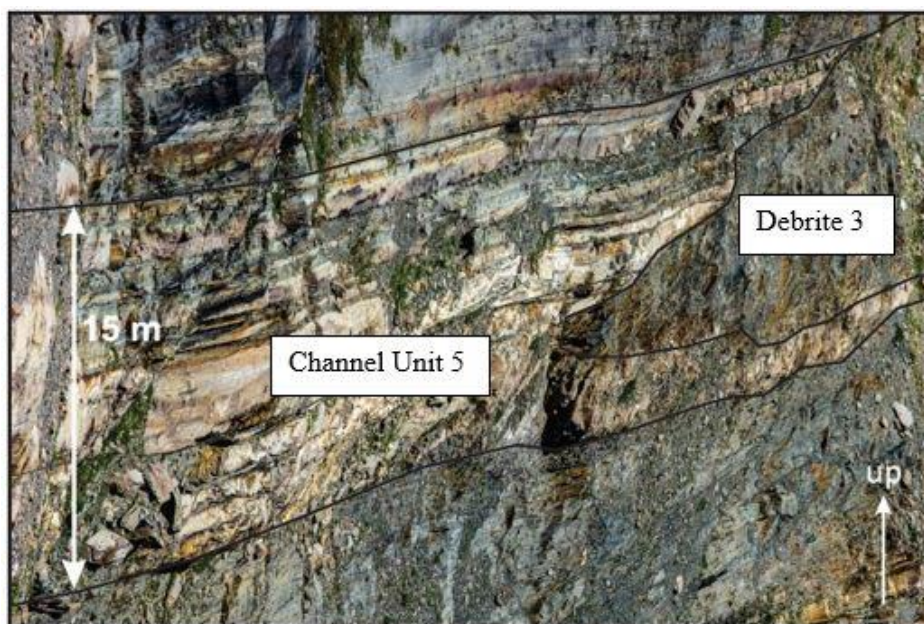


Figure 5.8: On the north side of the southeast drainage outcrop, debrite D3 (F4a) plugs CH5 except at the top of the channel unit, where the channel fill onlaps the debrite. Photo from Terlaky, 2016.

Figure 5.9: Strata in the southeast drainage outcrop. Top is an aerial photograph showing the contacts between the channel units, levee deposits and mass transport deposits identified in the study area. Bottom is a correlation panel of the strata showing the detailed architecture of the channel complex. (Modified from Terlaky, 2016).

5.1.6 Channel Complex Interpretation

In the study area the channel complex that crops out in the southeast drainage comprises five channel units bounded laterally on their northern margins by levees; counterpart levees on the south side are not exposed. The erosional base of each channel unit is interpreted to represent the inception stage of each channel related to an initial period of net erosion. Early erosional flows excavated a conduit for sediment being transported (i.e., bypassed) further basinward. Later, less efficient flows deposited sediment in these channels during the channel infill stage.

Channel units 1 and 2 are composed mostly of normally-graded, coarse-grained Ta turbidites (F1) deposited from high-concentration turbidity currents rich in quartz pebbles, carbonate-cemented sandstone and limestone clasts. Carbonate-cemented sandstone and carbonate clasts suggest sourcing from a most probably recently deactivated mixed siliciclastic-carbonate shelf system, followed by long distance downslope transport. The predominance of highly amalgamated beds in these channel units indicates that turbidity currents were energetic and erosive with their upper, finer-grained portion overspilling the margins of the channel and building up coeval levees (see section 5.2).

Channel units 3, 4 and 5 each comprise two lithologically distinct channel fills. Amalgamated, normally-graded, upper coarse- to medium-grained Ta turbidites in the lower channel fills with local mud-clast breccia were deposited by highly erosive, stratified, sand-rich, high-concentration turbidity currents. The upper part of channel units 3 and 5 consists of thin-bedded, fine-grained, upper division turbidites (F2); whereas in channel unit 4 it consists mostly of mudstone and siltstone, suggesting a dramatic decrease in flow energy. The upper fine-grained part of channel units 3 and 5 are interpreted to represent the partial abandonment of the channel, where high-concentration, sand-rich turbidity currents were redirected away from the channel and

only lower-concentration, less competent turbidity currents flowed through the channel. Channel unit 4, on the other hand, is interpreted to have been abandoned and later was filled by the fallout of clay and silt from flows that overspilled a nearby channel.

The fills of channel units 1, 2 3 and 5 are consistently coarsest in the southern part of the study area and then fine laterally towards the north, interpreted to reflect a channel axis to channel margin transect. These strata are then sharply bounded by generally finer-grained levee deposits along their northern margins (see section 5.2).

5.2 Levees

Levee deposits interpreted to be genetically related to the channel complex described previously in section 5.1 crop out in the southeast drainage (~ 863 m wide), southern (~382 m wide), and northern (~323 m wide) outcrops (Fig. 5.9). Although separated by moraine (~190 m wide) and the Castle Creek Glacier (~819 m wide), intercalated mass transport deposits, or MTDs, allow strata to be confidently correlated across the full width of the study area -- a distance of 2.6 km (Fig. 5.10). Based on their lithological makeup they have been divided into three separate types of levees which are described and interpreted next.

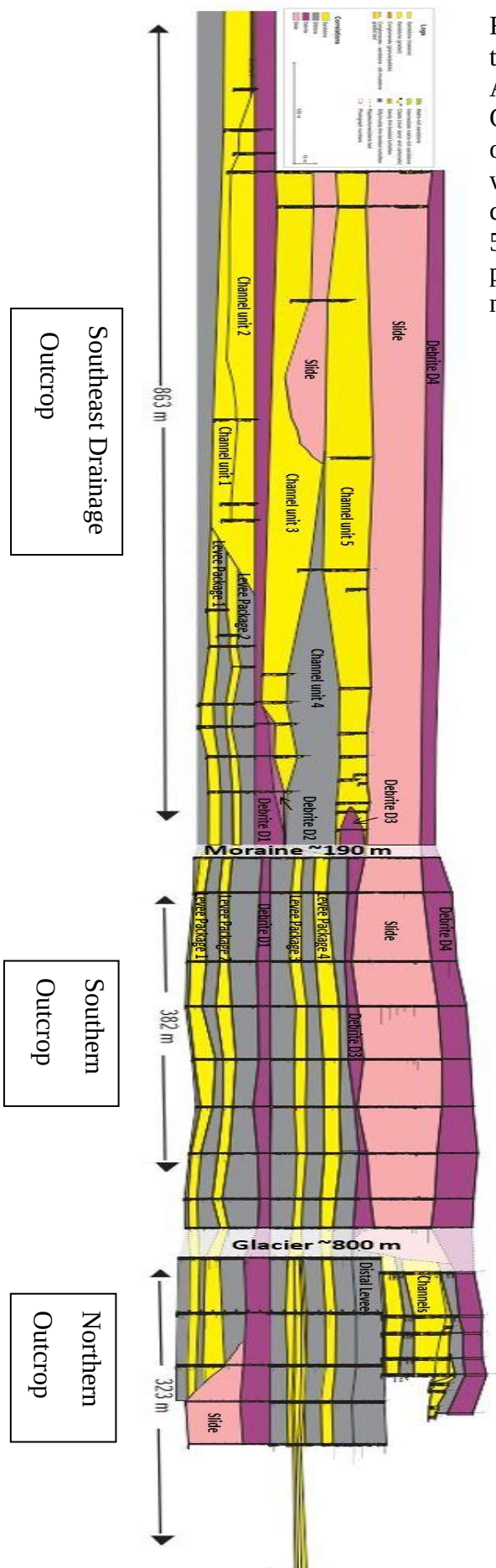


Figure 5.10: Correlation panel of the strata in all three outcrops of the study showing the detailed architecture of the channel complex. Although separated by moraine (~190 m wide) and the Castle Creek Glacier (~819 m wide), intercalated mass transport deposits, or MTDs, allow strata to be confidently correlated across the full width of the study area -- a distance of 2.6 km. The channel complex crops out in the southeast drainage outcrop (see section 5.1) and the genetically-related levees crop out on the northernmost part of the southeast drainage outcrop, across the southern and northern outcrops.

5.2.1 Type 1 Levee Deposits: Description of Vertical Trends

Type 1 levee deposits make up the majority of levee strata in the study areas. The vertical trends observed in these levee deposits in all three study areas (southeast drainage, southern and northern outcrops) are similar, and will be discussed (this section) and interpreted (section 5.2.3) together.

In the southern and northern study areas, type 1 levee deposits are divided vertically into four correlatable stratal packages (P1-P4); in the southeast drainage study area only P1 and P2 crop out (Fig. 5.10). Packages range from 6 to 21 m thick and can easily be subdivided into lower and upper parts, the transition being sharp and marked by a dramatic decrease in bed thickness, grain size and sandstone content (Fig. 5.11, 5.12). In addition, but at a larger scale, the four packages exhibit a general upward fining trend (Fig. 5.15).

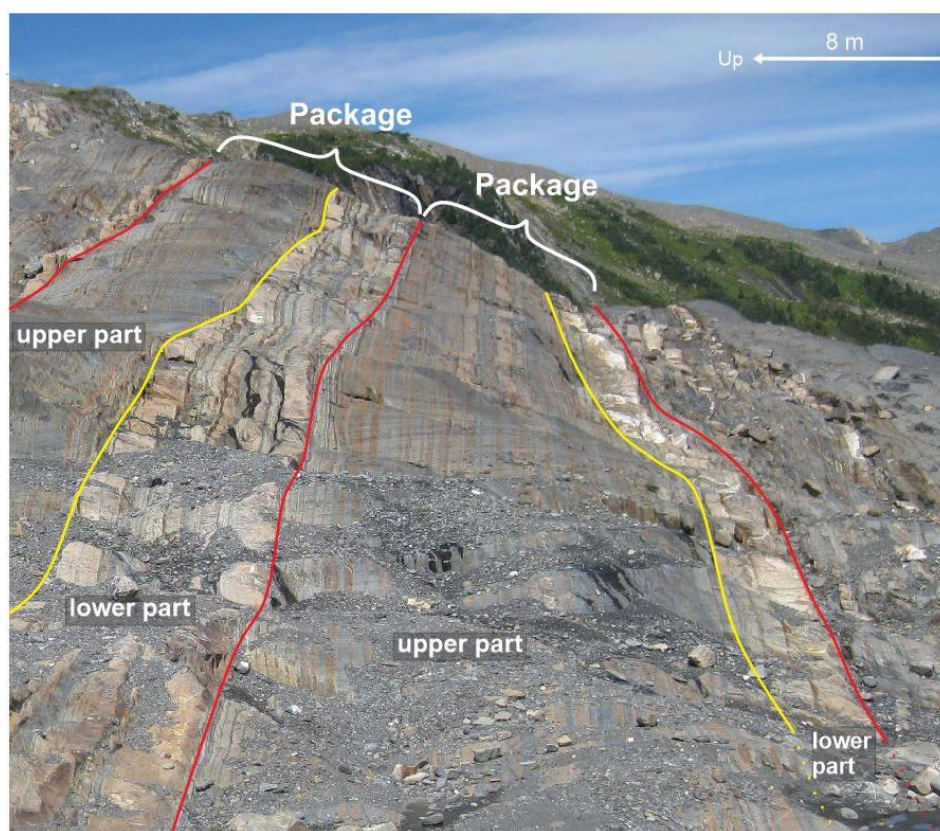


Figure 5.11: Type 1 levee deposits are divided vertically into four correlatable stratal packages, two of which are shown here. Packages can be easily subdivided into lower and upper parts, the transition being sharp and marked by a dramatic decrease in bed thickness, grain size and sandstone content. Red lines mark the base of each package, yellow lines mark the contact between the two parts.

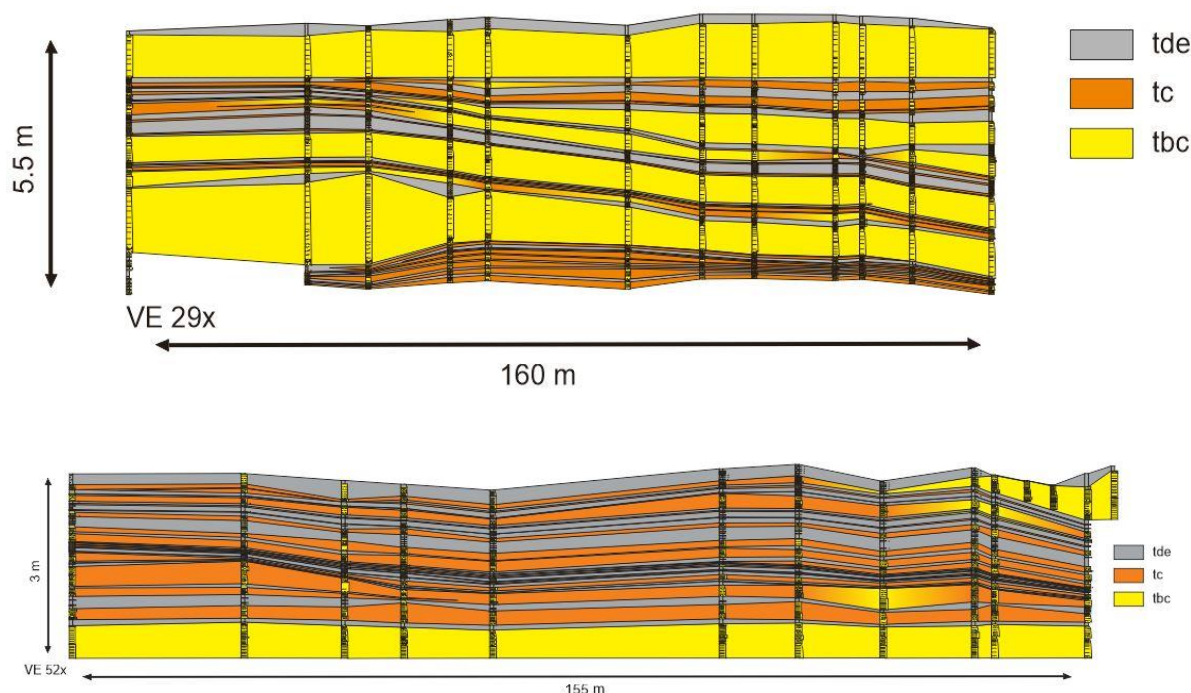


Figure 5.12: (Top) The quality of the northern outcrop is excellent with almost 100% exposure. Correlation panel of extra logs spaced 5-20 m apart to document small-scale lateral changes in individual beds in the lower part of a package. (Bottom) Correlation panel of extra logs spaced 5-20 m in the northern outcrop to document small-scale lateral changes in individual beds in the upper part of a package.

The lower part of packages (LP1-LP4) ranges from 3 to 10 m thick and consists mostly of medium- to thick-bedded, upper medium- to coarse-grained, lower division turbidites intercalated with very thin- to thin-bedded, fine-grained, upper division turbidites (Fig. 5.13). The upper part of packages (UP1-UP4) ranges from 2 to 16 m thick and consists mostly of very thin- to thin-bedded, fine-grained, upper division turbidites intercalated with medium- to thick-bedded, upper medium-grained, lower division turbidites (Fig. 5.14). Significantly, the thickness of very thin- and thin-bedded turbidites decreases stratigraphically upward whereas the thickness of intercalated medium- and thick-bedded turbidites changes little in the upper part (Fig. 5.15). Very thin- and thin-bedded turbidites are abundant throughout the upper part however the abundance of medium- and thick-bedded turbidites decreases upward (Fig. 5.15). The specific stratal makeup of packages is described next.

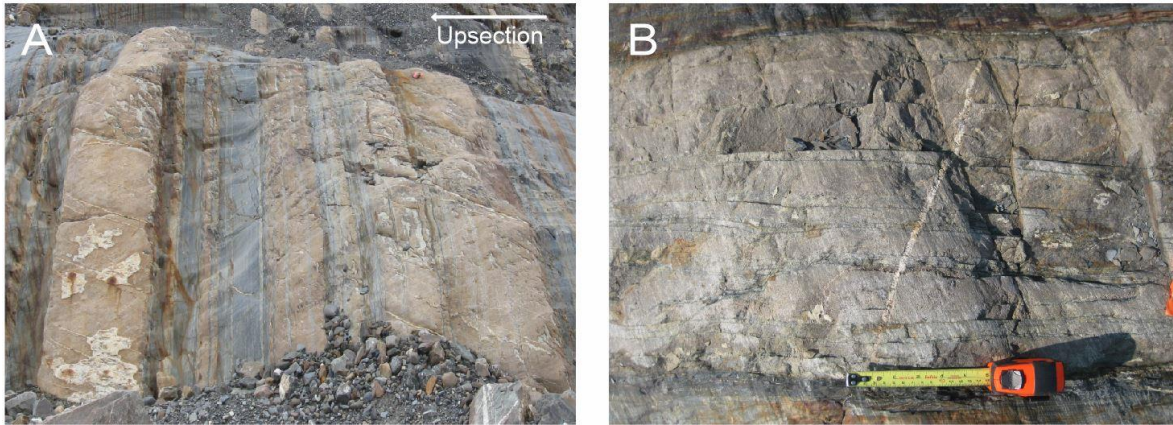


Figure 5.13: (A) The lower part of packages consists mostly of medium- to thick-bedded, upper medium- to coarse-grained, lower division turbidites intercalated with very thin- to thin-bedded, fine-grained, upper division turbidites. (B) Close up of a thick-bedded, upper medium-grained Tbcd/e turbidite in the lower part of a package.

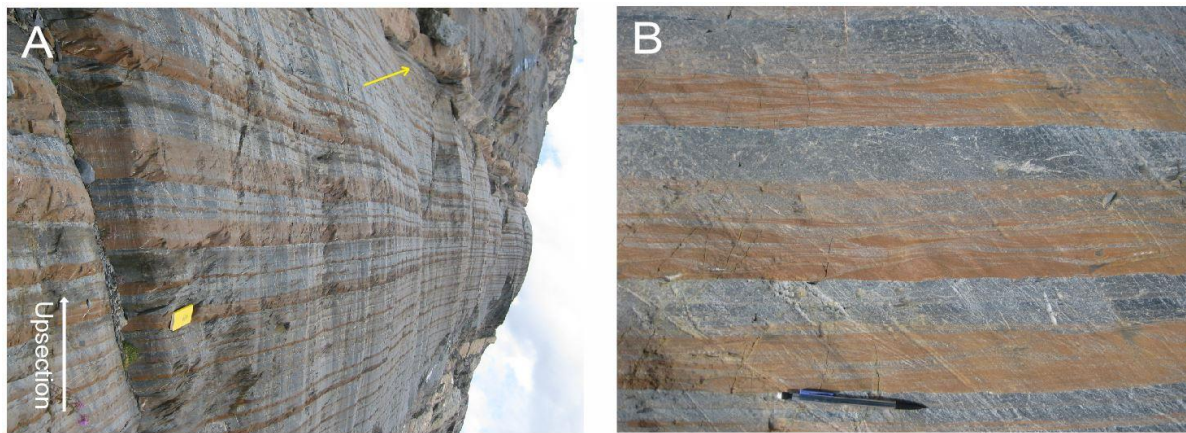


Figure 5.14: (A) The upper part of packages consists mostly of very thin- to thin-bedded, fine-grained, upper division turbidites intercalated with medium- to thick-bedded, upper medium-grained, lower division turbidites. Yellow arrow points to a medium-bedded, upper medium-grained Tbcd/e turbidite. (B) Close up of thin-bedded, fine-grained Tcd/e turbidites in the upper part of a package.

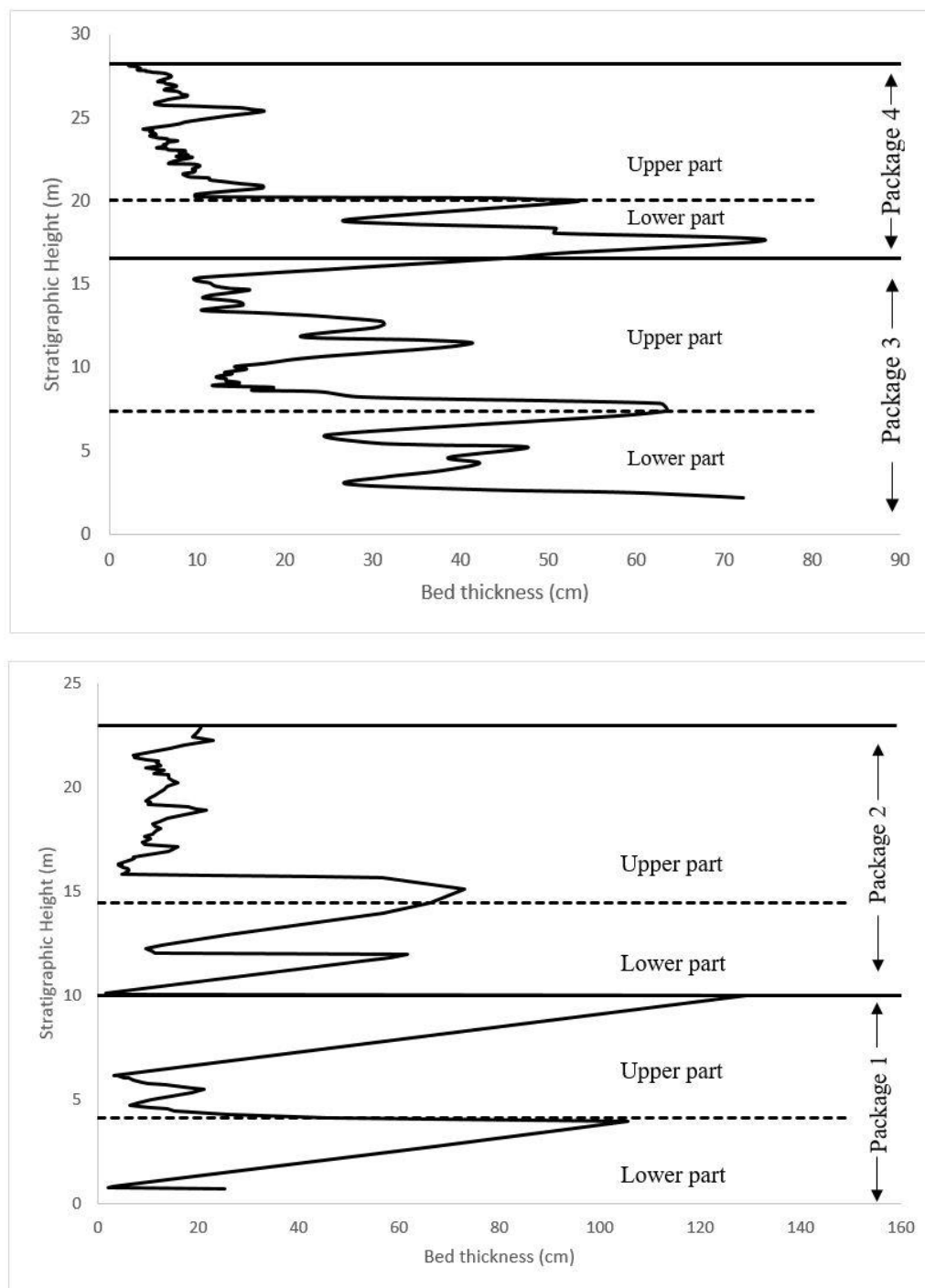


Figure 5.15: Thickness data of the 295 beds that make up packages 1-4 in the northern outcrop shows how thickness varies upwards. Data is from 1 log and these plots use a 3-bed moving average. Note the dramatic decrease in bed thickness at the contact between the lower and upper parts of a package. At a large scale, the four packages exhibit a general upward fining trend. At the small scale, note how the thickness of beds in the upper part decrease stratigraphically upward. Also, medium- and thick-bedded turbidites rarely crop out in the upper portion of the upper parts.

5.2.2 Type 1 Levee Deposits: Description of Lateral Trends

Although type 1 levee deposits in all three study areas (southeast drainage, southern and northern outcrops) exhibit a similar vertical packaging (section 5.2.1), lateral trends in thickness, grain size and bed structures differ and are described here. Based on similarities, this section groups the trends observed in the two lowermost packages (P1 and P2), and those in the two uppermost packages (P3 and P4) (Fig. 5.10). Also, trends observed in the southeast drainage and southern outcrops are described first followed by a description of their correlative packages in the northern outcrop.

In the lower part of packages, thickness, grain size, and structure of each bed were recorded. It was difficult to differentiate between the Td and Te portion of each turbidite, therefore these portions were lumped and measured together as a single Td/e portion. Due to outcrop exposure in the southern outcrop and time constraints, the same logging methods were not used for the upper part of packages. Instead, the upper parts were subdivided into units ranging between 50 cm to 2 m thick based on lithological similarities. Within these units several representative beds were measured and from these, minimum, maximum and average bed thickness of single set Tc, multiple set Tc, and Td/e turbidites were recorded. In addition, grain size and any general trends noted within the unit were also recorded (for example: upward fining and thinning of multiple set Tc turbidites). Nevertheless, all thick-bedded Tbc turbidites were measured in the upper part. Outcrop exposure is excellent in the northern outcrop, and therefore logs were spaced 5-20 m apart in order to document lateral changes of 20 beds in a lower part of the package and 19 beds in the upper part.

Levee Packages 1 and 2: southeast drainage and southern outcrops

Packages 1 and 2 (P1 and P2) are the only two levee packages exposed in the southeast drainage outcrop and here extend outward from the margin of their genetically related channels (CH1 and CH2) for a distance of ~300 m. They can then be correlated across the moraine (width of ~190 m) to P1 and P2 in the southern outcrop where they extend laterally for another ~382 m (Fig. 5.10).

The lower part of P1 and P2 (LP1 and LP2, respectively) thicken and then thin laterally away from the margins of CH1 and CH2 across the southeast drainage and southern outcrops (Fig. 5.16). More specifically, it is the medium- to thick-bedded, coarser-grained Tb/Tbc turbidites that exhibit this trend (Fig. 5.13B), and collectively control the thickness trends in the lower part of each package. In LP1 and LP2 individual beds thicken over the first 662 m and 604 m, causing the thickness of LP1 and LP2 to increase by 5.21 m and 3.51 m, respectively. Upon reaching their maximum thickness, beds then thin, and the total thickness of LP1 and LP2 decreases by 5.56 m over 175 m and 2.79 m over 233 m, respectively (Fig. 5.16).

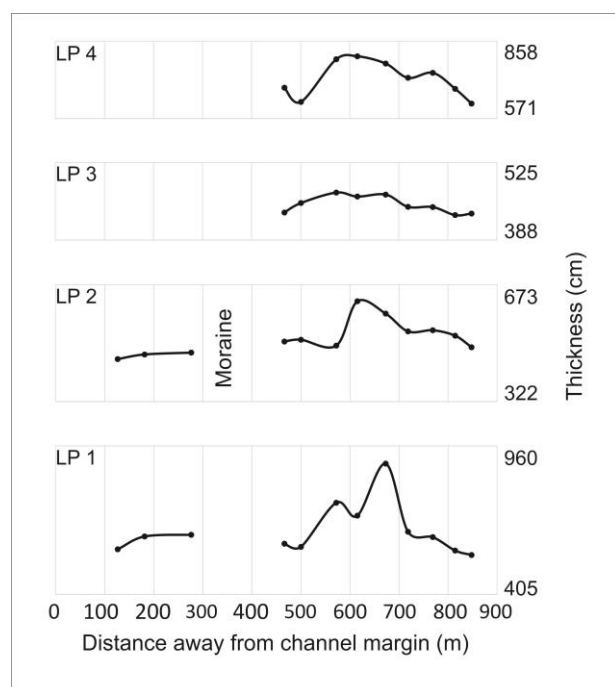


Figure 5.16: Lateral thickness (black lines) changes in the lower part of packages (LP) that extend from the southeast drainage outcrop (left) across the moraine to the southern outcrop (total lateral distance of 872 m). Packages are labelled LP1 to LP4. Minimum and maximum thickness of each LP are indicated by the numbers on the right vertical axis.

In the southeast drainage outcrop the basal part of LP1 and LP2 consist of very thick-bedded, very coarse- and coarse-grained Ta/Tab turbidites that average 179 cm in thickness. Stratigraphically upwards, beds thin to thick-bedded, coarse-grained Ta turbidites (average 55 cm thick) interstratified with medium-bedded, coarse- to upper medium-grained Tb/Tbc turbidites that average 21 cm thick. Almost all beds are overlain by a siltstone/mudstone Td/e cap, except where removed by erosion, which average 12 cm in thickness. The ratio of total sandstone thickness to total stratal thickness, or net-to-gross, of LP1 and LP2 in the southeast drainage outcrop averages 90% (Fig. 5.17).

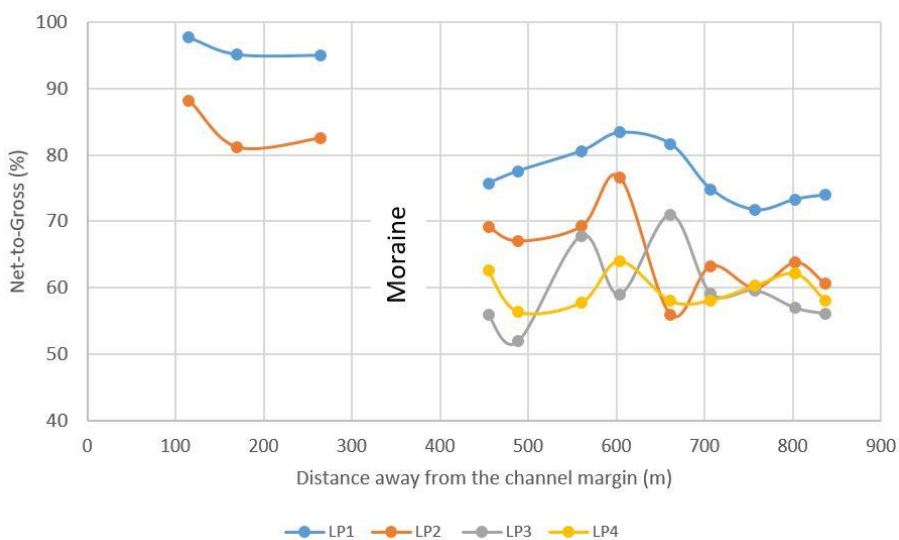


Figure 5.17: Lateral change in net-to-gross of the lower parts of packages (LP1 to LP4) extending from the southeast drainage outcrop (far left) across the moraine to the southern outcrop (total lateral distance of 872 m). The range of net-to-gross (%) for the lower parts are shown on the left vertical axis.

Across the moraine (~190 m wide) in the southern outcrop, the lithological makeup of LP1 and LP2 changes little upward. Here LP1 and LP2 consist mostly of thick-bedded, upper medium- and coarse-grained Tb/Tbc turbidites that average 65 cm in thickness. These strata are intercalated with very thin-bedded, fine- and lower medium-grained Tc turbidites that average 1.4 cm thick (single set) and 3.4 cm thick (multiple (2-3) set). Almost all beds are overlain by a

siltstone/mudstone Td/e cap, except where removed by erosion, which on average is 10 cm thick. The lower division turbidites thicken laterally to an average of 133 cm over the first 662 m and 604 m away from the channels, until LP1 and LP2 reach their maximum thicknesses, respectively. Upon reaching their maximum thickness these beds then thin to thick-bedded turbidites that average 55 cm thick. The grain size of these lower division turbidites changes little laterally across the southern outcrop. Intercalated Tc turbidites change little in grain size or thickness across the width of the southern outcrop. The net-to-gross averages 73 % proximal to the moraine, increases to an average of 80 % where the lower part of packages achieve their maximum thickness, and then decreases as they thin to an average of 67 % adjacent to the glacier (Fig. 5.17).

Strata forming the upper parts of P1 and P2 (UP1 and UP2) thin as underlying LP1 and LP2 thicken laterally away from the margins of CH1 and CH2. Conversely, they then thicken as LP1 and LP2 thin after reaching their maximum thicknesses (Fig. 5.18, 5.19).

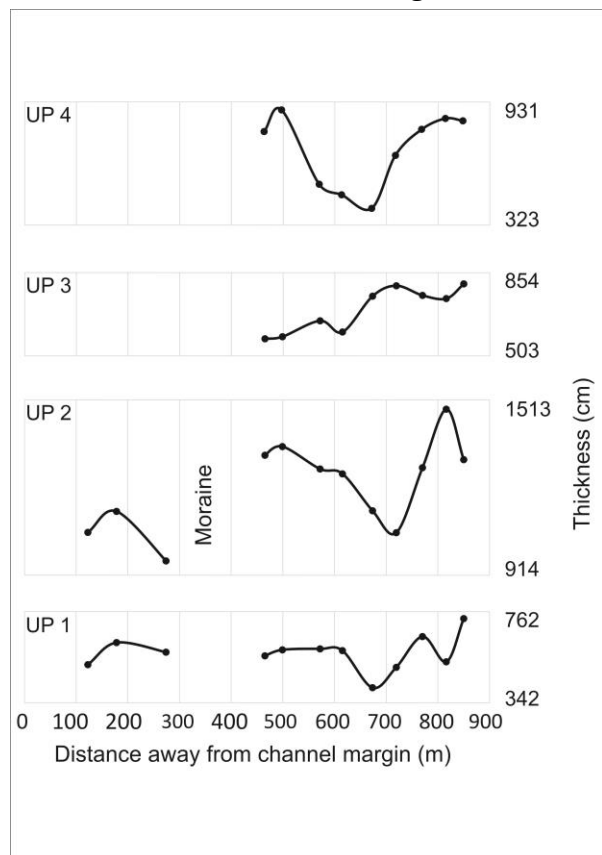


Figure 5.18: Lateral thickness (black lines) changes in the upper part of packages (UP) that extend from the southeast drainage outcrop (left) across the moraine to the southern outcrop (total lateral distance of 872 m). Packages are labelled LP1 to LP4. Minimum and maximum thickness of each UP are indicated by the numbers on the right vertical axis.

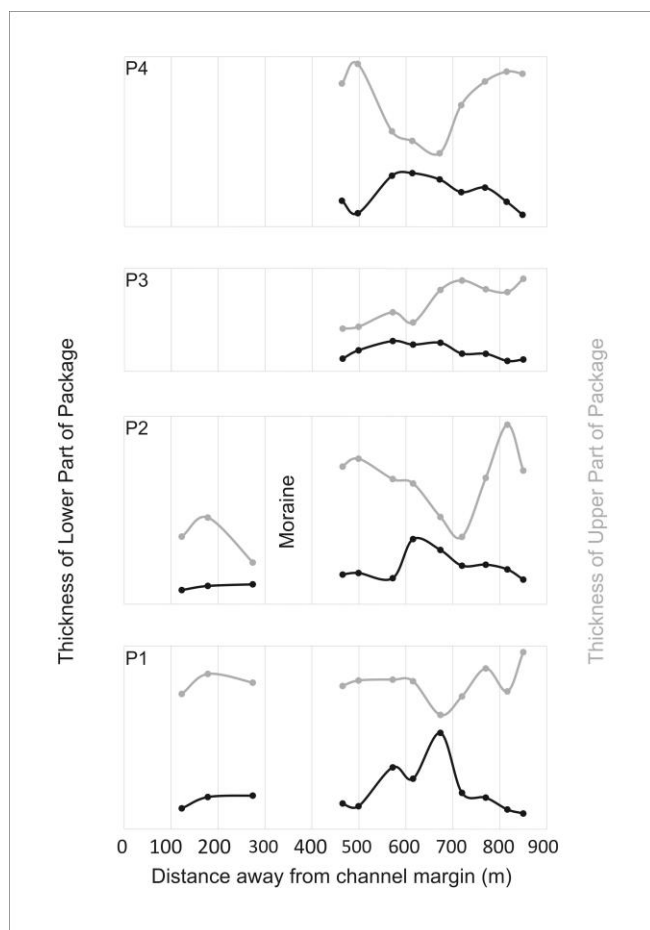


Figure 5.19: Lateral changes in thickness of the lower (black lines) and upper (grey lines) parts of packages extending from the southeast drainage outcrop (left) across the moraine to the southern outcrop (a total lateral distance of 872 m). Packages are labelled P1 to P4. Note the antithetic thickening and thinning patterns of the lower and upper parts of each package.

In the southeast drainage outcrop UP1 and UP2 consist mostly of thin-bedded, lower medium-grained Tb/Tbc turbidites that average 8 cm thick. These strata are intercalated with medium- and thick-bedded, upper medium- to coarse-grained lower division turbidites that average 19 cm and 52 cm thick, respectively. Almost all beds are overlain by siltstone/mudstone Td/e caps, except where removed by erosion, and average 19 cm thick. On average the net-to-gross initially increases from 52% to 62% over the first 170 m laterally away from the margins of CH1 and CH2 before decreasing to 48% over the next 95 m towards the moraine (Fig. 5.20).

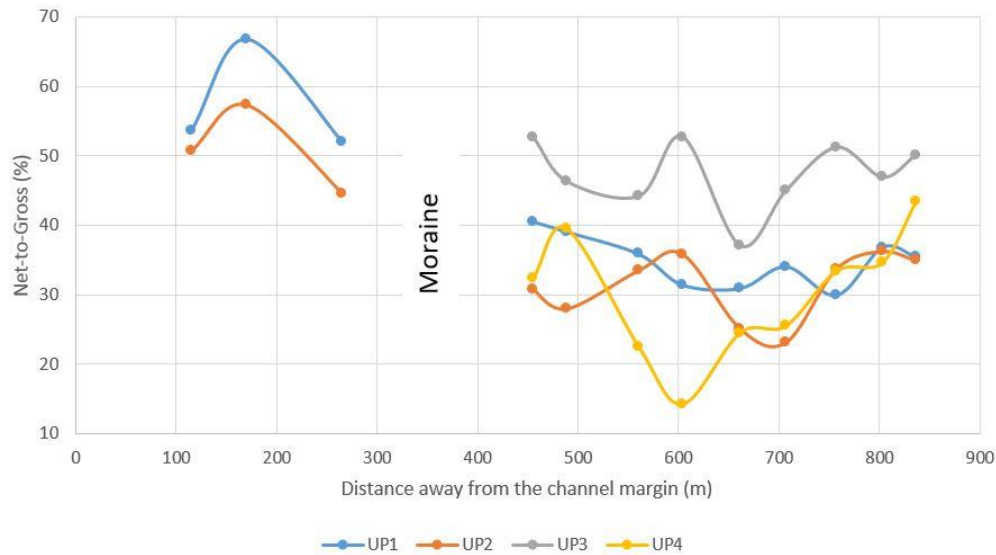


Figure 5.20: Lateral change in net-to-gross of the upper parts of packages (UP1 to UP4) extending from the southeast drainage outcrop (UP1 and UP2) (left) across the moraine to the southern outcrop (total lateral distance of 872 m).

In the southern outcrop (~382 m) UP1 and UP2 consist mostly of very thin- and thin-bedded, fine-grained Tc turbidites that average 1.4 cm (single set) and 5.7 cm (multiple (3-4) set) thick, respectively. These strata are overlain by siltstone/mudstone Td/e caps that average 15 cm thick. These upper division turbidites are intercalated with medium-bedded, upper medium-grained Tbc turbidites that average 25 cm thick. Significantly, the thickness of multiple (3-4) set Tc turbidites and of Td/e caps on average thin by 2 cm and by 5 cm, respectively, as LP1 and LP2 thicken to their maximum. Beyond this point, multiple (3-4) set Tc turbidites and Td/e caps then thicken, on average, by 1 cm (Fig. 5.21). In addition, the grain size of these strata changes little across the width of the outcrop. In comparison, the thickness of single set Tc turbidites and medium-bedded Tbc turbidites remains similar across the outcrop, as does their grain size. Average net-to-gross is 33 % and generally changes little laterally across the outcrop (Fig. 5.20).

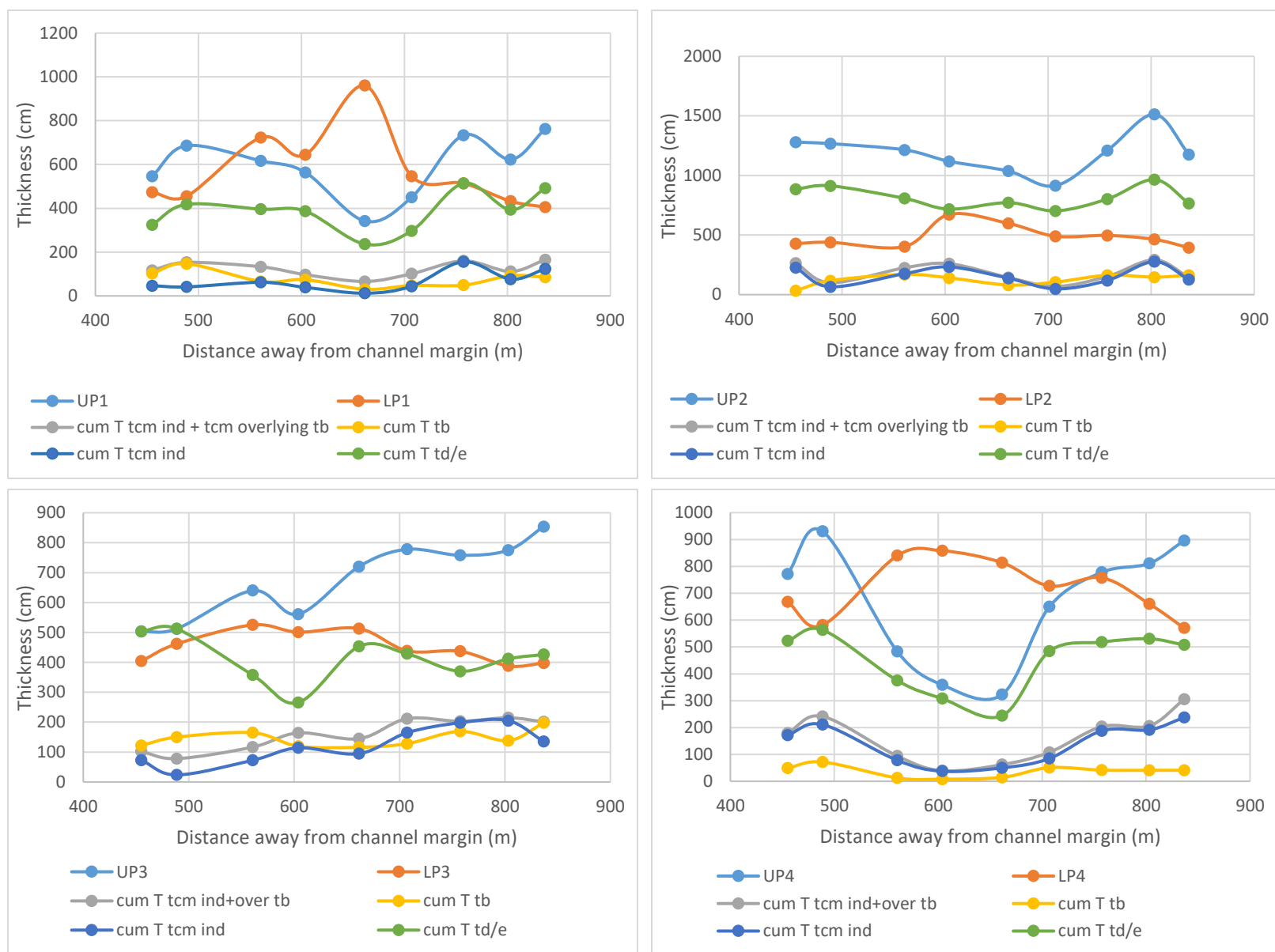


Figure 5.21: Lateral changes in thickness of the lower (orange) and upper (light blue) parts of packages extending across the southern outcrop for packages 1 to 4. Note the antithetic thickening and thinning patterns of the lower and upper parts of each package. Cumulative thickness of Tb turbidites (yellow), individual multiple set Tc turbidites (dark blue), individual multiple set Tc turbidites plus multiple set Tc turbidites overlying Tb turbidites (grey), and Td/e turbidites (green). Note that it is the Td/e turbidites, and, to a lesser extent, the multiple set Tc turbidites that control the antithetic thickening and thinning of the upper parts.

Levee Packages 1 and 2: northern outcrop

P1 and P2 in the southern outcrop extend across the glacier (~819 m wide) to the northern outcrop, where they are exposed over a lateral distance of ~110 m. Here the thickness of LP1

changes little whereas the thickness of LP2 decreases by ~3.33 m towards the glacier (southeast) (Fig. 5.22).

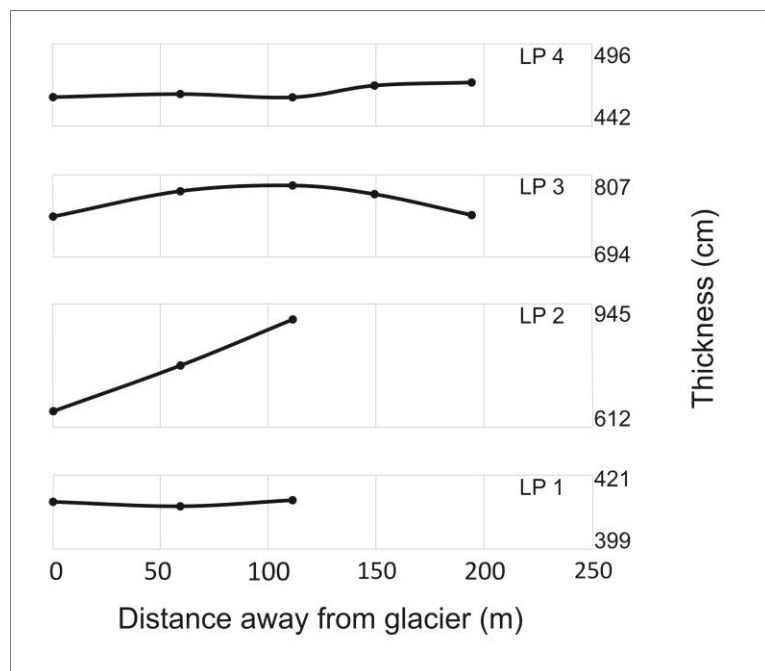


Figure 5.22: Lateral thickness (black lines) changes in the lower parts of packages (LP) that extend from the glacier (left) across the northern outcrop (total lateral distance of 194 m). Packages are labelled LP1 to LP4. Minimum and maximum thickness of each LP are indicated by the numbers on the right vertical axis.

Like in the southern outcrop, the stratal makeup of LP1 and LP2 changes little vertically. Here LP1 and LP2 consist principally of thick-bedded, coarse- and very coarse-grained Tb/Tbc turbidites that average 52 cm thick. These strata are intercalated with very thin-bedded, fine- and lower medium-grained Tc turbidites that average 1 cm thick (single set) and 5 cm thick (multiple (3-4) set) intercalated with rare very thick-bedded, coarse-grained Tb/Tbc turbidites that average 111 cm thick. Almost all beds are overlain by siltstone/mudstone Td/e caps, except where removed by erosion, which average 10 cm thick. About half of the lower division turbidites which make up the LPs thin laterally towards the glacier while the other half either thicken or remain similar in thickness (Fig. 5.23). In addition, some of the beds which thin towards the glacier also fine. Average net-to-gross is 86 % and changes little laterally across the outcrop (Fig. 5.24).

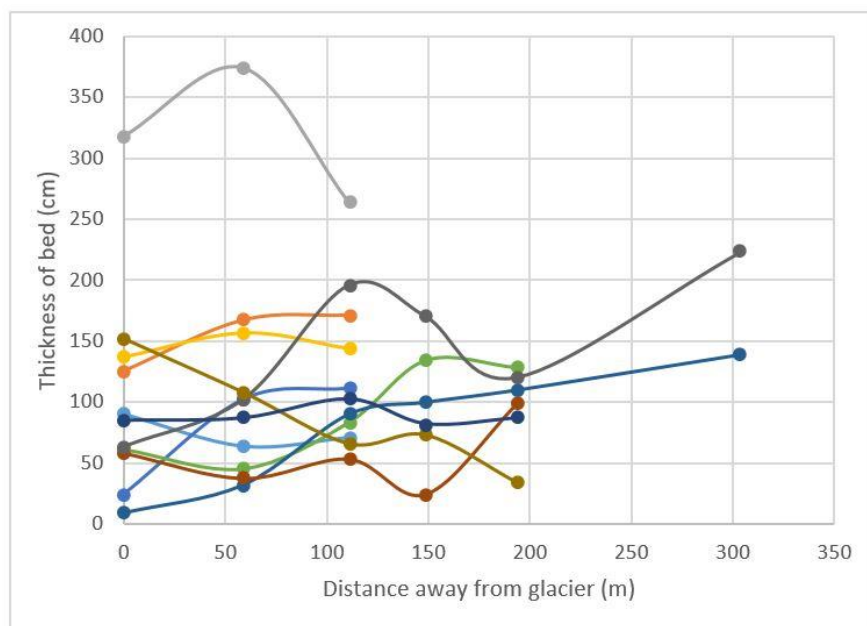


Figure 5.23: Thickness of select lower division turbidites from LP 1 to 4 in the northern outcrop. Graph on left shows how their thickness changes towards the glacier. Graph on right spaces them out in order to better show the thickness trends. About half of the beds thin towards the glacier while the other half either thicken or remain similar in thickness. Note that in general, beds do not thicken or thin linearly (smoothly).

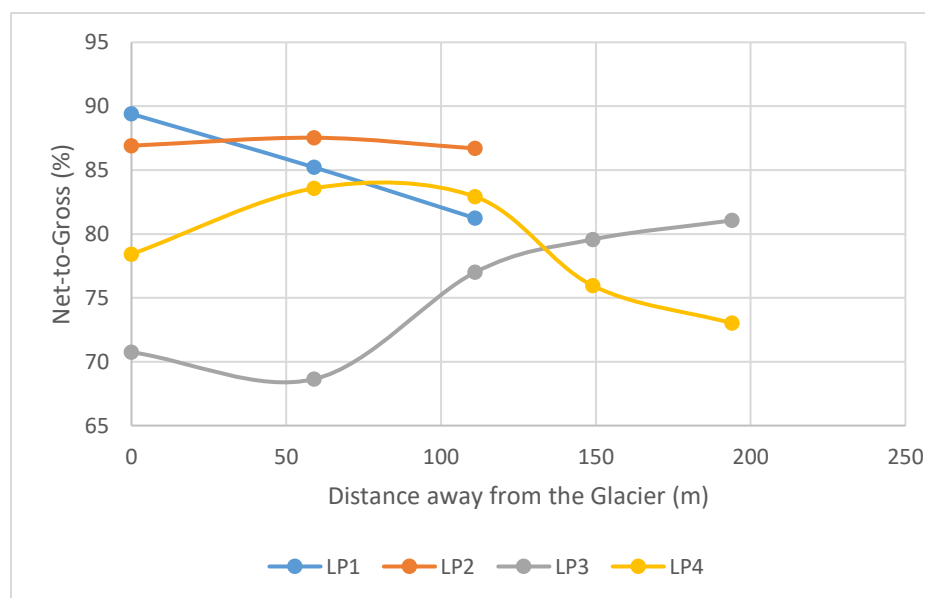
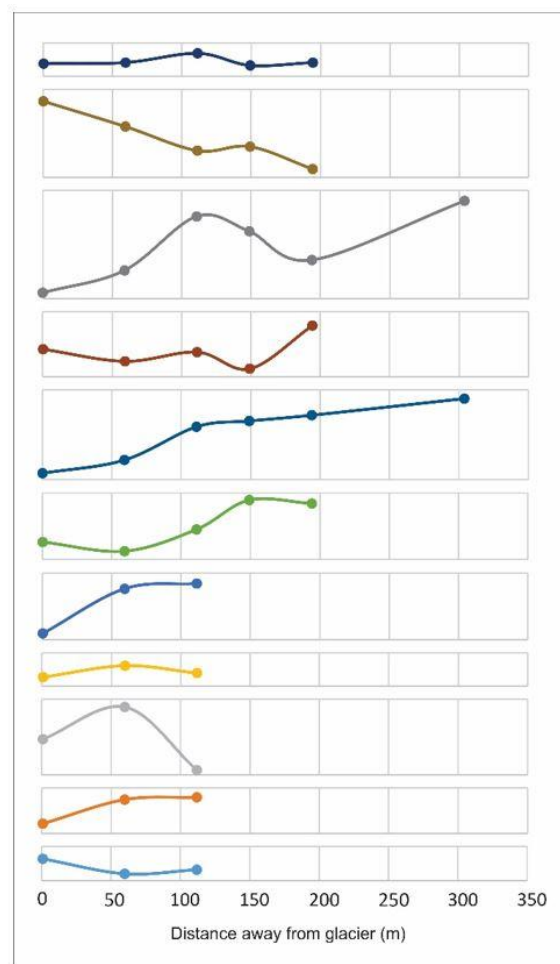


Figure 5.24: Lateral change in net-to-gross in the lower part of packages (LP1 to LP4) that extend from the glacier (far left) across the northern outcrop (total lateral distance of 194 m). The range of net-to-gross (%) for the lower parts are shown on the left vertical axis.

In the northern outcrop, UP1 changes little in thickness whereas UP2 thins as LP2 thickens (Fig. 5.25, 5.26). Like in the southern outcrop, UP1 and UP2 consist primarily of very thin- and thin-bedded, fine-grained Tc turbidites that average 1.2 cm (single set) and 5.8 cm (multiple (3-4) set) thick, respectively. These strata are overlain by siltstone/mudstone Td/e caps that average 7.5 cm thick. These upper division turbidites are intercalated with medium-bedded, upper medium-grained Tbc turbidites that average 26 cm thick. Average net-to-gross is 38 % and changes little laterally across the outcrop (Fig. 5.27). Upper division turbidites in the UPs remain similar in thickness and grain size across the width of the outcrop. The cumulative thickness of multiple set Tc turbidites and of Td/e turbidites changes little across the outcrop in UP1 but thickens towards the glacier in UP2 as LP2 thins (Fig. 5.28). In addition, medium-bedded, upper division turbidites also remain similar in thickness and grain size across the width of the outcrop.

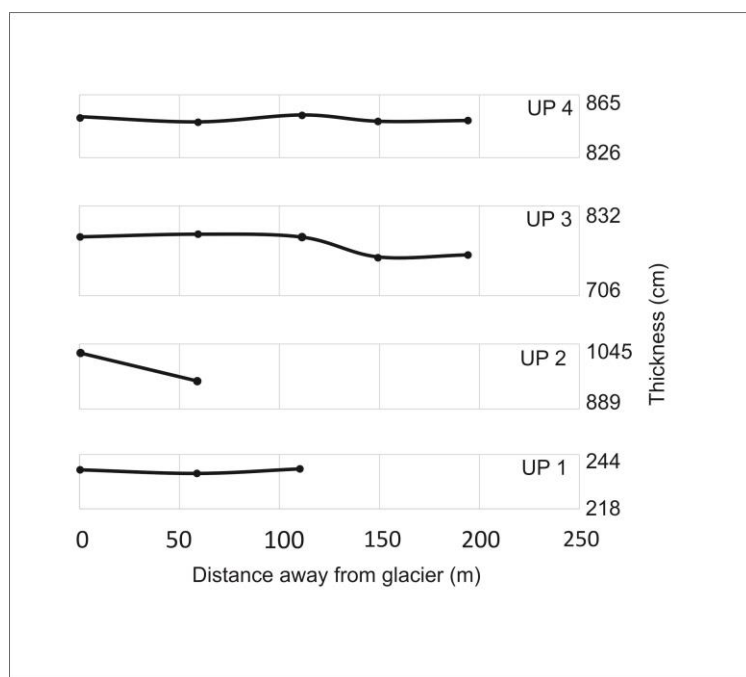


Figure 5.25: Lateral thickness (black lines) changes in the upper parts of packages (UP) that extend from the glacier (left) across the northern outcrop (total lateral distance of 194 m). Packages are labelled LP1 to LP4. Minimum and maximum thickness of each UP are indicated by the numbers on the right vertical axis.

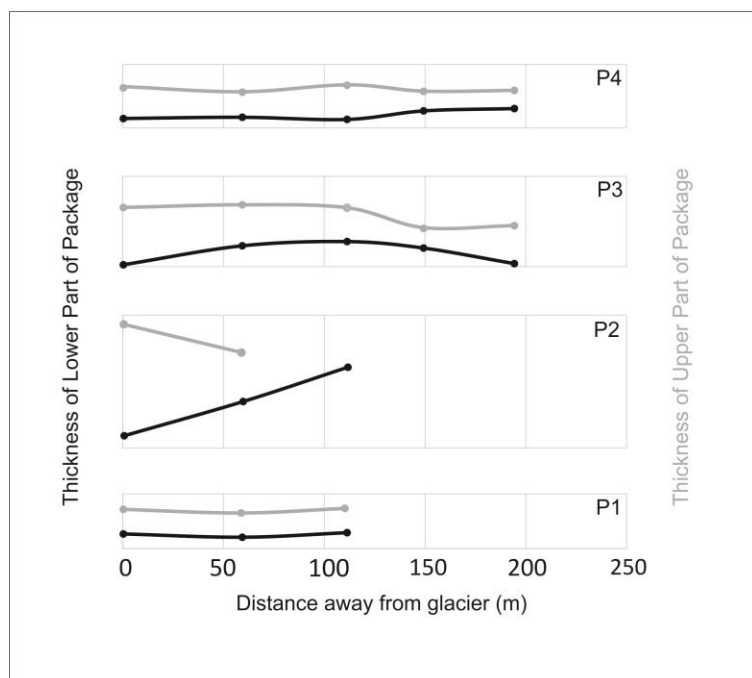


Figure 5.26: Lateral change in thickness of the lower (black lines) and upper (grey lines) parts of packages that extend from the glacier (left) across the northern outcrop (total lateral distance of 194 m). Packages are labelled P1 to P4. Note the antithetic thickening and thinning of LP2 and UP2. Other packages only show subtle antithetic thickening and thinning trends.

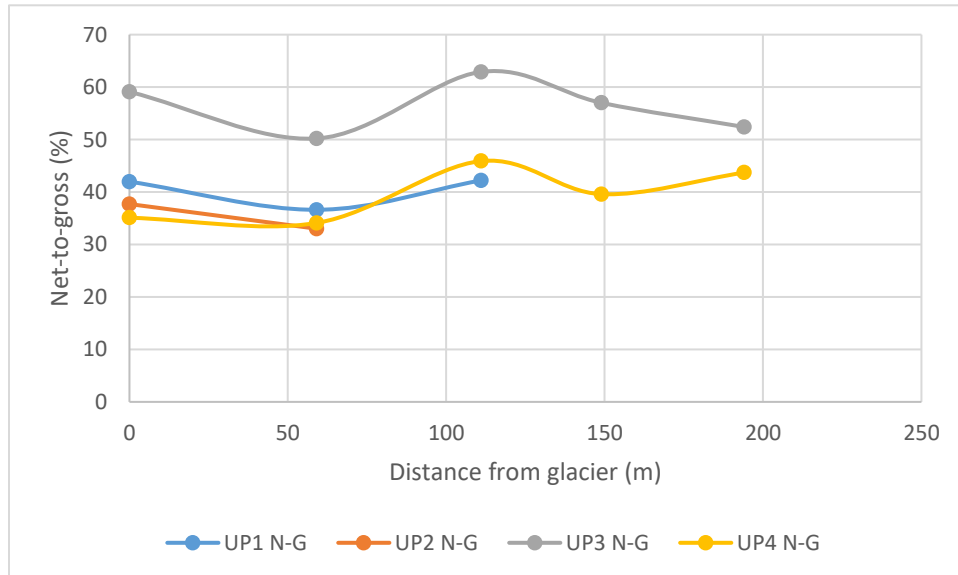


Figure 5.27: Lateral change in net-to-gross in the upper part of packages (UP1 to UP4) that extend from the glacier (left) across the northern outcrop (total lateral distance of 194 m). Note that UP2 extends laterally for 59 m then becomes truncated by a slide (see Fig. 5.10).

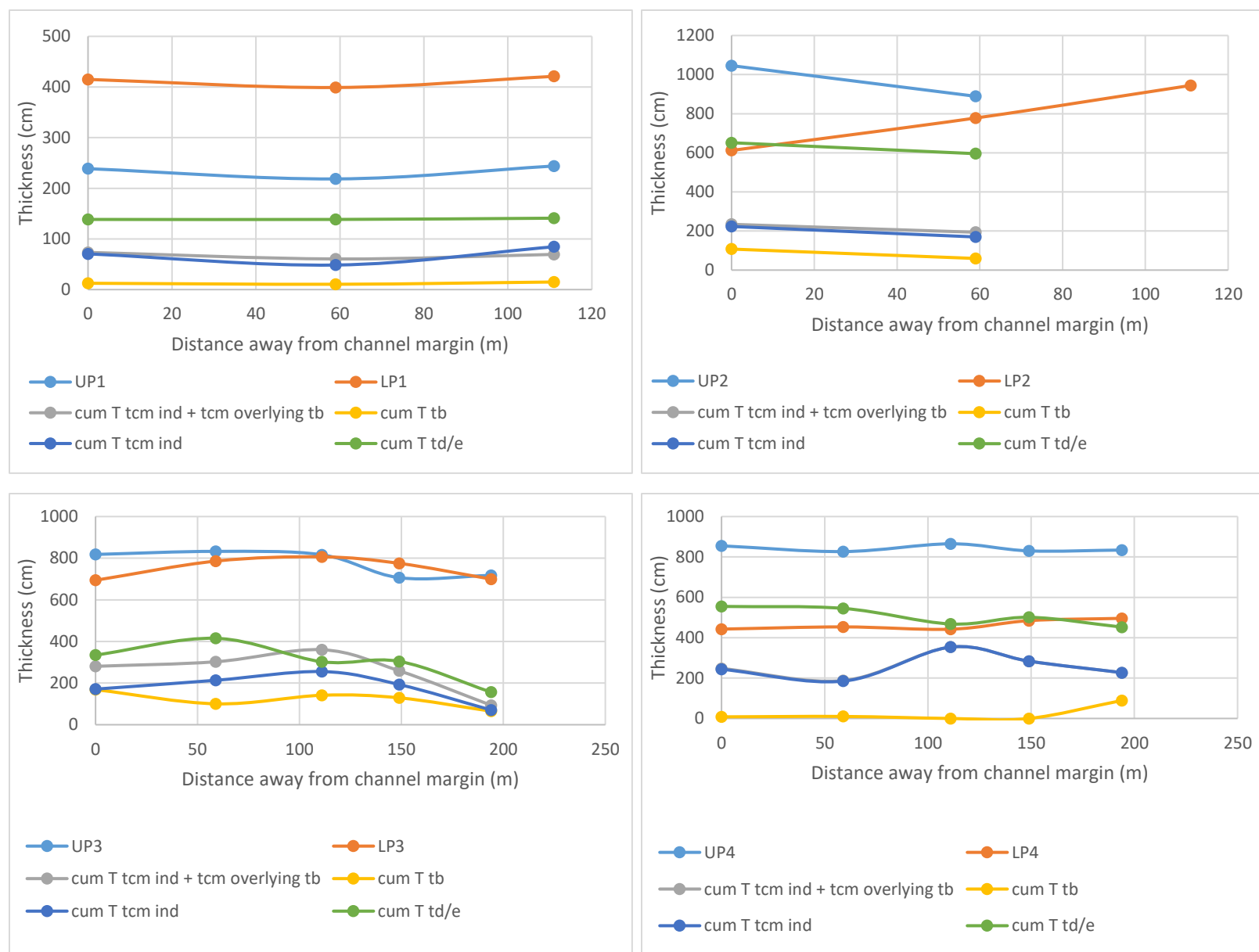


Figure 5.28: Lateral changes in thickness of the lower (orange) and upper (light blue) parts of packages extending across the northern outcrop for packages 1 to 4. Note the antithetic thickening and thinning patterns of the lower and upper parts of each package. Cumulative thickness of Tb turbidites (yellow), individual multiple set Tc turbidites (dark blue), individual multiple set Tc turbidites and multiple set Tc turbidites overlying Tb turbidites (grey), and Td/e turbidites (green). Note that it is the Td/e turbidites, and, to a lesser extent, the multiple set Tc turbidites that control the antithetic thickening and thinning of the upper parts.

Levee Packages 3 and 4: southern outcrop

Packages 3 and 4 (P3 and P4) are exposed across the moraine from their genetically related channels, CH3 and CH4, respectively, in the southeast drainage. Here they crop out in the southern outcrop over a distance of ~382 m (Fig. 5.10).

The lower part of P3 and P4 (LP3 and LP4, respectively) thicken and then thin laterally away from the margins of CH3 and CH4 across the southern outcrop (Fig. 5.16). Like in LP1 and LP2, it is the medium- to thick-bedded, coarser-grained Tb/Tbc turbidites that exhibit this trend (Fig. 5.13B), and collectively control the thickness trends in the lower part of each package. In LP3 and LP4 individual beds thicken over the first 560 m and 604 m, causing the thickness of LP3 and LP4 to increase by 1.21 m and 1.91 m, respectively. Upon reaching their maximum thickness, beds then thin, and the total thickness of LP3 and LP4 decreases by 1.27 m over 276 m and 2.87 m over 233 m, respectively (Fig. 5.16).

In the southern outcrop, the lithological makeup of LP3 and LP4 changes little upward. Here LP3 and LP4 consist mostly of thick-bedded, upper medium- and coarse-grained Tb/Tbc turbidites that average 41 cm in thickness. These strata are intercalated with very thin-bedded, fine- and lower medium-grained Tc turbidites that average 1.2 cm thick (single set) and 5.5 cm thick (multiple (2-3) set) and with medium-bedded, coarse-grained, medium-scale cross stratified turbidites that average 25 cm thick. Almost all beds are overlain by a siltstone/mudstone Td/e cap, except where removed by erosion, which on average is 12 cm thick. The lower division turbidites thicken to an average of 55 cm thick laterally over the first 560 m and 604 m away from the channels until LP3 and LP4 reach their maximum thicknesses, respectively. Upon reaching their maximum thickness these beds then thin to an average of 34 cm thick. The grain size of these lower division turbidites changes little laterally across the southern outcrop. Intercalated Tc turbidites change little in grain size or thickness across the width of the southern outcrop. In addition, intercalated medium-bedded dunes repetitively thicken and thin across the outcrop, however they remain upper medium- or coarse-grained. The net-to-gross averages 59 % proximal to the moraine, increases to an average of 67 % where the lower part of packages achieve their

maximum thickness, and then decreases as they thin to an average of 57 % adjacent to the glacier (Fig. 5.17).

Strata forming the upper parts of P3 and P4 (UP3 and UP4) thin as underlying LP3 and LP4 thicken laterally away from the margins of CH3 and CH4. Conversely, they then thicken as LP3 and LP4 thin after reaching their maximum thicknesses (Fig. 5.18, 5.19).

In the southern outcrop (~382 m) UP3 and UP4 consist mostly of very thin- and thin-bedded, fine-grained Tc turbidites that average 1.0 cm (single set) and 4.2 cm (multiple (3-4) set) thick, respectively. These strata are overlain by siltstone/mudstone Td/e caps that average 9 cm thick. These upper division turbidites are intercalated with medium-bedded, upper medium-grained Tbc turbidites that average 24 cm thick and with medium-bedded, coarse-grained, medium-scale cross stratified turbidites that average 14 cm thick. The cumulative thickness of multiple (3-4) set Tc turbidites and, significantly, of Td/e caps decreases as LP3 and LP4 thicken to their maximum, beyond which they then increase in thickness (Fig. 5.21). In addition, the grain size of these strata changes little across the width of the outcrop. In comparison, the cumulative thickness of single set Tc turbidites and medium-bedded Tbc turbidites remains similar across the outcrop, as does their grain size (Fig. 5.21). Intercalated medium-bedded dunes repetitively thicken and thin across the outcrop, however they remain consistently upper medium- or coarse-grained sandstone. Average net-to-gross of UP3 is 47 % and changes little laterally across the outcrop. UP4 is significantly less sand-rich than UP3 and has an average net-to-gross of 30 % that changes little laterally across the outcrop (Fig. 5.20).

Levee Packages 3 and 4: northern outcrop

P3 and P4 in the southern outcrop extend across the glacier (~819 m wide) to the northern outcrop, where they are exposed over a lateral distance of ~194 m. Here the thickness of LP3 and

LP4 changes little across the outcrop. The quality of the outcrop is excellent with almost 100% exposure. As a result, extra logs spaced 5-20 m apart were measured to document lateral changes in individual beds in both the lower and upper parts of each package (Fig. 5.10).

Like in the southern outcrop, the stratal makeup of LP3 and LP4 changes little vertically. Here LP3 and LP4 consist mostly of thick-bedded, coarse- and very coarse-grained Tb/Tbc turbidites that average 68 cm thick. These strata are intercalated with very thin-bedded, fine- and lower medium-grained Tc turbidites that average 1.3 cm thick (single set) and 7.1 cm thick (multiple (3-4) set). Almost all beds are overlain by siltstone/mudstone Td/e caps, except where removed by erosion, which average 11.1 cm thick. About half of the lower division turbidites that make up the LPs thin laterally towards the glacier while the other half either thicken or remain similar in thickness (Fig. 5.23). In addition, some of the beds that thin towards the glacier also fine. Average net-to-gross is 77 % and changes little laterally across the outcrop (Fig. 5.24).

In the northern outcrop, UP3 and UP4 change little in thickness laterally across the outcrop (Fig. 5.25). Like in the southern outcrop, UP3 and UP4 consist principally of very thin- and thin-bedded, fine-grained Tc turbidites that average 1.0 cm (single set) and 5.6 cm (multiple (3-4) set) thick. These strata are overlain by siltstone/mudstone Td/e caps that average 5.7 cm thick. These upper division turbidites are intercalated with medium-bedded, upper medium-grained Tbc turbidites that average 35 cm thick. Upper division turbidites in the UPs remain similar in thickness and grain size across the width of the outcrop (Fig. 5.29). Average net-to-gross is 48 % and changes little laterally across the outcrop (Fig. 5.27).

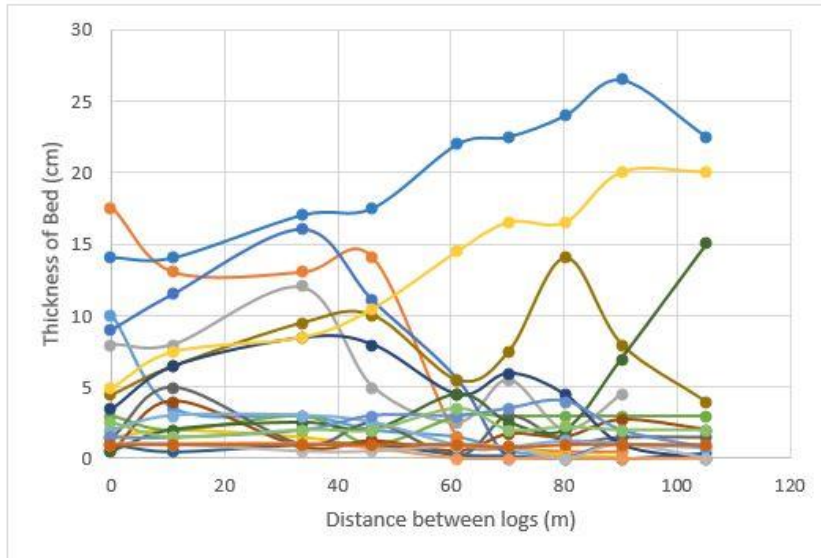
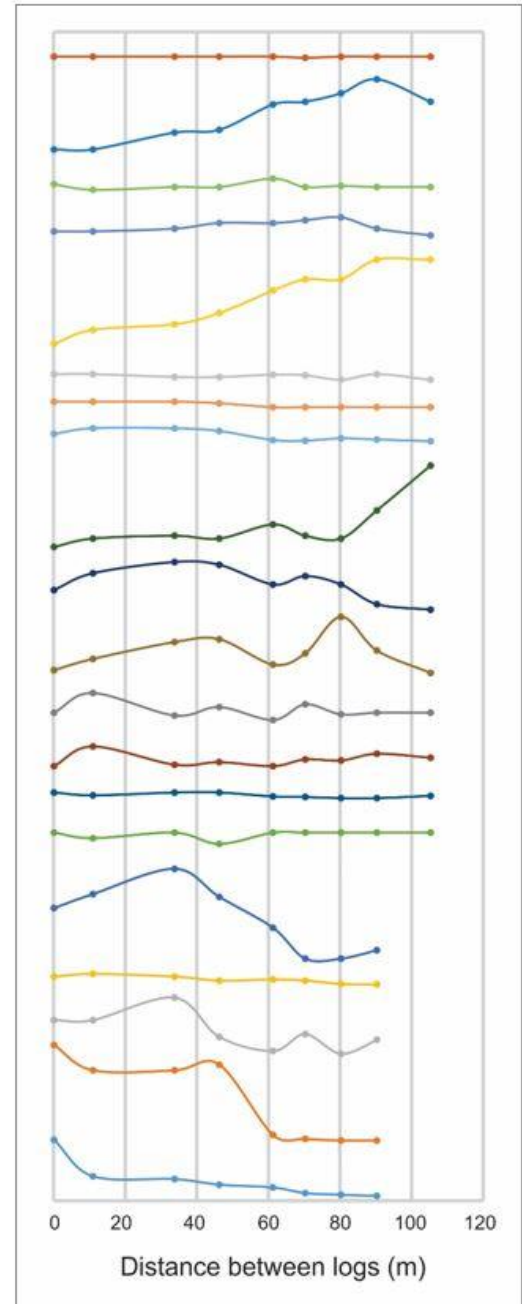


Figure 5.29: Thickness of upper division turbidites from UP4 in the northern outcrop. Graph on left shows how their thickness changes towards the glacier (towards the left). Graph on right spaces them out in order to better show the thickness trends. About half of the thin-bedded, upper division turbidites remain similar in thickness, only varying by 1-2 cm in thickness. The other half of upper division turbidites rhythmically thicken and thin by as much as 10 cm. Interstratified medium-bedded, upper division turbidites also rhythmically thicken and thin. Note that in general, beds do not thicken or thin linearly (smoothly).



5.2.3 Interpretation of Vertical Packaging in Type 1 Levee Deposits

Vertical trends in levee deposits are related to characteristics of the overspilling flows, including flow thickness, velocity, and vertical sediment concentration profile, in addition to levee relief (i.e., distance from channel floor to levee crest) (Kane et al., 2007; Khan, 2011) and channel

curvature (Morris et al., 2014). The vertical packaging observed in type 1 levee deposits of this study is interpreted to be related to recurring, systematic changes in these parameters.

Recent experimental work by de Leeuw et al. (2016) showed that as an initially unconfined turbidity current travels down a featureless slope it deposits two subparallel ridges, and between which the axis of the flow bypasses (Fig. 5.30). This depositional pattern occurs due to spatial differences in sediment transport/deposition, specifically lower rates of deposition within the axis compared to the margins of the flow (de Leeuw et al., 2016). The sharp contact at the base of each package, overlain then by coarse-grained turbidites, is interpreted to mark the base of the incipient levees (initial ridges) that formed during channel inception and helped to confine the axial parts of subsequent flows that travelled through the now better-defined channel (Fig. 5.31).

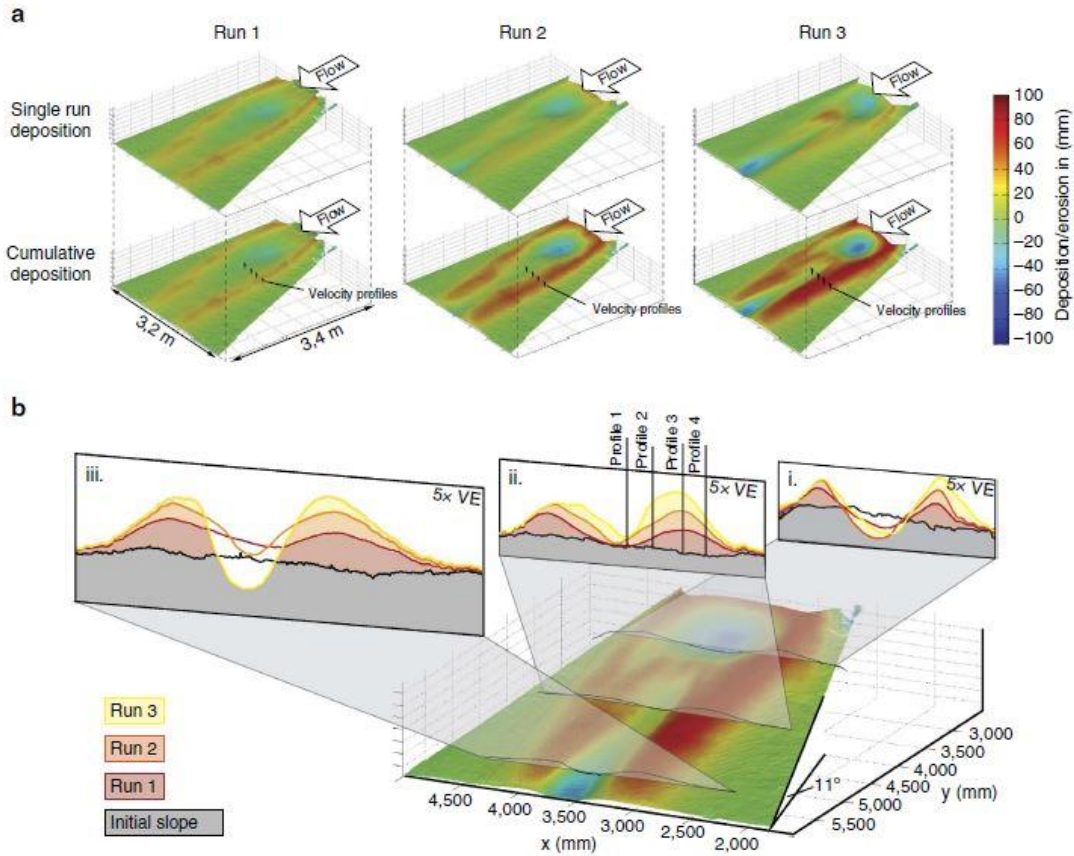


Figure 5.30: (A) deLeeuw et al. (2016) successively released three similar turbidity currents down an initially featureless slope (run 1-3). Digital elevation models of the deposits formed by these currents show that the first turbidity deposited two subparallel ridges between which the axis subsequent flows bypassed. This depositional pattern occurs due to spatial differences in sediment transport/deposition, specifically lower rates of deposition within the axis compared to the margins of the flow. Colours on the models indicate the thickness of the deposits or the depth of erosion. The amount of erosion/deposition after each individual run is shown (top) as well as the cumulative erosion/deposition. (B) Digital elevation model of the deposits after all three runs. Cross-sections show the amount of erosion/deposition on the lower, middle and upper slope. Profiles 1-4 shown on cross-section ii indicate the position of the profilers used to measure downstream velocity. From deLeeuw et al., 2016.

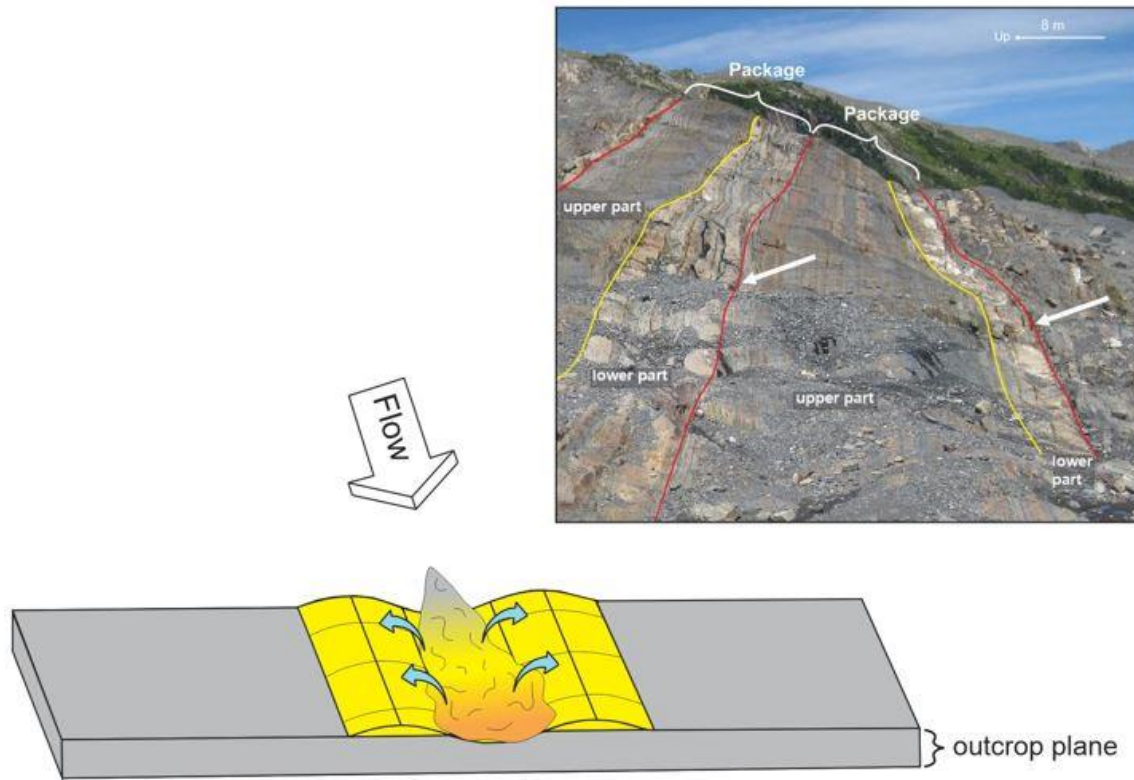


Figure 5.31: The sharp contact at the base of each package, overlain then by coarse-grained turbidites, is interpreted to mark the base of the incipient levees (initial ridges) that formed during channel inception and helped to confine the axial parts of subsequent flows that travelled through the now better-defined channel. White arrows in the photograph point to this sharp contact (highlighted by red lines) at the base of levee packages.

Initially the lower, coarse-grained, dense part of flows was able to overtop the aggrading levees and deposited thick-bedded, coarse-grained, lower division turbidites that built up the lower part of packages. To overspill, the velocity maximum in the dense lower part of the average channelized flows was necessarily above the height of the aggrading levee crests. Intercalated thinner-bedded, finer-grained, upper division turbidites reflect less energetic, smaller magnitude channelized flows whose coarse basal part remained confined to the channel and only the upper, fine-grained part overspilled.

With time the levees continued to aggrade and conditions in the channel continued to be dominated by flow bypass with negligible net sedimentation. Eventually relief between the channel

floor to levee crest exceeded the height of the velocity maximum (de Leeuw et al., 2016) after which only the upper, fine-grained parts of currents overspilled while the lower, coarse-grained parts remained confined to the channel. The sharp contact between the lower and upper parts of each package is interpreted to mark this condition followed by deposition of mostly fine-grained, thin-bedded turbidites intercalated increasingly less commonly upward with thicker, coarser turbidites deposited by uncommon, anomalously large flows. Levee package 1 is interpreted to have formed this way during the development of channel 1, levee package 2 during the development of channel 2, levee package 3 during the development of channel unit 3, and levee package 4 during the development of channel unit 4.

The velocity maximum, which here is interpreted to control the vertical packaging of levee deposits, is more fundamentally controlled by the vertical density profile of the channelized flows. Experimental and theoretical work suggests that turbidity currents consist of two end-member vertical density profiles that are controlled by grain size and grain sorting: 1) a plug-like structure where density changes little upward but then decreases sharply, and 2) an exponential structure with density decreasing continuously upward (Fig. 5.32) (Tilston et al., 2015; Kneller et al., 2016). The nature of the interface between the current and the overlying ambient fluid is controlled by the ratio of the buoyancy flux (measure of the resistance to mixing) to velocity shear forces (source of mixing), which can be described by a Richardson gradient number (Ri) (Tilston, 2017):

$$Ri = -\frac{g'}{\rho} \cdot \frac{\partial \rho / \partial z}{(\partial u / \partial z)^2}$$

where ρ is the density, u is the velocity and $g' = g(\rho_1 - \rho_2)/\rho$ is the reduced gravity. The lower part of levee packages are composed almost exclusively of coarse-grained, lower division turbidites suggesting the channelized flows were similarly well sorted and coarse grained. In a

coarse, well-sorted, non-depositional turbidity current, particles would have similar settling characteristics, and as a consequence would interfere with particle settling and promote the development of a plug-like density structure with negligible vertical particle stratification. In these flows the height of the velocity maximum is about half the flow thickness (Tilston et al., 2015) and the density gradient so steep ($Ri \geq 1$) that interfacial mixing would be inhibited and momentum maximized, thereby promoting flow bypass within the channel (Fig. 5.33). As noted above, once the height of the velocity maximum was exceeded by the height of the levee crest only the much finer grained sediment above the density interface overspilled from average channellized flows (Fig. 5.34).

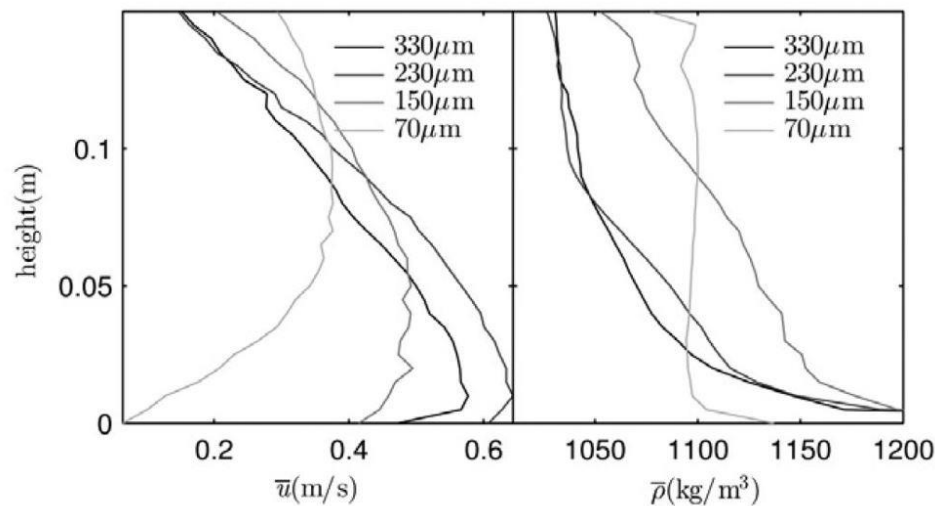


Figure 5.32: The velocity maximum is controlled by the vertical density profile of the flow. Experimental work by Tilston et al. (2015) suggests that turbidity currents consist of two end-member vertical density profiles controlled by grain size: 1) a plug-like structure where density changes little upward but then decreases sharply, and 2) an exponential structure with density decreasing continuously upward. Above are time-averaged velocity (u) and concentration (ρ) profiles for four grain sizes at a sediment concentration of 17.5% by mass. The finer grained flows have a plug-like density structure with a corresponding velocity maximum at about half the flow thickness. On the other hand, the coarser grained flows have an exponential density structure with a corresponding velocity maximum at about 20% of the flow thickness.

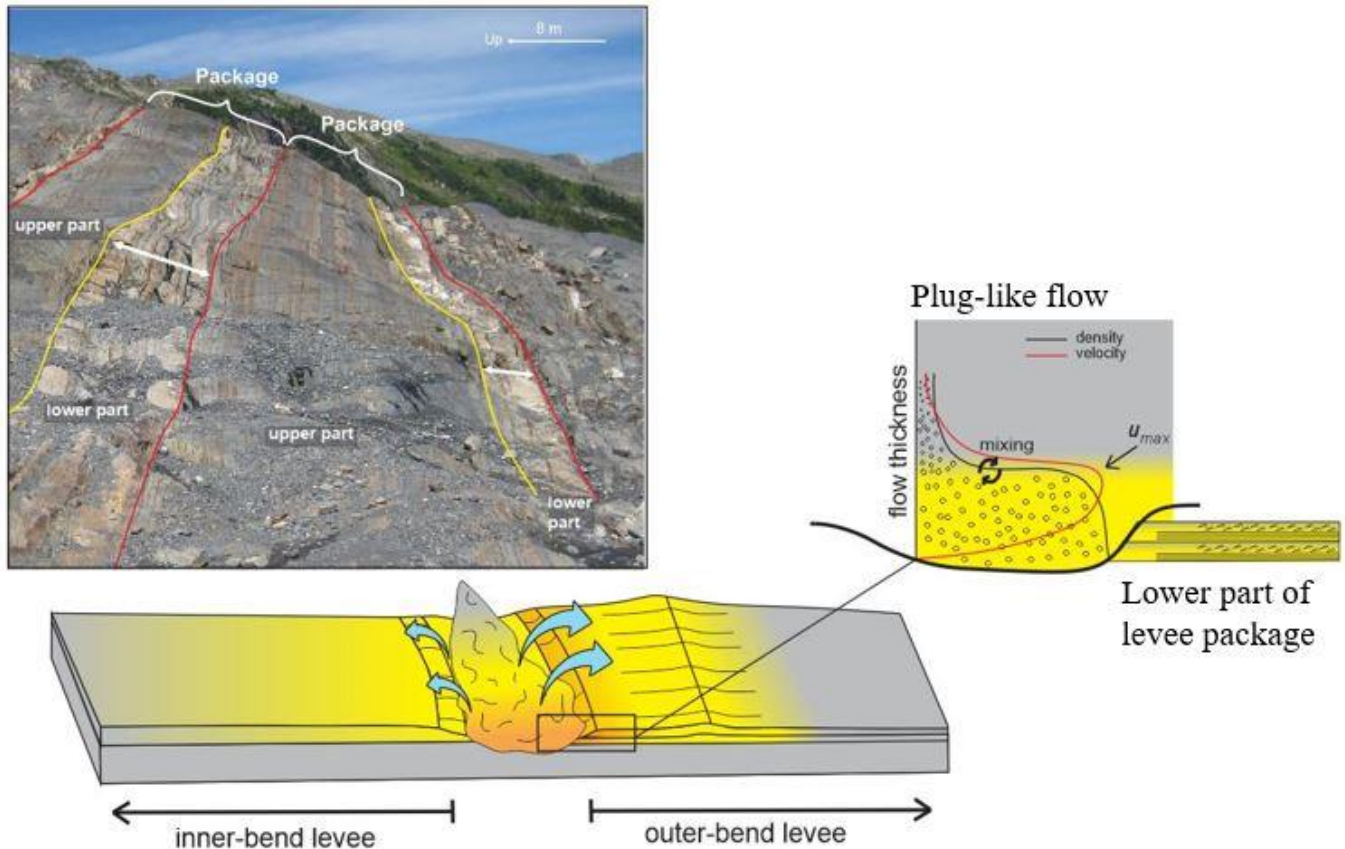


Figure 5.33: The almost exclusively coarse-grained make-up of beds in the lower parts of packages suggest that channelized flows were similarly coarse grained and well sorted. These flows would have had a plug-like density structure with negligible vertical stratification and high flow efficiency. Also, the height of the velocity maximum would have been at roughly half the depth of the flow. The velocity maximum of the channelized flows occurred above the height of the incipient channel margins allowing the lower, coarse-grained, dense part of flows to overspill and deposit thick-bedded, coarse-grained turbidites in the lower part of each package. White arrows in the photograph point to the lower parts of packages.

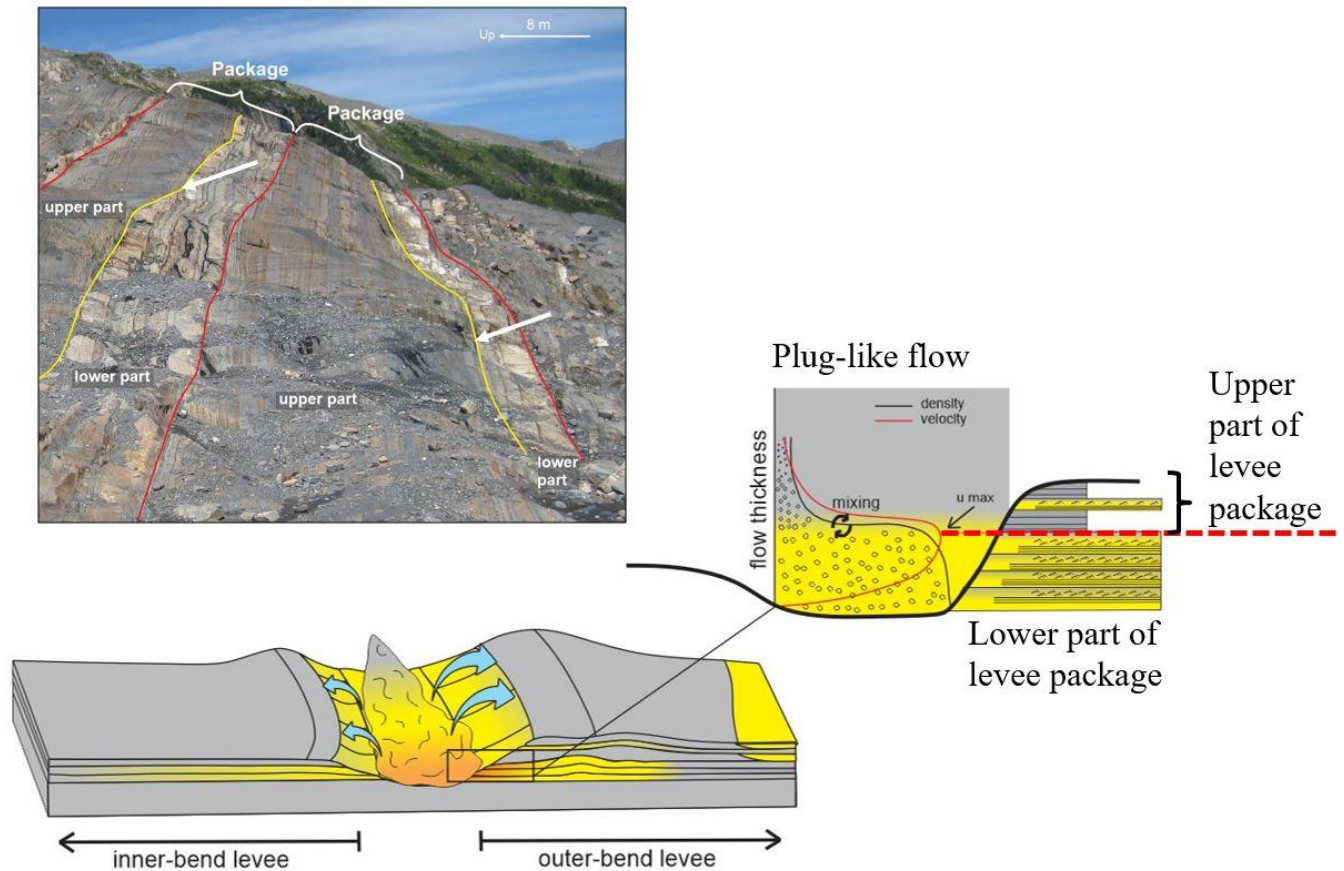


Figure 5.34: With time the levee continued to aggrade and conditions in the channel continued to be dominated by flow bypass with negligible net sedimentation. Relief between the channel floor to levee crest progressively increased. Eventually relief between the channel floor to levee crest exceeded the height of the velocity maximum of the average channelized flows after which only the upper, fine-grained parts of currents typically overspilled, building up the upper parts, while the lower, coarse-grained parts remained confined to the channel. White arrows point to the sharp contact between the lower and upper parts of the packages marks this condition (highlighted by yellow lines).

At some point during deposition of the upper parts of packages, the make-up of the sediment supply changed, specifically an increase in the sand fraction that caused throughgoing currents to become less well sorted and as a consequence more vertically stratified (large particles preferentially settled into the lower part of the flow). The resulting exponential density profile in the throughgoing currents significantly depressed the height of the velocity maximum to about 20% of the flow thickness (Tilston et al., 2015) -- an elevation now well below the height of the levee crest. The progressive upward thinning of thin-bedded, upper division turbidites and

decreasing abundance of thick-bedded, lower division turbidites within the upper parts of packages is interpreted to reflect the depression of the velocity maximum and overspill of the even more dilute, fine-grained upper part of the flows. Significantly, however, the exponential density profile would have promoted interfacial mixing and rapid energy loss, and accordingly deposition within the channel (Fig. 5.35). These depositional conditions were then repeated with the formation of the next levee package, signalling the possible influence of recurring changes of relative sea level on the make-up of the sediment supply, and in turn, characteristics of continental slope sedimentation (see discussion).

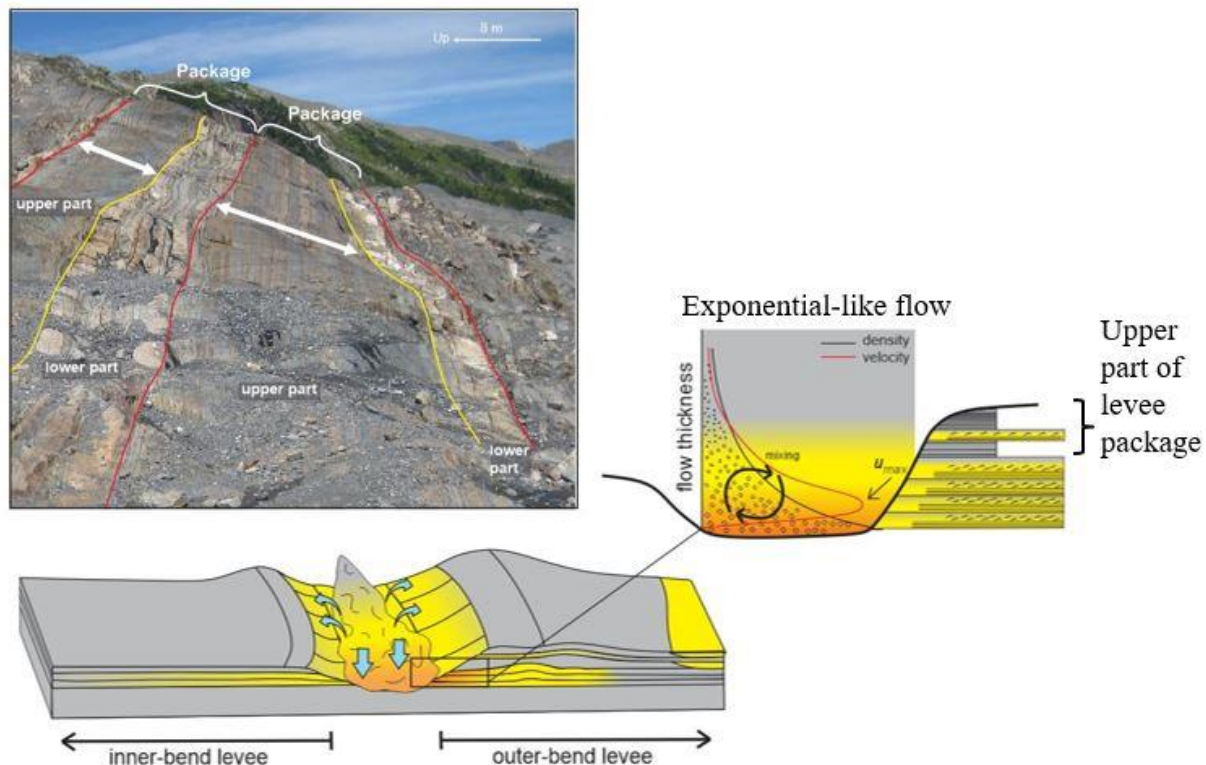


Figure 5.35: At some point during deposition of the upper part of the package (white arrows point to upper parts in photograph), the make-up of the sediment supply changed, specifically an increase in the sand fraction. The result was the development of an exponential density structure in throughgoing currents with a velocity maximum height significantly depressed to an elevation well below the height of the levee crest. The progressive upward thinning of thin-bedded, upper division turbidites and decreasing abundance of thick-bedded, lower division turbidites within the upper part of the package reflects this depression of the velocity maximum. The exponential density profile promoted interfacial mixing and rapid energy loss, and accordingly deposition within the channel.

5.2.4 Interpretation of Lateral Trends in Type 1 Levee Deposits

Levee deposits exposed in the southeast drainage, southern, and northern outcrops are interpreted to have been deposited along the outer bends of their genetically related channels. Owing to enhanced flow overspill (see section 3.4) outer-bend levee deposits tend to be thicker and more sand-rich compared to their inner-bend counterparts (Posamentier, 2003; Arnott, 2010; Khan and Arnott, 2011; Morris et al., 2014). Moreover, they generally consist of medium- to thick-bedded, medium- to coarse-grained Tbcde turbidites intercalated with thin-bedded, fine-grained Tcde turbidites and thick-bedded, coarse-grained Tade turbidites (e.g., Khan and Arnott 2011). Coarse-grained Ta turbidites are the least common and in this study are typically confined to channel-proximal areas in the southeast drainage, making them a useful diagnostic feature of channel-proximal levee deposition (Kane et al., 2007). Multiple (3-4) ripple cross-stratified set Tc turbidites are characteristic of levee deposits (e.g., Walker, 1985; Kane et al., 2007; Khan and Arnott 2011) and are abundant throughout the study areas. Paleocurrent measurements obtained from multiple set Tc turbidites indicate that the overspilling flows were generally directed towards the north-northwest.

Levees bordering deep-marine channels are typically reported to progressively thin away from the channel (Skene et al., 2002; Beauboeuf, 2004; Arnott, 2010). Less commonly, however, levee deposits are reported to initially thicken and then thin, for example in the Rosario Formation, Mexico (Kane et al., 2007) and in outer-bend levees in the Windermere Supergroup, Canada (Khan and Arnott, 2011; the adjacent inner-bend levees were reported to continuously thin away from the channel).

Experimental work has showed that as a channelized flow loses confinement the position of maximum sedimentation rate will vary depending on flow characteristics (Baas et al., 2004). In

high-concentration, high-energy flows, for example, the position occurs farther away from where the flow becomes unconfined compared to more dilute, less-energetic flows, which simply is a consequence of greater inertia (Baas et al., 2004; Khan and Arnott, 2011). The initial thickening in the lower part of packages in the southeast drainage and southern outcrops (Fig. 5.36) is interpreted to be related to the maximum sedimentation rate of the high-concentration, high-energy overspilling flows being displaced laterally from the channel margin (Khan and Arnott, 2011), which here is about 560-660 m away from the margins of their genetically-related channels (in the southeast drainage). Beyond the zone of maximum sedimentation, flows, which now have become depleted in much of their coarse-grained sediment, and hence flow density, begin to wane. Deposition becomes dominated by progressively finer, more slowly settling particles and the development of thinner, finer-grained turbidites, which at a larger scale causes the total thickness of the lower part of packages to thin laterally. In contrast, intercalated thin-bedded, fine-grained Tc turbidites within the lower part of packages change little in thickness or grain size laterally, which is interpreted to indicate sufficiently fine-grained overspilling flows in which particle settling and subsequent bed-load transport was more or less spatially uniform (Khan and Arnott, 2011).

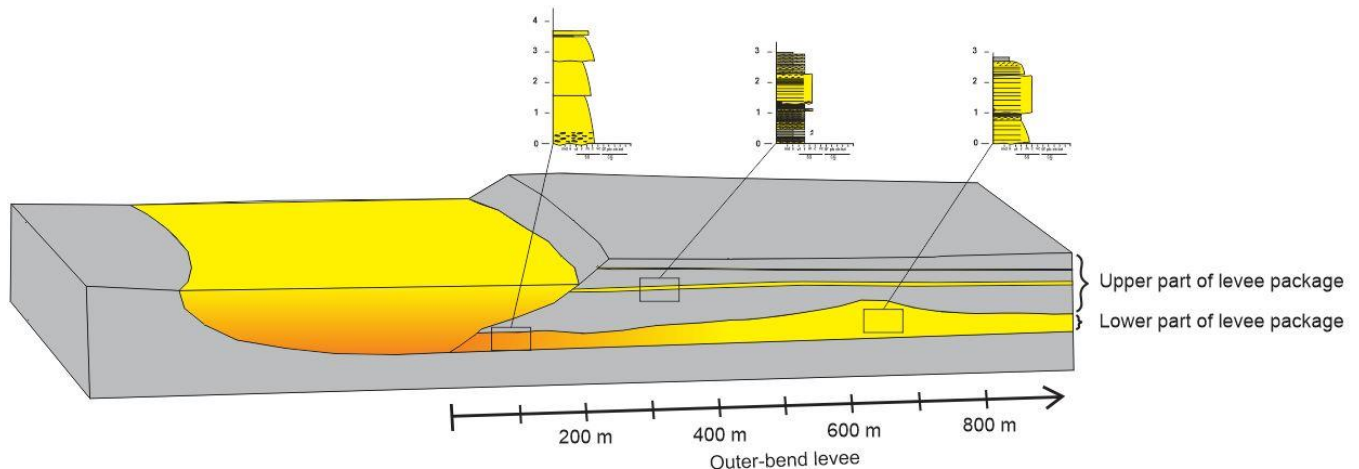


Figure 5.36: Schematic showing the initial thickening of a lower part of a package away from the channel margin across the southeast drainage and southern outcrops. This trend is interpreted to be related to the maximum sedimentation rate of the high-concentration, high-energy overspilling flows being displaced laterally from the channel margin (Khan and Arnott, 2011), which here is about 560-660 m away from the margin of the channel. Beyond the zone of maximum sedimentation, flows, which now have become depleted in much of their coarse-grained sediment, and hence flow density, begin to wane and the thickness of the lower part of the package thins.

In contrast to the southeast drainage and southern outcrops, the thickness of the lower part of packages changes little across the northern outcrop, with the exception of lower package 2 (LP2), which over 110 m thins by 3.3 m towards the glacier. It is important to note that where LP1 and LP2 crop out the exposure is only ~110 m wide before being truncated by a slide deposit (Fig. 5.10, see section 4.4.4) whereas LP3 and LP4 crop out over a total lateral distance of 194 m (Fig. 5.10). Nevertheless, if the lower part of packages did display thickening then thinning trends like those in the southeast drainage and southern outcrops, then maybe the exposure is simply too narrow to capture the changes. It is interpreted that the slide which truncates P1 and P2 formed by local collapse of the inner margin of a sinuous channel, a process which is commonly observed in modern (e.g., Gamberi and Rovere, 2011) and ancient (e.g., Kane et al., 2007) systems. In modern systems, the inner walls of sinuous channels are much steeper than the slope on the levee backside, and as a result are more prone to gravitational failure (Kane et al., 2007). In the Rosario Formation, for example, slides in the proximal levee are up to 100 m thick (Kane et al., 2007). This indicates

that the genetically-related sinuous channels of P1 and P2 were located to the northwest of the outcrop, which is supported by the thinning of LP2 in the opposite direction (i.e., towards southeast) (Fig. 5.37).

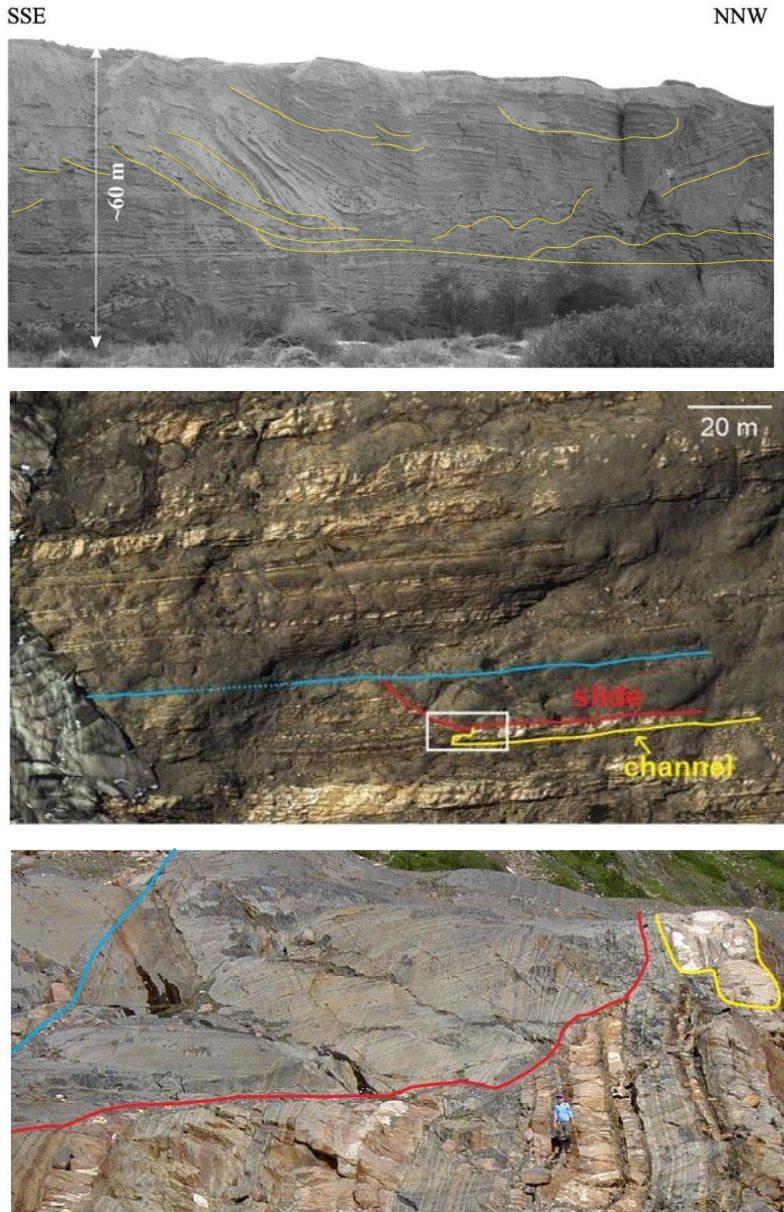


Figure 5.37: In modern systems, the inner walls of sinuous channels are much steeper than the slope on the levee backside, and as a result are more prone to gravitational failure. Top: Slump deposit in a channel-proximal location within the Upper Cretaceous Rosario Formation, Baja California, Mexico (modified from Kane et al., 2007). Some convoluted bedding contacts have been outlined in yellow. Middle: Aerial photograph showing the slide deposit (contact in red) that crops out in the northern outcrop of this study underlain by the channel fill deposit (contact in yellow) identified by Marion (2005). Blue contact marks the base of debrite 1 (D1). It is interpreted that this slide formed by local collapse of the inner margin of a sinuous channel indicating that the genetically-related sinuous channels of P1 and P2 were located to the northwest of the outcrop. Bottom: Photograph taken by Marion (2005) in the field of the area marked by the white rectangle in the aerial photograph. Slide contact outlined in red, channel in yellow and debrite D1 in blue.

The levee packages exposed in the southeast drainage and southern outcrops crop out adjacent to their genetically related channels. This is not the case for the LPs in the northern outcrop, which are sandwiched between the glacier to the southeast and moraine to the northwest

(Fig. 5.10). Here about half of the thick-bedded, coarser-grained, lower-division turbidites in the LPs thin laterally towards the glacier whereas the other half either thicken or change little in thickness (Fig. 5.23). Importantly, individual beds do not continuously thicken or thin as is often reported in the literature, but instead repetitively thicken and thin. A number of the beds that thin towards the glacier also fine. For example, a very thick-bedded, lower-division, very coarse-grained sandstone turbidite with dispersed quartz and carbonate pebble clasts thins and fines to a medium-bedded, upper medium-grained sandstone over a distance of 300 m toward the glacier (Fig. 5.38). Thick-bedded, lower-division turbidites that change little in grain size across the outcrop are typically coarse and upper-medium sand. Based on roughly half of the lower-division turbidites in the LPs thinning towards the glacier, along with a number of them also fining, it is interpreted that the genetically related outer channel bends to these proximal levees were located to the northwest of the northern outcrop. Moreover the slide deposit that truncates P1 and P2 is interpreted to represent collapse along the inner wall of a sinuous channel (see above), which supports the location of the channel to the northwest of the outcrop.

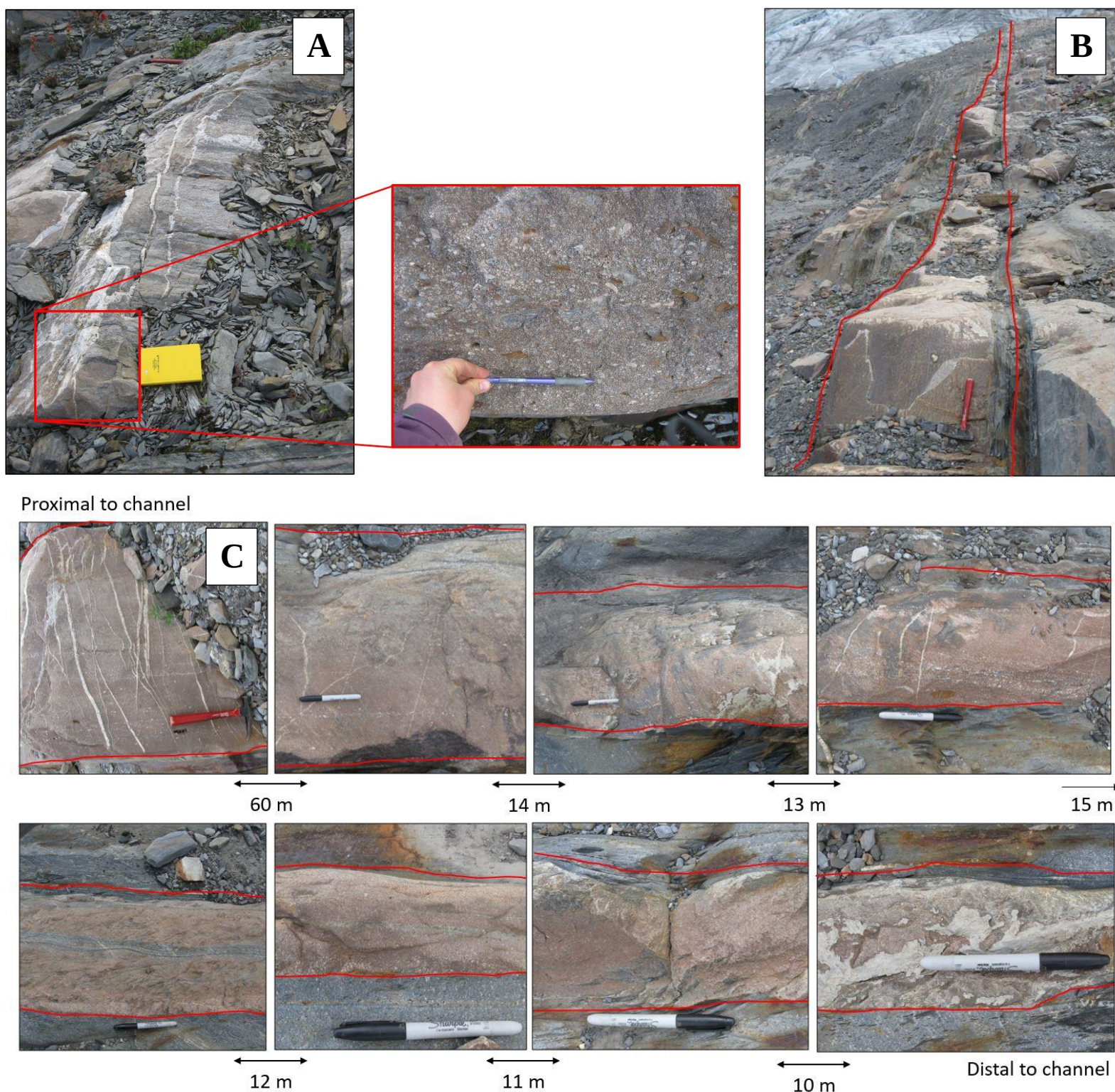


Figure 5.38: (A) Very thick-bedded, lower-division, very coarse-grained sandstone turbidite with dispersed quartz and carbonate pebble clasts thins and fines to a medium-bedded, upper medium-grained sandstone over a distance of 300 m toward the glacier. Photo on left shows the bed ~300 m from the glacier (proximal to channel) and on right is a close-up of the base showing the dispersed quartz and carbonate pebble clasts. (B) Photo of bed thinning towards the glacier ~190 m from the glacier. (C) Photos of this bed taken over the last 135 m towards the glacier show how it continues to thin and fine.

In contrast to the lower part of packages (LPs), the upper parts (UPs) are interpreted to have been deposited by less energetic, dilute flows that, after becoming unconfined, had more or less spatially uniform rates and characteristics of sediment transport and deposition. This resulted in deposition of mostly very thin- and thin-bedded, upper division turbidites, with coarser turbidites having been deposited by uncommon, anomalously large flows.

In the southern outcrop, the thickness of multiple (3-4) set Tcd/e turbidites in the UPs thin as the LPs thicken to their maximum, beyond which the Tcd/e turbidites then thicken as the LPs thin (Fig. 5.21). Although the Tc portion and overlying Td/e caps individually thicken and thin by only a few centimetres, collectively they act to level (i.e., heal) the underlying topography created by the thickening and thinning of the LPs – specifically, thickening or thinning to fill or flatten negative or positive topography, respectively. Interestingly, when comparing the cumulative thickness of the Tb, single set Tc, multiple set Tc, and Td/e portions of each turbidite that make up the UPs to the total thickness of the UPs, it is the upper Td/e part of turbidites that not only makes up most of the thickness of the UP, but changes most in thickness laterally (Fig. 5.21). The multiple set Tc turbidites also exhibit antithetic thickening and thinning, but to a much lesser extent. The thickness of the Tb turbidites and single set Tc turbidites, on the other hand, show little change across the outcrop suggesting that depositional patterns were little affected by seafloor topography. Systematic lateral changes in the thickness of Td/e and multiple set Tc turbidites, therefore, indicate that as flows moved into topographic lows and expanded, sedimentation rates increased and slightly thicker beds were deposited, whereas slightly thinner beds were deposited over topographical highs where flows accelerated.

In the northern outcrop, there is little topography created by the LPs. However, just like in the southern outcrop, it is the cumulative thickness of the Td/e turbidites that control the thickness

of the UPs, and to a much lesser extent the cumulative thickness of the multiple set Tc turbidites (Fig. 5.28).

5.2.5 Type 2 Levee Deposits: Description

Directly overlying the first mass transport deposit in the study area is a second type of levee unit that extends laterally across the southern outcrop (~382 m wide) and can be correlated across the glacier into the northern outcrop where it extends laterally over ~194 m (Fig. 5.10). The lithological makeup of this unit differs from both the lower (LP1-LP4) and upper parts of packages (UP1-UP4) of the previously described type 1 proximal levee deposits (see sections 5.2.1, 5.2.2), and is described here.

In the southern outcrop the unit thins as the underlying debrite thickens and then thickens as the debrite thins, ranging from 7.6 m to 11.4 m in thickness (Fig. 5.39). Strata consist mostly of medium-bedded, upper medium- and coarse-grained Tb/Tbc turbidites (average 27 cm thick); interbedded with less common medium-bedded, dune cross-stratified, coarse- and very coarse-grained sandstones (average 27 cm thick) and thick-bedded, very coarse-grained Ta turbidites (average 39 cm thick). Elongate mudstone intraclasts (average 5 cm x 1 cm) occur locally and in the basal part of beds are oriented subparallel to the base of the bed. Intercalated throughout this stratal succession are very thin-bedded, fine- and lower medium-grained Tc turbidites that average 1.3 cm (single set) and 3.9 cm thick (multiple (2-3) set). Almost all beds, irrespective of kind, are overlain by a siltstone/mudstone Td/e cap, except where scoured by the overlying bed, which on average is 12 cm thick. The cumulative thickness of Tb and Td/e turbidites, and to a lesser extent Ta turbidites, thicken and thin across the study area whereas it changes little for Tc turbidites and dunes (Fig. 5.39). Net-to-gross increases from 58% to 67% laterally across the outcrop towards the glacier (Fig. 5.40).

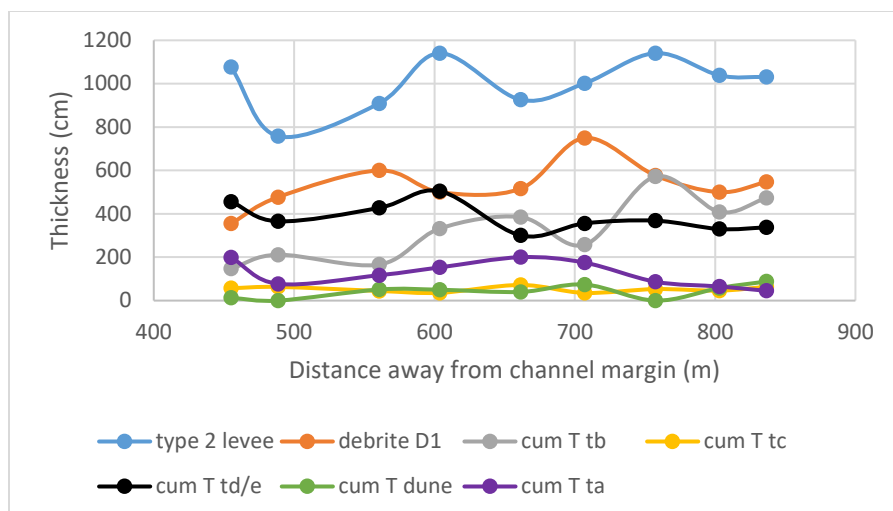


Figure 5.39: Lateral changes in thickness of the type 2 levee deposit (blue) and the underlying debris D1 (orange) extending across the southern outcrop. Note the antithetic thickening and thinning patterns of the levee and debris. Cumulative thickness of Ta turbidites (purple), Tb turbidites (grey), Tc turbidites (yellow), Td/e turbidites (black) and dunes (green) that make up the levee are shown. Note that it is the Tb and Td/e turbidites, and, to a lesser extent, the Ta turbidites that control the antithetic thickening and thinning of the levee. Tc turbidites and dunes change little in thickness.

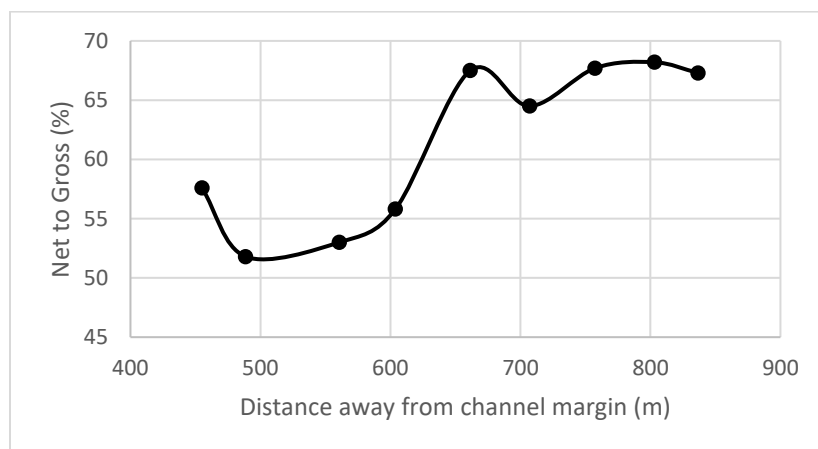


Figure 5.40: Lateral change in net-to-gross of the type 2 levee deposit extending across the southern outcrop.

In the northern outcrop, the unit changes little in thickness, ranging from 6.6 m to 7.9 m thick. Strata comprise thick-bedded, upper medium- and coarse-grained Tb/Tbc turbidites (average 41 cm thick) interbedded with very thin-bedded, fine- and lower medium-grained Tc turbidites (average 1.0 cm thick (single set) and 5.2 cm thick (multiple 2-3 set)) and less common medium-bedded, very coarse-grained Ta turbidites (average 25 cm thick). Like in the southern outcrop,

most beds are overlain by a siltstone/mudstone Td/e cap that averages 8.2 cm thick. The cumulative thickness of Ta, Tb, Tc and Td/e turbidites, in addition to the average net-to-gross (51%) (Fig. 5.41), changes little across the outcrop.

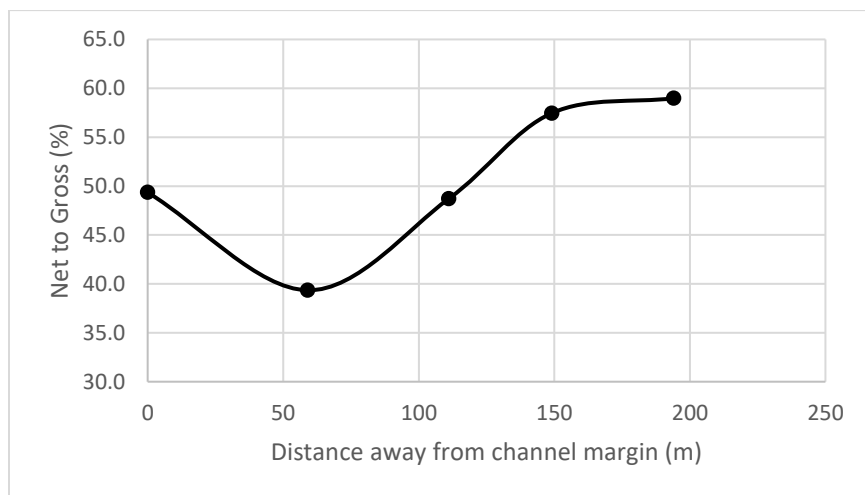


Figure 5.41: Lateral change in net-to-gross of the type 2 levee deposit extending across the northern outcrop.

5.2.6 Type 2 Levee Deposits: Interpretation

As mentioned, the lithology of this unit differs from the lower and upper parts of proximal type 1 levee packages previously described. In comparison to the lower part of other packages, lower division turbidites in this unit are generally thinner-bedded, coarser-grained and more commonly contain mudstone intraclasts. Also, and in the southern outcrop, dune cross-stratified sandstone is more common (Fig. 5.42). In terms of the upper part of other packages, medium- to thick-bedded, lower division turbidites are comparatively more abundant and coarser grained. Therefore, this unit is also interpreted to represent proximal levee deposits, but that experienced slightly different depositional conditions.



Figure 5.42: In the southern outcrop, dune cross-stratified sandstone is more common in type 2 levee deposits compared to type 1.

Mass transport deposits have been shown to control the distribution and deposition of overlying sediment-gravity flow deposits, for example in the Upper Cretaceous Tres Pasos Formation, Chile (Armitage et al., 2009), and in the Windermere Supergroup, Canada (Khan and Arnott, 2011). More specifically, their rugose surface can significantly influence the pathway of throughgoing turbidity currents, which in this case eventually stabilized with the development of channel 3 in the southeast drainage outcrop.

Compared to the proximal levee packages described previously (see sections 5.2.1, 5.2.2), strata that make up this unit contain more mudstone intraclasts, which most probably were sourced from erosion of the underlying mud-rich debrite – a source component apparently absent in the other levee packages. In 1985, Walker reported that levee strata commonly exhibit three attributes: climbing ripples, convolute lamination and mud rip-up clasts, terming them CCC-turbidites. Perhaps Walker's levee deposits were similar to those described here and are associated with an unexposed mud-rich mass transport deposit, or at the very least some other local source of mud.

Although rare in type 1 levee packages, dune cross-stratified, coarse- and very coarse-grained sandstone is comparatively common (Fig. 5.42). Like the ubiquitous ripple cross-stratification observed in all proximal levees packages, coarse-grained dunes formed by bed-load transport in unidirectional flows with a high concentration near-bed sediment layer that rapidly decayed upward. This then allowed for the development of the requisite hydrodynamic instability to form bed defects (Venditti et al., 2006; Arnott, 2012; Tilston et al., 2015) that then amplified into angular bed forms like ripples or dunes. The common occurrence of dunes, therefore, suggests that these conditions commonly developed when flow speeds were in the dune stability field.

In the southern outcrop, lateral changes in the thickness of Tb and Td/e turbidites, and to a lesser extent Ta turbidites, in the channel-bounding type 2 levee deposit acted to level (i.e., heal) the topography along the top of the underlying debrite – specifically, thickening or thinning to fill or flatten negative or positive topography, respectively (Fig. 5.39). Interestingly, when comparing the cumulative thickness of the Ta, Tb, Tc, and Td/e portions of each turbidite and the dunes that make up the unit in proportion to the total thickness of the unit, it is the Tb and Td/e parts of turbidites that not only make up most of the thickness of the unit, but show the greatest change in thickness laterally. Notably, Tc turbidites and dunes show little change in thickness across the outcrop suggesting that depositional patterns were little affected by the underlying topography. Systematic lateral changes in the thickness of Tb and Td/e turbidites, therefore, indicate that as flows transited areas of low topography they expanded, which elevated local sedimentation rates and deposited slightly thicker beds. Conversely, over areas of positive topography flows accelerated, sedimentation rates were reduced and thinner beds deposited. In the northern outcrop, the thickness of the debrite changes little, and, as a result, the thickness of overlying turbidites changes little.

5.2.7 Type 3 Levee Deposits: Description

Directly overlying the upper part of package 4 (UP4) in the northern outcrop (Fig. 5.10, 5.43) is a 12-14 m thick unit that differs from the upper part of type 1 proximal levee packages (UP1 to UP4) in a number of ways including: 1) thickness of very thin- and thin-bedded turbidites changes little stratigraphically upwards; 2) lower division turbidites are common throughout the unit; and, 3) net-to-gross is commonly about 10-15 % higher. This unit consists principally of very thin- and thin-bedded, fine-grained Tc turbidites that average 1.2 cm (single set) and 5.6 cm (multiple (3-4) set) thick, respectively. These upper division turbidites are intercalated with thick-bedded, upper medium-grained Tbc turbidites that average 60 cm thick. All strata are overlain by siltstone/mudstone Td/e caps that average 6.3 cm thick. The grain size of upper division turbidites remains similar across the width of the outcrop, as does their thickness. Similarly, the grain size of individual lower division turbidites changes little, but thickness either changes negligibly or thickens across the width of the outcrop. Lower division turbidites that do thicken do not do so in any preferred direction (Fig. 5.44). Average net-to-gross of this unit is 53% and changes little across the outcrop (Fig. 5.45).



Figure 5.43: A thick-bedded, upper medium-grained Tbcd/e turbidite marks the base of the type 3 levee deposit (yellow line). This deposit consists mainly of very thin- and thin-bedded, fine-grained Tc turbidites. It is overlain by a laterally-accreting channel deposit (white line, Dumouchel, 2015).

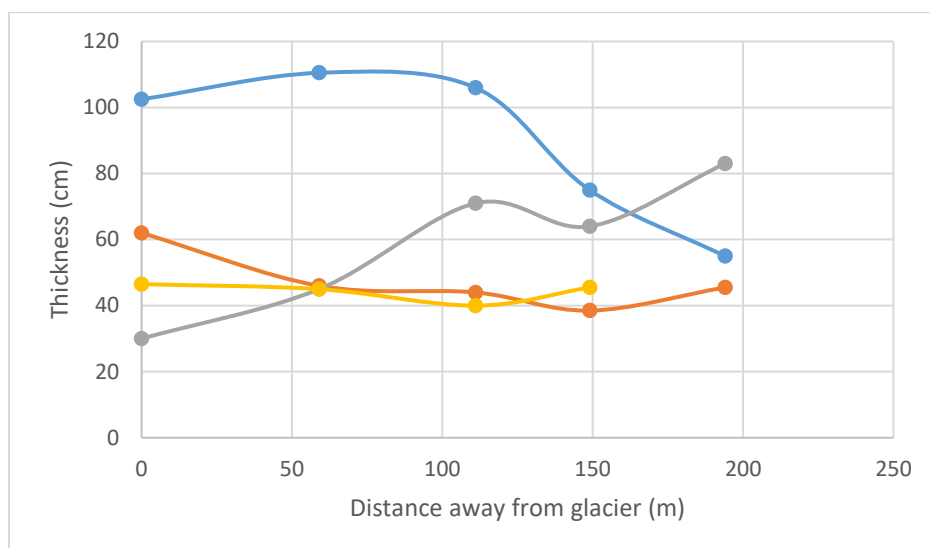


Figure 5.44: Graph showing how the thicknesses of four thick-bedded, upper medium-grained Tbc turbidites either change negligibly or thicken/thin across the width of the outcrop. Lower division turbidites that do thicken/thin do not do so in any preferred direction (blue, grey lines).

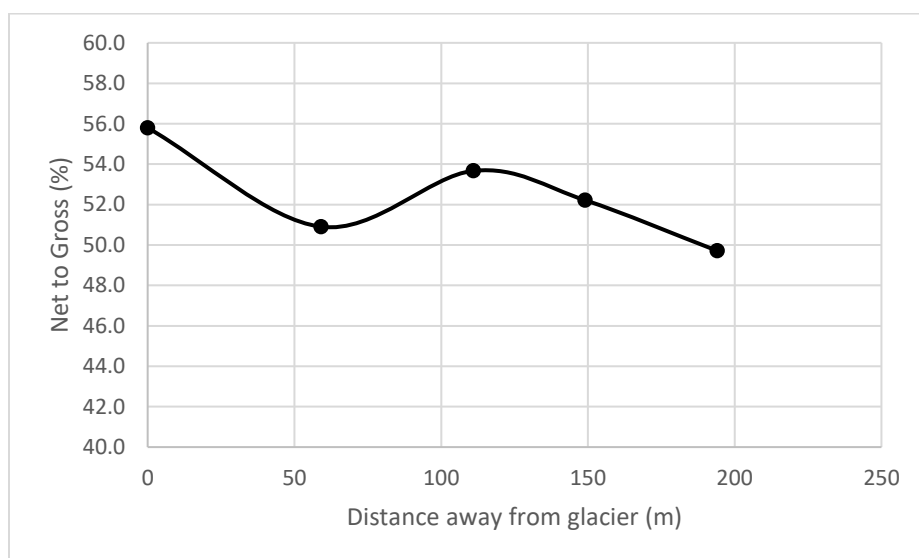


Figure 5.45: Lateral change in net-to-gross in the type 3 levee deposit extending from the glacier (left) across the northern outcrop (total lateral distance of 194 m). The range of net-to-gross (%) values are shown on the left vertical axis.

5.2.8 Type 3 Levee Deposits: Interpretation

In levee deposits, thick-bedded strata tend to thin rapidly away from the channel whereas thin-bedded strata show little lateral change (Kane et al., 2007; Khan and Arnott, 2011). As a result, distal levee deposits typically consist of thin-bedded, very fine- to fine-grained

sandstone/siltstone multiple (1-3) non-climbing ripple set turbidites (e.g., Khan and Arnott, 2011). The uppermost unit exposed in the northern outcrop, therefore, is interpreted to have been deposited distal to the outer-bend of its genetically related sinuous channel, which here is channel unit 5 exposed in the southeast drainage outcrop (Fig. 5.10). As overspilling flows moved away from the outer-bend they became progressively depleted of much of their coarse-grained sediment due to deposition over the proximal levee. In the distal levee then, deposition became dominated by finer, more slowly settling particles in which particle settling and subsequent bed-load transport was more or less spatially uniform (Khan and Arnott, 2011). These conditions are marked by the very thin- and thin-bedded, fine-grained, upper division turbidites that change little in thickness or grain size laterally (Fig. 5.46). Moreover the occurrence of discontinuous ripple trains that extend laterally for a few metres, or isolated ripples that have foundered into underlying silt/mud Td/e turbidites (Fig. 5.46) suggest conditions of limited sediment availability and low energy (Kane et al., 2007), which are consistent with a distal levee setting.

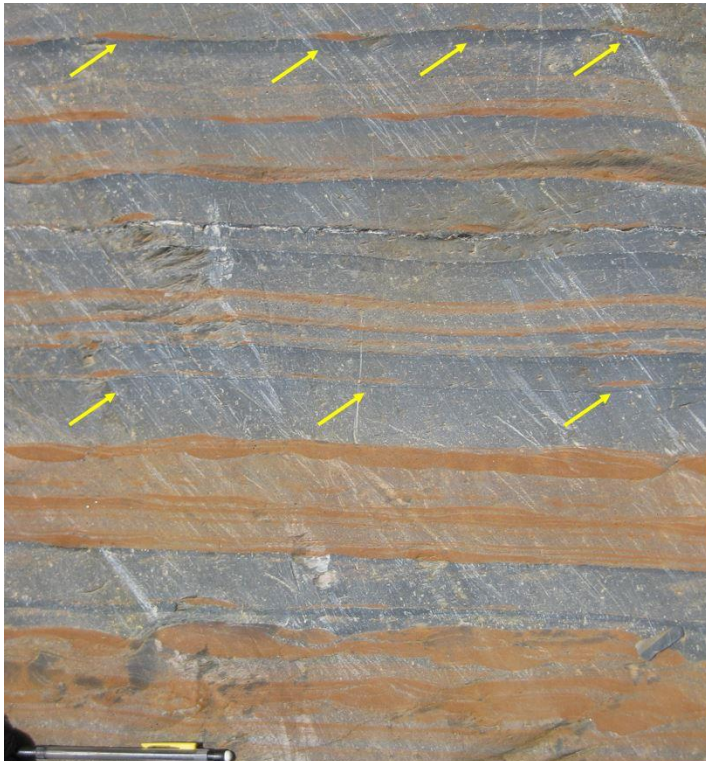


Figure 5.46: Deposition over the distal levee was dominated by finer, more slowly settling particles in which particle settling and subsequent bed-load transport was more or less spatially uniform (Khan and Arnott, 2011). These conditions are marked by the very thin- and thin-bedded, fine-grained, upper division turbidites that change little in thickness or grain size laterally. Yellow arrows point to discontinuous ripple trains which suggest conditions of limited sediment availability and low energy (Kane et al., 2007), consistent with a distal levee setting.

With time, continued overspill increased levee height, particularly in the proximal levee, which progressively increased the angle of the levee backslope (Khan and Arnott, 2011). Seismic profiles from the modern Amazon Fan, for example, show that backslopes can be as steep as 5°, which accordingly could cause overspilling flows to accelerate (Pirmez and Imran, 2003; Kane et al., 2010). Here, the intercalated thick-bedded, lower division turbidites in the uppermost unit are interpreted to have been deposited by coarse grained, overspilling flows that, because of the steepened slope, were able to bypass the proximal levee and subsequently deposit over the more shallowly inclined distal levee (Fig. 5.47). Such deposits were termed overbank splays by Khan and Arnott (2011). Similarly, finer-grained overspilling flows probably also bypassed the proximal levee, and, because of increased flow speed and increased interfacial mixing, would also have preferentially deposited over the distal levee as thin-bedded, upper division turbidites (Khan and Arnott, 2011). Over time, preferential deposition on the distal levee would have progressively reduced the slope on the proximal levee backside, and consequently reduced bypass. This, then, would have promoted deposition on the proximal levee, which in turn would have progressively steepened the levee backslope and once again encouraged bypass and deposition on the distal levee (Khan and Arnott, 2011).

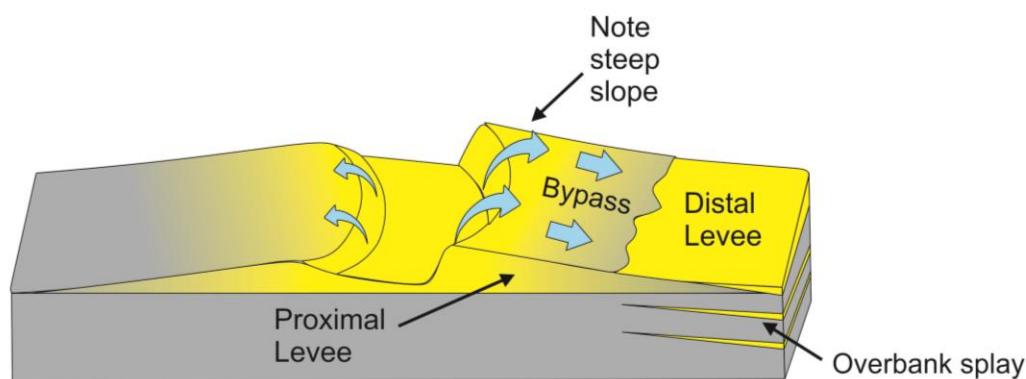


Figure 5.47: Schematic diagram showing coarse-grained overspilling flows that accelerate down the steep slope of the proximal levee and deposit thick-bedded, coarse-grained, lower division turbidites (yellow) over the more shallowly inclined distal levee. Such deposits were termed overbank splays by Khan and Arnott (2011). Finer-grained overspilling flows probably also bypassed the proximal levee but in this case deposited thin-bedded, fine-grained, upper division turbidites (grey). Figure modified from Davis, 2011.

6 Discussion and Conclusions

6.1 Temporal and Spatial Relationships between Channel and Levee Sedimentation

Five channel units crop out in the southeast drainage outcrop. These channel units are intercalated with five thick, easily identified mass transport deposits that in some cases extend across the entire study area (> 2.5 km), and therefore are used as stratigraphic markers to correlate between outcrops separated by moraine and/or the glacier. Levee deposits crop out adjacent to these channel units on their northern margins and extend over a lateral distance of up to 2 km; counterpart levees on the south side are not exposed (Fig. 5.10).

Generally, strata that make up the channel units exposed in the southeast drainage outcrop thin and fine upwards. Additionally, they also thin and fine laterally towards the northwest (see section 5.1). Levee deposits that crop out adjacent to them are divided vertically into 4 packages, each consisting of a lower and upper part (Fig. 5.11). Medium- to thick-bedded, lower division turbidites that make up the lower part of packages (Fig. 5.13) transition from Ta turbidites in areas adjacent to the channel units (southeast drainage outcrop) to Tbc turbidites moving laterally away from the channels (southern outcrop). These levee strata also fine laterally away from the channel units towards the northwest, in addition to displaying a consistent thickening then thinning trend (Fig. 5.16). Paleocurrent measurements obtained from levee strata indicate that the overspilling flows were generally directed towards the north-northwest.

In the northern outcrop, medium- to thick-bedded, lower division turbidites that make up the lower part of packages consist exclusively of Tbc turbidites. Moreover, the thickness and grain size in the lower part of packages generally change little laterally, with the exception of lower package 2 (LP2), which over 110 m thins by 3.3 m (9.4 m to 6.1 m) towards the glacier (Fig. 5.22). With that being said, about half of the lower division turbidites that make up the LPs thin

laterally towards the glacier (southeast) whereas the other half either thicken or remain of similar thickness (Fig. 5.23). In addition, some of the beds that thin also fine towards the glacier. For example, a bed in the lower part of package 3 exposed at the northwestern-most end of the outcrop changes progressively from a 139 cm thick, upper very coarse-grained Tb turbidite with abundant quartz pebbles and carbonate clasts to a 10 cm thick, upper medium-grained Tbc turbidite over a distance of ~300 m (Fig. 5.38). Paleocurrent measurements are also generally directed towards the north-northwest.

Once a flow over spills its channel margin, it loses confinement and expands, which increases its cross-sectional area and reduces unit energy. As a result flow velocity abruptly decreases, which in turn abruptly reduces transport capacity and competence and causes much sediment to be deposited adjacent to the channel margins --- the rate of deposition decreasing away from the channel (Kane et al., 2007; Arnott, 2010; Khan, 2012). Levee deposits in the lower parts of packages of this study consist of medium- to thick-bedded, upper medium- to coarse-grained turbidites, suggesting that they were deposited proximal to the margin of their genetically related channel. In addition, coarse-grained Ta turbidites typically crop out adjacent to channel deposits in the southeast drainage, and are reported to be useful indicators of channel-proximal levee deposition (Kane et al., 2007). Ta beds were deposited from high-density turbidity currents with their structureless nature indicating high rates of sedimentation (Arnott and Hand, 1989; Leclair and Arnott, 2006). This indicates that as flows overspilled the channel margins and became unconfined, flow capacity was exceeded resulting in rapid fallout from suspension and bed aggradation (Kane et al., 2007).

In Khan and Arnott's (2011) study of a sinuous deep-water channel-levee system, both the inner-bend and outer-bend margins of their channel units were exposed. They reported that

medium- and thick-bedded strata in their proximal outer-bend levees displayed a lateral thickening then thinning trend, similar to the one described in this study. They interpreted the trend to be related to the location of maximum sedimentation rate of high-concentration, high-energy overspilling flows, which because of flow inertia was displaced ~100 m beyond the outer-bend channel margin (see also Baas et al., 2004). In this study, the lower part of packages thicken over the first 560-660 m indicating that this is the position of maximum sedimentation rate; the greater distance might relate to the obliquity of the exposure relative to the channel margin. Beyond the zone of maximum sedimentation, flows, which now have become depleted in much of their coarse-grained sediment, and hence flow density, begin to rapidly wane. Deposition becomes dominated by progressively finer, more slowly settling particles and the development of thinner, finer-grained turbidites, which at a larger scale causes the total thickness of the lower part of packages to thin laterally (see section 5.2.4) (Khan and Arnott, 2011). Therefore, proximal levee deposits that crop out in the southeast drainage and southern outcrops are interpreted to have been deposited along the outer-bends of their genetically-related sinuous channels.

In contrast, determining on which side of the channel proximal levee deposits in the northern outcrop overspilled is more difficult as their related channel deposits are not exposed. However, near the base of the outcrop a 23 m thick slide crops out (see section 4.4.4) and is interpreted to be the result of failure along the typically steep inward-facing margin of a sinuous subaqueous channel – a feature commonly observed in both modern (e.g., Gamberi and Rovere, 2011) and ancient (e.g., Kane et al., 2007) deep-marine turbidite systems. It is likely that a portion of the inward-facing levee failed and formed an irregular slide scar. This depression was then filled, quite possibly during the same failure event, by a large, only slightly deformed block of thin-bedded, fine-grained, upper division turbidites interpreted to be mud-rich distal levee

deposits. Notably also is a channel unit that directly underlies the slide in the northern outcrop (Fig. 4.17) (Marion, 2005). Therefore, it is likely that the genetically-related but unexposed channels to the proximal levee deposits in the northern outcrop occur a short distance to the northwest. Approximately half of the lower division turbidites that make up the LPs thin and sometimes fine laterally towards the glacier (southeast), especially the extreme example of the thick-bedded, upper very-coarse grained sandstone with quartz pebbles described above (Fig. 5.38). This suggests that the nearby, but nevertheless unexposed channels were located to the northwest of the northern outcrop. The typical thickening then thinning trend displayed in outer-bend proximal levee deposits is not observed in the northern outcrop, possibly because the exposure is simply too narrow to capture the change. However, based on the coarse grain size that makes up many of the beds in the lower parts of packages these levees are also interpreted to have been deposited along the outer-bends of their unexposed genetically-related channels. Enhanced overspill processes along the outer-bend of a channel would likely have been required to allow such coarse sediment (up to pebbles) to escape the confines of the channel.

A meaningful modern analogue for understanding the three-dimensional architecture of the channel-levee system studied in this thesis is the modern Amazon turbidite system (Pirmez and Imran, 2003). Like the Windermere turbidite system, the Amazon Fan is a passive margin turbidite system of similar dimension (Ross and Arnott, 2007) and which recently has been the focus of extensive study (e.g., ODP Leg 155, Flood et al., 1995). In the Amazon Fan, average channel width (distance between levee crests) decreases rapidly from ~13 km in the canyon to an average of ~1.2-1.5 km in the sinuous leveed channels on the middle fan. Channel relief (distance from channel floor to levee crest) follows a similar trend, decreasing downslope from ~450 m to ~80 m on the middle fan. In general, the difference in relief between levees on the outer versus inner channel

bend ranges from 20-60 m on the upper fan and decreased to as low as 2-5 m on the lower fan (Pirmez and Imran, 2003). Generally, average channel wavelength decreases from 9 km on the upper fan to 3 km on the lower fan (Pirmez and Imran, 2003), but can be as low as 2 km (Wynn et al., 2007).

Based on grain size, bed thickness, sedimentary structures, and importantly, how these parameters change laterally away from the margins of the exposed channels in the southeast drainage outcrop and the unexposed channels to the northwest of the northern outcrop, the levee deposits of this study were interpreted to represent deposition on the outer-bend side of mid-fan sinuous channels (*sensu* Pirmez and Imran, 2003). Moreover, paleocurrents indicate that their formative flows were directed towards the north-northwest. Using the modern Amazon turbidite system as a three-dimensional analogue, a three dimensional reconstruction of the Windermere channel-levee complex can be developed from a two dimensional dataset (face of the exposed outcrop) (Fig. 6.2). Therefore, the 2.6 km distance separating outer-bend levee deposits in the southeast drainage with those in the northern study area is consistent in terms of lateral spacing between adjacent outer-bend deposits along a single sinuous channel in the Amazon system.

Additionally, it is interpreted that all five channels had a similar planform geometry and stacked vertically with a slight lateral offset pattern within a larger sinuous channel belt, and ultimately built up a single large-scale leveed-channel complex (Fig. 6.2) (e.g., Lien et al., 2003; Rongier et al., 2017). In the Ross Formation, Ireland, for example, four to five individual sinuous channels each ranging up to 100 m wide and 5-10 m thick stack to form channel belts that are 2.5-5 km wide and 25-35 m thick (Lien et al., 2003). A similar pattern of vertically nested channels is reported from seismic images of slope channels in the Nile Delta, Egypt (Fig. 6.1) (Samuel et al., 2003). In addition, these channels migrate laterally with successive channels laterally offset,

stacked at their channel bends. Individual channels typically range from 50-250 m wide and are of the order of 15 m thick, and stack to form composite units that range from 0.5-6 km wide and 50-200 m thick (Fig. 6.1) (Samuel et al., 2003).

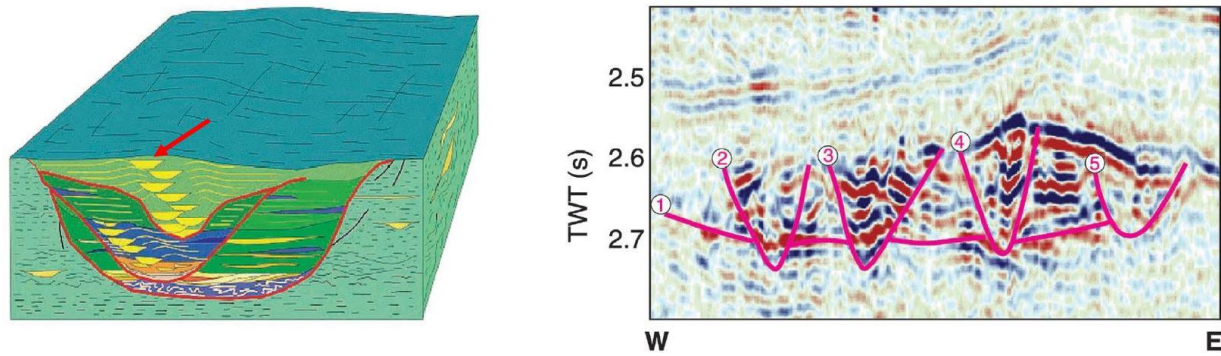


Figure 6.1: Block diagram (left) and seismic section (right) to show the development of channels in the Nile Delta, Egypt. Note how the channels in the block diagram (red arrow) stack vertically and are laterally offset. Also, they are aggradational and confined by their levee deposits. Seismic horizon 1 is interpreted as the base of the slope valley while seismic horizons 2-5 are interpreted to be channel-reincision surfaces. Modified from Samuel et al., 2003.

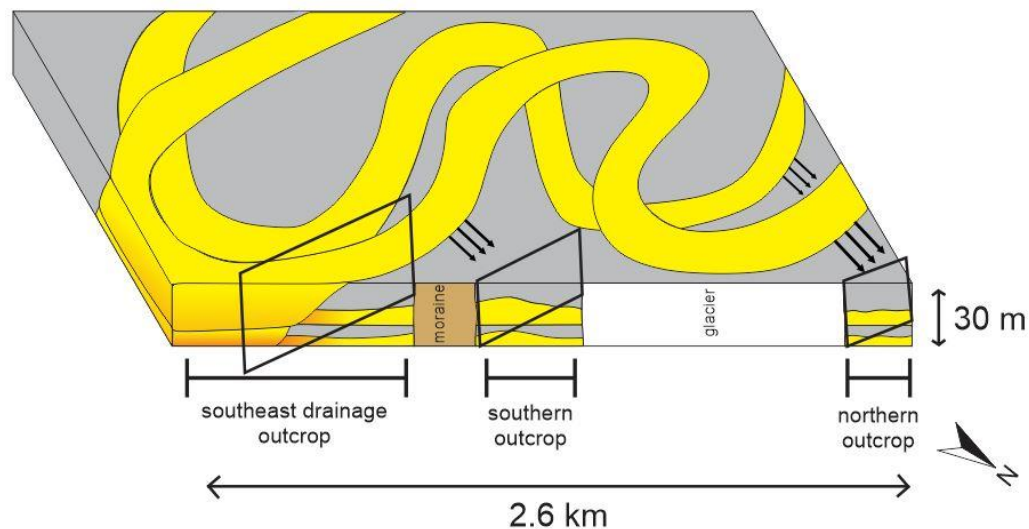


Figure 6.2: A three dimensional reconstruction of the two lowermost channel-levee packages in the Windermere channel-levee complex developed from a two dimensional dataset (face of the exposed outcrop). Black boxes indicate one possible face of the outcrop-- recall that the outcrop exposure is interpreted to be oriented oblique to the channel margins. The 2.6 km distance separating outer-bend levee deposits in the southeast drainage with those in the northern study area is consistent in terms of lateral spacing between adjacent outer-bend deposits along a single sinuous channel in the Amazon system. Moreover, it is interpreted that all five channels had a similar planform geometry and stacked vertically with a slight lateral offset pattern within a larger sinuous channel belt, and ultimately built up a single large-scale leveed-channel complex.

Initially as unconfined turbidity currents travel down a featureless slope, they build up two subparallel ridges between which flows then generally bypass. This depositional pattern occurs due to spatial differences in sediment transport/deposition, specifically lower rates of deposition within the axis compared to along the margins of the flow (Stage A Fig.6.3) (de Leeuw et al., 2016). In this study, the sharp contact at the base of each package overlain then by coarse-grained turbidites, is interpreted to mark the base of the incipient levee (initial ridges) that formed during channel inception and helped to confine the axial parts of subsequent flows that travelled through the now better-defined channel. As weakly-confined, coarse grained flows overspilled these initial ridges they deposited medium- and thick-bedded, upper medium- and coarse-grained lower division turbidites in the lower part of the proximal levee package that thickened over the first few hundred metres laterally away from the channel before thinning (Stage B Fig. 6.3). As these overspilling flows moved away from the outer-bend they became progressively depleted of much of their coarse-grained sediment due to deposition over the proximal levee. In the distal levee then, deposition became dominated by finer, more slowly settling particles in which particle settling and subsequent bed-load transport was more or less spatially uniform (Khan and Arnott, 2011). These conditions are marked by very thin- and thin-bedded, fine-grained, upper division turbidites that change little in thickness or grain size laterally.

With time the levee continued to aggrade and conditions in the channel continued to be dominated by flow bypass with negligible net sedimentation (Stage C Fig. 6.3). Relief between the channel floor to levee crest progressively increased as the lower part of the levee package built up to its maximum thickness, which was of the order of 3 to 10 m (undecompacked). Deposition over the proximal levee then abruptly changed, becoming dominated by less energetic, dilute, overspilling flows that deposited very thin- to thin-bedded, fine-grained, upper division turbidites

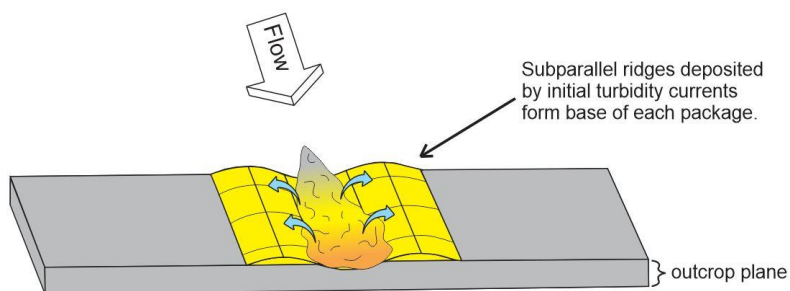
in the upper part of the package (Stage D Fig. 6.3). Intercalated thick-bedded, coarser-grained turbidites in the upper part record overspill by uncommon, anomalously large flows. The thickness of multiple (3-4) set Tcd/e turbidites in the upper part of the package thin as the lower part of the package thickens to its maximum, beyond which the Tcd/e turbidites then thicken as the lower parts thin. Although the Tc portion and overlying Td/e caps individually thicken and thin by only a few centimetres, collectively they act to level (i.e., heal) the underlying topography created by the thickening and thinning of the LPs – specifically, thickening or thinning to fill or flatten negative or positive topography, respectively.

The vertical packages in the proximal levees of this study are interpreted to have formed because of changes in both autocyclic and allocyclic controls. Specifically, they are interpreted to be related to recurring changes in the vertical density structure of channelized flows, which in turn was controlled by grain size and grain sorting. Initially, channelized flows were coarse grained and moderately well-sorted, causing them to have a plug-like density structure with negligible vertical stratification and high flow efficiency. In addition, the velocity maximum generally occurred well above the height of the incipient channel margins, thereby allowing the lower, coarse-grained, dense parts of flows to continue to overspill and deposit thick-bedded, coarse-grained turbidites that build up the lower part of each package. The sharp contact with the upper part of each package marks the point when channel relief exceeded the height of the velocity maximum and only the upper, finer-grained parts of currents overspilled. Later the make up of the sediment supply changed, specifically an increase in the previously underrepresented sand fraction, which in turn caused flow to adopt an exponential density profile (Stage E Fig. 6.3). This kind of density structure promotes interfacial mixing with attendant energy loss, which here

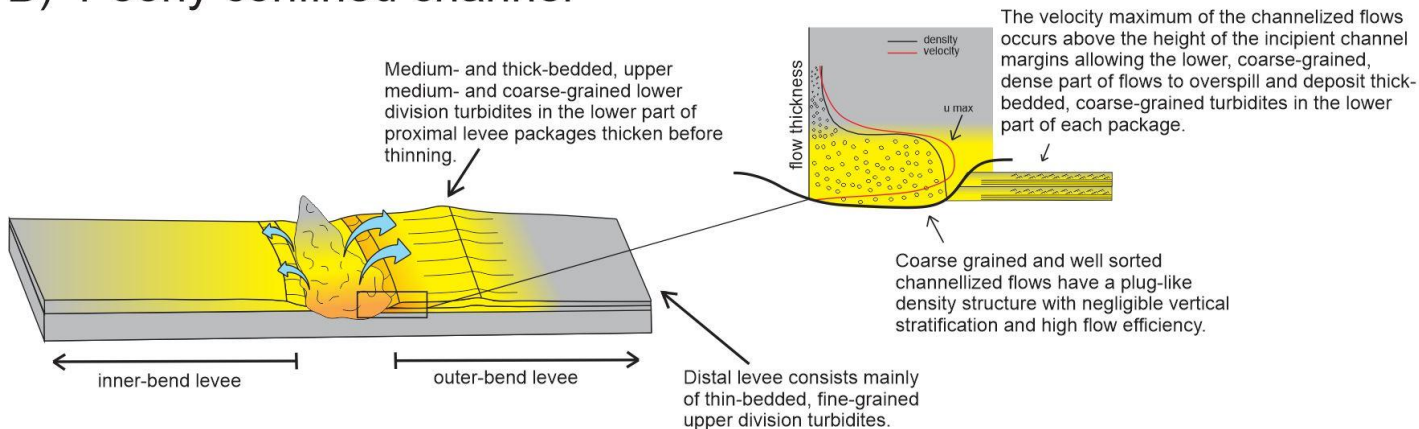
resulted in deposition within the channel. This, then, was succeeded by a return to flows with a plug-like density structure and the inception of the next channel-levee package (Stage F Fig. 6.3).

Importantly, the question arises as to why the channellized flows that formed the lower parts of packages consisted of moderately well-sorted, coarse-grained sediment, which here is interpreted to be a consequence of a short-term transgression. As the shoreline retrograded landward a significant part of the sand flux became preferentially sequestered into landward-migrating transgressive features like barrier islands. During the transgressive systems tract and possibly into the highstand and ensuing falling stage systems tract, the primary source of sediment for basinward-flowing turbidity currents was palimpsest and relict coarse sediment that had become stranded on the shelf, in addition to omnipresent fine grained sediment. This supply of fine, but much more importantly, coarse, moderately well-sorted sediment, would have resulted in highly efficient (i.e., limited interfacial mixing) flows with a plug-like density profile. Later as the sediment supply became more polydispersed, possibly indicating lowstand conditions and the arrival of the hinterland sediment supply, flows developed an exponential density profile with intense interfacial mixing that rendered them less capable of efficient sediment transport and hence promoted deposition within the channel. Therefore, each of the four levee packages observed in this study are interpreted to have formed as a result of short term recurring changes in relative sea level that profoundly affected the granulometric makeup of the sediment supply, which in turn controlled the spatial and temporal conditions of deposition (Fig. 6.7).

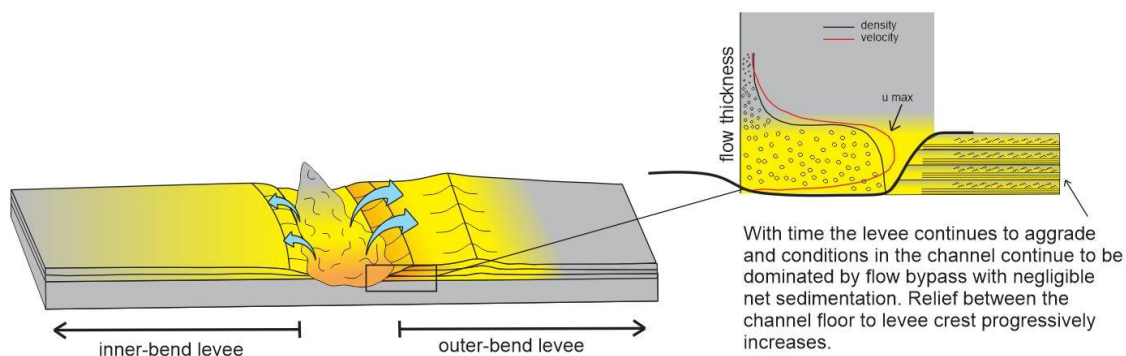
A) Incipient conditions



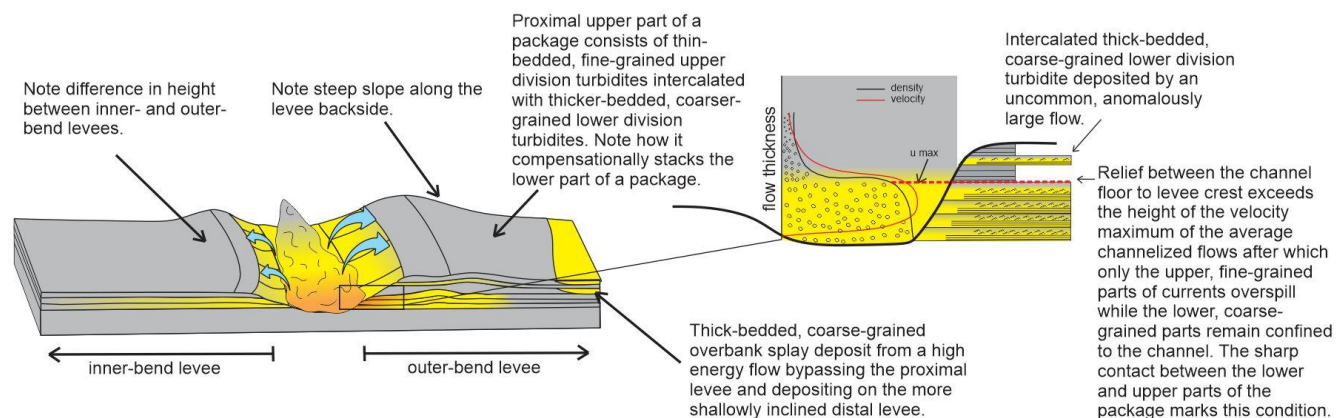
B) Poorly confined channel



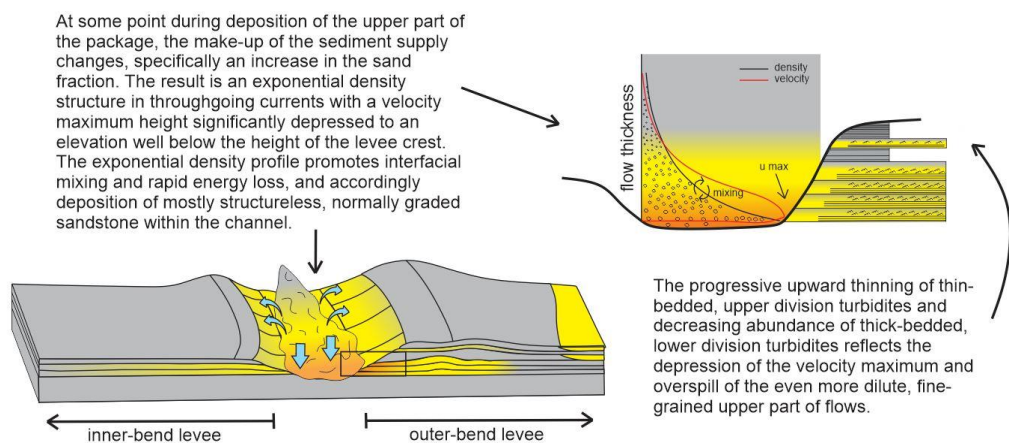
C) Increased channel confinement



D) Highly confined channel



E) Completion of the channel-levee system



F) Inception of next channel-levee package

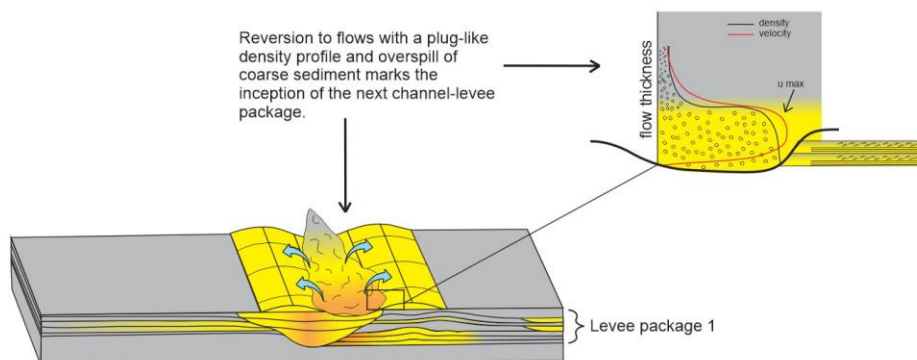


Figure 6.3: Depositional model for the channel-levee packages observed in this study.

6.2 Deposition of the leveed-channel complex

Due to the lack of age control within the study area, or much of the Windermere sedimentary pile in the southern Canadian Cordillera for that matter, it is not possible to accurately determine the length of time it took to form a single channel-levee package, and consequently, the duration over which the entire leveed-channel complex formed. Nevertheless, the composition of strata in the study area, specifically the occurrence of carbonate, provides evidence for long-term changes in sea level. Strata in the study area exhibit a 135 m thick upward change from siliciclastic with carbonate fragments to exclusively siliciclastic to mixed carbonate-clastic. At the base of the succession, abundant clasts of carbonate-cemented sandstone and limestone in channel unit 1 indicate the basinward resedimentation of sediment eroded from a pre-existing carbonate platform, most likely associated with a fall of relative sea level and deactivation of carbonate production. The ensuing lowstand consisted exclusively of a mud to gravel siliciclastic system. This, then, was succeeded by the re-introduction of carbonate sediment, signalling the re-initiation of the carbonate platform most probably associated with rising (transgressive) and/or highstand conditions of relative sea level. Superimposed on this long term allogenic change were higher order changes in relative sea level that are recorded in the channel-levee packages 1 to 4 (see previous section).

6.2.1 Time-Stratigraphic Development of Channel-Levee Packages 1 to 4

At the base of the section lowstand conditions resulted in the erosion of a now exposed shallow-water carbonate platform. Channel units 1 and 2 (Fig. 6.6, 6.7) were filled mostly with siliciclastic sand and dispersed carbonate-cemented sandstone and limestone clasts, including uncommon stromatolite and oolite fragments. The genetically related levees, on the other hand, were composed of finer-grained siliciclastic sand and mud. During the ensuing eustatic rise, quite possibly related to a transition from glacial to interglacial conditions, a shallow-water platform

became flooded and re-initiated carbonate production. Resedimentation of carbonate detritus onto the slope and attendant early seafloor carbonate cementation and lithification (van Der Kooij et al., 2010), steepened the slope beyond that more typical of their siliciclastic counterpart (Ross et al., 1994). Oversteepening, and consequent gravitational instability, resulted in widespread mass wasting that is interpreted to have caused the deposition of mud-rich debrite 1 and carbonate-rich debrite 2 (Fig. 6.6, 6.7). The occurrence of carbonate-cemented sandstone clasts, but more significantly oolite and stromatolite fragments, suggests that instability extended upslope and included mass wasting of even shallow-water parts of the carbonate platform.

Directly overlying the debrites in the southern and northern outcrops are proximal outer-bend levee deposits with anomalously abundant mudstone intraclasts and sets of coarse-grained dune cross-stratified sandstone (southern outcrop only) (Fig. 6.6, 6.7). These anomalous features are interpreted to be a consequence of the underlying debrites, firstly by providing a local source for mud intraclasts, and secondly high-angled slopes that allowed overspilling flows to achieve dune- rather than more typical ripple-forming flow speeds.

These strata are then overlain by channel-levee packages 3 and 4 (Fig. 6.6, 6.7). Thin-bedded, fine-grained, upper division turbidites in their upper part are a distinctive red in the field (Fig. 6.4), which in thin section is shown to be related to the oxidation of an iron-rich carbonate cement (Fig. 6.5) formed from the recrystallization of detrital carbonate grains. The abundance of resedimented carbonate grains indicates a well-developed shallow-water, mixed carbonate-clastic platform, and accordingly transgressive to highstand conditions.

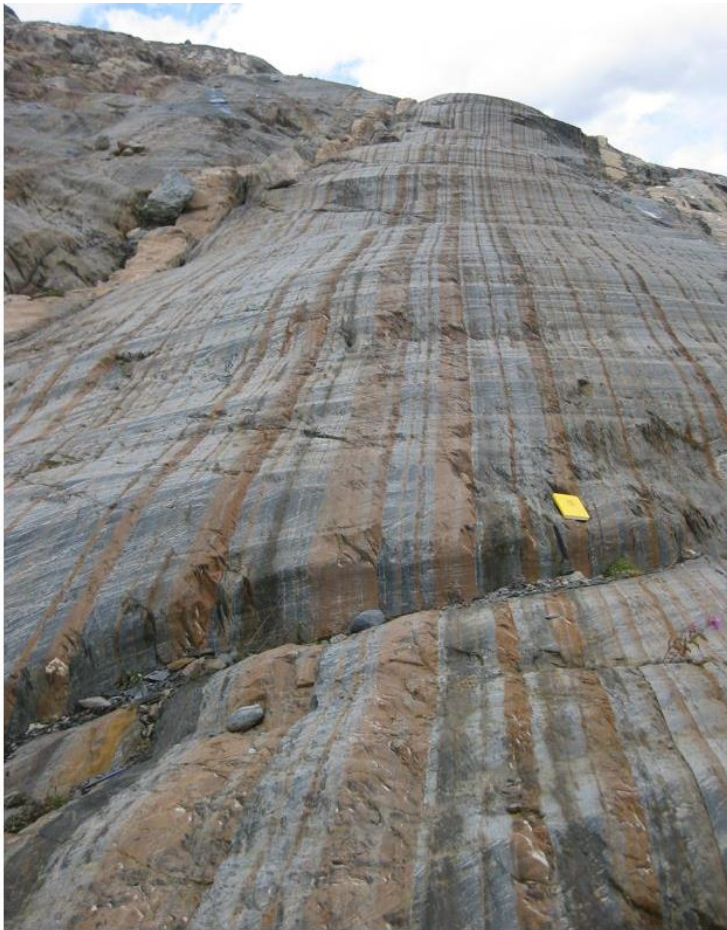


Figure 6.4: Thin-bedded, fine-grained, upper division turbidites in the upper parts of packages 3 and 4 are a distinctive red in the field.

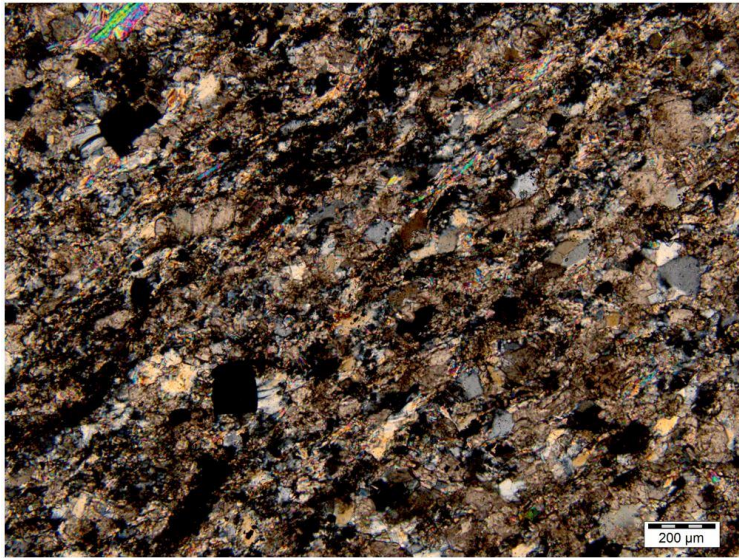


Figure 6.5: Photomicrograph of a thin-bedded, fine-grained upper division turbidites from the upper part of package 3 in cross-polarized light. Iron-rich carbonate cement makes up approximately 60% of the total stratal volume.

Subsequently, sea level began to fall (falling stage systems tract) with attendant slope instability indicated by the emplacement of debrites 3 and 4 and the thick slide that crops out in the southeast drainage and southern outcrops (Fig. 6.6, 6.7). Only 30 metres above these strata is Isaac channel 5, and a return to exclusively siliciclastic sedimentation associated with lowstand (Schwarz and Arnott, 2007).

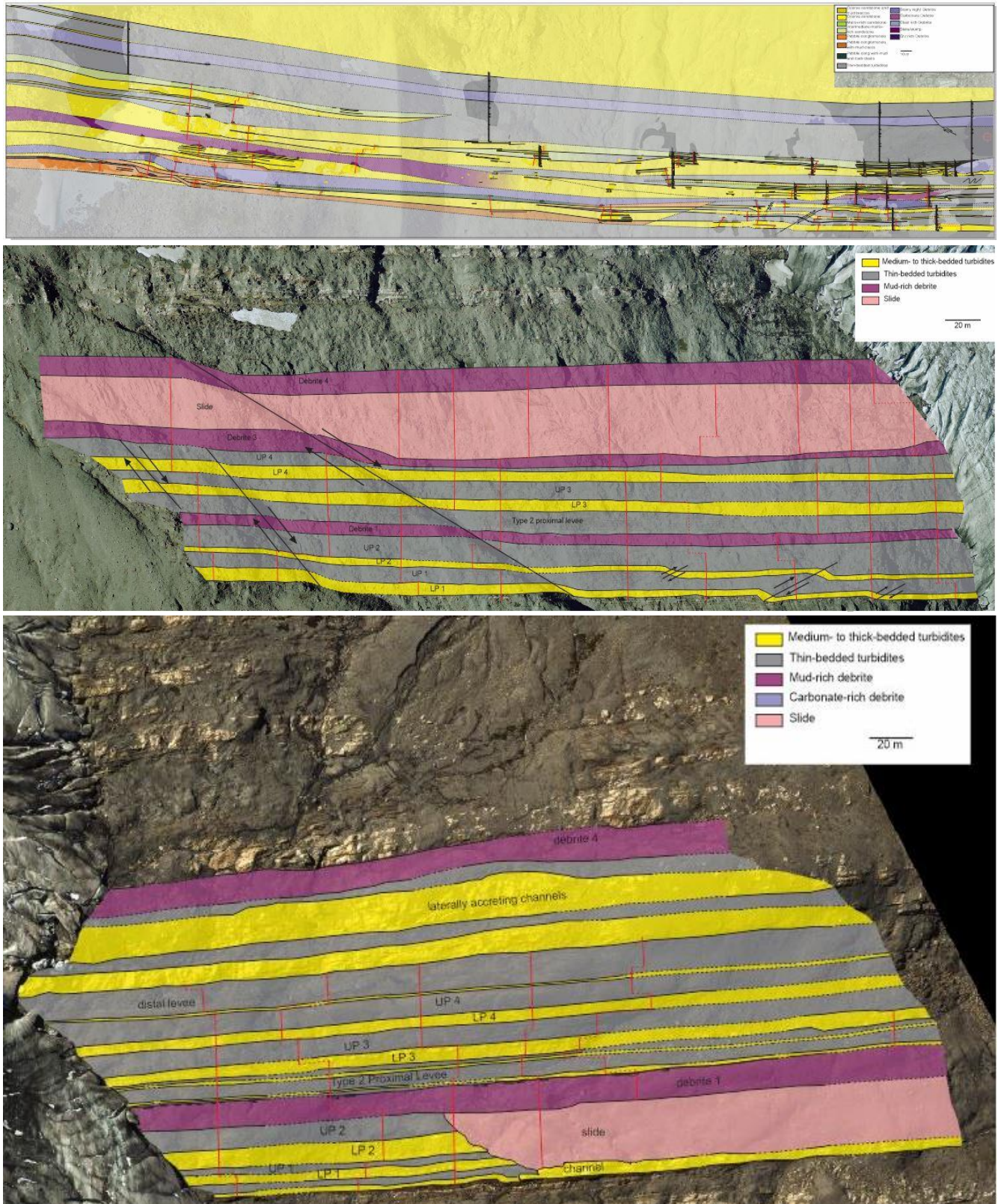


Figure 6.6: Aerial photographs showing the interpreted channel, levee, slide and debris deposits in the study areas (top: southeast drainage outcrop (from Terlaky, 2016), middle: southern outcrop, bottom: northern outcrop). Log locations are shown in red.

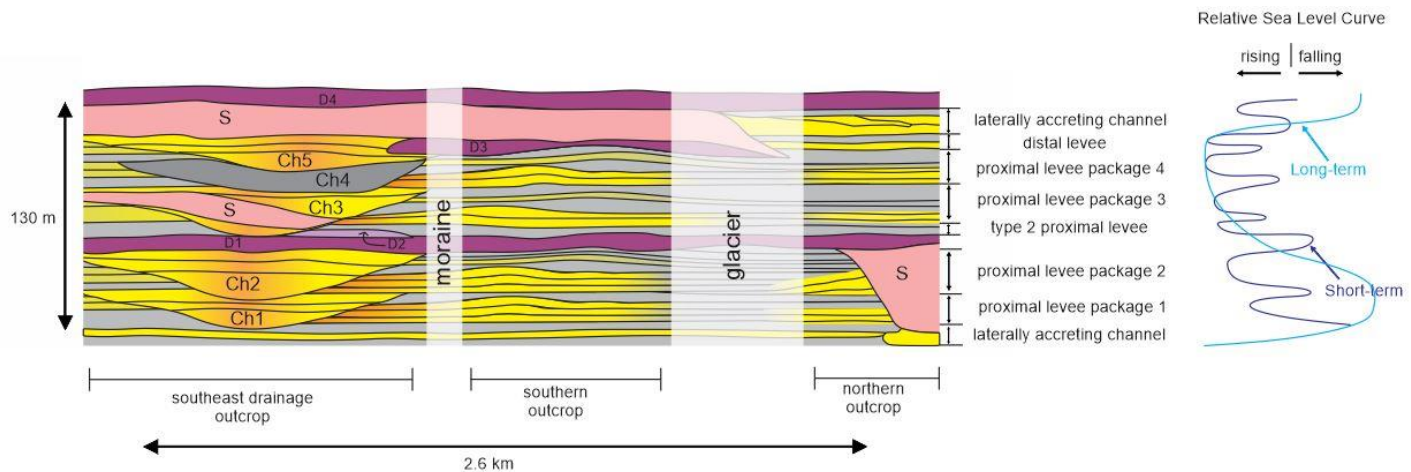


Figure 6.7: Model showing the time-stratigraphic development of the entire channel-levee complex. Short-term changes in relative sea level were shown to influence the development of the channel-levee packages whereas long-term changes controlled the overall mineralogy of the strata and the occurrence of mass-wasting deposits. Channel units 1 to 5 are labelled CH1 to CH5, debrites 1 to 4 are labelled D1 to D4 and the various slides are labelled S.

6.3 Implications for hydrocarbon exploration

Submarine fans are the largest depositional features on Earth (Curry et al., 2002). On the continental slope the principal depositional element in these systems is levee deposits associated with slope channels. Owing to their large geographical extent and high sedimentation rates levees represent major depocenters of inorganic and also organic detritus – levees in the modern Bengal (submarine) Fan, for example, act as major mechanisms for sequestering organic carbon. For example, the modern Bengal Fan alone is estimated to bury 10-20% of the total annual global flux of organic carbon (Galy et al., 2007). As a result, levees form a potentially significant part of a complete deep-water petroleum system, not only as a reservoir (e.g., Weimer et al., 2000; Clemenceau et al., 2000), but potentially also as an important source of petroleum, and therein the impetus for developing a highly resolved understanding of the stratal architecture and lithological makeup of an ancient deep-marine levee system. This thesis, therefore, describes the lateral and vertical trends in bed thickness, structure and grain-size distribution in a succession of levee deposits from the Neoproterozoic Windermere Supergroup, which are crucial parameters for

estimating and risking the spatial characteristics of reservoir quality and continuity in subsurface hydrocarbon reservoirs. Additionally, it explains how and why reservoir quality decreases away from deep-marine sinuous channels and also the relative timing of sedimentation on the levee and in the adjacent channel. Also, the sand-rich, generally coarse-grained lower part of packages in proximal levees of this study were found to be laterally continuous with little change in grain size, potentially making them excellent reservoirs with good horizontal connectivity that also hydraulic communication with good reservoir-quality channel sandstones. In addition, some mud- and sand-rich levee deposits are rich in residual carbonaceous matter (residuum of degraded primary organic carbon), suggesting that in addition to being potentially good reservoir rocks, mid-fan levees can also serve as source rocks for hydrocarbon generation.

6.4 Areas for future research

As described in chapter 5, due mainly to time constraints, not every individual thin-bedded turbidite in the upper part of proximal levee packages was measured. Instead, the upper parts of packages were subdivided into units based on lithological similarities and, within these units, minimum, maximum and average bed thicknesses of single set T_c, multiple set T_c, and T_{d/e} turbidites were recorded. It would be useful to measure each individual bed in the upper parts of packages in order to further investigate the antithetic thinning and thickening trends between the lower and upper parts of packages and determine if topographical flattening by differential sedimentation is in fact a general condition of levee sedimentation. Also, in multiple set T_c turbidites, the cross-stratification commonly consists of a distinctive interlamination of very fine-grained sandstone and siltstone. Due to time constraints these strata were not thoroughly investigated but future work could focus on describing and interpreting the physical basis of this enigmatic intercalation.

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8 Appendix

In this thesis, only the $\delta^{13}\text{C}$ of organic carbon was analyzed. Had the inorganic $\delta^{13}\text{C}$ from the carbonate cements been analyzed, for example Al-Mufti (2013), the following explanation would have been used to determine their timing and assess any possible microbial influence during their precipitation.

Published $\delta^{13}\text{C}$ values for carbonates precipitated from normal Neoproterozoic seawater using the Vienna Pee Dee Belemnite (VPDB) standard (Table 2) typically range from -5 to +6 ‰ (Kaufman and Knoll, 1995; Hayes et al., 1999; Jacobsen and Kaufman, 1999). Previous work in the Castle Creek study area concluded that the $\delta^{13}\text{C}$ values have not been reset by metamorphism, and therefore are representative of primary isotopic signatures (Meyer, 2004). As a consequence, $\delta^{13}\text{C}$ values for inorganic cements, such those of Al-Mufti (2013), which ranged from -21.42 to -16.41‰, are significantly depleted (in ^{13}C). Therefore, they must have been derived from sources other than normal seawater.

Reference	Approximate Age (Ma)	$\delta^{13}\text{C}$ VPDB
Kaufman and Knoll, 1995	600- 545	-5 to +6 ‰
Hayes et al., 1999	600- 590	-3 to +5 ‰
Jacobsen and Kaufman, 1999	600- 590	-5 to +4 ‰

Table 2: Approximate $\delta^{13}\text{C}$ values for carbonates precipitated from normal Neoproterozoic seawater obtained from the literature. Discretion should be used when referencing these ages due to poor age constraints and discrepancies in isotopic values in the Neoproterozoic (Meyer, 2004).

Carbon isotopic values may be altered by a variety of microbial or inorganic reactions during early diagenesis. These reactions include 1) the breakdown of organic matter and 2) mineral precipitation (Meyer, 2004). Organic matter is an unstable reactive constituent in marine sediments that is readily metabolized by bacteria during early diagenesis. Organic matter oxidation reactions

begin immediately after deposition and alter the chemical composition of pore waters (Hesse, 1990). Microbial organic matter decomposition and conversion into carbon dioxide and methane is carried out by five different populations of oxidant-specific bacteria. Each type of bacteria is present in a unique organic-matter decomposition zone that depends on local chemical conditions, which typically is controlled by depth below the sediment-water interface (Fig. 8.1) (Hesse and Schacht, 2011). Mineral precipitation associated with diagenesis can include pyrite and/or carbonates (Meyer, 2004). Pyrite is formed as a byproduct of sulphate-reducing bacteria (Ross et al., 1995) while pH controls the precipitation of authigenic carbonates (Hesse and Schacht, 2011; Meyer, 2004).

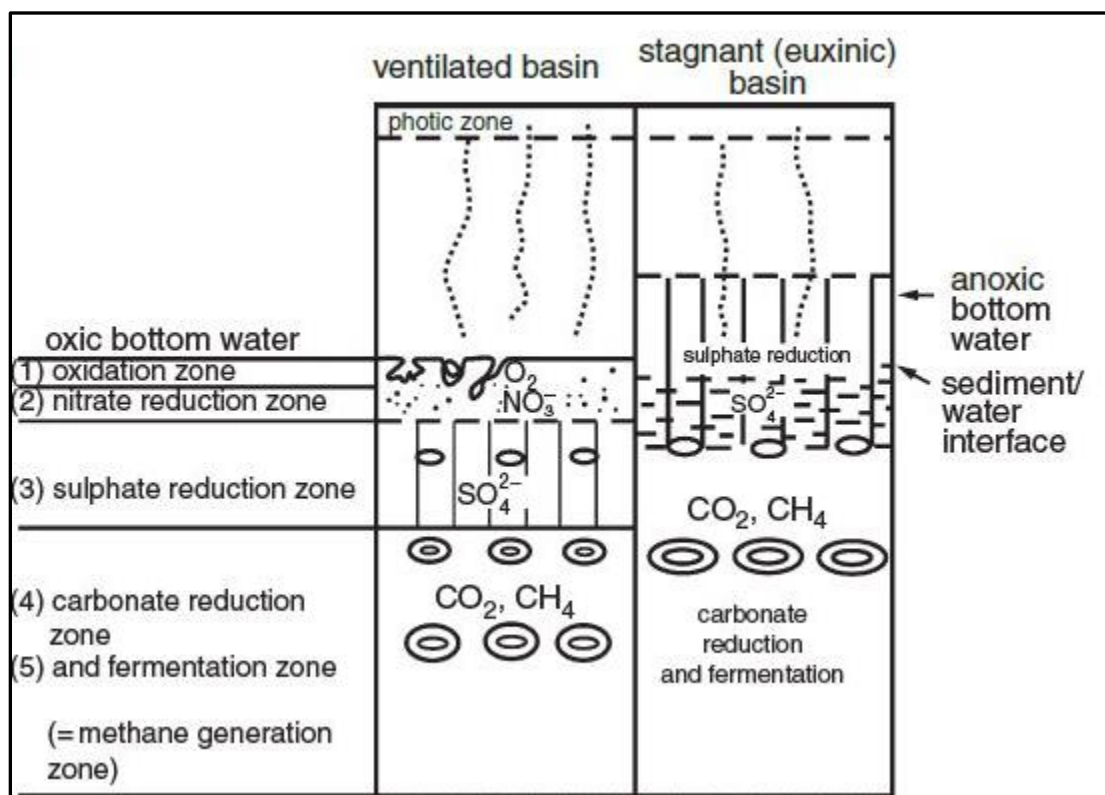


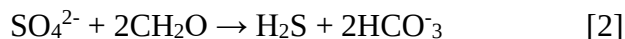
Figure 8.1: The five stages of organic-matter oxidation in anoxic sediments in ventilated and stagnant basins (Hesse and Schacht, 2011).

Depleted $\delta^{13}\text{C}$ values are the result of isotopic fractionation: the mass-dependent separation of isotopes during physical and chemical processes (Clark, 2015; Meyer, 2004). Bacteria and archaea can maximize the energy received from reduction-oxidation (redox) reactions by preferentially using the light isotope because its bonds are easier to break (for example phototrophs use $^{12}\text{C-O}$ instead of $^{13}\text{C-O}$ during photosynthesis) (Clark, 2015; Arthur et al., 1983). It is for this reason that biologically mediated redox reactions have high fractionation factors, and result in depleted $\delta^{13}\text{C}$ values (Clark, 2015; Meyer, 2004). For example, the $\delta^{13}\text{C}$ signature of marine plankton ranges from -15 to -28 ‰ (Arthur et al., 1983; Meyer, 2004). Methanogenic bacteria impart an even stronger carbon isotope fractionation, producing $\delta^{13}\text{C}$ signatures ranging between -50 to -100 ‰ (Arthur et al., 1983; Meyer, 2004).

Importantly, precipitation of early carbonate cements record a similar isotopic signature to the dissolved bicarbonate (HCO_3^-) in the pore water because of limited carbon isotope fractionation. In addition, the isotopic composition of the bicarbonate is related to the isotopic signature of the original carbon source, therefore, the carbon isotope values measured in the cements of this study record a similar isotopic signature to the original carbon source (Meyer, 2004).

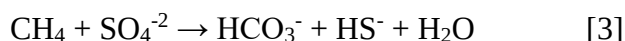
Strata in the Isaac Formation contain pyrite, commonly in amounts of ~1% of the total stratal volume, either as large crystals or as fine-grained, stratabound crystalline layers. Sulphur (^{34}S) isotopic analyses of pyrite in the Isaac Formation were interpreted to represent primary values unchanged by metamorphic processes (Ross et al., 1995). In the study area, $\delta^{34}\text{S}$ values are significantly depleted and indicate pyrite formed as a byproduct of bacterial sulphate reduction (BSR) (Ross et al., 1995). Thus, depleted $\delta^{13}\text{C}$ values from the study area may result from organic carbon oxidation via BSR (Dela Pierre et al., 2012; Meyer, 2004) and/or methane oxidation via

BSR (Meyer, 2004). In the anoxic sulphate-reduction zone, bacteria oxidize organic matter and reduce sulphate derived from pore water (Arthur et al., 1983; Meyer, 2004):



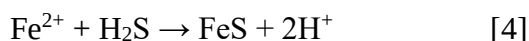
The isotopic composition of the bicarbonate (HCO_3^-) formed during BSR will be approximately the same as the original organic carbon source that the sulphate reducing bacteria consumed (Arthur et al., 1983). Since the original phototropic organic matter was already enriched in ^{12}C (depleted in ^{13}C) the isotopically light carbon became incorporated into the bicarbonate, resulting in bicarbonate with depleted $\delta^{13}\text{C}$ values (Arthur et al., 1983; Meyer, 2004).

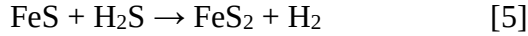
In addition, upward-diffusing methane (from the underlying methane generation zone, Fig. 3.15) may also be oxidized by bacteria in the BSR zone, producing ^{12}C -enriched HCO_3^- and possibly also contributing to the depleted $\delta^{13}\text{C}$ values of this study (Meyer, 2004). Methane is oxidized by bacteria in the BSR zone via the reaction (Arthur et al., 1983):



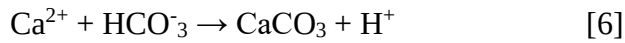
Like organic matter oxidation, methane oxidation increases alkalinity and if combined with reactions that maintain a high pH (like Fe reduction), the increase in alkalinity promotes the precipitation of carbonates (Hesse, 2011).

The hydrogen sulphide (H_2S) generated from the breakdown of organic matter via BSR [2] will either: migrate upwards in the sediment column into the oxidation zone (Fig. 4) where it will be oxidized back to sulphate (Al-Mufti, 2013) or under low Eh conditions and in the presence of ferrous iron (Fe^{2+}) will react to form iron sulfides, such as the pyrite observed in strata of this study (Smith, 2009). Pyrite is precipitated via the following reactions (Al-Mufti, 2013):





To precipitate calcite the ^{12}C -enriched bicarbonate (HCO_3^-) formed from BSR [2, 3] requires a source of calcium. Possible sources may include diffusion of calcium from overlying seawater or calcium released from feldspar alteration, which will react with the bicarbonate to form calcite via the reaction (Al-Mufti, 2013):



Calcite cements in this study are iron-rich, indicating that there was abundant iron in the pore waters (Al-Mufti, 2013). It is interpreted that there was limited Ca^{2+} availability and thus calcite incorporated excess iron into its structure.

The stable carbon (^{13}C) isotopes of strata analyzed by Al-Mufti (2013) range from -21.42 to -16.41‰. These $\delta^{13}\text{C}$ values do not match the $\delta^{13}\text{C}$ signature of carbonates precipitated from normal Neoproterozoic seawater (~ -5 to $+6$ ‰), from bicarbonate (HCO_3^-) produced from the oxidation of organic matter via BSR (~ -20 ‰) or from HCO_3^- produced from the oxidation of methane via BSR (~ -50 to -100 ‰) (Arthur et al., 1983; Meyer, 2004). Therefore, it is interpreted that the $\delta^{13}\text{C}$ values from carbonate cements of that study result from the mixing of HCO_3^- from different sources. These sources include HCO_3^- from normal seawater ($\delta^{13}\text{C}_{\text{sw}}$), pore waters enriched in HCO_3^- from the oxidation of organic matter via BSR ($\delta^{13}\text{C}_{\text{bsr}}$) and pore waters enriched in HCO_3^- from the oxidation of methane via BSR ($\delta^{13}\text{C}_{\text{m}}$):

$$\delta^{13}\text{C}_{\text{carb}} = \delta^{13}\text{C}_{\text{sw}}(X) + \delta^{13}\text{C}_{\text{bsr}}(Y) + \delta^{13}\text{C}_{\text{m}}(Z) \quad [7]$$

where X, Y and Z represent the amounts of $\delta^{13}\text{C}$ from each source contributing to the total $\delta^{13}\text{C}$ of the calcite cements (Meyer, 2004), the exact proportions being beyond the scope of this thesis.