DEPOSITIONAL ARCHITECTURE OF A NEAR-SLOPE TURBIDITE SUCCESSION: UPPER KAZA GROUP, WINDEREMERE SUPERGROUP, CASTLE CREEK, BRITISH COLUMBIA, CANADA

By

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Abstract

An expansive panel of well exposed (periglacial) strata of the Upper Kaza Group permitted a detailed study of the stratal architecture of proximal basin floor deposits in the Neoproterozoic Windermere turbidite system. Detailed stratigraphic and petrographic analyses identified six lithofacies: poorly-sorted, clast-rich mudstone (F1), thin-bedded siltstone and mudstone (F2), thick-bedded, massive sandstone (F3), medium-scale, cross-stratified sandstone (F4), mudstone-clast breccia (F5), and medium-bedded turbidites (F6). The spatial distribution of these facies identify five architectural elements: heterolithic feeder channel deposits (FA1), thin-bedded intralobe turbidites (FA2), terminal splay deposits (FA3), distributary channel deposits (FA4), and isolated scours (FA5). FA 1-4 are genetically related and form the basic building blocks of large-scale basin floor depositional lobes. FA 5, which is isolated to the stratigraphic top of the study area, is interpreted to have formed in a base-of-slope setting, and its superposition on FA 1-4 suggests the long-term progradation of the Windermere turbidite system.

Résumé

Un affleurement bien exposé des strates du groupe de Kaza supérieur a permi une étude détaillée de l'architecture stratigraphique des dépôts de plancher de bassin océanique dans le système de turbidite Néoproterozoïque Windermere. Des analyses stratigraphiques et pétrographiques détaillées ont identifié six faciès : mudstone mal-trié avec clastes abondants (F1), siltstone et mudstone à lits minces (F2), grès massif à lits epais (F3), grès à stratification entrecroisée d'échelle moyenne (F4), brèche à clastes de mudstone (F5), et turbidites à lits moyens (F6). La distribution spatiale de ces faciès identifient cinq éléments architecturaux : dépôts hétérolithiques de chenal d'alimentation (FA1), turbidites à lits minces d'intralobe (FA2), dépôts d'épandage terminal (FA3), dépôts de chenal de distribution (FA4), et affouillements isolés (FA5). FA 1-4 sont génétiquement connexe et forment les blocs fonctionnels de base des lobes dépositionnels de plancher de bassin océanique à grande échelle. FA5, qui est localisé stratigraphiquement au-dessus du secteur d'étude, s'est probablement formé dans un environnement en bas de pente, et sa superposition sur FA 1-4 suggère la progradation à long terme du système de turbidite de Windermere.

Author's Declaration

I declare that the work in this dissertation was carried out in accordance with the Regulations of the University of Ottawa. The work is original, except where indicated by special reference in the text, and no part of the dissertation has been submitted for any other academic award. Any views expressed in the dissertation are those of the author.

SIGNED: ....

DATE: ....

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1: Introduction

Introductory Statement

Rocks of the deep marine, typically situated in inaccessible areas, have been studied largely via indirect methods such a two and three-dimensional seismic. This has allowed entire deep marine systems (1000's of km^2) to be imaged and expanded our understanding of the deep marine. The main drawback of seismic imaging is that it is unable to resolve features smaller than approximately 10 metres. The only method to study features of those dimensions is to study them in outcrop. Unfortunately, most outcrops are several orders of magnitude smaller than the architectural elements of a typical deep-marine system.

In the Castle Creek study area (Cariboo Mountains, east-central B.C.), deep-marine proximal basin floor and base-of-slope deposits of the Neoproterozoic Windermere Supergroup (WSG) are well-exposed over an area of several km^2 (seismic-scale). Here, the recently-deglaciated landscape of vertically-dipping strata allows for the observation of mm- to km-scale lithological characteristics and vertical & horizontal trends. This type of outcrop provides the ideal bridge between typical outcrop studies and seismic imaging by allowing multiple architectural elements and a large extent of an ancient deep-marine system to be observed directly.

Southern Canadian Cordillera

Between 755 – 700 Ma (Neoproterozoic), Laurentia separated from the supercontinent Rodinia (Colpron *et al.*, 2002) and created a passive margin along its present-day western coast. Accumulation of sediment within the developing ocean basin (proto-Pacific Ocean) was interpreted by Monger & Price (1979) to be similar to the situation in the present-day Atlantic

Ocean. Thermally-driven subsidence in the newly-formed basin led to the accumulation of a thick succession (>6km) of deep marine sediments within the miogeocline (Ross *et al.*, 1995; Ross & Arnott, 2007). A second episode of subsidence that began during the Early Paleozoic led to the accumulation of over 10 km of mostly shallow marine carbonate rocks (Monger & Price, 1979; Devlin *et al.*, 1988). Passive margin sedimentation was terminated during the Jurassic as the passive margin was converted to a continental arc system. Accretion of micro-continents and related deformational events led to the creation of the southern Canadian Cordillera (Reid *et al.*, 2002). Presently, the Canadian Cordillera comprises five distinct geological provinces; the Rocky Mountain, Omineca Crystalline, Intermontane, Coast Plutonic, and Insular belts. The Rocky Mountain and Omineca Crystalline belts are composed of deformed marine sediments deposited in the passive margin miogeocline. The latter three represent either a collage of accreted terranes or a period of subduction-related magmatism during mountain-building. This present-day geological configuration is thought to have been largely complete by the Late Cretaceous, although uplift and deformation continued well into the Early Tertiary (Monger & Price, 1979; Cant & Stockmal, 1989).

Windermere Supergroup in the southern Canadian Cordillera

The Windermere Supergroup (WSG) is a succession of Neoproterozoic sedimentary rocks that extends from northern Canada to northern Mexico; a strike length of over 4000 km (see Figure 1.1). In the southern Canadian Cordillera, deep-water sedimentary rocks of the Windermere Supergroup are 6-9 km thick and exposed over an area of at least 35,000 km^2 (Ross *et al*., 1995). At the base of the succession, magmatic rocks intercalated with siliclastic rocks

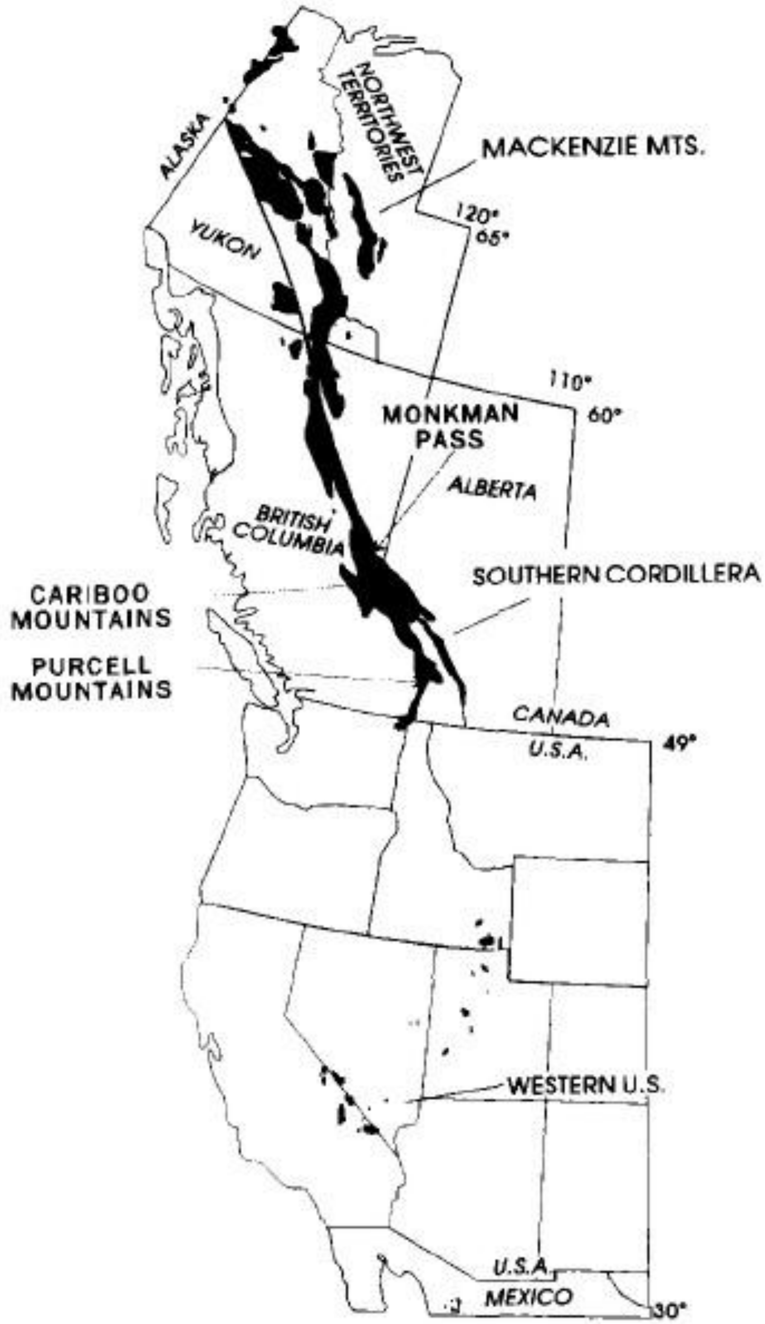

Figure 1.1: Distribution of Neoproterozoic WSG strata in western North America (from Ross *et al.*, 1995)

have been interpreted to represent rift facies related to continent break-up. These, in turn, are overlain by a succession of deep-water sedimentary rocks that chronicles the full extent of a classic passive margin system, shallowing upwards from basin floor to slope to shelf deposits.

The age of the WSG in the southern Canadian Cordillera is poorly known and constrained by three widely spaced data points. Data from the rift assemblage at the base of the WSG, specifically mafic volcanic rocks of the Irene Formation, have been dated at approximately 750-720 Ma (Colpron *et al*., 2002). An intermediate age from within the WSG is from a mostly fine-grained stratigraphic marker unit between the Middle and Upper Kaza groups called the Old Fort Point Formation(Smith, 2009). A horizon of black, organic-rich shale within the Old Fort Point is of stratigraphic significance because it is interpreted to represent both a maximum flooding surface and a period of deep-ocean anoxia (Kendall *et al*., 2004; Smith, 2009). Moreover, this unit has been dated by the Re-Os technique to 607.8 ± 4.7 Ma (Kendall *et al*., 2004). The youngest age for deposits of the WSG is bracketed by volcanic rocks within the Lower Cambrian Gog Group, which have been dated at 569.6 ± 5.3 Ma (Colpron et al., 2002). These rocks, however, overlie the Windermere Supergroup, and therefore represent a minimum age for Windermere sedimentation.

Although deep-water rocks of the Windermere Supergroup currently crop out over an area of at least 35,000 km^2, Mesozoic (Cordilleran) faulting and deformation has greatly reduced its original size. Palinspatic reconstruction of the Windermere turbidite system suggests that the system may have been at least 80,000 km^2, and therefore similar in scale to the modern-day deep-water Amazon and Mississippi Fans (Ross & Arnott, 2007). Paleoflow within the deep-water part of the WSG was generally toward the northwest. Finally, studies of detrital zircons within the WSG suggests that sediment deposited in the Windermere turbidite system was derived from the southern Canadian shield (Ross & Arnott, 2007).

Cariboo Mountains

The study area is in the Columbia Mountains of east-central British Columbia, which form part of the Omineca Belt in the southern Canadian Cordillera. Local structure is dominated by km-scale Jurassic folds that are bounded by a series of strike-slip faults (Reid *et al*., 2002). Metamorphic grade in the Cariboo Mountains is sub-greenschist in the northern part, and increases toward the south.

Neoproterozoic sedimentary rocks in the Cariboo Mountains comprise the Kaza and Cariboo Groups (Reid *et al*., 2002). In the Cariboo Mountains the Cariboo Group is subdivided, stratigraphically upwards, into the Isaac, Cunningham, and Yankee Belle Formations (see Figure 1.2). Regionally, the Kaza Group is 2-3 km thick and comprises mostly coarse-grained sandstone and conglomerate interbedded with fine-grained sandstone, siltstone, and shale (Reid *et al*., 2002; Ross *et al*., 1995). Rocks of the Kaza Group have been interpreted to represent stacked depositional lobes deposited in a basin floor setting (Meyer & Ross, 2007). The Kaza Group also includes the only regionally extensive marker unit within the WSG, the Old Fort Point Formation. Immediately overlying the Kaza Group is the Isaac Formation (~2.5 km thick), a unit made up of deep-marine slope shale interrupted locally by laterally discontinuous coarse-grained channel-fill deposits (Reid *et al*., 2002; Navarro *et al*., 2007). These, in turn, are overlain by the 800 m-thick Cunningham Formation, which is dominated by carbonate-rich upper slope to outer shelf deposits. Finally, the Yankee Belle Formation is approximately 900m thick and consists of a sequence of alternating carbonate and siliciclastic rocks interpreted to have been deposited in a high-energy shelf environment (Ross *et al*., 1995).

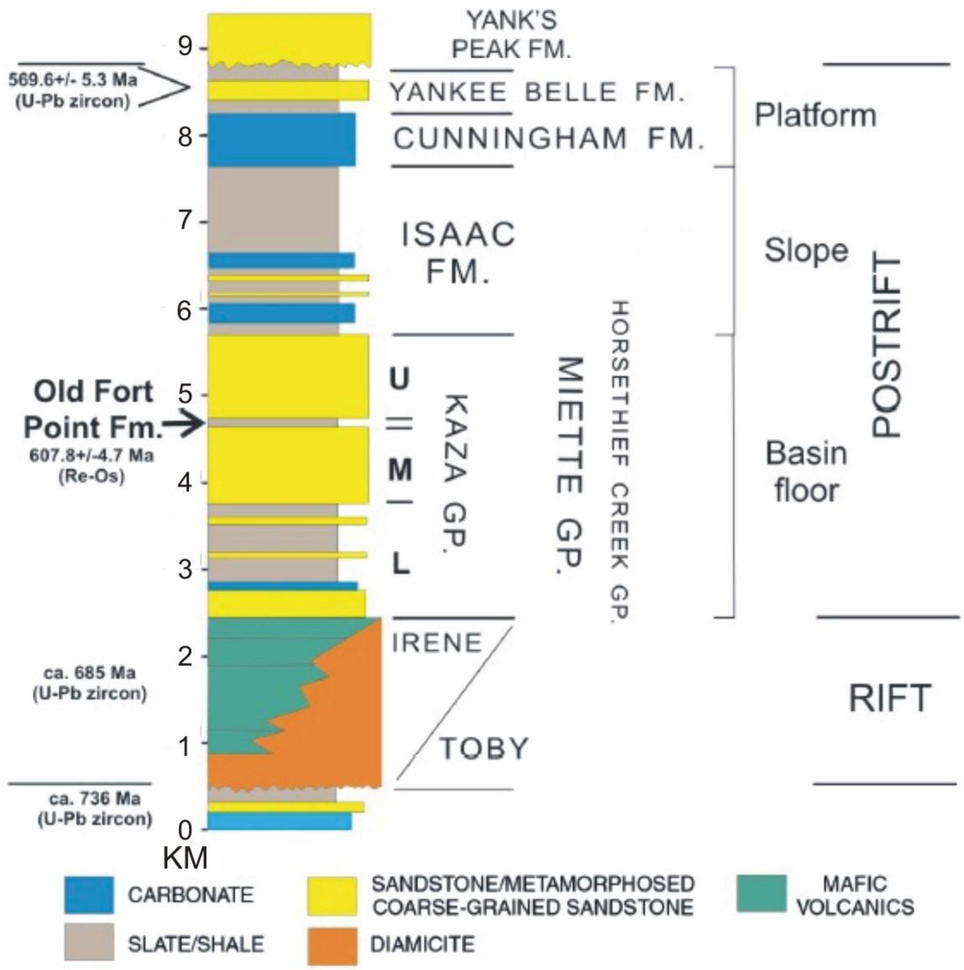

Figure 1.2: Regional stratigraphy of the WSG in the Cariboo Mountains of the Southern Canadian Cordillera. Main age controls and interpreted depositional environments shown. (Modified from Ross & Arnott, 2007).

Field Location: Castle Creek

The study area is located in east-central British Columbia, near the headwaters of Castle Creek, in the northern part of the Cariboo Mountains (~53º03'N, 120º26'W). At this location there is a km-scale exposure of Windermere Supergroup stratigraphy consisting of the Upper Kaza Group and the Isaac Formation. These rocks have been only recently exposed by the retreat of a nearby alpine glacier and there is little weathering and plant overgrowth, which accordingly allows them to be studied in detail. In addition, the exposed rocks have experienced only low-grade metamorphism (sub-greenschist facies), which has pristinely preserved primary sedimentary structures. Near-vertical bedding allows easy access to the entire stratigraphic section at the site. Access to the field location is limited to airlift by helicopter due to steep terrain and lack of roads. The area studied was a ~800 m x 160 m portion of the Upper Kaza Group in the greater Castle Creek study area.

Objectives & Methodology

The main objective of this project is to describe the stratal composition and architecture of rocks in the Upper Kaza Group near the contact with the Isaac Formation. Fieldwork was conducted over the course of two summer seasons, each two months long. Plotting and correlating of vertical section were assisted by mapping onto aerial photographs (~1:2500) of the study area. Outcrop quality was best in the stratigraphically lowermost portion of the study area and became patchier in the uppermost part (~top 50 m).

A total of 6 main and 15 mini stratigrapic sections were measured in detail, including descriptions of: grain size, nature of bedding contacts, sedimentary structures, and lithology. Correlation of bed contacts and significant stratigraphic horizons was done by physically walking out the surfaces in question. In this way the main architectural elements and their constituent lithologies could be studied. A multispectral scintillometer was also used to gather data for each section. Data were collected in 75cm increments and later compiled to create continuous gamma ray profiles.

In total, 56 rock samples were collected in the field, either using a rock hammer and chisel, or a portable rock saw. Thin sections were then made and used to examine mineralogy, grain size, sorting, structure, and matrix/cement. Thin sections suspected of containing carbonate minerals were stained with Alizarin Red and potassium ferrocyanide (Dickson, 1966). All the sections were studied using an Olympus BX41 optical microscope and a SZ61 stereo microscope.

Previous work in the Upper Kaza Group, Southern Canadian Cordillera

The first 1:250,000 geological mapping of the Upper Kaza Group in the Cariboo Mountains was published in 1973 by Campbell *et al*. A more detailed map (1:50,000), but still with limited sedimentological and stratigraphic detail, was published later by Ross & Ferguson (2003).

A detailed study of Upper Kaza Group Strata at Castle Creek was first published in 2004 (Meyer, Msc. Thesis). It sought to use detailed stratigraphic correlations and lateral as well as vertical facies changes to reconstruct part of the Windermere deep-water turbidite system. Meyer concluded that the sand-rich strata in the study area were deposited in an unconfined basin floor setting within a prograding turbidite system. In addition, the majority of Upper Kaza Group strata were deposited in the mid-fan part of a larger-scale basin floor fan.

Later studies of the Upper Kaza Group at Castle Creek have focused on comparatively smaller windows of stratigraphy. These include a highly detailed correlation of the internal architecture of a single depositional lobe (Privett, K., 2008), as well as a stratigraphic study of the interpreted slope to basin floor transition zone at the Kaza Group – Isaac Formation contact (Navarro, PhD. Thesis, ongoing).

Statement of Contributions

The research described in the following chapters adds a level of stratigraphic detail to an important portion of the Upper Kaza Group, one that is interpreted to represent the beginnings of the transition from proximal basin floor to base of slope. This transitional zone has not been much studied in outcrop and has been the object of little research in modern systems. Therefore, a detailed study of the stratigraphy, facies, and the way they change both laterally and vertically, is useful for future research concerning deep-water turbidite systems.

Furthermore, this study provides information which may be of some use for the exploration of turbidite deposits for the purpose of hydrocarbon exploitation. Many of the arctichectural elements studied at Castle Creek can be isolated in seismic imagery. However, internal architecture is not visible in seismic images, where an entire stratal element many metres thick is indicated by a single line (see Fig. 1.3). Stratigraphic detail provided by this study can indicate the connectivity of stratal elements and point out possible barriers to oil extraction.

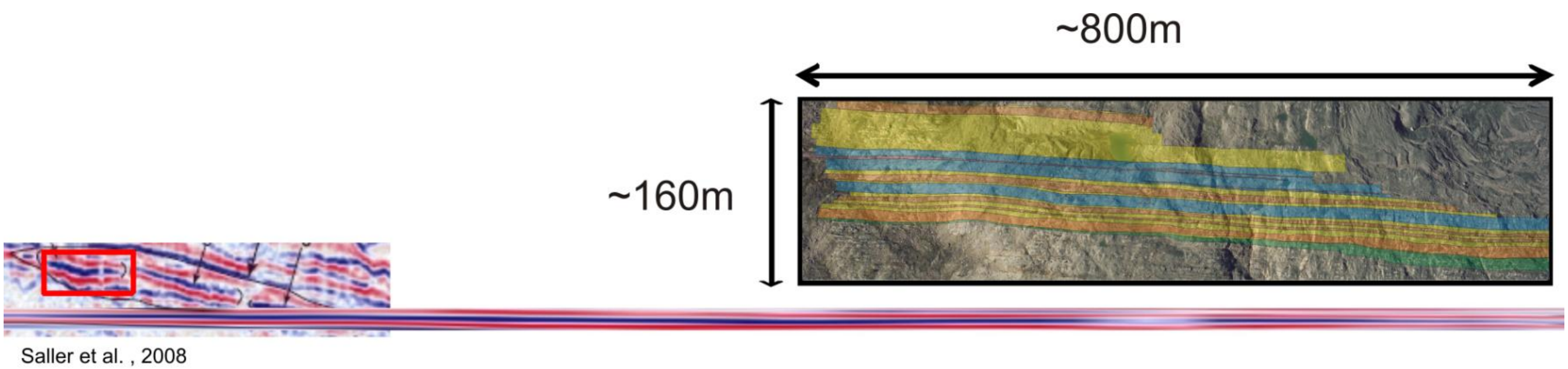

Figure 1.3: A comparison between a recent seismic image of a modern system (highlighting a single terminal splay) and the study area for this thesis (showing architectural elements). Bottom bar is the seismic image adjusted to the scale of the Castle Creek outcrop image. Note that even the coarsest internal details of a stratal elements are not visible on the seismic image.

2: Facies & Facies Associations

Introduction to sediment-gravity flows

Sediment-gravity flows are flows that move due to a density difference between it and the surrounding fluid. This allows the flow to displace the lower density fluid and move. Two types of sediment gravity flow will be introduced in this section: debris flows and turbidity currents.

A debris flow is a sediment-gravity flow that moves via plastic deformation because of its cohesiveness. This characteristic is due to a matrix composed of silt, clay, and the surrounding ambient fluid (air or water). The strength of this matrix allows a debris flow to support coarser particles and large clasts in suspension for its entire distance of travel. Deposition of sediment in a debris flow occurs abruptly when the frictional forces at any part of the flow's boundary exceeds the gravitational force acting upon it (Arnott, 2010). This causes the flow to freeze en masse and immediately form a debris flow deposit, or debrite. This rapid deposition prevents any sorting from taking place, resulting in very heterogeneous deposit. In many debrites, large, rafted clasts are oriented parallel to the basal contact of the deposit. This is interpreted to be a result of alignment with internal shear surfaces during transport (Arnott, 2010).

Turbidity currents are sediment gravity flows in which the main mechanism of particle support is fluid turbulence. A typical turbidity current is made up of three components: head, body, and tail (see Fig. 2.1). The head is the part of the flow where the sediment concentration is the highest, while the tail is where the turbidity current is most dilute. Via extensive mixing within the flow and differences in the settling rates for differently- sized particles, the turbidity current becomes both vertically and horizontally sorted.

Turbidites are deposited in an idealized sequence from assumed waning flow conditions (see Fig 2.2). This sequence was initially interpreted by Bouma (1962) and separates the deposits into 5 turbidite divisions (Ta → Te). The first in the depositional sequence, the Ta division, typically consists of massive, often graded, coarse to very coarse sandstone. It has been interpreted to form from rapid suspension fallout of the coarsest sediment in a highly concentrated turbidity current (Lowe, 1982). The Tb divison consists of parallel-laminated medium sand and is interpreted to be equivalent to upper flow regime plane bed. As the flow decelerates further, lower flow regime conditions become dominant and the deposition rate lessens. Next in the depositional sequence is the Tc division consisting of ripple-laminated fine sand. The Td division consists of faintly planar-laminated silt & mud, whose lamination has been interpreted to be due to differences in settling velocity between silt grains and mud flocs (Stow & Bowen, 1978). Lastly the Te divison consists purely of mud and is interpreted to represent slow deposition from suspension.

The classic Bouma turbidites are interpreted to be deposited by a low density turbidity current, one with a volumetric sediment concentration of approximately 9%. A turbidity current with a sediment concentration higher than this value is considered to be a high density turbidity current, whose resulting deposit may not follow the Bouma sequence. Such high sediment concentration is interpreted to result in a massive, poorly sorted Ta/S3 bed (Lowe, 1982) and has been supported via experimental flows created in a sediment flume (Leclair & Arnott, 2005; Arnott & Hand 1989). This type of flow deposits extremely rapidly (≥ 4cm/min aggradation) from suspension after undergoing a hydraulic jump, preventing lamination and trapping some of the finer grained portion of the flow in the resulting deposit.

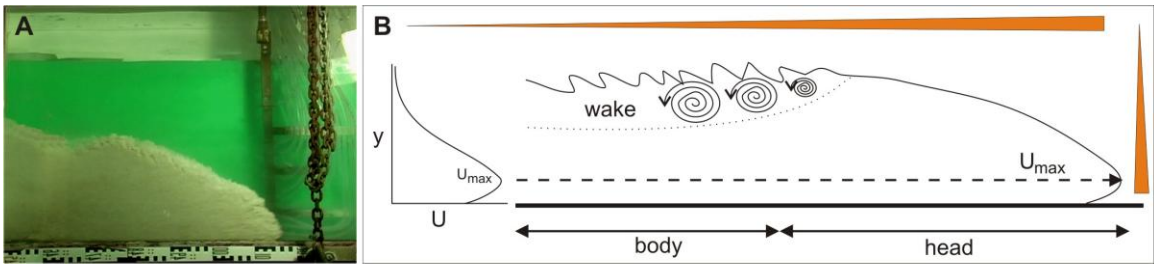

Figure 2.1: (A) Turbidity current in a flume, showing the head of the flow, (B) Schematic of a turbidity current showing the mixing with in the flow and its velocity profile. Orange triangles show the grading trends within the flow, both horizontal and vertical. (Modified from Arnott, 2010).

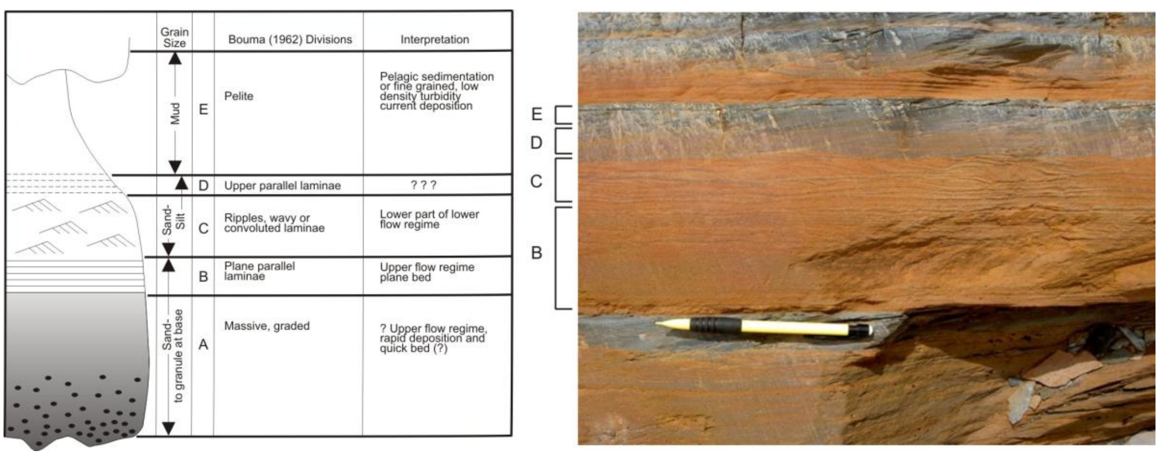

Figure 2.2: Diagram showing the idealized Bouma sequence turbidite (from Arnott, 2010).

Introduction to contour currents & contourites

Contour currents (also termed bottom currents) are unidirectional deep-sea currents that follow seafloor topography. These quasi-permanent currents typically flow parallel to the continental slope, but may change direction or velocity due to localized variations in topography (Arnott, 2010). Contour currents are capable of re-working seafloor sediments and preventing deposition, which can create gaps in the sedimentary record (Stow *et al*., 2008). However, when not accelerated by a bathymetric constriction, contour currents are unable to transport grains larger than fine sand.

Sediments deposited or modified by contour currents are called contourites (Stow *et al*., 2008). The most common large-scale example of contourite deposition is a sediment (or contourite) drift, which can cover 50 to 10^6 km^2. At the bed scale, contourites occur in an ideal sequence 20 to 30 cm thick (see Fig. 2.3). This sequence consists of a upward-coarsening basal interval from muddy to silty then sandy contourites, followed by a upward-fining interval that is approximately the inverse of the basal one (Stow & Faugères, 2008). Muddy contourites tend to be poorly sorted with indistinct or diffuse lamination. They can also contain lenses of sand or silt. Silty contourites may also contain faint or discontinuous laminae, and rarely contain cross-laminated intervals. They are also poorly sorted and may contain lenses of coarser-grained material (Stow & Faugères, 2008). These lenses of coarser material have been interpreted to be the result of repeated episodes of erosion, winnowing, and deposition (Stow *et al*., 2008). Sandy contourites are rare and occur most commonly as thin discontinuous layers or interbedded within finer-grained contourites. Planar or cross-lamination is rare and the beds are poorly sorted. Where lamination is preserved it is typically diffuse. In general, most contourites are

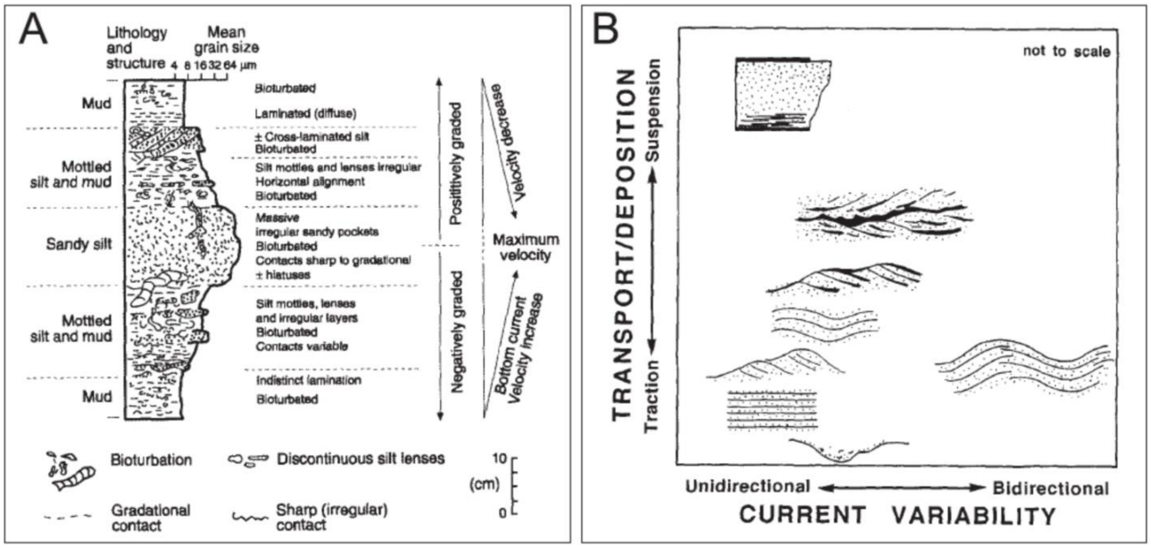

Figure 2.3: (A) Idealized contourite sequence. Note the abundance of silt/ mud, and the paucity of medium sand or coarser (Stow et al., 2002). (B) Example of the types of lamination in contourite facies (Ito, 1996).

characterized by the absence of distinct lamination, even in cases where bioturbation is absent (Stow *et al*., 2008).

In outcrop, it can be difficult to differentiate strata deposited by contour currents or turbidity currents, respectively, contourites versus turbidites. However, contour currents generally flow parallel to the slope whereas turbidity currents typically flow downslope. This should result in a bimodal distribution of paleocurrent measurements in outcrop (Arnott, 2010). Other possible criteria for contourite identification include abundant internal erosion surfaces and a mottled appearance due to cm-scale lenses of fine sandstone. In addition, a paucity of distinct laminae and the absence of coarser sandstone may also be diagnostic of contourite facies.

At the Castle Creek study area, fine-grained strata that might be candidates for contourites are conspicuously and distinctly cross-laminated. Laminae are commonly continuous and highlighted by pervasive calcite cementation. There is also a distinct lack of the mottled texture described in the literature (Arnott, 2010; Stow & Faugères, 2008). Small lenses of sediment and abundant small-scale internal erosion surfaces are also notably absent. In addition, the average paleoflow measured within the study area is ~250° (discounting a single outlier), which is similar to the regional paleoflow direction. This suggests that the influence of contour currents on sediment deposition and distribution at the Castle Creek study was most likely negligible.

Facies

Strata in the Castle Creek study area have been subdivided into six facies based on: grain size, lithology, sedimentary structures, sorting, thickness, and bedding contacts (see Table 2.1). Bed thickness terminology follows the classification of Mckee & Weir (1953): thin-bedded - 3-10 cm thick, medium-bedded - 10-30 cm thick, and thick-bedded - 30-100 cm thick. In addition, small- and medium-scale cross-stratification is defined as a set that is <5 cm and 5-30 cm thick, respectively.

Facies 1: Poorly-sorted clast-rich mudstone

Strata of this facies are typically massive and generally consist of variably-sized clasts dispersed in a silty mudstone matrix. There is no readily apparent pattern to the distribution of the clasts vertically within the deposit, but they tend to be oriented sub-parallel to the base of the deposit. Clasts are commonly composed of mudstone and include both laminated and unlaminated kinds, with laminated clasts tending to be larger and more angular. Laminated clasts are up to ~1 m long with an almost tabular shape, and typically consist of a similar lithology to the subjacent strata (see Figure 2.4). Unlaminated mudstone clasts range from 1 cm to 13 cm long, and are 0.3 cm to 2.5 cm thick. Some of the unlaminated mudstone clasts are black, in contrast to the green-grey mudstone most common in the study area. Carbonate-cemented coarse-grained sandstone clasts were also observed. These distinctively red-brown coloured clasts are typically rounded and ~25 x 8 cm in size. In some cases the clasts are internally stratified, which commonly is contorted. Another type of clast observed in this facies was variably-sized rounded quartz grains, which on average are upper coarse sand to granule in size.

#	Facies Name	Lithology	Sedimentary Structures	Bouma/ Lowe Division	Contacts	Thickness	Geometry	Other Features	Interpretation
1	Poorly-sorted, clast-rich mudstone	Quartz granules & pebbles dispersed in a mudstone matrix	N/A	N/A	Planar base; upper surface often reworked by a dune x-stratified bed	Up to 4 m	Sheet-like	$CaCO_3$-cemented sst clasts, mudstone intraclasts	Debris flow deposit (debrite)
2	Thin-bedded siltstone & mudstone	Siltstone & mudstone beds interstratified with rare fine to medium-grained sandstone beds	Faint horizontal lamination; ripple & planar lamination	Mostly Tde, rare Tb and Tc beds	Planar bases and tops	< 5 cm; Tc is 1-2 cm; Tb up to 15 cm	Sheet-like	Pyrite-rich	Suspension fallout from dilute flows; hemipelagic sedimentation; rare high energy flows
3	Thick-bedded, massive sandstone	cL–vcL sub-arkose sandstone	Massive or coarse-tail graded; amalgamated beds	Ta/S3	Shallowly-scoured bases; sometimes overlain by a thin siltstone bed or a single dune set	0.6 to 2.5 m	Lenticular to sheet-like	Flames and mudstone intraclasts along bedding contacts; $CaCO_3$-cemented pillar structures	Rapid suspension deposition from high concentration flows
4	Medium-scale, cross-stratified sandstone	mU-vcU subarkose sandstone	Dune cross-stratification	Tt	Shallowly-scoured bases; sometimes “foundered” into the top of underlying Ta/S3 bed	5 to 30 cm	Lenticular	$CaCO_3$-cemented	Sediment reworking by the tail of a turbidity current
5	Mudstone-clast breccia	Abundant mudstone clasts dispersed in a cU-granule conglomerate matrix	N/A	N/A	Erosive basal contact, top not always discernible; mostly present as a ‘zone’ within beds of Facies 3.	Up to 1 m	Irregular	Present usually at bedding contacts of sandstone beds or dispersed in an amalgamated bed (likely marking a cryptic bed contact)	Sourced by updip erosion; deposited in scours as a lag; from the breakup of a siltsone layer at an amalgamation surface
6	Medium-bedded turbidites	mU sandstone -> fU sandstone-> siltstone -> (mudstone)	Normal-grading; planar lamination; ripple cross-stratification; laminated siltstone	Tabcde	Undulose to planar base; top typically planar	15-60 cm	Sheet-like	Partial Bouma sequences are common; uncommon $CaCO_3$-cemented ripples	Deposition from a waning, low-density turbidity current

Table 2.1: Summary of the six facies identified in proximal basin floor strata at the Castle Creek study area.

Facies 1 occurs as two end member kinds: continuous and discontinuous. The continuous type (of which there is only one occurrence) is several metres thick (2.5 – 4.0 m) and overlain by medium-scale cross-stratified coarse-grained sandstone. Boulder-sized clasts of both mudstone (laminated and unlaminated) and sandstone are dispersed in a poorly-sorted matrix of mostly clay minerals (70%), rounded quartz grains from lower fine to upper coarse sand size (20%), and minor chlorite, muscovite, partially altered plagioclase feldspar, and calcite (see Figure 2.5). The laminated mudstone clasts are preferentially oriented sub-parallel to the base of the deposit. The discontinuous type is up to 30 cm thick and contains smaller mudstone clasts compared to the continuous type. Laminated mudstone and carbonate-cemented sandstone clasts were not observed and the matrix is generally coarser, consisting of more abundant silt. The matrix is composed of: quartz grains (70%) ranging from lower fine to lower coarse sand, calcite (10%), and minor clay minerals (5%), cubic pyrite, muscovite, plagioclase feldspar, and chlorite.

Continuous strata of Facies 1 can be traced laterally across the width of the study area (approx. 800 m) with only minor local variations in thickness. Despite the abundance of clasts, the base of the unit is surprisingly planar with little indication of scouring. Discontinuous strata, on the other hand, typically persist for no more than 10 m laterally with only a few notable exceptions that were traceable for up to 150 m. Local scours up to 5 cm deep were observed. Where sufficient outcrop continuity is present (5 m or more), these beds have been observed to thin and grade laterally into thin-bedded siltstones and mudstones (see Facies 2 next).

Interpretation

The heterogeneity and incorporation of many different types and sizes of clasts into a fine-grained matrix is consistent with the general description of debris flow deposits (Haughton

Figure 2.4: Large laminated mudstone clast in a thick bed of Facies 1.

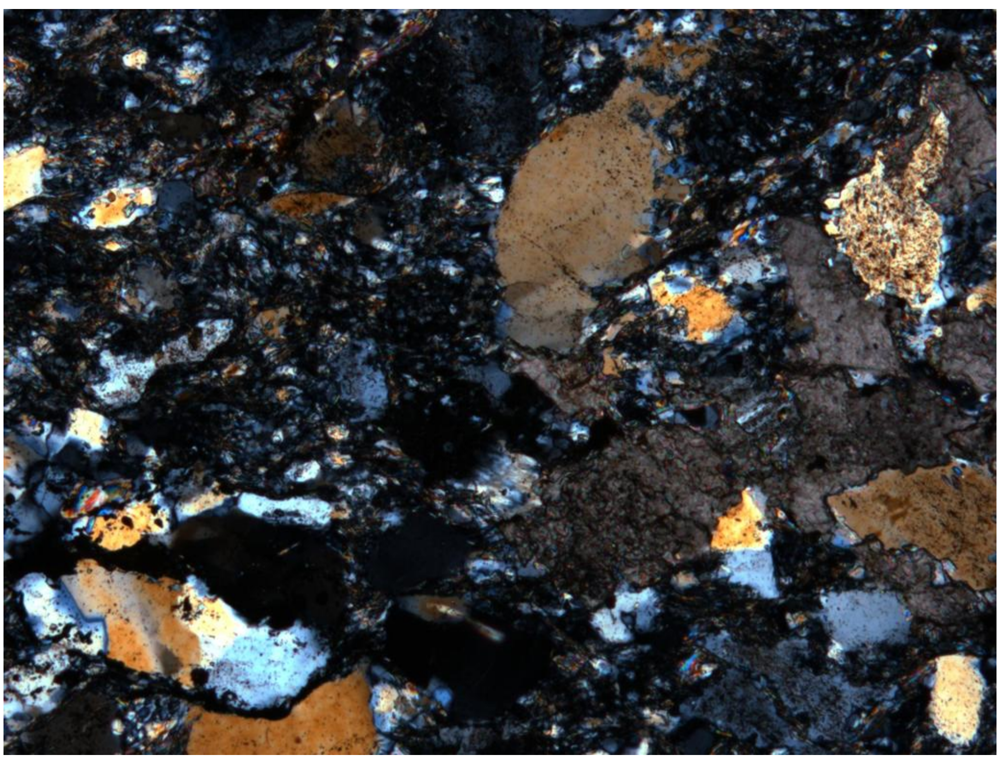

Figure 2.5: Microphoto of a sample of Facies 1. Note the poorly-sorted nature and the abundance of a fine-grained matrix. (This microphoto is approximately 1.8 mm wide).

et al., 2003). Debris flows are a kind of sediment gravity flow in which a cohesive silt/clay-water matrix supports the submerged weight of suspended larger particles as the flow moves (Lowe, 1982). The chaotic distribution of the clasts indicates that preferential settling of larger particles did not occur, either during flow or following deposition, and that deposition occurred by *en masse* freezing (Talling *et al*, 2004). Lack of relief along the base of the continuous deposit implies that the erosive incorporation of the boulder-sized mudstone and sandstone clasts occurred further upslope and that the study area was a zone of deposition. Moreover, it suggests that at least locally, no large clasts were dragged along the seabed, like those that form the erosional furrows characteristic of the base of some debris flow deposits (kms long, up to 15m deep) (Posamentier & Kolla, 2003). The flat, featureless bases might also indicate that the debris flow hydroplaned on a thin layer of water, which prevented it from interacting directly with the bed (Mohrig *et al*., 1998), and also helped minimize internal fluid turbulence. The layer of water forms when fluid at the front of the flow is unable to move upwards and over it. The water is either forced forward or penetrates as a wedge beneath the head of the flow (Mohrig *et al*., 1998). This basal layer of water would allow the flow to travel long distances even on a very shallow slope by reducing basal friction and increasing flow velocity.

The presence of black, unlaminated mudstone clasts in the continuous deposits also implies that the debris flow travelled a significant distance, due to the paucity this type of mudstone in the study area (these clasts, therefore, were eroded from a more proximal location and moved downslope). The discontinuous debris flow deposits, with their lack of lateral continuity and comparative thinness, may result from the transition of a turbidity current into a localized debris flow. This transformation is reported to occur when a turbidity current erodes a mud-rich substrate and as a result causes the flow to increase its internal cohesiveness and bulk

Figure 2.6: Thick unit composed largely of beds of Facies 2. Top is to the right of the picture, note the rusty weathering due to incorporated pyrite crystals.

density (Ito, 2008). Even a relatively low volume of clay (~5%) may be sufficient to cause turbulence to become damped and clasts to become increasingly buoyant and hence more easily suspended (Lowe, 1982). Their interpreted short distance of transport may be related to several factors, including low matrix clay content and hence low bulk permeability. Such conditions would have inhibited hydroplaning and elevated friction along the base of the flow, which in turn would have encouraged deposition.

Facies 2: Thin-bedded siltstones & mudstones

Strata of Facies 2 consist of thin-bedded mudstone interstratified with siltstone and fine sandstone. Beds of faintly interlaminated mudstone and siltstone are typically < 5 cm thick and tend to have planar bases commonly demarcated by a patchy veneer of fine sand grains. In addition, basal layers commonly consist of laterally discontinuous, upper fine cross-laminated sandstone up to 2 cm thick. Rare planar-laminated medium sandstone up to 15 cm thick is present. Rarely, there is a thin (<1 cm – 2cm) layer of unlaminated black mudstone present atop the laminated grey-green mudstone. In the study area, strata of Facies 2 form bed sets up to 10 m thick that generally overlie a planar basal contact (see Figure 2.6). Mineralogically, strata are dominated by clay minerals (50-60%) and quartz (30-40%), with minor pyrite, chlorite, muscovite, and calcite (see Figure 2.7). The distinctive deep red iron-stain seen in outcrop and on aerial photographs is due to the weathering oxidation of pyrite crystals.

Interpretation

The siltstone & mudstone beds are approximately equivalent to the upper-division Bouma turbidites (Td, Tde), while the thin rippled sandstones and thicker planar-laminated ones would be equivalent to the Tc and Tb divisions, respectively (Bouma, 1962). Laminated siltstone

Figure 2.7: Microphoto of a sample of Facies 2. Silt-sized quartz grains and recrystallized clay minerals are abundant. (This microphoto is approximately 1.8 mm wide)

and mudstone represent hemipelagic fallout or the suspension deposition of the low-energy parts of dilute turbidity currents (Lowe, 1982). Uncommon discontinuous, cross-laminated fine sandstone layers represent the migration of trains of sediment-starved current ripples by dilute low-energy flows, most probably just in excess of the critical bedload shear stress. Rare basal layers of planar-laminated, medium sandstone indicate the passage of anomalous high-energy turbidity currents capable of upper plane bed transport of medium sand. This basal layer would then be overlain by cross-laminated fine sandstone followed by laminated siltstone under waning flow conditions. The rare unlaminated black mudstone most probably indicates suspension fallout of fine clay particles with no transport or sorting (Lowe, 1982).

Facies 3: Thick-bedded massive sandstone

Facies 3 consists of coarse-grained, thickly-bedded, sub-arkose sandstone. Quartz content is high, averaging 70%, with a matrix composed of mica (muscovite, chlorite, etc.) and clay minerals making up about 18%. Minor plagioclase feldspar (~5%), calcite, apatite, and detrital zircon are also present (see Figure 2.9). Beds are usually weakly normal coarse-tail graded but can also be ungraded. Grain size ranges from very coarse to lower coarse sand, and it is common to observe granule and/or pebble-sized quartz clasts at the base of beds or rarely, dispersed throughout. Beds of Facies 3 are thick, ranging from 0.6 to 2.5 m. Strata are typically amalgamated, making it difficult to follow beds laterally. Where bedding contacts are visible, shallow scours are common, as are flame structures and mudstone clasts (see Figure 2.8). Carbonate-cemented pillar structures up to 5 cm long ~0.5 cm wide occur locally but generally are uncommon. Rarely, beds are overlain by a thin (< 5 cm) laterally discontinuous siltstone unit.

Figure 2.8: Flame structure at the contact between two massive sandstone beds.

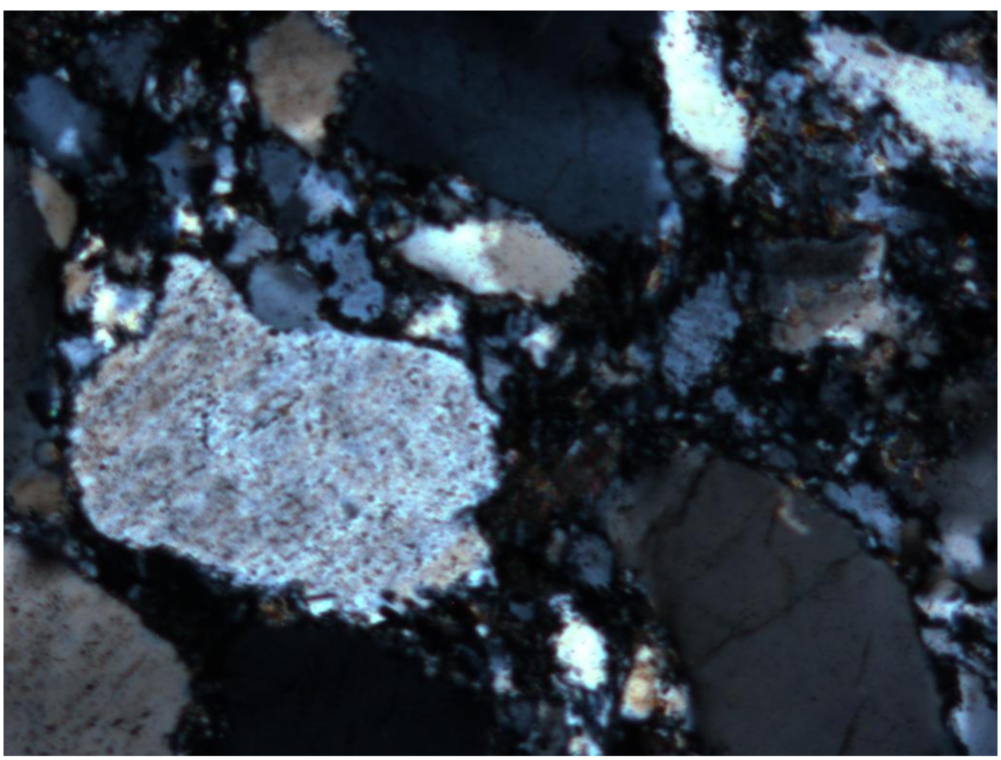

Figure 2.9: Microphoto of a sample of Facies 3. Quartz is the most abundant mineral and grains are poorly-sorted. Note the dissolution at grain boundaries. (Microphoto is approximately 1.8 mm wide)

Interpretation

Massive sandstones, like those of Facies 3 are common in the deep water sedimentary record. The Annot sandstone in southeastern France is a good example of this facies, where individual beds range in thickness from 10 cm to 4 m (Posamentier & Walker, 2006).

The massive sandstones of Facies 3 are equivalent to Bouma's (1962) Ta division and Lowe's (1982) S_3 division. One possible depositional mechanism for massive sandstones is rapid suspension sedimentation from a high-density turbidity current that experiences a hydraulic jump (Stow & Johansson, 2000; Lowe, 1982). Under these circumstances, the flow expands rapidly and loses a great deal of velocity, forcing it to immediately shed its coarsest fraction. This mechanism is supported by the experimental flume data of Arnott & Hand (1989), where migrating bedforms were unable to sort the bed sufficiently to form lamination at high bed aggradation rates ($\geq$ 4 cm/min). They suggested that migrating bedforms a few grains high, which would otherwise form planar lamination, were buried too rapidly by high sediment fallout. Alternatively Kneller & Branney (1995) propose that massive sand beds are deposited by sustained flow through gradual aggradation, rather than rapid sediment fallout from a collapsing high-density flow. They envisioned a two-part flow, a turbulent upper portion and a non-turbulent basal portion. The lower part of the flow in which grain interactions and hindered settling are the dominant sediment support mechanisms would be constantly supplied by a steady downward flux of sediment from the top of the flow (Kneller & Branney, 1995). This implies that the flow would deposit continuously at its base while being replenished by grains from the upper part of the flow. The lack of a defined interface between the flow and the underlying static bed would inhibit bed-load transport until the sediment flux fell below the threshold required to suppress it. The resulting deposit thickness would bear no relation to thickness of the flow that formed it, but rather the length of the flow (Kneller & Branney 1995). Although appealing, the experimental work of Leclair & Arnott (2005) suggests that bed load transport along a discrete interface remains active even at very high sediment concentrations (35% volume sediment concentration). This implies that the massive sandstones of Facies 3 were rapidly deposited from suspension from a high concentration flow rather than slowly aggraded under steady flow conditions.

Pipe structures observed in the study area are interpreted to be de-watering features related to high rates of sedimentation. Pillars form during bed consolidation because of forceful and sometimes explosive vertical escape of water (Lowe & Piccolo, 1974). The sediment within these features is usually better-sorted than the host bed, the finer grain sizes having been elutriated during fluid escape. The abundance of ferroan calcite within these features implies enhanced diagenetic fluid flow caused by higher permeability and hence preferential cementation. Although pillars would have originally been sub-vertical, their consequent obliquity to the basal bed contact is related to strain and re-orientation associated with Mesozoic folding.

Flame structures along some bedding contacts also suggest rapid deposition and insufficient time for water to escape from the consolidating sediment. These features are typically observed where there is a major increase in grain size above the bedding contact (see Figure 2.8). Flames form where the vertical flow of water from compacting sediment is impeded by a laterally continuous layer of impermeable material, usually clay-rich sediment. This causes the topmost sediment of the underlying bed to liquefy and attempt to flow upward, while at the same time parts of the overlying denser sediments are sinking into the underlying finer-grained material (Dasgupta, 1998). Density-driven deformation of the sediment takes place and results in a pattern of pockets of coarser sediment separated by thin protuberances of upward displaced finer-grained sediment.

Facies 4: Medium-scale cross-stratified sandstone

Facies 4 consists of medium to coarse-grained cross-stratified sub-arkose sandstone. The mineralogical makeup of the sandstone varies significantly; quartz 40-70%, clay mineral & mica matrix 5-40 %, and ferroan calcite cement 2-25% (see Figure 2.11). Minor plagioclase feldspar, non-ferroan calcite, biotite, pyrite, and detrital zircon are also present. The weathering of ferroan calcite, which is concentrated in thin zones aligned parallel to the cross-stratification, gives these rocks a distinctive pattern of reddish-brown stripes. Strata of Facies 4 are well-sorted and range from medium to very coarse upper sandstone, with rare granule conglomerate. Cross-stratified sandstone usually occurs as single medium-scale sets sharply overlying strata of Facies 2 or 3, although rare cosets made up of two stacked sets are observed. Beds range in thickness from 5 to 30 cm, have shallowly-scoured basal contacts and are laterally continuous over ~50-100 m. In places, beds thin so much that the result is a correlatable horizon of isolated mound-shaped cross-stratified units up to 2 m in length. Facies 4 occurs also as localized clasts embedded within strata of Facies 3 and 1. In both, the internal lamination is convolute and deformed, and in some cases completely overturned. Laminated clasts within the structureless sandstone of Facies 3 are always located near the top of the beds, typically within 10 cm of the top, and correlate laterally with an overlying cross-stratified sandstone horizon (see Figure 2.10).

Figure 2.10: Dune cross-stratified sandstone foundered into the top of a structureless sandstone bed.

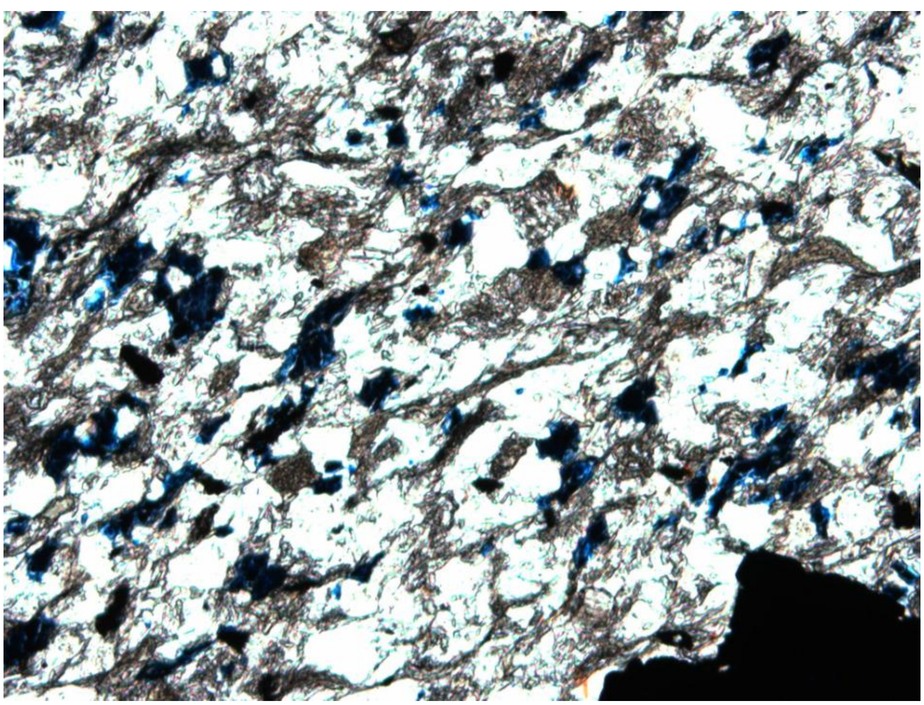

Figure 2.11: Microphoto of a sample of Facies 4 in plain light. The blue grains are stained iron-rich calcite crystals. Note that the minor calcite cement and the recrystallized clay mineral matrix is aligned parallel to the cross-stratification. (Microphoto is approximately 1.8 mm wide)

Interpretation

Medium-scale cross-stratification is not included in the classic turbidite succession described by Bouma (1962) and is rarely reported in deep water literature (Lowe, 1982; Arnott & Hand, 1989). Medium-scale cross-stratified sandstones of Facies 4 are interpreted to be the result of migrating dunes and are analogous to Lowe's (1982) T_t division. However, dune cross-stratification is relatively common in the study area, and the Upper Kaza in general (Meyer, 2004). Typically, dunes are the stable bedform over a wide range of flow velocities and grain sizes, from fine sand and coarser (Southard & Boguchwal, 1990). Assuming the presence of adequate grain size and flow speed, the most critical element required for the formation of dunes is flow separation. This occurs when a bed defect causes the boundary layer of the flow to detach from the bed (Middleton & Southard, 1984). An example of this process is immediately downstream of a ripple or dune crest, where the bed elevation drops abruptly. Some distance downflow, the boundary layer then re-attaches itself to the bed, typically on the stoss slope of the next bedform downstream. Intense turbulence is created along the interface between the detached boundary layer and the underlying low-energy zone characterized by weak eddy current motion (Middleton & Southard, 1984). Erosion is maximized at flow re-attachment, serving to deepen the trough between individual ripples and dunes. If flow separation is unable to take place and allow bed defects to amplify, dunes will not form. The question, therefore, is how can flow separation be prevented? It has been theorized that when a fluid is accelerated flow separation may be delayed (Allen, 1984). However strata of Facies 4 occur at the top of beds, which suggests a decreasing energy. Other mechanisms that have been suggested to suppress dune formation include flow unsteadiness and insufficient time for dune formation. Flow unsteadiness is considered to be negligible due to the well-graded nature of classic turbidites, and experiments have shown that dunes require relatively little time to form, far less than the estimated duration of most turbidity currents (Arnott & Hand, 1989). High sediment concentration in a turbidity current may prevent flow separation by increasing viscosity and dampening near-bed turbulence (Allen & Leeder, 1980). Under such conditions, dunes would be unable to form and upper plane bed would remain the stable bedform. By the time sediment concentration is low enough for adequate flow separation and dune formation, the available sediment is either too fine or the current too slow to form them.

Within the study area, dune cross-stratification was formed either by reworking sediment deposited during an earlier stage of the same flow event, or by a later flow. Where the cross-stratified sandstone and the underlying part of the bed are of similar grain size the dunes are due to reworking (Lowe, 1982). Although there remains the possibility that the dunes are associated with a later flow, the common foundering of Facies 4 dunes into underlying Facies 3 beds suggests a close temporal relationship between the two, which by extension would be the same event (see Figure 2.7). The depth of foundering, or conversely the lack thereof, is as a first approximation, an indication of the degree of consolidation of the bed at the time of dune migration. In turn, it implies that the structureless sandstone beds of Facies 3 must have dewatered at varying rates, and that the rate varied laterally within a single deposit.

The isolated nature of the sets and the exposure of the underlying strata in dune troughs suggests that the migrating dunes were sediment starved (Tuijnder *et al*, 2009). If dunes migrate over a non-erodible substrate, they are unable to reach equilibrium dimensions and become sediment starved. The spacing between dunes will be regular and similar to the equilibrium spacing, but bedform height will be less than normal (Flemming, 2000). However, if the sediment supply is insufficient to completely mantle the bed surface, isolated dunes form, with irregular spacing and bedform height determined by flow velocity (and flow depth in a fluvial setting) rather than sediment supply (Carling *et al*, 2000). The impoverished supply of coarse sand may be because the dunes migrated over a sandy substrate that hadn't yet completely dewatered, limiting the thickness and volume of material available to be reworked. Whereisolated dunes overlie mudstone or siltstone, the lack of appropriately-sized grains would cause them to be sediment starved.

Staining of thin sections of Facies 4 using the procedure outlined by Dickson (1966) and in the Introduction revealed the cement to be ferroan calcite (stain was royal blue and appeared fractured). The presence of a pervasive ferroan calcite cement in strata of Facies 4 suggests that these deposits were anomalously permeable due to their well-sorted nature. Moreover, the calcite cement fills large pores, indicating that the sediment was deposited with an open grain framework, which has been preserved. This, in turn, suggests that the carbonate-rich fluid that infiltrated these deposits began the process of cementation before compaction took place (i.e. shallow burial diagenesis).

Facies 5: Mudstone-clast breccia

Facies 5 consists of generally abundant mudstone clasts dispersed in a upper coarse sandstone to granule conglomerate matrix, and forms a breccia that is matrix supported in most cases. Clasts usually have a tabular shape with rounded edges and range from 2-50 cm in length and 0.5-10 cm in thickness. In most cases, clasts are unlaminated and dark coloured, although some rare large laminated clasts have been observed (~1 m x 1m). Mudstone clasts occur largely in three places within the study area: (1) at highly erosive basal contacts or scour surfaces (often overlying or interbedded with cross-stratified sandstone), (2) "floating" within structureless sandstone (Facies 3), or (3) at a vertical facies change (sandstone/mudstone contacts, usually at the top or base of sandstone beds).

Interpretation

Mudstone clasts present within deep erosional scours are likely to have been sourced from updip erosion and transported to their current location. Where a mudstone clast breccia is overlying or interbedded with medium-scale cross-stratified sandstone (Facies 4), this may indicate the deposition of scour-lag clasts (Johansson & Stow, 1995). These clasts are eroded from up-flow and tend to be densely deposited in deep scours.

Mudstone clasts "floating" within beds of structureless sandstone (Facies 3) most probably indicate a cryptic internal surface of erosion within the unit. In some places, horizons of mudstone clasts correlate laterally with a discrete siltstone bed between two beds (see Figure 2.12). This indicates amalgamation of two beds and the erosion of mud-rich strata between them (Posamentier & Walker 2006, Johnson *et al*. 2001). The clasts are analogous to the amalgamation clasts of Johansson & Stow (1995).

Mudstone clasts that occur at vertical facies changes (i.e., at the base or near the top of sandstone beds) are likely simple rip-up clasts that have been entrained at the base of or within a turbidity flow. In the rare case of clasts near the top of a sandstone bed, it is possible that the rip-ups were segregated to the top of the flow and remained there due to *en masse* deposition (Haughton *et al*., 2003). Another possibility is that tabular mudstone clasts were transported along a discrete interface between portions of the flow with different densities (Postma *et al.*, 1988). The clasts remain at this interface due to several mechanisms, which in addition to

Figure 2.12: Mudstone clast breccia horizon between two sand beds. Correlates laterally to a thin set of thin-bedded turbidites in one direction and thins in the other direction.

turbulence, include hindered settling, interactions between grains, and the tabular shape of the clasts that reduces their settling rate.

Facies 6: Medium-bedded turbidites

Facies 6 is a normally graded sequence, which stratigraphically upward consists of: (1) moderately-sorted, upper coarse to medium sandstone, (2) parallel-laminated well-sorted medium-grained sandstone, (3) small-scale cross-stratified, well-sorted medium to fine-grained sandstone, and finally capped by (4) laminated siltstone and mudstone.

The stratigraphically lower, graded part of the bed overlies a shallowly-scoured or undulose contact and ranges from 5-20cm thick. Mineralogically they are composed of 40-70% quartz; 5-40% recrystallized clay minerals; 5% partially altered plagioclase feldspar; and minor muscovite, biotite, chlorite, calcite, and pyrite – subarkose composition. In addition, small mudclasts and/or granule-sized quartz clasts are rarely observed.

Typically, sharply overlying the graded sandstone is medium-grained sandstone characterized by mm-scale planar laminae. This interval is composed of approximately 50-60% quartz, 15-30% recrystallized clay minerals, up to 25% calcite, and minor amounts of other minerals (chlorite, muscovite, biotite). The planar-laminated sandstone is better-sorted than the underlying interval and is 3 to 12 cm thick.

The next interval upwards, which ranges from 1- 3 cm thick, consists of well-sorted medium to fine-grained sandstone. It is characterized by a single unstacked set of small-scale cross-lamination with a reddish-brown weathering color. Mineralogically, this sandstone is composed of 50-75% quartz, 20-30% recrystallized clay minerals, up to 10% partially altered plagioclase feldspar, up to 10% calcite, and some minor minerals (muscovite, chlorite, pyrite, biotite – see Figure 2.14). Commonly, this interval overlies a shallow scour surface eroded into the planar-laminated sandstone.

Finally, the succession is usually overlain by a 1 to 4 cm layer of faintly laminated siltstone and mudstone which is rarely capped by a thin veneer of very dark-colored mudstone (see Figure 2.13).

The full sequence, as just described, is rarely present in its entirety in any one individual bed. Typically, one or two of the intervals is missing. In addition, there appears to be no discernable pattern of which intervals occur, or conversely, are missing.

Interpretation

The full sequence described above is equivalent to Bouma's (1962) turbidite sequence (Ta-e). This sequence is deposited by a waning, low-density turbidity current in which deposition occurs first by rapid suspension fallout of the coarsest fraction to form normally graded, coarse-grained sandstone, the Ta division (Posamentier & Walker, 2006; Lowe, 1982). Subsequently, sedimentation rate slows and as a result the bulk of the sand-sized grains are moved along the aggrading bed by traction, producing parallel or small-scale cross-laminations. These are the Tb and Tc divisions, respectively. Although better sorted than the underlying Ta, clay content can vary between 10-35% in these two divisions (Lowe & Guy, 2000), and is similar to percent estimates in this study (see appendix). Parallel laminae in the Tb division are equivalent to upper stage plane bed and have been shown experimentally to form from a few mm-high, migrating bed forms (Best & Bridge, 1988; Arnott & Hand, 1989). The Bouma Tc division's cross-stratified sandstone is due to the migration of a single non-climbing ripple set, and implies a very

Figure 2.13: Stacked beds of Facies 6, mainly Tbcd turbidites

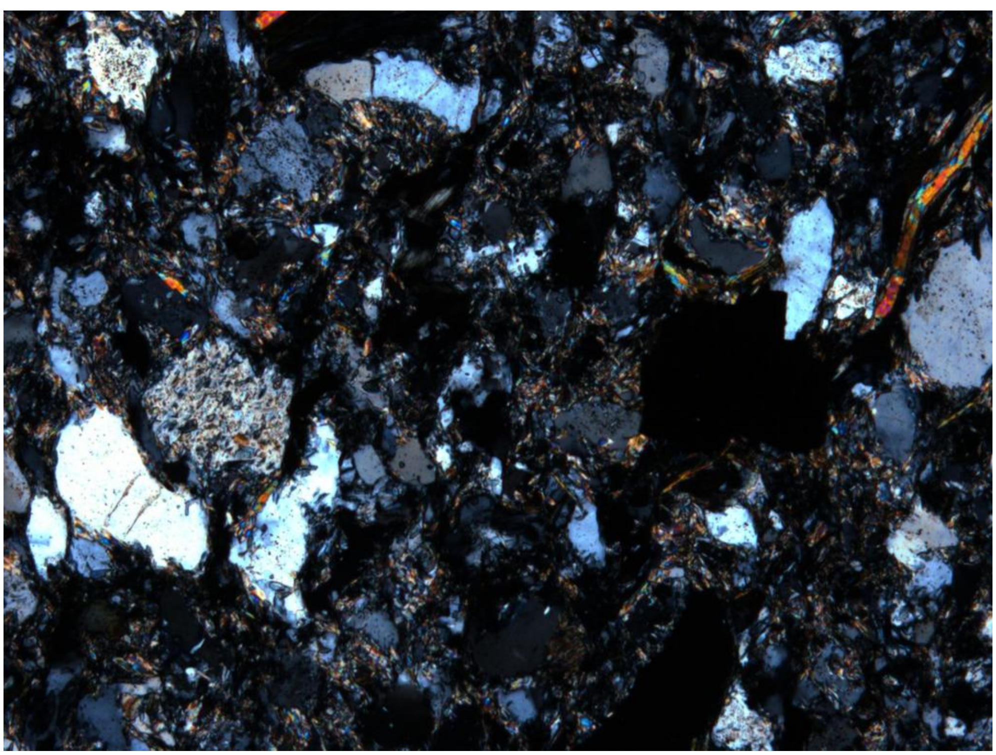

Figure 2.14: Microphoto of a sample of Facies 6, note the abundance of matrix and the smaller average grain size relative to Facies 3. (Microphoto is approximately 1.8 mm wide

low sedimentation rate. The overlying laminated siltstone and mudstone unit is equivalent to the Bouma Td and Te divisions, respectively. Both divisions are deposited largely from suspension, although there are some near-bed processes that give rise to the faint lamination present in Td intervals (Lowe, 1982). One possible mechanism is the formation of flocs by clay particles in suspension. These flocs would have the same settling velocity as the suspended silt grain. As these grains reach the near-bed layer, shear forces disaggregate the flocs and release their constituent clay particles back into suspension, allowing only the silt to deposit (Stow & Bowen, 1978). Eventually, a sufficiently high concentration of suspended clay particles overcomes the shear forces near the bed and a mud layer is deposited rapidly (suggested by the absence of silt grains in this layer). This results in the alternating silt and mud lamina characteristic of the Td division. The Te division is composed purely of hemipelagic and or pelagic mud and is assumed to deposit entirely from suspension sedimentation (Posamentier & Walker, 2006).

The presence of incomplete Bouma sequences usually involving the absence of either the Tb or Tc divisions suggests that the processes of traction sedimentation may have been suppressed or that flow conditions were not ideal. If a turbidity current is moving too slowly, upper plane bed would not form, resulting in a succession such as a Tacde turbidite. A succession without ripples could result from flow speeds that are too high and/or bypass, or subsequent erosion of the bed. If high sediment concentrations were present near the bed to suppress traction transport, the formation of planar laminae would be prevented even if flow speeds are within their stability field.

Facies Associations

Based on geometry, lateral & vertical facies changes, and unit contacts, five facies associations, or architectural elements, have been identified in the study area. Each element consists of two or more of the facies described in the previous section (see Table 2.2). A comparison between dimensions of these architectural elements measured in the study area and literature examples is presented in Table 2.3.

Facies Association 1: Isolated Scours

Facies Association 1 is composed largely of thick-bedded massive sandstone (Facies 3) with some mudstone clast breccia (Facies 5), medium-bedded turbidites (Facies 6) and rare medium-scale cross-stratified sandstone (Facies 4).

Units of Facies Association 1 typically form interbeds within fine-grained intralobe deposits (F.A. 2). Units are up to 5 m thick with scoured, highly erosive basal contacts and are laterally continuous over 100's of metres (see Figure 2.20). Beds are up to 2 m thick with common quartz granules or mudstone intraclasts at their base. Locally, mudstone intraclasts are evenly dispersed. Beds are typically ungraded or coarse-tail graded with uncommon carbonate-cemented pillar structures. Dune cross-stratified sandstone occurs either as isolated beds or at the base or top of a Bouma Ta/Lowe S3 bed (Facies 3). Typically beds consist of a single dune set, but rare stacked cosets (up to 2 sets thick) are observed.

#	Facies Association	Constituent Facies	Thickness	Description	Interpretation
1	Isolated scours	3, 4, 5, 6	≤ 5 m	Form units ≤ 5 m thick in a succession of thin-bedded intralobe turbidites. Scours are filled with thick-bedded, coarse-grained sandstone with some mudstone-clast breccia and rare interbeds of medium-bedded turbidites and/or dune cross-stratified sandstone. Laterally continuous over 50 - 300 m.	Isolated scours created by large turbulent eddies at a location where flow confinement decreases dramatically. These scours were later infilled by succcessive lag deposits from bypassing flows.
2	Heterolithic feeder channel deposits	2, 3, 4, 5, 6	2.5 m - 16.5 m	Nested, erosionally based stratal units, each filled with a unique heterolithic assemblage of lithofacies. Unit 1 (Facies 4 & 5), Unit 2 (Facies 2 & 6) and Unit 3 (Facies 2 with rare beds of Facies 3).	Erosionally-confined lobe feeder channel . These channels convey flows to the downflow depositional lobes.
3	Thin-bedded intralobe turbidites	1, 2, 4, 6	10 m - 30 m	Thick intervals of thin-bedded turbidites interbedded with rare medium-bedded turbidites and isolated dune cross-stratified sandstone beds. One unit of this facies association contains a debrite. Tabular geometry, with negligible lateral facies change.	Local shutdown (avulsion) of the sediment transport system. Deposition by infrequent, dilute, mud-rich turbidity currents.
4	Terminal splay deposits	3, 4, 5	5 m - 13 m	Composed largely of amalgamated coarse sandstone beds 0.8 - 4.5 m thick. Basal bedding contacts are planar or shallowly scoured, occasionally with a horizon of mudstone-clast breccia that correlates laterally to a mudstone parting or dune horizon. Quartz granules and flame structures may also be present at the bases of beds. Dewatering structures are present. Tabular geometry, little or subtle lateral facies change.	Sheetlike sandy terminal splay formed by flow expansion at the terminus of a distributary channel. Local depocentre for the lobe complex and is typically the site of rapid deposition.
5	Distributary channel deposits	2, 4, 6	2 m - 6.5 m	Intercalated with terminal splays. Composed of one or more upward-fining units 1 - 3 m thick that are typically as follows: one or two partial to full Bouma sequence turbidites, followed by an interval of thin-bedded Tcd turbidites with rare dunes and or Tb/Tbc turbidites. Irregular to tabular geometry, ratio of sand:mud may change laterally.	Fill of distributary channels that feed downflow terminal splays. Heterolithic upward-fining fill indicates bypassing flows followed by gradually decreasing activity. Proximal parts may resemble terminal splays, but are less tabular.

Table 2.2: Summary of the 5 facies associations identified within the study area.

#	Architectural Elements	Measured Dimensions	Literature Dimensions	Source
1	Isolated Scours	Thickness: ≤ 5 m Width:100 - 250 m	Thickness: < 10 m Width: > 50 m	Wynn *et al*., 2002
2	Feeder Channel	Thickness: 16.5 m (max) Width: ≥ 800 m (outcrop width)	Thickness: 8- 80 m Width: 100 - 800 m	Saller *et al*, 2008 ; Wynn *et al*., 2002
3	Intralobe Turbidites	Thickness: 10 - 30 m Width: n/a	Thickness: n/a Width:n/a	n/a
3	Distributary Channels	Thickness: 2 - 6.5 m Width: Unable to isolate individual channels	Thickness: < 15 m Width: < 100 m	Saller *et al*., 2008
5	Terminal Splays	Thickness: 5 - 13 m Width:100s of m (width of outcrop)	Thickness: avg of 9.5 m Width: 400 - 3000 m	Saller *et al*., 2008

Table 2.3: Comparison between measured strata elements and those observed in modern systems.

Units similar to Facies Association 1 are rarely described in the literature. In part this may be because their thinness makes them difficult to resolve in most seismic images and their association with fine-grained lithologies most probably causes them to be poorly exposed. However, scours observed at Castle Creek resemble large-scale isolated scours observed in seismic images of the channel-lobe transition zone on the modern Rhone Fan and also features near the basinward terminus of both the Lisbon and Agadir canyons (Wynn *et al*, 2002). Dimensions of features identified as isolated scours in ancient outcrop examples from Italy's Piedmont Basin and others, are 1-5 m deep and 50 m wide, and in modern examples are inferred from seismic to be < 10 m deep and 100's of m wide (Wynn *et al*., 2002). Wynn and others also observed that in outcrop, such features would resemble shallow channels filled with irregularly bedded and amalgamated sandstones. Similarities in characteristics between these examples from the literature and strata of F.A. 1 observed at Castle Creek are easily drawn, and include similarity in dimensions, scoured basal surface and a fill of laterally discontinuous beds of amalgamated sandstone.

As a result, units of F.A. 1 are interpreted to be shallow isolated scours infilled by coarse lag deposits of bypassing flows that partially removed and/or reworked a portion of a previous lag deposit. These isolated scours formed in the **T**ransition **Z**one, or CLTZ, between slope-**C**hannel complexes and basin floor depositional **L**obes. In that part of the system, rapid loss of confinement would lead to rapid flow expansion, creating large, turbulent eddies that would incise the local basin floor. While the CLTZ is largely a zone of bypass, these newly created topographic lows, or channels, allowed for at least some limited net deposition.

Figure 2.15: Example of an isolated scour. Note the highly erosive basal contacts and the lack of structure. Stratigraphic top is to the left.

Facies Association 2: Heterolithic feeder channel deposits

Facies association 2 is highly complex and composed of three stacked intervals. The lowermost interval consists of dune cross-stratified sandstone (Facies 4) interstratified with mudstone clast breccia (Facies 5). This, then, is overlain by an interval of medium-bedded turbidites (Facies 6) interbedded with laminated siltstones and mudstones (Facies 2). The uppermost interval consists of laminated siltstones and mudstones (Facies 2) with rare isolated dunes (Facies 4).

To date, Facies association 2 has been recognized in only two locations in strata of the Upper Kaza at Castle Creek, and only one of those occurs within the study area. This example ranges from 2.5 to 16.5 m thick (thickens from SE to NW) and can be traced laterally across the width of the outcrop (~800 m). The interpreted channel form is not completely contained in the extent of the outcrop implying that it is either significantly wider, or that the outcrop intersects the channel obliquely. Although confined to two paleocurrents measured from dune troughs (221º & 251º) the outcrop appears to be oriented slightly oblique to the trend of the channel.

Facies association 2 is interpreted to represent the fill of a deeply incised (trunk) feeder channel that conveyed sediment to a downflow depositional lobe complex. Although rarely reported in outcrop, feeder channels have been described from seismic and include the intralobe distributary channels of the East Kalimantan fan (100-300 m wide, 8.2-16.8 m thick; Saller *et al*., 2008) and the main lobe complex feeder channel in the channel-lobe transition zone of the modern Rhône Fan (500-800 m wide, 10-80 m deep; Wynn *et al*., 2002). Although dimensions of the trunk channel studied at Castle Creek are similar to these seismic examples, details about the nature of their fill are lacking. Nevertheless, work at Castle Creek suggests that trunk channels are filled with a heterolithic assemblage of poor reservoir quality (low net to gross), mud-prone strata.

The base of this facies association is marked by a deep, throughgoing erosion surface that locally is up to 14 m deep and erodes an underlying amalgamated sandstone unit that is ~10 m thick. The basal surface has a terraced morphology that suggests at least four distinct episodes of incision. The most spectacular example of this occurs between sections Ms6 and Ms7, where the base incises 7 m deep over 16 m laterally (see Figure 2.15). Commonly overlying the scoured base is a unit of laterally discontinuous dune cross-stratified sandstone interbedded with mudstone clast breccia and some rare thin-bedded turbidites that range from 1.5 to 8 m thick (see Figure 2.16). This is interpreted to be a complex lag deposit emplaced when the channel was an active sediment-transport conduit with flows bypassing most of their sediment load to the more distal depositional lobe complex. Mudstone clasts within the breccia vary greatly in size and are subangular to subrounded, with some retaining a tabular shape, implying a relatively short distance of transport between updip erosion and deposition at their current location. Clasts in the intraclast breccia were most probably not sourced from slumps along the channel margins because they consist mostly of sand and show little evidence of internal deformation. This implies that they are sourced entirely from updip erosion. The very coarse-grained, dune cross-stratified sandstone, on the other hand, implies episodes of reworking of previously deposited lag deposits by the tails of bypassing flows.

Immediately overlying the lower unit is a 1 to 10 m interval of interstratified medium-bedded turbidites and laminated siltstones and mudstones. The basal contact of this middle unit

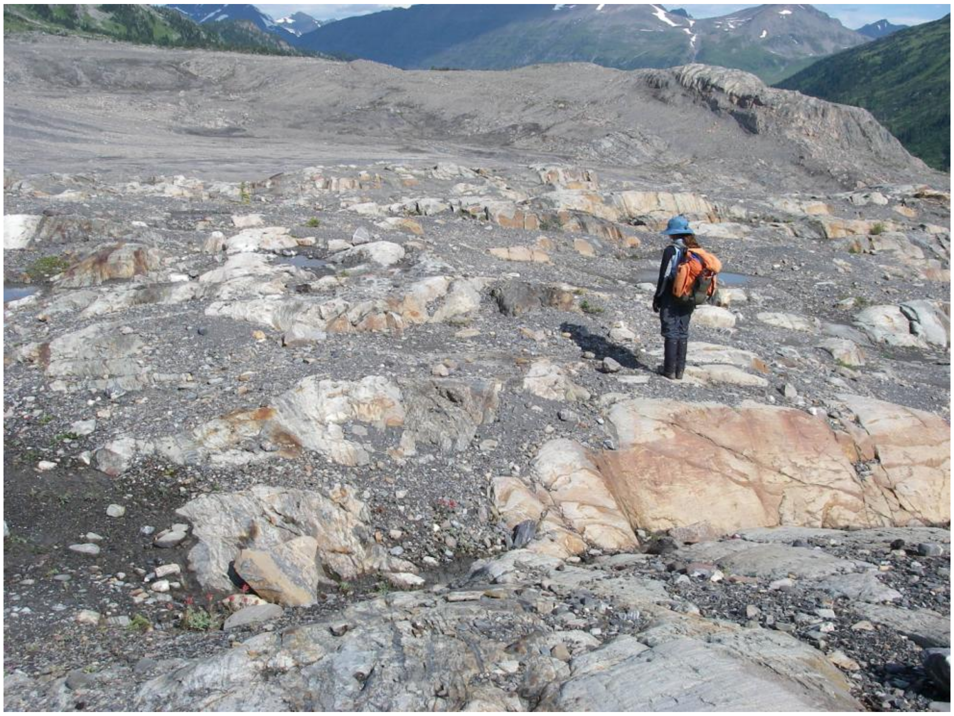

Figure 2.16: A major downcutting surface at the base of the feeder channel unit. The surface is partly defined by the intraclast breccia lag deposit and continues near the feet of the person in the mid-ground. Stratigraphic top is to the left.

Figure 2.17: A portion of the complex lag deposit at the base of the feeder channel. In this location, it is composed of mudstone-clast breccia and locally dune cross-stratified sandstone.

is a scour surface that locally erodes the underlying lag deposit completely (up to 5 m deep over a lateral distance of 26 m). This terraced surface represents a reactivation of the channel and a period of full bypass. The lack of extensive breccia and coarse-grained dune cross-stratified sandstone indicates that the channel began to backfill with deposits from low-density turbidity currents sometime afterwards. These strata onlap the basal erosion surface and extend ~200 m laterally to the northwest limit of the exposure.

These strata are then sharply overlain by a heterolithic assemblage 2.5 to 6 m thick of thin-bedded turbidites interstratified with lenses of coarse-grained sandstone up to 1 m thick and a few tens of meters long. Locally theses sandstones are dune cross-stratified. The base of this uppermost unit is irregular and includes a 3 m-deep scour over a lateral distance of 47 m. Stratigraphically-upward the upper interval fines, which is marked by an increasing abundance of thin-bedded Tde and Tcde turbidites with fewer intercalated lenses of coarse sandstone. The abrupt change in the nature of sedimentation beginning at the base of this unit indicates that the channel ceased to be a major sediment transport pathway for a downdip depositional lobe complex, possibly as a result of an upflow channel avulsion. Figure (2.17) suggests that the channel had become backfilled almost to the level of the adjacent extrachannel seafloor, and as a consequence reduced its hydraulic efficiency (i.e. favorability). The abrupt facies change beginning immediately above the irregular basal surface however, suggests an interval of full sediment bypass and erosion, followed by subsequent turbidity currents that lacked sufficient energy to erode further into previously deposited sediment. Southeastern portions of the irregular basal surface, particularly where it is also the base of the channel, may be due partly to initial

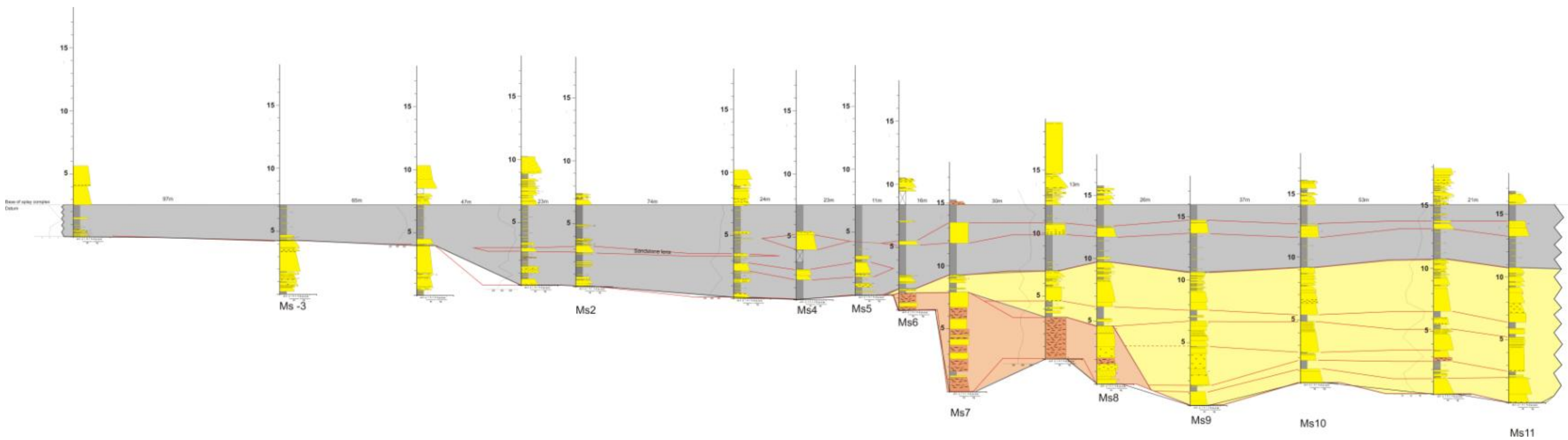

Figure 2.18: Correlation panel for the one example of Facies Association 2 in the study area at Castle Creek. Note the terrace-like geometry of the basal surface and the three distinctly different depositional intervals highlighted. Basal lag deposit (orange), backfill deposits (yellow), and abandonment deposits (grey).

incision of the channel. The upward fining trend in the unit and the general paucity of coarse-grained sediment is interpreted to represent a progression toward full abandonment of the channel. Lenses of coarse-grained sandstone represent rare high energy sandy flows during a period of otherwise low energy turbidity currents and hemipelagic sedimentation. This likely indicates a deactivation of the channel and downflow lobe complex, first by an initial upflow channel avulsion that abruptly terminated the supply of coarse-grained sediment, followed thereafter by a gradual decline in the frequency and/or size of turbidity currents or major shut-down in the upslope sediment staging area.

Facies Association 3: Thin-bedded intralobe turbidites

Facies association 3 is composed mostly of thin-bedded siltstones and mudstones (Facies 2), interbedded uncommonly with medium-bedded turbidites (Facies 6), rarely with isolated medium-scale cross-stratified sandstone (Facies 4) and poorly sorted clast-rich mudstone (Facies 1).

Units of Facies association 3 typically sharply overlie distributary channel units (F.A. 5), and are sharply overlain by either distributary channel units or lobe deposits (F.A. 4). Units of Facies association 3 are 10 to 30 m thick and composed mostly of thin Tcde/Tde turbidites interbedded with uncommon and laterally discontinuous, isolated sandy Tbc turbidites up to 15 cm thick. Bed contacts are generally non-erosional, although shallow scours up to 15 cm deep occur locally. These scours are typically filled with mudstone-clast breccia (Facies 5) or medium-grained dune cross-stratified sandstone, both of which tend to be less than 30 cm thick. Debrites are rare but occur as two end member kinds; the first is dominated by granule and uncommon pebble quartz clasts and are 15 – 30 cm thick, and the second type, which is isolated to a single occurrence, is up to 4 m thick and extends the full width of the outcrop (~800 m). This thick debrite is silt-rich and contains clasts of calcite-cemented sandstone as well as large laminated mudstone clasts.

Facies association 3 is interpreted as intralobe deposits. The occurrence of a thick succession of thin-bedded turbidites suggests a long period of relative quiescence during which silt-dominated, low-density turbidity flows and suspension deposition prevailed (e.g. Andersson & Worden, 2004). Rarely, however, these conditions were interrupted by sandier, higher energy turbidity currents that deposited isolated, medium-bedded turbidites. In addition, even rarer bypass flows eroded previous deposits and left behind thin localized lags of mud-clast breccia or trains of starved dunes. Both kinds of debrite contain clasts that are generally not observed in basin-floor deposits in the study area. For example, quartz granules and pebbles are rare but do occur in a single stratal unit at the top of the study area. These observations imply that the debrites consist, at least in part, of sediment sourced from an upflow, more proximal part of the Windermere transport system.

The abrupt superposition of thin-bedded, fine-grained strata of F.A. 3 indicates a dramatic decrease in sediment gravity flow energy, which possibly relates to a local shutdown of the transport system due to upflow trunk channel avulsion or submarine canyon switching. Alternatively the shutdown may reflect a larger scale deactivation caused by a eustatic rise and a dramatic reduction of the entire deep-water turbidite system. Attendant slope instability related to such a change may explain the occurrence, albeit uncommon, of debrites containing exotic clasts.

Facies Association 4: Terminal splay deposits

Facies association 4 is composed mostly of thick-bedded massive sandstone (Facies 3), interbedded with rare medium-scale cross-stratified sandstone (Facies 4) and mudstone clast breccia (Facies 5).

Units of Facies association 4 typically occur overlying and/or underlying distributary channel deposits (F.A. 5, see next). Units of Facies association 4 are 5 to 13 m thick with basal contacts that are either horizontal or shallowly scoured (usually less than 10 cm deep over 1 or 2 m laterally). These units are laterally continuous over the width of the outcrop (≤ 800 m), although lateral changes in sandstone to mudstone (e.g. net to gross) are common. In addition, units consist of one or more subtle upward-fining intervals 0.8 to 4.5 m thick separated by irregular scour surfaces marked by an abrupt increase in grain size, occasionally accompanied by a thin horizon of mud-clast breccia. These intervals are composed mostly of massive or coarse-tail graded sandstone beds 0.7 to 2.5 m thick that vary in amalgamation laterally (see Figure 2.18). Dewatering pipes are common, as are flame structures along bedding contacts. Most horizons of mudstone clast breccia correlate laterally to mudstone partings between sandstone beds where beds are less amalgamated. Rare diffuse planar lamination or medium-scale cross-stratified sandstone is present at the top of these units.

Facies association 4 is interpreted to represent terminal splay deposits that display a sheetlike geometry at outcrop scale. These are interpreted to be equivalent to sheetlike splay elements identified on seismic images from lobes in the East Kalimantan fan (Saller *et al*., 2008). These elements are sand-rich features that form part of a larger-scale lobe complex and are estimated to have an average maximum thickness of about 9.5 m, which is similar to the interpreted splays in the study area.

The abrupt replacement of a precursor unit of distributary channel deposits by those of Facies association 4 may indicate a rapid switching of splay deposition within the local lobe complex. The abundance of amalgamated beds of massive sandstone and the presence of dewatering pipes and flame structures implies high rates of local deposition, most probably from high concentrations of suspended sediment. Horizons of mudstone clast breccia are the remnants of thin fine-grained layers that draped the seafloor and were eroded by a later flow. These are equivalent to the amalgamation clasts of Johansson & Stow (1995). The widespread lack of classical turbidites and the dominance of the lowermost Ta interval is possibly due to the removal of the top part of the previous bed by the next incoming flow, making almost every upper bed contact a potential scour surface. Mudstone partings between beds, where present, indicate a period of time between depositional events, potentially following removal of the largely absent Tb and Tc intervals and a period of full bypass. Rare medium scale cross-stratified sandstone at the top of splay units indicates a reworking of the uppermost parts of the deposit. The usually abrupt shift to distributary channel deposits implies a change in location of the main depocenter and an increasing volume of bypassing sediment.

Each splay deposit was the major depocenter of an active lobe. Location of the active splay(s) may depend on the migration of distributary channels in response to the infilling of seafloor topography.

Figure 2.19: An example of a sand-rich terminal splay near the axis of deposition. Note the total lack of fine-grained sedimentary rock within the interval pictured. Stratigraphic top is to the right.

Facies Association 5: Distributary channel deposits

Facies association 5 is composed largely of medium-bedded turbidites (Facies 6) interbedded with medium-scale cross-stratified sandstone (Facies 4) and thin-bedded siltstone and mudstone (Facies 2).

Units of Facies association 5 typically overlie and/or underlie terminal splay deposits (F.A. 4). Units are 2 to 6.5 m thick with sharp, generally horizontal basal contacts (see Figure 2.19) and usually made up of one or more upward-fining sequences. These sequences are approximately 1 to 3 m thick and in general comprise two or more partial to full Bouma turbidites (Facies 6) that are 0.3 to 0.8 m thick with shallowly scoured basal contacts (< 10 cm depth over several meters laterally). Overlying these is an interval composed mostly of thin-bedded Tcd turbidites with uncommon interbeds of Tb/Tbc turbidites, and rare single-set dune cross-stratified sandstone (Facies 4 & 6, respectively). Where present, Ta divisions are 5 to 40 cm thick, typically well-graded, with local mudstone chips and/or clasts at the base. Single sets of medium-grained, dune cross-stratified sandstone are rarely present at the top of the Bouma Ta division in the beds mentioned previously. These dune sets commonly founder into the underlying bed (usually no deeper than 10 cm), although the depth of foundering often varies over a few meters laterally in the same bed (see Facies 4). The Ta division may also contain rare flame structures at its base or pillar structures internally. Tb and/or Tc divisions are almost always present, even in partial turbidites. The Td division is often preserved and may persist laterally for up to 100 m before being eroded by an overlying bed.

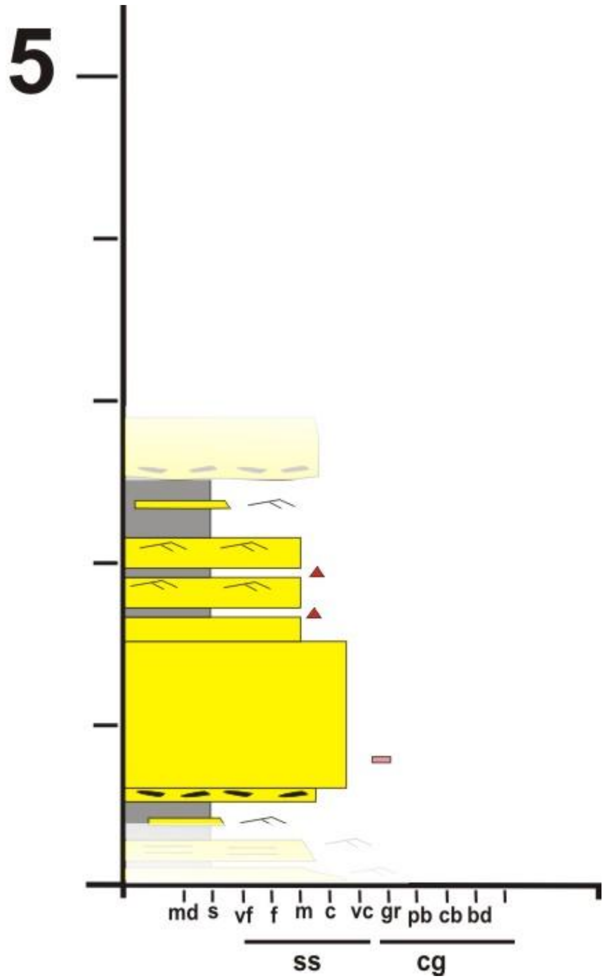

Figure 2.20: An example of one of the fining-upwards packages that compose units of F.A. 5. Note the mudstone breccia at the base.

Facies association 5 is interpreted to be the fill of intralobe distributary channels, which which transition downflow into terminal splay deposits (Facies Association 4). Seismic images of terminal splays show that they are commonly mantled, generally at their proximal end, by a network of small channels, which tend to be sinuous or even braided (Posamentier & Walker, 2006). Channels are typically 100 m or less in width, and less than 15 m deep (Saller *et al*., 2008). Although portions of the outcrop in the study area extend laterally for up to 800 m, local cover makes recognizing individual distributary channels problematic. In addition, the outcrop trend (~330°) is slightly oblique with respect to the average paleoflow of approximately 250°. However, no interval of Facies association 5 even approaches the thickness of distributary channel fills imaged in seismic so they may represent a more distal expression of the same feature.

Multiple upward-fining units within Facies association 5 indicates that individual channels within the distributary network may be reactivated multiple times, possibly as a response to infilling of topography lows. The relative thinness of these intervals suggests that either most of the sediment bypassed the channels into the terminal splay or that individual channels are active for only a relatively short period of time.

The abundance of partial to full Bouma sequence turbidites indicates deposition mostly by moderate to low-density turbidity currents. Flame and pillar structures in some Ta divisions suggest high rates of deposition, at least locally. Moreover, dune cross-stratified sandstone beds indicate the re-working of underlying deposits, most probably by bypass flows. Lateral changes in the sand/mud ratio of strata of Facies association 5 are likely an indicator of proximity to the distributary channel axis. Thin-bedded turbidite intervals represent a period of relative quiescence due to distributary channel migration or avulsion. Where these deposits are overlain by interlobe mudstones (F.A.3), the implication is that there was a full deactivation of the local splay/lobe complex.

3: Synthesis

Overview of Deepwater Systems

The classic model for a deepwater turbidite system consists of three main parts: shelf, slope, and basin floor (see Figure 3.1). The shelf forms a featureless, almost horizontal surface, which basinward steepens to about 0.8º to 2.1º on the slope and thereafter gradually diminishes to approximately 0.1º at the most distal edge of the basin floor (Saller *et al*., 2008; Pirmez *et al*., 2000). Sediment sourced from continental weathering is first deposited on the shelf upon reaching the marine environment. Where that deposition takes place in relation to the shelf-slope break depends on several factors, most notably the position of relative sea level, tectonic setting, and proximity of submarine canyon heads to deltaic and littoral transport systems (Covault *et al*., 2007). Where and when conditions are favourable, sediments are delivered directly to the shelf-slope break, marking a time of elevated sedimentation, particularly of sand and gravel, in the adjacent deep-marine.

Submarine canyons are the principal conduit through which clastic sediment is initially transported downslope and eventually out onto the basin floor. Canyons can be over ten kilometres wide and up to a kilometre deep and fully confine the through-going mass-wasting and sediment-gravity flows (Posamentier & Walker, 2006). Downslope, canyons shallow and flows eventually overspill their lateral confinement, forming channels with laterally extensive depositional levees.

Leveed channels are the dominant sand-rich stratal element on the lower part of the slope. Channels vary from a few hundred metres to a few kilometres in width (Posamentier & Kolla, 2003), and typically are moderately to highly sinuous in planform. Channels are bordered by

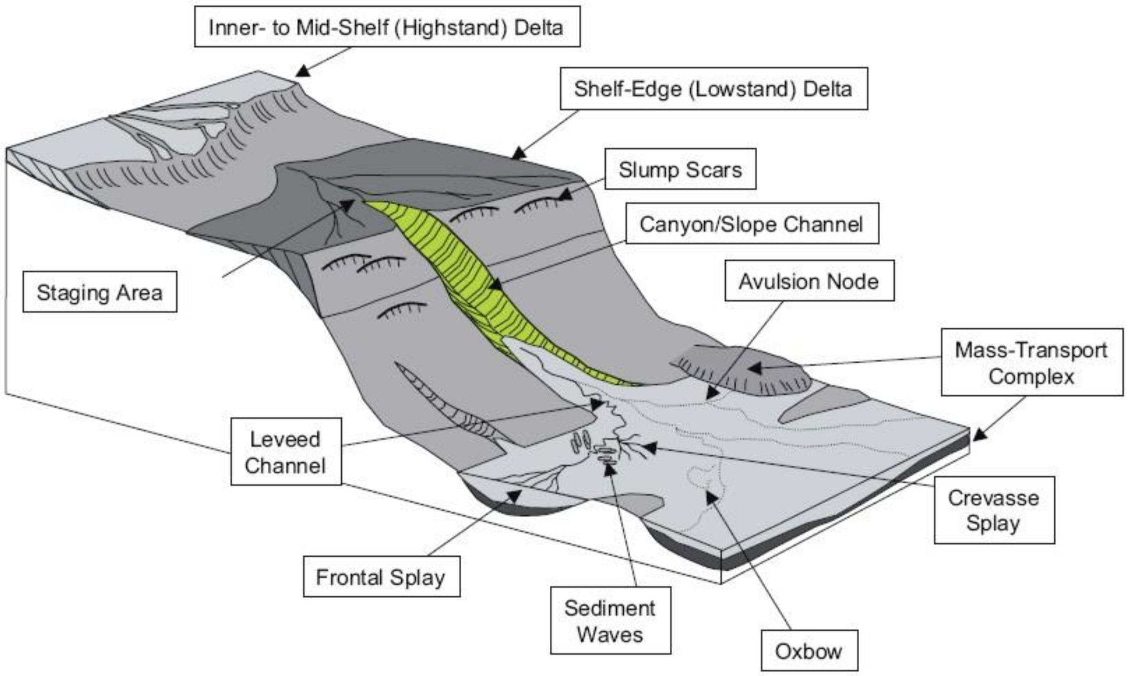

Figure 3.1: Schematic of a passive margin submarine fan system with major sedimentary components labelled (from Posamentier & Walker, 2006).

well-developed levees, which are laterally extensive wedges of generally fine-grained sediment formed by overspill of the upper, comparatively dilute part of turbidity currents passing though the adjacent channel. In contrast to fluvial levees which are a few metres high and few to several tens of metres wide, levees of submarine channels are commonly several kilometres wide and tens to hundreds of metres high (Deptuck *et al*., 2004; Khan, 2010, pers. comm.). This disparity in scale is due to the difference in density between the current and the surrounding ambient fluid. In the case of a marine turbidity current, the density contrast is very small, and as a consequence enhances buoyancy that allows the upper surface of the flow to be elevated far above that of a fluvial system. Moreover, greater buoyancy also allows submarine overspills to travel and deposit much farther from their parent channel.

Further basinward, the constant upslope overspill and progressive loss of fine-grained sediment causes the channel-bounding levees to become gradually reduced in height. Now on the basin floor the flow is confined only by the depth of channel incision. Nevertheless channels remain large, up to several hundreds of meters wide and a few tens of meters deep (Saller *et al*., 2008, Wynn *et al*., 2002). These channels act as the main conduit of sediment to downflow depositional lobes and are characterized by almost full sediment bypass while active. Downflow, flow in the feeder channel expands radially into a network of small, shallow (100-300 m wide, 8-16 m deep) channels termed a distributary channel network. This network of channels is similar to the planform in a river delta but here channels progressively shallow in the downflow direction. Eventually flows become fully unconfined, leading to very rapid expansion and deposition of a terminal splay (see Figure 3.2). Terminal splays are sand-rich mounds 6 to 13 metres thick and a few hundreds of metres to a few kilometres long and wide, respectively (Saller *et al*., 2008). During splay deposition, finer-grained sand, silt, and mud accumulate in the

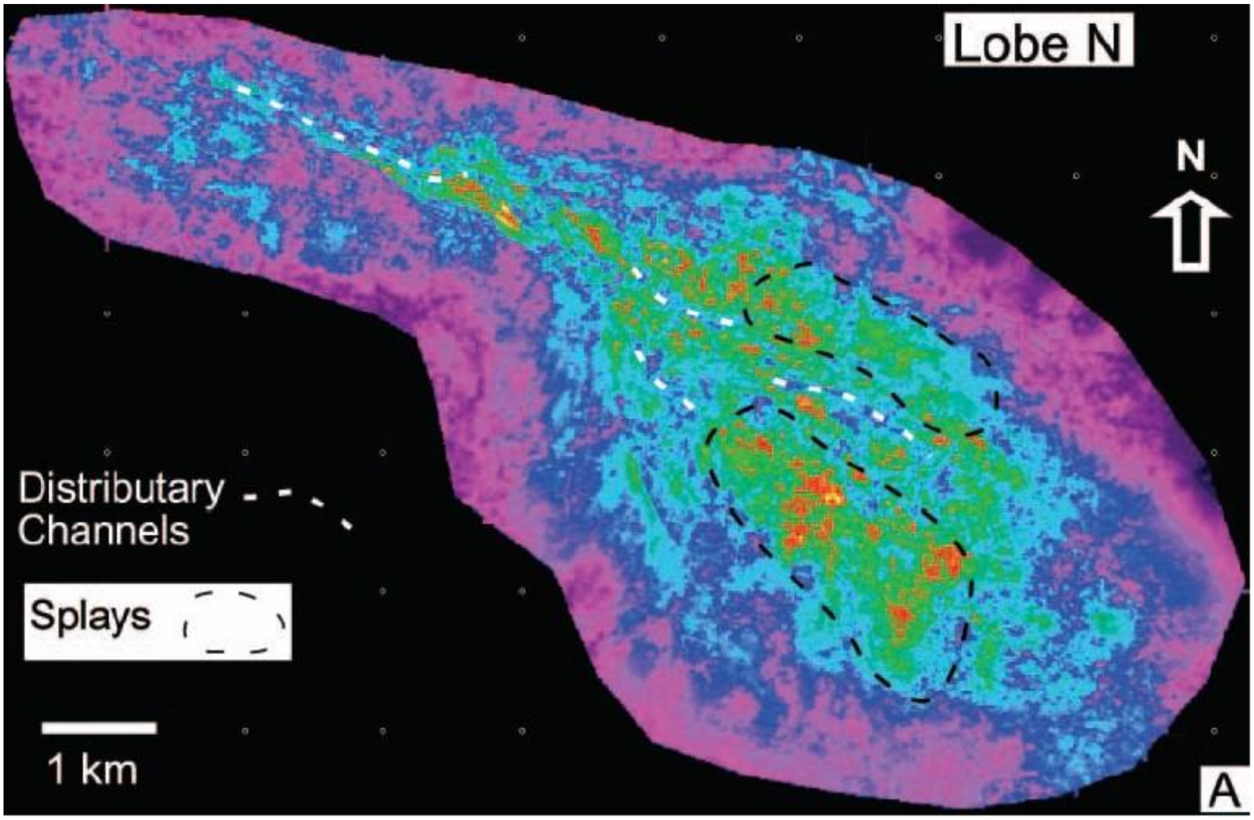

Figure 3.2: Seismic time-slice of a single depositional lobe showing a simple distributary channel network and two terminal splays (dashed line ovals). Colours indicate abundance of sand; orange and red indicate sand-rich deposits, indigo and purple are mud-rich (taken from Saller *et al*., 2008).

area between splays. These are collectively termed intrafan mudstones (Andersson & Worden, 2004), but in this thesis called intralobe deposits.

On the basin floor, the entire system of sediment transport and deposition, from the point a leveed channel becomes an erosionally confined feeder channel to the downflow terminal splay, form the building blocks of a depositional lobe. Lobes comprise multiple terminal splay deposits that stack compensationally and form a depositional element that ranges from 20 to 50 metres thick and 3 to 9 kilometres long and wide (Saller *et al*., 2008). Periodically the feeder channel (and by extension the distributary channel network) migrate laterally, shifting the locus of deposition and initiating the formation a new lobe. Multiple stacked lobes form a lobe complex, or basin floor fan, which can span across an area of hundreds of square kilometres.

Synthesis of Outcrop Data

Deposits of four of the elements of a deep marine fan system described above were identified in the Castle Creek study area: feeder channel (erosional channel), intralobe deposits, terminal splay deposits, and distributary channel deposits. In addition, a fifth depositional element not usually described in conventional models, isolated scours, was identified.

Within the study area, a single feeder channel was identified at the base of the section. It extends across the outcrop (~800 m) and ranges in thickness from 2-3 m (at the SE end) to ~16 m (at the NW end). The width of the channel, however, is unknown; firstly because only one channel margin is partly exposed and secondly the paucity of paleocurrent data precludes an estimate of the orientation of the channel in relation to the strike of the outcrop. Nevertheless, similar features interpreted from seismic images are of the order of 250 to 500 metres wide (Saller *et al*., 2008). The channel is characterized by a three part fill consisting of a basal lag deposit, backfill deposits, and a mud-rich drape, and is similar to the heterogeneous three-phase reported from erosionally-confined turbidite channels by Labourdette (2007). The following describes and illustrates (Figure 3.3) the sedimentary history, comprising alternating episodes of erosion and deposition, of feeder channel strata in the study area:

1. upward-fining trend that may indicate a progressive decrease in the energy of flows in the channel.
2. Another episode of channel widening (by at least 100 metres laterally) and erosion of the upper part of the backfill deposits. This is interpreted to indicate a second major reactivation of the channel system by energetic bypass flows. This second episode of channel reactivation, along with the first, created the terraced morphology observed along the base of the feeder channel.
3. Third episode of channel fill, in this case by mostly low energy flows and hemipelagic sedimentation punctuated by rare high energy flows. The succession is 2.5 to 6 m thick and made up of thin-bedded turbidites interstratified by three shallowly scoured lenses of coarse-grain sandstone up to a metre thick. A drape of mostly thin-bedded turbidites appears to overlap the SE edge of the channel margin and represents the deactivation of the channel due to an upflow avulsion leading to a spatial shift in the main locus of sediment transport and deposition. The intercalated lenses of coarse-grained sandstone most probably represent three discrete episodes, although each consisting of multiple flow events, of partially diverted flows from a laterally adjacent focus of flow activity.
4. Rejuvenation of the channel system causing the complete removal of the basal lag deposits in the northwestern part of the study area and widening of the channel by at least

15 metres (outcrop dimensions, see correlation panel in Chapter 2). This is interpreted to be another period of sediment bypass.

5. Deposition of backfill deposits up to 10 metres thick, consisting of mostly medium-bedded turbidites interstratified with uncommon thin-bedded turbidites. There is a weak upward-fining trend that may indicate a progressive decrease in the energy of flows in the channel.
6. Another episode of channel widening (by at least 100 m laterally) and the erosion of the upper part of the backfill deposits. This is interpreted to indicate a second major reactivation of the channel system by energetic bypass flows. This second episode of channel reactivation, along with the first, created the terraced morphology observed along the base of the feeder channel.
7. Third episode of channel fill, in this case by mostly low energy flows and hemipelagic sedimentation punctuated by rare high energy flows. The succession is 2.5 to 6 m thick and made up of thin-bedded turbidites interstratified with three shallowly-scoured lenses of coarse-grained sandstone up to a metre thick. A drape of mostly thin-bedded turbidites appears to overlap the SE edge of the channel margin and represents the deactivation of the channel due to an upflow avulsion leading to avulsion leading to a spatial shift in the main locus of sediment transport and deposition. The intercalated lenses of coarse-grained sandstone most probably represent three discrete episodes, although each consisting of multiple flow events, of partially diverted flows from a laterally adjacent focus of flow activity.

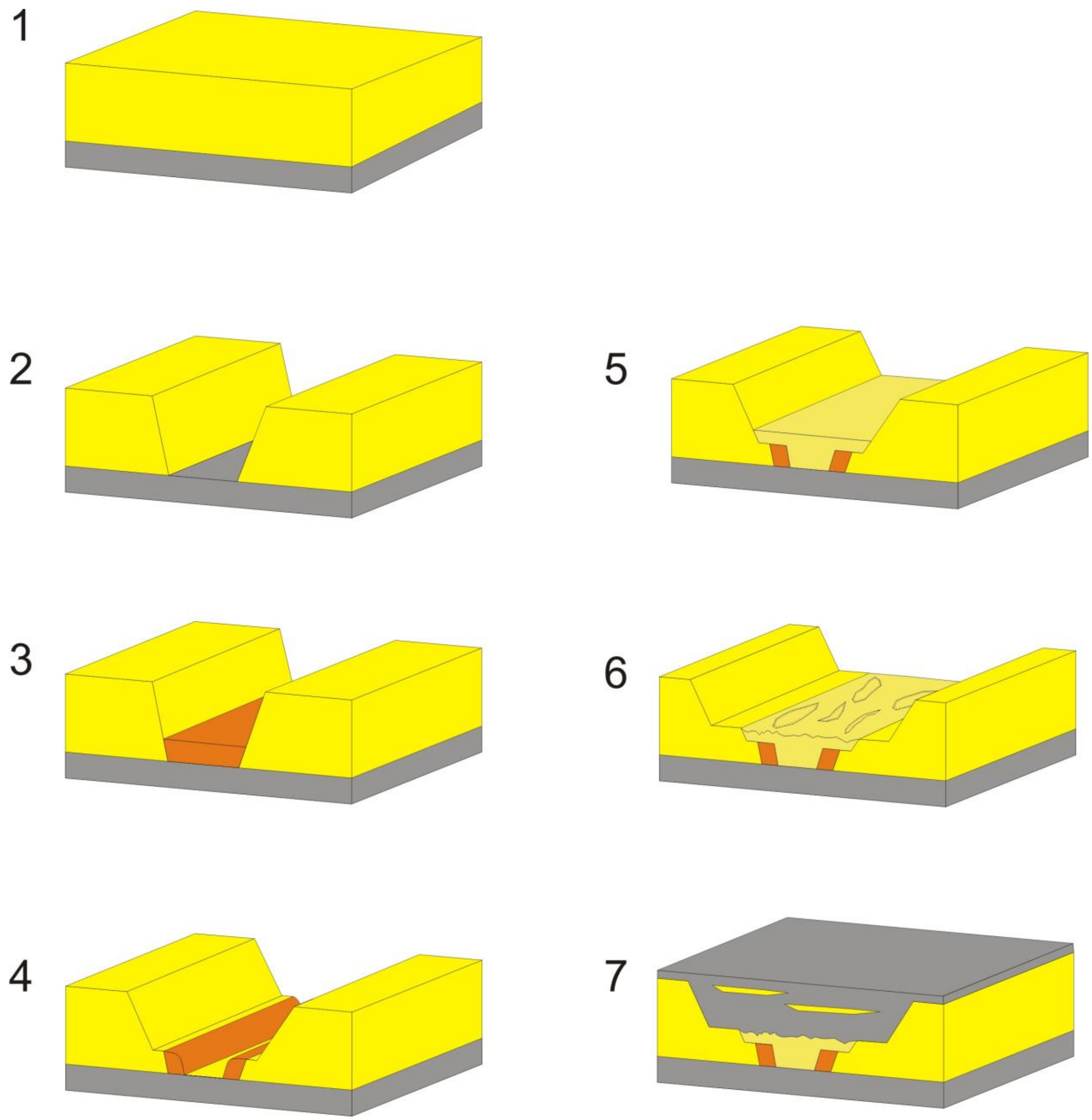

Figure 3.3: Interpreted history of feeder channel incision and fill. Note that the diagram is not to scale. Yellow = sand-rich splay deposits, grey = mud-rich deposits, orange = lag deposit, beige = backfill deposits. (1) Initial seabed conditions before channel incision, interpreted as intersplay mudstones overlain by a terminal splay. (2) First incision episode, that formed a relatively narrow channel incised through older splay deposits and bottomed-out on a lower mudstone layer. (3) Emplacement of a complex lag deposit formed by multiple episodes of deposition, reworking, and bypass. (4) Reactivation of the channel and incision that removed most of the basal lag and widened the channel. (5) Gradual decrease in energy and backfill of the channel. (6) Second reactivation of the channel marked by erosion along the top of the backfill deposits and a significant widening of the channel. (7) Abrupt deactivation of the channel that was then filled mostly by fine-grained turbidite and suspension fallout deposits. These strata are locally intercalated with isolated coarse-grained deposits representing energetic influx from an adjacent active system.

Overlying the feeder channel is a complex of alternating distributary channel network deposits and terminal splay deposits 31 to 35 m thick (see Figure 3.4). The base of this complex is defined by interstratified sandstone beds (Facies 6) with common mudstone intraclasts. Terminal splay deposits either abruptly overlie or are overlain by distributary channel network deposits. There are at least three splay deposits that range from 3 to 12 m thick (lowest is 9-12 m thick, middle is 2.5-3 m thick, and upper is 5-7 m thick) and extend laterally across the outcrop (up to 800 m).

The lowest splay thins on average from northwest to southeast, the middle remains roughly the same thickness throughout the study area, while the upper thins and becomes more mud-rich from southeast to northwest. Splays consist mostly of thick, amalgamated, coarse-grained structureless sandstone (Facies 3) interstratified with rare mudstone partings that usually correlate laterally to mudstone breccia horizons. Common de-watering structures and soft sediment deformation suggest that local rates of deposition were high. These strata are interrupted locally by rare dune cross-stratified sandstone (Facies 4), indicating at least periodic episodes of prolonged traction transport and reworking of the underlying bed. Distributary channel deposits, in contrast, are typically thinner (2 to 6.5 m thick), composed mostly of well-graded Bouma turbidites interstratified with thin-bedded upper division turbidites and rare dune cross-stratified sandstones, and form upward-fining sequences. These sequences are 0.6-2 m thick and typically separated by an erosion surface. The relative thinness (see previous, compared to 8 to 17 m thick in Saller (2008)) of the distributary channel deposits may indicate that they represent the distal part of the network where it transitions into the terminal splay. Such a setting would be prone to frequent changes in sedimentation style. This is suggested by the

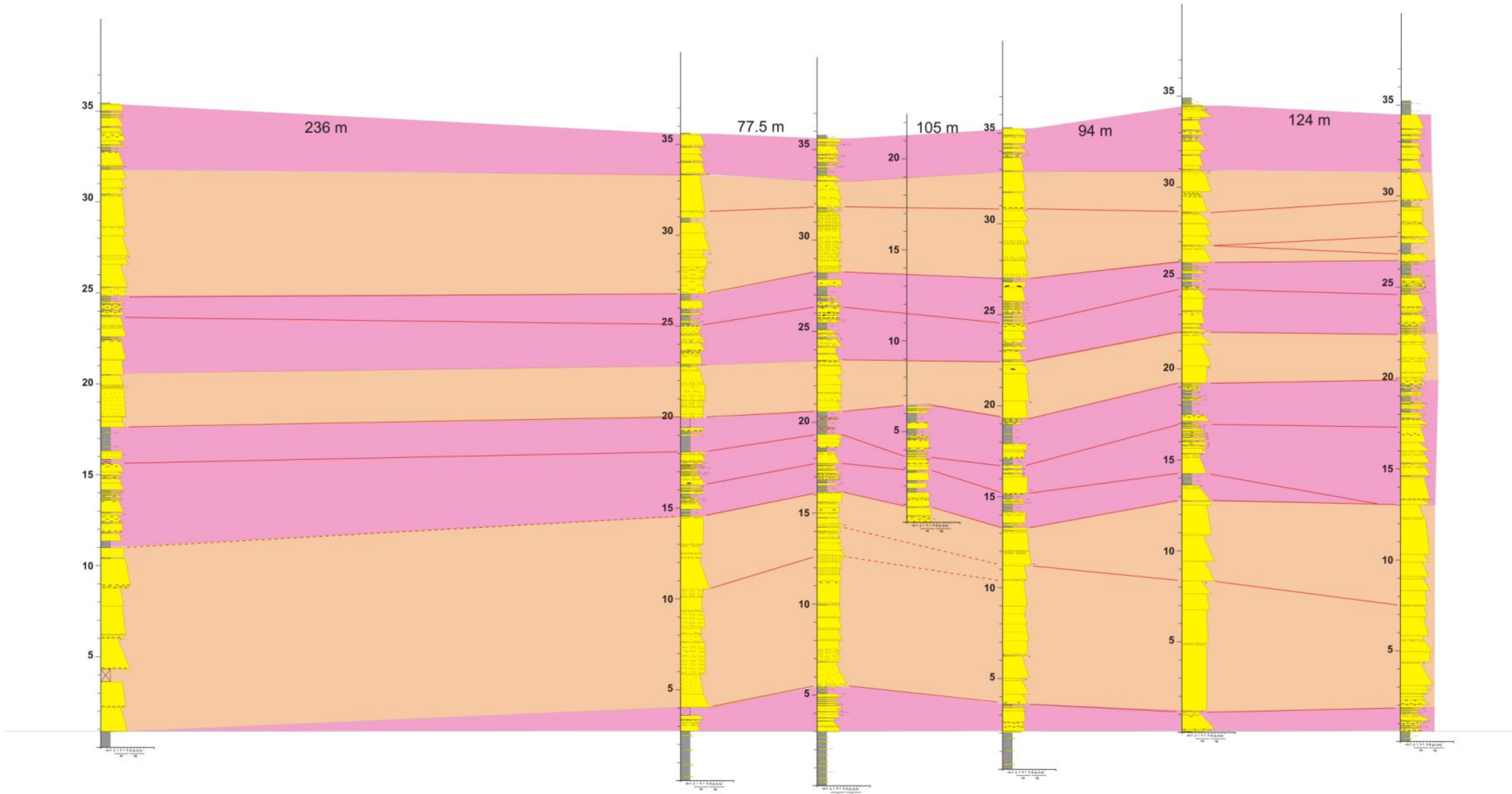

Figure 3.4: Correlation panel showing a depositional lobe composed of both terminal splays (beige) and distributary channel deposits (magenta). The interval underlying this unit is the previously described feeder channel.

alternation of splay and distributary channel deposits, which together form a larger-scale depositional element termed a depositional lobe.

Abruptly overlying the distributary channel network deposit at the top of the aforementioned depositional lobe is a unit of intralobe deposits (see Figure 3.5). These strata are 13 to 17 m thick and composed mostly of thin-bedded fine-grained turbidites uncommonly interstratified with medium-bedded turbidites and rare dune cross-stratified sandstone. This interval is unique in the study area because it contains a 2.5 to 4.5 m thick debrite. This debrite is mud-rich (~70% clay minerals) with abundant but dispersed calcite-cemented sandstone, laminated mudstone, and quartz granule clasts. The abrupt superposition of intralobe deposits immediately above a thick splay complex indicates a major decrease in flow activity, at least locally. This deactivation may be due to a major avulsion of the upflow feeder channel, followed by a largely quiescent interval, punctuated by rare higher energy flows depositing medium-bedded turbidite beds. Conversely, it might also result from a major shut-down of sediment supply to the deep-marine Windermere basin, possibly related to a major rise of relative sea level.

Stratigraphically upwards the thick succession of fine-grained strata is abruptly overlain by two, laterally discontinuous matrix-rich structureless sandstone beds. These beds are approximately 12 cm thick with sharp bases and contain some small mud clasts locally. These distinctively green-coloured, coarse-tail graded beds consist of about 40-50% (clay & silt) with minor (2-3%) chlorite and capped by about 2 cm of unlaminated siltstone. Similar beds have previously been observed at Castle Creek in the Isaac Formation and interpreted to be crevasse splay deposits (Arnott, 2007). Deposition is thought to have occurred rapidly and immediately downflow of an internal hydraulic jump (Leclair & Arnott, 2003). As such these beds herald the

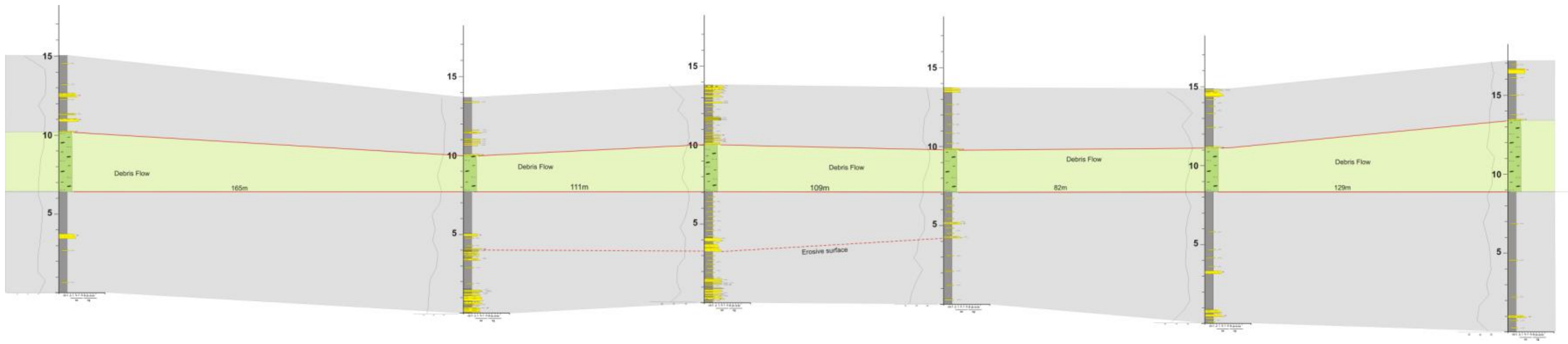

Figure 3.5: Correlation panel showing a unit of intralobe deposits containing a laterally continuous debris flow (green).

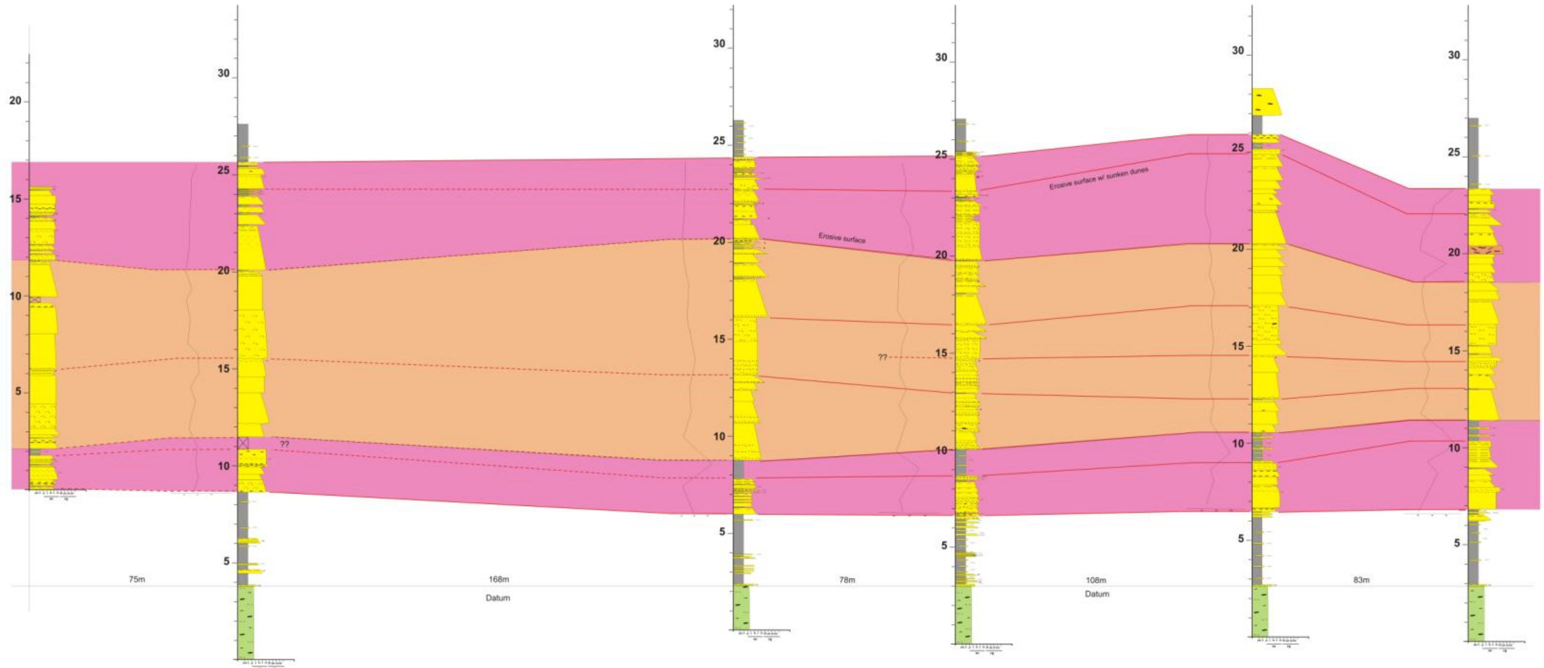

Figure 3.6: Correlation panel showing a depositional lobe composed of a thick terminal splay between two intervals of distributary channel deposits.

onset of high energy flow and sediment-rich depositional conditions within the study area. Overlying these strata is a 16 to 20 metre thick complex consisting of a 5 to 10 metre thick terminal splay deposit sandwiched between two distributary channel units (see Figure 3.6). This may represent a relatively thin depositional lobe.

Abruptly overlying the lobe deposit is a unit of intralobe deposits 28 to 32 m thick that contains at least two different styles of deposition (see Figure 3.7). The basal 15 to 20 m is similar to the underlying intralobe deposit being composed largely of thin-bedded turbidites interbedded with rare medium-bedded turbidites (usually Tbc). Within this interval are two laterally discontinuous, mudstone intraclast-rich very coarse to coarse sandstone units up to 1.5 metres thick. These units also contain uncommon dune cross-stratification and upper plane bed, and most probably represent the fill of shallow isolated scours. These scours were eroded by turbulent eddies created by a sudden reduction in flow confinement. It was subsequently filled by mudstone intraclast-rich sandstone and traction deposits (plane bed and dune cross-stratification) most likely representing a lag deposit related to episodes of sediment reworking and incomplete bypass.

Stratigraphically upwards lies a unit 9 to 14.5 m thick whose basal 1 to 4.5 m is a laterally continuous interval composed of laterally discontinuous coarse to very coarse-grained sandstone beds. Locally, the base of the interval is marked by a single to two set thick dune cross-stratified medium to coarse-grained sandstone or a thin (1 to 4 cm) layer of quartz granules. Nevertheless, the basal unit is dominated by normally graded medium to thick sandstone beds with common mudstone intraclasts and local pillar structures. The overlying 8 to 10 m thick stratal unit comprises mostly medium-bedded Tbc turbidites interbedded with thin-bedded turbidites. There is some variation in the ratio of sand to mud laterally, as shown by the

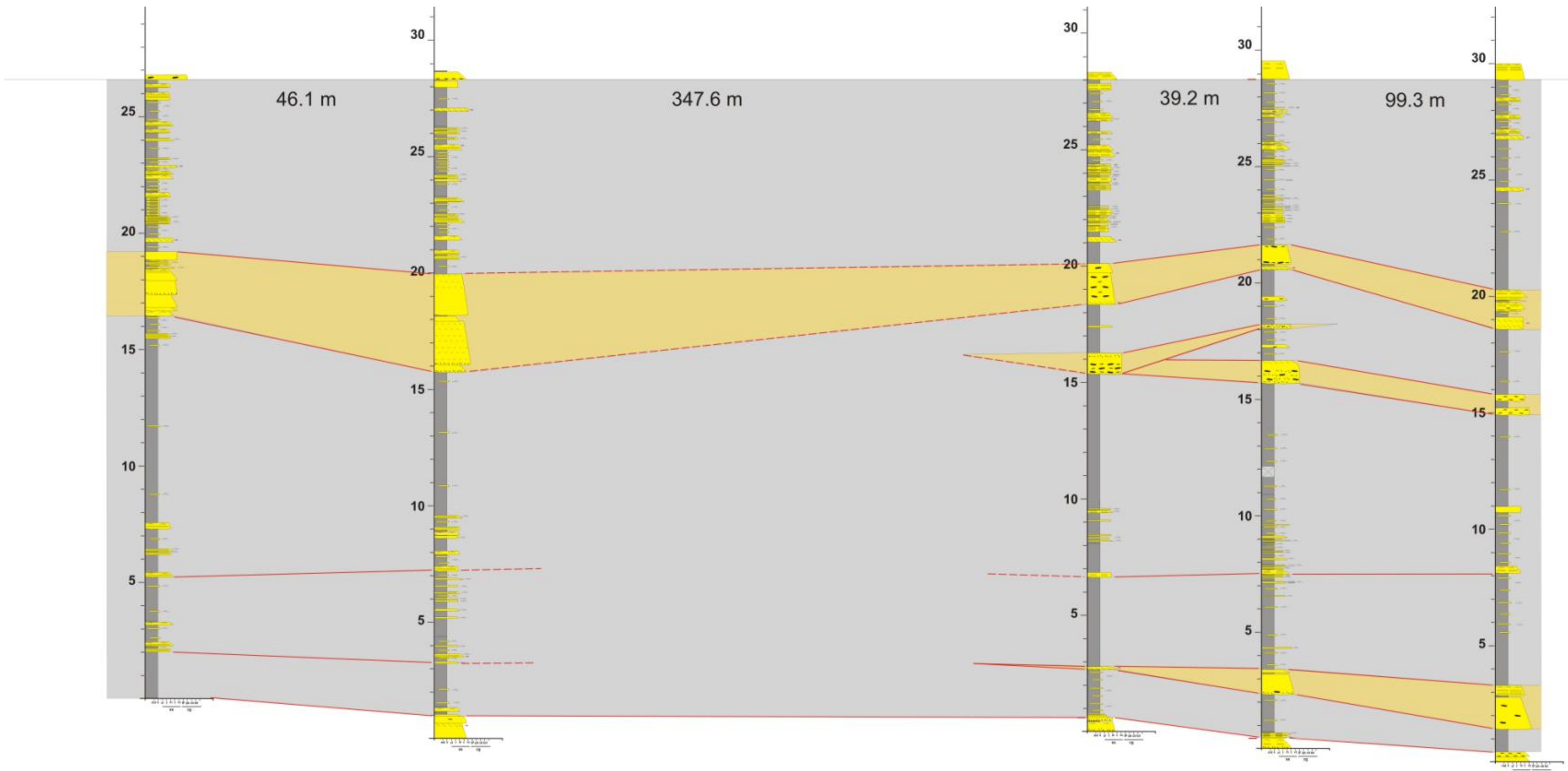

Figure 3.7: Correlation panel showing a unit of intralobe deposits containing multiple isolated scours filled in with sandy lag deposits. Note the change after the final scour.

more common occurrence of thin-bedded turbidites in the northwest part of the study area. The basal part of this interval is interpreted to be a shallow, isolated scour which was later infilled by lag deposits. Subsequently, the remainder of the interval is interpreted to be the result of periods of bypass and sediment reworking alternating with more quiescent intervals of background sedimentation.

The uppermost stratal succession is 58 to 60 m thick but covered mostly by modern glacial deposits. Nevertheless three possible terminal splay deposits are observed, the first two are near the base of the section and are 2.5 to 4.5 m and 4 to 6 m thick, respectively (see Figure 3.8). The third is 10 to 12 m thick and occurs at the top of the interval. The first is composed of two amalgamated sandstone beds interstratified with some thin-bedded turbidites. This unit thins from southeast to northwest. The second comprises mostly thick-bedded coarse to very coarse sandstone with uncommon thin mudstone partings and thins from northwest to southeast. The third consists mostly of thick bedded very coarse sandstone with rare mudstone partings and has a tabular shape. It also contains local pillar structures and dune cross-stratified sandstone. The remainder of the succession is dominated by laterally discontinuous thick-bedded sandstones interbedded with medium-bedded turbidites and rare thin-bedded turbidites. Sandstones are the coarsest in the study area, particularly within the splay deposits, where thin layers of granule and pebble conglomerate (2 to 6 cm thick of quartz clasts) are common at the base of many sandstone beds. Coarse and very-coarse dune cross-stratified sandstone beds are also observed. This interval is interpreted to represent distal slope/toe-of-slope deposits, largely dominated by the earliest/most proximal parts of the distributary channel network. In this environment, the distributary channels may backfill with coarser-grained sediment than they do closer to their termini, causing some to resemble terminal splay deposits. Only the uppermost tabular sand unit

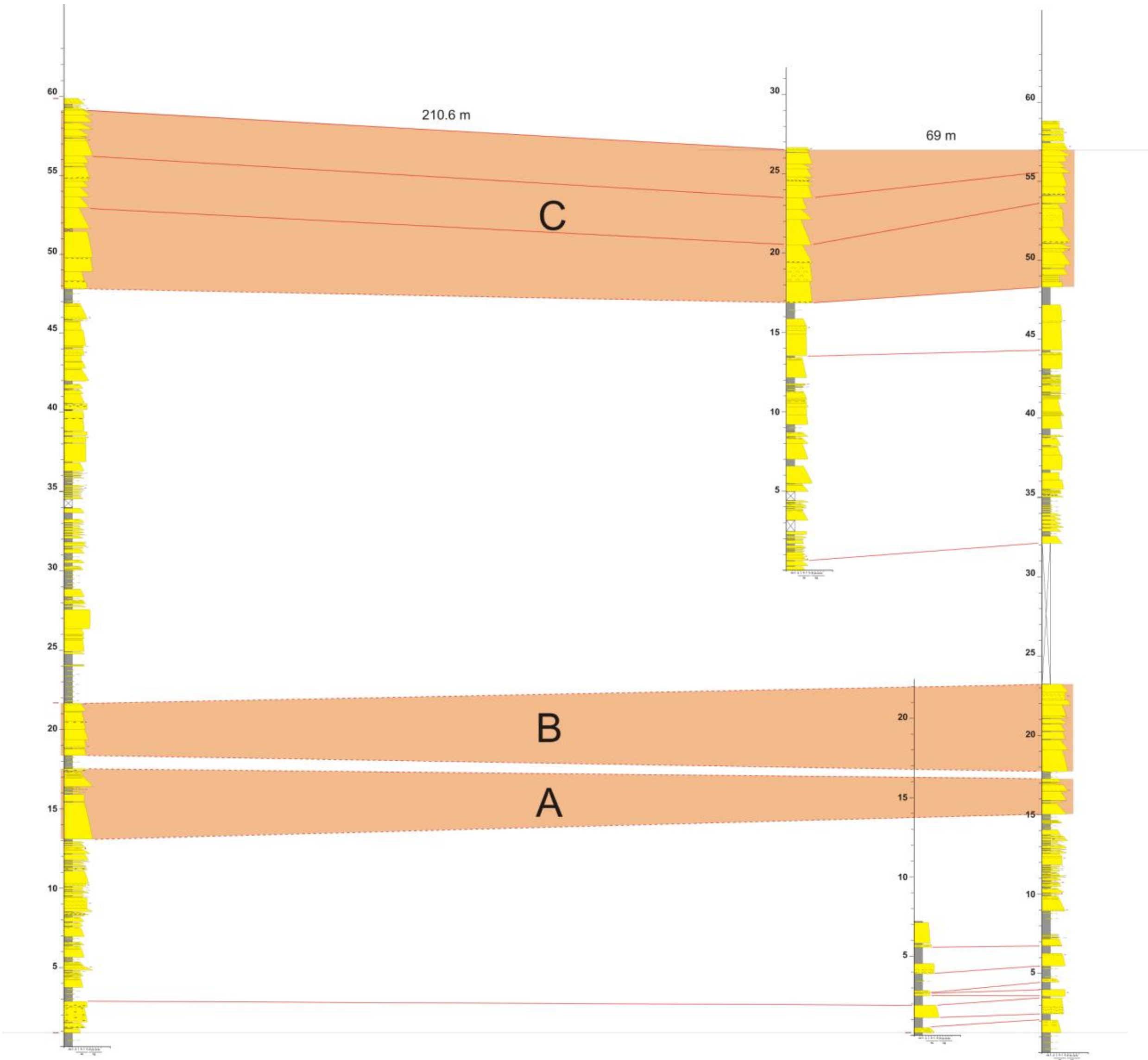

Figure 3.8: Correlation showing three possible splays within the uppermost unit of the study area. A & B are interpreted to be the proximal end of distributary channel, while C is interpreted to be a terminal splay. This unit as a whole is interpreted to be dominated by distributary channel deposits.

is likely to be a terminal splay, while the two lower ones are sandy distributary channel backfill. These events would have taken place near to the channel to lobe transition zone, which is an area of full bypass situated near the toe-of-slope, where the gradient shallows rapidly (Wynn *et al*, 2002). Importantly, this also represents an important change in depositional style relative to the rest of the study area.

Depositional Model

This section outlines the depositional history of strata in the study area. In addition, possible links to broader processes involving the larger-scale Windermere turbidite system will be discussed.

1. Principal depocentre shifted basinward and the main feeder channel incised into older lobe deposits to transport sediment to a more distal locale and create a new lobe. This channel was active for a relatively long period of time, as evidenced by the multiple terraces of the feeder channel suggesting multiple episodes of incision, bypass, deposition, and sediment reworking (see previous section and F.A. 1).
2. The feeder channel became largely abandoned due to upflow channel avulsion and began to be filled with hemipelagites and low-energy turbidites. This quiescence was episodically interrupted by sand-rich flows, most probably overspilled from an adjacent active system, which temporarily exploited the channel as a preferential sediment conduit and incised small, shallow channels in the mudstone-rich abandonment deposits.
3. Shift of depocentre into the study area and initiation of a new lobe. The first phase was the incision of a shallow distributary channel, which then was backfilled, partially eroded, and later overlain by a splay deposit.

4. The system prograded and the distributary channel prograded over the splay deposited in step 3. The distributary channel network shallowly eroded underlying deposits and left behind thin clast-rich lags and reworked sediment in fining-upward packages with mud-rich tops. During this time, the coarsest sediment is interpreted to have bypassed and deposited in the downflow terminal splay.
5. Shift of the depocentre back to the Castle Creek area and deposition of thin sand-rich splay (2.5-3 m thick).
6. Progradation of the distributary channel network and incision of a shallow channel later backfilled by a small number of upward-fining packages.
7. The outcrop area becomes the main locus of sedimentation and a thick splay (5-6.5 m thick) characterized by amalgamated sandstone with dispersed quartz granules is deposited. Strata become less amalgamated laterally toward the northwest, suggesting proximity to the lobe fringe. This is followed by progradation of another distributary channel network and deposition of more stacked, upward-fining channel-fill deposits.
8. The upflow feeder channel avulsed, shifting the depocentre out of the study area. What followed was an interval characterized by hemipelagic sedimentation punctuated occasionally by sandy low density flows that deposited mostly thin-bedded Tcde turbidites. Rare dune cross-stratified sandstone indicates episodic high energy, but low density turbidity currents.
9. Possible slope failure generated a debris flow that incorporated a variety of clast types including quartz granules, carbonate cemented sandstone, and mudstone clasts. Based on the abundant carbonate-cemented sandstone clasts, the debris flow most likely originated higher on the slope and then travelled a significant distance to deposit on the basin floor.

Such a long distance of transport was likely permitted by hydroplaning on a wedge of ambient fluid trapped beneath the flow (e.g., Mohrig *et al*., 1998).

10. A period of general quiescence similar to 8, defined largely by hemipelagic sedimentation and rare sandy turbidites and local sediment reworking.
11. The interval was capped by a small number of matrix-rich sandstone beds which herald a major rejuvenation of coarser-grained sedimentation in at least in this part of the Windermere basin.
12. This was then overlain by a new depositional lobe. It started with the formation of a distributary channel network that incised into the underlying matrix-rich sandstones. This channel was later backfilled by an upward-fining package 3-4.5 m thick overlain by a splay deposit 8.5-11 m thick.
13. Later, the depocentre shifted out of the plane of the study area, but one final distributary was incised into the splay, then became backfilled by several upwards-fining sequences of sediment, and represents the last depositional element in that particular depositional lobe.
14. A prolonged period of silt-rich, low-density, low energy turbidity currents and hemipelagic deposition. Episodically (twice, possibly three times), this quiescence was interrupted by the incision of shallow scours filled with very coarse grained, intraclast-rich lag deposits. In addition to these scours, rare dunes and Tbc beds indicate temporary episodes of high-energy activity dominated by sediment bypass.
15. Incision of a laterally continuous isolated scour that is about 1.5-2 m thick in the northwest part of the study area and 3.5 to a maximum of 5 m thick in the southeast. Flows exploiting this conduit were largely bypass flows. The scour was later filled by

numerous laterally discontinuous coarse-grained sandstone beds with common mudstone intraclasts and rare thin (1-2 cm) quartz granule lags. Laterally discontinuous bedding and traction transport stratification is present, suggesting continual incision and reworking of sediment during a time period of full bypass.

16. The style of deposition changed dramatically as the turbidite system became more active locally. The study area was the site of frequent bypassing flows, alternating with intervals of relative quiescence. This resulted in the deposition of medium-bedded Tbc turbidites alternating with intervals of thin-bedded turbidites.
17. Deposition of 13 to 15 m of laterally discontinuous coarse sandstone beds alternating with intervals of thin and medium-bedded turbidites. Scoured basal surfaces, ripples, and dune cross-stratification are common, suggesting the presence of multiple isolated scours. This is probably the distal-most expression of the channel-lobe transition zone, which is dominated by small scours and is where the system re-organizes itself into a distributary channel network (Wynn *et al*., 2002).
18. Two thin sandy splays were deposited sequentially (2.5 to 4.5 m and 4 to 6 m thick), separated by a relatively short time period of quiescence marked by the deposition of 30 to 50 cm of Tcd turbidites.
19. The distributary channel network was re-established in the outcrop area and deposited approximately 15 m of distributary channel fill.
20. The system locally has an abundance of very-coarse grained sand, along with quartz granules and pebbles, which are deposited in thin lags at the base of many beds. The sequence ends with the deposition of a thick splay deposit that is later incised by a distributary channel before the turbidite system shuts down locally.

Summary

Stratigraphy in the study area shows that at some point after the deposition of the second lobe (steps 12 & 13 above), during an assumed “quiet” period of time, conditions change. The style of deposition and relative abundance of stratigraphic elements changes dramatically, even the range of grain sizes changes and includes some of the coarsest sediment observed in the study area. This is interpreted to represent a shift from a lobe complex-dominated style of deposition to a more proximal style of deposition dominated by scours and thin splays. This may be due to a progradation of the Laurentian continental mass into the Proto-Pacific miogeocline, which in the Castle Creek study area is manifested by the upward change from splay & distributary channel dominated basin floor deposits of the Upper Kaza Group to the leveed-channel complexes of the Isaac Formation (e.g. Ross & Arnott, 2007).

4: Conclusions

Details of the deep marine sedimentary record are poorly known. In part this is because modern systems, which form the comparative basis for reconstructing ancient depositional systems, are largely inaccessible. In modern systems, therefore, seismic imagery is the principal investigative tool. Although excellent in terms of areal extent of investigation, high resolution (cm-scale) is only possible with shallow penetration, while achieving high vertical penetration reduces resolution, enabling only decameter-scale features to be resolved. The ancient record, on the other hand, provides excellent vertical resolution (down to mm-scale) but generally suffers in its lateral and in many places vertical dimension. This disparity in scale has made it difficult to accurately merge datasets from the modern and ancient deep marine sedimentary record. Of great benefit, therefore, would be the study of seismic-scale outcrops. These would allow a variety of stratigraphic scales to be observed concurrently and confidently integrated. A km-scale exposure of deep marine sedimentary rocks at the Castle Creek study area in east-central B.C., provides just such an opportunity. In this study the deep marine, basin-floor strata of the Upper Kaza Group are described in order to outline their stratal constituents and composite stratal architecture.

In deep-marine strata of the Upper Kaza Group strata at the Castle Creek study area a total of six facies were identified based on: grain size, sorting, lithology, sedimentary structures, thickness, and bedding contacts. Facies include: poorly-sorted, clast-rich mudstone (Facies 1); thin-bedded siltstone and mudstone (Facies 2); thick-bedded, massive sandstone (Facies 3); medium-scale, cross-stratified sandstone (Facies 4); mudstone-clast breccia (Facies 5); and medium-bedded turbidites (Facies 6). Facies 1 consists of quartz granules, pebbles, and mudstone clasts dispersed in a mudstone matrix and are interpreted to be debrites deposited by cohesive sediment gravity flows. Facies 2 consists of siltstone and mudstone beds interstratified with rare fine- to medium-grained sandstone beds. It is interpreted to be mostly upper division turbidites deposited by episodic, dilute, low-density turbidity currents. Facies 3 consists of coarse- to very coarse-grained amalgamated sandstone and indicates rapid suspension deposition from turbidity currents with high sediment concentration. Facies 4 consists of medium- to coarse-grained, commonly calcite-cemented, dune cross-stratified sandstone. The combination of dune cross-stratification, which is locally deformed, and the granulometric similarity with the top of the immediately underlying bed suggests the local reworking of the sea bed by the low-concentration tail of the same turbidity current that deposited the underlying bed. Facies 5 consists of mudstone clasts dispersed in a coarse matrix, typically upper coarse sandstone to granule conglomerate, are interpreted to be lag deposits sourced by updip erosion or the result of the local break-up of a siltstone layer along an amalgamation contact. Facies 6 consists of sandy, medium-bedded Bouma sequence turbidites interpreted to be the result of deposition from waning, low-density turbidity currents.

At Castle Creek the natural grouping of two or more facies defined five architectural elements, which are mappable, 3-dimensional geomorphic features that build up the depositional system. Here the architectural elements include: heterolithic feeder channel deposits (F.A. 1), thin-bedded intralobe turbidites (F.A. 2), terminal splay deposits (F.A. 3), distributary channel deposits (F.A. 4), and isolated scours (F.A. 5). These architectural elements form an idealized depositional continuum from the base-of-slope to basin floor (Fig 4.1).

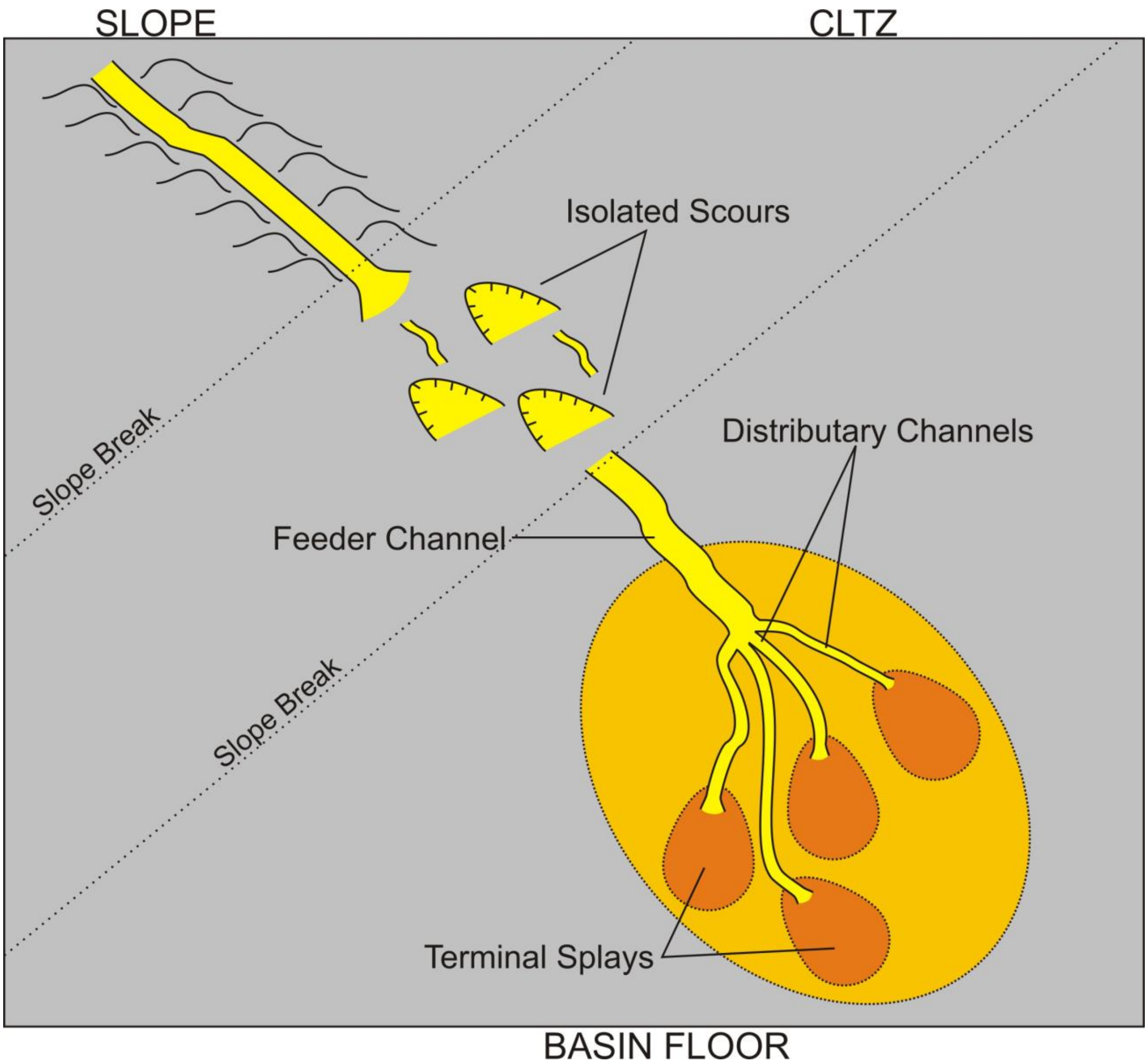

Figure 4.1: Simplified diagram of the Windermere turbidite system showing the architectural elements identified in this thesis. Note that intralobe mudstones would lie within the lobe boundaries or in close proximity to depositional elements in the CLTZ.

The most proximal architectural element is the isolated scour. It consists of local erosion surfaces that incise up to 5 m into underlying strata, which in most places are thin-bedded intralobe turbidites. Scours are filled mostly with thick-bedded sandstone and mudstone-clast breccia, and are interpreted to occur immediately downflow of the terminus of a leveed slope channel. Flows entering this area undergo a rapid loss of confinement, causing rapid flow expansion and increase of turbulence and its intensity. Accordingly, the sea bed is deeply eroded locally and deep scours form. These scours are later infilled by discontinuous lag deposits left by flows that largely bypassed the area and deposited most of their sediment further downflow. This area coincides with what is termed the channel-to-lobe transition zone, or CLTZ in modern systems, which according to Wynn and others (2002) is an area characterized mostly by sediment bypass and scour.

Downflow of the CLTZ the system organizes itself into a typical distributive architectural style (see Chapter 3). The most upflow element in a downflow continuum of architectural elements is the lobe feeder channel. At Castle Creek the feeder channel is an erosionally confined feature ranging from 2.5 to 16.5 m deep. Its fill comprises three erosionally-based stratal units, each made up of a unique assemblage of lithofacies. The first stratal unit consists of dune cross-stratified sandstone intercalated with mudstone-clast breccia, the second comprises mostly medium-bedded turbidites interbedded with thin-bedded turbidites, and the third by thin-bedded turbidites with rare coarse-grained sandstone beds. Each of the three stratal units is preceded by a major rejuvenation of the system and attendant erosion, followed thereafter by deposition. The first stratal unit is interpreted to represent an interval characterized mostly by bypass flows but local deposition of discontinuous lag deposits. The second is interpreted to represent the gradual backfilling of the channel, while the third is represents channel abandonment.

Downflow, the feeder channel shallows and divides into multiple shallow distributary channels. These channels are 2 to 6.5 m deep, 100's of m wide and flows passing through them were largely unconfined. The bases of the channels are shallowly scoured to nearly planar, although proximal parts may be more deeply incised. Channels were later filled with one or more upward-fining stratal units. These units typically consist of one or two partial to full Bouma sequence turbidites overlain abruptly by an interval of thin-bedded turbidites with rare dunes or Tb/Tbc turbidites. This style of fill is interpreted to represent an initial period of bypassing flows followed by gradually decreasing activity, then a local abandonment/avulsion. Collectively, all the distributary channels associated with a single depositional lobe are termed the distributary channel network and form the link between the lobe feeder channel and the outboard terminal splays.

At the terminus of each distributary channel lies a terminal splay. These are depositional features 5 to 13 m thick composed largely of amalgamated sandstone. Splays are planar-based with a tabular or sheet-like geometry and extend across the full width of the outcrop (≤ 800 m). Lateral facies changes are negligible to subtle over 100's of metres. Terminal splays are interpreted to result from flow expansion at the terminus of the distributary channels and represent the principal depocentre within the larger-scale depositional lobe, a feature that comprises an assemblage of terminal splay and distributary channel deposits.

Thin-bedded intralobe turbidites form units 10 to 30 m thick, consisting mostly of thin-bedded upper division turbidites interbedded with rare medium-bedded turbidites and dune

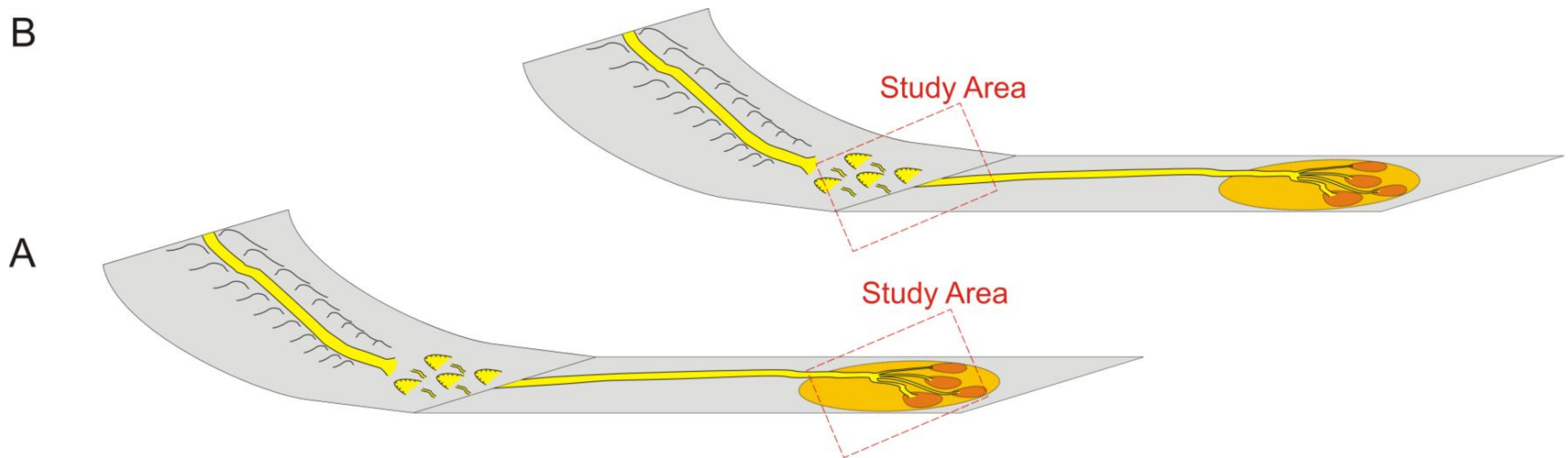

Figure 4.2: Schematic showing the progradation of the Windermere turbidite system across the study area: (A) deposition of "classic" basin-floor strata, (B) deposition of more erosional transition zone strata.

cross-stratified sandstone. Debrites up to 4 m thick occur, but are rare. Intralobe turbidites are interpreted to be the result of a local shutdown of the sediment transport system, which likely is the result of an upflow channel avulsion. These deposits are interpreted to in the area between and surrounding the distributary channels and terminal splays.

Lastly, the vertical stacking pattern of architectural elements in the Castle Creek study area suggests that the entire Windermere turbidite system was prograding when these strata were deposited. Specifically, the upward change in depositional style from sandy, stacked basin floor depositional lobes, fine-grained intralobe and rare feeder channel deposits to bypass, erosion, and rapid facies changes in the channel-to-lobe transition zone (CLTZ) (see Figure 4.2).

Future Work

The transition from basin floor to slope has been very poorly documented in the ancient or modern sedimentary records, and this thesis only captures the beginning of this transition. Future work should focus on the detailed description of the strata immediately above those in this study, documenting their composite facies, their vertical and lateral distribution, and ultimately, identify the constituent architectural elements. These elements should show widespread evidence of significant sediment bypass and erosion, but also local deposition. In addition, within basin floor strata, more detailed correlation of distributary channel deposits would help better define the stratal make-up and internal geometry of this important architectural element. This work could be conducted at Castle Creek but also in the area around Valemount, B.C. where expansive, well-exposed mountainside sections of sandstone-rich strata of the Upper & Middle Kaza Group crop out.

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				Grains					Cements				Matrix				Average Grain	Average			
Section #	Sample #	Facies	Qz	Plag	Pyr	Cal	Other		Qz	Cal		Bio	Mus	Chl	Clay min.	TOTAL	Size (mm)	Grain Size	Sorting		Comments
1	JR-1	Debrite	20	1	2	2							1	3	71	100	0.24	fine sand	very poor		undulatory ex
1	JR-2	Debrite	73	2	5	10	1						3	1	5	100	0.48	med. sand	poor		Pyr-cubic w /
1	JR-3	TBT	30		2	1							3	1	63	100	0.14	fine sand	well		polycrystallin
1	JR-4	TBT	40		2								1	2	55	100	0.1	v. fine sand	well		Qz-undulator
1	JR-5	MBT	30	5								1	3	2	59	100	0.34	med. sand	well		Qz-undulator
1	JR-6	Ta	80	5	1								2	1	11	100	0.55	coarse sand	poor		Qz-undulator
1	JR-7	Dune	58	10	1					25			1		5	100	0.38	med. sand	well		Ca-iron-rich c
1	JR-9	Ta	60	5	2	2	2						3	1	25	100	0.37	med. sand	poor		Qz-undulator
1	JR-10	MBT/Wft.	50	5	1		1						2	1	40	100	0.19	fine sand	well		Qz-undulator
1	JR-11	Dune	55	1	1		tr			25		tr	1	1	16	100	0.1	v. fine sand	well		Qz-undulator
1	JR-13	Dune/MBT	40	tr	5					15		tr	tr		40	100	0.29	fine sand	well		Preferred ori
1	JR-15	MBT/Wft.	40	5	2	3	tr						4	2	44	100	0.17	fine sand	well		Qz-undulator
1	JR-16	MBT	50	1	1							tr	3	2	43	100	0.17	fine sand	well		Planar lamina
1	JR-18	MBT?	45	4	2	1						tr	2	1	45	100	0.22	fine sand	poor		Qz-undulator
1	JR-19	MBT?	35	1	3								5	4	52	100	0.09	v. fine sand	well		Qz-undulator
1	JR-20	MBT	59	5	2					2		tr	2	tr	30	100	0.3	fine sand	moderate		Calcite cemer
1	JR-21	Ta	65	1	1	5				10		tr	1	1	16	100	0.33	med. sand	poor		Pg-polysynth
1	JR-22	Ta	65	2	1					10		tr	1	1	20	100	0.18	fine sand	well		Diffuse undu
2	JR-23	Ta	75	5	1	3	tr						3	1	12	100	0.34	med. sand	poor		Qz-undulator
1	JR-24	MBT	30		5							tr	2	3	60	100	0.08	v. fine sand	well		Qz-undulator
1	JR-25	Ta	65	7	1					5		tr	1	1	20	100	0.26	med. sand	poor		Qz-undulator
1	JR-26	Ta	71	5	1		tr			10		tr	2	1	10	100	0.23	fine sand	poor		Qz-undulator
1	JR-27	Ta	65	3	2							tr	1	1	28	100	0.41	med. sand	poor		Qz-undulator
4	JR-29	MBT	50	10	2							tr	3	2	33	100	0.18	fine sand	well		Qz-undulator
4	JR-30	MBT/Wft.	61	3	1		tr					tr	tr		35	100	0.31	med. sand	poor		
4	JR-32	Ta	75	2	2							tr	1	tr	20	100	0.29	med. sand	poor		Appears to b
4	JR-33	MBT	48	1	1								tr	tr	50	100	0.29	med. sand	poor		Pyr-cubic.
4	JR-34	MBT	50	1	2	5							1	1	40	100	0.4	med. sand	poor		
4	JR-36	Dune	40		2					15		tr	1	1	41	100	0.07	v. fine sand	well		Ca-has a disp
4	JR-37	Debrite	15		2	2							1	2	78	100	0.12	v. fine sand	poor		Weak preferr
4	JR-39	MBT	55	1	2							tr	2	2	38	100	0.16	fine sand	well		
3	JR-42	MBT	30		1							tr	5	5	59	100	0.07	v. fine sand	well		Faint planar la
3	JR-43	MBT	40		1	tr						tr	2	2	55	100	0.09	v. fine sand	well		Faint planar la
6	JR-44	Dune/MBT	60	2	1					10		tr	1	tr	26	100	0.21	fine sand	well		Visibly lamina
6	JR-46	MBT/Wft.	45	3	2								1	1	48	100	0.22	fine sand	poor		
6	JR-47	MBT	40	1	2							tr	2	3	52	100	0.11	v. fine sand	poor		Faint laminati
6	JR-49	Debrite	25	5	2		tr						2	2	64	100	0.2	fine sand	very poor		Weakk prefer
6	JR-52	Ta	62	5	1								2	tr	30	100	0.23	fine sand	moderate		
6	JR-51	Dune/MBT	71	3	tr					15		tr	1		10	100	0.28	fine sand	well		No visible lan
6	JR-54	Ta	69	2	1							tr	2	1	25	100	0.35	med. sand	moderate		
6	JR-55	MBT	20		2	tr						tr	5	5	68	100	0.08	v. fine sand	well		No visible lan
6	JR-56	MBT	35	tr	3							tr	3	2	57	100	0.09	v. fine sand	well		Visibly lamina

Paleoflow Measurements					
1	261				
2	252				
3	254				
4	224				
5	272				
6	249				* All meas
7	205				
8	279				
9	263				
10	271				
11	189				
12	267				
13	248				
14	251				
15	83		Outlier		
Average excluding outlier:			248.9286		