

**ORIGIN AND ARCHITECTURE OF DEEP-WATER LEVEE
DEPOSITS: INSIGHT FROM THE ANCIENT ROCK
RECORD AND EXPERIMENTS**

Zishann Khan

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Ottawa-Carleton Geoscience Centre
Faculty of Science
University of Ottawa

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Abstract

Although levee deposits make up a significant part of modern and ancient deep-marine slope systems, details of their internal lithological composition and stratal architecture remain poorly documented. At the Castle Creek study area, strata of the Neoproterozoic Isaac Formation (Windermere Supergroup) crop out superbly in a kilometre-scale section through a sinuous deep-water channel-levee system (ICC3).

Levee deposits near the outer bend of the channel consist of sandstone-rich (sandstone-to-mudstone ratio of 68:42), medium- to thick-bedded turbidites interstratified with thinly-bedded turbidites. Structureless sandstone (T_a), planar laminated sandstone (T_b), non-climbing ripple cross-stratified sandstone (T_c) and massive and laminated siltstone (T_d) are common. Thick beds generally thicken and then thin and fine laterally over about 300 m. Thin-bedded strata, in contrast, thin and fine negligibly over similar distances. In the distal part of the outer-bend levee (up to 700 m laterally away from the channel) strata consist predominantly of thin-bedded T_{cd} turbidites with a much lower sandstone-to-mudstone ratio (35:65). On the opposite side of the channel, inner-bend levee deposits are mudstone-rich, locally as low as 15:85, and consist mostly of thin-bedded, T_{cd} turbidites, although thicker-bedded, T_{a-d} turbidites are more common in the lower part of the section. Lateral thinning and fining of beds is more rapid than their outer-bend counterpart.

Levee deposits of ICC3 comprise three stacked decametre-scale upward-thinning and -fining successions. Each is interpreted to record a depositional history consisting of lateral channel migration, levee deposition, channel filling, and distal levee deposition. During the early stage of increasing levee relief it is proposed that the termini of individual beds progressively backstep towards the channel margin resulting in an overall lateral thinning of the stratal profile. This interpretation notably contrasts the common assumption that levee morphology is the result of the vertical stacking of beds that dip.

In addition to field studies, laboratory experiments were conducted to determine the depositional threshold of non-climbing ripple cross-stratification, which is common in levee strata of ICC3. It was determined that non-climbing ripples form when bed aggradation rates are less than 0.015 cm/sec, and most probably in flows made up of poorly sorted sediment.

Résumé

Malgré que les levées soient un élément important des systèmes turbiditiques profonds modernes et anciens, les détails quant à leur composition lithologique et architecture interne demeurent mal documentés. La formation Néoprotérozoïque d'Isaac (supergroupe Windermere) affleure sur une distance verticale de l'ordre du kilomètre dans la région de Castle Creek au sud-ouest de la cordillère canadienne. L'objet de la présente thèse est d'étudier l'un des systèmes sinueux de chenal-levée (ICC3) faisant partie d'un dépôt plus vaste de cône sous-marin d'environnement marin profond.

Les dépôts de levées le long de la rive externe du chenal sont des dépôts turbiditiques d'épaisseur moyenne à épaisse à haute teneur en grès (le ratio grès:mudstone est de 68:42), interstratifiés avec des turbidites finement litées. Des grès sans structures (Ta), à laminations planaires (Tb) ou à stratifications entrecroisées formées par des rides non-grimpantes (Tc), ainsi que des siltstones massifs et laminés (Td) y sont aussi communément observés. Les lits plus épais ont tendance à s'épaissir puis à s'amincir et à granodécroître sur une distance latérale d'environ 300m. Les lits plus minces, quant à eux, s'amincissent et granodécroissent que très peu sur des distances semblables. Les strates dans la partie distale de la levée situées sur la rive externe du chenal (jusqu'à 700 m au-delà du chenal) se composent principalement de turbidites finement-litées (Tcd) ayant un plus faible ratio grès:mudstone (35:65). Les dépôts de levées de la rive interne du chenal, malgré qu'également principalement composés de turbidites finement litées Tcd, sont quant à eux plus riches en mudstone (ratio grès:mudstone aussi faible que 15:85) et fréquemment sus-jacents à des dépôts de turbidites plus grossièrement litées (Ta-d) se retrouvant dans la partie inférieure de la section. Finalement, l'amincissement et la granodécroissance latérale des lits sont plus rapides dans ces dépôts de rive interne que dans ceux de la rive externe du chenal.

Les dépôts de chenal-levée (ICC3) sont constitués de l'empilement de trois successions dépositionnelles d'échelle décamétrique qui s'amincissent et granodécroissent vers le haut de la section. Chacune de ces successions dépositionnelles est interprétée comme étant la somme des événements suivants : 1) la migration latérale du chenal; 2) la construction des levées; 3) le remplissage du chenal et le dépôt de levée distale. Durant le début de la phase de construction des levées, il est proposé que les extrémités des lits se succédant migrent progressivement vers la marge du chenal. Ceci résulterait en un amincissement latéral général du profil des strates. Cette dernière interprétation s'oppose à l'hypothèse courante selon laquelle la morphologie des levées est le résultat de l'empilement vertical de lits inclinés.

De plus, des essais en laboratoire ont été menés afin de déterminer les conditions de formation des rides non-grimpantes communément observées des les dépôts de chenal-levée (ICC3). Lors de ces essais, des rides non-grimpantes ont été générées lorsque le taux d'aggradation du lit était inférieur à 0.015 cm/sec. Finalement, il est aussi postulé que ces structures sédimentaires se formeraient en présence d'une charge sédimentaire mal triée.

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CHAPTER ONE

Thesis introduction

1.1. Thesis rationale

Deep-water channel-levee systems are major conduits for delivering sandy sediment to the floor of the ocean where it accumulates and forms some of the largest sedimentary deposits on earth (i.e. submarine fans). These channels are impressive in dimension, measuring up to several hundred meters deep, several kilometres wide, and extending basinward hundreds to thousands of kilometres. Over time channels commonly become filled with sand-rich sediment that in the subsurface can form high-quality hydrocarbon reservoirs (Leonard et al., 2000; Navarre et al., 2002; Sprague et al., 2002; Weimer and Slatt, 2007). As a consequence, submarine channels and their deposits have garnered significant academic and economical interest over the past several decades. However, in many deep-water systems the finer-grained levee deposits are volumetrically more significant and have also proven to be important hydrocarbon reservoirs (e.g. Clemenceau et al., 2000; Kendrick, 2000). The paucity of studies that focus on levees in the literature is in part a consequence of the severe weathering and/or vegetation overgrowth typically associated with these deposits. Consequently, deep-water levees remain poorly understood and further investigation is needed to better comprehend their architectural attributes and formative processes.

Levee deposits are generally interpreted to aggrade by deposition from the overspill of throughgoing channel flows. This process commonly results in a relatively uninterrupted depositional record that reveals important temporal and spatial changes in channel-levee

development (Clark and Pickering, 1996; Hiscott et al., 1997). Furthermore, much has been inferred about the interplay between channel and associated levee deposition by studying well exposed channel margins (Beaubouef, 2004). It is rare, however, to preserve an ancient channel complex that exposes levee strata on both sides of the channel and, therefore, little is known about the stratal differences between these counterpart deposits. As a result, many channel-levee growth models described in the literature assume symmetry in facies and thickness of levees deposited on opposite sides of the channel (e.g. Eschard et al., 2003; Grecula et al., 2003; Beaubouef, 2004). High resolution seismic images of modern systems, however, show that the majority of submarine channels exhibit a sinuous planform and opposing levees typically vary significantly in height, especially along channel bends (Posamentier, 2003; Posamentier and Kolla, 2003; Deptuck et al., 2003). This illustrates how seismic data are advantageous over outcrop studies in determining large-scale channel-levee geometries, but they lack the vertical resolution needed to delineate small-scale internal stratal attributes. What is needed therefore, are well-exposed, large-scale outcrops to better understand stratal heterogeneity that is below seismic resolution.

A kilometre-scale, well exposed outcrop of stacked channel-levee complexes and other base-of-slope deposits of the Isaac Formation is present in the Castle Creek study area of east-central British Columbia. The objective of this thesis is to provide detailed descriptions and interpretations of inner- and outer-bend levees associated with one of the channel complexes. This thesis also presents the results of a laboratory investigation into the deposition of non-climbing ripple cross-stratification, a common sedimentary structure in the proximal levees of the Castle Creek channel-levee complexes. From an industry perspective, this study discusses details about lateral and vertical trends in bed thickness, grain-size distribution, sandstone-to-mudstone trends, and connectivity, which are properties essential

in accurately assessing risk in hydrocarbon exploration and production. Moreover, owing to the quality and expanse of the exposure, both vertically and laterally, a number of hypotheses are proposed about the formative processes of levees. These hypotheses challenge conventional norms and offer new insight into how deep-water channel-levee systems are deposited.

1.2. Geological setting and regional stratigraphy

Rocks of the Windermere Supergroup (WSG) crop out over most of the length of western North America; however, exposure and facies type vary considerably (Fig. 1.1). Throughout northwestern Mexico and the southwestern United States, strata crop out discontinuously and consist mainly of continental and shallow-marine sedimentary rocks (Link et al., 1993). This contrasts the fairly continuous outcrop belt that extends from southwestern Canada to the Yukon-Alaska border. In the southern Canadian Cordillera rocks are comprised mostly of deep-water sediment gravity flow deposits, whereas to the north in the Mackenzie Mountains shelf and upper-slope facies are exposed (Ross et al., 1995). Collectively, these rocks are interpreted to represent deposition along a continental margin from shelf to basin floor (Fig. 1.2).

Deep-water WSG rocks in the southern Canadian Cordillera are up to 9 km thick and exposed over an area of 35 000 km² (Fig. 1.3). These rocks make up part of the Foreland Fold and Thrust and Omineca belts (Fig. 1.4) – two of the five morpho-tectonic belts created by the collision and accretion of allochthonous terranes along the western margin of Laurentia (ancestral North America) during the Mid-Jurassic to Early Cenozoic (Price, 2000). These events led to pervasive deformation and metamorphism in Mesozoic and older

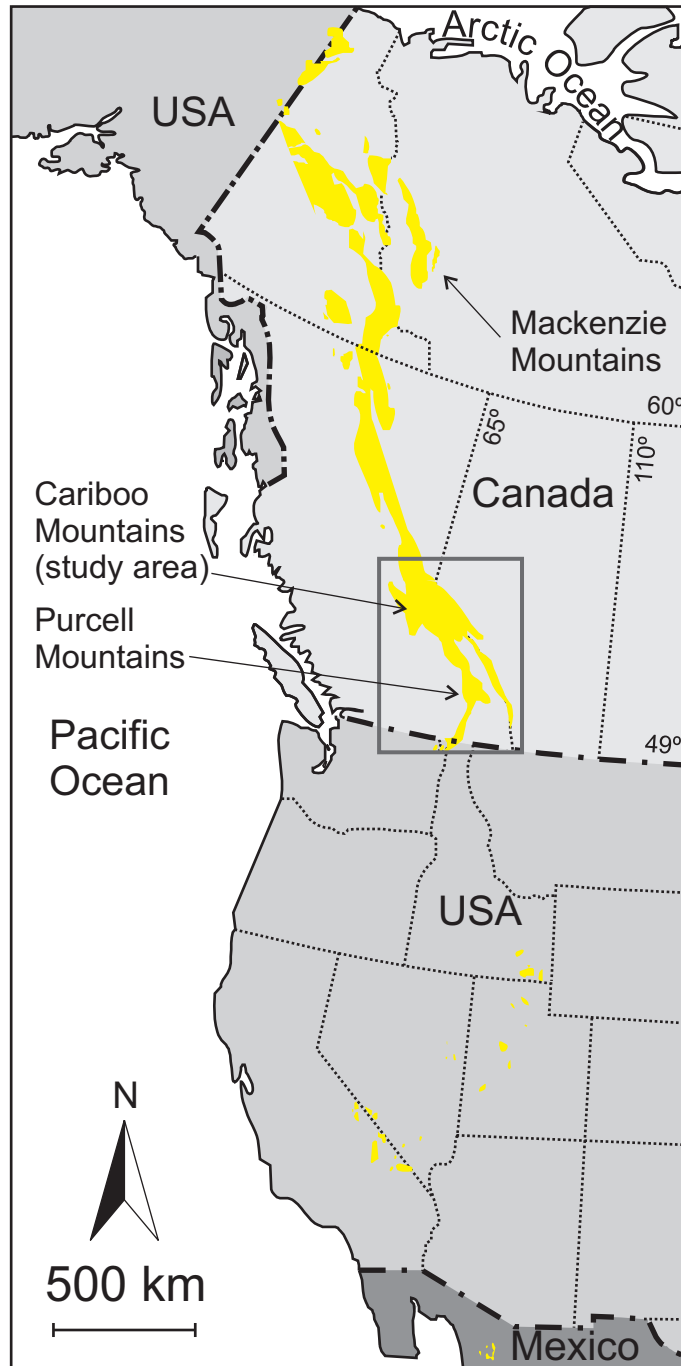


Fig. 1.1. Map showing the Windermere Supergroup outcrop belt in western North America. The deep-water part is outlined by square (after Ross, 1991).

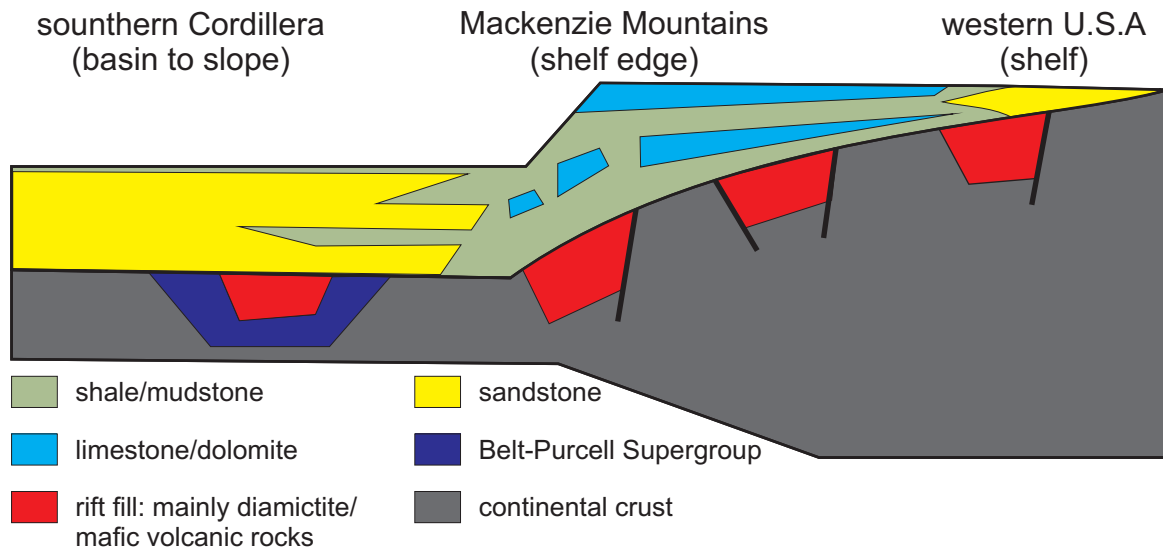


Fig. 1.2. Schematic showing interpreted cross-section of the western North America margin ca. 600 Ma (redrawn from Ross, 1991). The occurrence of carbonates in basinal facies of the southern Cordillera are correlated with shallow water carbonates in the Mackenzie Mountains. Correlation of shelf strata in the western United States of America with those in the Canadian Cordillera remains unknown. Although stratigraphic correlation between these areas remains uncertain, age and continuity of Neoproterozoic stratigraphic elements (e.g. glaciogenic diamictites and mafic volcanic rocks) along western North America suggest a once continuous basin (Eisbacher, 1985; Ross, 1991). Shallow water and continental equivalents that may have occurred in the Canadian Cordillera were eroded by sub-Cambrian erosion (Ross, 1988).

rocks in the southern Canadian Cordillera (Reid et al., 2002). Nevertheless, the degree of metamorphism and structural deformation tends to be relatively low in the Foreland Fold and Thrust Belt, especially east of the Southern Rocky Mountain Trench (Ross and Arnott, 2007). This includes regions around Lake Louise, Jasper, and Mount Robson. In the Omineca Belt, in contrast, rocks of the WSG are typically characterized by high metamorphic grade and complex structural deformation. However, in the Northern Cariboo and Purcell mountains within the Omineca Belt, local areas of relatively low metamorphic grade and simple structure occur. These areas preserve primary sedimentary textures and structures (Ross and Arnott, 2007). Accordingly, detailed sedimentological studies carried out in these areas and in the Foreland Fold and Thrust Belt have fostered the development of a robust regional stratigraphic framework for rocks of the WSG in southwestern Canada

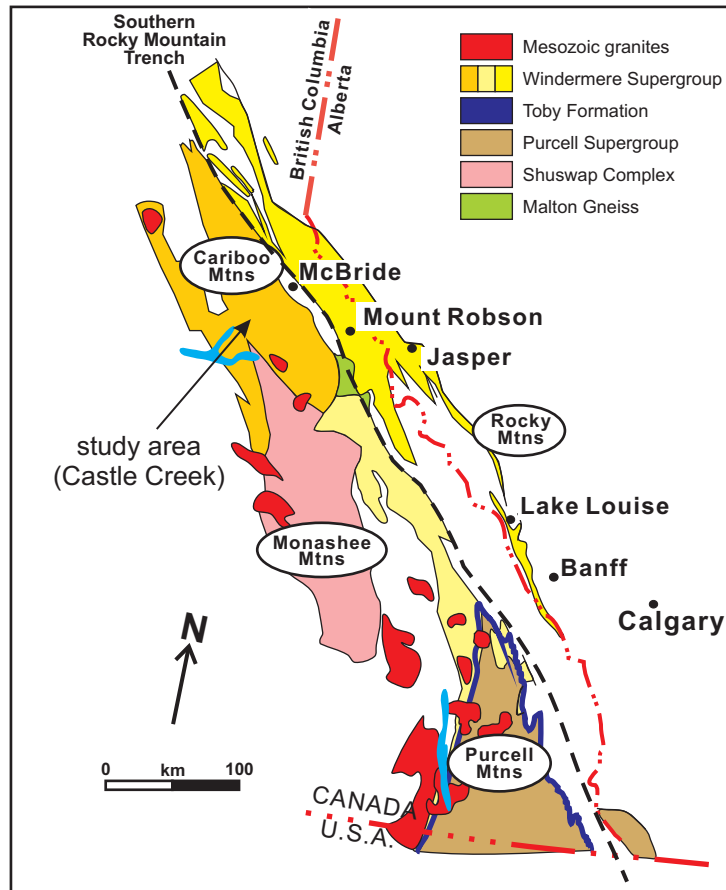


Fig. 1.3. Geological map of the southern Canadian Cordillera showing the distribution of the deep-water part of the Windermere Supergroup and surrounding units (after Young et al., 1973; Ross and Murphy, 1988).

(Ross et al, 1989, 1995; Ross, 2000).

In the southern Canadian Cordillera, strata of the WSG are interpreted to have been deposited in a deep-water ocean basin created during break-up of the supercontinent Rodinia at ca. 750 Ma (Bell et al, 1987; Ross, 1991; Ross et al., 1995; Karlstrom et al., 2001). At that time, it is postulated that a large continental mass rifted away from the western margin of Laurentia, resulting in the creation of the proto-Pacific Ocean and the development of an extensive passive margin (Ross et al., 1995). Consensus on the conjugate landmass has yet to be established but possible candidates include Siberia (Sears and Price, 1978; Condie and Rosen, 1994; Frost et al., 1998; Rainbird et al., 1998), East Antarctica (Moore, 1991;

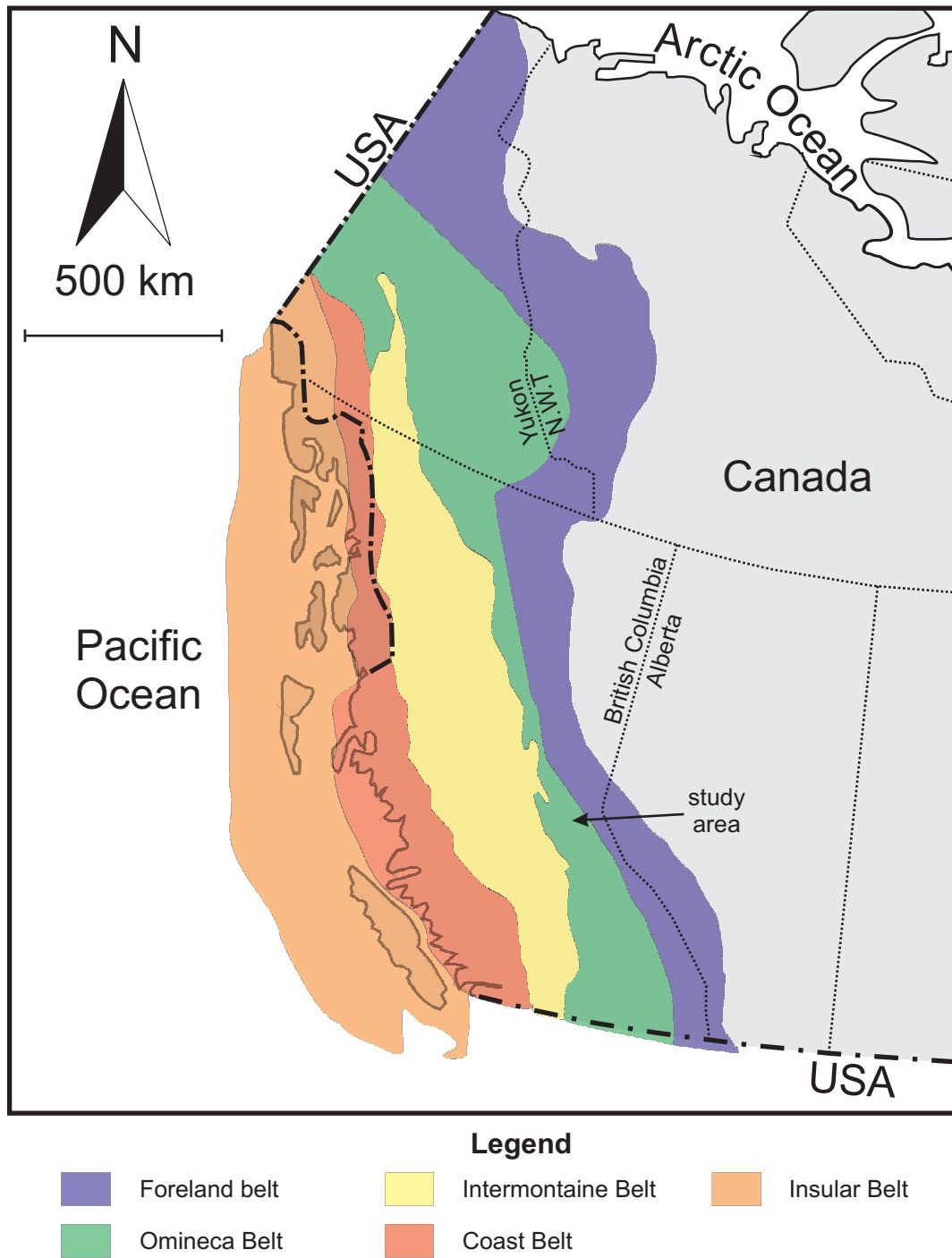


Fig. 1.4. Map showing the five morpho-tectonic belts of the Western Canadian Cordillera. Belts trend northwest-southeast and reflect mountain building processes related to collisional tectonism between 185 and 50 Ma (after Wheeler and McFeely, 1991).

Dalziel, 1991; Borg and DePaolo, 1994), Australia (Brookfield, 1993; Burrett and Berry, 2000; Wingate et al., 2002; Fioretti et al., 2005), and South China (Li et al., 1995). The lower part of the WSG in the southern Canadian Cordillera locally consists of an intercalated assembly of coarse grained siliciclastic strata of the Toby Formation and mafic volcanic rocks of the Irene Formation (Aalto, 1971; Ross et al., 1995) (Fig. 1.5). These rocks crop out locally in the Purcell Mountains and are interpreted to have accumulated in localized rift basins associated with the disaggregation of Rodinia (Stewart, 1972; Ross, 1991). Currently there is no direct geochronological data available to constrain the time of rifting in the southern Canadian Cordillera; however, potential rift correlative volcanic rocks in northern British Columbia and Idaho yield dates ranging from 762–684 Ma (Devlin et al., 1988; Ferri et al., 1999; Lund et al., 2003; Fanning and Link, 2004). These ages correlate well with suspected timing of the initial break-up of Rodinia and support the idea of syn-rift deposition.

Rift facies rocks of the WSG are succeeded by a 5-km thick accumulation of predominantly deep-water siliciclastic rocks with subordinate carbonate rocks. This post-rift succession is termed the Miette Group in the western Rocky Mountains (Charlesworth et al., 1967; Mountjoy, 1980), Horsethief Creek Group in the Purcell Mountains (Reesor, 1973), and the Kaza Group and Isaac Formation in the Cariboo Mountains (Campbell et al., 1973) (Fig. 1.6). These rocks, in turn, are overlain by shallower water strata of the Cunningham and Yankee Belle formations. Collectively, these post-rift rocks are interpreted to represent the passive-margin phase of the WSG, which is manifested by a km-scale upward-shoaling succession from basin-floor to outer shelf deposits (Ross et al., 1989; 1995; Ross and Arnott, 2007). This upward-shoaling succession is interpreted to represent basinward progradation of

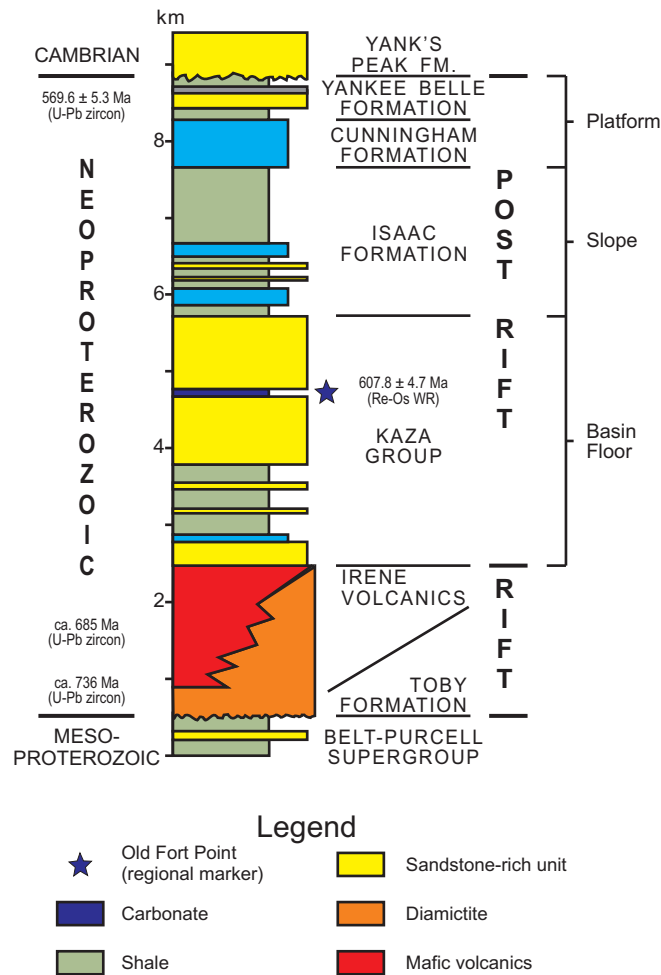


Fig. 1.5. General stratigraphic log of the Windermere in the southern Canadian Cordillera (after Ross et al., 1995). Geochronological dates from Ross et al. (1995), Lund et al. (2003), Kendall et al. (2004), and Colpron et al. (2002).

the western continental margin of Laurentia into the proto-Pacific ocean in response to diminishing rates of thermally-driven subsidence (Ross, 1991).

The timing and age of WSG deposition in the southern Canadian Cordillera is difficult to constrain because of sparse geochronological control (Fig. 1.5). Nonetheless, rare dates exist. Uranium-lead (U-Pb) zircon dates ranging from 740–728 Ma were obtained from volcanic rocks in the crystalline basement that unconformably underlies the WSG (Parrish and Armstrong, 1983; Evenchick et al., 1984; Parrish et al., 1988). Accordingly, these dates

		Cariboo Mountains		Purcell Mountains	Rocky Mountains	Eastern Rockies/ Lake Louise	
		Campbell et al., 1973 Ferguson, 1994		Kubli, 1990	McDonough, 1989 Hein & McMechan, 1994	Aitken, 1969	
Neoproterozoic	Lower Cambrian	Gog Group		Hamill Group	Gog Group	Upper Gog Group	
	Windermere Supergroup	Cariboo Group	Yankee Belle Formation		Jasper Member	Lower Gog Group	
			Cunningham Formation		unconformity		
			Isaac Formation	Horsethief Creek Grp.	Upper Clastic Division	Miette Group	Upper (East Twin Fm.)
			Carbonate Div.				
			Slate Division				
		Kaza Group	Upper	Upper Grit Division	Middle (McKale Fm.)		
			Old Fort Point Fm.	Baird Brook Div.	Old Fort Point Fm.	Taylor Lake Mbr.	
			Middle	Lower Grit Division	Lower (Crushing Creek Fm.)	Corral Creek Formation	
			Lower				
			Mica Creek gneiss	Toby Formation	unconformity	base not exposed	
	Gold Creek gneiss		Belt-Purcell Supergroup	Malton gneiss			

Windermere turbidite system

Fig. 1.6. Chart showing stratigraphic terminology to describe rocks of the WSG in different parts of western Canada. Dashed grey lines in the Upper Twin Formation and Isaac Formation represent the first and second regional carbonate markers (from Ross and Amott, 2007). Copyright ©2007 by The American Association of Petroleum Geologists, reprinted by permission of the AAPG whose permission is required for further use.

represent a maximum age for WSG deposition. A Re-Os age of 608 ± 5 Ma has been reported from black shales in the Old Fort Point Formation (Kendall et al., 2004) that provides the first direct age from within the WSG sediment pile. Rhyolites in the Gog Group that unconformably overlie the WSG have yielded a U-Pb age of 569 ± 4.6 Ma (Colpron et al., 2002), which provides a minimum age for the termination of WSG deposition.

Due to the age of the WSG, biostratigraphic markers are all but absent, making regional correlation a challenge. Fortunately, distinct, regionally extensive shale and

carbonate units exist and make excellent lithostratigraphic markers. The most widespread of these is the Old Fort Point Formation, which is up to 150 m thick and consists predominantly of thin-bedded siliciclastic and carbonate turbidites, with local occurrences of purple and green slates and black shales (Smith, 2009; Smith et al., in press). Two other regional markers occur above the Old Fort Point Formation and are informally termed the first and second Isaac carbonate markers. These units are dominated by interstratified thin- to medium-bedded limestone turbidites and massive calcareous mudstone, with subordinate conglomerate beds, breccias, and large carbonate slump blocks. Units range from 10–250 m thick and are interpreted to be deep-water clastic carbonates deposited during periods of sea level highstand and high rates of shelf carbonate production (Ross et al., 1995; Ross and Arnott, 2007).

Provenance of the deep-water part of the WSG has been interpreted based on by U-Pb dates of detrital zircon grains, which tend to either exceed 2.6 Ga or cluster around 1.9–1.75 Ga (Ross and Bowring, 1990; Ross and Parrish, 1991). These ages suggest that siliciclastic strata of the WSG were derived primarily from fluvial systems that drained the southern Canadian Shield and northwestern United States (Ross et al., 1995; Ross and Arnott, 2007). Moreover, regional mapping of the WSG and paleocurrent data, obtained mainly from current-ripple and dune cross-stratification, suggest transport in a general west-northwest direction (Ross et al., 1995; Ross and Arnott, 2007), although local variants have been reported (Arnott and Hein, 1986; Schwarz and Arnott, 2007a; Khan and Arnott, 2011).

Regional correlation, map patterns, paleocurrent data, and homogeneity of zircon provenance suggest that the deep-water rocks of the WSG were deposited by a single turbidite depositional system of substantial dimension. In fact, when taking into account tectonic shortening during the Mesozoic, the WSG turbidite system may have been of the

order of 80,000 km² or more (Ross and Arnott, 2007). This makes the WSG turbidite system one of the largest known ancient turbidite systems, comparable in size to the modern Amazon and Mississippi fans (Fig. 1.7).

1.3. Castle Creek study area

One of the best exposures of the WSG turbidite system occurs at the Castle Creek study area located in the northern Cariboo Mountains (Fig. 1.3). This area was affected by at least four phases of deformation (Murphy and Rees, 1983; Murphy, 1987a; Reid et al., 2002). The first phase of deformation (D₁) resulted in east-verging, meter-scale folds and a cleavage (S₁) that locally is sub-parallel with bedding. A second phase of deformation (D₂) formed up to kilometre-scale, southwest-verging folds, which represent the dominant regional structure preserved in the area. D₂ also resulted in an axial planar cleavage (S₂) that produced a prevalent crenulation cleavage throughout the area (Murphy, 1987a). A minimum age for D₁ and D₂ is provided by the crosscutting relationship with intrusive igneous rock dated at 174 Ma (Pigage 1977; Gerasimoff, 1988). The third deformation event (D₃) resulted in dextral strike-slip faults, kilometre-scale folds (northeast-verging), and an axial planar cleavage that crenulates S₂ cleavage (Reid et al., 2002). A fourth deformation event (D₄) resulted in locally occurring meter-scale kink folds (Reid et al., 2002). Moreover, metamorphism in the area began during, or shortly following, D₁ and peaked between D₂ and D₃ (Murphy, 1987a). During this period rocks in the Castle Creek area were metamorphosed to sub-greenschist to greenschist grade.

The Castle Creek area occurs on the overturned, nearly vertically-dipping limb of a southwest-verging anticline that formed during D₂ (Fig. 1.8). Despite the regional

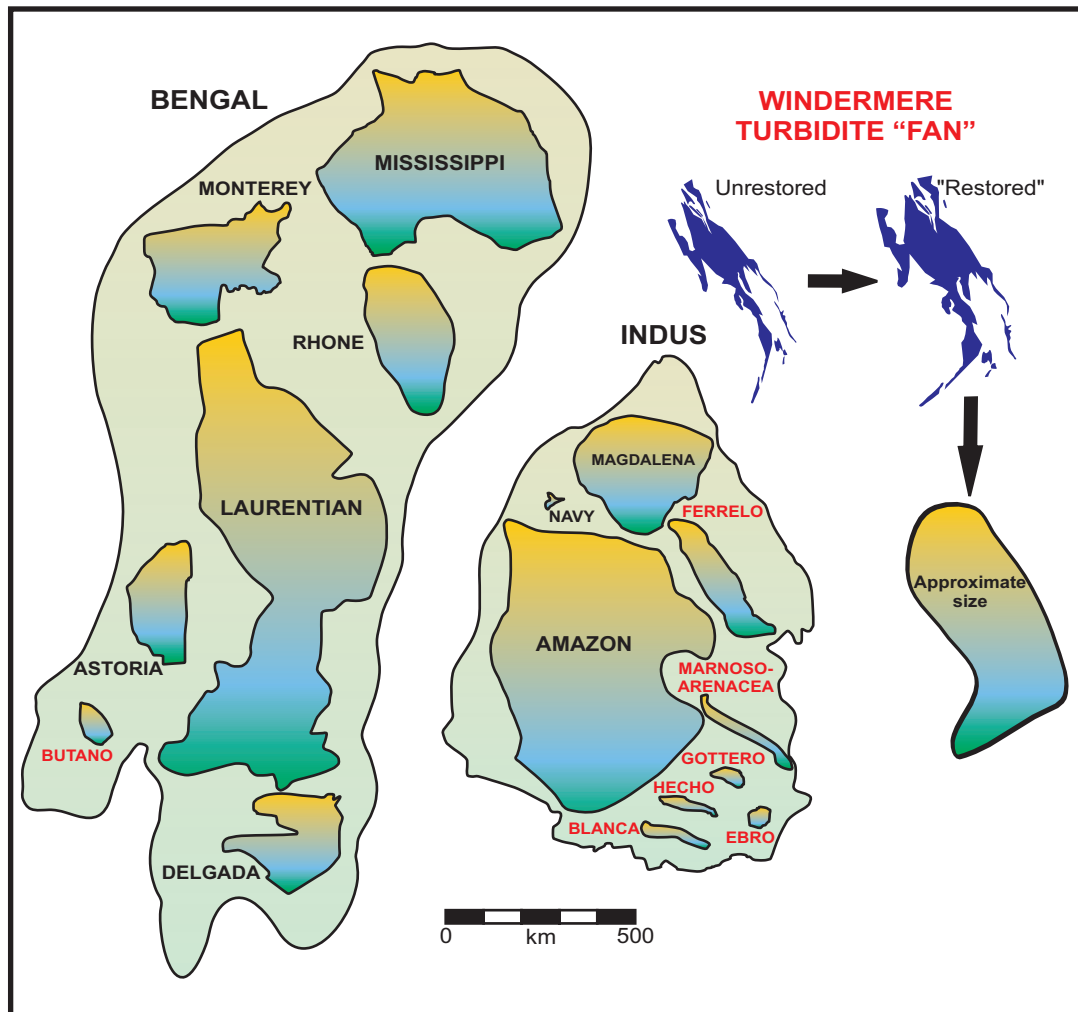


Fig. 1.7. Hypothetical reconstruction of the Windermere turbidite system in relation to other modern (in black text) and ancient (in red text) turbidite systems (redrawn from Ross, 2000, after Barnes and Normark, 1985).

deformation and low-grade metamorphism, primary sedimentary structures are well preserved, with mm-scale features easily visible. Accordingly, rocks in the Castle Creek area are commonly defined using sedimentary and not metamorphic terminology. At Castle Creek, WSG strata crop out over an area 7 km parallel to bedding and 2.4 km perpendicular to bedding. Figure 1.9 shows a 5 km × 2.4 km section of the Castle Creek area. A small alpine glacier, informally called the Castle Creek glacier, divides the study area into south and north regions. A third area occurs farther to the north and is informally termed the hill

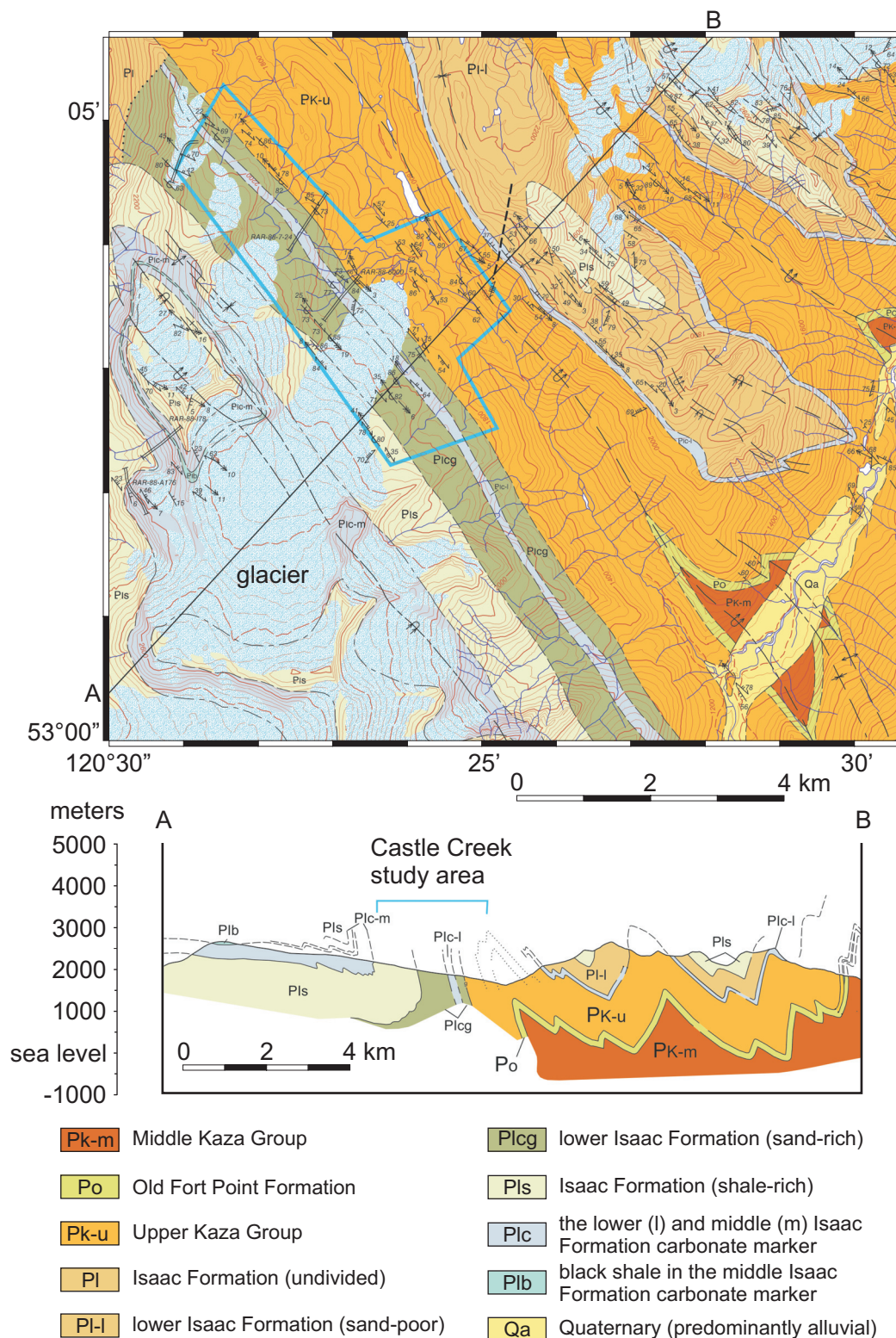


Fig. 1.8 Geology of the Castle Creek area (outlined in blue) and surrounding areas (from Ross and Ferguson, 2003a). Note that the study area occurs on the overturned limb of a large-scale southwest-verging anticline.

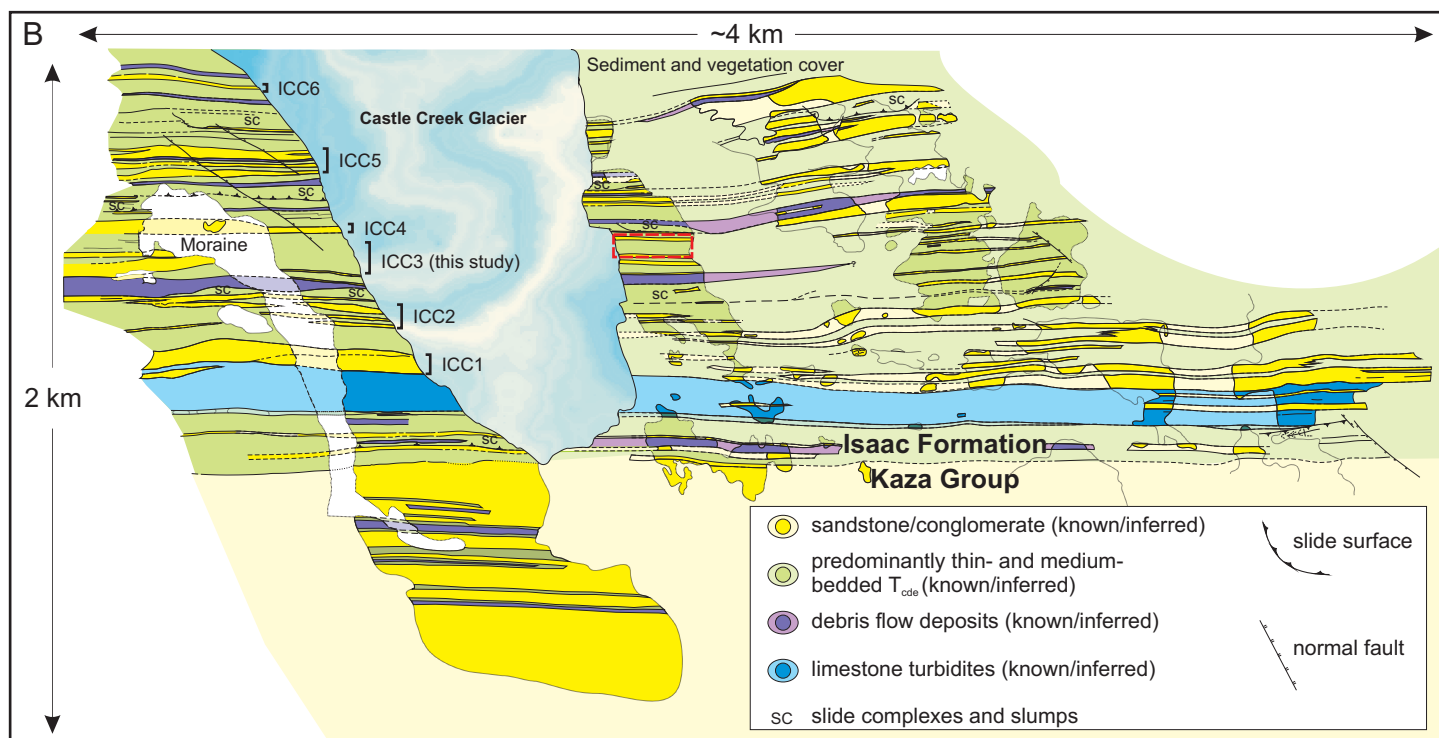
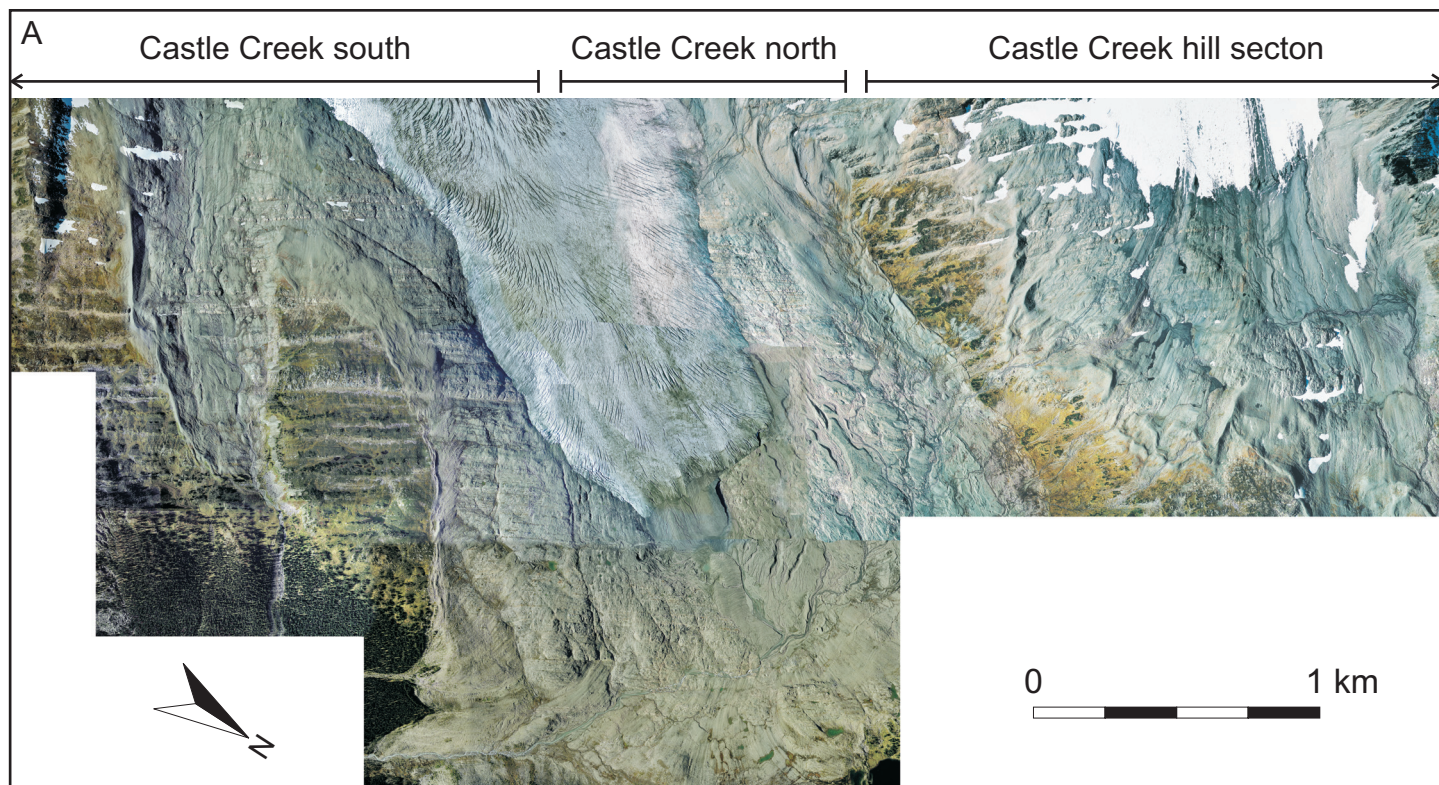


Fig. 1.9. A) Photomosaic of the Castle Creek study area. Isaac Channel B) Interpreted panel of the Castle Creek area. Isaac channel complexes 1–6 (ICC1–6) are indicated (after Ross and Arnott, 2007). Copyright ©2007 by The American Association of Petroleum Geologists, reprinted by permission of the AAPG whose permission is required for further use.

section. The glacier is retreating and freshly exposed strata along its periphery have been polished subglacially and are free of vegetation. Away from the glacier, local debris covers the outcrop and vegetation locally overlies finer-grained rocks.

The lower 800 m of strata exposed at Castle Creek comprises the upper Kaza Group, the lower half of which consists predominantly of massive, normally graded, coarse-grained sandstone beds. Sandstone beds are commonly vertically stacked and form quasi-tabular elements interpreted to represent basin-floor distributary lobes (Meyer, 2004; Meyer and Ross, 2007a, b). Near the top of the Kaza Group, in contrast, wide shallow channel fills with little to no associated levee development become common. In addition, intervals up to 15 m thick composed of thinly bedded, fine-grained turbidites and uncommon debrites that average 2–5 m thick are observed (Rocheleau, 2011; L. Navarro, personal communication).

The Isaac Formation conformably overlies the Kaza Group and is exposed over a stratigraphic thickness of 1.6 km in the study area. It contains several decametre-thick coarse-grained sandstone units that are several hundred meters to a few kilometres wide. These sandstone units are encased on all sides by finer-grained material consisting of interbedded medium to very thin-bedded sandstone and blue to grey siltstone with subordinate thicker (up to 5 m) amalgamated sandstone units. The coarse-grained sandstone units are interpreted to be leveed-channel fills, whereas the surrounding finer grained material is interpreted to represent associated overbank deposits (Ross and Arnott, 2007). Collectively, these channel-levee units are informally termed Isaac Channel Complex 1–6 (ICC1–ICC6). Also observed in the Isaac Formation are common slump, slide, and debris flow deposits up to 135 m thick (Arnott et al., 2011). These deposits are interpreted to reflect the gravitational instability inherent to deep-water slope to base-of-slope settings.

1.3.1. Isaac Channel Complex 3

This thesis focuses on documenting and interpreting levee deposits that occur adjacent the coarse-grained channel-fill of ICC3 (Fig. 1.9). This channel-levee system is interpreted to crop out as an almost perpendicular-to-flow cut through the bend of an 80 m thick, more than 1.8 km wide sinuous deep-water channel-levee system (Navarro et al., 2007a, b) (see Chapter 4).

The interpretation that the finer-grained sediment adjacent the channel-fill strata are levee deposits is supported in part by measured paleocurrent directions (Table 1.1). Specifically, paleocurrent directions along the NW side of the channel fill average 28° , whereas those along the SE side average 80° , both of which are oblique to the main paleocurrent channel trend of 40° (Navarro et al., 2007a, b). The divergence in paleocurrent between the channel and levees is consistent with levee formation resulting from channel flows overspilling the channel banks and continuing in a direction oblique to the main channel trend (DeVries and Lindholm, 1994; Hesse and Dalton, 1995; Hiscott et al., 1997; Hickson and Lowe, 2002; Fildani and Normark, 2004; Hubbard et al., 2008; Kane et al., 2010a). Paleocurrents were measured predominantly from the axes of current ripple cross-stratification. These measurements were made where the outcrop was locally cut forming a 3-D face. The interpretation that ICC3 was most likely sinuous is based on two key observations. Firstly, channel-fill units are vertically stacked and laterally offset, creating a terraced geometry along one side of the channel that sharply separates levee and channel-fill strata (Navarro et al., 2007b) (see Chapter 4). This architecture suggests discrete periods of erosion associated with the systematic lateral shifting of a sinuous channel, followed channel filling (Campion et al., 2000; Beaubouef, 2004; Samuel et al., 2003; Navarro et al., 2007b). Secondly, the levee deposits occurring along one side of the channel (outer bend) are

Table 1.1

Paleocurrent measurements from inner- and outer-bend levee deposits of ICC3

Inner-bend (°)*	Outer-bend (°)**
115	
105	50
85	48
85	45
84	35
83	32
70	30
70	30
54	25
50	22
	20
	18
	15
	10
	5
avg. = 80.1	avg. = 27.5
stdev. = 20.3	stdev = 13.8

* measured from beds within lower 5 m of levee

**measured from beds within proximal levee package 2 (POB2; see Chapter 4)

significantly sandier and thicker than those on the opposite side (inner bend) (Navarro et al., 2007b) (see Chapter 4). This is likely the result of thicker, higher energy flows that preferentially spilled over the outer bend channel margin due to inertial effects (i.e. superelevation and inertial run-up; Imran et al., 1999; Straub et al., 2008).

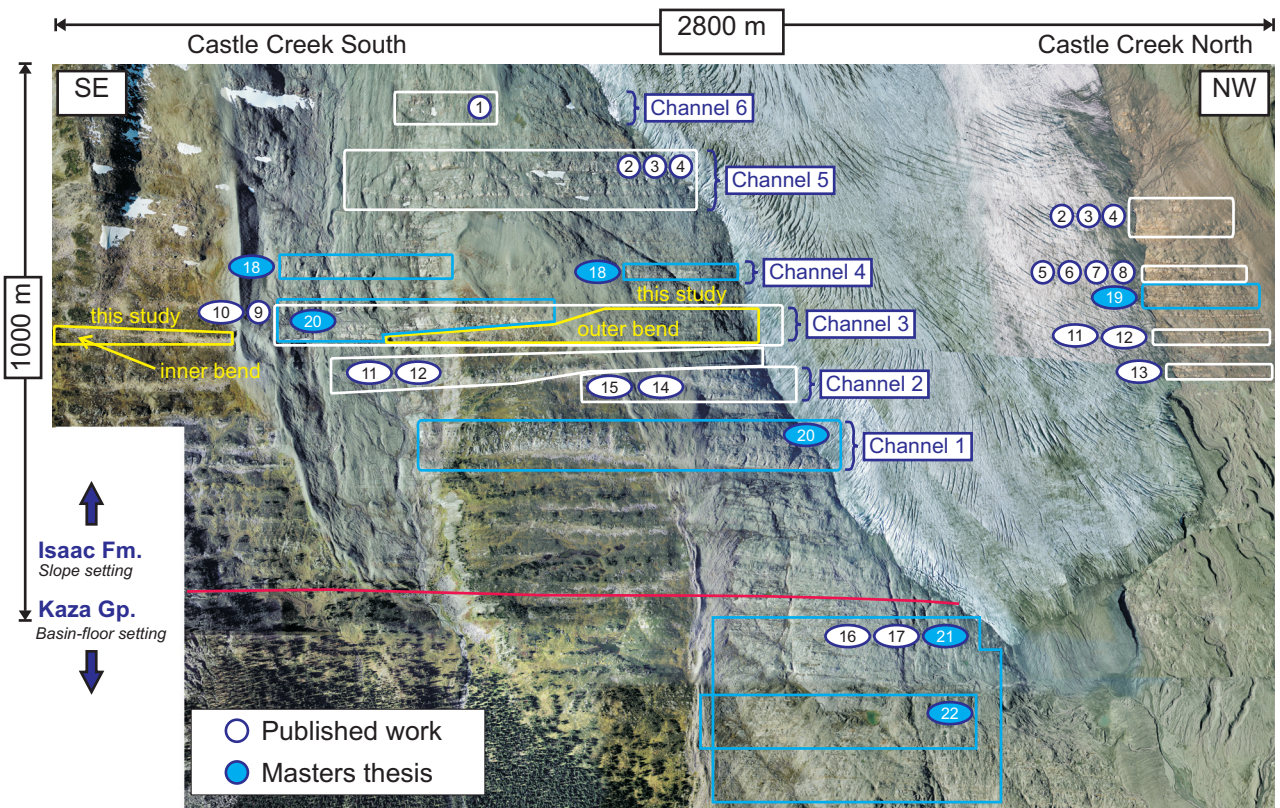
1.4. Previous Studies

Campbell et al. (1973) were the first to map and study the geology of the Castle Creek study area. This study included a 1:250 000 scale map and descriptions of the regional stratigraphy, deformation, and metamorphism. Later, Ross and Ferguson (2003a, b, c, d) mapped the area at a scale of 1:50 000 (Fig. 1.8). In addition to these studies there has been a number of other studies conducted in the Castle Creek and surrounding areas that focused primarily on regional geology and structure (e.g. Murphy and Rees 1983; Murphy, 1987a, b; Ross and Murphy, 1988; Aitken and McDonough, 1990; Ross 1991, 2000; Gabrielse and

Campbell, 1991; Hein and McMechan, 1994; Ross et al., 1995; Reid et al., 2002; Reid, 2003).

Since the beginning of the Windermere Consortium, an industry and government supported research program based at the University of Ottawa, there has been an explosion in the number of studies documenting the detailed sedimentology, stratigraphy, and architecture of deep-water strata in the Castle Creek study area (Arnott and Ross, 2007; Ross and Arnott, 2007). This documentation includes a wide range of basin-floor deposits (e.g. Meyer, 2004; Meyer and Ross, 2007a, b; Rocheleau, 2011), base-of-slope channel-levee complexes (Navarro, 2006; Arnott, 2007a, b; Barton et al., 2007; Gammon et al., 2007; Navarro et al., 2007a, b; O'Byrne et al., 2007a, b; Schwarz and Arnott, 2007a, b, c; Mussa-Caleca, 2008), overbank deposits (Arnott, 2007c, d; Davis, 2011), and mass transport deposits (Laurin et al., 2007; Arnott et al., 2011) (Fig. 1.10). Architectural attributes of Castle Creek have also been used to develop reservoir models (Buckley et al., 2010), which provide important insight to many hydrocarbon charged, subsurface deep-water leveed-channel systems.

Levee deposits have been discussed by several workers from the Castle Creek study area (e.g. Navarro, 2006; Arnott, 2007a, b, c, d; Barton et al., 2007; Gammon et al., 2007; Navarro et al., 2007; O'Byrne et al., 2007b; Schwarz and Arnott, 2007a, b, c; Davis, 2011). However, the detailed documentation and interpretation of these strata was never the primary focus of these investigations (see Fig. 1.10), which is ironic because levee deposits make up most of the exposed strata in the Isaac Formation. This thesis, therefore, focuses entirely on levee deposits within the Castle Creek study area and provides details about their architectural make up and depositional origin.



area	investigator	partial title of work
①	Gammon et al.	Architectural relationship between channels, levees, and debris flows (AAPG Studies in Geology, 2007)
②	Schwarz and Arnott	Anatomy and evolution of a slope channel-complex set: implications for reservoir characterization (JSR, 2007)
③	Schwarz and Arnott	Outcrop characterization of a passive-margin channel-complex set (AAPG Studies in Geology, CD-ROM, 2007)
④	Schwarz and Arnott	Stratigraphic and depositional architecture of a slope channel system (AAPG Studies in Geology, 2007)
⑤	Arnott	Stratigraphic architecture and depositional evolution of a levee to proximal crevasse splay to channel-fill succession (AAPG Studies in Geology, 2007)
⑥	Arnott	Stratigraphic architecture and depositional processes of a proximal crevasse splay and genetically related, sinuous channel fill (AAPG Studies in Geology, CD-ROM, 2007)
⑦	O'Byrne et al.	Architecture of a laterally migrating channel complex (AAPG Studies in Geology, 2007)
⑧	O'Byrne et al.	Architecture of a laterally migrating channel complex (AAPG Studies in Geology, CD-ROM, 2007)
⑨	Navarro et al.	Architecture of a deep-water channel-levee complex (AAPG Studies in Geology, 2007)
⑩	Navarro et al.	Depositional architecture and evolution of a deep-marine channel-levee complex (AAPG Studies in Geology, CD-ROM, 2007)
⑪	Arnott et al.	Stratal architecture and temporal evolution of a passive margin mass-transport deposit (SEPM Special Publication, 2011)
⑫	Laurin et al.	Stratigraphic anatomy and depositional history of a mass-transport complex (AAPG Studies in Geology, 2007)
⑬	Barton et al.	Architectural relationship between channels, levees, and debris flows (AAPG Studies in Geology, 2007)
⑭	Arnott	Architecture of lateral-accretion deposits in two stacked, deep-water, sinuous channel fills: relationship between coarse channel-fill and adjacent inner-bend levee deposits (AAPG Studies in Geology, 2007)
⑮	Arnott	Stratal architecture and origin of lateral accretion deposits (LADs) and conterminous inner-bank levee deposits in a base-of-slope sinuous channel (Marine and Petroleum Geology, 2007)
⑯	Meyer and Ross	Channelized lobe and sheet sandstones (AAPG Studies in Geology, 2007)
⑰	Meyer and Ross	Channelized lobe and sheet sandstones (AAPG Studies in Geology, CD-ROM, 2007)
⑱	Mussa-Caleca	Architecture and depositional history of a deep-water slope channel complex (2008)
⑲	Davis	Architecture of deep-marine interchannel deposits (2011)
⑳	Navarro	Depositional architecture and evolution of deep-water base-of-slope and slope channel complexes (2006)
㉑	Meyer	Internal architecture of an ancient deep-water, passive margin, basin-floor fan system (2004)
㉒	Rocheleau	Depositional architecture of a near-slope turbidite succession (2011)

Fig. 1.10. Photomosaic of the Castle Creek study area showing different areas of previous research. This thesis focuses on the inner- and outer-bend levee deposits associated with Channel 3 (after L. Navarro, unpublished).

1.5. Thesis structure

This thesis is a combination of a monograph thesis and a series of articles. Specifically, chapters 2 and 3 are literature reviews whereas chapters 4 to 6 are written in paper format. Chapter 2 reviews the different types of gravity flows common to deep-water settings, with special emphasis on turbidity currents. Chapter 3 briefly reviews the history of deep-water fan models and the elements that are most common in mud-rich submarine systems, especially in their overbank areas. Chapter 4 documents the sedimentology and stratal attributes of both inner- and outer-bend levee deposits and uses these details to construct a channel-levee evolution model. Chapter 5 expands on Chapter 4 and presents a novel model for the origin of levees. Additionally, this chapter discusses the effects of compaction on ancient stratigraphy, and how the dipping reflections observed on seismic images may in fact be produced by non-dipping strata. Chapter 6 presents a series of flume experiments that document the conditions necessary to produce climbing and non-climbing ripple cross-stratification. The results from these experiments are compared to observations in the ancient rock record. Chapter 7 combines the main conclusions of the previous chapters into a succinct levee evolution model and discusses areas for future research.

1.6. Contribution of co-authors

Chapter 4 is an expanded version of:

Khan, Z. A., and R. W. C. Arnott, 2011, Stratal attributes and evolution of asymmetric inner- and outer-bend levee deposits associated with an ancient deep-water channel-levee complex within the Isaac Formation, southern Canada: *Marine and Petroleum Geology* 28, p. 824–842 (see Appendix A).

This paper is based principally on fieldwork carried out by Zishann Khan (ZK) with assistance in the field from Marc Beauchamp, Claude Farley, Steven Gould, Diana Kuiper, Joel Laurin, Melisa Martel, and Benoit Saumur. Data were processed and figures were produced by ZK except where the original publication is referenced. The initial manuscript was written by ZK and later revised through discussions with Dr. Arnott. This paper also benefited from discussions with Lilian Navarro, specifically during the writing of a previous co-authored AAPG publication (Navarro et al., 2007a, b). The Marine and Petroleum Geology manuscript was submitted on the 29th of November, 2009, and underwent further modification during the peer-review process and after revisions was accepted on the 13th of July 2010.

Chapter 5 reproduces work included in a paper submitted to the American Association of Petroleum Geologists Bulletin:

Khan, Z. A., R. W. C. Arnott, and A. Pugin, 2011, An alternative model of producing topography in the crest region of deep-water levees: AAPG Bulletin, v. 95, p. 2085–2106 (this article is not included as an appendix due to copyright restrictions).

This paper expands on the Marine and Petroleum Geology paper and uses the same original data set collected by ZK with assistance in the field from Marc Beauchamp, Claude Farley, Joel Laurin, Melisa Martel, and Benoit Saumur. The seismic model (fig. 10) shown in this paper was developed by Dr. Pugin using parameters defined by ZK. The remaining figures were drafted by ZK except where referenced in the manuscript. Discussions between ZK and Dr. Arnott helped shape the central ideas of this text. The manuscript was written by ZK and later modified using the suggestions of Dr. Arnott. This research paper was submitted the 1st of March, 2010, a first revision of the paper was submitted the 19th of October, 2010, a second revision of the paper was submitted 28th of February, 2011, and was accepted on the 28th of March, 2011.

CHAPTER TWO

Deep-water sedimentation processes and their deposits

2.1. Introduction

Like outer space, the deep-sea has always sparked the curiosity of humankind, providing an outlet for uninhibited imagination as best epitomized in the works of Jules Verne. The reconnaissance voyage of HMS Challenger from 1872–1876 was the first to establish an accurate description of the general morphology of the world's ocean basins and confirmed that the seabed at depth is blanketed in fine-grained clayey sediment.

Accordingly, at the dawn of deep-sea exploration it was commonly thought that the deeper parts of oceans were rather calm settings where sediment accumulated largely through pelagic fallout and biologically related processes (Murray and Renard, 1891). Up until the end of the first half of the twentieth century it was widely accepted that sandstone deposits were generally exclusive to continental and shallow water environments. During the latter half of the twentieth century, however, it became increasingly evident that voluminous amounts of sand and coarser sediment are transported into deep-water basins by various subaqueous gravity-driven processes (Kuenen and Migliorini, 1950; Ericson et al., 1951).

Downslope movement of enormous volumes of shallow-water sediment are commonly initiated by failure along the shelf edge (Shanmugam, 2006). This condition occurs when the shear strength of a mass of sediment, or in many cases a particular stratal layer within it, is exceeded by the downward force of gravity. Loss of shear strength is initiated by a variety of mechanisms, including 1) high slope gradients (e.g. near the mouth

of the Magdalena River, Colombia; Heezen, 1956); 2) high rates of sedimentation (e.g. the Mississippi delta front, Gulf of Mexico; Coleman and Prior, 1982); 3) earthquake disruption (e.g. 1929 Grand Banks tremors, east coast Canada; Heezen and Ewing, 1952); 4) salt movement (e.g. intraslope basins, Gulf of Mexico; Tripsanas et al., 2004); 5) glacial loading (e.g. the Scotian margin, North Atlantic; Mosher et al., 2004); 6) tsunamis (e.g. Gutenberg, 1939); 7) storm waves (e.g. Henkel, 1970); 8) biological erosion (e.g. along submarine canyon walls; Warme et al., 1978); and 9) gas generation due to organic decay (e.g. Scripps submarine canyon head, west coast U.S.A; Dill, 1964).

Once a mass of sediment becomes mobile it commonly disaggregates progressively downslope and evolves into a fast moving sediment flow, such as a debris flow or a turbidity current (Fig. 2.1). In this chapter the classification scheme of Shanmugam (2006) is followed where subaqueous gravity-driven processes are subdivided into two kinds: mass movements and sediment flows (Table 2.1).

2.2. Mass movements

Shanmugam (2006) defines mass-transport processes, or mass movements, as a general term that refers to the failure, dislodgement, and downslope transport of coherent material under the influence of gravity in both subaerial and subaqueous settings. These events mainly comprise slides, slumps, and rock falls. Rock falls do not make up a significant part of the deep-water record and are not discussed here.

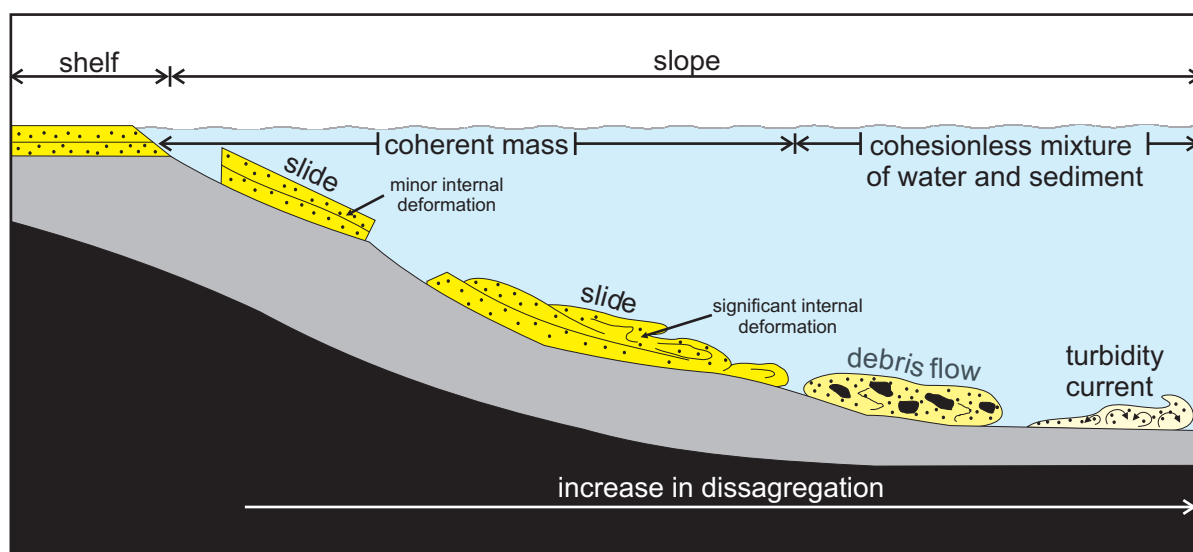


Fig 2.1. Schematic showing the typical downslope evolution of a gravity-driven sediment failure at the shelf edge from a coherent slides to a cohesionless water-sediment flow. Slides result from the failure of unstable material that moves downslope under gravitational forces. Internal deformation associated with slides is minimal and typically increases downslope where slides evolve into slumps. Uptake of water during downslope movement may lead to the transformation of slumps into debris flows. Increased fluid content causes debris flows to evolve into turbidity currents. Gravity-driven sediment failures do not always evolve in this manner and any one of these gravity-driven mass movements or sediment flows may evolve directly from sediment failure near the shelf edge, on the slope, and in the distal basinal settings (redrawn from Shanmugam et al., 1994). ©Copyright 1994. The American Association of Petroleum Geologists. All rights reserved. Reprinted by permission of the AAPG whose permission is required for further use.

Table 2.1

A Classification of subaqueous gravity-driven deposits (after Shanmugam, 2006)

Major type	Subtype	Nature of moving material	Fluid viscosity	Flow state
Mass movement	Slide	Coherent mass without significant internal deformation (mainly translational movement)	Not applicable	Not applicable
	Slump	Coherent mass with internal deformation (translational and rotational movement)	Not applicable	Not applicable
Sediment flow	Debris flow	Incoherent body matrix supported sediment in sediment-water slurry	Plastic fluid	Laminar
	Turbidity current	Incoherent body suspended sediment in sediment-water flow	Newtonian fluid	Turbulent

2.2.2. *Slides*

Submarine slides occur when the shear strength of a coherent mass of semi-consolidated to unconsolidated sediments is overcome by gravitationally-driven shear stress and moves along a planar glide plane. Shear forces are typically localized along the basal slip plane and as a result structures and fabrics within the slide material show minimal evidence of deformation apart from some minor faulting and folding. Sufficient overloading of sediment is one way in which shear stress can exceed the shear strength of material along a discrete plane. This condition can be achieved during episodes of sea-level fall and tectonic uplift where rapid sedimentation superimposes fine-grained material over underconsolidated shelf sediments (Stow et al., 1996). This favours the build-up of excess pore pressures along weak layers to the point of failure. Failure of these sorts commonly occurs on the front side of deltas where sedimentation rates are high due to input from sediment-laden rivers (e.g. Prior and Coleman, 1984; Cohen and McClay, 1996). Failure can also be triggered by earthquakes and, in shallow water, storm waves. Moreover, the production of biogenic gases in organic-rich slope sediments and the release of methane from crystallized gas hydrates within sediments have been reported to locally diminish shear strength, causing large mass movements (e.g. Haq, 1993; Paull et al., 1996; Maslin et al., 1998). Slides move downslope until frictional resistance at the base of the mass exceeds the downward directed force of gravity.

Slides can involve local movement of small, meter-sized blocks or involve the displacement of kilometre-scale masses over hundreds of kilometres. For example, slide deposits observed in the Nuuanu Slide moved 230 km seaward from the shore of Hawaii (Normark et al., 1993). Similarly, slide blocks off the coast of northwestern Africa have travelled nearly 300 km from the shelf edge (Jacobi, 1976). Equally impressive as the long-

distance transport are the dimensions of deposits associated with some debris flows. For example, the massive Storegga Slide deposit on the southern flank of the Voring Plateau offshore Norway reaches a thickness of 430 m and a length of more than 800 km (Bugge, 1983; Jansen et al., 1987).

2.2.3. Slumps

Slumps involve the movement of a coherent mass of unconsolidated material and share the same triggering mechanisms as slides. In fact, slumps commonly evolve from slides and therefore the boundary between these two mass movements is not easily defined. In general, slumps are distinguished from slides by their higher degree of internal deformation (Stow et al., 1996). The higher degree of deformation associated with slumps is caused primarily by the development of multiple shear planes, which cause parts of the slump to move at different speeds relative to other parts. This results in local folding and faulting. Local deformation within a slump also occurs due to loading, which results from density inversions between layers of weak sediment. Slump deposits occur in a variety of scales similar to slides.

2.3. Sediment flows

Sediment flows are mixtures of water and sediment grains that move downslope under the influence of gravity. They are initiated when particles within a mass of sediment become suspended in the interstitial fluid, creating a sediment dispersion that is denser than the surrounding fluid. Gravity acts on the density difference between the two fluids and moves the denser one down-gradient, which on a dipping surface is downslope. The manner

in which particles are supported in a flow provides a basis for subdividing sediment flows into two common kinds – debris flows and turbidity currents (Dott, 1963; Sanders, 1965; Middleton and Hampton, 1973; Dingle, 1977; Shanmugam et al., 1994; Shanmugam and Miola, 1995). In debris flows, sediment concentration is high and multiple support mechanisms, including buoyancy, dispersive pressure, cohesion, and hindered settling, operate to maintain clasts in suspension (Middleton and Hampton, 1973; Hampton, 1975; Pierson, 1981). In turbidity currents, on the other hand, sediment concentrations are relatively low, and particles are held in suspension mainly by the upward component of fluid turbulence (Sanders, 1965; Leeder, 1999; Shanmugam, 2006).

2.3.1. Fluid viscosity

Viscosity is another useful parameter to sub-divide sediment flows into Newtonian fluids, represented by turbidity currents, and non-Newtonian fluids, represented by debris flows (Dott, 1963). Viscosity, or more specifically shear viscosity, describes the relationship between applied shear stress (τ) and the rate of internal shear strain. The rate of shear strain is manifested as a velocity gradient formed perpendicular to the direction of shear, and is commonly denoted as du/dy . Fluids with no inherent strength begin to deform immediately after a shear stress is applied and continue to deform so long as the shear stress exceeds zero (Middleton and Wilcock, 1994). Such fluids are considered to be Newtonian when the rate of strain (du/dy) is directly proportional to the amount of applied shear stress (τ) (Fig. 2.2).

This behaviour can be described as

$$\tau = \mu \left(\frac{du}{dy} \right) \quad (2.1)$$

where the constant of proportionality (μ) is the dynamic viscosity. Non-Newtonian plastic flows, in contrast, initiate only after a critical yield stress (τ_0) is exceeded and come to rest once the applied shear stress falls back down below the yield strength of the material (Pierson and Costa, 1987; Phillips and Davies, 1991; Leeder, 1999). If deformation is linear after the yield stress of the plastic is exceeded then the material is considered to be an ideal Bingham plastic (Fig. 2.2) and can be described as:

$$\tau = \tau_0 + \mu \left(\frac{du}{dy} \right) \quad (2.2)$$

2.3.2. Flow state

During the late nineteenth century Osborne Reynolds demonstrated through a series of experiments the occurrence of two flow states – laminar and turbulent (Reynolds, 1883; Leeder, 1999). These experiments involved passing water through a section of a smooth pipe at various speeds. Reynolds observed that the pressure loss over a length of pipe increased with velocity but at a certain flow speed losses increased more rapidly. This observation suggested that at this critical velocity the pattern of flow changed. Reynolds confirmed this qualitatively by injecting a dye into the flowing water. At low flow velocities the thread of dye travelled in a straight line. This flow pattern, wherein macroscopic mixing is negligible across layers perpendicular to the main flow direction, is called laminar flow. Conversely, at high velocities the dye streak dispersed into chaotic eddies and quickly coloured the entire flow. This chaotic type of flow, in which pervasive mixing occurs between parallel layers, was termed turbulent flow. Reynolds found that the change from laminar to turbulent flow occurred at a fixed ratio between the inertial and viscous forces acting on the fluid. Viscous forces resist deformation in the fluid whereas inertial forces represent the ability of a fluid

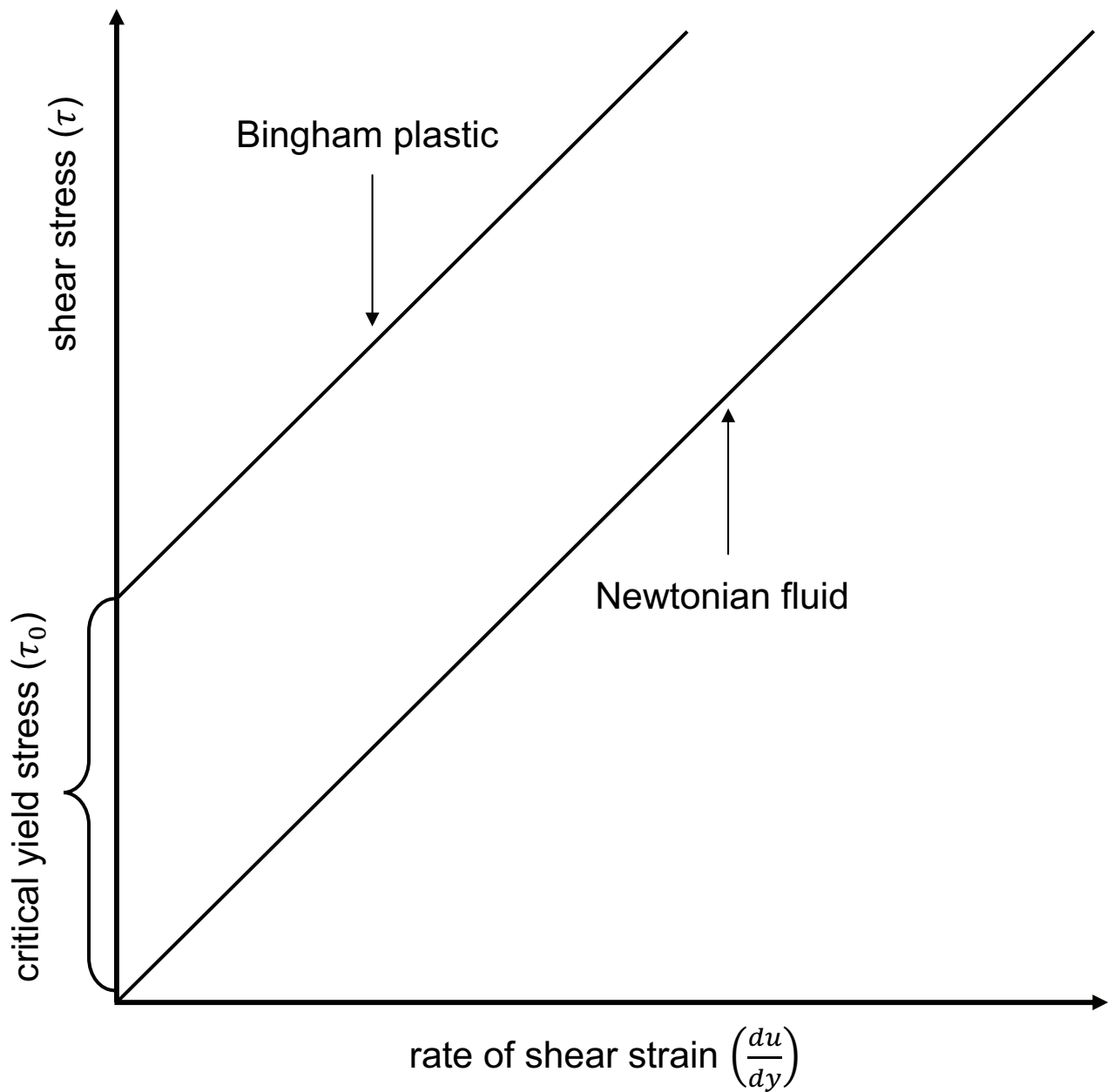


Fig 2.2. Plots of shear stress (τ) against rate of shear strain (du/dy) for Newtonian fluids (turbidity currents) and Bingham plastics (debris flows). The fundamental difference between these two flow types is that Newtonian fluids exhibit no strength and deform immediately when a shear stress is applied whereas plastic fluids only deform after a threshold shear stress is applied (critical yield stress).

mass to accelerate. The ratio between these two is termed the Reynolds number (Re) and is conventionally written as:

$$Re = \frac{\rho d \bar{u}}{\mu} \quad (2.3)$$

where ρ is the fluid density, $\bar{u}$ the mean flow speed, and d the mean flow depth in free-surface liquid flows (e.g. flow in a river channel). In general the transition from laminar to turbulent flow typically occurs between Reynolds numbers of 500 and 2000.

2.3.3. Eddy viscosity

Turbulent eddies characterize turbulent flow and have a 3-D form with constantly varying x-, y-, and z-components of velocity. As a result, when a sensitive flow measuring device like an Acoustic Doppler Velocity meter (ADV) is inserted into a turbulent flow it gives a fluctuating record of fluid velocity. Turbulent fluctuations (i.e. eddies) move volumes of fluid locally within the flow. This suggests that mass, and thus by extension fluid momentum, is being transferred within the flow. Of note is that the rate of momentum transfer is a force (i.e. $F = \frac{mv}{t} = ma$, where mv is momentum). Accordingly, this sets up stresses in turbulent flows that are additional to those created by dynamic viscosity. These additional turbulent stresses, which are created by the local velocity gradient within the flow, work against the mean velocity gradient and dissipate fluid energy, ultimately converting it to heat, and are referred to as the eddy viscosity (Boussinesq, 1877). In a turbulent flow the viscous shear stress equation is expressed as:

$$\tau = (\mu + \eta) \frac{du}{dy} \quad (2.4)$$

where η is the eddy viscosity. Unlike μ in a Newtonian flow, η is a variable dependent on the size and velocity components of the eddies in the flow. Eddy viscosity represents the observed increase in energy losses in Reynolds' experiments and differentiates turbulent from laminar flow.

2.3.4. *Debris flows*

Debris flows are plastic, laminar flows that typically consist of a varying proportion of clasts dispersed within a finer-grained matrix. Sediment and water are fully mixed and as a result there is rarely any relict bedding or lamination preserved except for in some larger intact clasts. Mud-rich debris flows that contain no or only a small number of clasts are commonly referred to as mudflows (Stow et al., 1996). The term olistostrome is used to refer to very thick, extensive debris flow deposits that contain decametre scale exotic blocks (Flores, 1955). Debris flows occur in variety of sizes, from centimetre to kilometre scale.

The cohesive strength of a debris-flow matrix is predominantly a result of friction and interlocking of clay-sized particles (Bryant, 2005). Accordingly, it has been argued that debris flows require a substantial clay-size fraction in order to provide sufficient shear strength to the matrix, or simply, matrix strength (e.g. D'Agostino and Jordan, 1997). Other authors, however, suggest that a clay component as little as 2% provides sufficient matrix strength (Hampton, 1975; Costa and Williams, 1984; Shanmugam, 2000). In addition to matrix strength, clasts in debris flows are supported by dispersive pressure from upward-directed intergranular collisions, buoyancy that results from the high density mud-water mixture, and hindered settling caused by the uprush of fluid associated with downward-settling grains (Shanmugam, 1996; Major and Iverson, 1999).

Debris flows commonly evolve from highly mobile slumps where progressive fragmentation and disintegration has led to the creation of a well defined mud-rich matrix. Debris flows also develop instantly when the shear strength of a porous mass is suddenly overcome by the overloading of other mass movements (e.g. slide, slumps) or debris flows originating higher up on the slope. This process has been interpreted to have caused a large mass movement in the southern Gulf of California (Moore et al., 1982). Here, coastal sand and gravel were initially transported as a debris flow downslope to a depth of about 2500 m where they overloaded a highly porous (70–80%) layer of siliceous silty clay. This triggered a second, significantly larger mud-rich debris flow that carried the deposit of the older mass flow further down slope to a depth of 3000 m. Impressively, this mass movement spread its load over an area of about 300 km², forming a sheet that locally is up to 100 m thick. Sudden loading by an older mass movement initiated upslope is also interpreted to have triggered the Canary debris flow, which traveled about 700 km from the West Africa slope to the Madeira abyssal plain (Mason et al., 1998).

The impressive run out distance of some debris flows is likely attributed to lubrication between grain contacts by the watery clay mixture of the matrix (Leeder, 1999). In addition, many mud-rich debris flows slide over a layer of water in a process termed hydroplaning (Mohrig et al, 1998; Shanmugam, 2000). This condition is set by the low permeability of the mud-rich make up of the flow which inhibits water intrusion. As a result, at the flow front water rides over the top but also penetrates beneath the debris mass separating it from the bed. This basal lubricating layer of water creates an almost frictionless surface over which the flow rides and enables long-distance downslope movement.

Local shearing commonly occurs through a debris flow; however, in highly viscous flows shear stress is sometimes only large enough to create a velocity gradient along the

margins and base (Reading, 1996). Consequently, the central part of these highly viscous flow moves along as a rigid plug. Furthermore, thinning at the lateral edges of many viscous debris flows reduces local shear stress and results in the deposition of levees that confine the central, more mobile part of the flow. Less viscous flows, in contrast tend to spread their load more radially forming extensive sheet deposits.

Turbulence generated along the leading edge of a subaqueous debris flow typically casts sediment into suspension, which in turn may spawn a turbidity current that overlies and ultimately overshoots the parent flow (Hampton, 1972; Mohrig et al., 1998). Elmore et al. (1979) described a modern deposit that is interpreted to have formed by such a process. This deposit occurs on the Hatteras abyssal plain in the western Atlantic and is 500 km long, 100 km wide, and up to 4 m thick. Its proximal reach consists of poorly sorted sand in a mud-rich matrix interpreted to be a debris flow deposit, whereas immediately overlying and around its periphery are sediments interpreted to have been deposited by an associated turbidity current.

2.3.5. Turbidity current

Turbidity currents are Newtonian fluids in which sediment is supported largely by the upward directed component of turbulent eddies (Bagnold, 1966; Leeder, 1983; Simpson, 1987) creating a sediment-water mixture that is denser than the surrounding ambient fluid, which in marine settings is seawater. As previously mentioned, gravity acts on the density difference between these two fluids and transports the denser fluid downslope, sometimes over several hundred kilometres (e.g. Heezen and Ewing, 1952; Heezen, 1956; Piper et al, 1988; Mulder et al., 1997). Turbidity currents are not restricted to deep-marine environments and occur also in lakes, fjords, and estuaries (Simpson, 1987). Powerful powder snow avalanches represent turbidity currents generated in air (Hopfinger, 1983).

Turbidity currents were first documented in lakes (e.g. Lake Geneva) as early as the late nineteenth century (Forel, 1885). Daly (1936) and Johnson (1938) suggested that erosion by turbidity currents during sea level lowstands was the main cause for the origin of continental-margin submarine canyons. Notwithstanding, the concept that turbidity currents transport and deposit large amounts of sand in deep water was only first introduced in 1950 by Kuenen and Migliorini. In their seminal publication, results from flume experiments on turbidity currents were used to suggest that normally graded sandstone beds in the Italian Apennines were deposited by such currents. Although Kuenen and Migliorini's idea sparked considerable debate (e.g. Walker, 1973), turbidity currents are now largely considered to be the principal transport agent of sand and coarse sediment into deep ocean basins.

2.3.6. Initiation of turbidity currents

Slope failures are suspected to be a common means by which turbidity currents become initiated (Normark and Piper, 1991). A spectacular example is the Grand Banks failure of 1929, where an earthquake generated a large slump that is interpreted to have progressively disintegrated downslope into a debris flow that then evolved into a turbidity current (Piper et al., 1985; Piper et al., 1999a). This flow sequentially broke submarine telegraph cables along its basinward flow path up to 13 hours after the initiation of the flow. The distance and time over which these breaks occurred suggests that the flow travelled up to 70 km/hr (Piper and Aksu, 1987; Piper et al., 1988; Hugues-Clarke et al. 1990). Also, it was estimated that the flow was upwards of several hundred meters thick and had a flow volume of about 185 km³. Remarkably, that is enough sediment to cover the entire land surface of Canada in a layer of sediment just over 2 cm thick. More recently, the Nice airport slope

collapse of 1978 generated a turbidity current that deposited sediment up to 200 km basinward of the shelf edge on the Var fan (Mulder et al., 1997).

Another initiating mechanism for turbidity currents is build-up of sediment transported by longshore drift and storm surges into the heads of nearshore canyons that trigger downslope sediment flows (Fukushima et al., 1985; Normark et al., 1998; Einsele, 2000). This process most commonly produces turbidity currents that contain a high proportion of sand. Dilute mud-rich flows, however, are also generated in these settings (Parsons et al., 2003). Longshore drift feeding sand directly to the heads of modern submarine canyons is well documented along the coast of California and southeast Australia (e.g. Covault et al., 2007; Boyd et al., 2008; Covault and Graham, 2010).

A third way in which turbidity currents originate is from hyperpycnal river plumes that feed sediment directly onto the slope (Mulder and Syvitski, 1995; Mulder et al., 1998). In this scenario, rivers with extremely high suspended sediment load at their mouths form inertial flows that transform seaward into density underflows (i.e. hyperpycnal flows) and ultimately into turbidity currents (Prior et al., 1987; Prior and Bornhold, 1988; Mulder and Syvitski, 1995). The modern Yellow river (Huang He) of China is an excellent example of a river that generates semi-permanent hyperpycnal flows (Wright et al., 1986; Van Gelder et al., 1994), although there is no evidence of substantial sand deposition in its delta front region (Mutti et al., 2003). Therefore, it is commonly argued that hyperpycnal flows can only generate silt- or mud-rich turbidity currents (Leeder, 1999). Turbidity currents derived from hyperpycnal flows can last up to weeks whereas those that evolve from mass movements can last up to tens of hours but, typically, have a much shorter duration (Normark and Piper, 1991).

Direct observation of turbidity currents remain rare but have been documented in the Zaire valley (Khripounoff et al., 2003), in a British Columbia fjord (Prior et al., 1987), and in several submarine canyons offshore California (e.g. Inman et al, 1976; Parsons et al., 2003; Paull et al., 2003). A consequence of the apparent rarity of turbidity currents (periodicity of years to tens of thousands of years) and the inherent difficulties in observing them directly is that much of our knowledge about these flows is derived from laboratory work, theory, and field observations.

2.3.7. Structure of a turbidity current

The structure of turbidity currents is commonly studied experimentally in flumes by releasing a dense fluid into a less dense ambient fluid, which is typically clear water. The dense fluid is usually a brine, cold water or a sediment-water mixture. Immediately upon entering the ambient water the dense plume moves along the base of the flume and forms a well developed head, body, and tail (Fig. 2.3). The head has an overhanging nose that forms in response to frictional forces exerted along the lower and upper boundary of the flow. Moreover, a brain-like array of bulbous lobes and clefts forms at the contact between the overhanging nose and lower boundary (Simpson, 1972). Ambient water mixes with the flow under the lobes and through the clefts (Fig. 2.4). Entrainment by this process, however, is considerably less significant than the mixing that takes place towards the rear of the head through a series of transverse vortices termed Kelvin-Helmholtz instabilities (Britter and Simpson, 1978) (Fig. 2.5).

The head area is typically thicker, and moves slower than the rest of the flow along slope gradients $>0.13^\circ$ (Komar, 1972). This enables the faster moving body of the flow to continuously feed sediment to the head, which helps the head counterbalance boundary

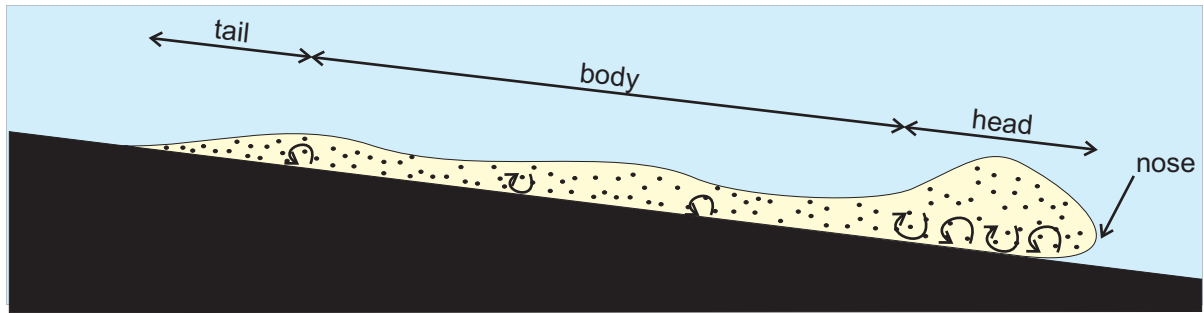


Fig 2.3. Illustration of a turbidity current showing the head, body, and tail regions. Note the overhanging nose at the front of the head that forms in response to frictional forces between the lower boundary of the flow and the static bed and the upper boundary of the flow and ambient fluid.

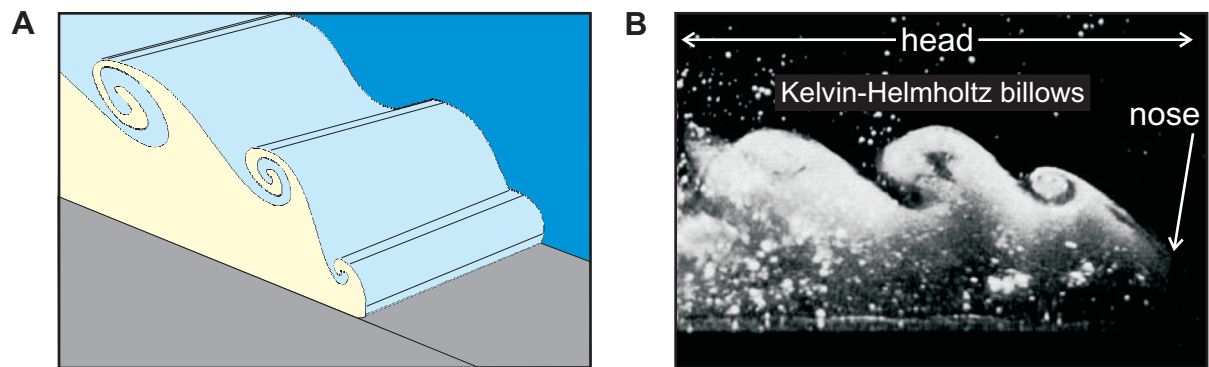


Fig. 2.4. Instability at the front of a turbidity current head. A) The brain-like lobes and clefts that form at the contact between the overhanging nose and solid boundary (redrawn from Simpson, 1987). B) A schematic view of lobes and clefts viewed from below (redrawn from Kneller and Buckee, 2000). Used by permission of John Wiley and Sons.

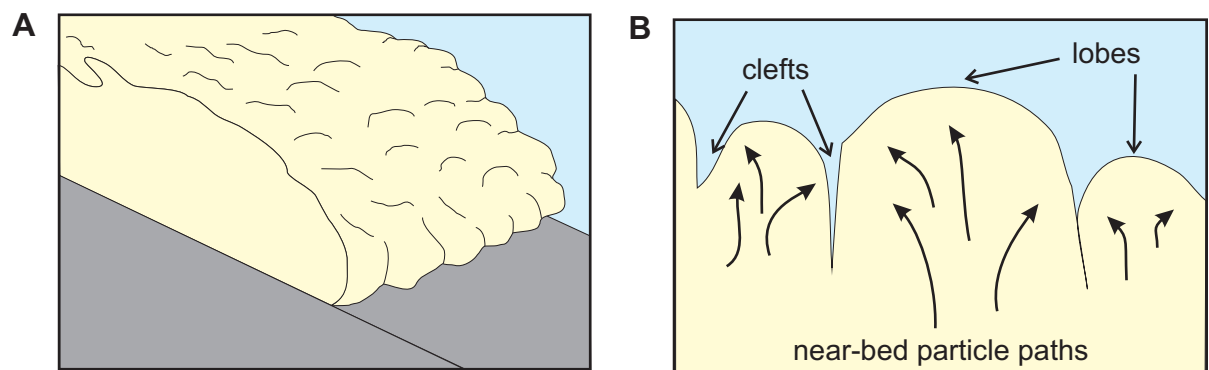


Fig 2.5. Instability that takes place between the upper boundary of the flow and the ambient fluid. A) Schematic illustrating the developing Kelvin-Helmholtz billows (redrawn from Simpson, 1987). B) Image of the head of a saline gravity current showing well developed Kelvin-Helmholtz billows (after Simpson, 1969). Used by permission of John Wiley and Sons.

friction effects and density loss due to ambient fluid entrainment (Hiscott, 2003).

Accordingly, a head that is supplied with a near constant influx of sediment from the body will flow steadily with time. River-supplied hyperpycnal flows are an example of a current that has a potentially constant influx of sediment. Conversely, surge-like flows typically decelerate because the body of the flow can only supply a finite amount of sediment causing the head to shrink and eventually completely dissipate. Some surge-like flows, however, may flow for long periods of time and over great distances by achieving autosuspension (Bagnold, 1962, 1963; Pantin, 1979; Parker et al., 1986). This condition occurs when turbulence derived from power expended against bottom drag is sufficient to support the suspended load. Pantin (1979) suggested that autosuspension occurs when

$$\frac{e_x \alpha u}{u_s} \geq 1 \quad (2.5)$$

where e_x is flow efficiency (the proportion of power expended against bottom friction available for suspension), α the bed slope, u is the forward flow velocity, and u_s the suspended grains fall velocity. Therefore, efficient, fast moving flows down a steep slope with a fine-grained suspended load favours autosuspension. In addition, erosion of the bed by the moving current, especially under its head, helps to maintain the integrity (i.e. its density contrast) of the current, and in doing so causes the current to accelerate or “ignite” (cf. Parker et al., 1986).

2.3.8. *Velocity of turbidity currents*

The velocity of turbidity currents is calculated by treating the surging dense fluid as a shallow-water wave. This implies that the conversion of potential energy to the kinetic energy of motion gives:

$$u \propto \sqrt{gd} \quad (2.6)$$

where d is the flow thickness, and g acceleration due to gravity. Velocity of the turbidity current will further depend on the density difference between the flow and the ambient water. This is taken into account by replacing g with:

$$g' = \frac{\Delta\rho}{\rho_a + \Delta\rho} g \quad (2.7)$$

where g' is the reduced gravity, ρ_a the density of ambient fluid, typically seawater, and $\Delta\rho$ difference in density between the turbidity current and ambient fluid. In addition, velocity will also be proportional to the bottom slope, and inversely proportional to the sum of basal (f_0) and interfacial (f_i) friction factors. Through a series of experiments, Middleton (1966) observed that the equation for the velocity of the body of a turbidity current with uniform density is:

$$u^2 = \frac{8g'}{f_0 + f_i} d \cdot \sin\alpha \quad (2.8)$$

2.3.9. Velocity profile of a turbulent open channel flow vs. turbidity current

Friction between a viscous fluid and stationary surface causes flow retardation. Accordingly, velocity in the lowermost part of the flow is commonly considered to be zero. Theoretically, this is only true below a certain critical shear stress when liquid molecules in the lower part of the flow are attracted to and remain in contact with solid particles in the stationary surface (Leeder, 1999). Immediately above this retardation layer viscous and inertial forces give rise to shear stresses that in turn form a velocity gradient. In theory, retarding effects die out progressively away from the flow boundary until at some point the velocity gradient is reduced to zero. Above this level, if the flow is sufficiently thick, the flow can be described by inviscid flow theory (potential flow zone) (Simpson, 1987). In

reality, however, there is never complete retardation of the flow at the flow boundary, or a zone away from the boundary that is entirely free from basal frictional effects.

In an open channel turbulent flow, friction at the air-water interface is negligible because air is significantly less dense and viscous than water. As a result, flow velocity will be zero at the base and highest at the air-water interface (Fig. 2.6A). Exactly how the velocity increases vertically will depend on both viscous and turbulent stresses. Eddy strength varies proportionately with velocity and therefore decreases towards the bed. Very close to the bed viscous stresses dominate in a layer called the viscous sublayer. The velocity profile in the viscous sublayer can be determined by integrating Newton's viscous stress equation (2.1):

$$u = \int \tau \frac{1}{\mu} dy = \frac{\tau y}{\mu} \quad (2.9)$$

The viscous sublayer does not extend far into the flow, is only up to a few millimetres thick, and only exists where the bed is hydraulically smooth (sediment < .6 mm diameter) (James, 2002). Experiments show that above the viscous sublayer the velocity gradient decreases logarithmically with height (Leeder, 1999). If it is assumed that the velocity gradient decreases linearly and inversely with height, then

$$\frac{du}{dy} \propto \frac{1}{y} \quad (2.10)$$

or

$$du = k \int \frac{1}{y} dy \quad (2.11)$$

where k is some constant, so that

$$u = k \ln(y) + C \quad (2.12)$$

where C is the constant of integration. This gives the logarithmic relationship required by experimental evidence. The fuller form of this basic equation is called the law of the wall (Prandtl, 1925), written as

$$u = \frac{u_*}{\kappa} \ln\left(\frac{y}{y_0}\right) \quad (2.13)$$

where u_* is shear velocity ($\sqrt{\tau/\rho}$), κ the experimentally derived von Kármán constant (equal to about 0.41; Hughes, 2007), and y_0 is the distance above the boundary at which u equals zero. Of note, is that this equation in reality applies only to the lower 20% of the boundary layer above the viscous sublayer but is nevertheless often applied to the entire flow (Wren et al., 2005).

Unlike in an open channel flow, fluid at the upper boundary of a turbidity current is of similar density and viscosity to the surrounding ambient fluid. As a result, there is a significant retarding effect at the interface between the turbidity current and ambient seawater and greater interfacial instability (as per Richardson number). Accordingly, velocity

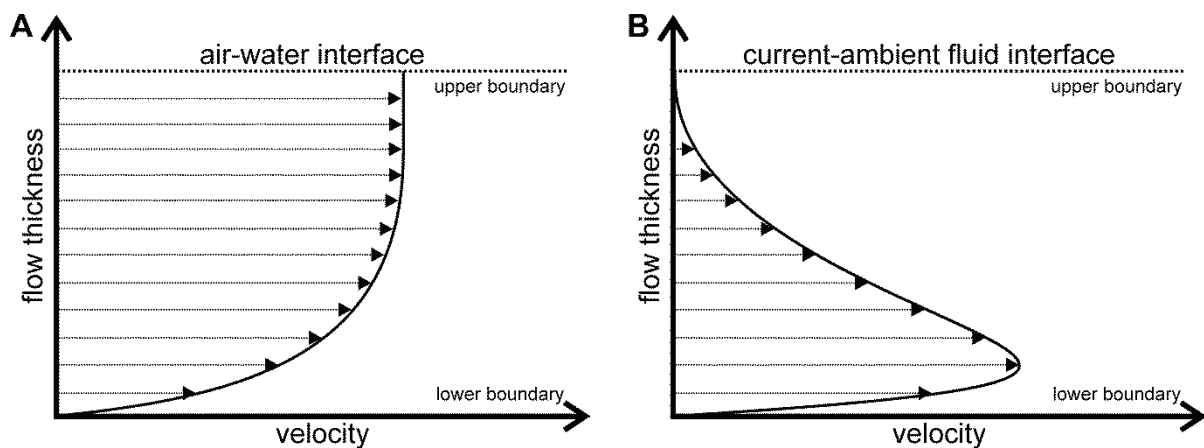


Fig. 2.6. Velocity profiles in an open channel flow (A) and turbidity current (B). Note that in an open channel flow the velocity gradient (shear stress gradient) is steepest near the base whereas near the top of the flow it changes little. In a turbidity current the velocity gradient between the lower boundary and velocity maximum is steeper than the velocity gradient between velocity maximum and upper boundary.

in a turbidity current increases away from the lower and upper boundary surfaces. Shear stress, however, is significantly higher at the base of the flow and consequently the change in the rate of du/dy is appreciably lower away from the upper bounding surface. This causes the high-velocity core to occur closer to the base of the flow (Fig. 2.6B), typically at about 20–30% of the flow thickness (Kneller and Buckee, 2000).

2.3.10. Concentration profile of turbidity current

The sediment concentration profile in a turbidity current is in part controlled by turbulent stress, which can be written as

$$\tau = \rho u_*^2 = \rho A_y \frac{du}{dy} \quad (2.14)$$

where A_y is the turbulent diffusivity (note that $A_y = \eta/\rho$). From the law of the wall

$$\frac{du}{dy} = \frac{u_*}{\kappa} \frac{1}{y} \quad (2.15)$$

so

$$\rho A_y \frac{u_*}{\kappa} \frac{1}{y} = \rho u_*^2 \quad (2.16)$$

or

$$A_y = \kappa u_* y \quad (2.17)$$

Importantly, particles do not always perfectly track patterns of turbulence and this is taken into account in the following expression:

$$K = \beta A_y \quad (2.18)$$

where K is the eddy diffusivity and β a coefficient commonly assumed to be close to one (Kneller and McCaffrey, 1996; Wren et al., 2005). Specifically, β would be one if particles were completely passive and perfectly tracked patterns of turbulence. Particles that settle on

their own in spite of turbulence, however, suggest that β be less than one. Conversely, particles that are given inertia by turbulent stress enhance mixing and suggest that β is greater than one. Empirical observation shows that β , at least in the lower half of flows, is indeed close to one (Bennett et al., 1998).

Ultimately, the concentration gradient in a turbidity current (dc/dy) depends on the relation between particle settling (u_s) and the diffusive flux of particles upwards due to turbulence. This can be expressed by

$$u_s c = -K \frac{dc}{dy} \quad (2.19)$$

or

$$\frac{dc}{c} = -\frac{u_s}{\beta \kappa u_*} \frac{dy}{y} \quad (2.20)$$

This expression is integrated in terms of a relative concentration (i.e. comparison of concentration with that at some reference level)

$$\int_{c_{ref}}^c \frac{dc}{c} = -\frac{u_s}{\beta \kappa u_*} \int_{y_{ref}}^y \frac{dy}{y} \quad (2.21)$$

to give

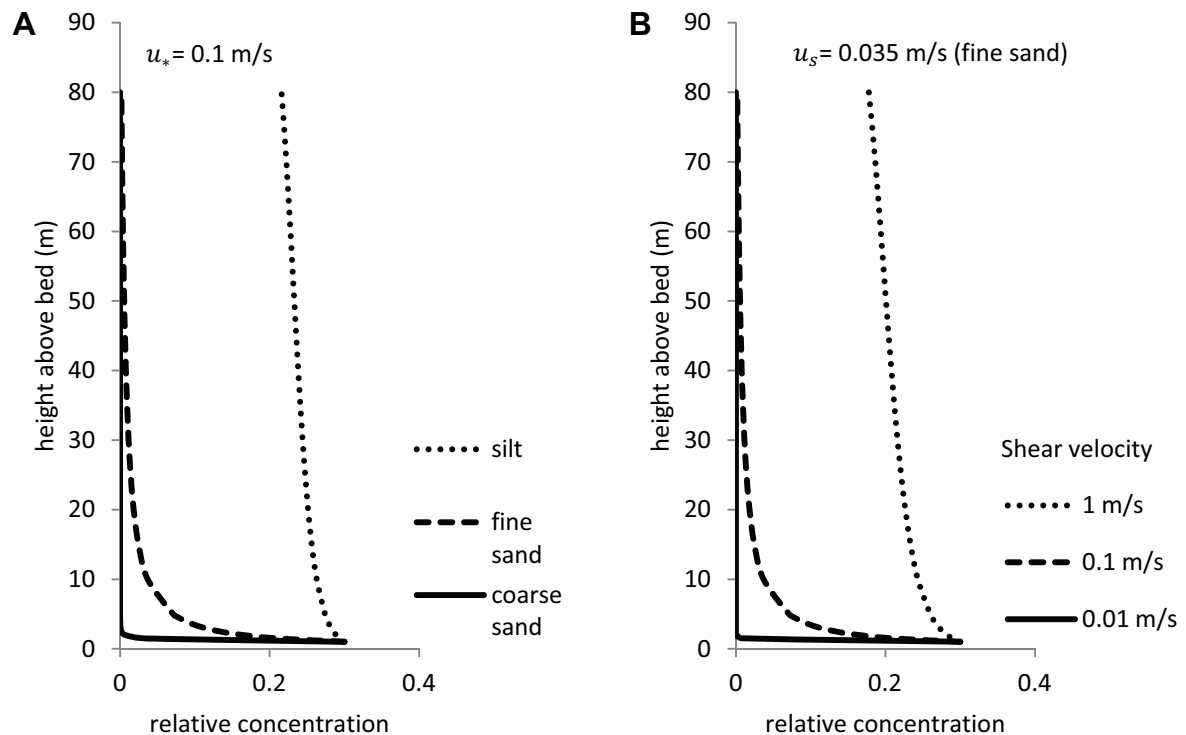
$$\ln \frac{c}{c_{ref}} = -\frac{u_s}{\beta \kappa u_*} \ln \frac{y}{y_{ref}} \quad (2.22)$$

or

$$\frac{c}{c_{ref}} = \left(\frac{y_{ref}}{y} \right)^{\frac{u_s}{\beta \kappa u_*}} \quad (2.23)$$

This expression is the Rouse equation and the exponent is termed the Rouse number or parameter (Middleton and Southard, 1984). In practice, the reference level is established by measuring concentration and size distribution near the base of the current in order to estimate

the profile at all levels in the flow. Settling velocity (u_s) depends on grain size and by extension so too does the Rouse parameter. Specifically, fine-grained particles are distributed more uniformly through a flow compared to coarser-grained particles that tend to concentrate near the bed (Fig. 2.7A). The Rouse parameter is also dependent on shear velocity (u_*) wherein an increase in shear results in a more uniform distribution of particles (Fig. 2.7B).



u_* = Shear velocity
 u_s = Settling velocity

Fig. 2.7. Concentration profile based on the Rouse equation. A) The effect of grain-size on concentration profile. Finer-grained particles are distributed more uniformly than their coarser-grained counterparts. B) The effect of shear velocity (u_*) on concentration profile. Higher shear velocity distributes fine sand particles (settling velocity (u_s) = 0.035 m/s) more uniformly than lower shear velocity.

2.3.11. *Turbidites*

Turbidity currents deposit beds called turbidites, which are ubiquitous in the deep-marine rock record. These deposits range from a few millimetres to several metres thick and consist of mud- to gravel-sized sediment. Bouma (1962) was the first to describe the ideal sequence of sedimentary structures that occur in normally graded turbidites, now called Bouma sequences. This sequence includes five intervals that from base to top are termed the T_a – T_e divisions (Fig. 2.8). The T_a division consists of massive to graded, structureless, coarse to medium sand, commonly with flute and tool marks at its base. These beds are commonly interpreted to form in the upper plane flow regime (Walker, 1965); however, high sedimentation rates are thought to prevent planar lamination from developing (Walton, 1967). Arnott and Hand (1989) confirmed this assumption experimentally by showing that high suspension fallout suppresses the development of planar lamination.

The T_b division is similarly deposited in the upper flow regime, but where lower sedimentation concentrations and rates promote the development of well defined planar lamination with primary current lineations. Grain size in this interval typically ranges from medium- to fine-grained sand. The T_c interval is deposited in the lower flow regime and is made up of small-scale cross-laminated fine to very fine sand. Climbing ripple cross-stratification forms when sedimentation rates are high relative to bed load transport rates (Allen, 1971). The T_d interval comprises silt and interlaminated silt-mud couplets that form through both suspension and traction processes (Piper, 1978; Stow, 1979; Stow and Shanmugam, 1980). The T_e interval consists of homogeneous, structureless mud deposited completely from suspension (Kelts and Arthur, 1981).

Lowe (1982) extended the Bouma sequence to include gravel deposited by sandy high-density turbidity currents (Fig. 2.9). The lowest interval (R_2) in Lowe's sequence

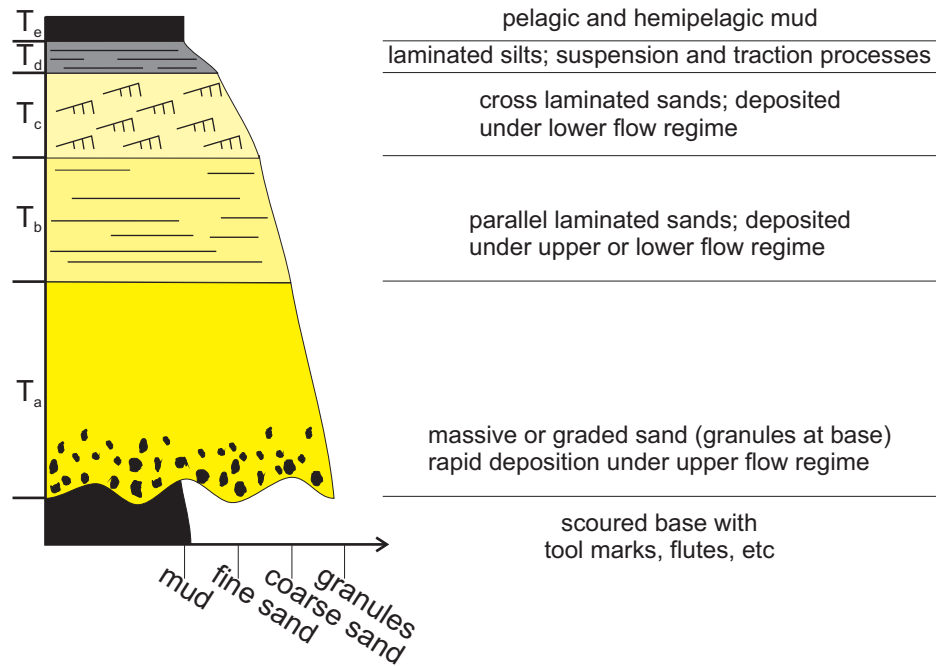


Fig. 2.8 Idealized sequence of sedimentary textures and structures deposited from a turbidity current (after Bouma, 1962).

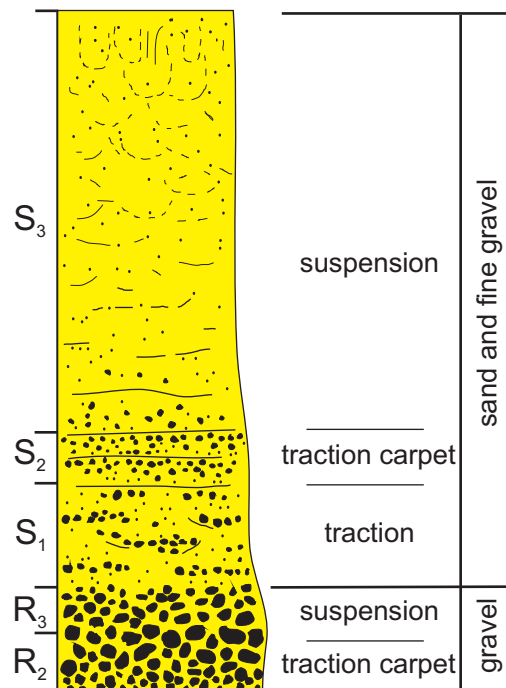


Fig. 2.9 Sequence of sedimentary textures and structures deposited by high-density, coarse-grained turbidity currents. S₃ is equivalent to Bouma T_a interval (redrawn from Lowe, 1982). Used by permission of SEPM (Society for Sedimentary Geology).

involves traction carpet sedimentation of the coarsest gravel. Traction carpet deposition occurs when the concentration of coarse particles near the bed increases to the point where turbulence in the bed load layer becomes suppressed and grain collisions (i.e. dispersive pressure) become the dominant support mechanism (Bagnold, 1956; Dzulynski and Sanders, 1962). R_3 , on the other hand, represent suspension sedimentation of coarse gravel. Progressive deposition causes the current to become progressively finer grained and accordingly, S_1 involves bed load transport of the finer gravel and sand. Subsequent increase in sedimentation rates and near-bed concentration form a second interval involving traction carpet sedimentation comprising finer gravel (S_2). S_3 involves suspension sedimentation of sand which ensues once traction carpet processes terminate. S_3 is equivalent to Bouma's T_a interval.

2.3.12. *Flow steadiness and uniformity*

The concept of sediment continuity (Smith, 1977; Middleton and Southard, 1978) affirms that a flow will either be depositional, erosional, or non-depositional depending on its steadiness and uniformity. Flow steadiness and uniformity are independent and, as a result, a number of different flow conditions arise.

Flow steadiness describes the rate of change in mean flow velocity with time (du/dt) at a specific point in space (Fig. 2.10A). If there is no change in du/dt then the flow is considered to be steady. Alternatively, an unsteady flow will either accelerate (waxing flow) or decelerate (waning flow) with time. Flow uniformity describes how flow velocity evolves in space (du/dx) at a moment in time (Fig. 2.10B). A uniform flow will show the same velocity as measured at two locations simultaneously. A non-uniform flow, in contrast, will either move slower (depletive flow) or faster (accumulative flow) at the downcurrent

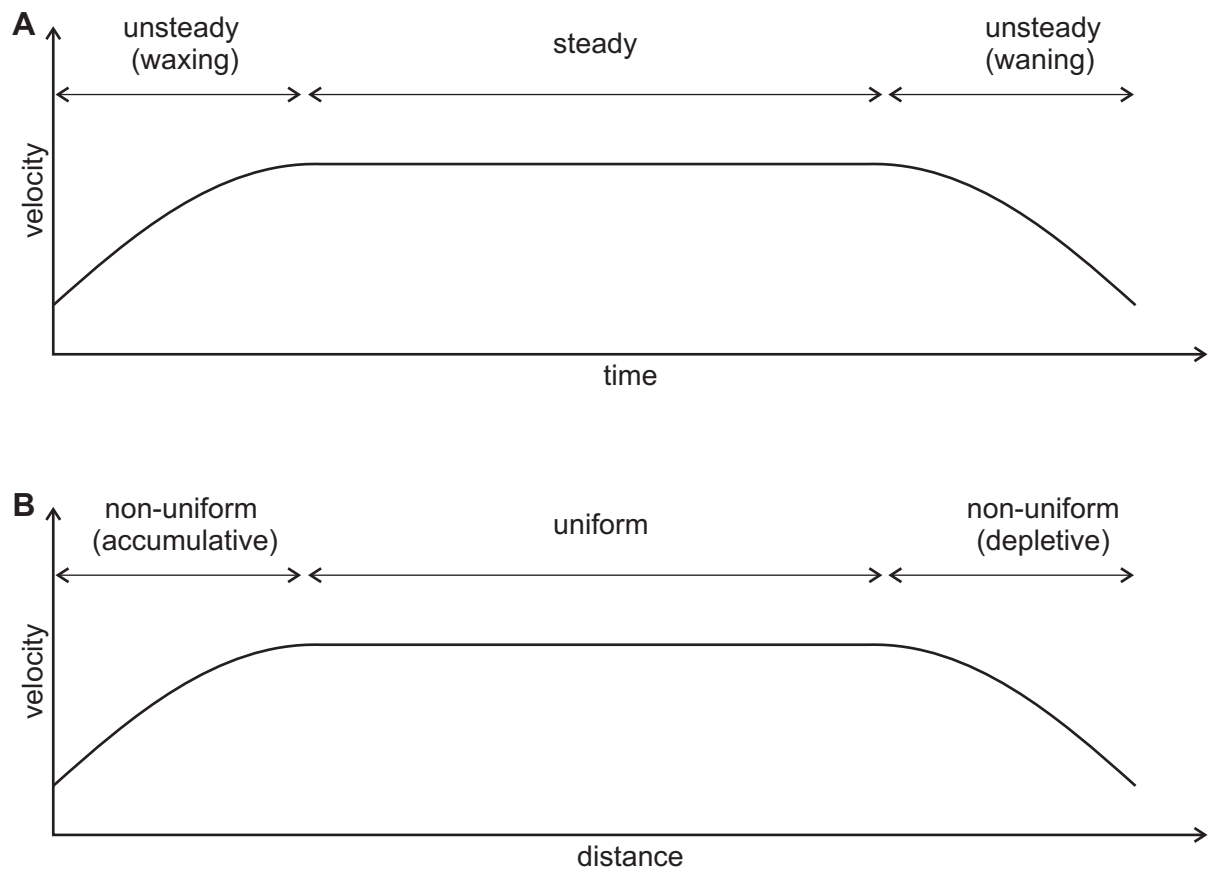


Fig. 2.10. Diagram of how flow varies in time when observed at a fixed point (A) and how flow varies in space when observed in an instant of time (B).

position. A depletive flow occurs when a flow expands areally, like at the mouth of a river. Depletive flows also occur where there is a decrease in slope gradient, like at the base of a slope. Accumulative flows, in contrast, develop where slope gradient increases, like at channel avulsion nodes or where flow is constricted, for example between interconnected mini basins.

Figure 2.11 is Kneller's (1995) conceptual model that predicts the depositional behaviour of a turbidity current based on flow steadiness and uniformity. In this model, waning flows deposit graded beds, although lateral changes will depend on flow uniformity. A waxing depletive flow deposits coarsening upward beds that fine downcurrent, whereas a steady depletive flow deposits beds that show no vertical grain-size change but fine

downcurrent. The sediment continuity equation specifies that all other types of flows result in erosion or non-deposition.

The ubiquity of partial to complete Bouma sequences in the deep-water rock record suggests that turbidites are generally deposited from flows in which there is a fallout of progressively finer-grained sediment combined with a decrease in velocity. Walker (1965) also suggested that turbidites become finer distally with a progressive loss of the lower divisions. This implies that flows responsible for the deposition of an ideal Bouma sequence are depletive and waning.

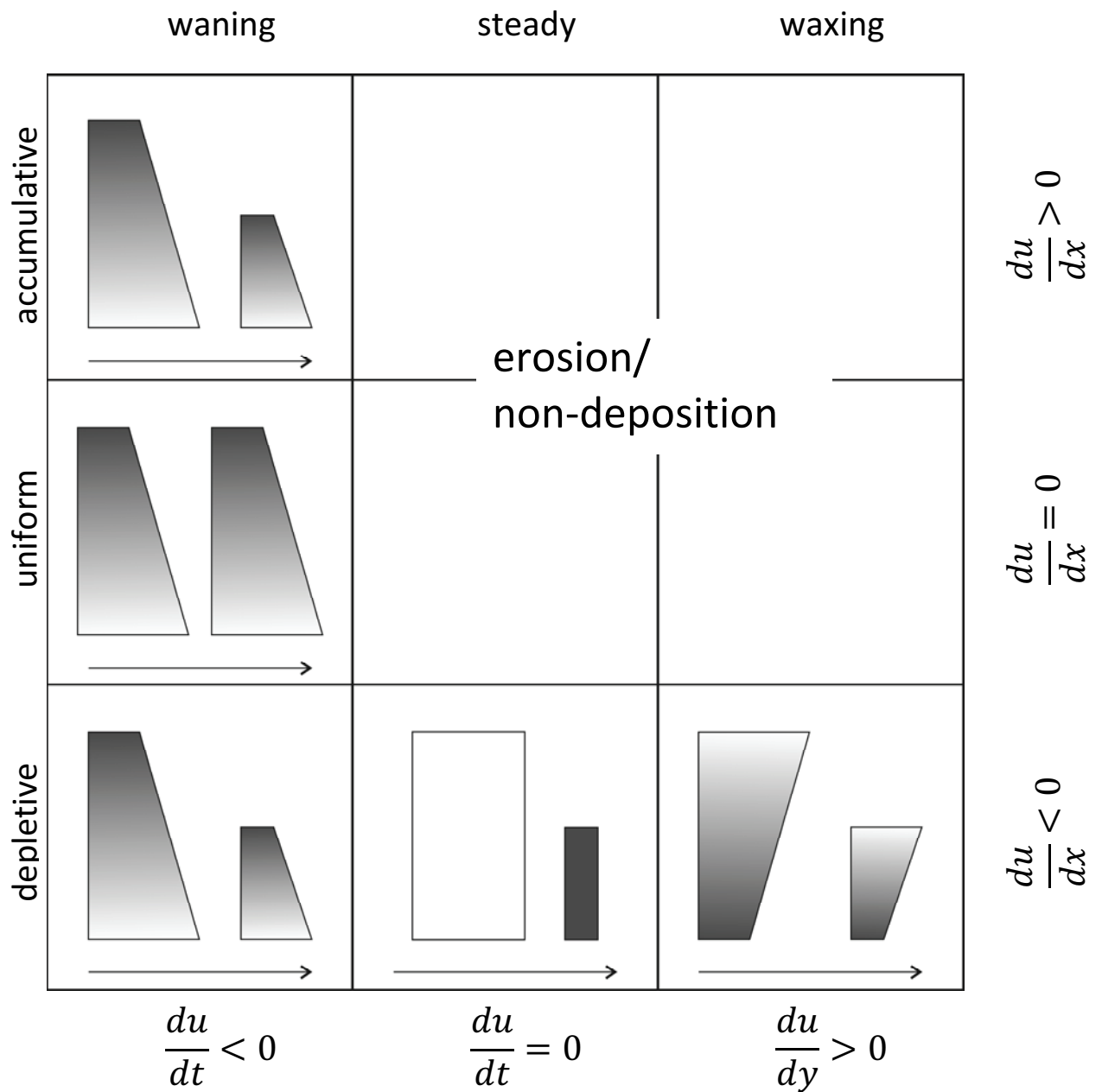


Fig. 2.11. Matrix that predicts whether a turbidity current will be erosional, non-depositional, or depositional based on temporal (du/dt) and spatial (du/dx) changes in flow velocity. Deposition only occurs in fields where velocity decreases with time ($du/dt < 0$) and/or in space ($du/dx < 0$). Schematic lithological logs in the depositional fields show how grain size changes vertically and laterally (redrawn from Kneller, 1995). Used by permission of the Geological Society.

CHAPTER THREE

Submarine fans

3.1. Introduction

Gravity currents transport enormous volumes of sediment from the shelf and slope to the deep ocean floor where it is deposited, forming large-scale depositional systems termed submarine fans (Menard, 1955; Mutti and Normark, 1987, 1991). The study of these deep-water features is of both academic and economic interest, and over the past forty years has resulted in a number of depositional models used by geologists to describe and interpret the nature of sediment transport, depositional processes, sediment distribution, and reservoir architecture associated with these deep-water systems. A major conclusion from these studies is that the controlling factors that influence the deep-marine sedimentary record are numerous and vary in both time and space. A consequence of this is that a single unifying classification scheme for submarine fans is all but impossible. Nevertheless, general fan models are valuable in that they provide useful guidelines on major trends and general characteristics of deep-water depositional systems. This chapter briefly discusses the controls on sediment transport to deep basins, the history of fan models, and provides an overview of mud-rich fan systems (cf. Reading and Richards, 1994). Levees and other overbank deposits are major elements within these systems and are of particular relevance to this thesis. Accordingly, the characteristics and formative processes associated with these elements are emphasised in this chapter.

3.2. Controls on sediment transport

The dominant variables that control the delivery of sediment to the deep ocean are sediment supply, tectonic regime, and sea-level (Reading, 1996) (Fig. 3.1). Sediment supply is a first-order control and influences the calibre and composition of the sediment, its volume and flux, and the number and location of input points. Sediment supply, in turn, is controlled by regional and local tectonics. For example, the tectonic regime affects orogenic processes, which then influence the gross sediment budget, climate, fluvial gradient, and shape and size of the coastal plain. Tectonics also influences the shape, size, and slope of the basin (i.e. shelf, continental slope, continental rise, and abyssal plain), which in turn profoundly influence the gross morphology of the deep-marine depositional system.

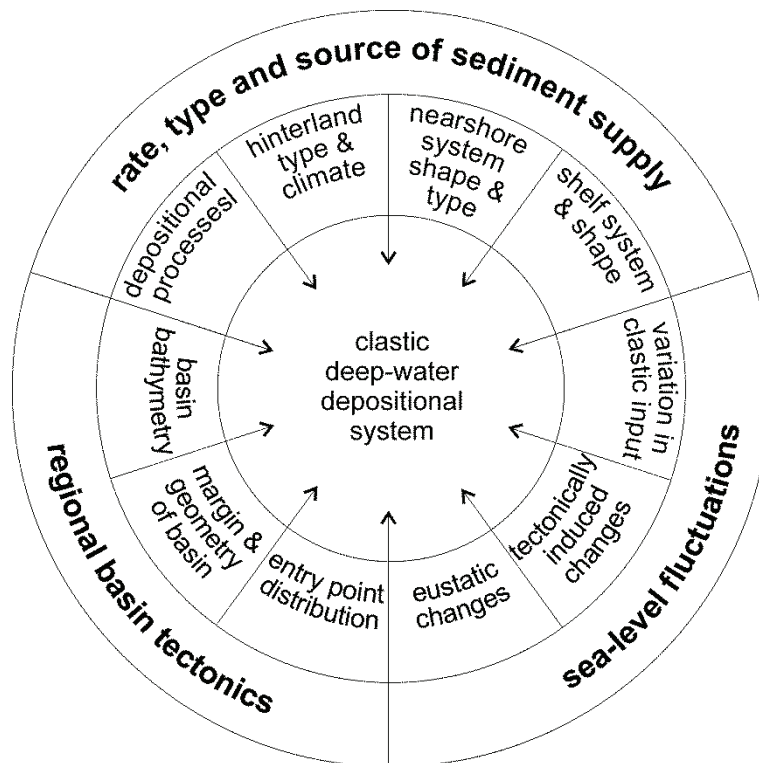


Fig. 3.1. Schematic showing the factors that control deep-water deposition (after Stow et al., 1996).

Fluvial sediment supply to the shelf edge, and by extension the slope and deep ocean, is controlled by sea-level fluctuations. As sea-level begins to fall, shelf delta systems are forced basinward towards the shelf-slope break. As discussed in Chapter 2, when these deltas reach the shelf edge, high pore pressure associated with the rapid accumulation of loosely packed delta sands promotes failure (Prior and Coleman, 1984; Cohen and McClay, 1996; Stow et al., 1996). Failure of shelf-edge deltas during falling sea-level triggers some of the largest and most efficient gravity currents (Reading, 1996). In contrast, fluvial sediment sources tend to move landward and hence farther away from the shelf edge during periods of rising sea level. As a result, delivery of fluvial sediment, especially of sand and gravel, to the shelf-slope break typically becomes reduced and suspension fallout becomes the dominant mode of deposition in the deep basin (Reading, 1996). Although highstands typically represent periods of reduced sedimentation in the basin there are exceptions. For example, active continental margins characterized by narrow shelves and submarine canyons that incise into the shallow littoral zone can sometime supply sandy sediment to deep marine settings (Covault et al., 2007; Covault and Graham, 2010). In addition, systems such as the modern Mississippi River whose rate and volume of sediment input exceeds the increase in accommodation space continue to remain stationary or even prograde during transgressive and/or highstand condition (Uroza and Steel, 2008, make sure this discusses Mississippi).

Although sediment supply, tectonics, and sea-level are the primary allocyclic controls governing deep-water sedimentation there are a number of other local factors that must be considered including basin salinity, organic supply, gas production within the sediment, erosional and depositional features of the sediment, synsedimentary deformation, diapiric movement, and differential compaction (Reading, 1996).

3.3. History of submarine fan models

Submarine fans are commonly divided into an upper, middle, and lower fan region based on their proximity to the shelf break (Fig. 3.2). The first submarine fan model was proposed by Normark (1970; see fig. 22) based on data from the sand-rich La Jolla and San Lucas fans off the coast of California. This model consisted of a leveed upper fan channel that bifurcated downflow into multiple distributary channels on the middle fan, and then into areally extensive unconfined sand-rich deposits (i.e. lobes) on the lower fan. A similar model was proposed by Mutti and Ghibaudo (1972) based on outcrop data from Spain and Italy. The many similarities between these two models suggested that the formative processes of ancient and modern sand-rich fans were similar. Later, studies of the Amazon (Damuth and Kumar, 1975) and Mississippi (Moore et al., 1978) fans established the first models

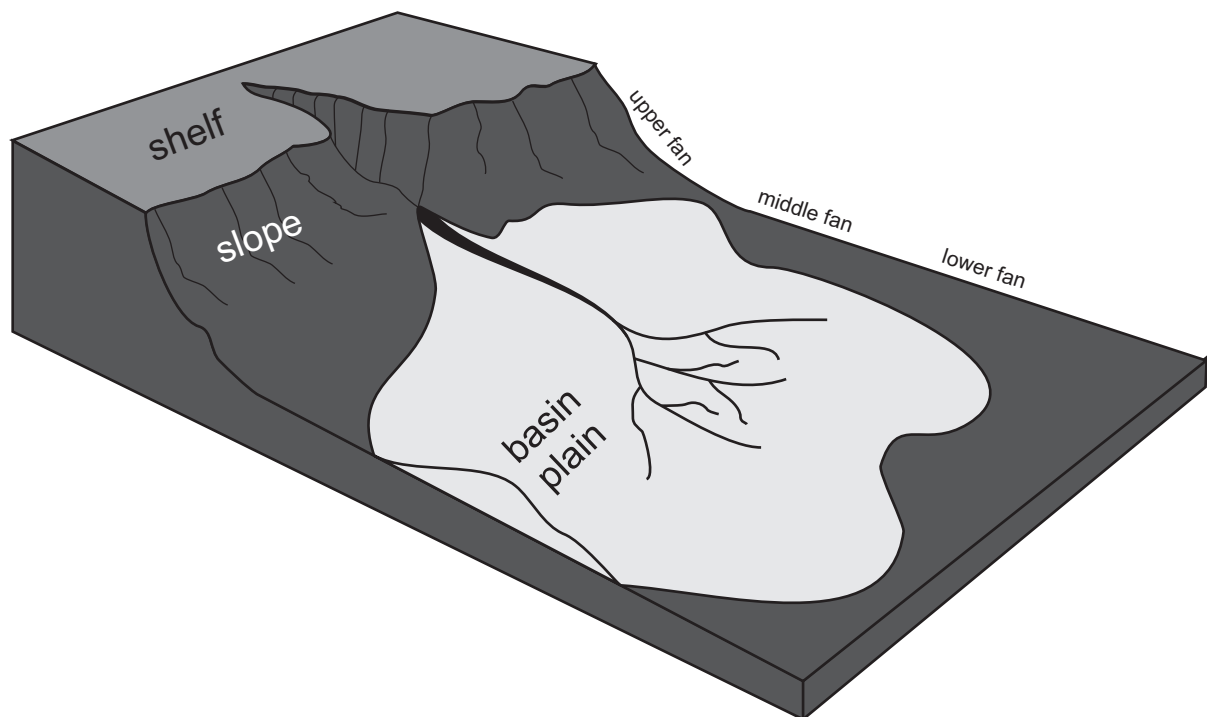


Fig. 3.2. Diagram illustrating the partitioning of a submarine fan into upper, middle, and lower units.

describing fine-grained, mud-rich, large-scale deep-water depositional systems. Walker (1978) then developed a model that merged data from both ancient and modern turbidite systems.

Although these early models remained popular throughout the seventies they did not adequately address the possible influences of sea-level fluctuation, tectonic setting, or grain size on fan development or architecture. As a result, a number of updated models were proposed to better categorize deep-water depositional systems. For example, Mutti (1985) introduced a tripartite depositional model influenced by the volume of the input gravity flows. A model based predominantly on tectonic setting was first introduced by Shanmugam and Moiola (1988) who classified fan systems into four categories: 1) immature passive-margin fans, 2) mature passive-margin fans, 3) active margin fans, and 4) mixed-type fans. Mutti and Normark (1987, 1991) introduced a hierarchical classification consisting of five orders, wherein higher orders represent smaller architectural elements. More specifically, a fan complex is considered a 1st order element whereas a single channel infill, levee, or lobe are 5th order elements. Ghosh and Lowe (1993) also introduced a hierarchical classification; however, in their classification higher orders represent increasingly larger deep-water elements starting with individual Bouma divisions (1st order) through to fan complexes (6th order).

All of the above models have merit but it was Reading and Richards (1994), subsequently updated by Stow and Mayall (2000), who provided arguably the most comprehensive classification scheme for submarine fans (Fig 3.3). This model separates deep-water systems based on whether the system is fed by a single source (submarine fans), multiple sources (ramps), or an essentially continuous linear source (slope aprons). Each of

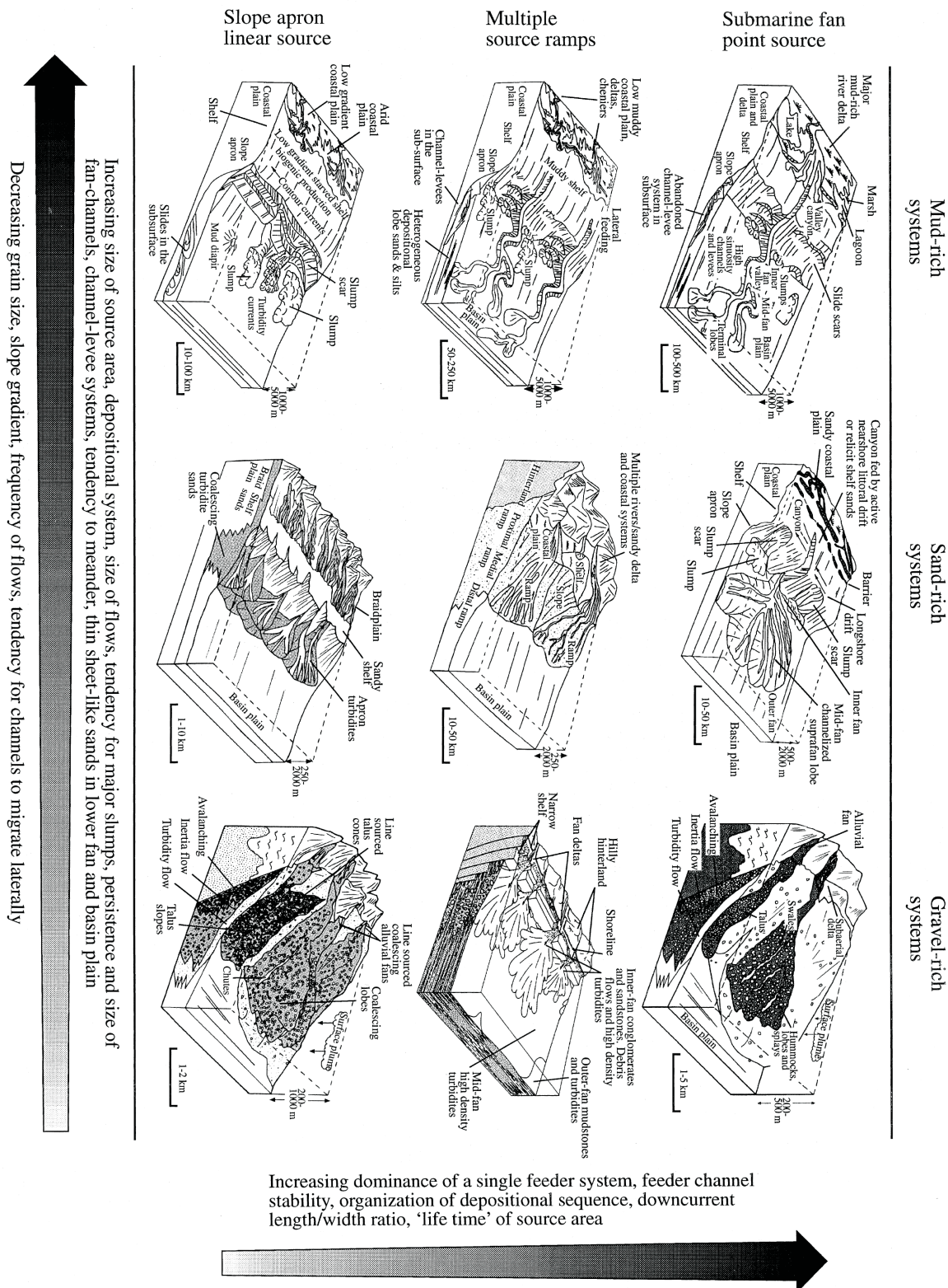


Fig. 3.3. Classification of deep-water systems based on grain size and nature of sediment supply: single point (submarine fan), multiple sources (ramp), or linear source (slope apron) (from Stow and Mayall, 2000; after Reading and Richards, 1994). Reprinted by permission of Elsevier.

these three classifications is further divided into mud-rich, sand-rich, or gravel-rich types based on the dominant grain size in the system.

3.4. Mud-rich submarine fans

The idealized mud-rich submarine fans described by Reading and Richards (1994) closely resemble depositional elements of the Windermere Supergroup deep-water system in terms of tectonic setting, area, and elements (Ross and Arnott, 2007). Modern mud-rich submarine fan systems typically form on passive margins (Reading, 1996) and range in size from the large 3 000 000 km² Bengal fan (Curry et al., 2003) to the much smaller 35 000 km² Astoria fan (Nelson, 1984). In mud-rich systems, as with sand-rich systems, sediment is typically fed to the shelf by major rivers (e.g. Stow et al., 1996). A dominant feature of submarine mud-rich systems is submarine canyons, which are large-scale erosional features that incise the shelf and slope, and represent the main upflow conduit for the transport of sediment to the base-of-slope and then basin floor (Fig. 3.4). Canyons are tens of kilometres to over a thousand kilometres long and range in width from tens to hundreds of kilometres (Shepard and Dill, 1966; Shepard, 1973; Carlson and Karl, 1988). The average relief of canyon walls is 900 m, but in the case of the Great Bahama Canyon the relief is up to 4285 m (Shepard, 1973). The slope of canyon walls are typically less than 10° and rarely more than 30° (Seibold and Hinz, 1974), whereas the average gradient of the canyon floor is ~3° (Shepard, 1973).

Canyons are formed by several processes that are not necessarily mutually exclusive, including 1) erosion at the mouths of rivers that incise the shelf during periods of lowstand (e.g. Spencer, 1903; Bourcart, 1938; Shepard, 1955; Spinelli and Field, 2001); 2) erosion

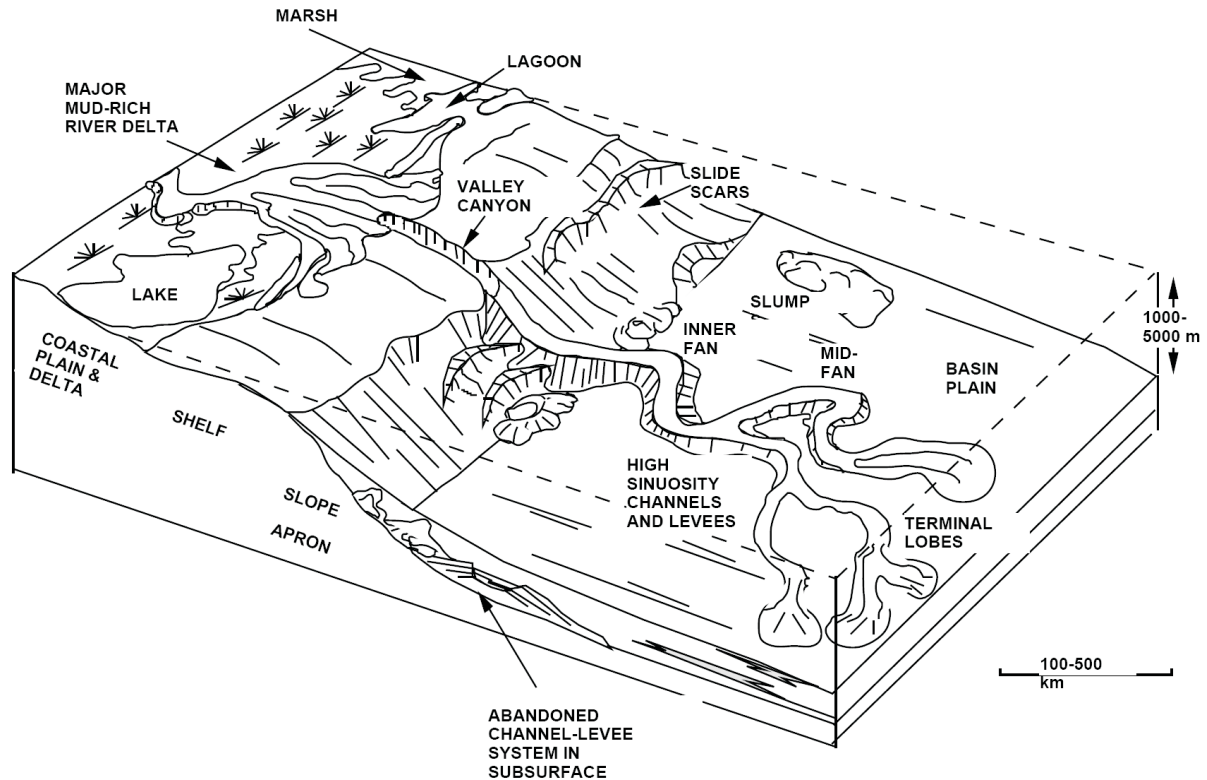


Fig. 3.4. Depositional model for a point-source, mud-rich system showing well developed, sinuous leveed channels in the mid-fan feeding sandy lobes basinward (from Reading and Richards, 1994). ©Copyright 1994. The American Association of Petroleum Geologists. All rights reserved. Reprinted by permission of the AAPG whose permission is required for further use.

from slides, slumps, and debris flows (e.g. Shepard and Dill, 1966; Piper et al., 1999a; Pickering et al., 1989); 3) erosional passage of turbidity currents (e.g. Daly, 1936; Pratson and Coakley, 1996; Mitchell, 2004); 4) glacial or glaciofluvial erosion (e.g. Shepard, 1981); 5) erosion by tidal currents (e.g. Bucher, 1940; Shepard, 1981); 6) biological activity (e.g. Shepard, 1981); and 7) growth faulting (e.g. Yerkes et al., 1967; Kenyon et al., 1978; Berryhill, 1981).

Canyons widen and shallow basinward in response to a reduction in slope gradient. Collectively these conditions promote deposition at the base-of-slope, which is typically characterized by erosionally stacked, wide, thin channel-fill units (Bouma, 2000). The

development of well defined levees is limited by confinement of the canyon walls that inhibits widespread overbank deposition. Away from the base-of-slope and towards the middle fan the system becomes less confined and as a result large channels with well developed levees form. The lower sand-rich part of turbidity currents are confined by the levees and continue to travel downslope but the upper, more dilute part of the current overflows the channel and contributes to the construction of levees (Hiscott et al., 1997). At Castle Creek, the Isaac Formation is characterized by thick channel-fills, extensive overbank deposits, and a low sandstone-to-mudstone ratio (30:70), which collectively suggests a mid-fan setting (Ross and Arnott, 2007).

Towards the lower fan, levees become progressively smaller as levee building material is continually exhausted updip (Posamentier and Kolla, 2003). Accordingly, channels become progressively less confined promoting bifurcation and the development of a distributive channel network (Posamentier and Kolla, 2003). Distributary channels are significantly narrower and shallower than their larger mid-fan counterparts. Downslope distributary channels continue to shallow and eventually become unable to confine the throughgoing channel flows, at which point deposition of large, elongate, unconfined sand lobes ensues. Depositional lobes are commonly tabular and amalgamated in their axis and laterally become progressively more interbedded with thinner-bedded strata. At Castle Creek, the Kaza Group is characterized by shallow, narrow distributary channels, depositional lobes, and a high sandstone-to-mudstone ratio (70:30) and therefore most likely represents a lower fan setting (Ross and Arnott, 2007).

In summary, the turbidite system that deposited the strata exposed at Castle Creek is similar to mud-rich fans because of the interpreted tectonic setting (i.e. passive margin), the estimated areal extent ($>80\,000\text{ km}^2$), and the interpreted basinward change in architectural

elements (i.e. sinuous leveed channels to depositional lobes) (Ross and Arnott, 2007). However, as previously mentioned, general fan models are unable to accurately predict every aspect of all fan systems, and the Windermere deep-water system is no exception. The most obvious difference between the Windermere fan deposits and most mud-rich fan deposits relates to the grain size of the sandstone fraction. In most mud-rich systems, the average grain size in the channels and on the lobes is fine-grained sand and sediment coarser than medium-grained sand occurs rarely (Bouma, 2000). At Castle Creek, on the other hand, pebble conglomerate beds are locally present in the otherwise sandstone-dominated channel fills and lobe deposits typically consist of coarse- and medium-grained sandstone (Navarro et al., 2007a, b; Meyer and Ross, 2007a, b; Ross and Arnott, 2007). Therefore, in regards to grain size, Castle Creek is seemingly more akin to a mud/sand-rich (intermediate between sand- and mud-rich systems; Fig. 3.5) and sand-rich systems (Fig. 3.6). Moreover, the areally extensive basinal sandstone deposits of the Kaza Formation are more akin to basinal deposits developed in sandier systems (Meyer, 2006; Meyer and Ross, 2007a, b). Leveed channels also occur sometimes within mud/sand-rich systems, but the interpreted sinuosity of the channels in the Isaac Formation is more similar to the sinuosity of channels in mud-rich systems (Khan and Arnott, 2011). It is difficult to ascertain whether the Windermere deep-water system was fed by a single or multiple upflow sources. Collectively, these observations illustrate how generalized fan models should only be a guide for interpretation, and that the development of a single unifying depositional model is probably unrealistic.

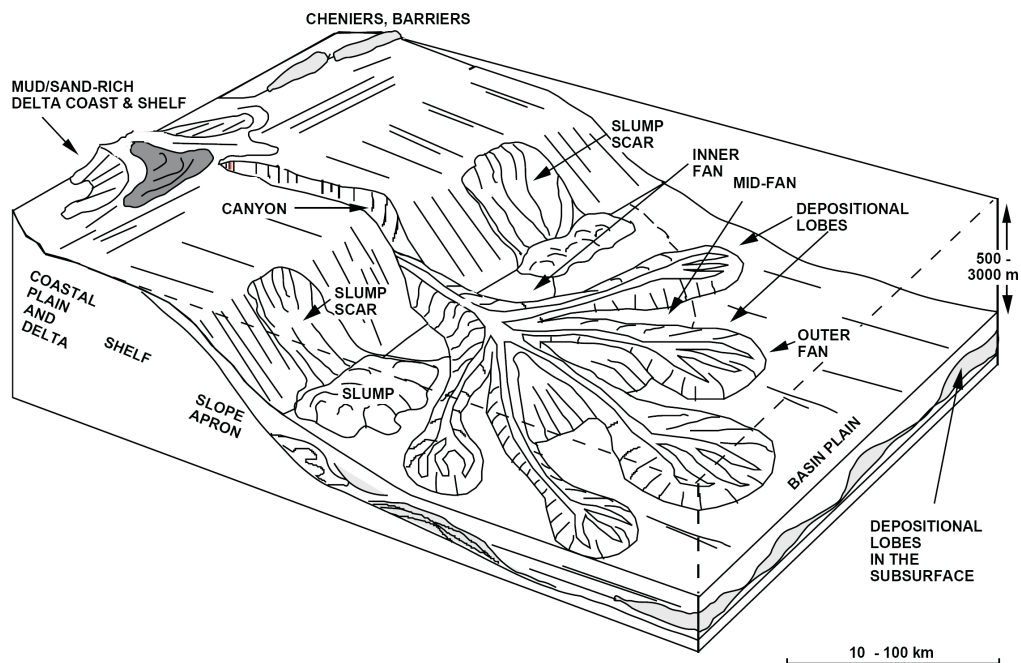


Fig. 3.5. Depositional model for a point source, mud/sand-rich system showing well develop, areally extensive sand lobes and leveed channels (after Reading and Richards, 1994). ©Copyright 1994. The American Association of Petroleum Geologists. All rights reserved. Reprinted by permission of the AAPG whose permission is required for further use.

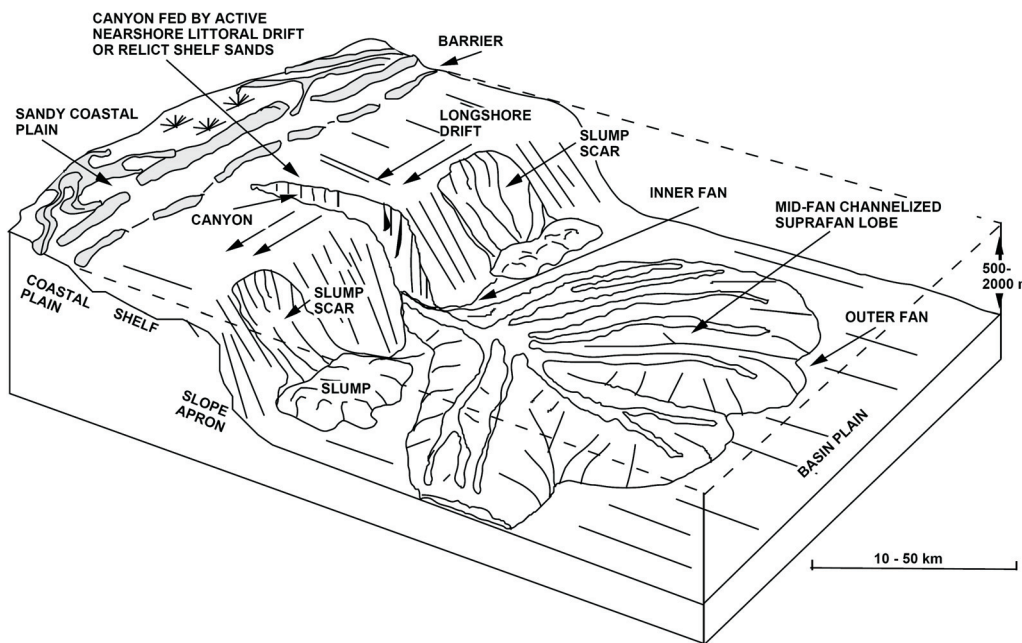


Fig. 3.6. Depositional model for a point source, sand-rich system showing well develop, areally extensive sand lobes (after Reading and Richards, 1994). ©Copyright 1994. The American Association of Petroleum Geologists. All rights reserved. Reprinted by permission of the AAPG whose permission is required for further use.

3.5. Channels

Leveed channels are the main elements observed in mid-fan settings of mud-rich submarine fans, and the processes associated with their development are of particular relevance to this thesis. In this study, a submarine channel is defined as a long-term conduit for the transport of sediment basinward (cf. Mutti and Normark, 1991). In this definition, “long-term” refers to the passage of multiple (several hundred to several thousand) gravity flows. Clark and Pickering (1996) proposed that channel initiation is a progressive process involving sustained periods of erosion related to the passage of many flows. Alternatively, Elliot (2000) suggested that some become initiated by high-magnitude, low-frequency events.

In modern settings, channels have run-out lengths that are up to several thousand kilometres long (e.g. Northwest Atlantic Mid-Ocean Channel; Chough and Hesse, 1976, 1980; Hesse and Chough, 1980). Similarly impressive is the depth and width of many deep-water channels. For example, channels on the Zaire, Rhône, Bengal, and Indus fans are locally 200–325 m deep and 1–20 km wide (Emmel and Curray, 1981a, b, 1984, 1985; Kolla and Coumes, 1984, 1987; Torres et al., 1997; Babonneau et al., 2002; Curray et al., 2003; Wynn et al., 2007). The axial thalweg slope in these systems ranges from 0.05–0.6°.

3.5.1. Longitudinal channel profile

Similar to fluvial systems, deep-water channels decrease in gradient downslope and have a concave up profile (Pirmez et al., 2000; Kneller, 2003). Also, like fluvial channels, submarine channels either erode or aggrade to achieve an equilibrium profile, with base level being the lowest point that a turbidity current can reach within the basin (Pirmez et al.,

2000). When the channel floor exceeds its equilibrium profile erosion ensues and sediment is transported basinward resulting in a flattening of the profile (Fig. 3.7A). Abreu et al. (2003) termed these kinds of channels erosionally confined, and in order to fill them the throughgoing turbidity currents must become less efficient and backfill part of or the entire channel. In contrast, channels that are below their equilibrium profile must aggrade in order to reach an equilibrium profile (Fig. 3.7B). These channels create their own confinement through the deposition of channel-bounding levees, which typically aggrade faster than or at least equally as fast as the channel floor aggrades. Accordingly, these types of channels are termed depositional channels (cf. Normark, 1970). A channel that is at equilibrium, or similarly “at grade”, will carry its sediment basinward with no erosion or deposition along its length (Fig. 3.7C). Although completely consistent, this condition most probably only describes the passage of “average” flows, because deposition most likely still occurs in smaller than average flows and, conversely, erosion will still occur in larger than average flows. In reality, topographic irregularities along most slopes require that channels be both erosional and aggradational at different locations along their length in order to achieve a smooth equilibrium profile (Fig. 3.7D). The equilibrium profile adjusts on multiple time scales as its controlling variables, like sediment calibre, flow magnitude, lobe deposition and position, and regional and local tectonics, change.

3.6. Levees

Submarine levees typically display a gullwing-shaped geometry in cross section (Fig. 3.8A). These deep-water elements are interpreted to build up through the deposition of throughgoing channel flows that spill out of the channel (Buffington, 1952; Hay, 1987a, b;

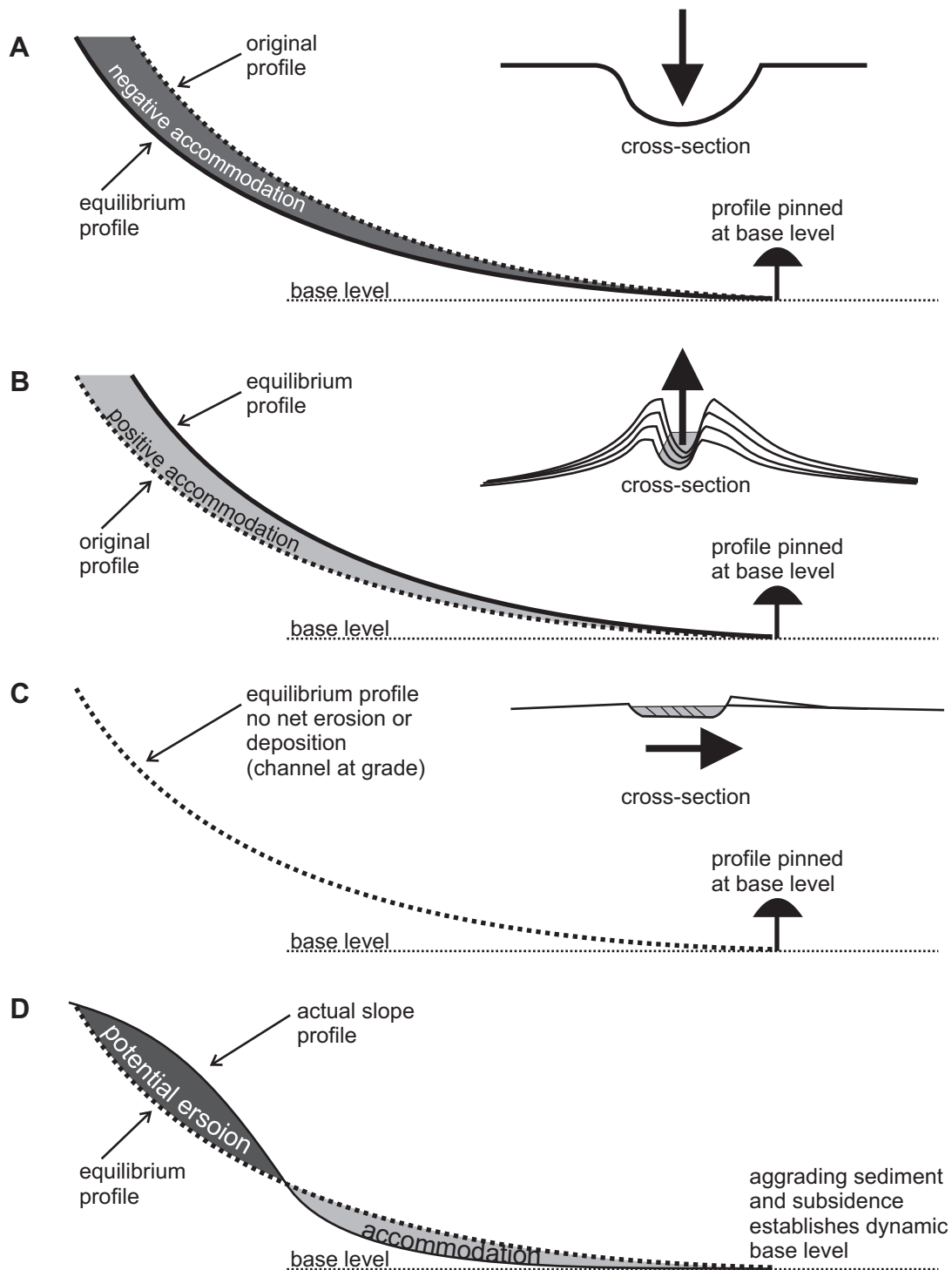


Fig. 3.7. Adjustment of a channel profile to attain transport equilibrium. A) Channels above equilibrium erode to flatten their profile (redrawn from Kneller, 2003); B) channels below equilibrium aggrade to steepen their profile (redrawn from Kneller 2003); C) channels at grade show no net aggradation or erosion but instead shift laterally and bypass sediment basinward (redrawn from Kneller, 2003); D) To achieve equilibrium channels may be both erosional and aggradational at different locations along their length (redrawn from Kneller, 2003; after Samuel et al., 2003). Reprinted by permission of Elsevier.

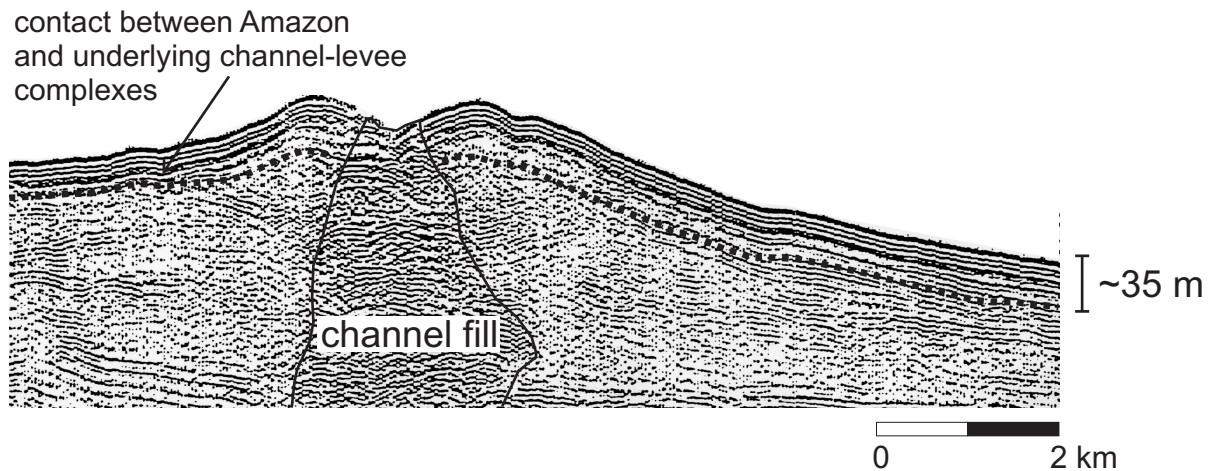


Fig. 3.8. Seismic image of the modern Amazon Channel and underlying channel-levee complexes. Note the well developed gullwing geometry (after Skene, 1998).

Hiscott et al., 1997; Skene et al., 2002). Levees are typically large features, and in some modern systems like on the Amazon, Mississippi, Bengal, and Indus fans locally range from tens of meters to 300 m high and up to 70 km wide (Emmel and Curray, 1981a, b, 1984, 1985; Bouma et al., 1984, 1985; Bouma and Coleman, 1985; Kolla and Coumes, 1984, 1987; Droz and Bellaiche, 1985; Kastens and Shor, 1985, 1986; Pickering et al., 1986; Pirmez et al., 1997; Piper and Normark, 2001; Deptuck et al., 2003; Curray et al., 2003). Levees are divided into proximal and distal regions based on their proximity to the channel axis (Fig. 3.9). In this thesis, the term outer- and inner-bend levee describes levees that occur along the outer- and inner-bend side of a sinuous channel, respectively (Fig. 3.10).

3.6.1. *Overspill*

Overspill refers to situations when flows travelling down a submarine channel overtop both of the associated channel banks (e.g. Clark and Pickering, 1996; Hübscher et al., 1997; Peakall et al., 2000) (Fig. 3.11). Once outside the channel, these flows deposit sediment and form levees. Accordingly, levees are strictly depositional features and,

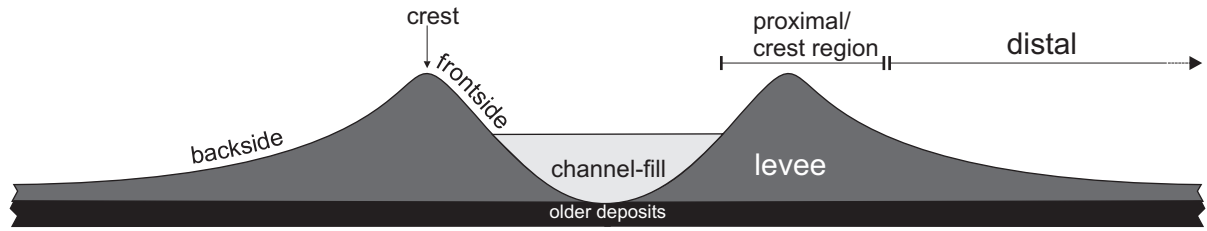


Fig. 3.9. Diagram showing levee terminology used in this thesis.

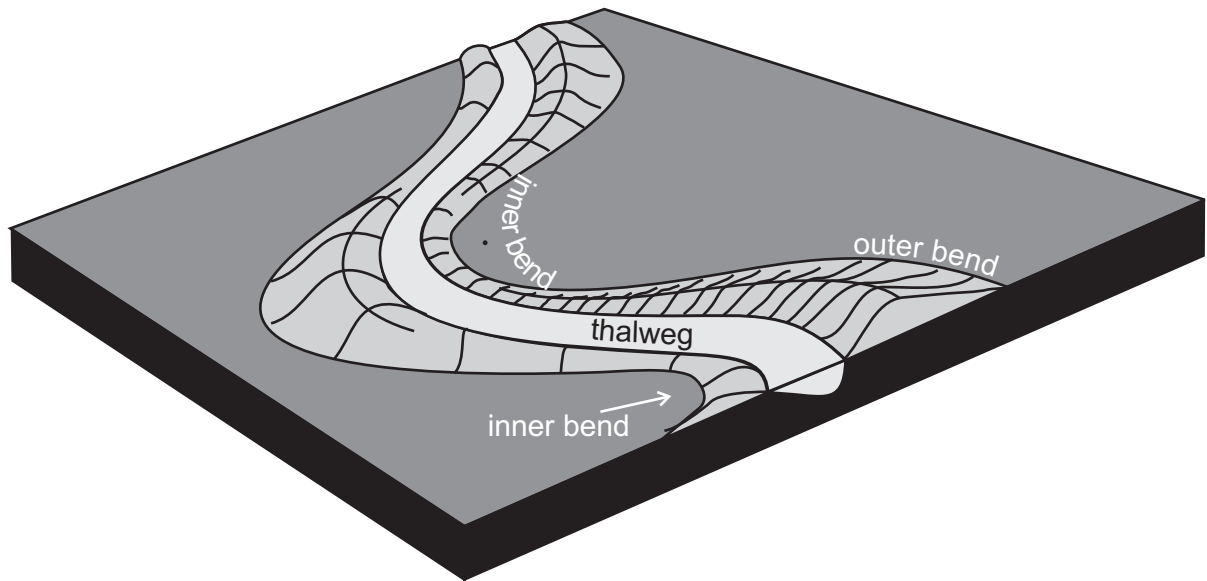


Fig. 3.10. Diagram showing outer- and inner-bend levees.

therefore, channel margins created by erosion are not considered levees. Subaerial levees are an order of magnitude or more smaller in both vertical and lateral dimension than their submarine counterparts (Cazanagli and Smith, 1998; Wynn et al., 2007). This is because overspill flows in subaerial channels are significantly more dense than the ambient fluid (air). As a result, overspilling flows collapse immediately and levee deposition is unable to extend any significant lateral distance away from the channel. In deep-water systems, the significantly lower density contrast between the turbidity flow and the ambient fluid allows the flow to easily thicken vertically and be transported far from the channel margin, forming

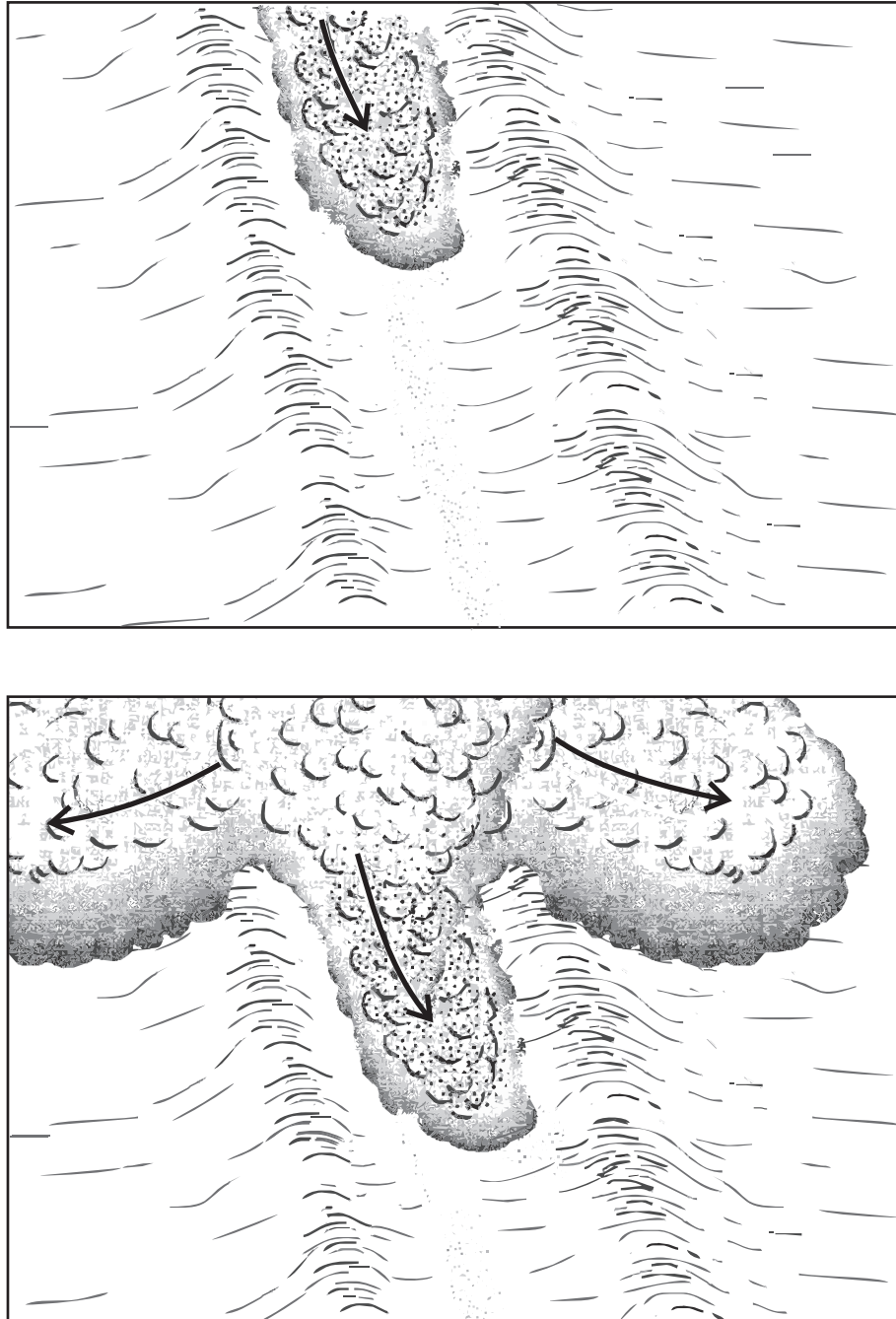


Fig. 3.11. Cartoon showing the process of overspill wherein the upper, dilute part of throughgoing channel flows deposit along both sides of the channel (after Peakall et al, 2000; Kane, 2007).

levees up to several hundred meters high and several tens of kilometres wide (Wynn et al., 2007). Accordingly, in terms of volume, deep-water levees are the most significant element in the mid-fan setting of mud-rich submarine fans (e.g. Prior et al., 1983; Damuth et al., 1988; Hiscott et al., 1997; Pirmez et al., 1997; Deptuck et al., 2003, Beaubouef, 2004).

3.6.2. Flow stripping

Piper and Normark (1983) noted that overbanking occurred preferentially along the outer bend of sinuous channels on the Navy fan. They reasoned that due to inertia the upper more dilute part of throughgoing channel flows became decoupled and overflowed the levee whereas the lower part of the current continued to follow the curving channel path (Fig. 3.12). Piper and Normark (1983) termed this process flow stripping.

Superelevation in channel bends also enhances flow stripping processes by increasing the thickness of the flow along the outer bend (Fig. 3.13). The sloped surface creates a pressure gradient along the outer-bend that balances the radial acceleration that occurs due to centrifugal effects:

$$\frac{g(\rho_c - \rho_a)\Delta h}{w} = \frac{\rho_t \bar{u}^2}{r_c} \quad (3.1)$$

and solving for the required superelevation:

$$\Delta h = w \frac{\rho_c \bar{u}^2}{r_c(\rho_c - \rho_a)g} \quad (3.2)$$

where ρ_c is the density of the turbidity current, ρ_a the density of ambient fluid (i.e. seawater), $\bar{u}$ the mean flow velocity of the turbidity current, h is flow height, w is channel width, and r_c is the radius of curvature. This equation shows that superelevation will be enhanced in wide, highly sinuous channels, and where there is little difference between the

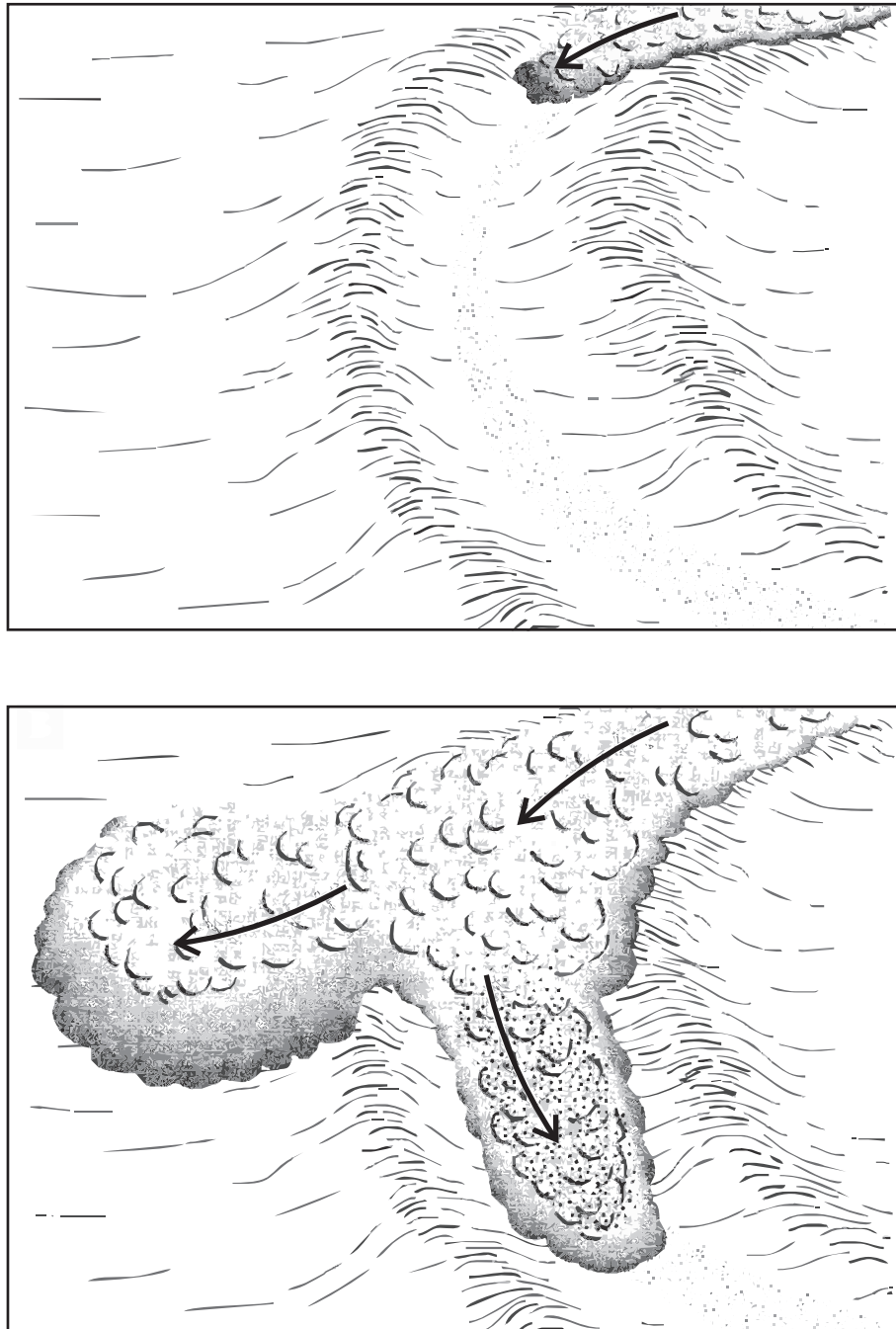


Fig. 3.12. Cartoon illustrating the process of flow stripping, wherein the upper, dilute part of the flow preferentially overflows along the outer-bend of the channel while the coarser-grained, lower part of the flows remains confined within the channel (redrawn from Peakall et al., 2000). Used by permission of SEPM (Society for Sedimentary Geology).

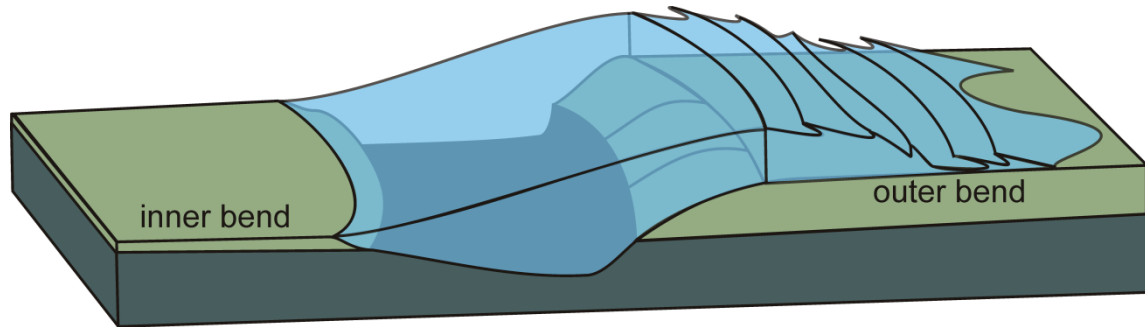


Fig. 3.13. Diagram showing the process of superelevation. As a result of fluid inertia (centrifugal effect) the flow thickens along the outer-bend side of the channel.

density of the current and ambient fluid. Accordingly, superelevation is significant in the deep-water realm and explains why outer bank subaqueous levees are commonly much higher than their inner bank counterparts (e.g. Posamentier, 2003; Posamentier and Kolla, 2003; Deptuck et al., 2003; Khan and Arnott, 2011).

3.6.3. *Inertial overspill*

Hay et al. (1983) remotely monitored channelized subaqueous flows within Rupert Inlet of British Columbia, and recognized a type of overbanking he later termed inertial overspill (Hay, 1987b). This process is similar to flow stripping in that it occurs mostly in sharp channel bends, but unlike flow stripping it describes a condition where the inertia of the throughgoing channel enables the majority of the flow to ride up and overtop the channel bank instead of just the upper, dilute part (Fig. 3.14). Recently, Straub et al. (2008) reported this phenomenon from a series of flume experiments and explained that a unit fluid volume in a current has kinetic energy (i.e. $K.E = \frac{1}{2} \rho_c \bar{u}^2$) that becomes converted to potential energy at bends, which enables particles to ride up and over the outer channel bank. The complete

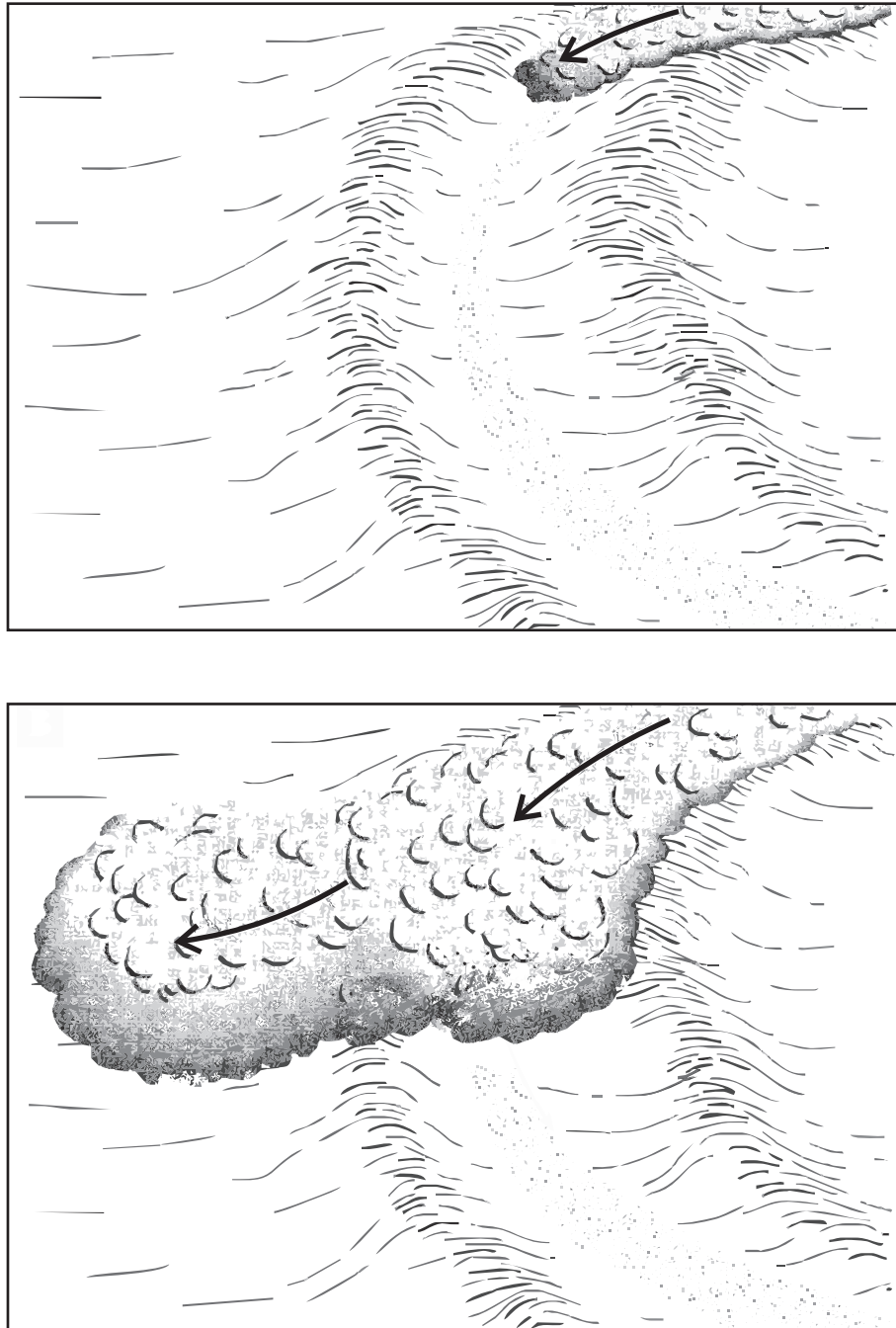


Fig. 3.14. Schematic showing the process of inertial overspill. Along the outer bend fluid inertia enables the flow, including the lower, sandy part, to ride up and over this side of the channel (after Kane, 2007).

conversion of kinetic energy to potential energy will yield a maximum particle run-up (Δh) that is expressed as

$$\frac{1}{2}p_c u^2 = (p_c - p_a)g\Delta h \quad (3.3)$$

solving for run-up height

$$\Delta h = \frac{p_c u^2}{2(p_c - p_a)g} \quad (3.4)$$

This suggests that dense, fast moving flows will experience more run-up than dilute, slow moving currents. Note that Δh in the above equation represents an upper limit because energy losses due to internal and bed friction are not considered.

3.6.4. *Levee deposition and spatial trends*

As channel flows overspill their banks they become less confined and increase in cross-sectional area. As a consequence, velocity of the overspilling flow abruptly decreases and the flow quickly loses transport capacity (i.e. volume of sediment able to remain in suspension) and competence (i.e. largest grain size that can be maintained in suspension). The specific style of deposition that occurs on levees is directly controlled by the amount of levee relief (i.e. channel base to levee crest), in addition to the thickness, velocity, and vertical sediment concentration profile of the throughgoing channel flow. Based on core from the Var turbidite system in the western Mediterranean, Migeon (2000) inferred a vertical profile of the grain-size distribution in turbidity currents wherein the coarsest sediment fraction occurs within the lower 20–30% of the flow and then decreases logarithmically upward. This profile corresponds with the vertical velocity profile measured in experimental turbidity currents, where the velocity maximum, or velocity core, is

commonly reported to occur at about 20–30% of the flow thickness and then rapidly decreases upward (Stacey and Bowen, 1988, Felix, 2002; Kneller and Buckee, 2000). Like velocity and grain size, concentration similarly decreases vertically within a turbidity current (Kneller and Buckee, 2000; Choux et al., 2005) (see Chapter 2). Consequently, poorly confined channels allow much of the flow to overspill, including the lower, sandier concentrated and faster moving basal part. As a result the lower, channel-proximal parts of levees typically consist of the thickest bedded and coarsest-grained deposits in the levee (e.g. Hiscott et al., 1997; Eschard et al., 2003; Beaubouef, 2004; Khan and Arnott, 2011). These thicker-bedded deposits often record complete or near-complete Bouma-type sequences and in some systems are over 100 cm thick with coarse-grained sandstone at their base (see Chapter 4).

In comparison to poorly confined channels, channels with high levee relief confine the sandy, coarse-grained lower part of the throughgoing flows within the channel. As a result, only the upper, more dilute, slow moving part of the throughgoing channel flow is able to overspill resulting in deposition of thinly-bedded, fine-grained T_{c-e} turbidites (e.g. Walker, 1985; Hiscott et al., 1997; Kane et al., 2007, Khan and Arnott, 2011) (see Chapter 4). Walker (1985) reported that thin-bedded levee strata commonly contain climbing ripples, convolute bedding, and mud rip-up clasts, which collectively are interpreted to reflect rapid sedimentation. Strata that display these features are referred to as CCC (Climbing ripples, Convolute bedding, and rip-up Clasts) turbidites and have been reported from many deep-water levee deposits (e.g. King et al., 1994; Lyons, 1994; Coleman, 2000; Cronin et al., 2000; Hickson and Lowe, 2002; Eschard et al., 2003; Grecula et al., 2003; Kane et al., 2007). However, at Castle Creek, CCC turbidites are rarely observed within levee deposits, which may suggest lower rates of sedimentation (see Chapter 6).

The discussion above illustrates how levee deposition depends on the degree of channel confinement, insofar as vertical trends in levees reflect temporal changes in confinement or, more specifically, changes in the ratio between channel flow height and levee height. Where channel flows are in equilibrium (i.e. bypass), episodes of overspill would cause levee relief to increase with time and accordingly strata to progressively fine and thin upward. This process of deposition has been interpreted from many modern and ancient levee deposits, including those in the modern Amazon Fan (Hiscott et al., 1997), the Cretaceous Cerro Torro Formation (Beaubouef, 2004), the Cretaceous Pab Formation (Eschard et al., 2003), and the Permian Karoo Basin (Grecula et al., 2003). The magnitude (height) of flows conveyed through the channel also varies systematically over long periods of time due to changes in relative sea-level, climate, and tectonics (Posamentier, 2003). These allocyclic processes could also potentially gradually reduce flow magnitude, causing similar upward-fining and -thinning levee packages to form. Although such stratal trends are rarely reported to be related solely to allocyclic processes, many authors do suggest that such processes play a role in generating upward-fining trends in levee deposits (e.g. Beaubouef, 2004; Kane et al., 2007). Thickening-upward trends are less commonly reported but have been observed in levees of the ancient Tinker Channel, Turkey (Cronin et al., 2000) and the Rosario Formation, Mexico (Kane et al., 2007). They are typically interpreted to be generated where the channel-floor maintains or outpaces deposition on the levee crest (Kane, 2007). In this scenario, a progressive increase in the magnitude of the overspill flow would result in an increase in the ratio between flow-height and levee-height causing progressively thicker, sandier flows to overspill and deposit on the levee.

In addition to vertical trends, levee strata also display distinctive lateral trends in which most beds thin laterally away the channel, although thick turbidites can initially

thicken and then thin (see Chapter 4). Also, thicker-bedded strata tend to thin more rapidly compared to their thinner-bedded counterparts (DeVries and Linholm, 1994; Beaubouef, 2004; Kane et al., 2007; Khan and Arnott, 2011). Thicker-bedded deposits are more common in the proximal levee and, as a result of lateral thinning; the distal levee consists mostly of thin-bedded turbidites. The lateral thinning of levee strata from the proximal to distal levee is reflected by a decrease in the sand-to-mud or sandstone-to-mudstone ratio and in some systems this decrease occurs rapidly over several hundred meters (e.g. Khan and Arnott, 2011), whereas in other systems this decrease occurs over several kilometres (e.g. Clemenceau et al., 2000). It is generally assumed that the vertical stacking of laterally thinning beds causes an upward increase in stratal dip (Clemenceau, 2000; Browne and Slatt, 2002; Kane et al., 2007; Straub and Mohrig, 2008a) and that these dipping strata generate the multiple dipping reflections commonly observed in seismic images of deep-water levees (e.g. Straub and Mohrig, 2008a; see figs. 15 and 17). This assumption is brought into question by observations made at Castle Creek (see Chapter 5).

3.6.5. Sediment waves

Sediment waves are large-scale, deep-water wavy bedforms with heights that range from a few meters to over a hundred meters and wavelengths of a few hundred meters to several kilometres (Fig. 3.15). Such features have been observed in continental slope, continental rise, and abyssal plain environments (e.g. Fox et al., 1968; Bouma and Treadwell, 1975; Jacobi et al., 1975; Embley and Langseth, 1977; Kolla et al., 1980; McCave and Tucholke, 1986; Manley and Flood, 1993). Sediment waves commonly develop on the flanks of levees, especially along the outer bend of sinuous channels (e.g. Normark et al., 1980; Flood, 1988; Nakajima et al., 1998; Migeon et al., 2000, 2001, 2004; Lewis and Pantin,

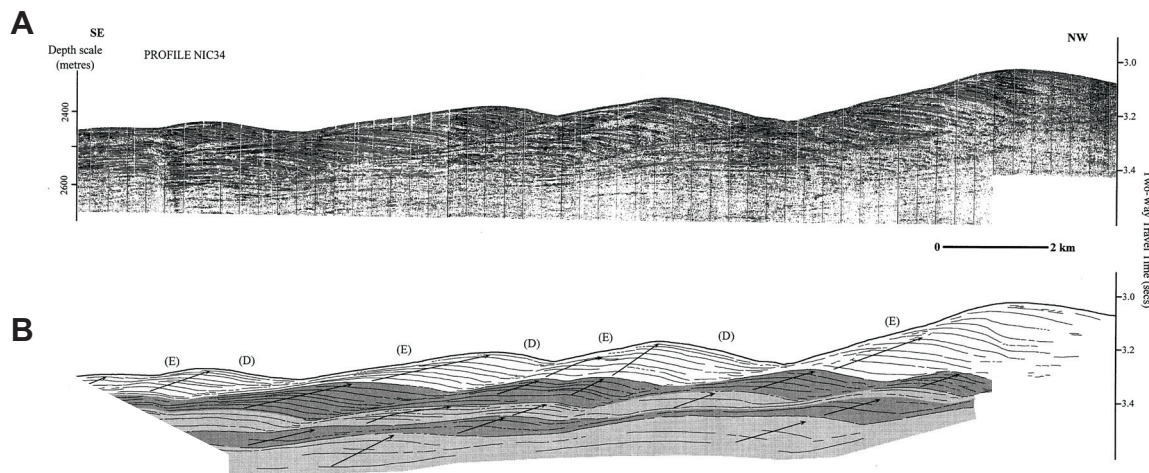


Fig. 3.15. Seismic image of sediment waves developed on the Var Sedimentary Ridge, SE France with interpretation. Note that deposition (D) mainly occurs on upstream flanks whereas erosion (E) is predominant on the downstream flanks. Arrows indicate direction of wave propagation, which is opposite to the general paleoflow (from Migeon et al., 2000). Reprinted by permission of Elsevier.

2002; Normark et al., 2002; Posamentier and Kolla, 2003; Fildani et al., 2006). Moreover, the crests of these features are generally oriented sub-perpendicular to the inferred overspill direction and, accordingly, wave development associated with levees is interpreted to be linked with overspill processes (Normark et al., 1980; Piper and Normark, 1983; McHugh and Ryan, 2000; Migeon, 2000). The exact way in which sediment waves are initiated remains poorly understood. One possibility is that pre-existing topography induced by salt diapirs or faults promotes sediment wave deposition (e.g. Migeon et al., 2000). A second proposed mechanism is that they are deposited by stationary lee waves (e.g. Flood, 1988), or simply internal waves, formed by currents flowing over positive topographical elements, such as the crest of a levee (e.g. Allen, 1982a). Another possible mechanism links them to large erosional structures formed by turbulent vortices within the flow, or the alternation of burst and sweep events at the base of a fully turbulent flow (Migeon et al., 2000; Fildani et al., 2006). Lastly, some sediment waves are interpreted to be initiated by preferential

deposition at the slope break between the proximal and distal levee (Fildani et al., 2006).

Here flow velocity decreases abruptly in response to the sharp change in slope and as a result sediment falls rapidly from suspension (Hiscott, 1994).

Regardless of the exact manner in which waves are initiated, it is generally thought that, once initiated, wave propagation and migration result from variations in the flow dynamics over the irregularity (Flood, 1988; Manley and Flood, 1993; Migeon et al., 2001, 2004; Fildani et al., 2006). More specifically, in the upstream trough the abrupt decrease in slope causes deposition, which in some cases may be accompanied by an internal hydraulic jump (Fig. 3.16). Deposition persists along most of the upstream flank as velocity and bed shear stress are progressively reduced in response to gravity effects caused by the increasing slope. However, before reaching the crest the flow overcomes these impediments and accelerates because of the reduced cross-sectional area of the flow. On the downstream flank,

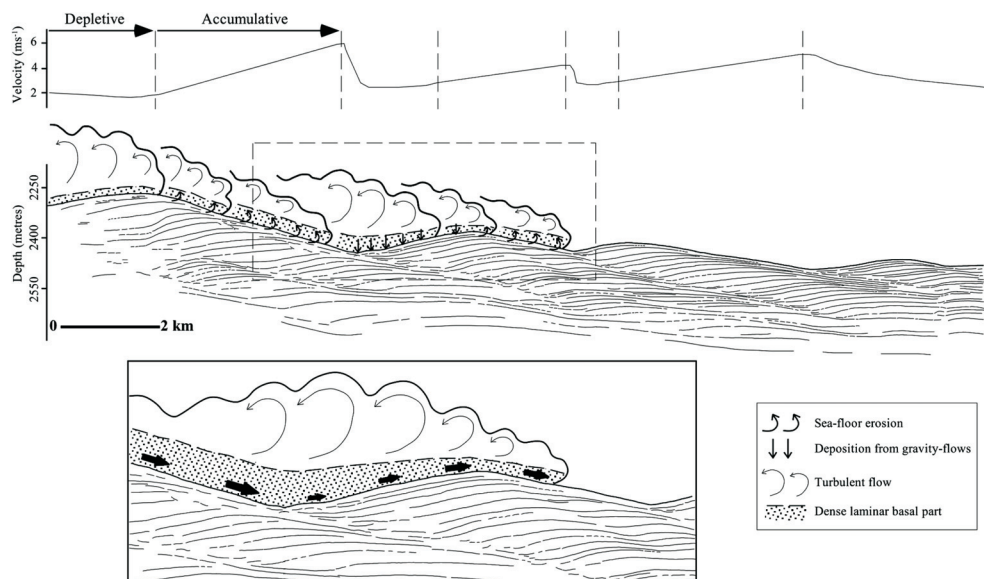


Fig. 3.16. Diagram showing how sediment waves potentially propagate following initiation. Flow velocity increases along the downstream flank from crest to trough inhibiting deposition and causing local erosion. Velocity along the upstream flank, on the other hand, progressively decreases from trough to crest, promoting deposition. Velocity profile and scale is deduced from data on the Var Sedimentary Ridge (from Migeon et al., 2001). Reprinted by permission of Elsevier.

flow velocity is maintained or even accelerates due to the favourable slope. As a result there is significantly less deposition, or even erosion, along the downstream flank of a sediment wave. Preferential deposition on the upstream flank of sediment waves causes upstream migration, which is commonly observed in deep-water settings (e.g. Normark et al., 1980; Manley and Flood, 1993, Migeon et al., 2000, 2001, 2004; Fildani et al., 2006).

3.6.6. Crevasse and overbank splay deposits

Crevasse splays are a type of overbank element that occurs at the terminus of a crevasse channel. The term crevasse channel refers to a conduit immediately downflow of a major levee breach. Crevasse channels typically range from tens to hundreds of meters wide and are hundreds of meters to a few kilometres long (e.g. Kastens and Shor, 1985; Devries and Lindholm, 1994). Crevasses are more likely to form along the outer bend of sinuous channels where flow inertia increases outward directed forces against the levee. The base of a crevasse is substantially lower than the levee crest and is therefore able to tap lower into the throughgoing channel flow. The longitudinal profile through the crevasse is initially decreasing and concave-up and requires erosion at its upstream end to achieve an equilibrium profile (Pirmez et al., 1997). Erosion scours coarse-grained sediment from the channel floor and lower part of the levee causing flows to be significantly sandier, denser, and coarser-grained in their lower part than simple overspill flows.

Crevasse channels commonly extend down the steep backside of the levee and terminate in the distal levee area at a sharp reduction in slope (Pirmez et al., 1997). At the terminus of the crevasse channel, flows expand and abruptly slow in response to the increased cross sectional-area and reduced slope. This causes an increase in the rate of deposition which, in turn, tends to result in the production of lenticular beds and shallow

channels characterized by frequent avulsion (Posamentier and Kolla, 2003). Crevasse splay units cover areas up to several tens of square kilometres and are up to several tens of meters thick (Fig. 3.17). Of note, is that some crevasse channels evolve into independent large scale channel-levee systems. This occurs when the crevasse channel achieves an equilibrium profile and the throughgoing flows begin to overspill and form their own levees. Eventually, the majority of the flows passing through the parent channel are diverted through the “new” channel, which might be more appropriately termed the avulsion channel (Posamentier, 2003; Posamentier and Kolla, 2003).

Overbank splays are similar to crevasse splays but are associated with sandy overspill flows that do not breach the levee. These sandy overspill flows are interpreted to bypass the steep backside of the levee. For this condition to prevail, the slope on the levee backside, and thus the downslope gravitational force, must be sufficient to overcome deceleration associated with flow expansion in the less confined levee setting (Hiscott et al., 1997). Seismic profiles from the Amazon Fan, Indus Fan, and deep-water part of the Niger Delta demonstrate that slopes along the backside of levees are up to 5° , which most likely provide the necessary conditions to induce some local bypass (Pirmez, 1994; Piper and Normark, 2001; Deptuck et al., 2003). Once these flows reach a significant break in slope, commonly in the proximal part of the distal levee region, they begin to rapidly deposit sediment forming a splay unit. Overbank splays are not sourced by a sustained supply of sandy sediment like crevasse splays. Accordingly, the sandy portions of overbank splay deposits are significantly smaller and typically cover areas no more than a few square kilometres and only up to several meters thick (Posamentier and Kolla, 2003; Davis, 2011) (Fig. 3.18).

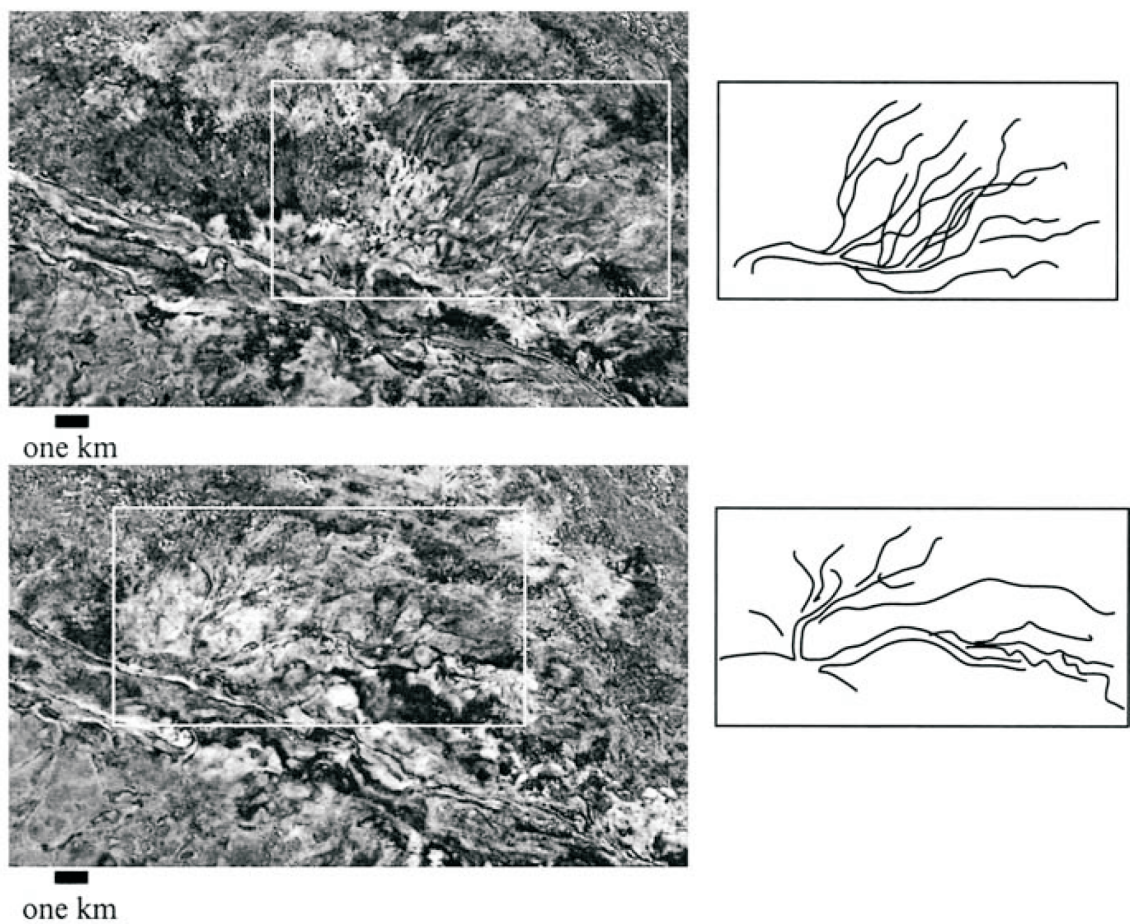


Fig. 3.17. Seismic images of crevasse splays from the Gulf of Mexico (from Posamentier and Kolla, 2003). Used by permission of SEPM (Society for Sedimentary Geology).

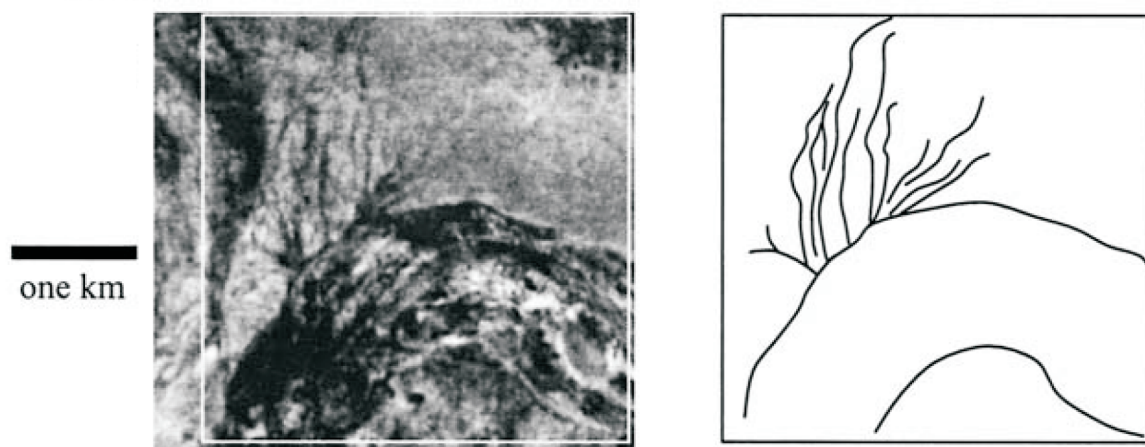


Fig. 3.18. Seismic image of an overbank splay from the Gulf of Mexico (from Posamentier and Kolla, 2003). Used by permission of SEPM (Society for Sedimentary Geology).

CHAPTER FOUR

Stratal attributes and evolution of asymmetric inner- and outer-bend levee deposits associated with an ancient deep-water channel-levee complex within the Isaac Formation, southern Canada

4.1. Introduction

Sinuuous channels are fundamental geomorphological features in both the continental (e.g. Leopold and Wolman, 1960; Walters and Simons, 1984; Brice, 1984; Törnqvist, 1993; Miall, 1996) and deep water realm (e.g. Garrison et al., 1982; Flood and Damuth, 1987; Abreu et al., 2003; Deptuck et al., 2003; Posamentier and Kolla, 2003; Deptuck et al., 2007). Channelization typically results from local erosion of the underlying substrate and/or by deposition of topographic highs along the sides of the conduit. Channel bounding depositional highs are termed levees and form mainly by three kinds of overbanking processes: overspill (Clark and Pickering, 1996; Hiscott et al., 1997), flow stripping (Piper and Normark, 1983), and inertial overspill (Hay, 1987a, b). Overspill results from over-fit flows that deposit simultaneously along both sides of the channel. Flow stripping, in contrast, preferentially takes place along channel bends where the upper more dilute part of flows becomes detached from the main throughgoing channel flow and deposits on the outer channel bank. This process is typically enhanced by the superelevation of flows along the outer bend due to fluidal inertia (Imran et al., 1999; Kolla et al., 2001; Straub et al., 2008). Inertial overspill occurs when the kinetic energy of sediment particles, specifically coarse sediment carried in the lower part of the flow, enables particles to ride up and over the outer

bank where they are then deposited (Straub et al., 2008). Unless otherwise specified, this paper will refer to all of the above processes collectively as overspilling or overbanking.

Overbanking processes are significantly enhanced in the deep-water realm compared to fluvial settings (Imran et al., 1999; Kolla et al., 2001). This is because fluvial flows are constrained at their upper boundary by the water air interface which is marked by an abrupt and significant change in density. Consequently, overbanking occurs mostly during floods that commonly deposit a thin layer of sediment on the floodplain (Cazanacli and Smith, 1996; Brierley et al., 1997). In the deep marine realm, on the other hand, turbidity currents are the main sediment transport agent and their upper boundary is marked by a significantly smaller density contrast. Accordingly, flows are able to easily transfer across the upper boundary because of buoyant support provided by the ambient fluid (i.e. seawater), allowing the flow to expand vertically, commonly exceeding the depth of the associated channel. Due to these conditions, channel bounding levee deposits make up a significant part of the sedimentary record of modern and ancient deep-water slopes (e.g. Prior et al., 1983; Damuth et al., 1988; Pirmez, 1994; Hiscott et al., 1997; Deptuck et al., 2003; Beaubouef, 2004; Navarro et al., 2007b), and have proven to be important hydrocarbon reservoirs (e.g. Tahoe and Ram/Powell fields in northern Gulf of Mexico; Clemenceau et al., 2000; Kendrick, 2000). Nonetheless, it is the volumetrically less significant channels and their associated fills that form the focus of most studies, particularly in the ancient record. A likely reason for the scarcity of levee documentation is that channel bounding strata, once exposed, are typically heavily weathered or overgrown by vegetation, due to their generally fine-grained nature, making it impractical to carry out detailed studies. Moreover, it is even less common to preserve an ancient channel complex that exposes strata on both sides of the channel and therefore there is little understanding of the stratal differences between these counterpart

deposits. Owing to a lack of such detail many channel-levee growth models described in the literature assume symmetry in facies and thickness of levees deposited on opposite sides of the channel (e.g. Eschard et al., 2003; Grecula et al., 2003; Beaubouef, 2004), even though high resolution seismic images of modern systems show that the majority of submarine channels are characterized by highly asymmetric channel-bounding levees, especially along channel bends (Posamentier, 2003; Posamentier and Kolla, 2003; Deptuck et al., 2003). This illustrates the advantage of seismic data over outcrop studies in determining large-scale channel-levee geometries. Nevertheless, seismic resolution still remains too coarse to resolve and differentiate small-scale internal stratal attributes. Therefore, large scale outcrop examples (e.g. Beaubouef, 2003; Eschard et al., 2003; Grecula et al., 2003; Kane et al., 2007; Crane and Lowe, 2008; Hubbard et al., 2008) are vital in narrowing the gap between large scale seismic models and limited but high resolution outcrop studies.

Isaac Channel Complex 3 (ICC3) in the Castle Creek study area is an 80 m-thick, and more than 1.8 km-wide seismic-scale outcrop of a submarine sinuous channel-levee system (Navarro et al., 2007b), in which levee deposits are well exposed along both channel margins and can be traced out continuously for hundreds of meters laterally. Also, the spatial relationships between levee strata and their association with related channel-fill deposits are easily observed (Navarro et al, 2007b). Outcrops like the one described in this manuscript are important to define reservoir architecture and lithofacies associations in different sub-environments of deposition, as well as sub-seismic geometries. The aim of this study, therefore, is to 1) describe the stratal attributes of inner- and outer-bend levees, and then use those characteristics to 2) construct a generalized model describing the stratigraphic evolution, in both space and time, of an ancient sinuous channel-levee system.

4.2. Geological setting and Castle Creek study area

Sedimentary rocks of the Neoproterozoic Windermere Supergroup (WSG) extend the length of western North America from northwestern Mexico to the Yukon-Alaska border – a distance of more than 4000 km (Fig. 4.1A). The deep-water part of the WSG is well exposed in the southern Canadian Cordillera (Fig. 4.1B), which forms part of an orogenic belt related to plate convergence in Mid-Jurassic to Early Cenozoic time (Reid et al., 2002). WSG strata here are interpreted to have been deposited along part of the Laurentian (present day North America) continental margin during break-up of the supercontinent Rodinia ca. 750 Ma (Brookfield, 1993; Ross, 1991; Ross et al., 1995). In the study area, the lowermost part of the WSG (i.e. Toby and Irene formations; Fig. 4.2) is interpreted to record deposition during active rifting and comprises conglomerate, glaciogenic diamictite, and local mafic volcanic rocks that are up to 1 km thick (Stewart, 1972; Ross, 1991). Overlying these rift related rocks, and representing the passive-margin phase of WSG deposition, is a 5–7 km thick pile of predominantly deep-water siliciclastic rocks with subordinate carbonates. Based on regionally extensive marker units (i.e. Old Fort Point Formation and two deep-water carbonate successions), the WSG deep-water marine basin may have been of the order of 80,000 km² or more (Ross and Arnott, 2007).

In the northern Cariboo Mountains of east-central British Columbia strata of the WSG are well exposed at the Castle Creek study area (Fig. 4.1). At this locality WSG strata crop out on the overturned, nearly vertically-dipping limb of a southwest-verging fold. Notwithstanding regional deformation and low-grade metamorphism in the area, effects on the preservation of primary sedimentary structures have been relatively minor, and even mm-scale features are well preserved. Total exposure of WSG in the Castle Creek area is about 7

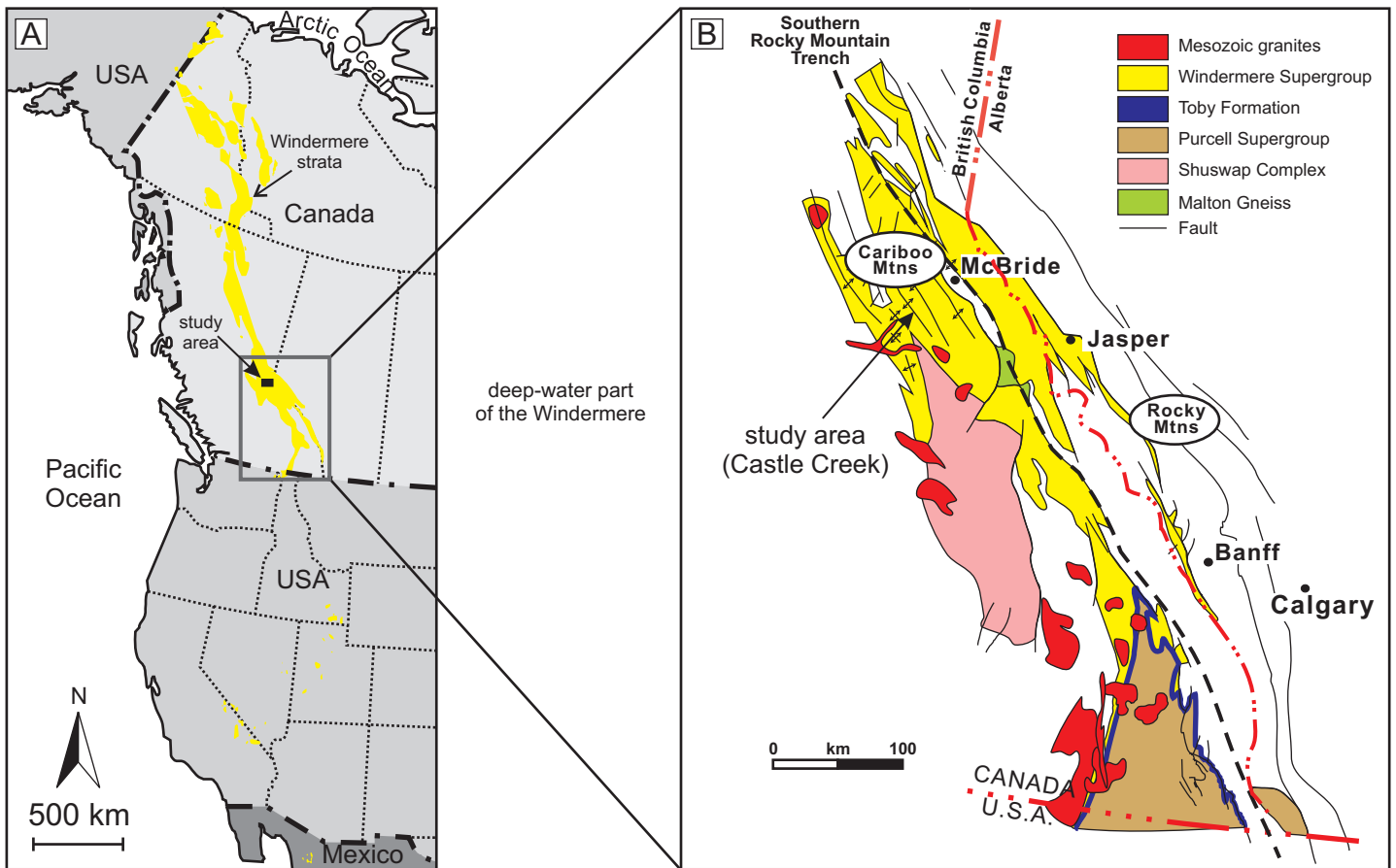


Fig. 4.1. Location of the study area in relation to (A) map of western North America (outcrop distribution from Ross, 1991) and (B) geological map of the deep-water part of the Windermere Supergroup and surrounding units in the southern Canadian Cordillera (modified from Ross and Murphy, 1988).

km parallel to bedding and 2.4 km perpendicular to bedding. Figure 4.3 shows a 2.2 km by 2.4 km portion of the Castle Creek study area.

In Figure 4.3 the lower 800 m of the section comprises the upper Kaza Group, which consists predominantly of massive to normally graded, medium- to very thick-bedded turbidites intercalated with subordinate intervals, up to 40 m thick, of very thin- to thin-bedded, fine-grained turbidites (Meyer and Ross, 2007a, b). Thicker-bedded strata comprise a mixture of rhythmically-stacked, quasi-tabular elements interpreted to represent basin-floor distributary lobes, or similarly, terminal splays (Meyer and Ross, 2007a, b).

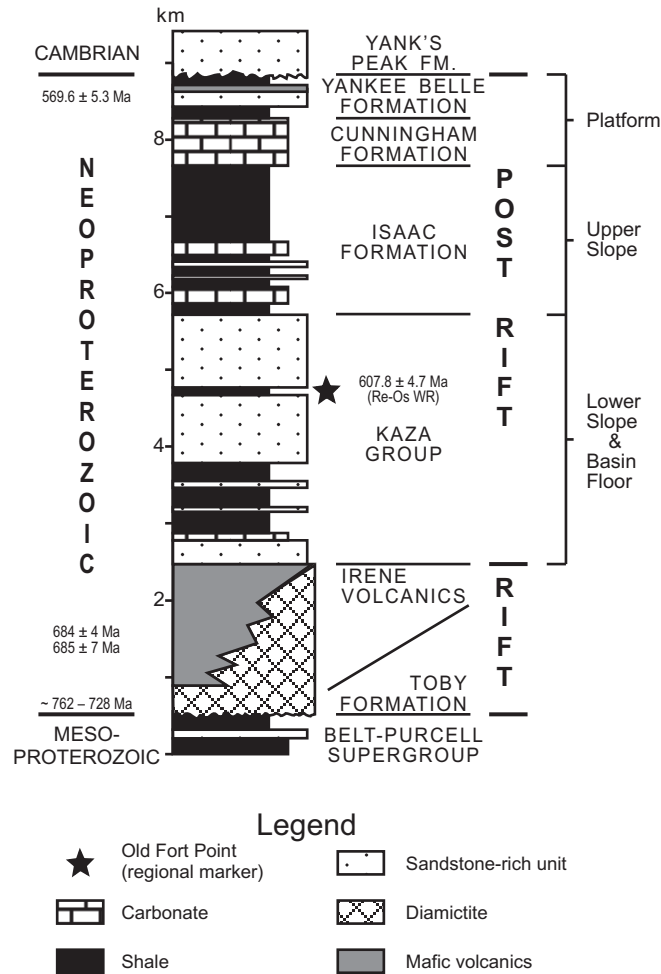


Fig. 4.2. General stratigraphic log of the Windermere Supergroup in the southern Canadian Cordillera (Modified from Ross et al., 1995). Age determinations from Ross et al. (1995), Lund et al. (2003), Kendall et al. (2004), and Colpron et al. (2002).

The Isaac Formation conformably overlies the Kaza Group and has a stratigraphic thickness of approximately 1.6 km in the study area. In contrast to the Kaza Group, coarse sandstone and conglomerate form major, but laterally discontinuous, stratal units up to about 100 m thick. These units exhibit channel geometries (Arnott, 2007a, b; Navarro et al., 2007a, b; Schwarz and Arnott, 2007a, b, c) and are informally termed Isaac Channel Complexes 1–6 (ICC1–ICC6). Channel complexes are encased on all sides by predominantly thin-bedded turbidites interbedded with subordinate yet thicker (up to 5 m) amalgamated sandstone units.

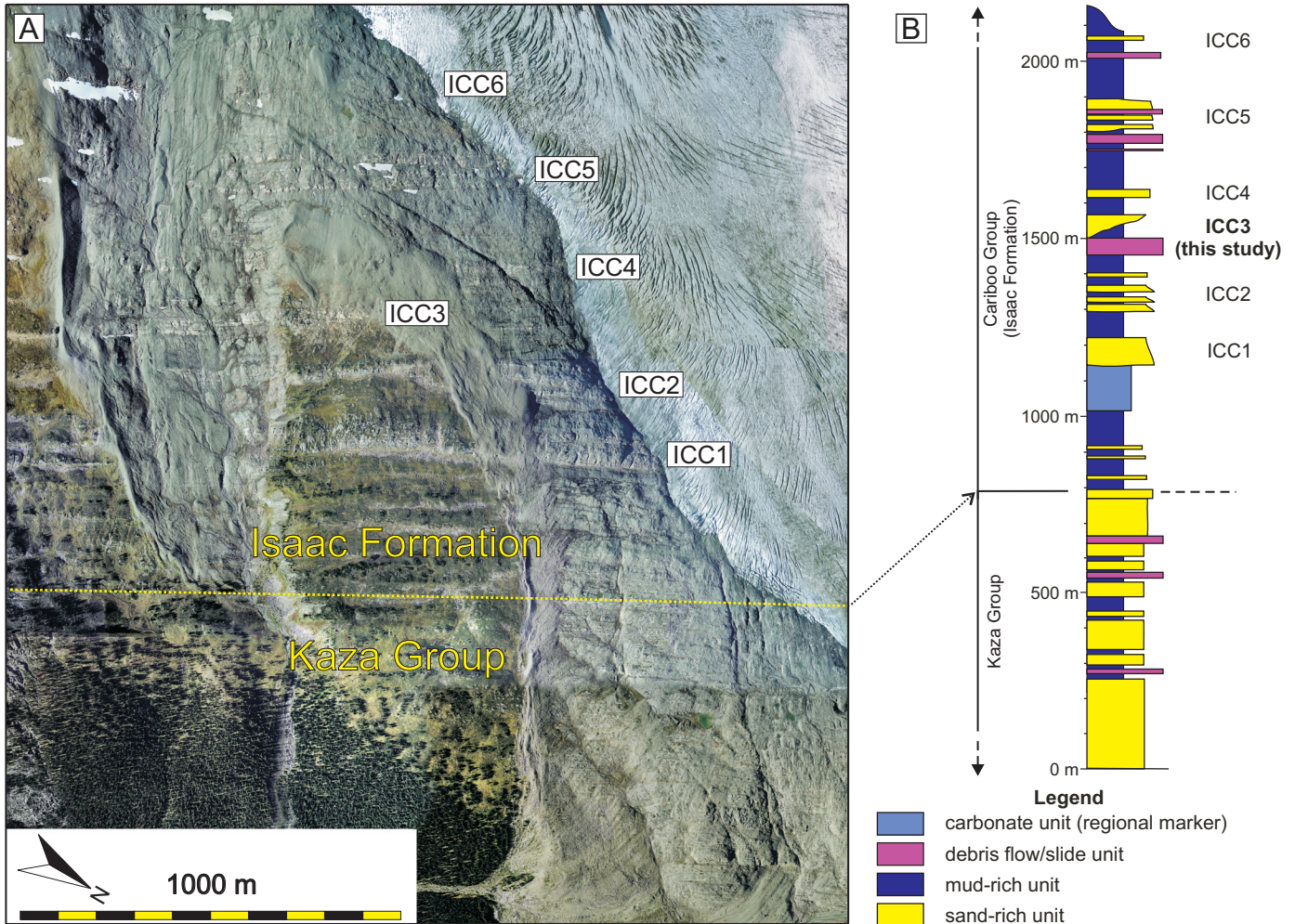


Fig. 4.3. A) Aerial photomosaic of the Castle Creek south study area. Beds strike SE – NW and are nearly vertical. The lower 800 m of stratigraphy comprises the upper member of the Kaza Group whereas the overlying 1.6 km of stratigraphy comprises the Isaac Formation. The location of Isaac Channel Complex Units 1 – 6 is shown (ICC1 – ICC6). B) Generalized stratigraphic log of the Castle Creek study area (modified from Schwarz and Arnott, 2007a).

These strata are collectively interpreted to represent channel-margin levee and other overbank deposits (Ross and Arnott, 2007). Also observed in the Isaac Formation are thick debris flow, slump, and slide units that, together, are considered to reflect gravitational instability common to deep-water slope settings (Arnott et al., 2011).

Overall, strata exposed at Castle Creek record a kilometre-scale upward-shoaling succession from basin-floor (Kaza Group) to base-of-slope (Isaac Formation) interpreted to

be the result of progradation of the Laurentian continental margin in response to diminishing rates of thermally-driven subsidence (Ross, 1991; Ross and Arnott, 2007).

4.3. Isaac Channel Complex 3

Isaac Channel Complex 3 is interpreted to crop out as an almost perpendicular-to-flow cut through the bend of an 80 m thick, more than 1.8 km wide sinuous deep-water channel-levee system (Navarro et al., 2007a, b) (Fig. 4.4). This channel complex is underlain by an up to 60 m thick, mud-rich deposit interpreted to be a debrite (Arnott et al., 2011). In the southernmost part of the study area, however, the channel-levee and debrite units are separated by a thin poorly exposed coarse-grained (up to pebble conglomerate), concave-upward unit interpreted to be an early stage channel-fill. A younger channel-levee system that is up to 50 m thick overlies ICC3 (Navarro, 2006). The channel-fill parts of ICC3 and this younger channel are separated by a unit, up to 50 m thick, composed of thin-bedded turbidites, however immediately adjacent to the moraine both channel units are in direct contact with one another. On the opposite side of the moraine the top of ICC3 is marked by the base of a slide complex. In this paper only the middle channel-levee complex (ICC3) is discussed.

4.4. Lithofacies of channel-fill strata in ICC3

The channel-fill portion of ICC3 is about 80 m thick and comprises four individual channel-fill units (C1–C4), which individually range from 7 to 50 m thick (Navarro et al., 2007b). The lower two channel-fill units in outcrop are at least 1.5 km wide whereas the

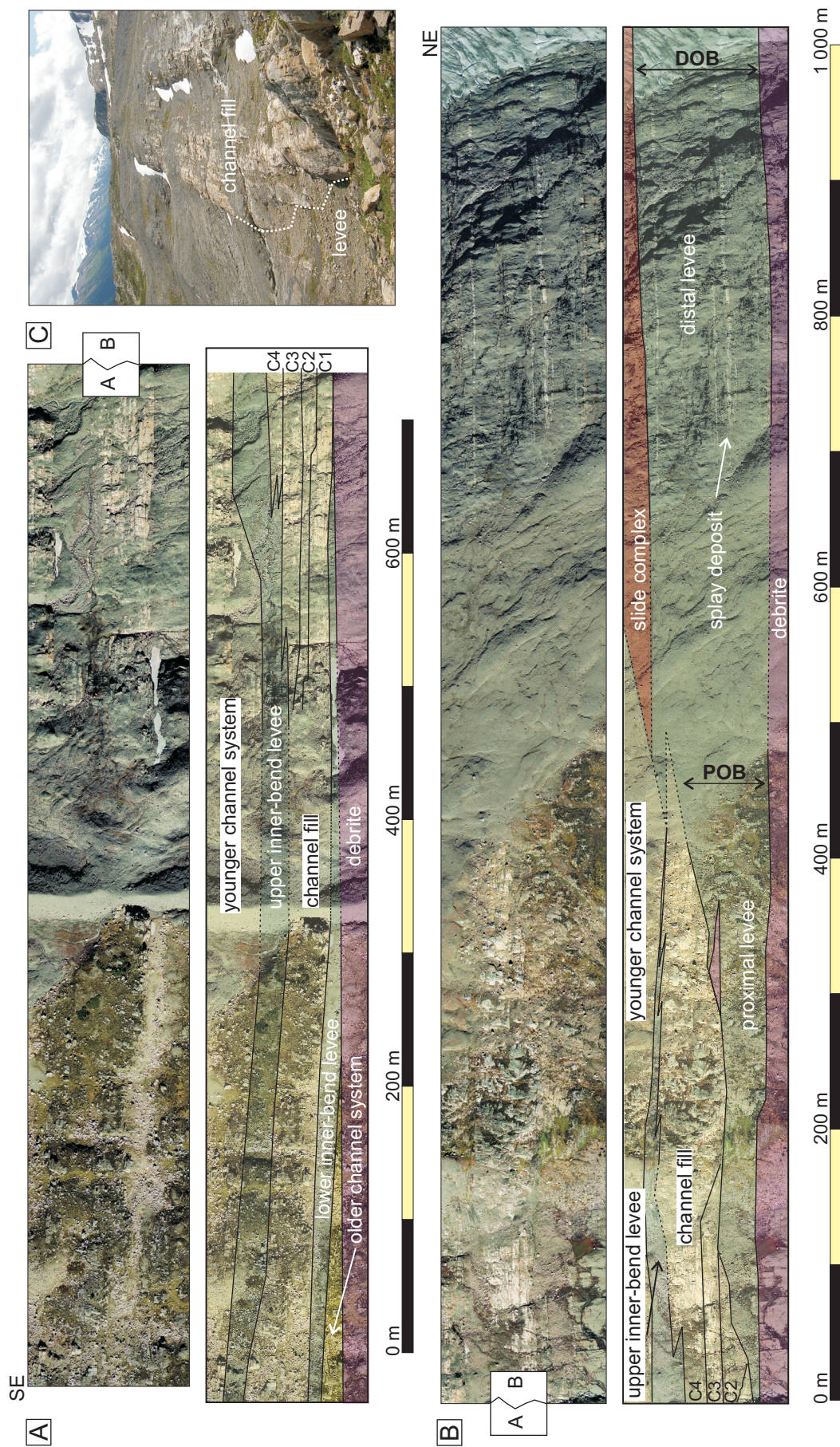


Fig. 4.4. Photomosaic of Isaac Channel Complex 3 showing the continuity of outcrop and exposure of levees along both sides of the channel. Channel-fill strata are divided into four units (C1 – C4). Levee strata along the inner-bend (SE) side of the channel in (A) are divided into lower and upper units. Strata in the inner-bend levee consists predominantly of thin-bedded turbidites. Strata in the lower unit are truncated by the shallow-dipping channel margin, whereas in the upper unit channel-strata thin and fine laterally away from the channel. In (B), levee strata in the outer-bend proximal levee range from very thin- to thick-bedded turbidites. Note the occurrence of an up to 10-m thick debrite within the outer-bend proximal levee. Collectively these deposits are sharply separated from channel-fill deposits along the terraced channel margin (C). Proximal levee strata transition laterally over hundreds of meters into principally thinner-bedded distal levee deposits that locally are intercalated with up to 5 m thick amalgamated overbank splay units. ICC3 is bounded at its base by a debrite and older channel system and at its top by a slide complex and younger channel unit. POB = proximal outer-bend levee, DOB = distal outer-bend levee.

upper two units are less than 500 m wide (Fig. 4.4); however, the apparent upward narrowing may be an artefact of the changing obliquity of the cut relative to the axis of the channel. Individual channel-fill units generally fine and thin upward and collectively form a large-scale upward-thinning and -fining trend. The coarsest strata, consisting of pebbly conglomerate, occur at the base of C1. Beds are up to 2.5 m thick and are laterally discontinuous over distances of hundreds of meters, typically because of scour by younger conglomerate beds. Upward, these strata grade into beds of crudely normally graded, coarse-grained sandstone, which are up to 2.5 m thick and are laterally continuous, in some cases over several hundreds of meters. Still further upward, these sandstone beds become progressively interbedded with thinly-bedded siltstone turbidites. In contrast, C2 consists of an extensive mud-clast breccia at its base and less common conglomerate beds. The upper channel fill units, C3 and C4, are generally finer grained than C1 and C2, and at their bases consist typically of very coarse sandstone and local breccia layers. The C3 and C4 units also fine- and thin-upward, but, unlike C1 and C2 grade, rapidly laterally over a few hundred meters from amalgamated coarse-grained sandstone into medium- to thick-bedded coarse sandstones turbidites interbedded with thin- to medium-bedded upper-division turbidites toward the channel margins.

4.5. Lithofacies of overbank strata in ICC3

In relation to the channel fill of ICC3, overbank strata are subdivided into 3 discrete kinds: proximal outer-bend (POB), distal outer-bend (DOB), and inner-bend levee (IB) (Fig. 4.4). The outer- and inner-bend levee interpretation is based primarily on stacking patterns and lithology (see Chapter 1). Note that POB and DOB are separated by an up to 250-wide

moraine that buries intervening strata. The IB levee is subdivided vertically into a lower (IB1) and upper (IB2) unit. Three main lithofacies are recognized in overbank strata of ICC3. These are 1) medium- to thick-bedded, coarse to fine sandstone T_{a-d} turbidites (F1), 2) very thin- to thin-bedded, fine sandstone to siltstone T_{cd} turbidites (F2), and 3) chaotic overbank deposits (F3). Facies are briefly described next and are summarized in Table 4.1.

Table 4.1
Characteristics of overbank lithofacies

Lithofacies	Lithology	Bed thickness*	Sedimentary structures	Bouma divisions	Bounding surfaces	Lateral trend	Depositional process
F1	Coarse- to fine-grained sandstone	10 – 150 cm	Massive, parallel laminated, and ripple cross-stratified	Most common T_{bc} , T_b , and T_c , less common T_a , T_{ab} , and T_{abc} . Beds are typically overlain by thin massive T_d cap	Sharp bases and tops	Thicken laterally over ~120 m away from the channel and thereafter thins over the next 100's of meters	Unconfined, depletive waning flows
F2	Fine-grained sandstone to siltstone	<1 – 10 cm	Generally ripple cross-stratified, rare massive to planar laminated. Siltstone cap are generally massive and rarely laminated	Most common T_{cd} , less common T_{abcd} , and T_{bcd}	Sharp bases and gradational tops	Grain size and bed thickness change negligibly laterally	Unconfined, uniform waning flows
F3a	Pebble clasts within a mudstone matrix	50 cm – 10 m	Structureless	N/A	Sharp non-erosive to erosive	Lenticular over 300 m to sheet-like	Debris flow
F3b	Commonly distorted F2 strata	Individual beds <1-10cm. Deformed stratal unit up to 5 m	Folds at the meter scale and faulting at the centimeter scale	Deformed T_{cd}	Sharp irregular bases and tops	Local deformation pockets less than 100 m wide	Local slumping of overbank deposits

*follows the nomenclature and classification proposed by Bouma (1962) and thickness terms of McKee and Weir (1953)

4.5.1. Description of medium- to thick-bedded, coarse to fine sandstone T_{a-d} turbidites (F1)

Facies 1 (F1) beds are most common in the POB but occur also in IB and DOB. Beds generally range from 20–70 cm thick although beds up to 1.5 m thick are observed. F1 beds are T_{a-d} turbidites that normally grade from moderately-sorted medium to fine sandstone; however, the basal parts of some beds consist of massive to subtly graded coarse-grained sandstone (Fig. 4.5). Beds are typically abruptly overlain by a 1–5 cm thick layer of coarse to fine siltstone marking a sharp grain size break. Less commonly beds lack an upper siltstone (T_d) cap and form amalgamated units up to 5 m thick, the thickest of which occur in the distal levee. The majority of basal contacts of F1 beds are sharp and planar; however, a small number conform to the shape of underlying ripple form-sets.

Table 4.2 shows the percentage of the different type of Bouma successions occurring within F1 sandstones in the POB 100 m and 350 m from the defined channel margin. Massive and normally graded T_a sandstones are observed in almost half of the beds 100 m away from the channel. In contrast, 300 m away from the channel T_a sandstones occur in only 23% of the beds, half of which are overlain by either planar or cross-laminated sandstone (Fig. 4.5A), whereas the other half are overlain by a subsequent T_a turbidite or a thin T_d layer. Parallel lamination (T_b) and non-climbing ripple cross-lamination (T_c) are the most common sedimentary structures observed in F1 strata occurring in approximately 84% and 90% of beds 100 m and 350 m away from the channel, respectively. Thirty three percent of planar laminated (T_b) divisions are overlain by a ripple cross-laminated (T_c) division forming a T_{abcd} or T_{bcd} Bouma turbidites (Fig. 4.5B). Sandstone beds characterized by T_{bcd} sequences range from 30 to 80 cm thick and there is no correlation between total bed thickness and the percent thickness of the T_b or T_c division. For example, the T_b division may account for 10 to 90% of a T_{bcd} bed regardless of the total bed thickness. Moreover, 52% of

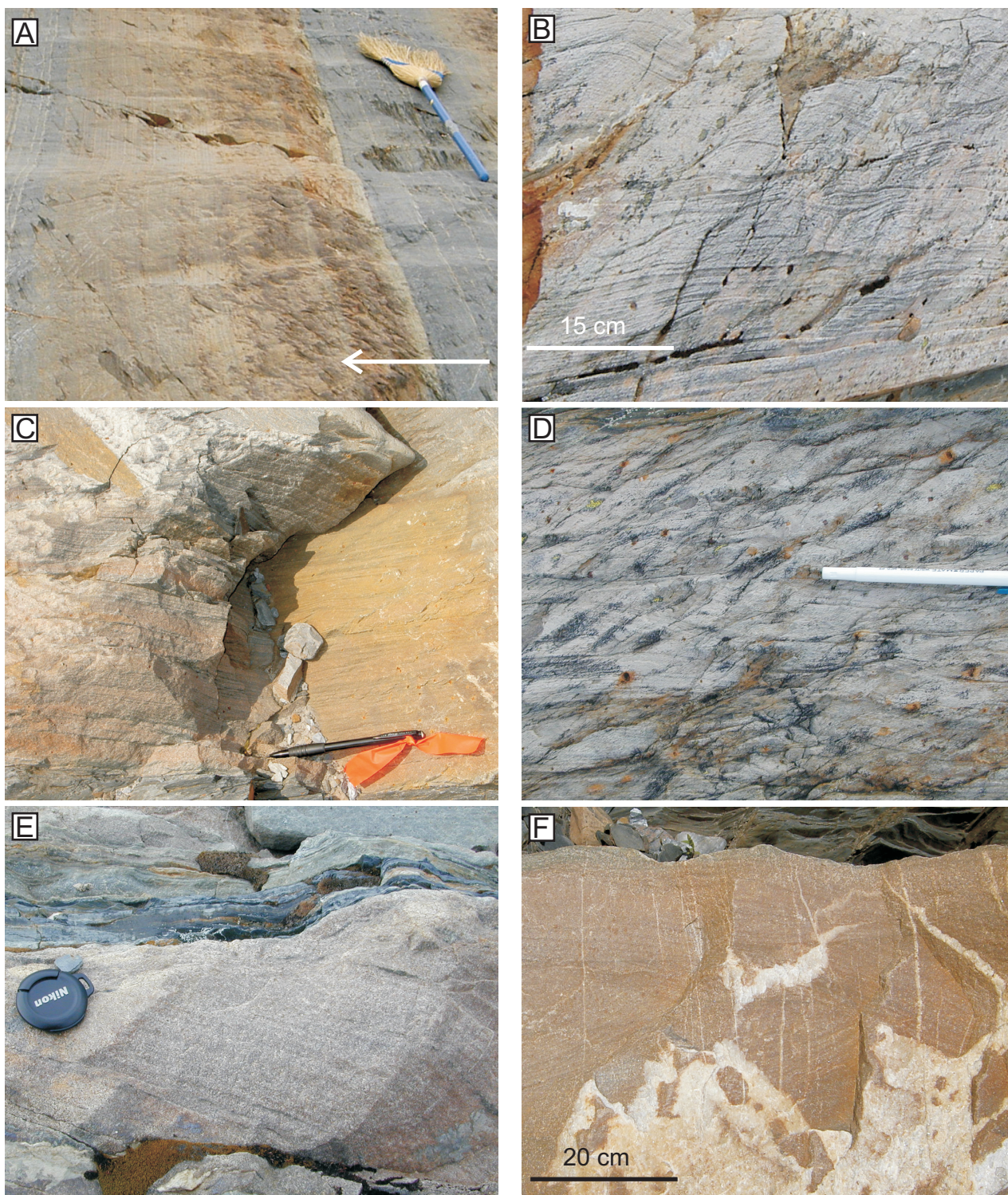


Fig. 4.5. Facies F1 strata (medium- to thick-bedded, coarse to fine sandstone Ta-c turbidites). A) Thick-bedded T_{ab} turbidite that fines upward (towards the left) from medium to fine-grained sandstone. The transition from massive to planar laminated occurs approximately at mid height of the bed. B) Thick-bedded T_{bc} turbidite that fines from medium to fine-grained sandstone. The planar laminated part of this bed occurs only in the lower 6 cm. C) Medium-bedded, medium-grained sandstone T_b turbidite with minor vertical fining. D) Thick-bedded, medium- to fine-grained T_c turbidite consisting of multiple sets of non-climbing ripples. Bed fines little vertically from medium-grained to upper fine-grained sandstone. E) Medium-bedded, medium-grained (no vertical grading), dune cross-stratified turbidite. F) Multiple medium-grained dune trough cross-stratified sets underlain by a coarse-grained sandstone, T_a turbidite. Note that the bed is also capped by a single ripple set.

Table 4.2

Percent occurrence of the different type of Bouma successions occurring within F1 sandstones (outer-bend proximal)

Bouma succession	100 m laterally from channel margin	350 m laterally from channel margin
Ta	16%	10%
Tab/Tac	13%	5%
Tabc	16%	7%
Tbc	23%	26%
Tb	16%	31%
Tc	16%	21%

Numbers in each column are based on 42 F1 beds

F1 beds consist exclusively of either planar laminated or ripple cross-laminated sandstone commonly capped by a thin T_d layer (Figs. 4.5C and D). Beds composed primarily of parallel laminated sandstone vary from 30–100 cm thick, whereas current ripple cross-stratified beds are up to 40 cm thick and, where visible, consist of sets 1.5 to 3 cm thick with ripple spacing of 15-20 cm. Dune cross-stratification is rare and only occurs in about five beds that range from 10 to 50 cm thick. These beds commonly consisting of a single dune cross-stratified set (Fig. 4.5E). Less commonly, these beds are underlain by a T_a or T_b division and at their upper contact preserve ripple bedforms (Fig. 4.5F).

4.5.2. Interpretation of medium- to thick-bedded, coarse to fine sandstone T_{a-d} turbidites (F1)

Structureless T_a divisions at the base of some F1 sandstone beds were most likely deposited rapidly from suspension (Lowe, 1988; Arnott and Hand, 1989; Sumner et al., 2008). Arnott and Hand (1989) showed experimentally that parallel lamination becomes progressively more poorly developed as the rate of suspension fallout, and accordingly bed aggradation, increased. Implied in this model is that the few-grain-high (low-amplitude) bedforms that form upper-stage plane bed eventually become buried so rapidly that their distance of down flow migration becomes so limited that lamination becomes imperceptible.

Sufficiently high rates of sediment fallout needed to suppress parallel lamination would easily be achieved as high concentration flows became unconfined as they overtopped the channel banks, resulting in competence- and possibly even capacity-driven deposition (e.g. Hiscott et al., 1994; Kneller and McCaffrey, 1996). The scarcity of erosional basal contacts suggests that overbanking flows in ICC3 most probably became immediately depositional. Rare amalgamated F1 units, however, likely indicate erosion by later flows that removed the uppermost part of the underlying bed (i.e. T_d layer). In addition, rare erosion is recorded locally by minor scour or rare rip-up clasts at the base of some beds. In comparison to T_a beds, planar laminated (T_b) divisions associated with F1 indicate deposition from flows where the rate of suspension fallout was sufficiently low that the distance of low-amplitude bedform migration became long enough that planar lamination became visible (Paola et al., 1989). Accordingly, the transition from T_a to T_b deposition indicates a temporal reduction in sedimentation rate.

Non-climbing ripples are common in levee deposits of ICC3 and either overlie T_a or, more commonly, T_b divisions, or form beds up to 40 cm thick composed entirely of the T_c division. The transition from T_b to T_c is generally thought to be associated with a gradual decrease in flow speed (Walker, 1965; Komar, 1985). However, according to standard stability bedform diagrams (e.g. Southard and Boguchwal, 1990) as flow speed decreases for bedload sediment coarser than very fine sand, upper-stage plane bed should be succeeded by dunes. Nonetheless the fact remains that dunes are seldom reported in levee deposits, or classical turbidites for that matter, even though sediments of sufficient grain size are widespread in many deep-water systems (e.g. Walker, 1985; Beaubouef, 2004; Eschard et al., 2003; Grecula et al., 2003; Kane et al., 2007). Therefore, factors in addition to reduced flow velocity must control the deposition of T_{bc} turbidites. A number of explanations have

been proposed, including lack of sufficient time for dune growth (Walker, 1965), near-bed turbulence suppression (Lowe, 1988), flow unsteadiness (Allen, 1970), and insufficient grain size (Allen, 1969). Recently Arnott (2007e) suggested that high sediment concentration inhibits the amplification of bed defects from which angular bedforms like ripples and dunes grow. Under such conditions plane beds remain stable even at flow speeds that, in a clean-water flow, would form dunes or ripples. The paucity of dune cross-stratification, but the ubiquity of ripple cross-stratification in the deep-water record, suggests that the concentration threshold when bed defects become amplified is typically attained in the ripple field consequently skipping the entire stability region of dunes. Moreover, observations of dune cross-strata without an underlying T_b division in ICC3 levee strata, suggest deposition from relatively steady, uniform, dilute turbidity currents beneath which bed defects could be sufficiently amplified to facilitate dune migration downflow. The rarity of dune cross-stratification suggests that such conditions were rarely met.

Facies 1 (F1) beds are commonly overlain abruptly by a thin layer of siltstone. It is unlikely that this sharp break in grain size reflects a deficiency of the missing grain sizes (i.e. lower fine and very fine sandstone fraction) because this grain size is widespread throughout the WSG deep-water system and commonly occurs immediately adjacent to F1 sandstone beds. More likely, and as suggested by Kane et al. (2007) for levee deposits in the Rosario Formation, the missing grain sizes reflect competent currents that maintained the grains in suspension and bypassed them to more distal areas. Therefore, the thin silt caps that overlie F1 sandstone beds are interpreted to have been deposited from suspension during the waning stages of the flow when the remaining sediment in the flow was silt whereas the missing grain size fraction was transported laterally during higher energy stages of the flow.

4.5.3. Description of very thin- to thin-bedded, fine sandstone to siltstone T_{cd} turbidites (F2)

Facies 2 (F2) sandstone beds occur in both proximal (inner and outer-bend) and distal levee deposits. Beds range from 1 to 10 cm thick and consist of T_{cd} turbidites with rare T_b bases (Figs. 4.6A, B). Beds commonly grade from fine sandstone to siltstone and, unlike F1, commonly show no sharp break in grain size. The basal T_c portion of these beds is up to 8 cm thick and comprises single to multiple (up to four) non-climbing ripple cross-stratified sets. Laterally discontinuous (starved) ripple trains occur in some thinner T_{cd} beds less than 5 cm thick. In the uppermost part of T_c divisions, individual ripple foresets are commonly interlaminated with finer-grained sandstone and silt drapes (Fig. 4.6C). The T_d divisions typically range from 2 to 8 cm thick and the thickest siltstone beds commonly overlie the thinnest T_c divisions. They are locally faintly parallel laminated but most commonly appear structureless. Where present, T_d laminae are commonly characterized by interlaminated thin (<2 mm) light grey coarse siltstone or very fine-grained sandstone and dark grey finer-grained siltstone (Fig. 4.6D). These beds lack a true mud-rich T_e cap as this grain size is underrepresented in the entire ICC3 system.

4.5.4. Interpretation of very thin- to thin-bedded, fine sandstone to siltstone T_{cd} turbidites (F2)

The deposition of F2 is interpreted to be largely from dilute, fine-grained overspilling turbidity currents. Similar to F1, the lower ripple cross-laminated portions of beds do not climb, indicating that sedimentation rates were consistently low. Also, the gradual upward change from fine sandstone to siltstone suggests a gradual reduction in flow competence. Common structureless T_d divisions in ICC3 levee strata are interpreted to represent the very fine-grained suspension settling toward the end of a flow event. Rare laminated T_d divisions

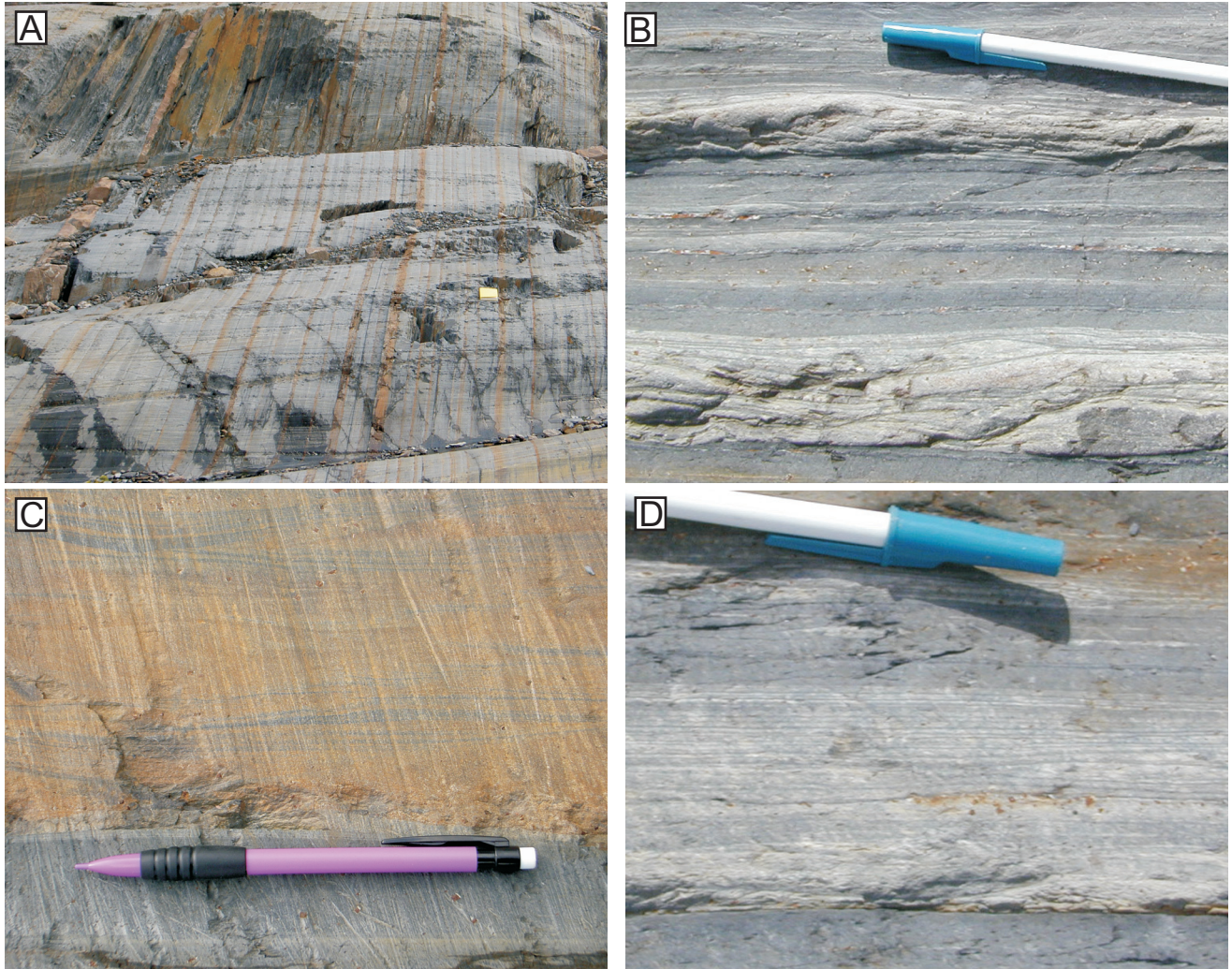


Fig. 4.6. Facies F2 strata (very thin- to thin-bedded, fine sandstone to siltstone T_{cd} turbidites). A) Predominantly thin-bedded, fine sandstone to siltstone T_{cd} turbidites (yellow notebook for scale). B) Close up of very thin- to thin-bedded T_{cd} turbidites. C) Fine-grained T_c turbidite with interlaminated finer-grained drapes. D) Interlaminated thin light grey very fine-grained sandstone and dark grey siltstone.

recorded as coarse siltstone or very-fine sandstone interlayers within finer-grained siltstone are interpreted to have been created by the presence of internal waves within the channelized turbidity current. This in turn potentially caused fluctuations (or pulses) in the hydraulic regime and grain-size make up of the overspilling flow (e.g. Hiscott et al, 1997; Gervais et al., 2001). Alternatively, Stow and Bowen (1980) suggest that the fine-scale interstratification of coarser and finer layers is the result of rhythmic changes in

flocculation/deflocculation and sediment settling processes within the viscous sublayer of a low-energy turbidity current.

4.5.5. Description of chaotic mud-rich deposits (F3)

Facies 3 (F3) is divided into two sub facies: poorly sorted muddy deposits (F3a) and deformed thinly-bedded deposits (F3b). The thickest F3a deposit is up to 10 m and occurs within the proximal levee where it overlies a succession of interbedded F1 and F2 strata (Fig. 4.4). This unit consists of a low percentage (10%) of pebble-size quartz clasts dispersed in a silty mudstone (Figs. 4.7A, B) and pinches out over 150 m laterally away from the channel margin and in the opposite direction is truncated by the channel margin. Over the entire length of the unit the base of the bed incises up to 3 m into the underlying stratigraphy. Facies 3a strata occur also as beds that range from 50 to 100 cm thick and are located in the distal levee. These beds consist of abundant quartz pebble clasts (up to 40%) dispersed in a siltstone-mudstone matrix (Fig. 4.7C) and, in some cases, clast abundance and size decreases upward. Bed bases are sharp and planar with only minor local erosion. Units are laterally continuous over 300 m but pinch and swell laterally.

Facies 3b (F3b) is also present in the distal levee and consists of units up to 5 m thick composed of deformed F2 thinly-bedded turbidites. Syn-depositional deformation of these units is expressed as tight, overturned to recumbent folds (Fig. 4.7D) or more open, non-overturned folds. Also, small-scale faulting with apparent reverse and normal sense of movement is recorded.

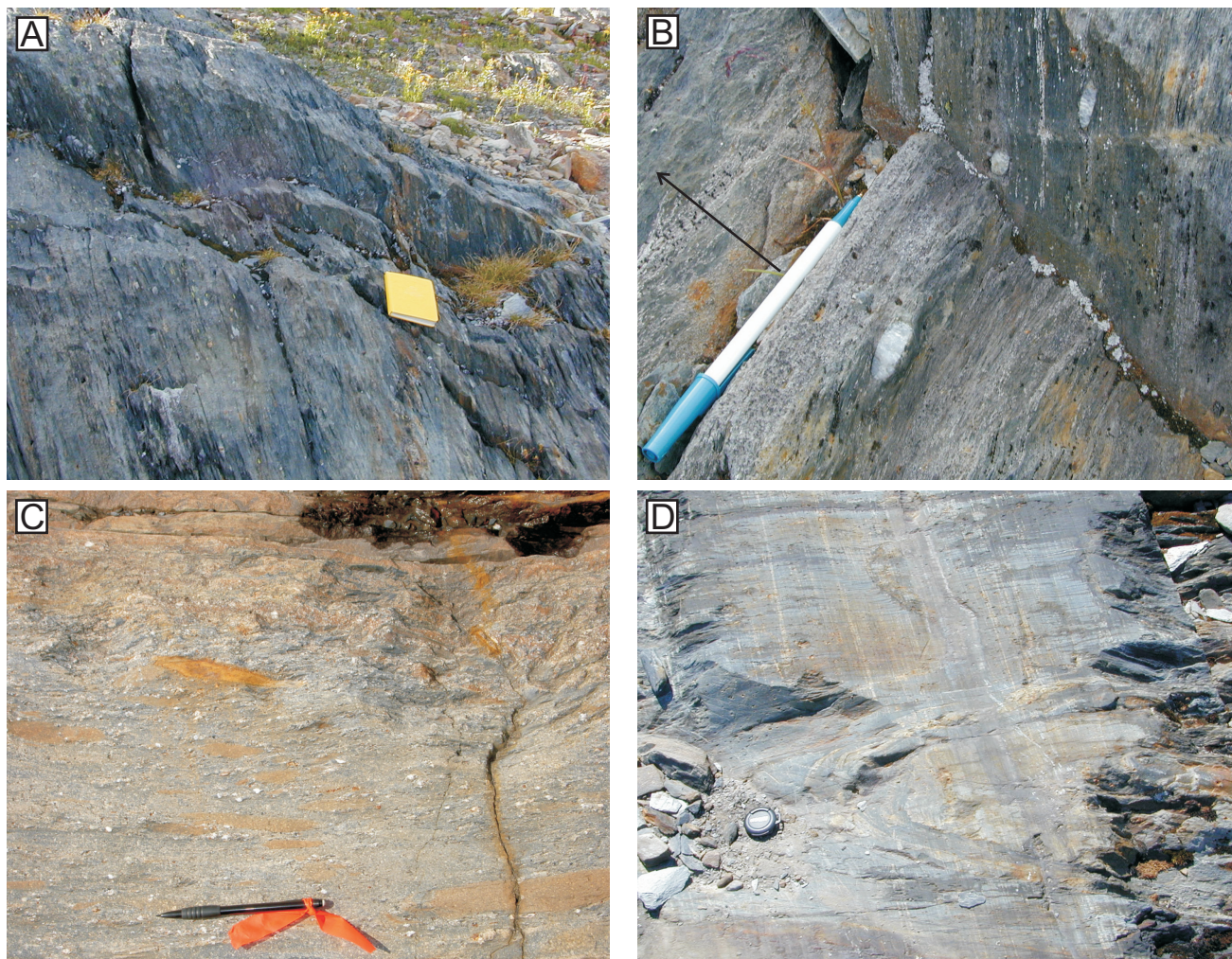


Fig. 4.7. Chaotic strata of facies F3. A) Part of a thick (up to 10 m) debris flow deposit occurring in the proximal outer-bend levee. Yellow notebook is located on contact between underlying interbedded levee strata and debris flow deposit (stratigraphic top is to the left). Note lack of stratification in the debris flow unit. B) Close up of debris flow unit in (A). Note the occurrence of pebble clasts dispersed in a mudstone matrix. C) Thinner debris flow units in the distal outer-bend levee. Note the higher abundance of quartz clasts compared to (B). D) Recumbent fold occurring locally within the distal outer-bend levee (lens cap for scale).

4.5.6. Interpretation of chaotic mud-rich deposits (F3)

The F3a deposit occurring immediately adjacent the channel margin is interpreted to have been deposited by a throughgoing debris flow that spilled over the channel banks and deposited in the proximal levee area. Thinner F3a deposits occurring in the distal levee are interpreted to have been deposited by more fluidal overspilling debris flows that bypassed the proximal levee. Alternatively, these deposits could have been deposited by debris flows

that descended the slope adjacent ICC3 and deposited in the distal levee lows. The upward decrease in clast abundance suggests also a comparatively more fluidal sediment mass with syn and/or post-depositional differential particle settling. Strata of F3b are only slightly deformed and bedding remains visible. Accordingly, strata of F3B are interpreted to be slump deposits resulting from local mass movement of semi-consolidated, thinly-bedded turbidites. The occurrence of these units in the distal levee suggests that movement took place on the sloping surface along the backside of the levee (Hickson and Lowe, 2002; Hubbard et al., 2008).

4.6. Geometries and spatial trends of overbank elements

4.6.1. Description of vertical packaging in outer-bend proximal levee

The proximal outer-bend levee is subdivided vertically into three stratal packages (POB1–POB3) that range from 12 to 28 m thick (Figs. 4.8A, B). Individually, each package is divided into lower and upper parts. The contact between these two units is clearly marked by an abrupt decrease in average bed thickness and sandstone proportion (Figs. 4.8B, C, 4.9A, B). Previously, Navarro et al. (2007b) divided the proximal levee of ICC3 into four units. The top unit of Navarro et al. (2007), however, lacks a well developed lower part and here makes up the majority of the upper part of POB3.

The lower part of packages range from 7 to 11 m and consist of intercalated medium- to thick-bedded and very thin- to thin-bedded turbidites of F1 and F2, respectively (Figs. 4.8B, C). Facies 1 (F1) strata account for about 60% of the total thickness of the lower part of packages and individual beds average about 40 cm in thickness. The basal sandy part of intercalated F2 strata, on the other hand, generally range between 1–2 cm thick, but on

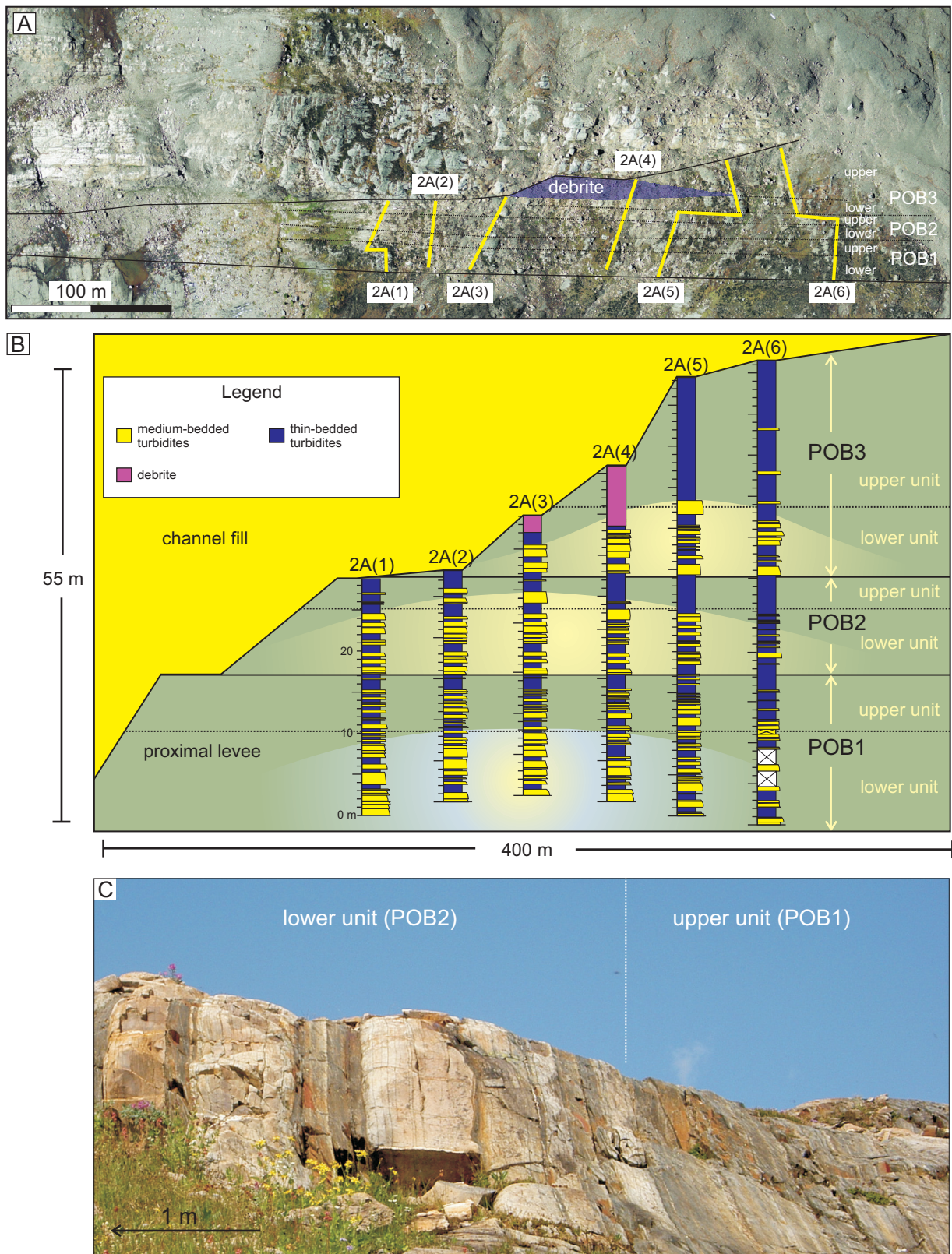


Fig. 4.8. Levee deposits in the proximal outer-bend levee. A) Aerial photo showing location of measured stratigraphic logs. B) Correlation panel showing the detailed architecture of proximal levee deposits. Proximal levee deposits comprise three distinct packages that are divided into a lower part that comprises interbedded thick-bedded (F1) and thinner-bedded (F2) turbidites and a more siltstone-rich, thin-bedded upper part (F2). Beds in the sandier lowermost parts of packages typically thicken then thin laterally. As a result, lower packages as a whole typically increase and then decrease in N:G laterally away from the channel margin (represented by the lateral shade grading). Note that levee packages POB1 to POB3 show a consistent lateral offset that is especially noticeable when comparing POB2 with POB3 – this pattern mirrors the direction of lateral migration of the channel. C) Photograph showing the sharp contact between the upper part of POB1 and the lower part of POB2.

average account for about 70% of the total number of sandstone beds recorded in the unit.

Accordingly, the average sandstone bed thickness (10–20 cm) is biased by the abundance of very thin- to thin-bedded F2 strata (Fig. 4.9A).

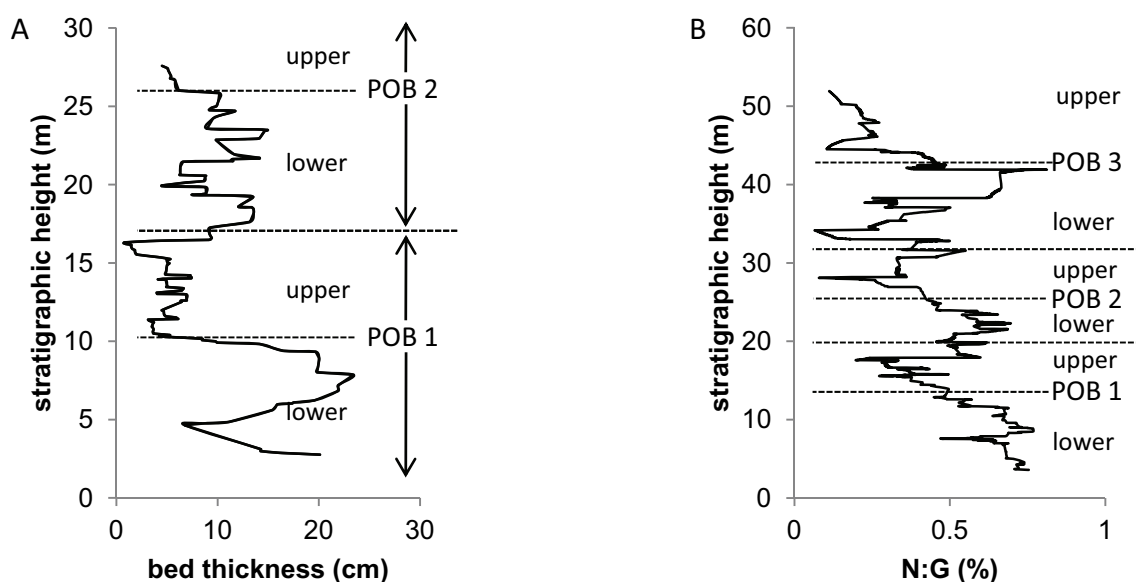


Fig. 4.9. A) Data displayed from 474 sandstone beds (not including siltstone tops) showing how bed thickness varies upward. Data for POB1 and POB2 is from Log A2(1) whereas data for POB3 is from Log A2(5.5). Log A2(5.5) was measured between logs A2(5) and A2(6) (see Fig. 8b). B) Data displayed from 524 beds (including siltstone beds) measured in Log A2(5.5) showing upward trend in N:G. A decrease in N:G indicates an upward fining trend. These plots use a 15-bed moving average.

The upper parts of levee packages range from 5–18 m thick and consist of at least 85% very thin- to thin-bedded F2 strata. The sandy part of F2 beds, like those in the lower part of packages, typically range from 1–2 cm thick. Although significantly less common, intercalated beds of F1 in the upper part of POB1 and POB2 average about 22 cm thick and account for about 30% of the total thickness. As a result average sandstone bed thickness is disproportionately increased to about 4.5 cm (Fig. 4.9A). Facies 1 (F1) strata in the upper part of POB3, in contrast, account only for 6% of the total thickness and accordingly have little impact on average bed thickness. Moreover, the ratio of sandstone thickness to total

stratal thickness, or simply N:G, progressively decreases upward (Fig. 4.9B) – note that sandstone is considered to consist of sediment coarser than very fine sand. Beds thin upward gradually from 7 to 3 cm thick in the upper parts of POB2 and POB3. A more crude upward thinning trend is recorded in the upper part of POB1 where beds average 7 cm and then thin rapidly in the upper 1.5 m to less than 3 cm. In addition, each proximal outer-bend package (lower + upper parts) displays an overall upward fining trend (Fig. 4.9B).

4.6.2 Interpretation of vertical packaging in outer-bend proximal levee

Upward trends in levees are interpreted to be linked to changes in the characteristics of overspilling flows, which, in turn, are controlled by the amount of levee relief (channel base to levee crest), in addition to the thickness, velocity, and vertical sediment concentration profile of throughgoing channel flows. Consequently, channels with low levee relief allow much of the flow to overspill, including the lower sandier, more concentrated, faster moving basal part, resulting in the deposition of thicker and coarser-grained turbidites. Channels with high levee relief, on the other hand, allow only the upper, more dilute, slow moving part of the throughgoing channel flows to overspill, resulting in deposition of thinly-bedded, fine-grained turbidites. Accordingly, in ICC3, periods of low channel confinement allowed relatively large-magnitude flows to deposit medium- to thick-bedded F1 strata on the outer bank and make up the lower parts of levee packages. Interstratified finer, thinner strata of F2 within the lower parts of packages were similarly deposited during conditions of low channel confinement, but by channel flows of significantly smaller magnitude in which the coarser basal part of the flow remained confined entirely within the channel. The intercalation of F1 and F2 strata in the lower parts of levee packages suggests that flow magnitude within the

channel commonly fluctuated. In the lower part of levee packages 50–80% of the beds are F2 implying that most overspilling flows were comparatively small.

The general vertical fining of strata in proximal levee packages in ICC3 (Fig. 4.9B) is interpreted to have resulted from the deposition of progressively finer-grained overspilling flows associated with increasing levee relief. Accordingly, the abrupt transition between lower and upper parts of levee packages suggests that channel confinement likely achieved a threshold height where after only the upper, more dilute portion of throughgoing channel flows typically overspilled. In the case of POB3 a locally occurring debrite that overlies the lower part of the levee package is interpreted to have been deposited by frictional freezing processes and is inferred to have abruptly added levee relief. This added relief further confined the lower, sandy parts of subsequent flows within the channel (Navarro et al., 2007b). Importantly, it is likely that during periods of high levee relief that small magnitude flows passing through the channel were confined completely within the channel or that only the very dilute uppermost part of the flow would be stripped and enter into the levee area. Such dilute overspilling flows, however, would lack sandy sediment resulting in the inability to form the sandy ripple cross-stratified basal part of F2 beds and consequently would not be differentiated in the field as individual events but instead considered part of an earlier single T_d layer. If a similar ratio of small to large magnitude flows calculated from the lower part of levee packages is assumed for the upper part, then it is possible that up to 80% of throughgoing channel flows during periods of high channel confinement did not overspill enough sandy sediment to be recorded within the levee as an individual event. This, therefore, contradicts the assumption that levees record the passage of almost all throughgoing channel flows (Clark and Pickering, 1996; Hiscott et al, 1997).

Sometime during the deposition of upper levee packages, flows within the channel are interpreted to have changed from mostly non-depositional to predominantly depositional. The consistent and gradual upward fining observed in all the channel fills is interpreted to be related to allocyclic processes, such as relative sea-level rise or reduced fluvial sediment flux, was the dominant driving force behind changes in flow characteristics and related sedimentation patterns. Rapid filling of the channel associated with, for example, each relative sea-level rise, would raise the channel floor at a faster rate than the levee aggraded and cause levee relief to become progressively reduced. An ensuing increase in flow magnitude would then once again allow the basal sandier part of throughgoing channel flows to overspill resulting in deposition of the lower part of the overlying levee package. All levee packages are interpreted to have aggraded in a similar fashion, which is manifest in the consistent vertical stacking of levee deposits in POB1–POB3.

4.6.3. Description of lateral trends in outer-bend proximal levee

In the outer-bend area levee and channel-fill strata are separated by an erosive and terraced surface that has an apparent local dip of up to 25° (Fig. 4.4). Here, F1 strata show a distinctive lateral thickening then thinning trend (Fig. 4.10). Specifically, individual beds typically thicken rapidly laterally over about 120 meters exhibiting an average thickening gradient of 0.3 (thickening (cm)/lateral distance (m)). After achieving their maximum thickness, beds begin to thin over the next 180 m with an average thinning ratio of 0.2. Once beds thin to about 10 cm or less, and now are considered to be F2, they show negligible lateral thinning over the remaining length of the outcrop (up to 100 m). Similarly, beds in the proximal levee that are initially less than 10 cm thick (F2) show negligible lateral thinning along their entire length. Although rare, some F1 and F2 beds are observed to pinch out

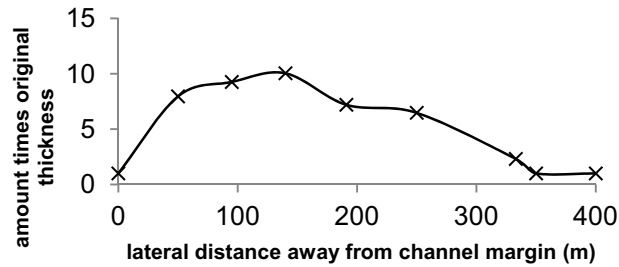


Fig. 4.10. Average thickening then thinning trend of medium-bedded sandstone in the lower part of levee packages. This trend is the average of 25 thick- and medium-bedded sandstones that display a well developed thickening then thinning trend in the lower part of levee packages.

entirely within the proximal levee. Moreover, lateral thinning of F1 strata generally coincides with a lateral fining trend wherein grain size typically decreases from medium- (up to coarse-) to fine-grained sandstone. In contrast, grain size in F2 strata display negligible lateral change.

The lateral trends in bed thickness directly control lateral changes in N:G. For example, Figure 4.11A shows that the sandier lower part of levee packages, which in terms of overall thickness are composed mostly of F1 sandstones, are at their maximum N:G about 120 m laterally away from the channel reaching values ranging from 0.6–0.8. Subsequently, N:G decreases rapidly over the next 180 m away from the channel to values as low as 0.38. In the upper part of levee packages, strata consist mostly of lower N:G F2 beds that show negligible lateral change in thickness or grain size, and therefore N:G. Nevertheless, it is the thicker but much less common F1 beds that control the lateral N:G trend. For instance, in the upper part of POB1 and POB2 N:G begins to decrease appreciably only at 300 m and 150 m (Fig. 4.11B), respectively, which coincides with the location where F1 strata rapidly transition laterally into thinly-bedded F2 strata (see Fig. 4.8B). The upper part of POB3, in contrast to underlying packages, consists of relatively few F1 beds and therefore N:G changes negligibly over 100 m (Fig. 4.11B). A single levee package (lower + upper part)

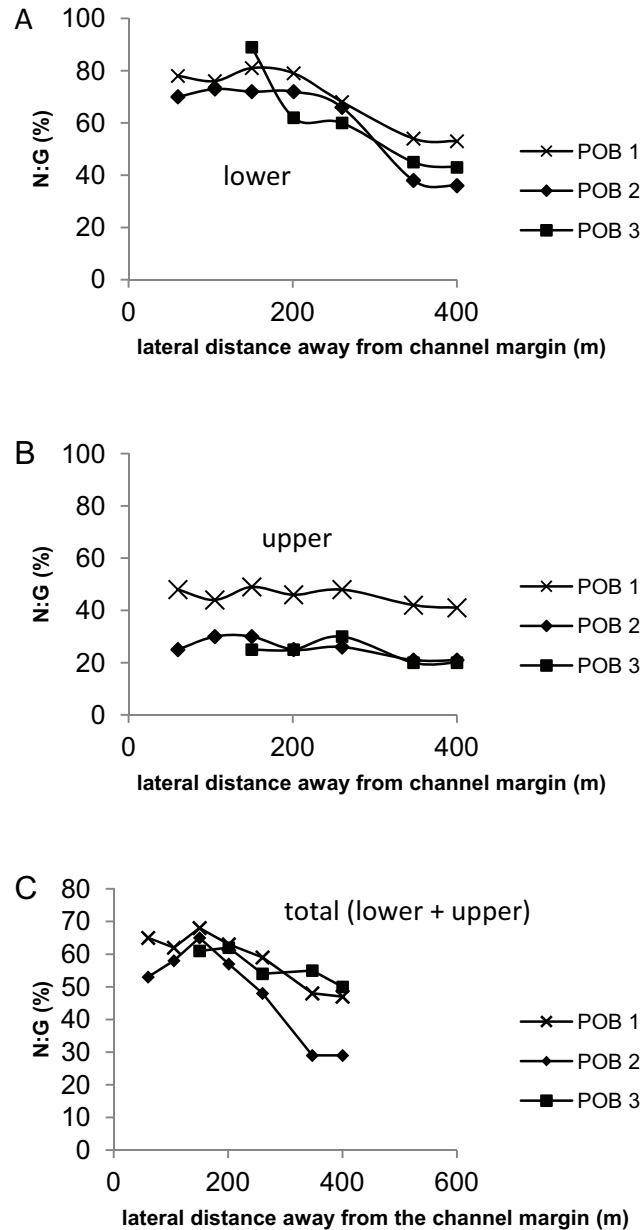


Fig. 4.11. Lateral variation of N:G in the proximal outer-bend levee (X = POB1; Δ = POB2; $\square$ = POB3) A) N:G trends in the lower part of packages. In places strata immediately adjacent the channel margin are not well exposed and as a result N:G could not be measured there. Nevertheless the existing database suggests that F1 beds thin toward the channel, which in turn decreases N:G. B) N:G trends in the upper part of packages. The paucity of F1 strata in the upper part of POB3, in comparison to POB1 and POB2, is reflected by a negligible lateral change in N:G. C) N:G trends of complete levee packages (lower + upper parts).

yields maximum N:G values up to 0.68 close to the channel margin and as low as 0.34 over the next few hundred meters laterally away from the channel margin (Fig. 4.11C).

4.6.4. *Interpretation of lateral trends in outer-bend proximal levee*

In the literature it is commonly reported that thick-bedded turbidites typically thin laterally faster compared to thin-bedded turbidites (DeVries and Linholm, 1994; Beaubouef, 2004). Thickening then thinning trends are less commonly reported but have, for example, been described from levee deposits in the Rosario Formation, Mexico (Kane et al, 2007). Straub et al. (in press) also reported an example from seismic images offshore Brunei Darussalam where mean levee thickness is interpreted to increase rapidly away from the channel, reaching a maximum at the levee crest, and then gradually thin until a certain thickness was achieved, after which thickness remained constant.

The initial thickening trend observed in F1 sandstones is interpreted to be linked to the overspilling flow characteristics of high energy, highly concentrated flows. Baas et al. (2004) showed experimentally that supercritical channel flows quickly become depositional upon becoming unconfined. These experiments also showed that the maximum rate of deposition associated with denser turbidity currents occurred further from the point of unconfinement compared to slower, less dense ones. This difference is most likely related to greater inertia of concentrated fast-moving overbanking flows and the outward displacement of the zone of maximum sedimentation, which in the case of the ICC3 levee deposits is of the order of 100–150 m (Fig. 4.10). Beyond the zone of maximum deposition, F1 beds begin to fine and thin laterally, suggesting a more progressive loss of competence and capacity. Under such conditions the coarsest particles with the highest settling velocities would concentrate towards the base and then fall from suspension. With the progressive loss of coarse sediment the flow becomes increasingly composed of finer-grained particles with lower settling velocities, and accordingly lower sedimentation rates. This lateral change in sedimentation rate and particle grain size is recorded by the lateral thinning and fining trend

observed in F1 strata. Furthermore, sandy flows responsible for the deposition of F1 strata would eventually become depleted laterally in their coarser grained fraction to the point where the remaining particles in suspension would be characterized by a sufficiently low settling rate and that turbulent mixing creates an almost uniformly distributed sediment dispersion (Kneller and McCaffrey, 2003; Baas et al, 2004). Deposition under such conditions resulted in the deposition of F2 strata wherein negligible lateral thinning or fining occurs. Therefore, the progressive loss of flow competence associated with initially sandy overbank flows explains the common lateral transition from F1 to F2 strata. In contrast, F2 beds occurring directly adjacent to the channels, and which change little laterally, were deposited by overspilling flows that were initially sufficiently fine-grained to allow uniform sediment distribution within the flow.

4.6.5. Description of outer-bend distal levee

The outer-bend distal levee is subdivided into a lower (DOB1) and upper package (DOB2; Fig. 4.12). The lower package (DOB1) is about 80 m thick and consists predominantly of fine-grained F2 strata interbedded with less common coarser grained F1 beds. Locally these strata are interbedded with contorted F2 units (i.e. F3b – slump deposits) up to 3 m-thick. The sandy bases of F2 beds in DOB1 range from 8 cm to <1 cm thick, but are typically 1–2 cm. Initially beds of F2 appear to be continuous across the width of the outcrop (300 m wide) and show negligible lateral thinning. However, based upon five closely-spaced (~40 m apart) stratigraphic logs (see Fig. 4.12), it is noted that the number of beds varies between logs (specifically, bed number between logs 1 and 2 is 24, increases to 29 in log 3, decreases to 26 in log 4, but then increases to 31 in log 5), suggesting that within this small window some beds pinch out whereas others become initiated. Interbeds of F1

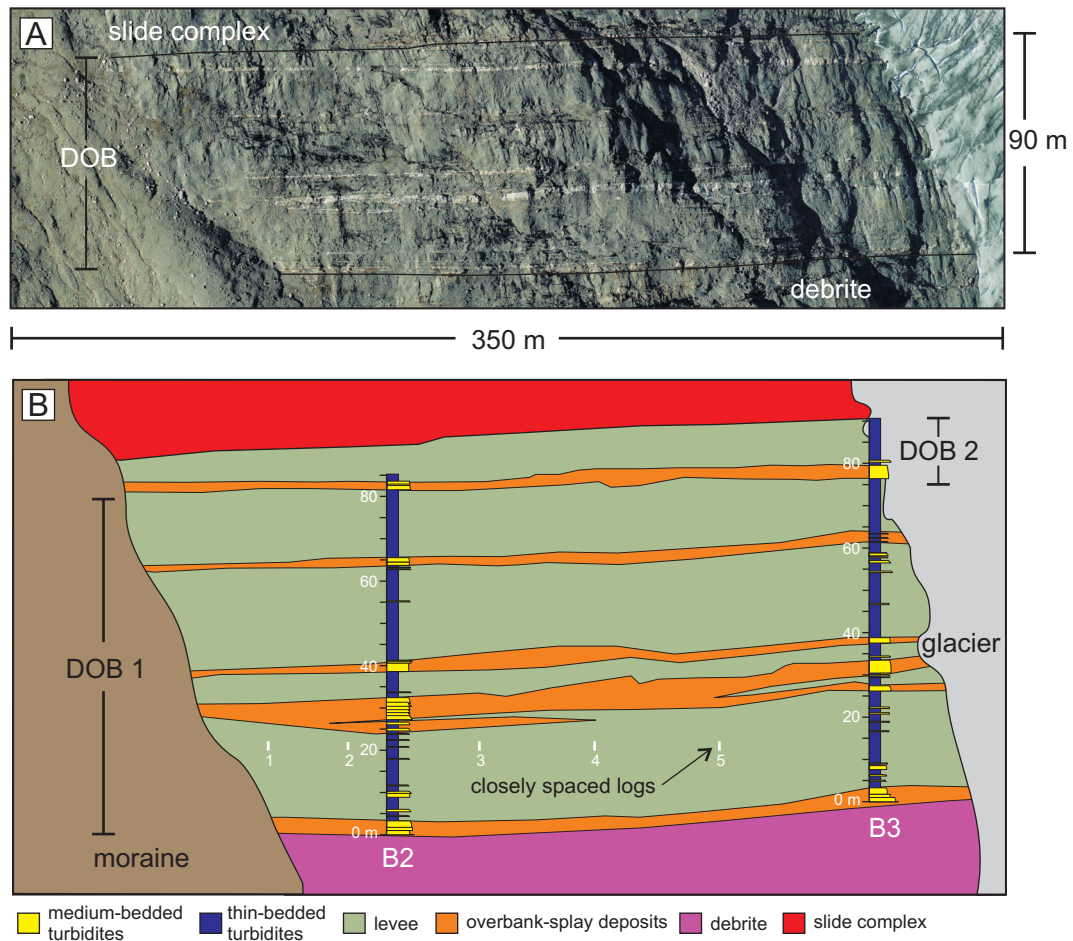


Fig. 4.12. Strata in distal outer-bend levee (DOB). A) Aerial photomosaic of DOB showing the contacts between the underlying debrite and overlying slide complex. B) Interpreted panel showing the distribution of overbank splay deposits and the contact between DOB1 and DOB2. Also included are two measured litho-stratigraphic logs and the location of the closely spaced logs used for lateral correlation (see Section 5.2.1).

strata range from 10–100 cm thick and average about 26 cm. Generally these beds thin away from the channel (i.e. toward glacier), although some thin and even pinch out toward the channel. There is no obvious upward thinning or fining trend within the lower 62 m of DOB1, although F1 beds become noticeably less common above about 45 m. Specifically, F1 strata account for 27% of the total number of event beds in the lower half but only 12% in the upper half. The uppermost 18 m of stratigraphy, on the other hand, exhibits a well developed upward thinning trend wherein average sandstone thickness decreases from 4 to less than 1 cm (Fig. 4.13). Collectively, sandstone bed thickness trends are recorded by the

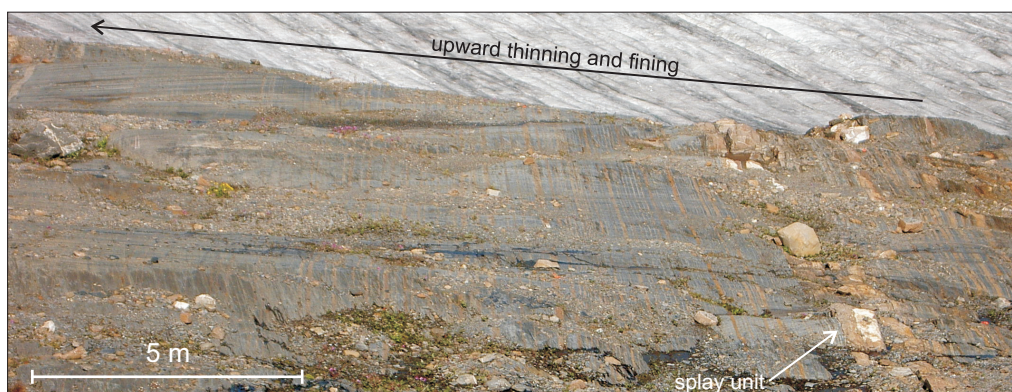


Fig. 4.13. Photograph showing the thinning- and fining-upward trend observed in the upper 18 m of the distal outer-bend levee (DOB1).

N:G, which in the lower half is about 0.44 and in the upper half averages about 0.29 (Fig. 4.14). Note also that the upward thinning trend in the uppermost 18 m is mirrored by a gradual upward decrease in N:G from about 0.4 to 0.08. Furthermore, F1 and F2 strata are intercalated with 2–5 m thick amalgamated units made up of distinctively thicker, coarser amalgamated F1 strata that lack intercalated siltstone layers. These units occur along four horizons: at the base of the levee succession and at about 28 m, 40 m, and 65 m above the base. Amalgamated units are laterally continuous along their entire exposed length, which is up to 300 m. Although the units are laterally extensive they generally thin appreciably towards the glacier. Internally, beds are typically laterally discontinuous, commonly being eroded by an overlying bed in less than 100 m. The uppermost package (DOB2) is bounded below by DOB1 and above by an erosion surface that marks the base of a thick slide deposit. This uppermost levee package is up to 12 m thick and consists of intercalated F2 and F1 strata with an average N:G of 0.48. In addition, a 2–3 m thick amalgamated unit occurs 2 m above the base.

The proximal and distal levee of ICC3 is separated by an up to 250 m-wide moraine and, therefore, direct correlation is impossible. Notwithstanding, the majority of F1 beds in

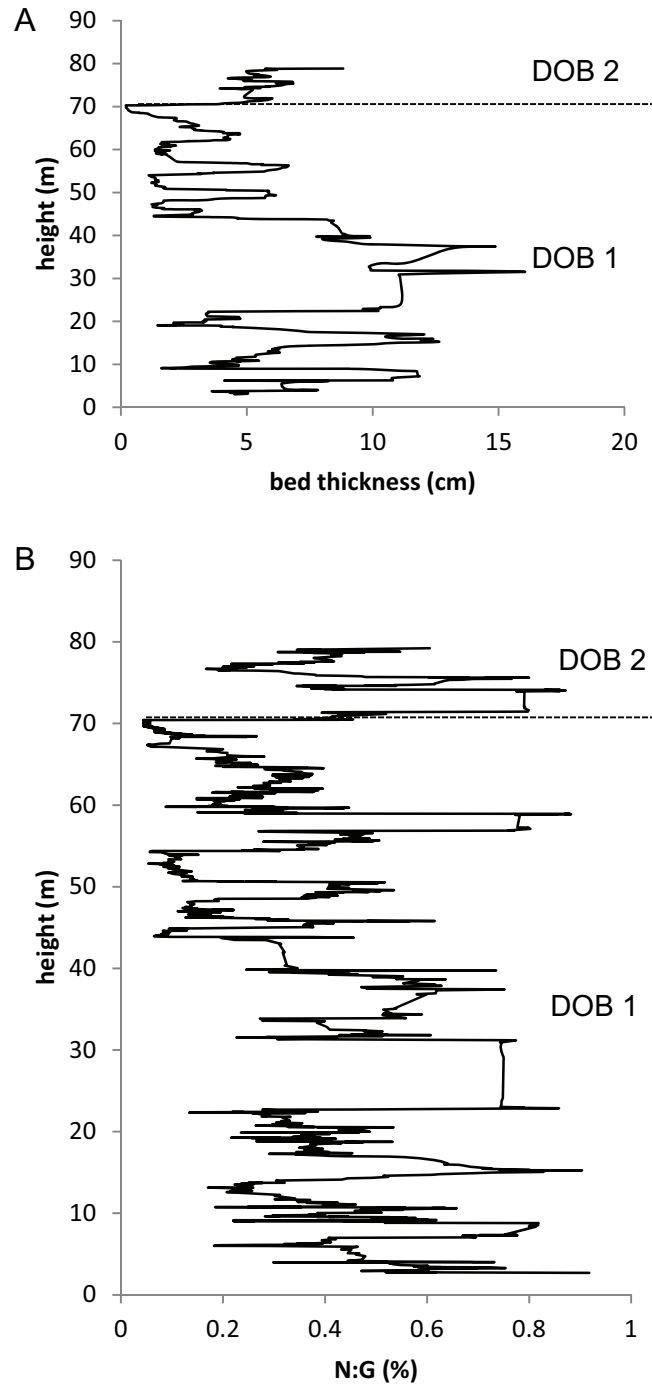


Fig. 4.14. Data displayed from 531 beds (including siltstone beds) measured in Log B2 showing upward trends in N:G. There is an abrupt decrease in average N:G above about 45 m. This coincides with a decrease in the frequency of F1 strata. In the upper 18 m of DOB1 N:G progressively decreases upward and coincides with a well developed thinning-upward trend (see Fig. 12). Note that a splay unit at about 65 m (stratigraphic height) causes a local spike in N:G. This plot uses a 15-bed moving average.

the proximal levee thin and fine to beds of F2 over less than 300 m laterally thus implying that the majority of F1 strata in the distal levee are not continuous beneath the moraine. In contrast, some if not many F2 beds in the proximal levee probably correlate across the moraine to F2 beds in the distal levee. Nevertheless, the fact that some F2 beds pinch out towards the moraine in both the proximal and distal levees suggests that at least some of these beds are physically discontinuous beneath the moraine.

4.6.6. Interpretation of outer-bend distal levee

Facies 2 (F2) beds in the distal levee that are interpreted to correlate with F1 beds in the proximal levee are inferred to represent continuous deposition from overspilling flows that progressively diminished in competence laterally away from the channel. As a consequence, deposits from these flows became progressively thinner and finer-grained. Rare F1 beds that are interpreted to extend from the proximal into the distal levee represent deposition from anomalously large magnitude overbanking flows that were able to spread coarser sediment over a significantly wider area than the more common flows that overspilled ICC3. Beds that begin in the distal levee with no proximal equivalent are interpreted to have been deposited by flows that locally bypassed the proximal levee. Alternatively, the minority that pinch out towards ICC3 may have been deposited by flows that overbanked an adjacent but unknown channel.

The conditions necessary to induce bypass along the backside of levees are shown theoretically using Middleton's (1966) equation for the velocity (u) of a turbidity current wherein particles are uniformly dispersed by

$$u^2 = g'/f_0 + f_i)d \cdot \sin\alpha \quad (4.1)$$

where

$$g' = (\Delta\rho/\rho + \Delta\rho) \quad (4.2)$$

where g is the gravitational constant, ρ the density of seawater, $\Delta\rho$ the density difference between seawater and the turbidity current, f_o the basal friction factor, f_i the interfacial friction, d the height of the turbidity current, and α the slope. Using this equation Hiscott et al. (1997) discussed how the velocity of an overspilling turbidity current would remain constant or even increase upon rapid lateral thinning so long as $d \cdot \sin\alpha$ remains constant or increases (i.e. an inverse relationship between d and $\sin\alpha$). Therefore, if the slope along the backside of the levee is sufficiently greater than the channel gradient then flow speed could be maintained or even accelerate down the backside of the levee. Seismic profiles from the Amazon Fan, Indus Fan, and deep-water part of the Niger Delta demonstrate that slopes along the backside of levees are up to 5° whereas slopes of the related channels are commonly an order of magnitude less (Pirmez, 1994; Piper and Normark, 2001, Deptuck et al., 2003). Therefore, levee backside slope is commonly significantly greater than channel slope and, accordingly, the bypass of flows down the levee backside is likely a common condition. In ICC3 the occurrence of slump deposits (F3b) in the distal levee suggests that at times a steep slope developed along the backside of the levee, therefore providing ideal conditions for flows to bypass locally. Dilute overspilling flows that bypassed the proximal levee locally deposited F2 strata in the distal levee, whereas sandier more concentrated flows collapsed more rapidly and deposited F1 beds.

Amalgamated F1 sandstone units represent deposition on the distal levee from successive large magnitude sandy flows, interpreted to have occurred due to both typical overbank-bypassing processes (i.e. overbank splays) and breaches in the levee (i.e. crevasse

channels and splays). Preferential deposition on the distal levee progressively reduced the slope along the backside of the levee locally, which in turn reduced later bypass. This negative feedback relationship has been reported along the backside of levees in a number of modern fluvial systems (e.g. Cazanacli and Smith, 1998; Brierley et al., 1997).

The uppermost distal levee package (DOB2) may represent more proximal levee deposition associated with the lateral migration of ICC3. This would explain the higher N:G in comparison with the upper half of the underlying package. Alternatively, the uppermost package may represent the proximal levee associated with the younger overlying channel system (see Fig. 4.4). Due to the local moraine cover neither scenario can be substantiated and, therefore, this unit is not considered in the evolution of ICC3.

4.6.7. Description of inner-bend levee

Deposits along the inner-bend levee of ICC3 are divided into lower and upper units, respectively about 17 m and 40 m thick (Fig. 4.4). The lower inner-bend levee (IB1) overlies an up to 60 m thick mud-rich debrite and consists of strata that are truncated by channel-fill deposits. Unlike in the outer-bend, the inner-bend channel margin is not terraced but instead shows a more consistent dip towards the channel axis of about 2–4° (Fig. 4.15A). The lower inner-bend levee unit (IB1) consists largely of F2 strata that, like those on the outer bend, show negligible lateral thinning or fining. Thicker F1 beds are observed in the lower 5 m of the section most proximal to the channel (Fig. 4.15B), but they are less common than in proximal outer-bend levee deposits. In contrast to the proximal outer-bend levee where there is a relatively abrupt vertical transition from F1 to F2 strata, the inner-bend levee is characterized by a more gradual upward-thinning trend (Fig. 4.15B). Average bed thicknesses are 25 cm near the base and less than 1 cm at the top (Fig. 4.15B). Moreover, F1

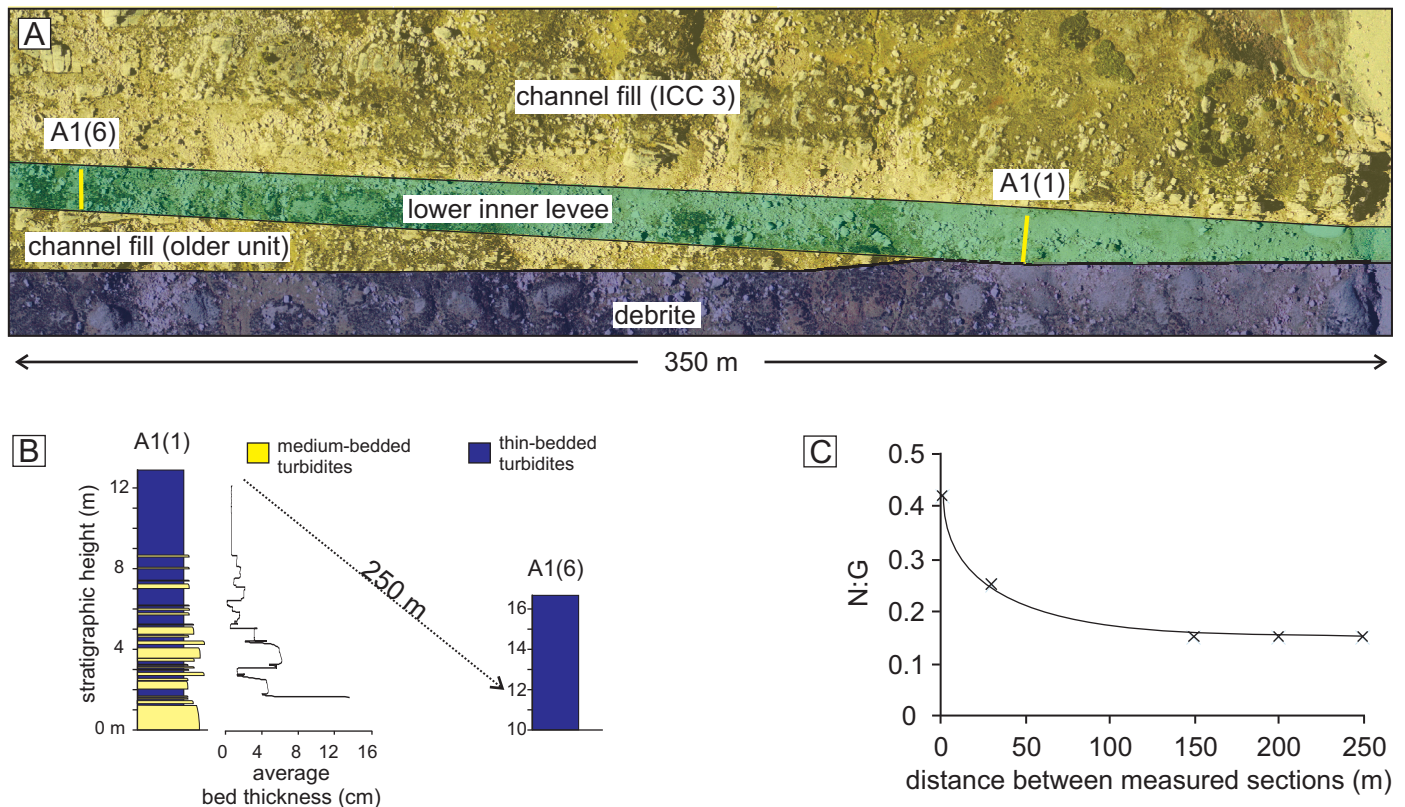


Fig. 4.15. Strata in the lower inner-bend levee (IB1). A) Aerial photomosaic of the lower inner-bend of ICC3 showing locations of measured litho-stratigraphic logs in (B). C) Trend in N:G laterally away from channel margin based on five measured logs (only the

beds at the base of the section thin away from the channel margin over a distance of less than 100 m from 23 cm to less than 10 cm thick, before onlapping the surface of an older channel unit (Fig. 4.15A). As a result of this lateral thinning, N:G decreases away from the channel from 0.40 to 0.15, and the majority of the decrease takes place within the first 50 m (Fig. 4.15C).

Similar to IB1, the upper inner-bend levee (IB2) is composed mostly of F2 strata. The channel geometry associated with IB2, however, differs markedly from that of IB1. Here some thick-bedded, coarse-grained, amalgamated channel-fill deposits in the axial channel grade laterally over about 140 m into fine, thinly-bedded turbidites that become progressively interbedded with other F2 inner-bend levee strata (Fig. 4.16). More

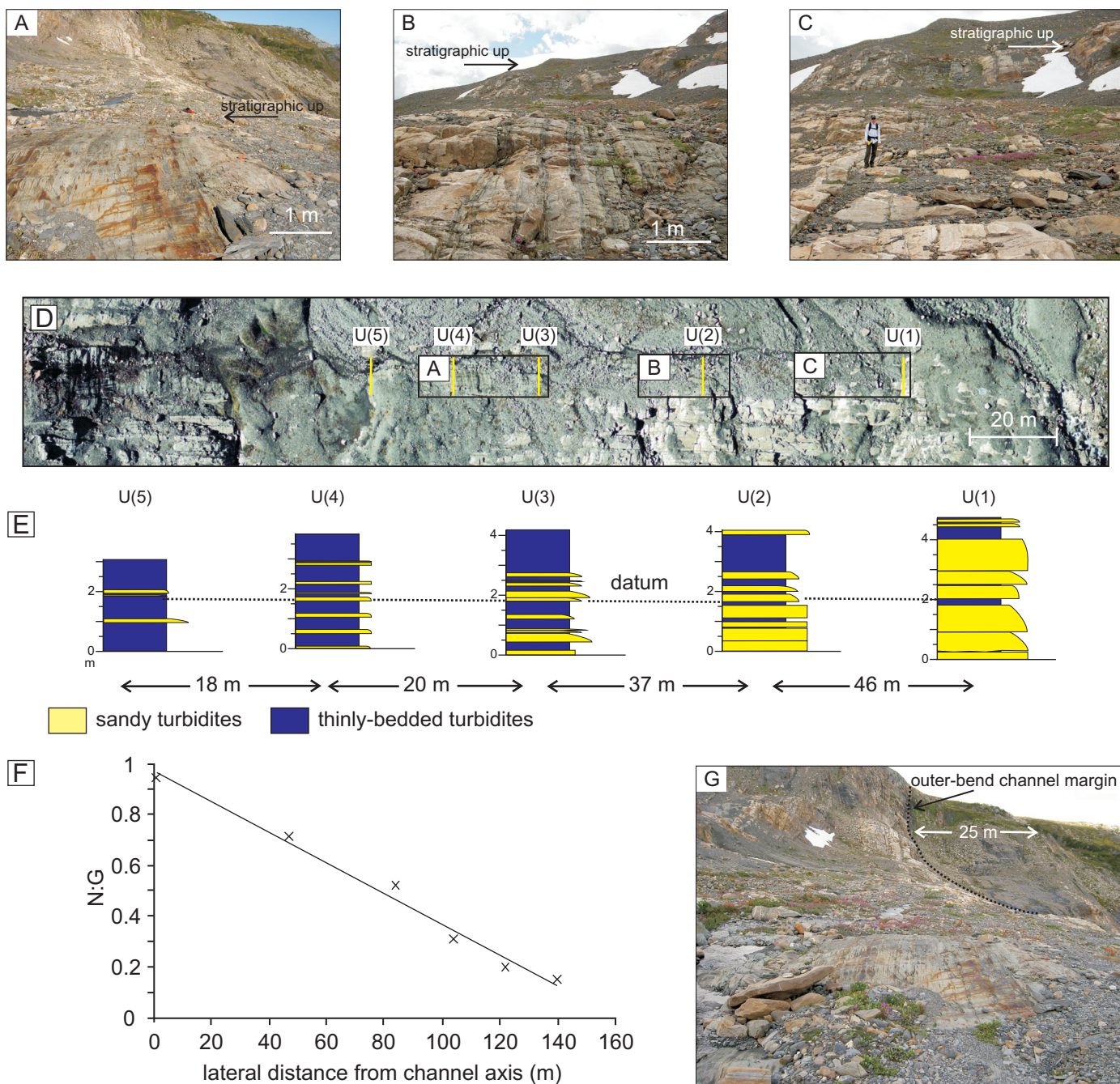


Fig. 4.16. Strata in the upper inner-bend levee (IB2). Photographs illustrate the lateral progression from amalgamated channel fill (C) into interbedded levee strata (B and A). D) Shows location of (A, B, and C) in addition to measured litho-stratigraphic logs in (E). F) Lateral trend in N:G from channel through inner-bend levee. G) Photograph from upper inner-bend levee towards outer-bend channel margin. Note levee strata in foreground and channel-fill deposits in background along the same stratigraphic horizon.

specifically, amalgamated channel-fill strata up to 2 m thick thin laterally into either a single or multiple F2 beds, forming a distinctive interfingering pattern (Figs. 4.16A–E). Net-to-gross (N:G) values reflect these lateral trends, decreasing from 0.9 immediately adjacent where channel-fill strata begin to split to 0.15 140 m laterally away from the channel, after which N:G remains constant (Fig. 4.16F).

4.6.8. Interpretation of inner-bend levee

The two end member stratal packages that form on the inner-bend (i.e. IB1 vs. IB2) are interpreted to be controlled completely by topography. Like the outer-bend levee, relief along the lower inner-bend levee (IB1) was built up by the overspill of thick channel flows that bypassed most sediment downslope. Initially, when the channel was weakly confined, the lower, sandy parts of flows overbanked and deposited medium-bedded F1 turbidites on the inner bank. These earliest deposits provided initial confinement that helped constrain the lower part of subsequent throughgoing channel flows within the channel, allowing only the more dilute, finer-grained upper part of flows to spill across the overbank area. The gradual upward thinning-fining trend observed in IB1 suggests that levee relief increased progressively. In contrast to outer-bend levee packages where F1 strata occur as high as 12 m above the base of the package, F1 strata in IB1 only occur in the lower 5 m. This implies that significantly less relief was needed along the inner bank levee to constrain the sandy, lower parts of flows within the channel compared to the outer bend. Accordingly, flows that overspilled on the inner-bend levee were generally smaller and lower energy compared to those that overspilled on the outer bend, resulting in the deposition of an overall finer-grained and thinner-bedded succession. This asymmetry in flow conditions is interpreted to have been linked to centrifugal effects of flows rounding a channel bend and becoming

superelevated along the outer bend. Subdued erosion along the inner-bend margin created a surface that dips gently ($\sim 4^\circ$) towards the channel, and on to which channel-fill strata onlap. Unlike the outer-bend, the inner-bend margin lacks terraced segments, suggesting that erosion associated with abrupt shifts of the thalweg did not affect it.

The predominantly fine-grained, thin-bedded style of deposition observed in the upper inner bend levee (IB2) suggests deposition from dilute overspilling suspensions. Unlike IB1, topography along IB2 was comparatively more subdued either because overspilling was less extensive, possibly related to a reduction in flow size, or alternatively the time between predominantly bypassing channel flows and depositional ones was too short to allow the levee to build up to appreciable height. The lack of significant levee relief on IB2 enabled throughgoing channel flows to expand laterally over the adjacent levee area without becoming detached from the main-channel flow. This resulted in stratal continuity between channel-fill and levee strata wherein beds thin and fine laterally and become progressively intercalated with other fine, thin-bedded strata of the levee. The interfingering pattern created by laterally thinning channel beds resembles observations from the Permian Lainsburg Formation in the Karoo Basin, South Africa (Grecula et al., 2003) and the Upper Carboniferous Ross Formation in the Shannon Basin, western Ireland (Lien et al., 2003). Both of those examples are interpreted to have resulted from low topographic relief present along the channel margin.

Similar to outer-bend deposits, the lateral thinning trend of thicker levee deposits in IB1 and IB2 are associated with overspilling flows that were initially well stratified and became progressively more uniformly distributed laterally as the coarser fraction transferred to the bedload layer. Deposition from such flows resulted in beds that initially thin rapidly laterally, developing a more constant thickness as the flows became more uniformly

distributed away from the channel. Unlike F1 strata in the outer-bend area, thicker-bedded deposits in the inner-bend levee lack a lateral thickening trend. This suggests that overspilling flows along the inner-bend had insufficient inertia to move the zone of maximum deposition appreciably away from the channel margin. Facies 2 (F2) strata observed in the inner-bend levee, like in the outer-bend strata, are interpreted to have been deposited by overspilling flows that were initially uniformly distributed and, accordingly, deposited beds that show negligible change away from the channel with respect to grain size and thickness. The laterally thinning part of channel-fill strata represent deposition from flows that expand laterally and deposited on the levee.

4.7. Levee evolution

Isaac Channel Complex 3 (ICC3) presents a rare opportunity to document the differences of levee strata on opposite sides of a sinuous channel. This section assembles the data presented earlier to illustrate the stratigraphic evolution of ICC3 and to highlight the asymmetry of outer- and inner-bend levee deposits. Although similar to the channel-levee model for ICC3 presented by Navarro et al. (2007a, b), this contribution focuses mainly on the evolution of the channel-bounding levees, and less on the channel fill.

The inception of ICC3 is interpreted to have been controlled by topography along the top of a thick debrite that caps an areally extensive mass transport deposit (Fig. 4.17A). Rugosity along the top of debrites and localization of overlying channel-levee systems has been reported from other deep-marine systems, including the deep-water part of the Niger Delta (Samuel et al., 2003), Indus Fan (Deptuck et al., 2003), and the ancient Tres Pasos Formation (Armitage et al., 2009). Following inception, the evolution of ICC3 occurred

during three major episodes of channel-levee growth (channel-levee units 1–3), which are all recorded in outer bend proximal levee packages POB1 to POB3 (Figs. 4.8A, B). Only the correlative inner-bend levee deposits of packages POB1 and POB3 are exposed, which respectively are IB1 and IB2. The inner-bend equivalent of POB2 is covered by vegetation. As for the channel, C1 is interpreted to be genetically related to POB1 and IB1, C2 to POB2, and C3 and C4 to POB3 and IB2. Individual episodes of channel-levee are characterized by three successive stages of levee growth and channel fill.

Correlation of the various levee packages and channel-fill units are based primarily on their spatial relationship to one another and their position relative to the stratigraphic datum, which is the top of the underlying debrite. For example, IB1 and POB1 are assumed to be correlative because they both immediately overlie the debrite. Moreover, stratigraphic trends, specifically the upward fining and thinning of strata in IB1, POB1, and C1 suggests a genetic relationship. Accordingly, the majority of the proposed correlations are inferred and not observed directly. An exception is the correlation between IB2 and the uppermost channel-fill units C3 and C4, where there is stratal continuity between units.

4.7.1. Development of Channel-Levee Unit 1 (CLU1): low levee relief (Stage 1)

Stage 1 of the first channel-levee unit (CLU1) began immediately after the inception of ICC3, at which point the channel is estimated to have been a few hundred meters wide and confined by topographic highs along the debrite surface no more than a few meters thick. This weakly confined system allowed the lower, coarse-grained part of relatively large magnitude turbidity currents to spill out of the channel, resulting in the deposition of medium- to thick-bedded, coarse- to fine-grained turbidites (F1) along both sides of the channel (Fig. 4.17B). These thicker-bedded deposits are recorded in the lower part of IB1

and POB1. The greater abundance and thickness of F1 strata on the outer-bend levee suggests that fluid and sediment inertia effects (i.e. superelevation and run-up) enhanced deposition there. Additionally, Straub et al. (2008) attributed the preferential emplacement of coarser grains on outer-bend levees not only to superelevation but also to “runup” forces. Runup occurs when kinetic energy enables particles to escape the confines of the channel. The coarse-grained nature of many beds within the proximal levee suggests that runup probably also influenced transport and deposition on to the outer levee.

The abundance of thicker-bedded strata along the outer-bend channel margin created an asymmetry in levee height. If it is assumed that each turbidite on the outer bend has an inner bend equivalent that was deposited contemporaneously during a single overspilling event, then it is possible to quantify the difference in levee height at any given time. For example, the lower part of POB1, interpreted to have accumulated entirely by the end of Stage 1, consists of 64 beds yielding a total thickness of 10.5 m. Accumulation of the first 64 beds in the lower inner-bend levee (IB1), in contrast, is 5 m thick. This implies that after the first 64 overspilling events of ICC3 the outer-bend levee was about two times higher than its inner-bend counterpart.

During this early stage of levee growth, flows in the channel are interpreted to have bypassed most of their sediment load basinward leaving behind coarse-grained lag deposits and thin-bedded, fine-grained tail-of-flow turbidites (Navarro et al., 2007b). Bypassed flows continuously scoured outer-bend levee deposits as the channel thalweg shifted laterally. This modified the channel margin facing into the channel, which locally dips up to 25°. Other researchers working on different turbidite systems have inferred that as levee height increases, so too does the slope along the levee backside (Straub and Mohrig, 2008b; Kane et al., 2010b) and that this can cause overspill flows to become supercritical, leading to the

formation of sediment waves (Migeon et al., 2001; Lee et al., 2002; Normark et al., 2002; Migeon et al., 2004) or large-scale scour features (Fildani et al., 2006). There is no clear evidence of such large-scale features in ICC3; however, it is interpreted that slope along the backside of the levee promoted bypass, albeit limited, and deposition on the distal levee. At the same time, but on the inner-bend levee, diminished flow height and strength resulted in reduced erosion forming a shallow dipping (up to 4°) ramp towards the channel axis.

4.7.2. Development of Channel-Levee Unit 1 (CLUI): high levee relief (Stage 2)

Increased levee relief during Stage 1 confined the approximately 500 m wide channel that was bound by 10.5 m high outer- and 5 m high inner-bend levees. This added relief along the outer-bend margin is interpreted to represent a threshold height, over which only the upper, more dilute part of relatively large magnitude throughgoing channel flows overspilled (i.e. flow stripping). Accordingly, the lower part of POB1 is abruptly overlain by predominantly thin-bedded, fine-grained sandstone turbidites (F2) whose deposition marked the beginning of Stage 2 (Fig. 4.17C). At the same time, deposition along the inner bend, which during Stage 1 had experienced comparatively minor levee aggradation, was marked by even thinner and finer-grained F2 strata emplaced by increasingly more dilute overspill flows. The gradual upward-thinning trend observed in the lower inner-bend levee (IB1) demonstrates that effects of levee aggradation on overspilling flow dynamics were not as dramatic compared to the outer-bend levee, where an abrupt facies transition occurs between Stage 1 and Stage 2 deposition.

Minor channel filling likely occurred throughout Stage 2; however, it is assumed that flow in the channel was largely in a state of bypass and, for the most part, continued to erode and modify the channel margins. Growth of the outer-bend levee is also interpreted to have

continued to outpace its inner-bend counterpart throughout Stage 2. Differences in levee relief along opposing channel margins is estimated by assuming that the upper part of levee package POB1, which consists of 90 beds and is about 7 m thick, accumulated largely by the latter part of Stage 2. Accordingly, the total thickness of levee package POB1 (upper + lower parts), which consists of 154 beds and a thickness of 17.5 m, is estimated to approximate levee relief along the outer bend towards the end of Stage 2. The first 154 beds of IB1 (including the 64 beds deposited during Stage 1), in contrast, have a cumulative thickness of 9.5 m. This implies that the outer-bend levee was slightly less than two times higher than its inner-bend counterpart by the end of Stage 2 suggesting that levee asymmetry changed little during the first two stages of channel-levee growth. Also, erosion along the outer and inner bend approximately doubled the width of the channel; however, due to increased levee relief the width to height ratio of the channel did not change significantly during Stage 2.

4.7.3. Development of Channel-Levee Unit 1 (CLU1): onset of channel-filling (Stage 3)

Stage 3 is interpreted to have been characterized by the onset of channel deposition that significantly outpaced levee aggradation and progressively reduced channel confinement (Fig. 4.17D). Also, as a result of increased levee relief developed during Stages 1 and 2, it is interpreted that the slope along the backside of the outer-bend levee was highest during the early part of Stage 3. The occurrence of slump deposits in the distal levee supports the existence of a steep slope along the backside of the levee (Hickson and Lowe, 2002; Hubbard et al., 2008). Therefore, it became increasingly common for overbanking flows to bypass the proximal outer-bend levee and deposit on the distal levee. Deposition from these bypassing flows is recorded in DOB1 as F1 and F2 strata. Furthermore, implied by these bypassing conditions is that the 17.5 m of levee thickness at the end of Stage 2 remained

relatively constant throughout Stage 3 (i.e. negligible deposition on the proximal outer-bend levee). Deposition on the inner-bend during Stage 3, in contrast, is interpreted to have continued unabated because levee relief and the style of deposition remained insufficient to locally develop a steep slope along the backside of the levee and change transport and depositional conditions.

Continued deposition on the inner-bend levee is interpreted to have coincided with mainly bypassing flows on the proximal outer-bend levee, which progressively reduced the difference in levee height along opposing channel banks. The thickness of the inner-bend near the end of Stage 3 is estimated by the total thickness of IB1, which is about 17 m. This value, compared to the 17.5 m thick levee along the outer-bend, illustrates how drastically the difference in levee height along opposite sides of the channel decreased during Stage 3. Moreover, the interpretation of flows bypassing the proximal outer-bend levee is supported by the number of event beds that is recorded along opposite sides of the channel. For instance, IB1 consists of 367 event beds whereas only 159 are present in POB1 thus implying that more than half of the flows bypassed the proximal and deposited only on DOB1. Moreover, it is interpreted that preferential deposition on the distal levee gradually reduced relief between the distal and proximal levees, and in doing so reduced the slope along the levee backside eventually creating a “new” near horizontal levee surface area.

4.7.4. Development of Channel-Levee Unit 2 (CLU2)

During the latter stage of CLU1 it is interpreted that flows within the channel transitioned from bypassing to depositional, possibly related to allocyclic controls that temporally diminished flow magnitude and/or increased the ratio of sand to mud. If correct, then it follows that some time later allocyclic controls would again progressively increase the

magnitude or sediment content of throughgoing channel flows. As a consequence, flows within the channel that previously were depositional would again become non-depositional (bypass) to erosive. These higher magnitude flows reoccupied the once again weakly confined channel of ICC3 marking the onset of Stage 1 of CLU2 (Fig. 4.17E). In addition these energetic flows, which locally eroded up to 12 m deep, scoured the outer bend and caused it to migrate laterally about 50 m forming a “flat” along the channel margin. Overall, Stages 1–3 of CLU2 (Figs. 4.17E–G) are similar to those of CLU1, where throughgoing channel flows constantly eroded and readjusted the channel-levee margin as levee relief increased locally, thereby forming surfaces that are inclined up to 25° toward the channel axis. The result is a “flat and step” (c.f. Eschard et al., 2003) morphology as observed along the outer channel bank. Deposition of CLU2 culminated in the generation of a “new” relatively flat proximal levee surface for the onset of the next channel-levee unit. Differences in levee heights that existed on opposite sides of the channel throughout the development of CLU2 could not be documented due to local cover by vegetation on the inner bend.

4.7.5. Development of Channel-Levee Unit 3 (CLU3)

Stages 1 and 2 of CLU3 were similar to those in the underlying unit (CLU2), where early throughgoing channel flows eroded up to 10 m into CLU2 deposits and added another flat about 100 m wide to the “flat and step” morphology of the channel margin (Figs. 4.17H–I). The channel-filling stage of CLU3 (Stage 3), however, differs significantly from previous channel-levee units. Here, channel-fill strata overlapped the outer-bend channel margin whereas relief along the inner bend of the channel was reduced to the point where it allowed larger than average channel flows to expand laterally over the levee without becoming detached from the main channel flow. These flows deposited thick-bedded, coarse-grained

sandstone turbidites that grade laterally into thinly-bedded F2 strata along the inner-bend channel margin and onto the levee (see IB2 description) forming a distinctive interfingering pattern (Figs. 4.16, 4.17J). This suggests that the later part of Stage 3 was characterized by alternating episodes of minor levee growth and channel filling driven by fluctuating flow magnitudes.

Based on interpretations of the inner-bend levee, it is likely that the levee along the outer-bend was similarly characterized by alternating episodes of levee growth and channel deposition. Unlike the inner-bend levee, however, relief remained sufficiently high and accordingly channel-fill strata continued to terminate abruptly against the channel margin. Also, depending on the steepness on the levee backside many overspilling flows likely bypassed the proximal levee throughout Stage 3. These flows are recorded by strata in the upper 30 m of DOB1 that fine and thin upward. This trend, combined with the upward fining and thinning of the equivalent channel-fill strata (i.e. C4), suggests a progressive shutting down of the sediment delivery system.

The lowermost part of the inner-bend levee is not exposed and, therefore, an accurate comparison of the differences in levee heights that existed early in the development of CLU3 (i.e. during Stages 1 and 2) could not be made. However, like the underlying channel-levee units, it is interpreted that the lower part of POB3, which is 7 m thick, represents levee relief that is comparable to that which existed along the outer bend of the channel at the end of Stage 1. As suggested earlier, this relief represents the threshold height that constrained the lower, sandy parts of flows to the channel allowing only the dilute upper parts of flows to overflow. Moreover, on-going erosion along the base of the channel further increased levee relief and, accordingly, 7 m is a minimum estimate of the threshold height developed during Stage 1. Also, it is assumed that during Stage 2 flows within the channel continued to bypass

sediment basinward as supported by local scour in C3 and erosion along the margins. This suggests that any overbank deposition during this time would only increase levee relief. Therefore, the lack of sandy, medium- to thick-bedded F1 strata in the upper part of POB3 suggests that the minimum threshold height was maintained throughout Stages 2 and 3. At the same time, but on the inner bend, levee relief developed during Stage 3 is estimated based on the distinctive interfingering geometry between channel-fill and levee strata. The thin-bedded deposits that occur between laterally thinning channel-fill deposits represent levee relief that existed before channel filling events that spilled out onto the levee (Fig. 4.17G). Where exposed, intervening thin-bedded units between laterally thinning channel fill deposits are up to 2 m, which is considered to represent the maximum amount of levee relief that existed on the inner bend during Stage 3. Therefore, the outer-bend levee was at least 3 times higher than its inner bend counterpart throughout most of Stage 3.

4.8. Summary of ICC3 levee growth

The three channel-levee units (CLU1–CLU3) of ICC3 stack in a vertically-aggrading offset pattern (Fig. 4.4) associated with the preferential lateral and upward shift of the axis toward the outer bank of the channel. Erosion associated with the lateral and then upward shift of the channel axis produced a terraced (flat and step) morphology along the outer-bend channel margin that sharply separates the channel-fill and adjacent levee. This geometry supports the interpretation that ICC3 was formed by at least three major episodes of channel-levee growth. Each episode of growth was characterized by a similar 3-stage evolution of levee development. These three stages are summarized as follows:

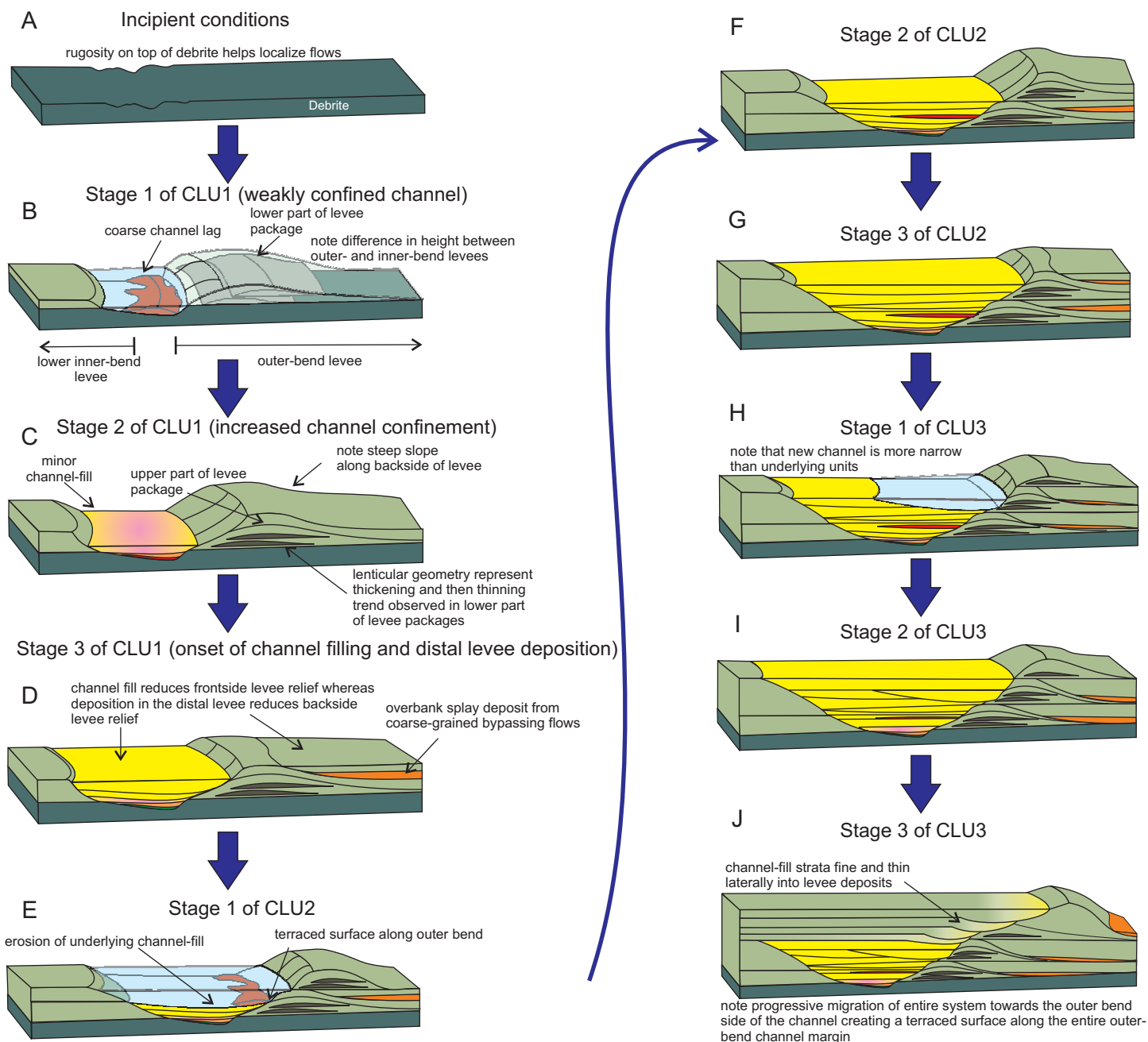


Fig. 4.17. Depositional model of ICC3 consists of three major episodes of channel-levee growth (CLU1 – CLU3), each of which are characterized by three similar growth stages (model not to scale). A) Rugosity along the top of this unit helped focus subsequent flows that ultimately localized and initiated ICC3. B) Stage 1 of CLU1 – low channel confinement and superelevation of turbidity currents caused preferential deposition of thicker-bedded and coarser-grained strata along the outer-bend resulting in the outer-bend being twice the height of its inner bend counterpart. At this time the channel was largely in a state of bypass. C) Stage 2 of CLU1 – increased channel confinement enabled only the upper, more dilute fraction of flows to overspill, resulting in the deposition of thinly-bedded turbidites over both inner- and outer-bend levees. D) Stage 3 of CLU1 – is marked by the onset of major channel filling with flow bypass on the outer-bend margin and preferential deposition on the outer-bend distal levee. In addition, continued deposition on the inner-bend levee reduced the height difference between the outer- and inner-bend levees. A consequence of intense channel filling was a reduction in the amount of levee relief enabling subsequent high energy flows to once again spill out of the channel and begin

4.8.1. Stage 1: Aggradation and channel by-pass

During Stage 1 the channel was weakly confined, and as a consequence, superelevation of turbidity currents as they rounded the bend aided in overspill of the lower, higher-density portion of flows along the outer channel-bend. This resulted in the deposition of sand-rich deposits (F1) on the outer-bend levee. Superelevation of the flow on the outer bend resulted in comparatively diminished flow on the inner bend side of the channel and deposition of thinly-bedded turbidites (F2). Preferential deposition of thicker-bedded, coarser deposits caused the outer-bend levee to be about two times higher than its inner-bend counterpart. During Stage 1 flows confined within the channel were largely bypassing sediment further downslope and deposited only lag deposits, including mudstone-clast breccias, pebble conglomerate, and tail-of-flow thinly-bedded turbidites in the axis of the channel. In addition, these bypassing flows were largely responsible for sculpting the channel form by eroding along the outer-bend channel margin.

4.8.2. Stage 2: Continued levee growth

Due to levee growth during Stage 1, generally only the upper, more dilute fraction of stratified flows spilled over the channel margins and deposited thinly-bedded turbidites (F2) on both the outer and inner bend levees. However, as a result of inertial effects, outer-bend levee aggradation continued to outpace its inner-bend counterpart. A consequence of increasing levee thickness along the outer bend was the development of a steep slope along the backside of the levee. Channel filling commenced during the latter part of Stage 2 but remained minor compared to levee growth. The outer-bend levee was also consistently about two times higher than its inner bend counterpart throughout Stage 2.

4.8.3. Stage 3: Channel aggradation/decreased levee growth

The early part of Stage 3 was marked by the onset of intense channel filling, likely related to a progressive decrease in flow magnitude and/or increase in sand:mud ratio. Initially channels filled with thick-bedded, normally graded and massive conglomerate and very coarse-grained sandstone. Stratigraphically-upward, thick amalgamated sandstone and interbedded thick-bedded coarse sandstone and thin-bedded turbidites became more common. Increased rates of channel aggradation coincided with diminished flow overbanking and a net reduction in levee relief. Also, it is likely that any overspilling flows that occurred along the outer-bend channel margin bypassed the proximal levee and deposited in the distal levee as a consequence of the inferred steep slope along the backside of the levee following Stage 2. This increase in distal levee deposition significantly reduced the differential topography between the proximal and distal levees, and formed a near horizontal surface area onto which strata deposited during Stage 1 of the next levee package.

Deposition on the inner bend during Stage 2, in contrast, is interpreted to have prevented the development of steep backside slopes locally and, as a result, overspilling flows did not commonly bypass the proximal inner-bend levee but instead continued to build levee relief. Accordingly, this reduced the height difference between the inner- and outer-bend levees during the development of CLU1 and CLU2. Stage 3 of CLU3, on the other hand, was characterized by alternating episodes of levee growth and channel filling. This allowed the outer-bend levee of CLU3 to remain at least three times higher than the inner bend throughout most of Stage 3.

4.8.4. Inception of next channel-levee unit

During the early part of Stage 1 of the next channel-levee unit, flows in the channel

transitioned from depositional to non-depositional/erosive as a result of a progressive increase in flow magnitude, which is interpreted to have been linked to allocyclic controls. These erosive flows scoured along the outer-bend channel margin shifting it laterally up to 100 m and created a distinctive terraced surface. As levees continued to aggrade throughout Stage 1 and Stage 2 channel flows become more localized and readjusted the channel facing side of the levee locally creating steeply dipping surfaces. Aggradation and lateral migration of ICC3 as a whole is also recorded by the vertically stacked and laterally offset pattern of POB1–POB3 associated with the preferential lateral shifting of the thalweg (Fig. 4.8B). This relationship is especially noticeable when comparing the lower parts of POB2 to POB3.

4.9. Discussion

4.9.1. Comparison with other deep-water channel systems

Although several studies provide criteria for recognizing levee deposits in the rock record (e.g. Kane and Hodgson, 2011; Campion et al., 2011), it is typically difficult to unequivocally assess the contact between levee and channel-fill deposits in the field. This is well exemplified by the numerous interpretations for channel systems in the Cerro Toro Formation, which include 1) channel-levee growth where fine-grained deposits are genetically linked with juxtaposed coarse-grained channel fills (Winn and Dott, 1979; Beaubouef, 2004; Hubbard et al., 2008), 2) cut and fill where channel flows erode into underlying, unrelated fine-grained deposits (Coleman, 2000), and 3) more complex evolution wherein the fine-grained interchannel and coarse-grained channel sediments were likely deposited concurrently by different flows or by the same flow that was decoupled upstream (Crane and Lowe, 2008). Similarly, alternative interpretations might fit the observations

from ICC3. This next section will compare the interpretation presented in this paper concerning the stratigraphic evolution of ICC3 with other common channel-levee models.

4.9.2. Cut and fill model

The erosive outer-bend channel margin of ICC3 superficially supports the cut and fill model (Coleman, 2000; Gardner and Borer, 2000), wherein flows responsible for finer-grained deposition are not associated with flows responsible for channel incision. However, in Castle Creek the terraced profile of one of the channel margins (outer bend), the in-step lateral migration of channel fill and levee facies (Fig. 4.8B), and divergent paleocurrent directions on either side of the channel, supports more strongly a genetic channel-levee relationship. Furthermore, on the other side of the channel (inner bend), channel fill strata in the upper part of ICC3 grade laterally into, and interfinger with, finer-grained levee deposits, clearly indicating their coeval nature (Fig. 4.16).

4.9.3. Large-scale build and fill model

The complete outer-bend levee of ICC3 has a thickness of about 80 m, making it comparable in thickness to large-scale channels such as those on the Amazon fan (Piper and Normark, 2001; Hiscott et al., 1997; Pirmez et al., 1997). Amazon fan levees commonly display an upward-thinning and -fining trend interpreted to be linked with increasing levee relief that causes flows to progressively overspill their more dilute upper part. The proximal outer-bend-levee in ICC3, in contrast, is not characterized by a single large-scale upward-thinning or -fining trend but instead is subdivided into three distinct levee packages. Moreover, fine sand rarely overspills onto levees higher than 25 m on the Amazon fan (Hiscott et al., 1997). In contrast, thick-bedded, up to coarse sandstone beds are common as

much as 40 m above the base of the levee in ICC3. This makes it unlikely that ICC3 represents a channel-levee system in which the levee built to appreciable height before channel filling began.

4.9.4. Multiple episodes of channel-levee growth model

This model involves multiple episodes of levee growth followed by channel fill on the decametre scale (e.g. Beaubouef, 2004), and is supported here by the consistent stacking of ICC3 levee packages in a direction that mirrors the lateral migration of the channel fill (Fig. 4.8B). Theoretically, channel filling is succeeded by abandonment and deposition of fine-grained strata (e.g. Eschard et al., 2003; Samuel et al., 2003). The lack of fine-grained deposits in channel-fill strata of ICC3 may be because 1) the channel system was never abandoned but instead alternated between periods of by-pass and coarse sediment deposition, or 2) intense erosion between the late and early stages of two channel-levee units erased any evidence of these strata. The latter hypothesis is supported by the presence of deeply scoured surfaces (up to 12 m deep) commonly overlain by thick mudstone breccia layers at the base of individual channel fills.

4.9.5. Sinuosity and lateral migration

The interpreted sinuous nature of ICC3 is supported by the greater levee height recorded along one side of the channel, which typically occurs along the outer-bend of sinuous channels (e.g. Deptuck et al., 2003; Posamentier, 2003; Straub et al., 2008). This does occur in straight segments of channels if flows are significantly influenced by the Coriolis effect (e.g. Gamberi and Marani, 2008; Cossu et al., 2010). In the case of ICC3, however, the preferential emplacement of generally coarser grained sediment along one side

of the channel suggest an association with the inertial effect of sediment-charged flows as they round a channel bend (Straub et al, 2008). Also, the step and flat geometry along the channel margin suggests a consistent directional shifting of the thalweg, which is a phenomenon commonly ascribed to meander loop expansion (c.f. Posamentier, 2003). It is especially difficult to estimate how the channel bend evolved in time; however the consistent asymmetry in the height of the levees on opposite sides of the channel, in addition to their contrasting internal stratal make up and architecture, suggests that the channel bend did not migrate so far downstream that the straight part of the channel cropped out in the study area.

4.9.6. Levee shape

In the literature, levee beds, and in particular thick levee beds, are commonly reported to exhibit a well developed lateral thinning trend (DeVries and Linholm, 1994; Beaubouef, 2004; Kane and Hodgson, 2011). Thickening- then thinning-trends are less commonly reported, but have, for example, been described from levees deposits in the Rosario Formation, Mexico (Kane et al., 2007). It is commonly assumed that the vertical stacking of laterally thinning levee strata give rise to the typical “gull-wing” levee morphology (Kane et al., 2007; Straub and Mohrig, 2008a; Kane et al., 2010b).

Based on the analysis of modern deep-marine levees, Skene et al. (2002) showed that levee thickness decayed exponentially away from the channel. Later, Kane et al. (2007, 2010b) suggested that (paleo)geographical position of the levee along the turbidite system was the primary control on levee shape, with levees on the continental slope and basin floor displaying power law and exponential decay, respectively. Based on theoretical arguments, Birman et al. (2009) proposed that levees showing power law profiles are associated with flows that have constant rates of fluid entrainment or sediment detrainment, whereas levees

associated with flows that experience negligible entrainment thin exponentially laterally. Lateral trends in ICC3 strata show both exponential and power law decay, in addition to logarithmic and linear decay. For example, the thinning trend (excluding initial thickening) of individual medium- and thick-bedded (F1) strata in the proximal outer-bend levee of ICC3, are equally well described by linear ($R^2 = 0.96$) and logarithmic ($R^2 = 0.95$) decay profiles. This lateral reduction in bed thickness is reflected in the lateral decrease in N:G, which is equally well described by logarithmic ($R^2 = 0.97$) and power law ($R^2 = 0.97$) decay functions. Moreover, the lateral trend in N:G of the lower inner-bend levee is best described by logarithmic decay ($R^2 = 0.99$), but a power law ($R^2 = 0.97$) relationship fits the data almost as well. In the upper inner-bend levee, on the other hand, N:G change is best described by a linear decay ($R^2 = 0.99$), only moderately by exponential decay ($R^2 = 0.92$), and poorly by logarithmic ($R^2 = 0.73$) and power law ($R^2 = 0.55$). It would, therefore, appear that lateral trends in channel margin levees at Castle Creek are not necessarily reliable indicators of (paleo)geographic location.

4.10. Conclusion

This paper documents the lithofacies and architecture of outer- and inner-bend levees associated with an up to 1.8 km wide by 80 m thick passive-margin, deep-water slope channel complex. The principal observations and interpretations are as follows.

- ICC3 levees consist of three main lithofacies: 1) thin- to medium-bedded, coarse to fine sandstone T_{a-d} turbidites (F1); 2) very thin- to thin-bedded, fine sandstone to siltstone T_{cd} turbidites (F2), and 3) chaotic overbank deposits (F3; debris flow and slump deposits). Facies 1 (F1) strata initially thicken over about 120 m (sometimes to

more than ten times their original thickness) and then thin laterally and grade into F2 beds over the next 180 m away from the channel. Thereafter beds show little lateral change in thickness or grain size. Facies 2 (F2) strata not associated with the thinning of F1 strata also show little change in grain size or bed thickness laterally.

- Along the outer-bend channel margin, channel-fill strata are erosively juxtaposed on levee strata. The proximal outer-bend levee is subdivided vertically into three stratal packages, each of which is subdivided into a lower (intercalated F1 and F2 strata) and an upper (predominantly F2 strata) part. The contact between these parts is clearly marked by an abrupt decrease in average bed thickness and sandstone content. The net to gross ratio (N:G) is up to 0.68 in the proximal levee and decreases over several hundreds of meters to about 0.35 in the distal levee, an area dominated by mudstone-rich F2 strata.
- The lower part of the channel-fill onlaps the predominantly fine-grained, inner-bend levee wherein N:G decreases over about 100 m from 0.4 to 0.15. Strata in the upper part of the channel, in contrast, grade laterally over 140 m into fine, thinly-bedded turbidites on the inner-bend levee. These strata become interbedded and are indistinguishable from other inner-bend levee strata. As a result, N:G decreases from 0.9, immediately adjacent where channel-fill strata begin to split, to 0.15 over 140 m laterally away from the channel. The upward change in channel margin geometry along the inner bend suggests that levee relief became more subdued over time, allowing channelized flows to overspill onto the levee without becoming detached from the main channel flow.

- The outer-bend channel margin forms a composite terraced surface formed by three discrete episodes of channel inception followed by channel-levee growth. During each episode the channel thalweg migrated progressively toward the outer bend. Moreover, each episode of channel-levee growth began with a period dominated by levee deposition and channel bypass. Enhanced overbank flow along the outer bend of the channel, related to inertial and buoyancy forces, enabled the outer-bend levee to become significantly higher and coarser grained compared to its inner bend counterpart. In some cases the asymmetry in levee height is interpreted to be at least 3:1. This was then followed by intense channel filling that caused a reduction in levee relief. These conditions “reset” the channel system and allowed the next episode of channel-levee growth to repeat the pattern of channel-levee inception, growth and fill.

CHAPTER FIVE

An alternative model of producing topography in the crest region of deep-water levees

5.1. Introduction

On seismic images the crest region of deep-water levees typically exhibits a mound-shaped geometry made-up of multiple internal reflections (e.g. Piper et al., 1999b; Kendrick, 2000; Pirmez et al., 2000; Piper and Normark, 2001). The reflections typically dip towards the channel between the channel axis and levee crest (levee frontside) and away from the channel thereafter (levee backside). In this chapter, the proximal levee, or crest region, includes the area immediately adjacent the channel in addition to the entire levee frontside and the initial part of the levee backside (Fig. 5.1). The proximal levee changes laterally away from the channel into the areally more extensive deposits, referred to here as the distal levee.

In modern deep-water levee systems, like those on the Amazon Fan, the crest region extends laterally over 1.0–1.5 km, whereas the distal levee extends laterally over several kilometers. Moreover, seismic reflections generated by deep-water levee strata typically dip at progressively steeper angles laterally away from the crest reaching maximum dip angles between 1 and 5° (e.g. Pirmez, 1994; Piper and Normark, 2001, Deptuck et al., 2003). In the distal levee, seismic reflections then decrease in dip and converge laterally away from the channel.

Generally, it is assumed that dipping reflections in levees are generated by packages of dipping levee beds that thin away from the channel from proximal to distal levee (e.g.

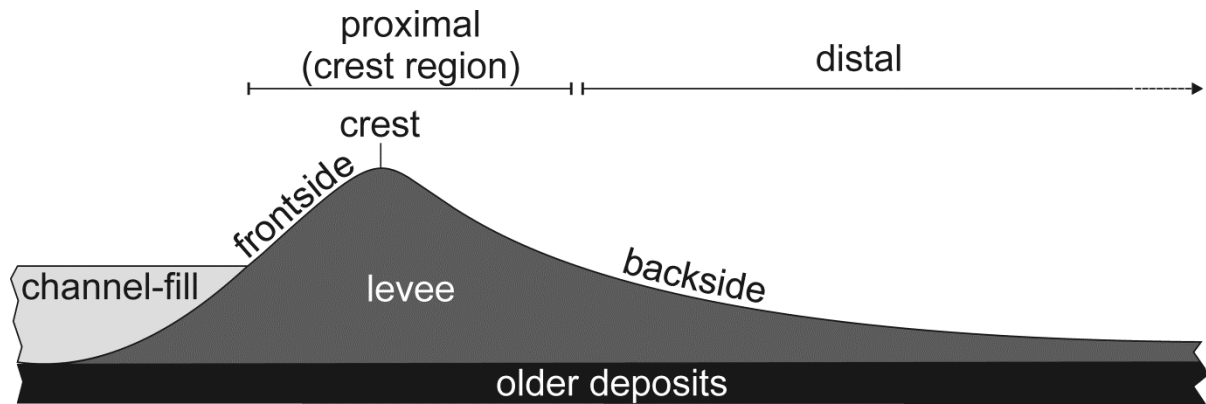


Fig. 5.1. Schematic diagram illustrating levee terminology used in this paper.

Kane et al., 2007, 2010b; Straub and Mohrig, 2008a). This commonly held perception, however, is not substantiated in the Neoproterozoic Isaac Formation where levee strata, particularly in the crest region, show no visible or measurable change in dip away from the channel based on lateral bed-by-bed correlations over several hundred meters. This therefore suggests that the vertical stacking of levee strata do not necessarily result in beds that dip at appreciable angles. The objectives of this chapter are to 1) document field relationships that confirm the lack of stratal dip in the crest region of an ancient levee system; 2) discuss the effects of compaction on stratal dip; 3) present an alternative model for creating levee topography, and 4) generate a synthetic seismic model that forms dipping reflections from non-dipping strata.

5.2. Geological setting and stratigraphy

The Isaac Formation makes up part of the Neoproterozoic Windermere Supergroup (WSG) of western North America (Fig. 5.2). The deep-water part of the WSG, which crops out extensively in the southern Canadian Cordillera, is interpreted to have been deposited on

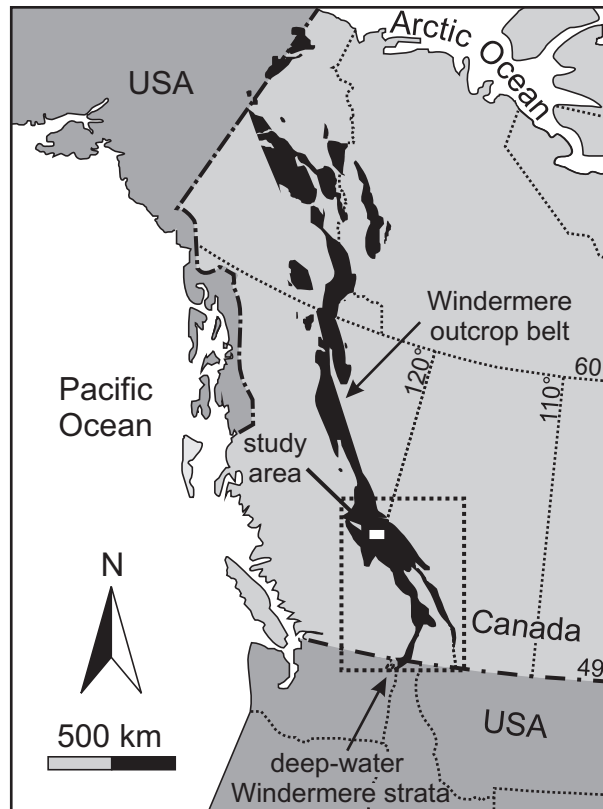


Fig. 5.2 Distribution of Windermere Supergroup strata in western Canada (modified from Ross, 1991). Study area is located in the southern Canadian Cordillera (dashed box) where the Windermere consists predominantly of deep-marine siliciclastic sedimentary rocks.

a passive margin following breakup of the supercontinent Rodinia at ca. 750 Ma (Ross et al., 1995). This ancient turbidite system was conservatively of the order of about 80 000 km² (Ross and Arnott, 2007), which makes it comparable in dimension and tectonic setting to modern systems like the Amazon and Mississippi deep-marine fans (Barnes and Normark, 1985).

5.2.1. Castle Creek

In the Castle Creek study area (Fig. 5.2), deep water rocks of the WSG are well exposed on the nearly vertically-dipping limb of a southwest-verging anticline. Despite regional low-grade metamorphism and tectonism in the area, effects on the preservation of

primary structures are relatively minor, even down to millimetre-scale details. Figure 5.3 shows a portion of the Castle Creek study area where strata crop out 2.2 km parallel to regional bedding and 2.1 km perpendicular to bedding. Here, the strata record a transition from deep-water basinal sandstone of the Kaza Group to base-of-slope deposits of the Isaac Formation. This succession is part of a kilometer-scale, upward-shallowing succession related to progradation of the continental margin of Laurentia (ancestral North America) into the proto-Pacific ocean (Ross et al., 1995; Ross and Arnott, 2007).

5.2.2. Isaac Formation

In Castle Creek, the Isaac Formation conformably overlies the Kaza Group and is exposed over a stratigraphic thickness of 1.6 km. This formation consists of six laterally discontinuous units that range from 30–100 m thick and consist predominantly of very coarse-grained sandstone turbidites with less common conglomerate and mudstone-clast breccia. These sandstone turbidite units are encased in interbedded sandstone and mudstone beds interpreted to have been deposited by turbidity currents. The coarser-grained units and enveloping finer-grained strata are interpreted to represent filled channels and their associated levee deposits, respectively. Channel complexes are informally termed Isaac Channel Complexes (ICC) 1–6 (Fig. 5.3) (Arnott, 2007a, b; Navarro et al., 2007a, b; Schwarz and Arnott, 2007a, b, c). Other strata observed in the Isaac Formation include thick debris flow, slump deposits, and slide deposits (Arnott et al., 2011). Collectively, these mass transport deposits indicate gravitational instability commonly associated with deep-water base-of-slope settings (Ross et al., 1995; Ross and Arnott, 2007).

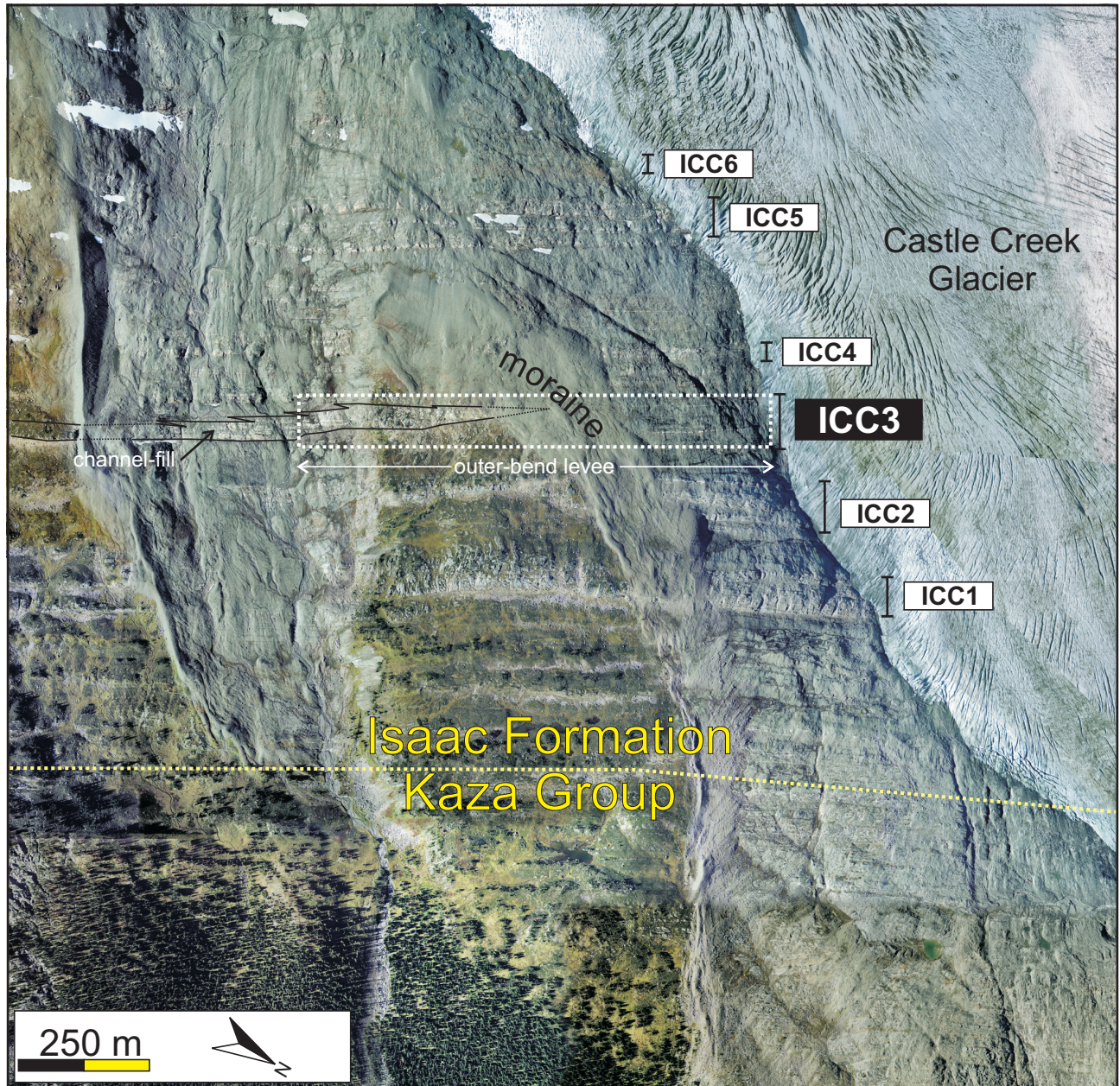


Fig. 5.3. Aerial photomosaic of the Castle Creek study area. Beds strike SE - NW and are nearly vertical. Locations of Isaac channel complex units 1–6 are shown (ICC1–ICC6). This study focuses on the outer-bend area of ICC3 (dotted box with outline of channel fill).

5.2.3. *Isaac Channel Complex 3*

Lateral trends from proximal to distal levee and the relationship between overbank strata and channel-fill deposits are best exhibited in ICC3 (Fig. 5.4). This channel complex, which abruptly overlies a mass transport deposit that is up to 135 m thick, crops out as an almost perpendicular-to-flow cut through a sinuous, deep-water channel-levee complex (Navarro et al., 2007b; Khan and Arnott, 2011). The channel-fill portion is 80 m thick and consists of four units that collectively narrow upward from more than 1000 m at the base to less than 500 m in the uppermost channel-fill units. Channel-fill deposits are bounded by levee deposits along both outer- and inner-bend margins (Khan and Arnott, 2011); however, exposure along the outer-bend margin (i.e. outer-bend levee) is better and will be the focus of this paper.

Outer-bend levee deposits of ICC3 are divided into proximal and distal areas based on their position relative to the channel (Fig. 5.4A). Note that the proximal and distal levee deposits are separated by a 250 m-wide moraine that buries the intervening strata (Fig. 5.4A). The proximal levee area occurs up to 450 m laterally away from the channel margin and reaches a maximum thickness of 60 m close to the moraine. In the opposite direction, proximal levee strata are progressively truncated by the terraced channel margin. Note that in Figure 5.4A the outer-bend margin of the channel exhibits offset terraces (“flats”) connected by steep channel banks (“steps”). This geometry suggests discrete periods of erosion followed by lateral migration along the outer-bend (cut bank) margin (Navarro et al., 2007; Khan and Arnott, 2011). Furthermore, the onlapping of channel-fill strata against the outer bank suggests that the levee built to a certain height prior to the onset of channel filling (Navarro et al., 2007; Khan and Arnott, 2011). The distal levee occurs 200–700 m laterally away from the channel margin and is continuously exposed vertically for 80 m and 300 m

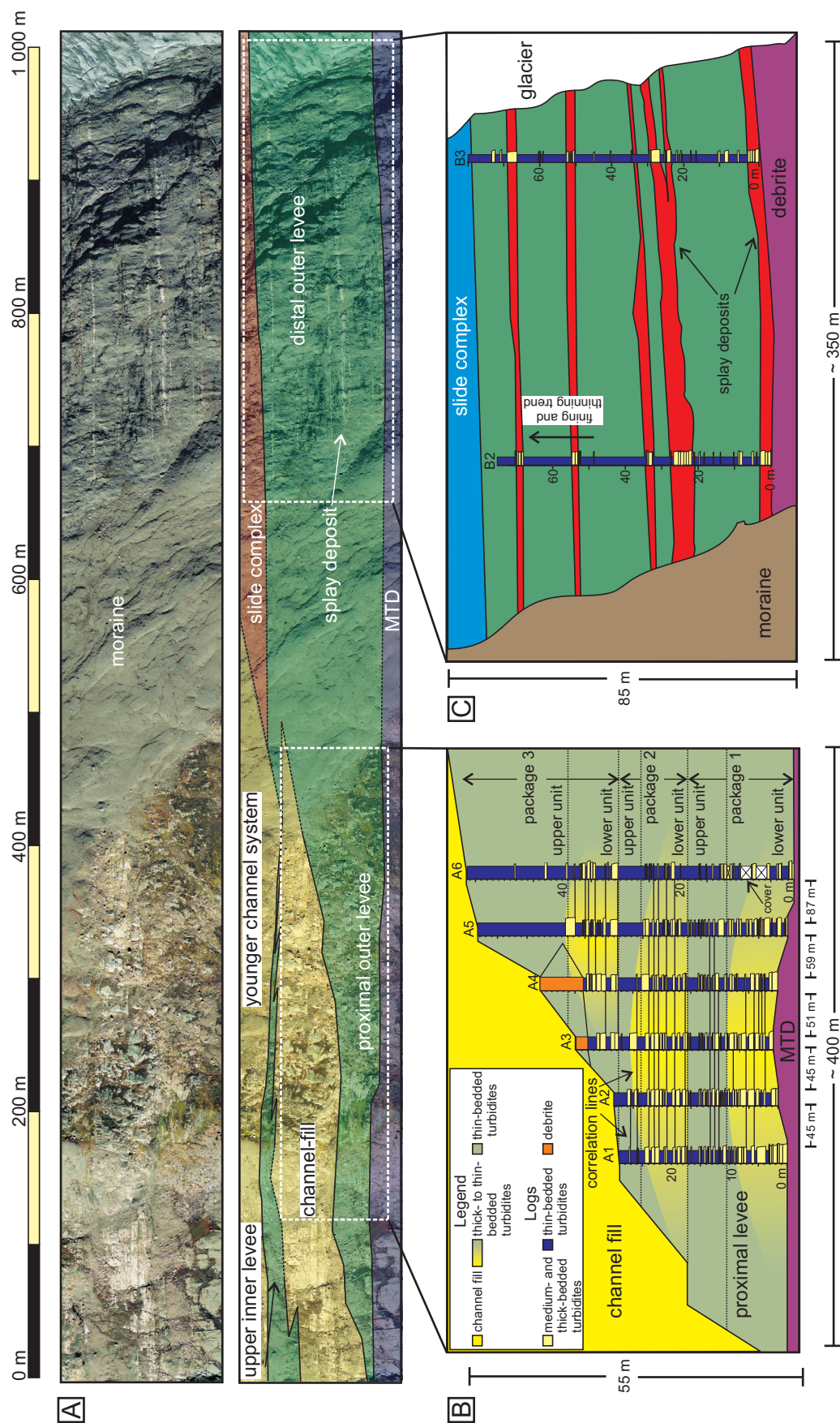


Fig. 5.4. A) Photomosaic of the outer-bend area of ICC3 showing continuity of outcrop and exposure of outer-bend levee deposits. B) Correlation panel showing detailed architecture of proximal levee deposits (modified from Khan and Arnott, 2011). Proximal levee packages are divided into sandstone-rich lower and mudstone-rich upper parts. Thicker-bedded deposits (F1) in the lower parts typically thicken then thin laterally. As a result, sst:mdst to about 42:58 over a distance of about 300 m). This lateral trend is represented schematically by the shading pattern. Notice that bed-by-bed correlation of proximal levee strata show little change in vertical position (i.e. negligible dip). C) Correlation panel showing detailed architecture in distal levee (modified from Khan and Arnott, 2011). Strata in the distal levee are mainly finer-grained and thinner compared to their proximal counterparts. Intercalated in the distal levee are five distinct stratal units that range from 2 to 5 m thick and are made up of discontinuous amalgamated F1 beds interpreted to be overbank splay deposits.

laterally before being covered by the Castle Creek Glacier.

5.3. Sedimentology of the outer-bend levee

Proximal levee strata are divided into three levee packages, which are further subdivided into lower and upper parts (Fig. 5.4B). The lower part ranges from 7–11 m thick and comprises interbedded medium- to thick-bedded (10–100 cm thick), normally graded, coarse to fine sandstone T_{a-d} turbidites (Facies F1; Table 5.1) and very thin- to thin-bedded (<10 cm), normally graded, fine sandstone to siltstone-mudstone T_{cd} turbidites (Facies F2; Table 5.1) (T_a – massive or normally graded sandstone; T_b – planar laminated sandstone; T_c – ripple cross-stratified sandstone; T_d – massive and subtly or well laminated siltstone-mudstone). The lower part is sharply overlain by the upper part, which, in contrast, consists almost exclusively of thin, fine-grained F2 beds. The upper part of packages ranges from 5–18 m thick. As a result of this upward change in facies, the boundary between the lower and upper parts of each levee package is marked by an abrupt thinning and fining of strata. Furthermore, F2 strata in each levee package become progressively more mudstone rich upward. Each package is interpreted to represent a discrete episode of levee growth that progressively confined the lower, coarse-grained part of flows to the channel (Khan and Arnott, 2011). Each episode of levee growth is interpreted to have been followed by significant channel filling that reduced the amount of levee relief and as a result enabled coarser-grained particles to once again overspill and deposit outside of the channel (Khan and Arnott, 2011). This process explains the consistent vertical stacking of proximal levee packages in the outer-bend region of ICC3 and shares similarities with previous interpretations from channel-levee systems in the Pab and Cerro Torro formations (Eschard

Table 5.1
General Characteristics of Levee Lithofacies

Lithofacies	Grain-size	Bed thickness	Typical Bouma successions	Lateral trend in bed thickness
F1	normally graded from coarse- to fine-grained sandstone	10–150 cm (4–60 in) avg. 40 cm (16 in)	- most common: T_{bc} , T_b , and T_c - less common: T_a , T_{ab} , and T_{abc} - overlain by thin (<5 cm/<2 in) siltstone-mudstone cap	- typically thicken laterally away from channel over 120 m (~400 ft) at about 3 mm/m (.03 in/ft) and thereafter thin laterally over the next 180 m (~600 ft) at 2 mm/m (0.02 in/ft). - typically grade from medium- (up to coarse-) to fine-grained sandstone.
F2	normally graded from fine-grained sandstone to siltstone-mudstone	<1–10 cm (<0.4–4 in) avg. 2–5 cm (0.8–2 in)	- most common: T_{cd} - less common T_{abed} and T_{bcd}	- beds show negligible change in thickness or grain size over 100's of meters (100's of feet).

et al., 2003 and Beaubouef, 2004, respectively).

The lower 62 m of strata in the distal levee is made up of F2 beds intercalated with F1 beds that become less common upward (Fig. 5.4C). The upper 18 m of the distal levee, in contrast, is composed predominantly of F2 beds that display a well developed upward-fining and -thinning trend. Collectively, strata in the distal levee are intercalated with four sharp-based stratal units that range from 2–5 m thick and are made up of discontinuous amalgamated F1 beds interpreted to be overbank splay deposits.

5.4. Levee slope

Seismic data from offshore Brunei Darussalam suggest that slope associated with levees temporally increases until a certain threshold height above the sea floor is obtained after which it remains relatively constant (Straub and Mohrig, 2008b). Based on modern deep-water systems this slope typically ranges from 1–3° (e.g. Piper and Normark, 2001; Deptuck et al., 2003; Posamentier and Kolla, 2003), but in some cases is up to 5° (e.g. Pirmez, 1994). Similar slope trends have been created experimentally (Straub and Mohrig,

2008a). In the experimentally generated levees, thick, coarse-grained beds, typically occurring in the lower part of the levee, exhibit the highest degree of lateral thinning and as a result the stacking of such beds yield a progressively steeper slope. In contrast, thin, fine-grained strata, which make up most of the upper parts of levees, exhibit much more gradual thinning trends and, therefore, their stacking does not significantly change the levee slope. The levee frontside occurs over a fraction of the distance compared to the levee backside and therefore is not the focus of most studies. Notwithstanding, Kane et al. (2007) noted that in the Rosario Formation the average bed thickness of levee strata initially increases laterally away from the channel and then decreases. Kane et al. (2007) suggests that the vertical stacking of beds that thicken and then thin gives rise to the typical mound-shaped geometry associated with levee crests.

5.4.1. Lack of stratal dip in Castle Creek levees

In the rock record, levee strata commonly exhibit the systematic lateral thinning trends discussed by Straub and Mohrig (2008a). The thinning trend is related primarily to the thicker-bedded, coarser-grained turbidites, which thin more rapidly away from the channel than their thinner-bedded, finer-grained counterparts (e.g. DeVries and Lindholm, 1994; Piper and Deptuck, 1997; Beaubouef, 2004; Hubbard et al., 2008). Thickening then thinning trends are less commonly observed, but as mentioned above, they have been reported from the Rosario Formation, Mexico (Kane et al. 2007). In that study, data were compiled from average bed thicknesses calculated from eight stratigraphic logs over a distance of about 1.5 km. Levee strata in the proximal outer-bend levee of ICC3 also display a lateral thickening and then thinning trend; however, here data are based on individual beds that have been walked out laterally. Specifically, F1 strata in the proximal levee of ICC3 typically thicken

laterally over 120 m exhibiting an average thickening gradient of about 3 mm/m. After achieving their maximum thickness, F1 beds begin to thin over the next 180 m by an average thinning gradient of 2 mm/m. The basal part of F1 beds also show a lateral fining over this 180 m lateral distance, typically grading from medium- (up to coarse-) to fine-grained sandstone. Facies 1 (F1) beds commonly thin to about 10 cm or less (i.e. they transition into F2 beds), after which they display negligible thinning or fining over the remainder of the outcrop, which in some cases is just over 100 m, beyond which the moraine buries the strata before reaching the distal levee region. Beds in the proximal levee that are initially less than 10 cm thick do not thin or fine appreciably laterally.

Theoretically, the vertical stacking of thickening then thinning strata should cause stratal dip to increase progressively upward on both the levee frontside and backside. Topography associated with the bed surfaces of laterally thickening then thinning F1 strata in ICC3, however, is commonly infilled by overlying fine-grained, thin-bedded deposits (F2) (see Fig. 5.5). This "healing" of local topography has also been reported from other deep-water systems (e.g. Piper et al., 1999b). In the case of ICC3, the elimination of topography causes levee beds to exhibit no detectable dip and as a result they remain consistently sub-parallel to regional strike (Fig. 5.4B). This suggests that topography, and thus stratal dip, associated with F1 beds in the crest region of the levee is either muted or eliminated over time, leading to the deposition of strata that are, on average, horizontal. This would appear to be at odds with the dipping reflections, presumed to be associated with dipping strata, which are observed in seismic images of modern deep-water systems.

The foregoing discussion is based only on observations from strata in the proximal levee region of ICC3. In the distal levee, correlation relies principally on thin-bedded deposits, which because of their thinness, abundance, and monotonous character makes them

difficult to trace laterally, even across minor cover. High-resolution aerial photos and correlation of meter-scale splay deposits in the distal levee (see Fig. 5.4C), however, show no evidence of stratal dip. Nevertheless, to deny the existence of stratal dip in the distal levee based on such a modest dataset is difficult and contrary to reports where individual turbidites in shallow cores have been confidently correlated and shown to be dipping (e.g. Piper and Normark, 1983). At the very least, for the study area, data from the proximal levee of ICC3 confirm the lack of stratal dip in the crest region, and possibly it is only on the laterally adjacent but more distant part of the levee that stratal dip becomes evident.

5.4.2. Common perceptions for lack of stratal dip in the rock record

Although stratal dip is commonly inferred to exist in modern levee deposits based on seismic data, it commonly appears lacking in outcrop exposures of ancient levee deposits (e.g. DeVries and Lindholm, 1994; Beaubouef, 2004; Kane et al., 2007). Apparent lack of dip in outcrop exposures has been attributed to 1) the low dip angle ($1\text{--}5^\circ$) commonly associated with levees, which would not be easily resolved in the field (H. Posamentier, 2008, personal communication); and 2) compaction, which would significantly reduce slope to imperceptibly small angles (Posamentier, 2003).

The assumption that slopes associated with levees are too small to be observed in the field is brought into question by the conceptual exercise presented in Figure 5.6. Here six hypothetical measured sections of a levee deposit are displayed. The strata are drawn to show how the internal stratigraphy of levees is envisioned by Kane et al. (2007). The levee, which is up to 20 m thick at its crest, is shown over a lateral distance of 300 m. Over the first 120 m laterally away from the channel (levee frontside), levee beds dip towards the channel, and increase in dip moving stratigraphically upwards, reaching a maximum slope of 2° . Over the

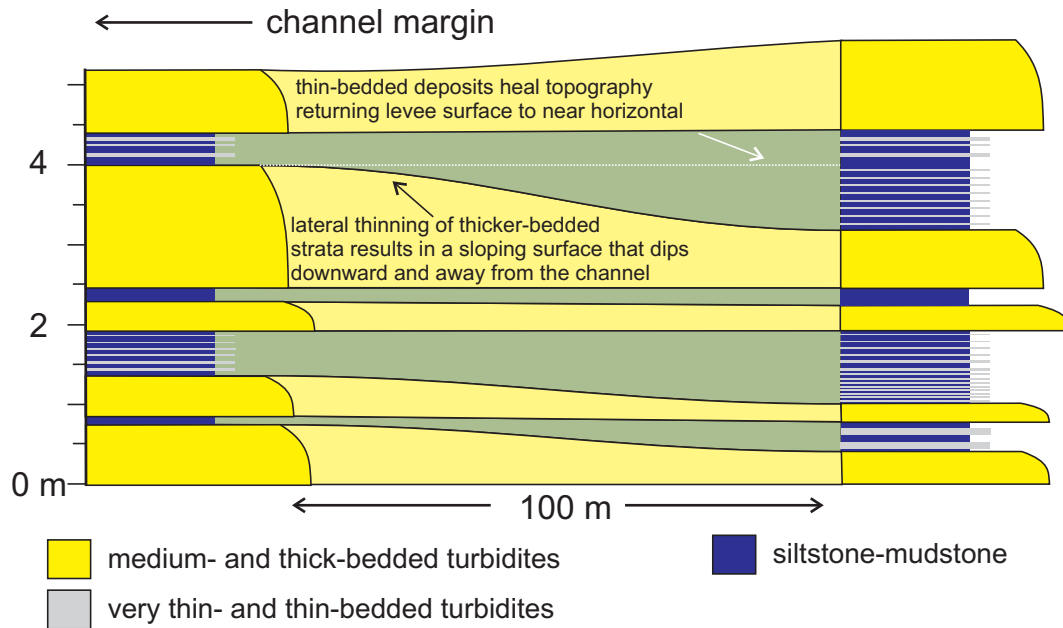


Fig. 5.5. Evidence of infilling (healing) of local topography in the proximal levee of ICC3 illustrated by two correlated detailed bed-by-bed measured sections. Sections are ~5 m thick and separated by ~100 m. These sections occur in the lower part of levee package 2 (see Fig. 5.4B). Note that a greater number of thin-bedded deposits are recorded laterally away from the channel margin. This suggests that thin-bedded strata are preferentially deposited where the underlying thicker-bedded strata are thinnest. This eliminates local topography associated with the lateral thinning of thicker-bedded deposits. Note also that the uppermost bed thickens laterally, but then further laterally begins to thin (not shown here), and therefore illustrates the common initial thickening geometry observed in many medium- to thick-bedded turbidites in ICC3. Moreover, the lower four thicker-bedded deposits thin laterally towards the channel (not shown here).

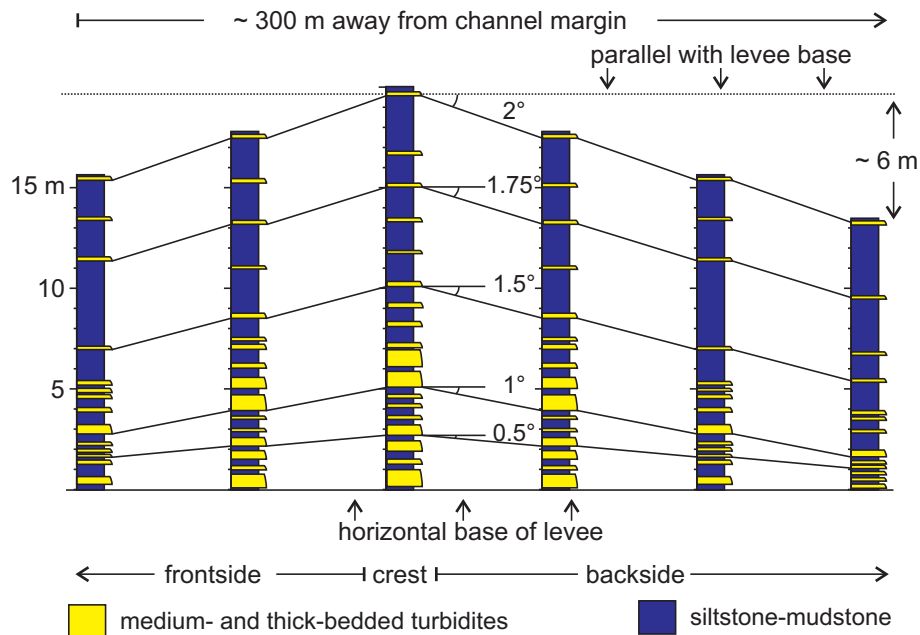


Fig. 5.6. Hypothetical measured sections of a levee in which beds at the base are horizontal and progressively increase in dip upwards (up to 2°). Note that correlative beds in the upper part of the section occur approximately 6 m stratigraphically lower in the last section compared to the section at the crest. In well exposed sections, like at Castle Creek, such a topographic change would be noticeable, but was not detected (see Fig. 5.4B).

next 180 m beyond the levee crest (levee backside), strata begin to dip away from the channel at a maximum slope of 2° . In the absence of healing phase deposits, the vertical stacking of thickening then thinning beds clearly gives rise to the mound-shaped crest geometry of the levee, and results in a drop in elevation of up to 6 m over a horizontal distance of less than 200 m. As shown previously, the strata in ICC3 do not show any appreciable drop over several hundreds of meters laterally (see Fig. 5.4B). This suggests that the basal contacts of strata in ICC3 do not dip towards or away from the channel at any measurable angle over the length of the outcrop.

The second argument that the lack of stratal dip in outcrop is related to compaction effects is addressed in Figure 5.7. Here a hypothetical levee on the sea floor is shown to thicken from 15.8 to 20.0 m over the first 120 m laterally away from the channel and thereafter thin to 13.7 m over the next 180 m. This variability in lateral thickness yields a

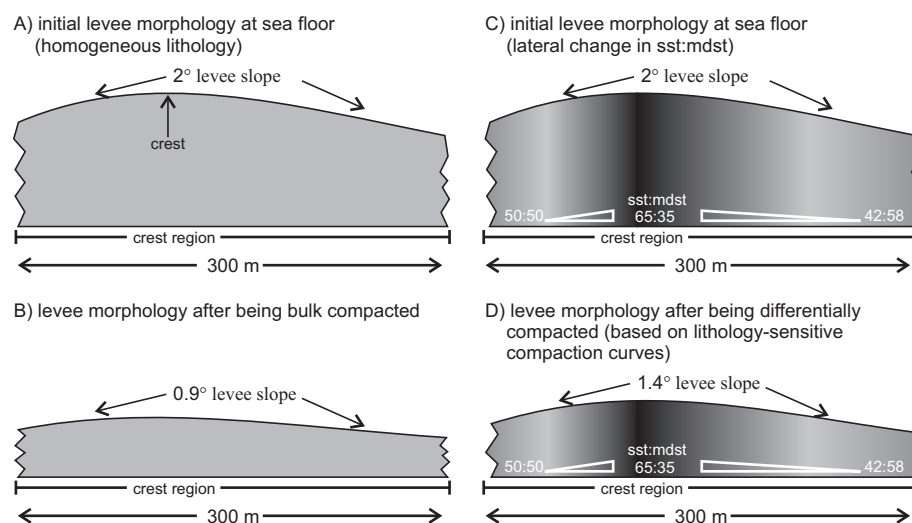


Fig. 5.7. Levee compaction model. A) A homogeneous muddy levee at the sea floor with a high initial porosity (0.6). The levee thickens from 15.8 to 20 m over the first 120 m laterally away from the channel and thereafter thins to 13.7 m over the next 180 m, which yields a frontside and backside slope of 2° . B) Levee in (A) is uniformly mechanically compacted until the porosity is reduced to 0.1, which causes the levee slope to decrease to 0.9° . C) A levee where sst:mdst increases over the first 120 m laterally away from the channel from 50:50 to 65:35 and thereafter decreases over the next 180 m to 42:58. D) The area closest to the channel, the crest, and the area 180 m beyond the crest in (C) are independently compacted to a depth of 3 km following Sclater and Christie (1980)(see text for details). Note that the decrease in slope is smaller compared to (B).

moderate slope of 2° along both the levee frontside and backside (Fig. 5.7A). After compaction the slope along this levee is reduced to 0.9° (Fig. 5.7B), which demonstrates how compaction could significantly reduce levee slope. Notwithstanding, a stratal dip of 0.9° would still be evident in the field as shown in Figure 5.6, provided that marker beds are present.

The compaction illustrated in Figure 5.7B assumes a homogeneous muddy lithology with an initial porosity of 0.6 that decreases to 0.1 after being mechanically compacted, resulting in a more than 50% reduction in levee thickness. However, a bulk compaction of a homogeneous lithology is a gross oversimplification of most natural levees, which typically vary significantly in facies laterally away from the channel. For example, based on both modern (e.g. Clemenceau et al., 2000) and ancient (e.g. Devries and Lindholm, 1994; Kane et al., 2007) data sets, proximal levees are up to five times sandier than their distal levee counterpart. When combining data from the three levee packages in ICC3 it is observed that the ratio of sandstone-to-mudstone (sst:mdst) increases over the first 120 m laterally away from the channel from 50:50 to 65:35 and thereafter decreases over the next 180 m to 42:58 – note that this sst:mdst trend reflects the changing lateral thickness of individual F1 strata. With this added complexity Figure 5.7C illustrates a more realistic levee with a frontside and backside slope of 2° and a lateral sst:mdst trend based on data from ICC3.

Because of the lateral facies trends discussed above, differential compaction is suspected to occur to deep-water levee deposits during burial. Sand has a lower initial porosity compared to mud (MacCary and Lambert, 1962; Cohen, 1965; Davis, 1969; Kitajima et al., 2005) and also experiences lower compaction rates (Sclater and Christie, 1980; Rowan et al., 2003). Accordingly, the amount of compaction that occurs during burial

at the levee crest in Figure 5.7C is less than that of muddier areas laterally away from the crest. In order to quantify exactly how differently the crest and areas lateral to the crest compact it is first necessary to calculate the initial porosity and compaction constant describing these different lithologies. Calculation of these two compaction parameters is based on Sclater and Christie's (1980) data from deep-water deposits in the North Sea where porosity loss with depth was estimated by the following equation:

$$\phi = \phi_0 e^{-cz} \text{ (Rubey and Hubbert, 1959)} \quad (5.1)$$

where ϕ_0 is the porosity at the surface, ϕ is the porosity at depth z (in km), and c is the lithological dependent compaction constant. In their paper different initial porosity and compaction constants for sand, shale, and shaley sand are presented (Table 5.2). Note that the shaley sand reported by Sclater and Christie (1980) had an initial porosity and compaction constant exactly between that of sand and shale. Assuming that the shaley sand reported by Sclater and Christie (1980) had a sand-to-mud ratio of 50:50 then the initial porosity and compaction constant values decrease linearly with increasing sand content as follows:

$$\phi_0 = -0.14\text{sst:mdst} + 0.63 \quad (5.2)$$

and

$$c = -0.24\text{sst:mdst} + 0.51 \quad (5.3)$$

where sst:mdst is the sand-to-mud ratio of a given lithology. Using equations (5.2) and (5.3) the initial porosity and compaction constant are calculated for the different areas of the proximal levee in Figure 5.7C. These values, in turn, are used to independently calculate the compaction effects at the crest and adjacent regions following the method described by Sclater and Christie (1980). In this method, the thickness contributed by pore space (T_ϕ) in a stratal column between depths z_1 and z_2 is calculated by integrating:

$$T_{\emptyset(z_2-z_1)} = \int_{z_1}^{z_2} \phi_0 e^{-cz} dz \quad (5.4)$$

giving

$$T_{\emptyset(z_2-z_1)} = \frac{\phi_0}{c} (e^{-cz_1} - e^{-cz_2}) \quad (5.5)$$

Assuming that all the pressure dissolved solid is precipitated locally (i.e. a closed system) then the volume occupied by solids does not change during mechanical compaction. If true then the compacted thickness T_C (i.e. $z'_2 - z'_1$) of a levee of thickness T_0 (i.e. $z_2 - z_1$) is defined as

$$T_C - \int_{z'_1}^{z'_2} \phi_0 e^{-cz'} dz = T_0 - T_{\emptyset(z_2-z_1)} \quad (5.6)$$

giving

$$T_C - \frac{\phi_0}{c} (e^{-cz'_1} - e^{-cz'_2}) = T_0 - T_{\emptyset(z_2-z_1)} \quad (5.7)$$

Using this method the effects of compaction at the crest of the proximal levee in Figure 5.7C at 3 km depth is calculated by setting $z_1 = 0$ and $z_2 = 0.02$ km (i.e. 20 m thick at the surface) and $z'_2 = 3$ km (i.e. burying the levee base to a depth of 3 km) giving

$$z'_1 + \frac{\phi_0}{c} (e^{-cz'_1}) = 3.53 \quad (5.9)$$

where z'_1 is solved numerically, giving a value of 2.9886 km. Therefore the compacted thickness (T_C), which is $z'_2 - z'_1$, is equal to 11.4 m. Compaction immediately adjacent the channel margin and 180 m laterally beyond the levee crest are similarly solved and give values of 8.4 m and 7.1 m, respectively. These values are combined in Figure 5.7D and result in a slope along the frontside and backside levee of $\sim 1.4^\circ$. This demonstrates that the reduction in slope during burial is not as significant in a heterolithic levee system compared to a homogeneous one. Furthermore, the initial porosity of sand used in this example is particularly high at 0.49 when typical values for modern deep-water sand range from 0.15 to

0.35 (Clemenceau et al., 2000; Kitajima et al., 2005). Accordingly, Figure 5.7D represents a maximum decrease in levee slope that could be expected after mechanical compaction at a depth of 3 km. Of note is that chemical compaction generally becomes the principal cause of porosity loss in silica-rich sediments beyond depths of 3 km (Ramm, 1992; Ramm and Bjorlykke, 1994; Giles, 1997). Moreover, in closed systems, where all the pressure dissolved solid is precipitated locally, the bulk geometrical structure following mechanical compaction is preserved (Schneider and Hay, 2001). Therefore, Figure 5.7D likely represents the geometrical effects of compaction at depths much greater than 3 km and thus acts as a proxy for effects of mechanical compaction in the ancient rock record.

Another method to estimate compaction effects is to use the porosity at the midpoint of a stratal interval (i.e. $\frac{z_2 - z_1}{2}$) as an estimate of the entire interval between $z_2 - z_1$. Using this method it is then possible to calculate the thickness contributed by the sediment (T_{solid}) by

$$T_{solid} = (z_2 - z_1) \left(1 - \phi_0 e^{-c \left(\frac{z_2 - z_1}{2} \right)} \right) \quad (5.9)$$

Assuming once again that T_{solid} remains constant with depth, then the ratio between T_{solid} and the total compacted thickness (T_c) at some depth must equal

$$T_c = \frac{T_{solid}}{1 - \phi_0 e^{-cz'}} \quad (5.10)$$

where (z') is the midpoint depth of the compacted interval. Of note is that the midpoint porosity used here only approximates the porosity of the entire interval and is not as accurate as integrating the porosity as in equation (5.7). Notwithstanding, this simpler method works exceptionally well on thin intervals, which is a consequence of the fact that e^{-cz} decreases almost linearly over small intervals. As a result, compacting the levee in Figure 5.7C using

equation (5.10) with $z' = 3$ km, gives values that are within 1% of those using equation (5.7). Furthermore, the difference in the values calculated from each method derives mostly from how the depth terms are defined. Specifically, setting $z' = 3$ km in equation (5.10) puts the midpoint of the compacted levee at 3 km whereas setting $z'_2 = 3$ in equation (5.7) puts the base of the levee at 3 km.

Figure 5.7 illustrates the effects of compaction on initial levee slope. The ancient rock record, in contrast, offers a view of vertical and lateral trends associated with levees that have already been compacted. Accordingly, a useful exercise is to use these details and analyze the effects of decompaction using a simple model. This is illustrated in Figure 5.8 where five laterally thickening and then thinning sandstone beds and intercalated mudstone-rich units (20:80) are decompacted. The stratigraphy in Figure 5.8A loosely resembles the situation observed in ICC3 (see Fig. 5.5). The decompaction presented in Figure 5.8 is purely theoretical and follows the principles developed in Figure 5.7, where changes in bulk volume are estimated to be reflected entirely by porosity loss due to mechanical compaction to a depth of 3 km. Accordingly, the decompaction factor used in Figure 5.8 is the ratio of the initial thickness (T_i) to T_c , where T_i is solved by:

$$T_i = \frac{T_{solid}}{1 - \phi_0} \quad (5.11)$$

but in this case T_{solid} is:

$$(z'_2 - z'_1) \left(1 - \phi_0 e^{-c \left(\frac{z'_2 - z'_1}{2} \right)} \right)$$

Finally, the decompaction factor is calculated by:

$$\frac{T_i}{T_c} = \frac{1 - \phi_0 e^{-cz'}}{1 - \phi_0} \quad (5.12)$$

where z' is 3 km. Using equation (5.12) compaction factors of $T_i = 1.53T_c$ and $T_i = 2.13T_c$ were respectively calculated for the sand and mud-rich (20:80) units. Based on these values the stratal unit in Figure 5.8A is decompacted in Figure 5.8B. Decompaction results in a scenario where over the first 100 m laterally from the channel the base of beds dip away from the channel and thereafter dip towards the channel. Also, stratigraphically upward the bases of beds dip increasingly more steeply. This creates an overall concave-up morphology and is counter to common perception where levee beds are typically inferred to thicken towards and dip away from the levee crest resulting in a concave-down morphology. Therefore, it is difficult to imagine the initial levee condition required under simple decompaction considerations to produce the parallel non-dipping strata observed in ICC3. This highlights the difficulty involved in properly assessing the effects of compaction and decompaction on ancient levees and demonstrates that other factors, for example early diagenetic cements, must also be considered when decompacting sedimentary rocks.

Table 5.2

Compaction constant and initial porosity used by Sclater and Christie (1980) for deep-water sediment in the North Sea

Lithology	Initial porosity (ϕ_0)	Compaction constant (c)	sst:mdst
Mud	0.63	$0.51 \times 10^{-5} \text{ cm}^{-1}$	0:100
Muddy sand	0.56	$0.39 \times 10^{-5} \text{ cm}^{-1}$	50:50
sand	0.49	$0.27 \times 10^{-5} \text{ cm}^{-1}$	100:0

plotting ϕ_0 vs. sst:mdst – $\phi_0 = -0.14(\text{sst:mdst}) + 0.63$

plotting c vs. sst:mdst – $c = -0.24(\text{sst:mdst}) + 0.51$

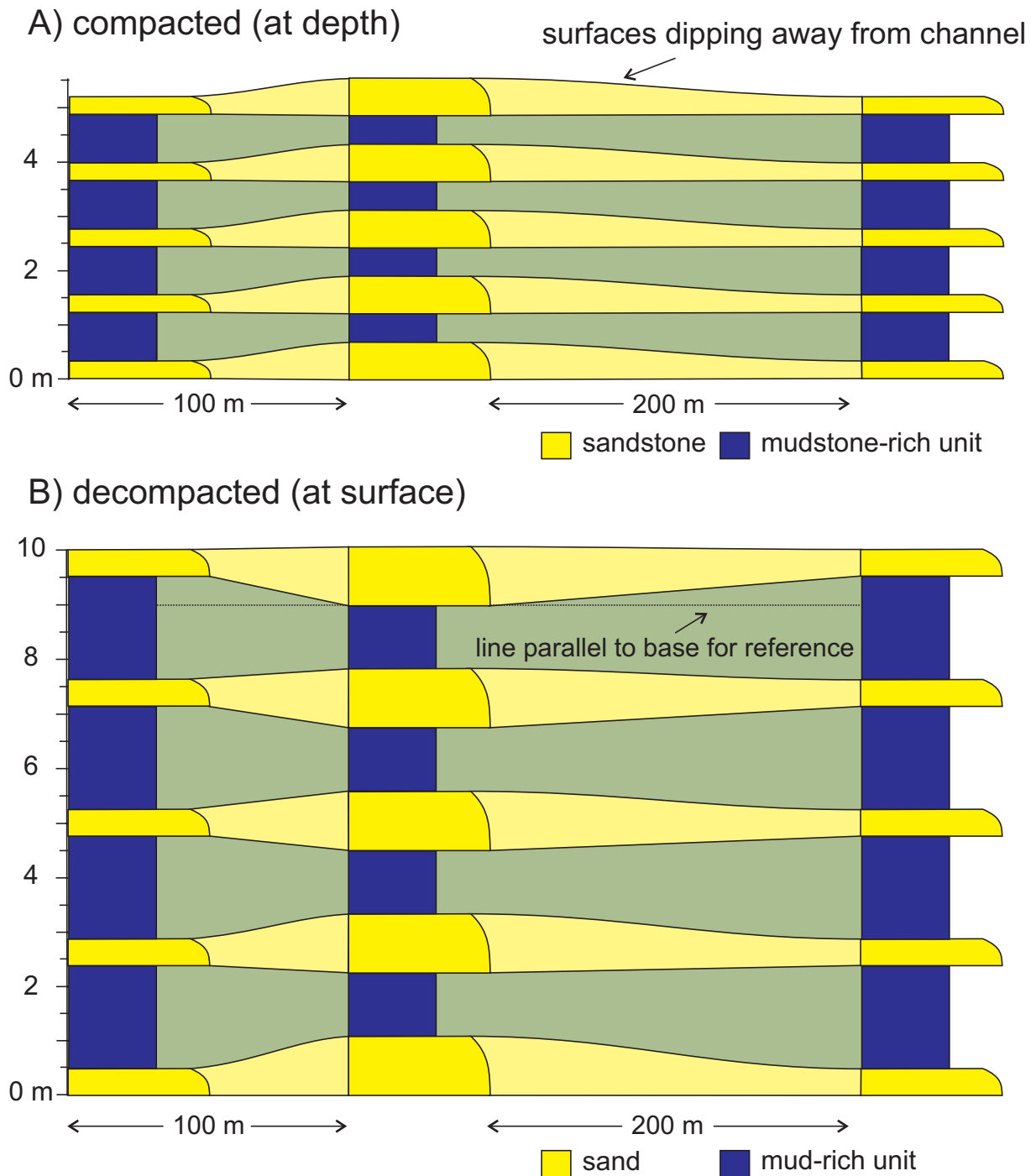


Fig. 5.8. Effects of decompaction on levee stratigraphy. Channel is located to the left. A) A ~5 m levee compacted at depth. Here five laterally thickening then thinning thick-bedded turbidites are encased in mudstone-rich (20:80) deposits and resembles the conditions observed in ICC3. B) Levee is decompacting using methods described in the text. Note that stratigraphically upward bases of beds progressively dip towards the “levee crest” (i.e. where beds are thickest). Beds dipping downward and toward the levee crest are not typically (if ever) reported from modern levees, and therefore highlight some of the inherent difficulties associated with applying simple decompaction models to the ancient rock record.

5.5. Alternative model of producing levee topography

The model presented below illustrates how thickening and then thinning beds deposited parallel to one another with no dip can create the typical mound-like levee geometry observed in the crest region of many deep-water systems. Levee topography in this model is interpreted to develop in three main stages based largely on differing degrees of channel confinement and levee slope away from the channel.

5.5.1. Stage 1

Initially the channel was weakly confined, allowing coarser sediment in the lower part of throughgoing flows to easily overspill and deposit F1 strata. The overbank area in ICC3 is interpreted to have been initially horizontal and, as a result, flows would have immediately begun to thin and decelerate upon overbanking the channel margin. Notwithstanding, the initial thickening trend of F1 beds was an inertial effect associated with high energy overspilling flows as they rounded the channel bend, which inhibited deposition immediately outboard of the channel. It is only after some distance away from the channel, in this case about 120 m, that flow speeds decreased and maximum rates of deposition ensued. After achieving maximum thickness, F1 strata began to thin and fine laterally as a result of decreasing flow competence and capacity wherein the coarser suspended sediment fraction was progressively lost to the bedload layer. Accordingly, remaining particles in the current became progressively finer grained, to a point where they were characterized by sufficiently low settling velocities where turbulent mixing created a uniformly distributed sediment population (Kneller and Buckee, 2000). Deposition from such flows resulted in thin-bedded, fine-grained F2 strata that show negligible lateral thinning or fining. Therefore, higher

energy, coarser-grained flows resulted in the deposition of coarse- to medium-grained F1 beds that transition laterally into F2 beds in the lower part of proximal levee packages. Consequently, these beds are the most laterally extensive within the levee. Furthermore, the distinct lateral thickening then thinning trend associated with thicker-bedded F1 beds resulted in subtle topography immediately following their deposition (Fig. 5.9A). The occurrence of intercalated F2 beds in the lower part of proximal levee packages imply that the magnitude of throughgoing channel flows fluctuated over time. Moreover, on average 70% of the total number of beds recorded in the lower part of levee packages are thin-bedded F2 beds, indicating that small magnitude overspill flows were the most common type during periods of low channel confinement. As a result, topography associated with F1 deposits was generally infilled by thin-bedded turbidites (F2) deposited by lower energy flows with shorter run-out distances (evidence of infilling is illustrated in Figure 5.5). These F2 beds show little change laterally suggesting deposition from overbanking flows that were initially well mixed and consisted of a narrow range of grain sizes.

A consequence of these more common flows, initially depositing preferentially in topographic lows where F1 beds were thinnest, was that they did not contribute to building local levee relief. Accordingly, the character of successive overspill flows that healed underlying topography did not change significantly, which in turn caused beds to terminate on average a similar distance away from the channel. The stacking of these thin-bedded deposits created a surface elevated above the surrounding sea floor (Fig. 5.9B). Importantly, where observed, F2 beds (at least the sandy basal part of beds) in the POB of ICC3 pinch out rather abruptly (over 10's of metres) and therefore it is interpreted that the progressive stacking of thin-bedded deposits that filled in local topographic lows created a steep slope downflow of their termini (see circle in Fig. 5.9B).

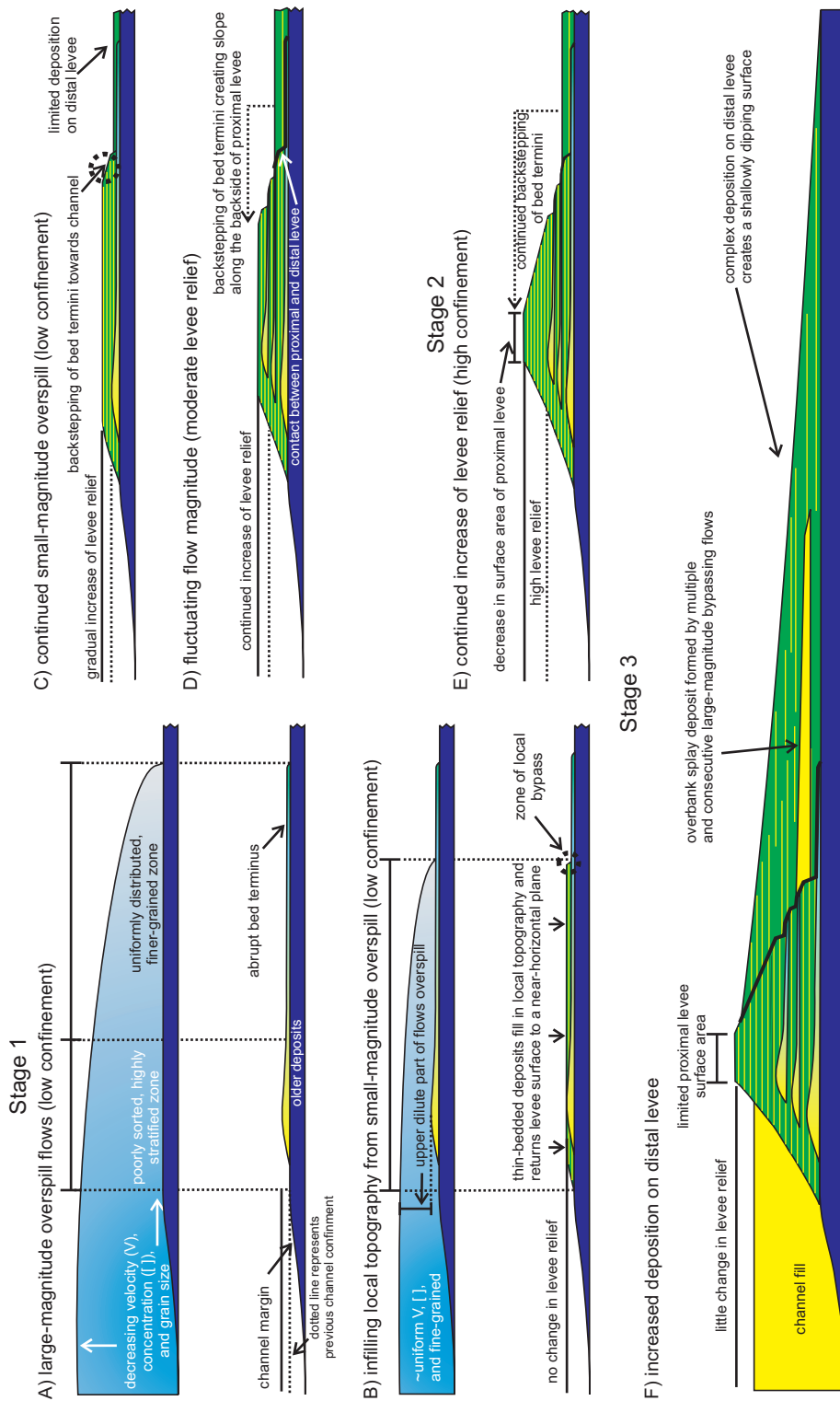


Fig. 5.9. Levee topography is interpreted to develop in three main stages in which each is characterized by differing degrees of channel confinement and levee slope away from the channel. **Stage 1:** A) Initially the channel was weakly confined allowing higher energy, sand-rich flows to deposit laterally thickening then thinning, thick-bedded (F1) turbidites. B) Topography along the surface of F1 strata was later infilled by thin-bedded turbidites (F2) deposited by lower energy flows with shorter run-out distances. The stacking of these thin-bedded turbidites created topographic relief and a steep slope downflow (circle). C) Once the topography of an underlying F1 bed was infilled levee relief continued to increase and overspilling flows became progressively more dilute and slower moving resulting in the backstepping of bed termini toward the channel (circle). D) Relief added by the deposition of subsequent F1 strata caused an abrupt offset of overlying thin-bedded bed termini towards the channel. This “stepped” slope acted as a zone of bypass for some flows promoting limited deposition on the distal levee. **Stage 2:** E) This stage was characterized by increasing channel confinement that progressively allowed finer, lower energy flows to overspill resulting in a systematic channelward backstepping of bed termini. **Stage 3:** F) This stage was characterized by the onset of channel filling decreasing the amount of channel relief and once again allowed higher energy flows to overspill. This, coupled with the fact that the surface area of the proximal levee was greatly diminished, and the slope along the backside of the levee was at its steepest, promoted a period of significant distal levee deposition. With time, this deposition reduced the angle and shortened the length of the proximal levee slope. This scenario caused the momentum of bypassing flows to gradually decrease, which in turn progressively reduced the run-out distance of flows over the distal levee. As a result beds progressively backstepped towards the channel creating a surface that sloped downward and away from the channel.

The upper surface of the proximal levee was returned close to horizontal once thin-bedded deposition healed the underlying topography with a steep slope created locally downcurrent, at the edge of the proximal levee (POB). This steep slope is inferred; it is either buried under the moraine or present but not resolvable in the mud-rich distal levee (DOB) deposits. This allowed subsequent overspill deposition to once again progressively increase local levee relief, which in turn caused small magnitude flows to progressively overspill only their upper more dilute and slower moving part. Consequently, these flows had progressively shorter run-out distances and caused successive bed termini to backstep toward the channel. This created a more subdued sloped surface along the backside of the levee compared to the one formed by the stacking of thin-bedded deposits that filled in topography (see circle in Fig. 5.9C).

A later large magnitude flow would be less affected by the increase in local levee relief due to its greater size and energy. In this case the flow would deposit sediment along the width of the proximal levee reaching the slope created previously by the stacking of thin-bedded turbidites. After reaching this slope, the flow would bypass the sloped proximal levee backside and deposit on the distal levee. For this condition to prevail the slope on the levee would have to be sufficiently steep to overcome the inherent deceleration associated with the lateral thinning of an overspilling flow (Hiscott et al., 1997). In ICC3, the occurrence of slump deposits in the distal levee suggests that at times a steep slope developed along the backside of the levee, therefore, providing ideal conditions for flows to bypass locally.

On the proximal levee, deposition from high energy flows would have abruptly increased local levee relief, in some cases by over 1 m (i.e. the thickness of some F1 beds). This quick addition of relief caused later small magnitude overspill flows to be appreciably more dilute and hence slower moving than previous flows. As a result, the thin-bedded

deposits that in filled the local topography created by the underlying F1 deposit were appreciably shorter (i.e. pinched out closer to the channel) than their underlying counterparts. This resulted in an overall large-scale stepped geometry with long flats and thick steps along the backside of the proximal levee (Fig. 5.9D). This demonstrates how an initially sub-horizontal levee surface rapidly increases its slope with time and deposition.

A consequence of the stepped geometry of the levee backside surface was the progressive reduction in surface area of the proximal levee. As a result, overspilling flows tended to increasingly bypass into the distal levee area as the height of the proximal levee increased during Stage 1. Deposition in the distal levee depended on the characteristics of the flow as it reached the backside slope of the proximal levee. For example, small magnitude bypass flows deposited thin-bedded F2 strata, whereas larger magnitude flows deposited thicker-bedded F1 strata on the distal levee. This model explains the occurrence of beds that pinch out away from the channel in the proximal levee and towards the channel in the distal levee. However, deposition on the distal levee was limited during this stage and was greatly outpaced by proximal levee aggradation. Accordingly, relief between the proximal and distal levee progressively increased.

5.5.2. Stage 2

As a result of the constant increase in levee relief throughout Stage 1 a threshold height was eventually reached where the lower, coarser-grained parts of most flows remained entirely contained within the channel. Consequently only the finer, more dilute upper part of the channelized flow typically overspilled and deposited F2 strata. This period of mostly thin-bedded, fine-grained deposition formed the upper part of proximal levee packages. In addition, the upward-fining trend observed in the upper part of levee packages suggests that

the channel became increasingly confined throughout Stage 2. Accordingly, overspilling flows generally became more dilute and slower moving resulting in the continued backstepping of bed termini towards the channel (Fig. 5.9E). Unlike in the lower part of packages, the paucity of intercalated F1 beds allowed thin-bedded termini in the upper part to backstep more systematically towards the channel.

Continued backstepping of bed termini during Stage 2 further decreased the surface area on top of the proximal levee. This allowed anomalously more energetic flows to deposit along the entire proximal levee and then descend the proximal levee backside slope and deliver their remaining sediment to the distal levee. Nevertheless, the paucity of F1 beds suggests that sedimentation on the proximal levee during Stage 2 was dominated by the fine grained, upper parts of throughgoing channel flows, and that anomalous large magnitude flows capable of delivering sediment to the distal levee were rare. The cumulative effect of these spatially varying depositional patterns continued to increase the differential relief between the proximal and distal levee.

5.5.3. Stage 3

Sedimentation rates were consistently greater on the levee than in the channel throughout stages 1 and 2 and, as a result, caused a continuous increase in channel relief, and, accordingly, confinement of the flows passing through the channel. Stage 3, in contrast, was characterized by the onset of channel filling that began to outpace levee growth. This progressive reduction in channel relief once again allowed the lower more energetic parts of flows to overspill onto the adjacent levee. However, due to the significantly reduced surface area of the proximal levee, in addition to the steep slope developed along its backside, the majority of overspill flows deposited only a small fraction of their sediment load

immediately outboard of the channel margin and bypassed most of their load on to the distal levee. These conditions contrast those of stages 1 and 2 where only the largest magnitude flows were able to deposit on the distal levee.

Deposition on the distal levee is interpreted to have resulted in stratal onlap onto the proximal levee. With time this eliminated the steep slope developed along the backside of the proximal levee while at the same time forming a more shallowly dipping, laterally extensive slope along the backside of the distal levee. Similar processes have been inferred previously for both deep-water (e.g. Kane et al., 2007) and fluvial channel systems (e.g. Cazanacli and Smith, 1998). At Castle Creek, it is interpreted that changes in the distal levee backside slope were controlled principally by the character of the bypassing flows, which in turn were influenced in part by the initial slope along the backside of the proximal levee. More specifically, the initial flows that bypassed on to the distal levee gained a significant amount of momentum by accelerating down the backside of the proximal levee, thus giving them greater run-out length. With time, deposition on the distal levee reduced the angle and shortened the length of the proximal levee slope, which as a result reduced the run-out distance of flows over the distal levee. Therefore, stratigraphically upward the upflow and downflow terminus of levee deposits in the distal levee are interpreted to have progressively backstepped towards the channel creating a surface that sloped downward and away from the channel. Of note is that the backstepping interpretation in the distal levee represents the average behavior of flows. In reality, deposition on the distal levee may have been more complex if local accumulation formed areas of local zones of topographic relief that later acted as areas of local bypass. Nevertheless, over time deposition on the distal levee muted these local irregularities and formed a smooth decreasing concave-up surface along the backside of the levee (Fig. 5.9F). In addition, during Stage 3 the surface of the proximal

levee progressively widened and flattened, ultimately creating a suitable surface on which the next channel-levee package formed.

Importantly, the interpreted mode of deposition on the distal levee is based on a laterally restricted (< 1 km-wide) dataset (compared to the typical km-scale of modern deep-marine levees) and, as a result is influenced by observations made in the proximal levee. Because easily traced sandstone marker beds are typically lacking in the mudstone-rich distal levee, the interpreted occurrence of horizontal strata and backstepping of bed termini in the distal levee can be considered speculative, and in need of further investigation. Accordingly, it is entirely possible that the full ~1 km-wide exposure in ICC3 reveals only the mostly horizontal crestal region of a larger levee complex, and that strata on the more distal levee flanks may in fact dip over several kilometers.

5.5.4. Recognizing levee backside slope in the field

Based on this study, the sloped surface that is theorized to form along the backside of the proximal and distal levee is not possible to delineate in the field. To do so requires the ability to walk along the majority of beds to their terminus. Even with the exceptional exposure in ICC3, this task is exceedingly difficult due to local cover, in particular the moraine that separates the POB from DOB. Furthermore, the occurrence of similar facies on either side of the hypothesized sloped surfaces makes recognition problematic in the field. Also, not all beds that terminate away from the channel would necessarily represent the location of either the proximal or distal backside levee surface. For example, unusually small magnitude, slow moving flows may not extend across the entire width of the levee and therefore terminate prematurely. Also, it is inferred that local topography developed on the levee, especially in the distal parts. Beds that terminate against this local topography would

not coincide with the location of the main sloped surface along the backside of the levee. Accordingly, the sloped morphology in Figure 5.9 represents only the overall average backstepping portion of bed termini due to increasing proximal levee relief and deposition on the distal levee.

5.6. Seismic modelling

The depositional model discussed above (Fig. 5.9) generates a well defined mound-shaped profile along the surface of a levee producing an overall geometry similar to that observed in seismic images. However, in this proposed depositional model strata are deposited sub-parallel to one another and to a paleohorizontal datum. Consequently, for this model to remain consistent with seismic data sets there must be other factors other than bed dip that produce the multiple internal dipping reflections commonly observed in seismic images.

One hypothesis for the development of these multiple, internal-dipping reflections, is that lateral changes in lithology and overall levee packaging are able to generate dipping seismic reflections. To test this hypothesis a schematic representing the main stratal attributes observed in Castle Creek was generated (Fig. 5.10A). Here, a 90 m-thick channel complex, made up of four vertically-stacked and laterally-offset channel-fill units, is juxtaposed against three 30 m-thick, laterally-offset levee packages. Each levee package consists of sand-rich lower and mud-rich upper parts. In the lower sand-rich part, individual thick-bedded turbidites thicken and then thin laterally away from the channel and transition rapidly into a mud-rich, thin-bedded lithofacies. This closely resembles the trends observed in ICC3 where thick-bedded F1 turbidites display a rapid lateral change into thin-bedded

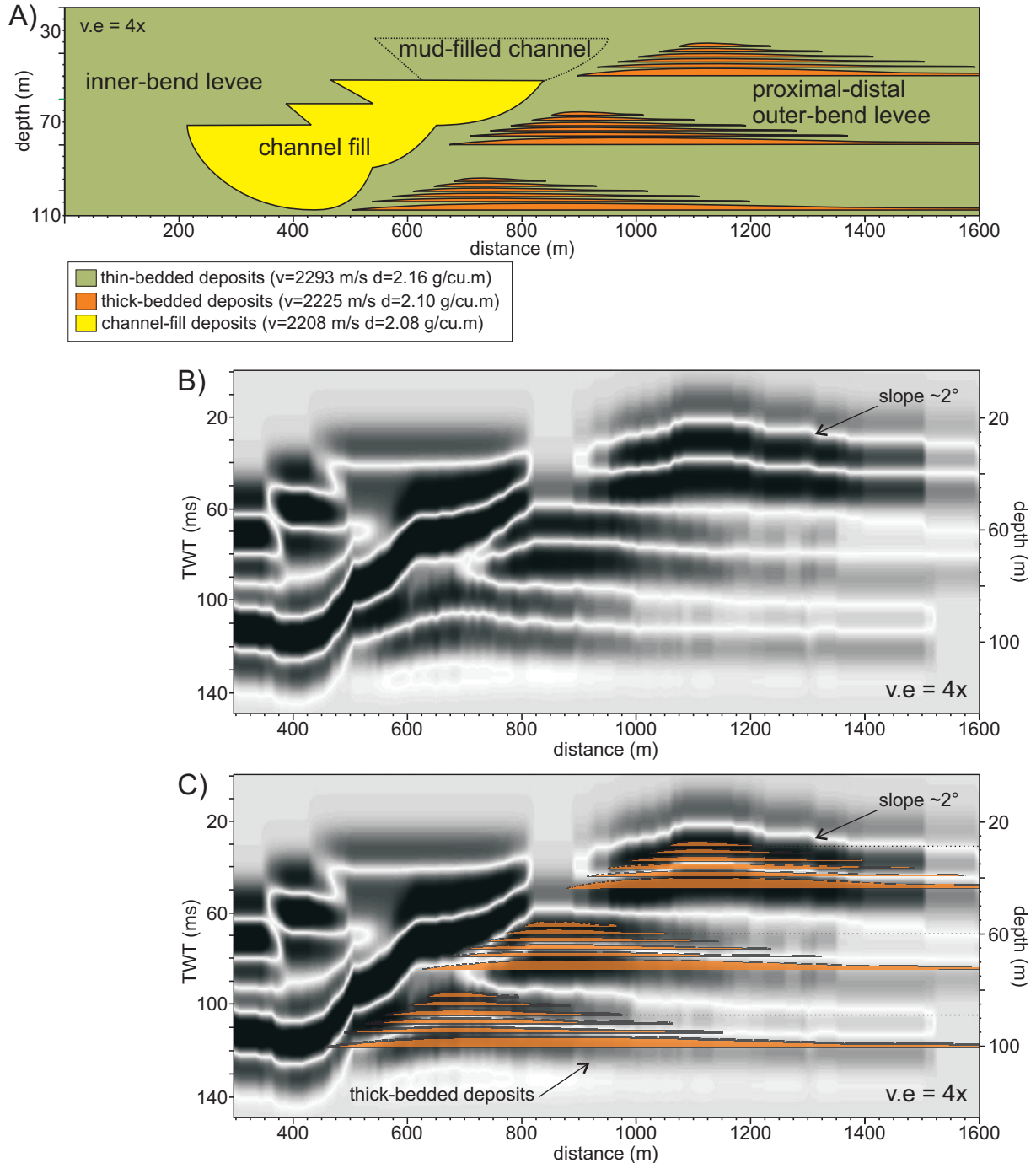


Fig. 5.10. A) Schematic illustrating main architectural attributes of ICC3 (compare with Fig. 5.4). Three lithologies were used: channel-fill (pure sand), thick-bedded (interbedded sand and mud 70:30), and thin-bedded (interbedded sand and mud 30:70). Each lithology was assigned different sonic velocities and densities based on forward modelling of ICC3 (E. Schwarz, 2008, personal communication) and are similar to values used in other seismic models (e.g. Batzle and Gardner, 2000). Note that thick-bedded deposits transition laterally into thin-bedded facies. B) Synthetic seismic section of (A) rendered with a 30 Hz centered frequency (SIGHT software developed by A. Pugin). Note that reflections locally dip at 2° . C) Interpretation outlining thick-bedded deposits in the lower part of levee packages. Note how seismic reflections cross-cut stratigraphy (indicated by dotted lines).

(F2) turbidites. As mentioned previously, the sandstone to mudstone ratio (sst:mdst) in the proximal levee of ICC3 is observed to initially increase up to 65:35 and then decrease to 42:58 over 300 m. Thereafter, sst:mdst continues to decrease in the distal levee to about 35:65. Therefore, on average sst:mdst decreases over 700 m from up to 65:35 in the proximal levee to as low as 35:65 in the distal levee. In Figure 5.10A this lateral trend is reflected by the upward reduced run-out length of the thick-bedded strata in the lower, sand-rich units, which in turn causes sst:mdst to initially increase away from the channel and then decrease over about 700 m.

For simplicity, each of the levee packages in Figure 5.10A have equal dimensions and are divided into upper and lower halves that are each 15 m thick. This does not completely mimic the complex stratigraphy observed in the levees of ICC3, where each package and its lower and upper parts vary in dimension, but it is a close approximation. Moreover, the entire levee succession in ICC3 occurs laterally adjacent to coarse-grained channel-fill deposits. In Figure 5.10A, on the other hand, only the lower two levee units are juxtaposed coarse-grained channel fills whereas the uppermost levee package occurs adjacent mud-rich channel-fill deposits. This latter situation simulates a levee that developed adjacent to a channel that was later abandoned and filled with mud. Prior to abandonment, however, throughgoing flows are interpreted to have deposited a levee package with a sand-rich lower part, and therefore similar to the two underlying levee packages. Including levee packages that are adjacent to both coarse-grained and mud-filled channels in the model was done to resolve any differences in seismic response between these two end-member scenarios.

In Figure 5.10A, the channel fill (pure sand), thick-bedded levee deposits (interbedded sand and mud 70:30), and thin-bedded mud-rich facies (interbedded sand and mud 30:70) were all assigned different densities and sonic velocities. Note that in the synthetic seismic model the margin of the channel is well defined and mimics the overall aggradation and lateral migratory trajectory of the channel (Fig. 5.10B). In the levee region each 30 m thick levee package generates two distinct reflections resulting in a total of six reflections for the entire levee succession. Each of these reflections share similar geometries where over the first few hundred meters laterally away from the channel they dip towards the channel and thereafter dip away from the channel toward the distal levee. These attributes are similar to those of the crest region observed in the deep-water seismic images of levees related to the Amazon Channel (Fig. 5.11). Importantly, the synthetic model produced in Figure 5.10B was generated from non-dipping strata and it is therefore interpreted that the individual reflections cross-cut stratigraphy (i.e. reflections cross-cut timelines) and do not represent true stratal dip (Fig. 5.10C). Instead, these reflections record the overall concave-down geometry produced by the lateral thickening and then thinning of the thick-bedded turbidites in the lower part of levee packages.

Another interesting observation in Figure 5.10B is that the channel margin appears to erode deeply into the lower two levee packages and as a result channelward-dipping reflections are not as well developed as those in the uppermost levee package. The main difference between the uppermost package and those below is that it does not occur immediately adjacent a coarse-grained channel-fill. Instead, it is laterally adjacent to mud which represents the fill of an abandoned channel. Therefore, the apparent erosion observed in the lower two units may in effect be an artefact of sonic wave interference between sand-rich channel margin strata and those in the levee. If the sharp density contrast between

channel and levee strata does cause some kind of interference, then it might explain the lack of well developed reflections dipping towards the channel like those associated with the lower channel-levee complexes in Figure 5.11. Alternatively, the lack of these reflections might simply record erosion of that part of the levee by the genetically-related, laterally-migrating channel.

The model depicted in Figure 5.10 does a good job in mimicking seismic attributes in the crest region of some deep-water levees; however, it does not accurately reproduce the long continuous dipping reflections typically imaged in the distal levee. This could result from the fact that the lateral sst:mdst trend in ICC3 is only documented over 500-1000 m due to local cover by a glacier. If, however, sst:mdst continued to decrease beneath the glacier for another few kilometers, and therefore be consistent with changes reported in the literature (e.g. DeVries and Lindholme, 1994; Clemenceau et al., 2000; Kane et al., 2007), then it is possible that the model would show multiple dipping reflections with a lateral scale and seismic character similar to those imaged along the backside of the levee in the Amazon Channel example (Fig. 5.11). Alternatively, it is possible that non-dipping strata occur only in the crest region of levees that in seismic images show a broad area of horizontal to sub-horizontal reflections. For instance, broad “flat” crests are recorded in levees associated with the Var deep-sea fan (Savoye et al., 1993) and even locally in the Amazon Channel system (Pirmez et al., 2000). Of note is that these horizontal reflections in these modern examples are interpreted to be continuous with dipping reflections in the distal levee.

The above two scenarios illustrate how the facies architecture of a levee at Castle Creek could generate seismic features typical of the crest region of modern deep-water levees. If this is in fact representative, then an important question that remains is, are strata in the proximal and distal levee physically continuous? If not, then a third possibility is that the

backstepping bed termini of horizontal, discontinuous strata proposed here have in fact been previously seismically imaged, for example the Danube Channel (Popescu, 2002, fig. 4.25a)

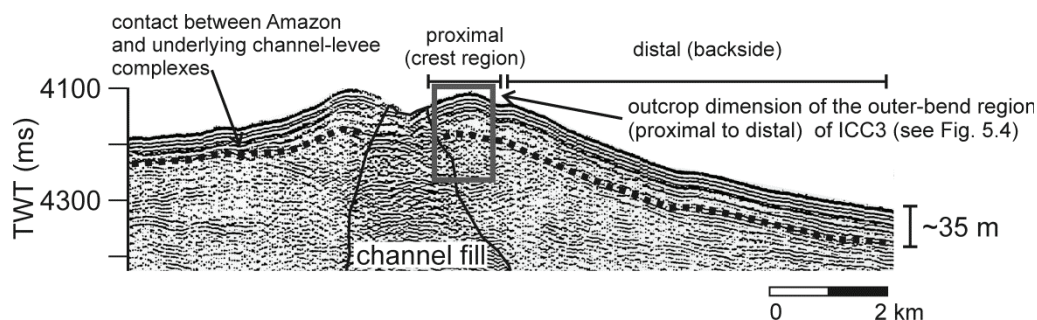


Fig. 5.11. Seismic reflection cross-section of the Amazon and underlying channel-levee complexes (modified from Skene, 1998). The levees adjacent the Amazon channel show a typical “gull-wing” geometry. Note that reflections in levees adjacent filled channels do not initially dip toward the channel like the ones associated with the younger Amazon channel.

and Laurentian Fan (the “hinge line” of Piper and Normark, 1982, fig. 3B). Note that in these two modern examples, the backstepping surface begins several kilometers away from the channel margin and thereafter extends for about 2–10 km. Notwithstanding the obvious differences in horizontal scale, the fundamental depositional processes described here are hypothesized to be applicable, at the very least locally, to depositional processes and architectures in some large-scale levee systems.

5.7. Conclusion

Dipping seismic reflections associated with deep-water levees are commonly interpreted to reflect the vertical stacking of beds that thin and fine progressively away from the channel. Field observations from the crest region of a Neoproterozoic deep-water levee example, however, suggest that topography associated with the lateral thinning of medium- to thick-bedded strata is infilled by fine, thin-bedded deposits and, as a result, levee beds

remain consistently sub-parallel to regional strike. To explain these observations, an alternative model for levee deposition, as proposed here, is that the mound-shaped geometry of levee crests is instead related to the upward and channelward backstepping of thin-bedded levee deposits. A synthetic seismic model was produced based on levee strata of ICC3 that successfully formed dipping reflections and, therefore, raises the possibility that dipping reflections in seismic data may not necessarily be the manifestation of dipping stratigraphy, but instead may record lateral changes in lithofacies within horizontal strata, such as those observed in ICC3 levee deposits. Moreover, this implies that important lateral changes in stratal continuity exist in channel-levee complexes, which in turn should have major implications on reservoir characterization.

CHAPTER SIX

Experimental modelling of beds composed of multiple non-climbing ripple cross-sets

6.1. Introduction

Channel-levee systems are significant geomorphic elements in the deep-water realm (e.g. Garrison et al., 1982; Prior et al., 1983; Flood and Damuth, 1987; Damuth et al., 1988; Abreu et al., 2003; Pirmez, 1994; Hiscott et al., 1997; Deptuck et al., 2003; Posamentier and Kolla, 2003) and their deposits are proven hydrocarbon reservoirs (e.g. Clemenceau et al., 2000; Kendrick, 2000). As a result, documentation of these deep-water features has garnered much attention. Although the majority of studies focus on channel fills, many contributions have been made detailing the sedimentology of levees (e.g. Beaubouef, 2004; Eschard et al., 2003; Grecula et al., 2003; Kane et al., 2007; Crane and Lowe, 2008; Hubbard et al., 2008; Kane and Hodgson, 2011; Khan and Arnott, 2011). A common sedimentary structure reported in levee deposits of both modern and ancient deep-marine systems is climbing ripple cross-stratification (e.g. Walker, 1985; King et al., 1994; Lyons, 1994; Coleman, 2000; Cronin et al., 2000; Basu and Bouma, 2000; Hickson and Lowe, 2002; Eschard et al., 2003; Grecula et al., 2003; Kane et al., 2007). Climbing ripples occur when rates of bed aggradation are high relative to rates of bedform migration (Allen, 1971, 1982a, b). These conditions are thought to be commonly achieved in levee settings, where throughgoing channel flows overtop the channel banks and become unconfined, resulting in rapid deposition (Kane et al., 2007). It is therefore somewhat perplexing that climbing and non-climbing ripple cross-stratification is scarce and abundant, respectively, in levee deposits in

the Castle Creek study area of the southern Canadian Cordillera, an area of much recent turbidite-related research (see Chapter 1) and the field area of the present study.

Numerous earlier studies investigating climbing ripples based on field observation, experiments, and theoretical consideration have been carried out in order to gain a better understanding of their formative processes (e.g. Jopling, 1960; McKee, 1965; Allen, 1971; Ashley et al., 1982; Rubin and Hunter, 1982; Jobe et al., in press). Allen's (1971) work is especially noteworthy in that it tested experimental and theoretical approaches and produced different kinds of climbing-ripple cross-stratified beds under steady flow conditions with constant rates of sediment fallout. In contrast, the experimental work of Ashley et al. (1982) generated thick climbing-ripple beds by varying both flow speed and sediment fallout rate in an attempt to simulate deposition from natural unsteady and non-uniform flows. The results of those studies clearly show that the steepness of climb depends on the ratio between sediment fallout and bedload transport. However, in the literature the presence of climbing-ripple cross-stratification is typically used as evidence of "high" rates of deposition, with little consideration of the other key variable, bedload transport rate (e.g. Grecula et al., 2003). Furthermore, the term "high" is qualitative and vague. How high, exactly, does the sediment fallout rate need to be before ripples start to climb? To gain insight into this question, the objective of this study is to conduct flume experiments at various sediment fallout rates and flow velocities and then use the results to estimate the time needed to deposit decimetre-thick beds of stacked non-climbing ripple cross-sets, similar to those observed in levee deposits at Castle Creek.

6.2. General geology of study area

The Castle Creek study area is located in the Cariboo Mountains (Fig. 6.1) where deep-water strata of the Neoproterozoic Windermere Supergroup are well exposed (Ross and Arnott, 2001). Vertically-dipping strata of the Isaac Formation, which makes up part of the Windermere Supergroup, are exposed in a section about 7 km parallel and 2.1 km perpendicular to regional bedding. Figure 6.2 shows only a 2.5 by 1.3 km portion of the Castle Creek study area. At Castle Creek the Isaac Formation consists of six major channel complexes bounded by genetically related overbank deposits (Ross and Arnott, 2007). Channel fills are up to 100 m thick and commonly consist of breccia, conglomerate, and coarse-grained sandstone turbidites. Levee deposits are composed mostly of interbedded sandstone and siltstone forming T_{a-d} turbidites. Channel complexes are informally termed Isaac Channel Complexes (ICC) 1–6 and are interpreted to form part of a base-of-slope succession in a deep-marine system (Ross et al., 1995; Ross and Arnott, 2007).

6.3. Isaac Channel Complex 3

Isaac Channel Complex 3 (ICC3) crops out as a roughly perpendicular-to-flow exposure through a highly aggradational deep-water channel where finer-grained strata are exposed on both sides of the channel-fill (Navarro et al., 2007; Khan and Arnott, 2011) (Fig. 6.3). The finer-grained strata adjacent the channel fills are interpreted to be levee deposits based on their sedimentology and divergent paleocurrent indicators (Khan and Arnott, 2011). This channel complex overlies an areally extensive 135 m thick mass transport deposit, the upper part of which consists of an up to 60 m thick mud-rich debrite (Arnott et al., 2011).



Fig. 6.1. Map of western Canada outlining the deep-water part of the Windermere Supergroup (shaded area). The Castle Creek study area is located in the Cariboo Mountains (circled).

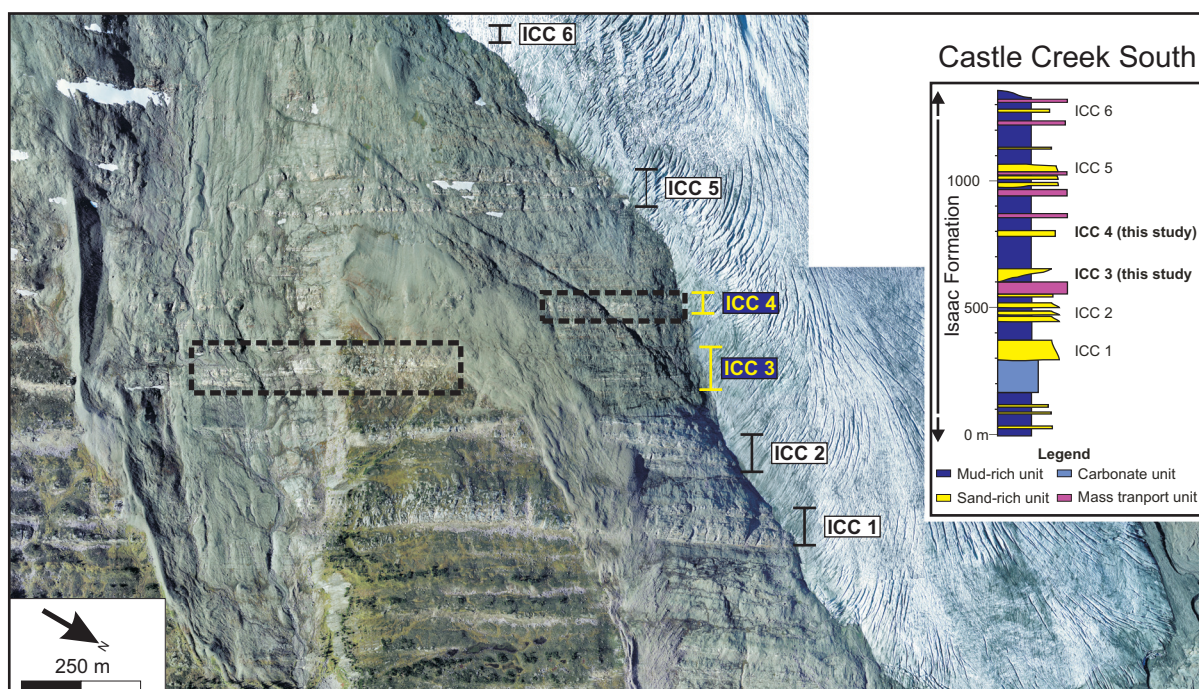


Fig. 6.2. Aerial photomosaic of the Isaac Formation exposed at Castle Creek south. Beds strike SE - NW and are nearly vertical. The Isaac Formation consist of channel-fill units encased in finer-grained levee deposits. The location of Isaac channel complex units 1-6 is shown (ICC1-ICC6). A dotted rectangle shows the location of the two channel complex units (ICC3 and ICC4) discussed in this chapter.

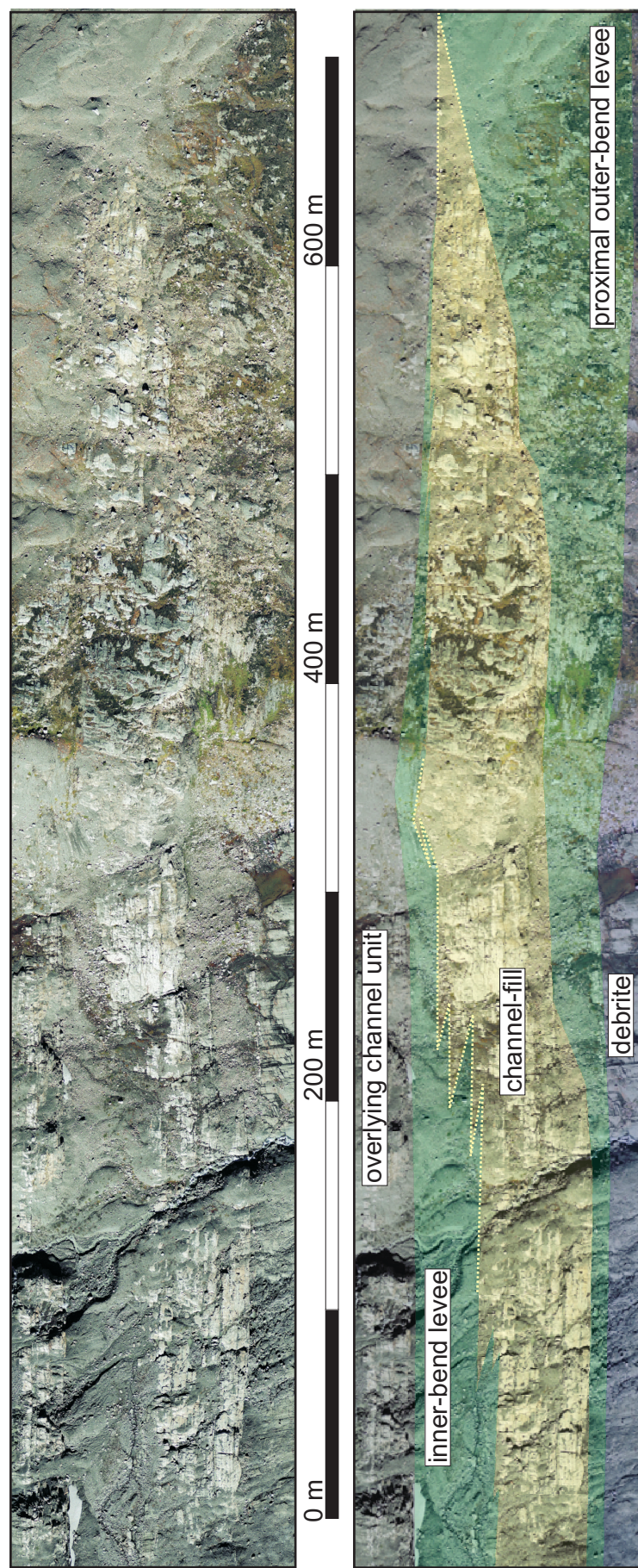


Fig. 6.3. Aerial photomosaic of Isaac Channel Complex 3 (ICC3). This unit crops out as a roughly perpendicular-to-flow cut through a sinuous, deep-water channel-levee complex where levee deposits are well exposed along both sides of the channel. Thick-bedded axial channel-fill deposits grade and interfinger laterally with inner-bend levee deposits that consist mainly of thin-bedded, T_{ed} turbidites. Along the outer-bend margin channel-fill deposits are in erosional contact with levee strata. Moreover, the terraced geometry of this erosional surface is interpreted to reflect the progressive lateral shifting of the thalweg towards the outer-bend side of the channel. Strata proximal to the outer-bend channel margin consist mainly of medium- and thick-bedded turbidites intercalated with thin-bedded turbidites. Thicker-bedded deposits thin rapidly over a few hundred meters away from the channel to thin-bedded deposits.

Based on the asymmetry of the facies architecture in the levees, ICC3 is interpreted to have been sinuous, and the deposits to the left and right of it are interpreted to have been deposited on its inner and outer-bend levees, respectively (Khan and Arnott, 2011). The inner-bend levee consists primarily of thin-bedded (<10 cm), T_{cd} turbidites whereas immediately adjacent the outer-bend margin medium- to thick-bedded (10-100 cm thick) T_{a-d} turbidites are commonly interbedded with thin-bedded T_{cd} turbidites (Fig. 6.4A). The abundance of thicker-bedded deposits along the outer- compared to the inner-bend margin is typically ascribed to fluid and sediment inertia that promotes the overspill of thick, high energy flows along the outer-bend of channels (Hay, 1987a,b; Piper and Normark, 1983; Imran et al., 1999; Straub et al., 2008). Thicker bedded deposits in the levee generally thin rapidly laterally over a few hundred meters and commonly become thin-bedded (<10 cm), T_{cd} turbidites that change little in thickness over hundreds of meters thereafter (see Chapter 4). Moreover, T_{cd} turbidites immediately adjacent the channel thin negligibly laterally.

Ripple cross-stratified (T_c) successions in ICC3 are common in thick- to thin-bedded levee turbidites, where they tend to overlie T_a or T_b sandstone divisions (Khan and Arnott, 2011). However, beds up to 40 cm thick consisting exclusively of multiple, vertically stacked, non-climbing ripple cross-stratification are also present (Fig. 6.4B). Individual ripple sets range from 1.5–3 cm thick and in near parallel-to-flow sections are spaced 10–20 cm apart. Ripple cross-stratified (T_c) divisions in ICC3 levees typically grade upward from medium- to fine-grained sandstone, and they are commonly overlain abruptly by a thin massive or laminated siltstone layer.

Thin-bedded deposits, which are the most common facies in overbank strata at Castle Creek, consist mainly of a 1–8 cm thick T_c division at their base (Fig. 6.4C). These T_c divisions comprise single to multiple (up to four) non-climbing ripple cross-stratified sets.

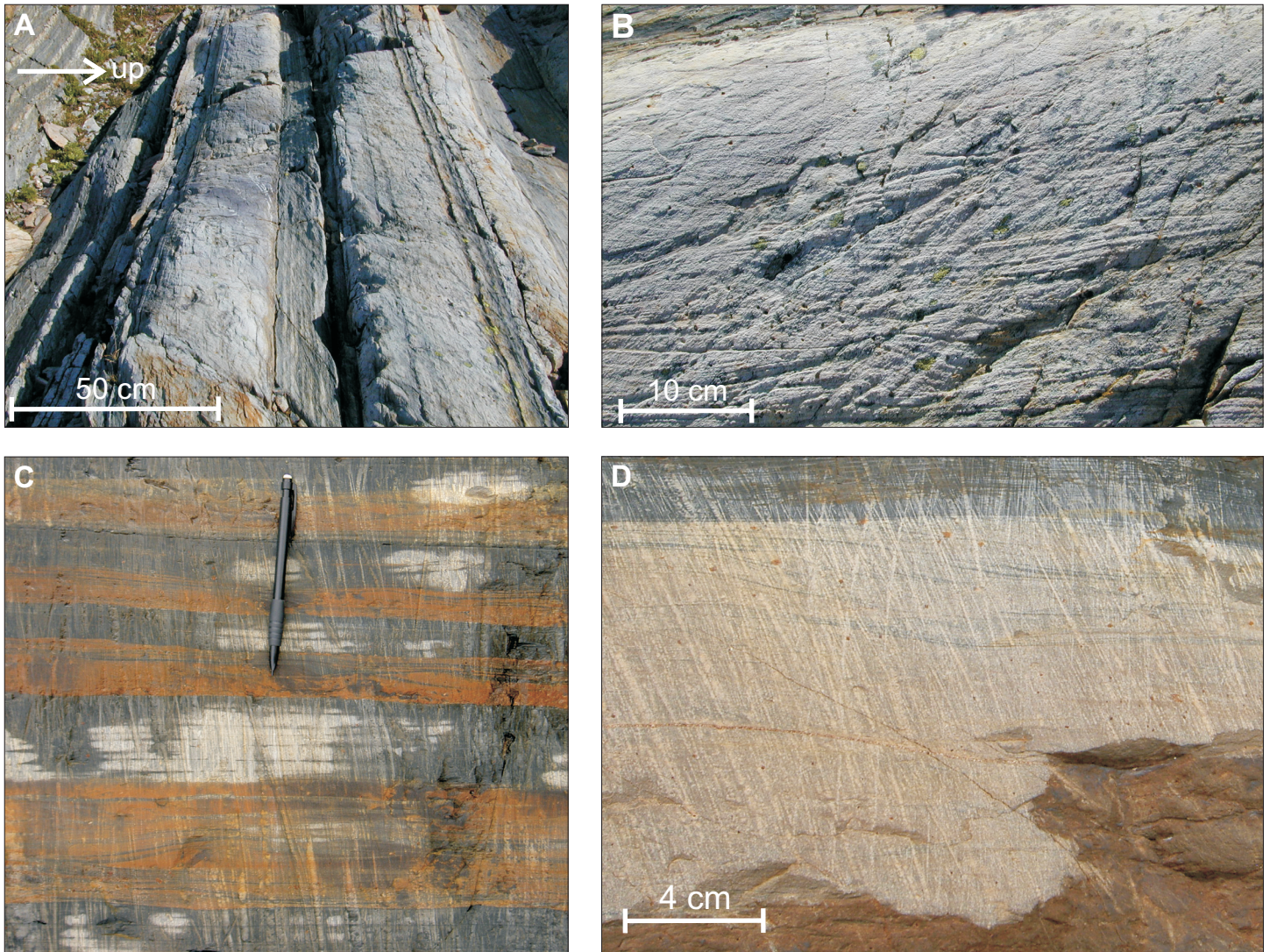


Fig. 6.4. Strata observed in the outer-bend margin of ICC3. A) Medium- and thick-bedded, T_{a-d} turbidites intercalated with thin-bedded, T_{cd} turbidites. B) A thick-bedded turbidite consisting of vertically stacked sets of non-climbing ripple cross-stratification. C) Thin-bedded, T_{cd} turbidites commonly observed a few hundred of meters away from the outer-bend channel margin in ICC3. This facies is also the most common in the Isaac Formation at Castle Creek. D) Thin-bedded T_{cd} bed that grades upward from fine-grained sandstone to siltstone. Note that in the uppermost part of the T_c division individual foresets are interlaminated with finer-grained sandstone or silt drapes.

Ripple cross-stratified (T_c) divisions in thin-bedded deposits typically grade upward from fine-grained sandstone to siltstone and in their upper part consist of sandstone foresets with common finer-grained sandstone or siltstone drapes (Fig. 6.4D).

6.4. Isaac Channel Complex 4

The upper part of ICC4 (Fig. 6.5) is interpreted to be an example of an oblique cut through a sinuous, highly aggradational channel that is similar to ICC3 (Mussa-Caleca, 2008). Unlike ICC3, however, the outer bend side of the channel is covered by a glacier. The inner bend channel margin, on the other hand, is well exposed and characterized by an interdigitate pattern of channel-fill strata that obliquely-upward pinch out abruptly into their conterminous levee deposits. A similar pattern is observed in ICC2 and is interpreted to result from processes associated with deposition along the inner bend of deep-water sinuous channels (Arnott, 2007b). Locally adjacent the inner bend channel margin of ICC4 there is a 8 m thick and 100 m wide levee succession that consist of very thin-bedded (<3 cm) T_{cd} turbidites intercalated with thicker-bedded T_{cd} turbidites that are up to 50 cm thick but on average are 10-20 cm thick. The thicker-bedded deposits thin rapidly laterally over about 100 m to less than 10 cm. Moreover, in this channel-proximal area the T_c division of these thicker-bedded strata makes up to 90% of the total bed thickness and consist of multiple climbing-ripple cross-stratified sets that individually measure about 2–4 cm thick (Fig. 6.6A and B). Moreover, the T_c divisions in these beds typically fine upward from fine-grained sandstone to siltstone and, like the thin-bedded deposits in ICC3, become progressively more interlaminated with siltstone drapes upward. Here the apparent angle of climb averages about 14° ; however, based on the common preservation of stoss-side lamination the original angle was probably on the order of at least 20° (see Allen, 1973). The lower angle now preserved in the field is most likely due to the oblique cut of the outcrop face and compaction effects. Also, as a result of the oblique cut of the outcrop the average measured ripple spacing of 15 cm is greater than the true value. The two-dimensionality of the outcrop made it impossible

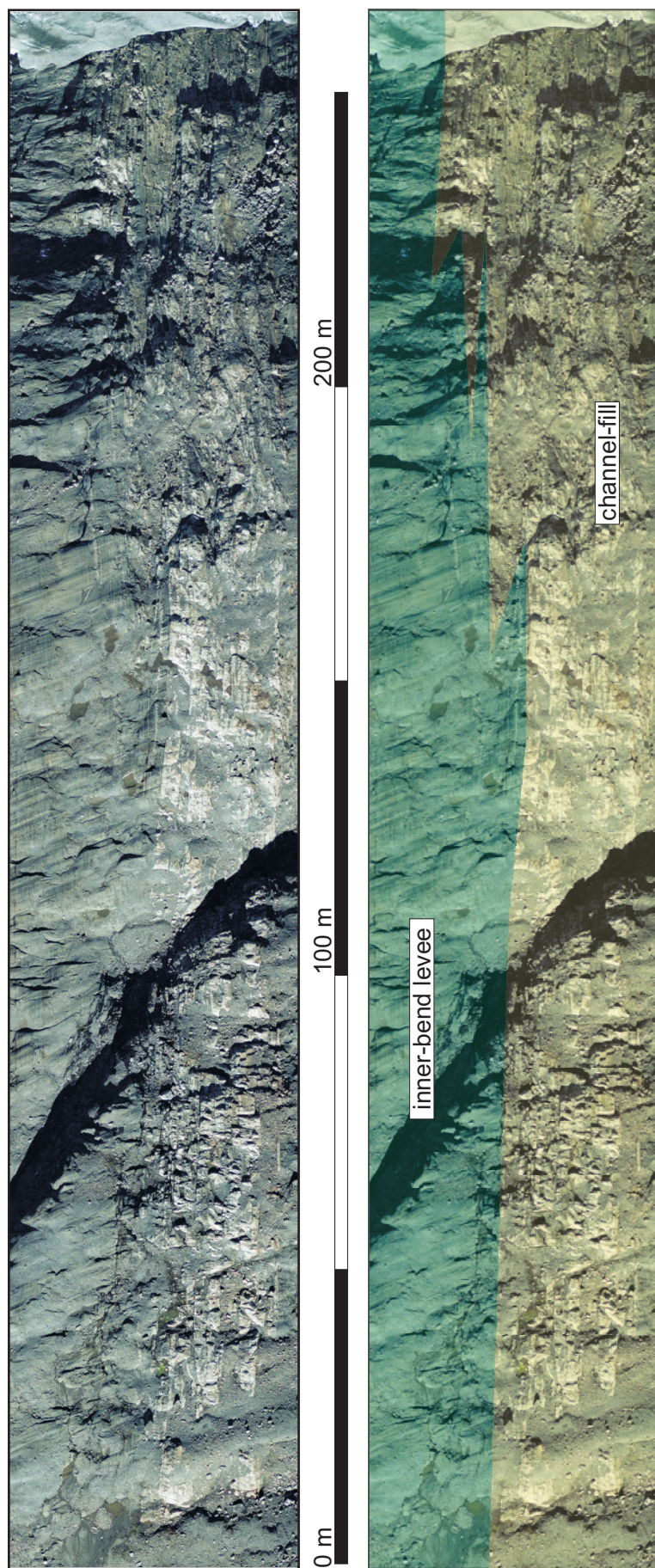


Fig. 6.5. Aerial photomosaic of the upper part of Isaac Channel Complex 4 (ICC4). This unit represents an oblique cut through a sinuous, deep-water channel-levee complex where levee deposits are well exposed along the inner-bend side of the channel. The inner-bend interpretation is based on the interfingering pattern of channel-fill and levee strata. Also, the channel aggrades and migrates laterally away from the interpreted inner-bend levee, suggesting lateral shifting of the channel axis towards the outer-bend side of the channel. Inner-bend strata consists locally of medium- and thick-bedded, T_{cd} turbidites interstratified with thin-bedded T_{cd} turbidites.



Fig. 6.6. Inner-bend levee strata of ICC4. A) Thick-bedded turbidite consisting of stacked sets of climbing-ripple stratification. B) Medium-bedded, T_{cd} turbidite that fines upward from fine sandstone to siltstone. Note that in the upper part of the bed, individual climbing foresets are interlaminated with finer-grained sandstone or silt drapes. The average angle of climb of ripple stratification in ICC4 is 14° , however, due to the obliquity of the outcrop face and compaction this angle represents a minimum value. Moreover, note the common preservation of stoss-side lamination, which suggests that the true angle of climb is at least 20° based on the studies of Allen (1973).

to accurately measure paleocurrents; nevertheless, ripples appear to trend roughly NW (toward the channel). Of note is that climbing-ripple cross stratification is rare throughout most of the Castle Creek study area and even in ICC4 it is only locally observed.

6.5. Climbing ripples

Allen (1971) has shown that the tangent of the angle of ripple climb (δ) is directly proportional to the rate at which the bed aggrades (V_Y) and inversely proportional to the speed of ripple migration (V_X) (Fig. 6.7). The bed aggradation rate (V_Y) and the ripple migration speed (V_X) are respectively measured normal and parallel to the generalized depositional surface. The vertical component (V_Y), or net deposition from suspension on the rippled bed, is expressed as

$$V_Y = \frac{R}{\gamma} \quad (6.1)$$

where R is the average sediment deposition rate measured in units of mass over unit time and area and γ is the bulk density of deposited sediment. Allen (1973) showed that stoss-side erosional climbing ripples form when the rate of deposition is low relative to bedload transport whereas high relative depositional rates result in preservation of stoss-side lamination (Fig. 6.7). Compared to the vertical component (V_Y), the velocity of travel of the ripple (V_X) is decomposed into two components denoted $V_{X(1)}$ and $V_{X(2)}$. The first of these, $V_{X(1)}$, represents bedload transport in the absence of net deposition upon the bed and is therefore related to the flux of sediment contained within the ripple through a fixed plane normal to the streamwise direction. Bagnold (1954) and Simons et al. (1965) showed that the bedform migration rate for ripples with a triangular profile (Fig. 6.7) is expressed as

$$V_{X(1)} = \frac{2J_b}{H\gamma} \quad (6.2)$$

where J_b is the bedload transport rate in mass per unit time and width of flow and H is the ripple height.

Allen (1971) justified the importance of including a second contribution to V_x which depends on the local non-uniformity of flow over a ripple. This local non-uniformity is created because flow is compressed downstream from trough to crest and therefore local mean flow velocity and bed shear stress increases (Raudkivi, 1963, 1966; Henry et al., 1964; Vanomi and Hwang, 1967; Migeon et al., 2000). In contrast, local mean flow velocity and bed shear stress decrease in the downstream direction from crest to trough. Therefore, providing there is available sediment in suspension for net deposition, the average rate of deposition on the upstream ripple face is smaller than on the downstream ripple face. Based on experimental observations, Guy et al. (1966) suggested that the equilibrium shape and dimension of ripples are independent of flow conditions. This theory has been supported by numerous other studies (e.g. Harms, 1969; Jopling and Forbes, 1979; Richards, 1980; Middleton and Southard, 1984; Allen, 1985; Arnott and Hand, 1989; Baas, 1994). Allen (1971) argued that if the shape of equilibrium ripples is independent of flow conditions then any additional sediment deposited on the leeward face must accumulate in such a way as to contribute additively to the downstream migration of the ripple. Hence

$$V_{X(2)} = \frac{L \cdot \Delta r}{H\gamma} \quad (6.3)$$

where L is the length of the leeward face, and Δr the difference in the mean rate of sediment deposition on the leeward and stoss side.

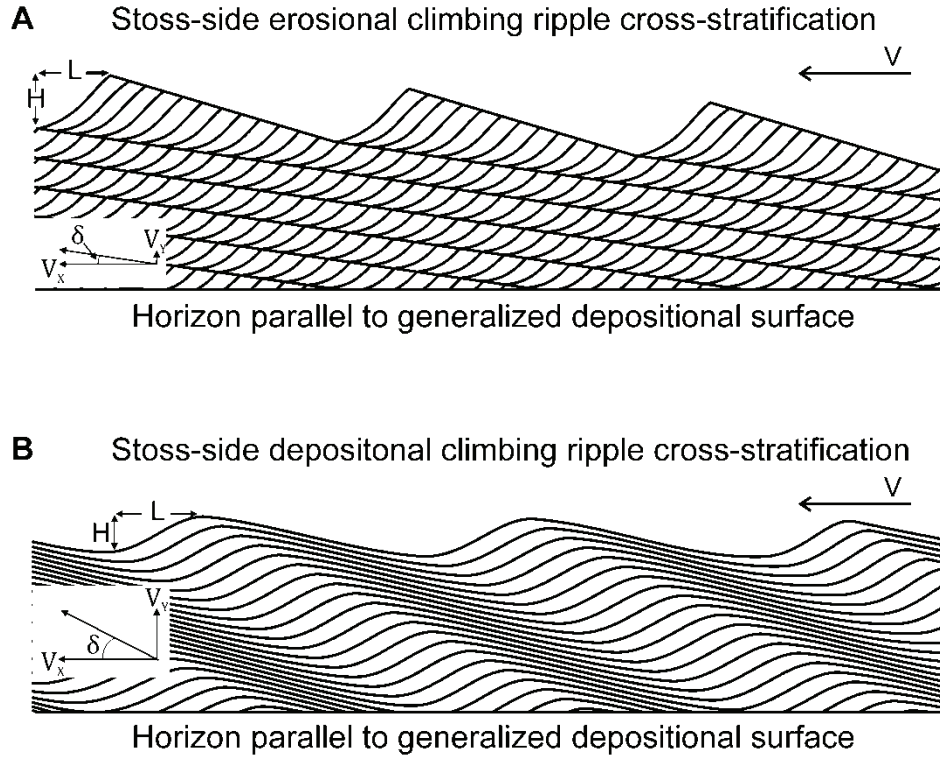


Fig. 6.7. Diagram illustrating climbing-ripple parameters. A) Cross-section of stoss-side erosional climbing ripples. B) Cross-section of stoss-side depositional ripples. V is direction of ripple migration; L is length of lee side; H is ripple height; V_y is depositional rate; V_x is the ripple migration rate.

6.6. Methodology

6.6.1. Equipment and set up

Seven climbing ripple runs were conducted in an approximately 10 m long by 0.6 m wide by 0.6 m high recirculating flume (Fig. 6.8). Water discharge, and accordingly flow speed, was controlled by adjusting the speed of the pump. Moderately well sorted ($\sigma = 0.53$), strongly skewed (0.60) sand with a median grain diameter of 0.25 mm was used in these experiments (Table 6.1). Prior to each run, dry sand was poured into four bins that rested along rails on top of the flume. Each bin was 1.2 m long, 0.6 m wide and 0.3 m high. A metal

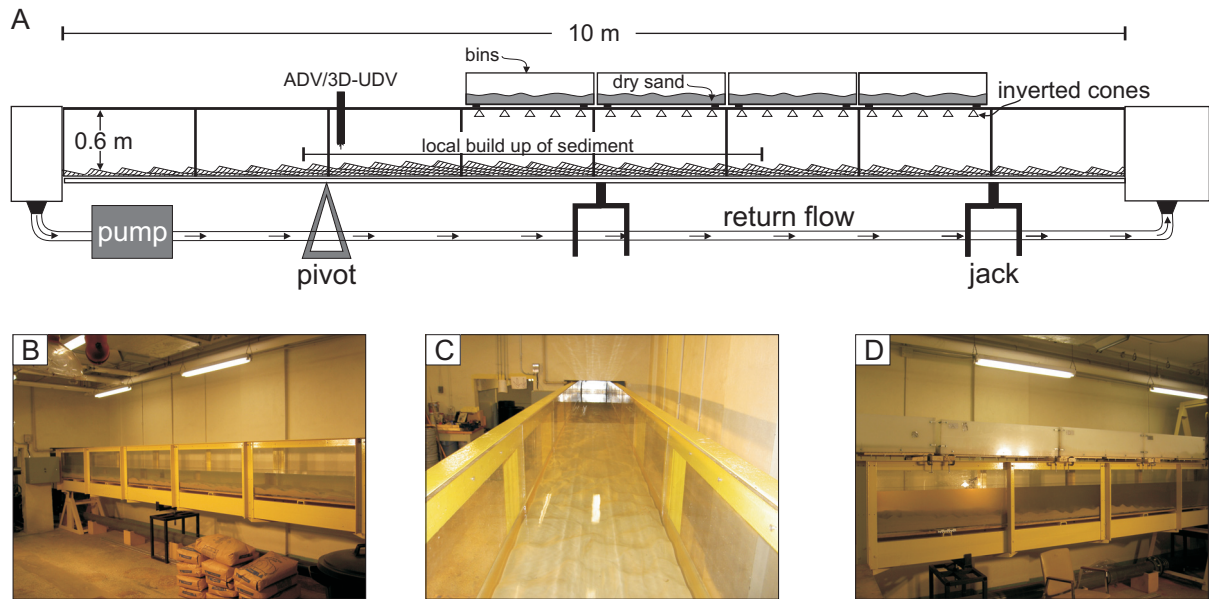


Fig. 6.8 Cross-sectional schematic and photos of the flume. A) Schematic illustrating experiment set-up. B) Side view of flume. C) Plane view of flume. D) Side view of flume with bins placed along the top.

Table 6.1
Grain size analysis of experimental sand

Class interval (Phi)	Class interval (microns)	Wt. %
<1.0	>500	0
1.0-1.5	500-354	12.17
1.5-2.0	354-250	49.84
2.0-2.5	250-177	19.89
2.5-3.0	177-125	12.34
3.0-3.5	125-88	5.36
3.5-4.0	88-63	0.4
>4.0	<63	0
Total		100
mean	250	

grate overlain by a thin sheet of plastic lined the base of the bins. These plastic sheets were pierced with eighty, more or less evenly distributed, 5 mm diameter holes. Various numbers of holes were covered to achieve the desired fallout rate, which ranged from 0.00009 cm/s and 0.003 cm/s (bed aggradation rate). A series of inverted cones were placed below the bins to help distribute sediment more uniformly before contacting the water surface. Additionally, sediment was further dispersed by flow turbulence, ultimately causing deposition to be.

relatively uniform over the bed in the streamwise direction. The bed-surface fallout area from the overhead bins, however, was about 10 cm less than the width of the flume and caused a slight increase in sediment accumulation away from the flume sidewalls. This was most notable during runs with high rates of fallout and for those runs data from the center of the flume were used instead of near the sidewalls. Also of note is that the combined length of the overhead bins was about 5 m and as a result sediment accumulated in a roughly trapezoidal shape (Fig. 6.8). Notwithstanding, uniform deposition rates were well maintained locally over an area of about 3–4 m long, and it is this area that was used for analysis.

Velocity for runs 2–6 were measured at the downstream section of the study area using an acoustic Doppler velocimeter (ADV) whereas runs 1 and 7 used a three-dimensional ultrasonic Doppler velocimeter (3D-UDV). A velocity profile was measured prior to runs 2–6 using the ADV, which then was fixed at the appropriate flow depth to record the mean velocity during the run. No water was added to the flume during the runs and as a result velocity gradually increased throughout the run as the bed aggraded and decreased the average mean depth. In runs with low bed aggradation rates the velocity was allowed to increase gradually and was at most 5% faster at the end of a the run (Fig. 6.9). During runs with higher aggradation rates the velocity was adjusted throughout the run to stay within 5% of the initial velocity. The 3D-UDV was used for Run 1 and Run 7, which had the lowest and highest aggradation rates, respectively. The advantage of the 3D-UDV is that it instantaneously measures a continuous velocity profile from the bed surface to near the water surface. Noteworthy, is that bed shear stress changed negligibly during Run 1, which had a low sediment fallout rate. In addition, adjusting the velocity during Run 8 was sufficient to keep bed shear stress to within 10% of the pre-run value. Extrapolating these results to the

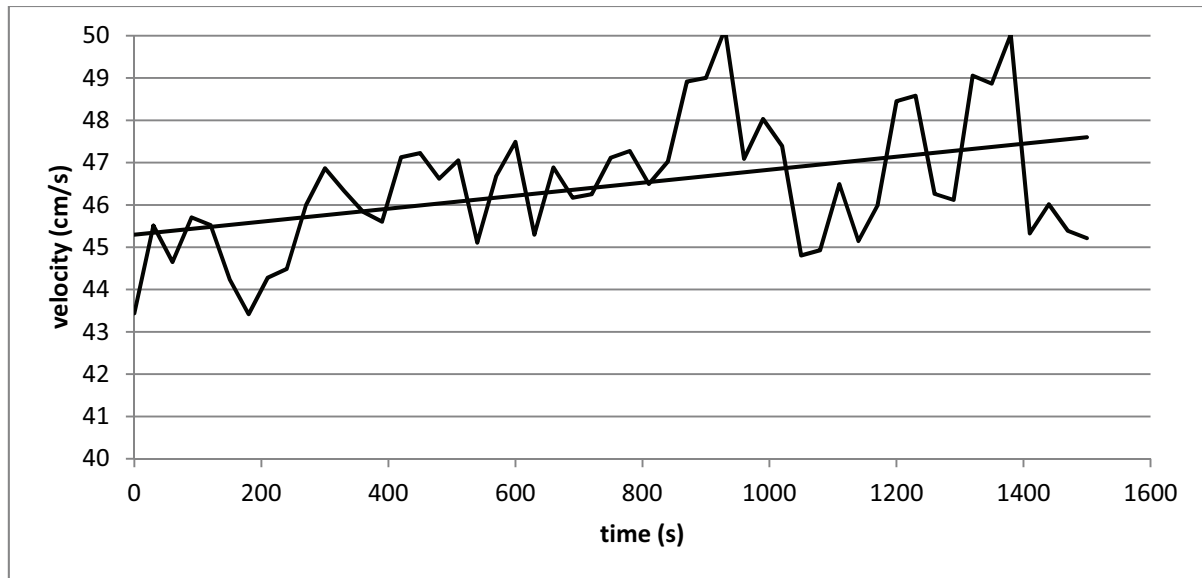


Fig. 6.9. Real time velocity profile of Run 3. Note that the velocity increases slightly throughout the run as a result of bed aggradation and decreased flow depth.

other runs suggests that bed shear stress and hence bedload transport rate varied little during individual runs.

Experiments in this study were not intended to replicate or confirm previous findings (e.g. Allen, 1971), but instead attempted to define the threshold conditions under which ripples start to climb at measurable angles. Intuitively, non-climbing ripples would most likely only occur under low rates of aggradation and high velocities. Accordingly, to simplify the experiment, and also lower the number of necessary runs, a narrow range of relatively high velocities (43-49 cm/s) was used. These flow speeds are 6–12 cm/s below the ripple to dune threshold for the flow depths used here (e.g. Southard and Boguchwal, 1990). As a result runs always remained within the ripple field. Of note, however, is that the observed relationship between flow speed and ripple migration rate did not correspond well with results from previous experiments (see Section 6.6.3). In addition to ripple migration and

aggradation rate, sediment concentration was measured 3–5 times during each run using a 500 ml bottle sampler.

6.6.2. Procedure

A sand bed approximately 1.5–5 cm thick was laid down along the flume prior to each run. Initial water depth was about 18 cm. Flow velocity was then set and equilibrium (i.e., 3D) ripples (cf. Baas, 1994) were allowed to develop (Fig. 6.10). After achieving equilibrium a pre-measured volume of sand was added to the overhead bins which then were monitored to ensure that holes remained open and that sediment fallout remained relatively constant. High definition camcorders were used to record each run. Aggradation rate was calculated by measuring the average temporal increase in bed thickness. Bedform migration rates were estimated from the average migration speed of 4 to 12 ripple crests during the runs as captured through the sidewall on video. At the end of each run the flow was stopped and cross-section photos of the bed taken. The flume was then slowly drained and allowed to stand for several days in order to reveal more detailed internal stratification. For high fallout runs, sediment between the sidewalls and center of the flume was excavated and then the exposed bed was photographed.

6.6.3. Ripple transport rate

Ripple migration rates measured in this experiment (Table 6.2) are lower than those measured by earlier workers that used fine- to medium-grained sand (e.g. Allen, 1971; Baas, 1994). The slower rates measured here are likely because ripple migration rate was measured through the sidewalls where sidewall effects reduced flow speed (Vanoni and Brooks, 1957; Williams, 1970; Cheng and Chua, 2005). Accordingly, in addition to the mean streamwise

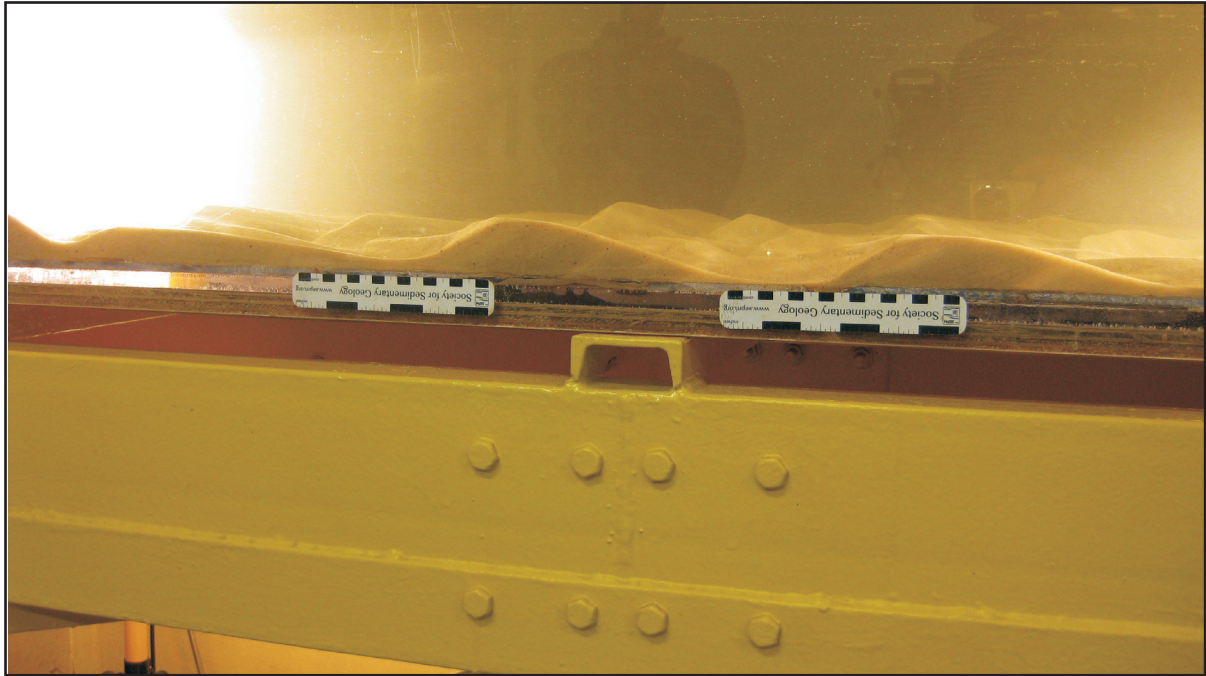


Fig. 6.10. An example of the equilibrium ripples formed prior to each run. Most runs started with a thicker initial sediment bed than shown here.

Table 6.2
Summary of experimental results

Run	Time (sec)	Flow velocity (cm/sec)			Ripple migration rate (cm/sec)	Depositional rate (cm/sec)	Climb angle (°)		Concentration by volume
		measured	adjusted	theoretical			mean	theoretical	
1	1800	46	31	25-36	0.038	0.000086	0	>1°	0.0002
2	2640	45	29	20-30	0.0216	0.0010	6	3°	0.0001
3	1800	49	28	19-28	0.0173	0.0010	4	3°	0.0002
4	1440	45	26	15-22	0.0093	0.00069	6	4°	0.0001
5	510	46	28	18-27	0.015	0.0034	14	13	N/A
6	1560	44	26	15-22	0.008	0.0023	13	16	0.0005
7	1200	46	26.5	16-24	0.0105	0.0034	17	18	0.0009

[†]velocity based on Allen's (1970) experimental results

^{††}velocity calculated using Bagnold's (1966) equation for bed load transport

velocity measured at the center of the flume, Table 6.2 also includes adjusted velocities based on experimental results from Allen (1971), where ripple migration rates were presumably measured away from the flume sidewall (Fig. 6.11). Migration rates in this experiment were adjusted using Allen's (1971) results because the grain size used in that experiment were similar.

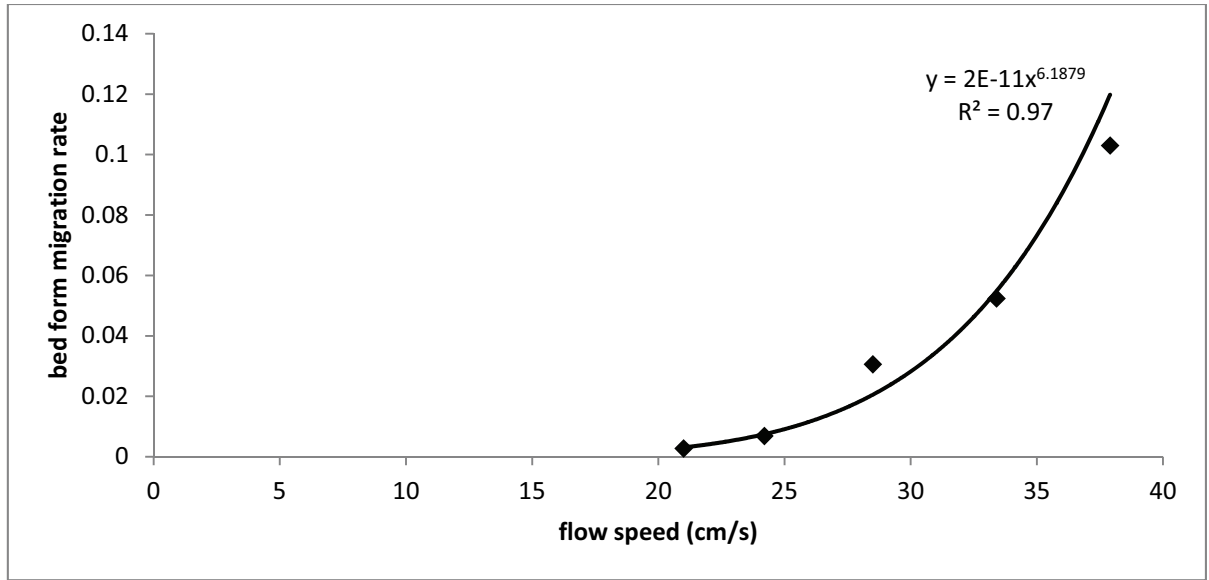


Fig. 6.11. Plot of Allen's (1970) bedform migration rates against flow speed. This plot is well described by a power-law relationship. Measured flow speeds in Table 7.3 were adjusted using this relationship.

In addition to the measured and adjusted mean flow velocities, Table 6.2 also includes a theoretical velocity range based on Bagnold's (1966) equation for bedload transport:

$$j_b = \frac{kf\sigma\rho u^3}{8(\sigma - \rho)g \cdot \tan\alpha} \quad (6.4)$$

where j_b is the bedload transport rate, k is the bedload efficiency factor, f is the Darcy-Weisbach bed friction coefficient, σ is the density of the sediment particles, ρ is the fluid density, u is the mean flow velocity, g is the acceleration due to gravity, and $\tan\alpha$ is the coefficient of friction of the transported particles. Equation 6.2 shows how bedload transport is related to bedform migration rate ($V_{X(1)}$). It should be noted that that $V_{X(1)}$ is a good estimation of the total bedform migration rate (V_X) for climb angles of less than 30° (cf. Allen, 1970), and therefore $V_{X(2)}$ is not considered here.

The grain size and velocities used in the experiments indicates a bedload efficiency factor of 0.14 (Bagnold, 1966). The Darcy-Weisbach bed friction coefficient for a bed of ripples at a flow depth of a few decimetres varies from 0.03–0.10 (Guy et al., 1966; Allen, 1971). Quartz sand was used in this experiment and hence σ is equal to 2.65 g/cm³ and ρ is assumed to be 1 g/cm³. Moreover, Bagnold (1966) demonstrated that for the grain sizes used in this experiment, the coefficient of friction is constant at 0.75. Collectively, these values have been used to calculate the theoretical mean flow velocities in Table 6.2. Note that the theoretical velocities calculated for different bedload migration rates agrees well with the adjusted velocities based on results presented by Allen (1971). One of the initial aims of this experiment was to use relatively high flow speeds to help minimize climb angle; however, as discussed above, the bedform migration rates measured here most likely corresponded with intermediate velocities (26–31 cm/s) thought to have occurred close to the sidewall window due to frictional retardation. Accordingly, the adjusted velocities in Table 6.2 are used in the following results and discussion instead of the actual measured mean flow velocities measured at the center of the flume.

6.7. Results

Table 6.2 summarizes data measured during each run. Based on climb angle runs were subdivided into three categories: 1) high climb (mean 13–25° – Runs 5, 6, and 7), 2) moderate climb (mean 4–6° – Runs 2, 3, and 4), and 3) low climb (mean <1° – Run 1). Generally, ripple dimensions changed negligibly during each run. The average ripple height and length over all runs combined were 1.6 cm and 14.8 cm, respectively. Runs 7, 3, and 1 are selected as being representative of each group and are discussed in detail below.

6.7.1. Group 1: high climb

Run 7 lasted 20 minutes with an adjusted flow velocity of 27 cm/s and a ripple migration rate of 0.011 cm/s. Observed fallout rate and suspended sediment concentration were 0.0034 cm/s and 0.0015% by volume, respectively. Generally, these conditions produced stoss-side-erosional ripples (cf. Allen, 1973). Laminae deposited and preserved on the stoss side were rare but where observed were noticeably thinner than those on the leeward side – an observation reported earlier by Allen (1971,1973). The observed climb angle of ripples averaged 17° ranged from 10–25° (Fig. 6.12). The theoretical angle of climb (V_Y/V_X) for this run was calculated to be 18°.

6.7.2. Group 2: moderate climb

Run 3 lasted 30 min with an adjusted flow velocity of 28 cm/s and an average ripple migration rate of 0.0173 cm/s. Fallout rates and suspended sediment concentrations during this run were 0.0010 m/s and 0.0003% by volume, respectively. This run created multiple sets of stacked ripples with no stoss side preservation. Angle of climb averaged about 4°, but ranged up to 10° (Fig. 6.13). In addition, several ripples had no discernible angle of climb whereas others migrated down at a negative angle of climb up to -6°. During the run it was difficult to visually discern that Group 2 ripples were indeed mostly climbing, a fact made evident only after post experiment ripple measurements. Difficulty in assessing the climbing nature of these bedforms was most likely due to the relatively low average angle of climb and also the simultaneous occurrence of both upward (positive) and downward (negative) migrating ripples.

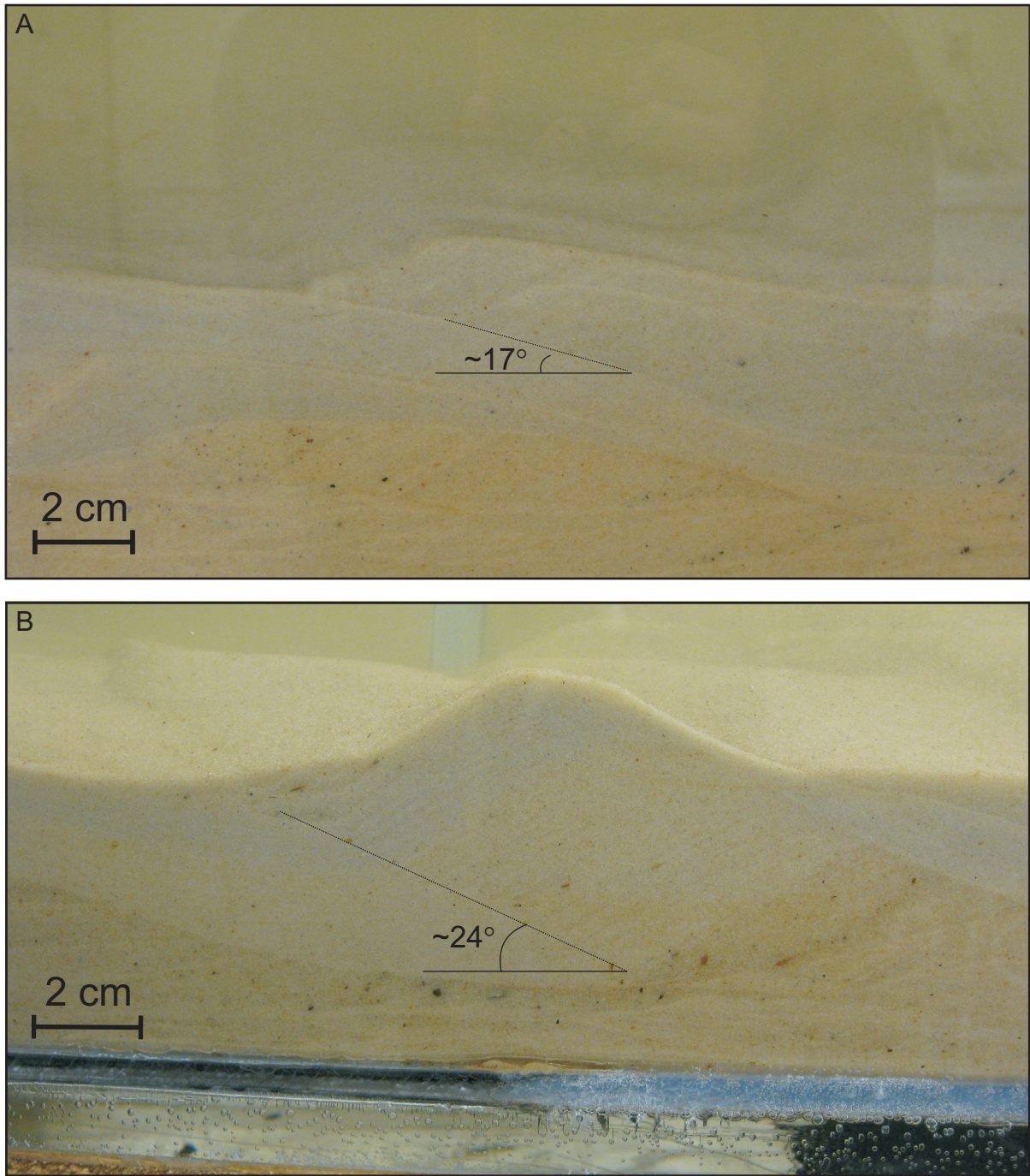


Fig. 6.12. Run 7 from Group 1 (high climb). A) Photo showing average climb of 17°. B) Photo showing a local climb angle of 24°. Note well developed forests. Lower ~2.5 cm of sediment represents the initial bed.

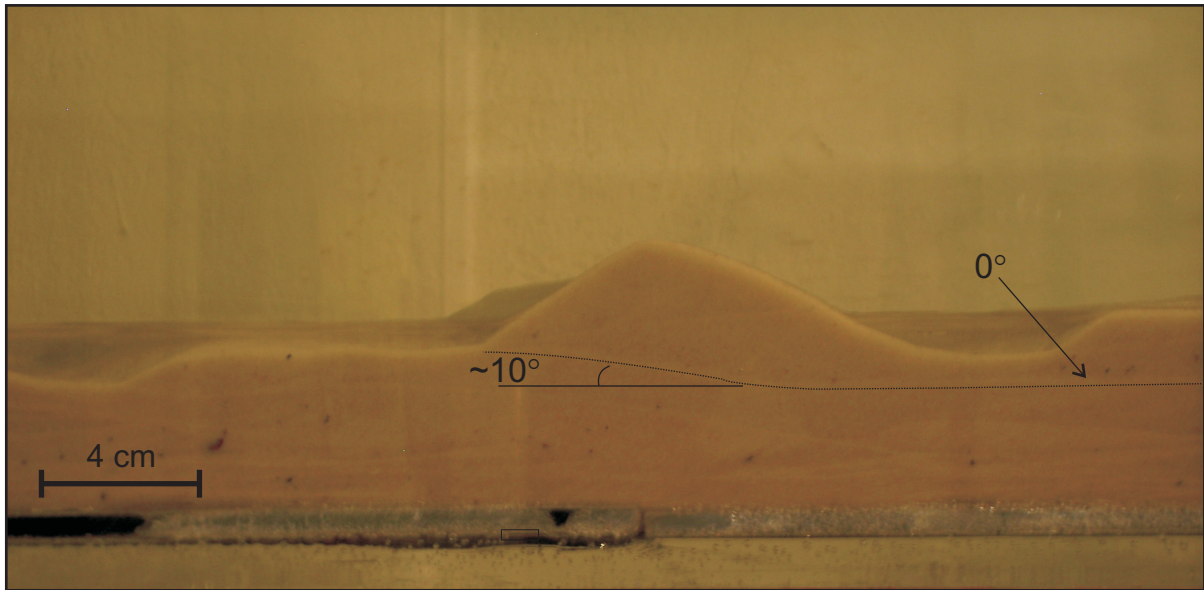


Fig. 6.13. Run 3 from Group 2 (moderate climb). Ripple in the middle of photo climbs at an angle of 10° , however, the ripple immediately up flow does not climb. The average climb angle measured in this run was 4° . Lower ~ 2 cm of sediment represents the initial bed.

6.7.3. Group 3: low climb

Run 1 lasted 30 minutes with an adjusted flow speed of 31 cm/s and an average ripple migration rate of 0.038 cm/s. Fallout rate was 0.000086 cm/s and measured suspended sediment concentration by volume 0.0003%. Such low rates of sediment fallout (an order of magnitude less than any other run) caused the stoss side of almost every ripple to be eroded by the ripple immediately upflow. As a consequence, an individual ripple set rarely extended more than 20 cm downflow. The majority of basal bounding surfaces of ripple sets were sub-parallel to one another with no apparent or measurable angle of climb (Fig. 6.14). Similar to Group 2, there were a few individual ripples that locally climb at both positive and negative angles. Moreover, the theoretical angle of climb of this run was less than 0.13° – an angle that would be difficult to discern even in the controlled environment of the flume.

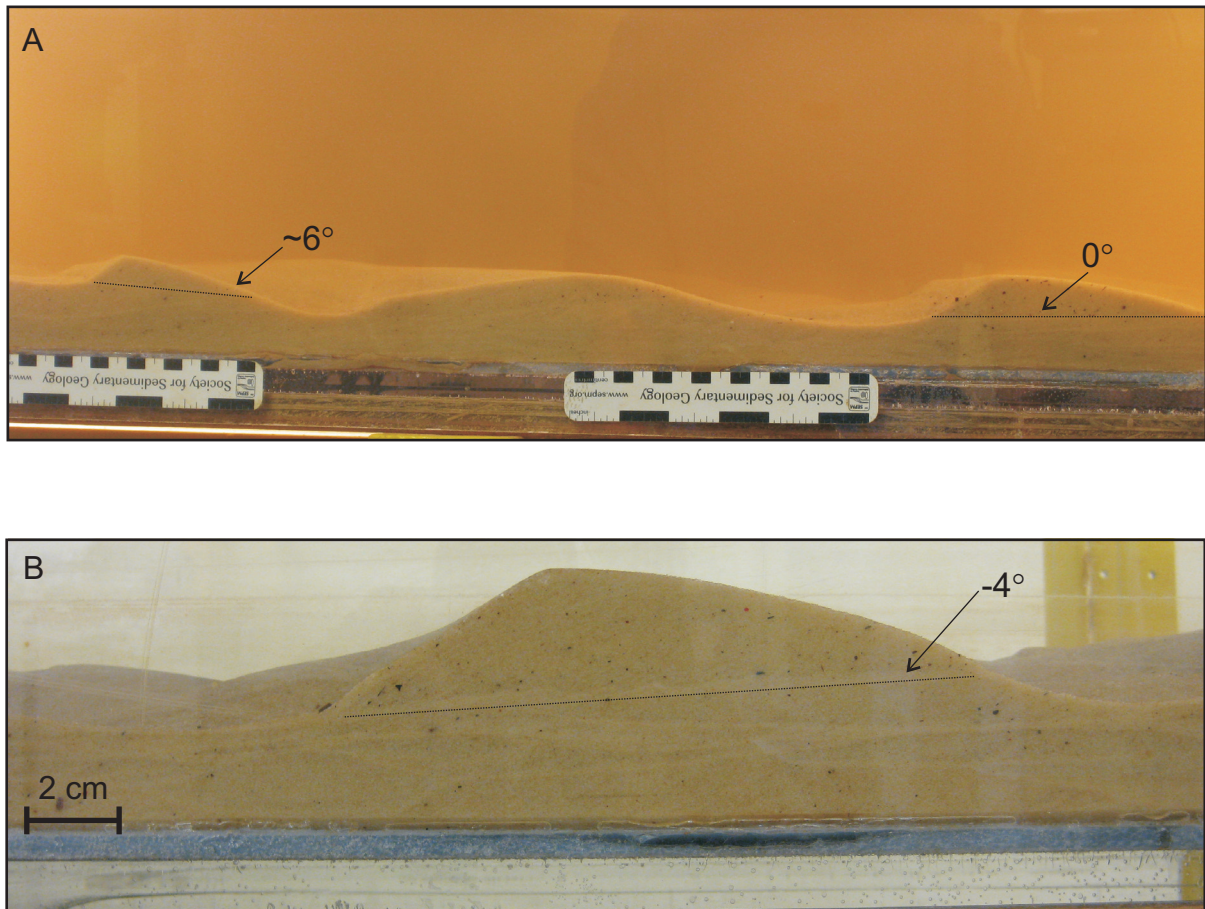


Fig. 6.14. Run 1 from Group 3 (low climb). A) Photo showing ripple bounding surfaces dipping locally at -6–6°. This is due to local erosion and deposition. B) Photo showing a ripple migrating down a dipping surface. The bed only aggraded 0.15 cm during this run and, therefore, the majority of sediment represents the initial bed that was subsequently reworked.

6.8. Discussion

The objective of the experiment was to simulate suspended sediment fallout in deep-water turbidity currents when transport competence and/or capacity had been exceeded. Such conditions commonly occur in deep-water channel-levee systems where throughgoing channel flows overspill, expand and deposit on their less confined levees. The narrow grain size distribution used in this experiment does not mimic the typical wide range of grain sizes in most natural flows. However, there are a number of ancient and modern deep-water levee

systems that are characterized by a narrow grain size distribution similar to those used in the experiments (e.g. Cronin et al, 2000; Browne and Slatt, 2002; Grecula et al, 2003; Eschard et al, 2003). Another more obvious difference is that an open channel flow was used rather than a turbidity current. Producing a sustained (i.e. 10-50 minute) turbidity current in the lab with upper fine sand is a monumental task, and beyond the capabilities of the facilities used here. Nevertheless, ripples formed beneath a turbidity current, like those in open channel flow, are a bed-load transport phenomenon controlled by flow velocity, or more specifically shear stress at the bed. Therefore, the runs presented here most likely simulate well the bed and near-bed conditions developed beneath a depositional, dilute turbidity current.

6.8.1. Climbing-ripple conditions

The high angle (14–25°) climbing ripple cross-stratification formed in the Group 1 experiments is similar to that observed in levee strata of ICC4. The average adjusted flow velocity and depositional rate in Group 1 experiments ranged from 27-28 cm/s and 0.002-0.003 cm/s, respectively. Under these conditions, a 40 cm thick bed, the thickness of the some thicker-bedded, cross-stratified levee deposits in Castle Creek, would be deposited in 4–6 hours. However, for flow depths used in this experiment, ripples are known to be stable at flow speeds ranging from 15–55 cm/s (Southard and Boguchwal, 1990). Therefore, the adjusted flow velocities of Group 1 runs fall into the middle part of the stability field for current ripples. Lower velocity flows would require lower fallout rates to produce the same style of stratification observed in Group 1 runs, whereas higher velocity flows would require higher rates of fallout.

It is difficult to put an upper limit on the time needed to deposit a climbing ripple succession because it involves estimating the minimum speed of ripple migration, which in

theory can be infinitesimally small (Ashley et al., 1982). However, natural sand-transporting turbidity currents are relatively fast moving (Dengler et al, 1984; Middleton, 1993; Khripounoff et al., 2003) and the requisite low mean flow velocities for low rates of ripple migration likely exist only for a short time during the final waning stages of flow. Even if such low velocity turbidity currents did exist they would need to be sustained for weeks to deposit decimetre thick beds at the required bed aggradation rate, and such time spans are much greater than those generally perceived to be involved in turbidite-bed deposition (Piper and Normark, 2009) Estimating the minimum amount time for deposition, on the other hand, is comparatively simple. This is done by finding the maximum possible bedform migration rate, which corresponds to the maximum possible velocity before dunes develop. This transition occurs at about 55 cm/s for decimetre thick flows. Using this flow speed in Equation (6.4), a bedload transport rate of 0.30 cm/s is obtained, which corresponds to a ripple migration rate 0.21 cm/s (see Equation 6.2). Using this bedform migration rate, it is possible to calculate the maximum fallout needed to produce a climb angle of 14° , the average angle of ripple cross-stratification in ICC4. This maximum fallout rate is calculated to be 0.06 cm/s and therefore a bed 40 cm thick would be deposited in about 13 minutes. This is consistent with earlier studies (e.g. Ashley et al., 1982) which suggested that decimetre-thick beds with high angles of climb are deposited in minutes to tens of minutes when ripples migrate at their maximum speed.

Another important observation of Group 1 runs is the fact that climbing angles of 14° – 25° were achieved in flows with suspended sediment concentrations of only 0.0009–0.0015% by volume. This suggests that the important parameter in forming climbing bedforms is the rate of bed aggradation. In other words, the rate that suspended sediment is transferred to the bedload layer which, in natural flows, is not necessarily a function of the

suspended sediment concentration. For example, a turbidity current with a suspended sediment concentration of 15% by volume may have sufficient competence and capacity to bypass its entire load downslope and therefore be non-depositional locally. Alternatively, bed aggradation will occur in a dilute flow (~1% by volume) whose capacity and/or competence has been exceeded. Accordingly, as long as a small percentage (ten-thousandths of a percent) of sediment in the flow is being transferred to the bedload layer then climbing bedforms will form irrespective of the parent flow concentration.

6.8.2. Non-climbing ripple conditions

Run 1 (Group 3) produced non-climbing ripple cross-stratification, which is commonly observed in ICC3. Experimental fallout rate and adjusted flow velocity were 0.00009 cm/s and 31 cm/s, respectively. Of note is that beds a few sets thick can form even if net deposition is zero (Paola and Borgman, 1991). This is a result of the fact that the depth of scour in the lee of equilibrium (3D) ripples varies and, as a consequence, a ripple upstream will not invariably completely erode the next ripple downstream. However, to create beds several sets thick requires net deposition. Moreover, to create beds consisting of multiple stacked sets of non-climbing ripple cross-stratification requires net deposition that to be sufficiently low in comparison to bedform migration rates, which is the case in Run 1. If the ripple cross-stratified beds observed in ICC3 were deposited under the conditions of Run 1, then it would have taken over five days to deposit a bed 40 cm thick. Although hyperpycnal flows have been reported to generate turbidity currents that can last up to several weeks (Mulder and Syvitski, 1995; Mulder et al., 1998), such flows and their related deposits are suspected to be composed predominantly of mud (Mutti et al., 2003). Sand-rich systems, like that exposed at Castle Creek, were more likely deposited from surge-type sustained flows

generated by retrogressive slope failures (Hugues-Clarke et al. 1990; Piper and Normark, 2009). Unlike hyperpycnal flows, these flows last up to several hours (Normark and Piper, 2009), and therefore based on the fallout rate of Run 1 would have been of too short duration to deposit the thick-bedded ripple beds observed in ICC3.

When viewed through the sidewall window, ripples formed in runs 2, 3 and 4 (Group 2) appear to produce non-climbing ripple cross-stratification despite the fact that upon closer inspection they yield an average measured angle of climb of $\sim 4^\circ$. Therefore, the ripple cross-stratified beds in Castle Creek may in fact climb but at angles that cannot be recognized in the field. Moreover, compaction effects and the orientation of the outcrop at Castle Creek, which is typically highly oblique to the direction of ripple migration, would result in apparent climbing angles less than true angle. Accordingly, if it is assumed that many of the ripples in Castle Creek do in fact climb, on average up to 4° , then the conditions of Group 2 could be used to estimate the time of deposition of beds in ICC3. Specifically, the adjusted velocity and bed aggradation rate of Group 2 runs ranged from 26-29 cm/s and 0.00069-0.001, respectively, and using these parameters a bed 40 cm thick would be deposited in about 11–12 hours. This timeframe is comparable with the currents generated by some retrogressive slope failures (Normark and Piper, 2009).

The preceding is based on a moderate mean flow velocity and accordingly, moderate ripple migration rate. If instead, flow speed was higher and migration rate closer to its maximum (0.21 cm/s), then an aggradation rate of 0.015 cm/s would be needed to maintain a climb angle of 4° . This rate of deposition would form a bed 40 cm thick in less than 45 min, suggesting that the thick-bedded deposits in ICC3 could also have been deposited by simple surge-type flows.

6.8.3. *Ripples in the deep-water record*

The scarcity of climbing ripple cross-stratification in Castle Creek is contrary to many other reports of levee deposits (e.g. King et al., 1994; Lyons, 1994; Coleman, 2000; Cronin et al., 2000; Hickson and Lowe, 2002; Eschard et al., 2003; Grecula et al., 2003; Beaubouef, 2004). Walker (1985) suggested that one of the distinctive characteristics of levee deposits is the occurrence of climbing ripple cross-stratification. This criterion is based on the notion that a flow that overtops its levee crest will be highly depositional, causing fallout rates to exceed bedload transport and instigating deposition of climbing ripples (Allen, 1982a, b). In the literature, many reported examples of climbing ripple cross-stratified turbidites are fine-grained, and, perhaps more importantly, very well sorted (e.g. Cronin et al, 2000; Browne and Slatt, 2002; Grecula et al, 2003; Eschard et al, 2003). A consequence of the good sorting is that as current speed wanes a threshold may be attained, where after fallout rates would increase significantly due to competence- and capacity-driven deposition, which, in turn, would overwhelm the bed-load transport system resulting in high angles of climb.

In ICC3, thick-bedded T_c turbidites typically fine upward from medium- to fine-grained sandstone, indicating a broad range of available grain sizes. Accordingly, T_c turbidites in ICC3 most likely represent progressive deposition from mainly competence-driven flows whose instantaneous rates of sediment fallout were never so high as to result in discernible angles of climb.

In ICC4, climbing-ripple cross-stratified beds also fine upwards and therefore do not appear to support the above interpretation that competence-driven deposition of a wide distribution of grain-sizes necessarily result in low aggradation rates. Particularly not worthy, however, is that the climbing bedforms observed in ICC4 migrate toward the channel axis. In

most levee systems paleocurrent are typically directed obliquely away from the channel (DeVries and Lindholm, 1994; Hesse and Dalton, 1996; Hiscott et al., 1997; Hickson and Lowe, 2002; Fildani and Normark, 2004; Hubbard et al., 2008; Khan and Arnott, 2011). On the other hand, channel-convergent paleocurrents have been reported from ancient levee deposits (Crane and Lowe, 2008; Kane et al, 2007, 2010a). These channel-convergent paleoflow trends have been interpreted to be associated with flow reflection along the channel margins or against local topography on the levee formed by mass transport processes (Kane et al., 2010a). Additionally, it has been proposed that channel-directed paleoflow may reflect regional slope gradient, which locally could be oriented towards the channel (Crane and Lowe, 2008). Such a scenario occurs in modern deep-water channel-levee systems such as the Amazon channel (Hiscott et al., 1997; fig. 3). In the case of ICC4, it is interpreted that overbanking flows updip were eventually directed down the regional slope, which might have been roughly perpendicular to the apex of the inner bend exposed in the Castle Creek area. It is further interpreted that channel-directed flows interacted locally with the overbanking flows of ICC4, which were directed away from the channel and over the inner-bend levee. These opposing flows caused local flow expansion, which in turn led to rapid flow deceleration and high rates of bed aggradation as recorded by the climbing ripple cross-stratification. Moreover, it is interpreted that at this time ICC4 was highly confined and that local overbank flows were weak and fine grained. Channel-directed flows, in contrast, were stronger and sustained by overbanking further up the channel system. Accordingly, ripples in ICC4 climbed, but under this assemblage of local conditions migrated toward instead of away from the channel axis.

6.9. Conclusion

Climbing ripples are common features in the modern and ancient deep-water record and are generally interpreted to indicate high rates of sediment fallout from suspension (Allen, 1971, Walker, 1985, Grecula et al., 2003), but how “high” is high? Theoretically, any flow that is net depositional will result in climbing ripple cross-stratification. Flume modelling described here, however, show that moderate flow velocities (26–31 cm/s) and extremely low sediment fallout rates (0.00009 cm/s) result in negligible angles of climb (0.13°), which would not be easily measured in the field. This study argues that a climbing angle of less than 4° may not be readily apparent in the field. At moderate flow velocities this threshold angle of climb is achieved with very low sediment fallout rates (0.0009–0.001 cm/s). At this higher fallout rate, a bed 40 cm thick would be deposited in 11–12 hours, which is consistent with the duration of some turbidity currents. However, for ripples formed at high flow velocities just below the theoretical threshold of dune development, a sediment fallout rate not exceeding about 0.015 cm/s is needed to maintain a climb angle of 4° or less. At this higher rate of fallout, a bed 40 cm thick would be deposited in less than 45 min, well within the estimated duration of time of simple surge-type flows. It is concluded that this fallout rate represents an upper limit under which shallow climbing ripple cross-stratification may be mistaken in the field to be non-climbing. At lower mean flow velocity however, a lower sediment fallout rate could result in easily recognized climbing ripple cross-stratification. For example, in the present experiment moderate velocities and a sediment fallout rate of 0.002 cm/s caused ripples to climb at angles between 14° and 25° . Therefore it is entirely possible that some non-climbing ripple beds in ICC3 formed under higher rates of sediment fallout than their climbing counterparts. This suggests that the term “high” is

essentially meaningless unless constrained by flow velocity and hence ripple migration speed. Moreover, it is suggested that well sorted flows are more likely to form climbing-ripples compared to those made up of a wide range of grain sizes. In the former the conditions of competence- and capacity-driven deposition causes instantaneous fallout rates to easily exceed the bed-load transport system, and result in climbing ripples.

Conclusions and areas for future research

7.1. Introduction

Deep-water levee deposits are significant elements within many submarine fans and are the most voluminous stratal component in base-of-slope settings of mud-rich deep-marine systems (Prior et al., 1983; Damuth et al., 1988; Hiscott et al., 1997; Pirmez et al., 1997; Deptuck et al., 2003, Beaubouef, 2004; Khan and Arnott, 2011). Nevertheless, detail analysis of these deep-water stratal units are underrepresented in the literature, especially from the ancient sedimentary record. This is most likely due to the fact that levee deposits, being generally fine-grained, are typically intensely weathered and/or poorly exposed in the rock record. In the Castle Creek study area, however, levee deposits are well exposed with negligible surface weathering and, therefore, offer a rare opportunity to study such deposits. This thesis takes advantage of these conditions by documenting in detail levee deposits through field investigation. In addition, experiments were conducted to better understand the depositional processes of ripple cross-stratification, a common feature in modern and ancient levees. This final chapter develops a generic depositional model of a sinuous deep-water channel-levee system based on key observations and interpretations presented in chapters 4 and 5. In addition, the key findings of chapter 4 and 5 and their implications for reservoir characterization and exploration are summarized (Table 7.1 and 7.2). The principal conclusions from Chapter 6, which describe the depositional properties of ripple cross-

stratified beds, are also reviewed. Lastly, this chapter discusses possible areas for future research based on the research presented in chapters 4–6.

7.2. Evolution of a sinuous channel-levee system – a synthesis

7.2.1. Incipient conditions

Deep-water channel-levee systems are commonly underlain by areally extensive debrites. This stratigraphic relationship suggests that topographic irregularities along the top of debrites helped focus subsequent turbidity currents downslope (Fig. 7.1A). Early flows locally eroded the underlying substrate and promoting the localization and progressive development of a large-scale (10's × 1000's m) sinuous channel-levee system.

7.2.2. Stage 1

During Stage 1 the channel was weakly confined (width to depth ratio of 100's to 1) allowing up to coarse-grained sand to overspill onto the adjacent levee areas (Fig. 7.1B). This resulted in the deposition of thick- to medium-bedded, coarse- to fine-grained turbidites along both sides of the channel. Thicker-bedded strata on the levees typically thicken and then thin over about 300 m laterally away from the channel and thereafter become thin-bedded deposits that show negligible lateral change in thickness or grain size. As a result of these lateral changes the sandstone proportion decreases away from the channel by 25–30%. Moreover, superelevation during this stage of channel development preferentially deposited coarser-grained, thicker-bedded strata along the outer-bend side of the channel, which caused the outer-bend levee to be significantly thicker than its inner bend counterpart. Noteworthy is that intercalated thin-bedded, fine-grained turbidites change little laterally away from the

Table 7.1

Key findings of Chapter 4 and their implications for reservoir characterization

Key findings		Implications
Levee beds		
- Thin beds <1 – 10 cm thick - fine-grained sandstone to siltstone - ripple cross-stratified sandstone, massive or planar laminated siltstone cap (T_{cd}) - bed thickness changes negligibly laterally - Medium and thick beds 10 – 150 cm thick - coarse- to fine-grained sandstone - massive, parallel laminated, and ripple cross-stratified sandstone, with massive or planar laminated siltstone cap (T_{a-d}) - thicken laterally over ~100 m away from channel and thereafter thin over next ~200 m to less than 10 cm thick		- Aid in recognizing and differentiating different levee sub-environments in core - Upward and lateral thinning and fining trends can be used to interpret levee deposits from down-hole geophysical well log - Help estimate rates of reservoir quality degradation with distance from channel - Evaluate lateral pressure communication between channel and levee deposits
Levee architecture		
Outer-bend proximal levee		
- Channel fills onlap outer-bend levee along erosive, terraced channel margin. Levee is made up of three levee packages (12 to 28 m thick) - Packages consist of lower and upper parts - lower part 7 to 11 m thick - medium- to thick-bedded turbidites interbedded with thin-bedded turbidites - fine and thin upwards and laterally - N:G decreases laterally from 0.6–0.8 to 0.38 over ~200 m (away from channel margin) - upper part 5 to 18 m thick - composed of interbedded thin-bedded turbidites - strata thin and fine upward - N:G changes negligibly laterally		- Help create forward seismic models to better interpret deep-water seismic - Data can be used to develop more realistic stratigraphic models describing the origin and stratal architecture of subaqueous levees
Outer-bend distal levee		
80 m thick - Mostly thin-bedded turbidites interbedded with uncommon medium- and thick-bedded turbidites in the lower 40 m - Local 2–5 m sandstone-rich overbank splay deposits - Upper 18 m fines and thins upward (no trend in lower 62 m) - N:G in lower 62 m averages about 0.44 and in the upper 18 m decreases from 0.4 to 0.08 (no measurable lateral change in N:G over 300 m)		
Inner-bend levee		
- Subdivided into two parts - lower unit 17 m thick - consists mostly of thin-bedded turbidites; uncommon medium-bedded turbidites occur in lower 5 m - fines and thins upward - N:G decreases laterally from 0.4 to 0.15 over 50 m - channel fills onlap levee - upper unit 40 m thick - thin-bedded turbidites - N:G averages 0.15 - channel-fill strata thin and fine laterally and then grade into the levee over a distance of about 140 m (N:G is up to 0.9 immediately adjacent where the channel-fill begins to intercalate with much finer and thinner levee deposits)		

Table 7.2

Key findings of Chapter 5 and their implications for reservoir characterization

Key findings	Implications
- Stacking of beds that change thickness laterally do not necessarily create noticeable stratal dip - Stratal dip as low as 1° should be resolved in detailed correlation panels - Compaction does not appear to sufficiently reduce slope to where it would not be resolved on correlated stratigraphic logs - Typical compaction models for levees that assume a homogeneous clay-rich lithology do not accurately apply to most levees that change lithology laterally - Levee backslope may result from the backstepping of bed termini toward the channel margin - Dipping reflections in seismic images of deep-water levees may instead be the manifestation of a lateral lithological change and not, as is commonly assumed, related to stratal dip - Seismic reflection of levees may cross-cut horizontal stratigraphy (i.e. timelines)	- Bed pinch-outs in levees may be more common than previously assumed and, therefore, lateral bed continuity of levee beds in reservoir models need to be re-evaluated - If levee beds do cross-cut seismic reflection then reservoir volumes based on previous models based on previous assumptions may need to be re-calculated (which presumably will increase volume estimates) - Compaction models may need to be re-evaluated to account for the internal stratigraphic complexity of levees

channel and were deposited by smaller magnitude flows from which only the upper more dilute part spilled over onto the levees.

Generally it is assumed that the gull-winged geometry of levees reflects stratal dip resulting from the vertical stacking of individual beds that vary laterally in thickness. In the Castle Creek study area, however, topography associated with the lateral thickening and then thinning of medium- and thick-bedded strata is infilled with thin-bedded deposits and therefore does not cause beds to dip appreciably in the levee. An alternative process of forming the mound-shaped geometry commonly exhibited by levees is the stratigraphically upward backstepping of levee strata toward the channel. This stacking pattern was the consequence of levee growth, which progressively diminished flow magnitude and caused beds to become reduced in run-out length. With time this created a surface along the backside of the proximal levee that dipped away from the channel. Episodically, however, high-magnitude flows were able to bypass some of their sediment load along this sloped

surface and onto the distal levee.

During this early stage of levee growth flows confined within the channel were largely bypassing sediment further downslope and deposited discontinuous lag deposits (e.g. pebbly conglomerate, mudstone-clast breccia, and tail-of-flow thinly-bedded turbidites) locally in the axis of the channel. In addition, it is likely these bypassing flows were largely responsible for sculpting and widening the channel-form by eroding along the channel margins.

7.2.3. Stage 2

Increased channel confinement during the early part of Stage 2 enabled only the upper, more dilute portion of flows to overspill (Fig. 7.1C). This resulted in deposition of mostly thin-bedded turbidites on both outer- and inner-bend levees. However, as a result of inertial effects, outer-bend levee aggradation continued to outpace its inner-bend counterpart. Overspilling flows during this stage progressively became more dilute and slower moving resulting in the continued backstepping of bed termini towards the channel. This progressively decreased the surface area on top of the proximal levee and lengthened the backside levee slope. At the same time, deposition on the distal levee remained limited and as a result was greatly outpaced by growth of the proximal levee. Channel filling commenced during the latter part of Stage 2 but remained minor compared to upward levee growth.

7.2.4. Stage 3

The early part of Stage 3 was marked by the onset of intense channel filling, likely related to a progressive decrease in flow magnitude and/or increase in sand:mud ratio (Fig. 7.1D). Initially channels filled with thick, amalgamated, normally graded and massive

conglomerate and very coarse-grained sandstone. Stratigraphically-upward these strata thin and fine slightly and become more interbedded with thin-bedded turbidites. Increased rates of channel aggradation resulted in a net reduction in levee relief.

During this stage overspilling flows along the outer-bend margin typically bypassed the proximal levee and deposited on the distal levee. This was mainly the result of depositional patterns during Stage 2 that appreciably reduced the surface area of the proximal levee. Successive large-magnitude, sandy bypassing flows emplaced splay deposits on the distal levee that increase the sand proportion locally by as much as 10%. These large-magnitude flows, however, were not common and so the majority of flows that extended into the distal levee area deposited thin-bedded, fine-grained turbidites contained only about 35% sand. Collectively, increased deposition on the distal levee reduced the differential topography between the proximal and distal levees. The surface area of the inner-bend levee was not sufficiently reduced during Stage 2 and therefore, during Stage 3, overspilling flows typically terminated on the proximal part of the inner-bend levee and added to its relief. Consequently, this reduced the difference in height between the inner- and outer-bend levee.

Although channel aggradation began to outpace levee growth during this stage it is interpreted that the outer-bend levee remained sufficiently high, thereby creating a margin onto which channel-fill strata overlapped. Strata in the lower channel fills also onlap the inner-bend levee suggesting that levee relief was initially relatively high. As the channel continued to fill, however, levee relief became more subdued until some flows that deposited thick, coarse beds in the channel axis continuously thin and fine laterally until they become indistinguishable from other fine, thin-bedded inner-bend levee strata.

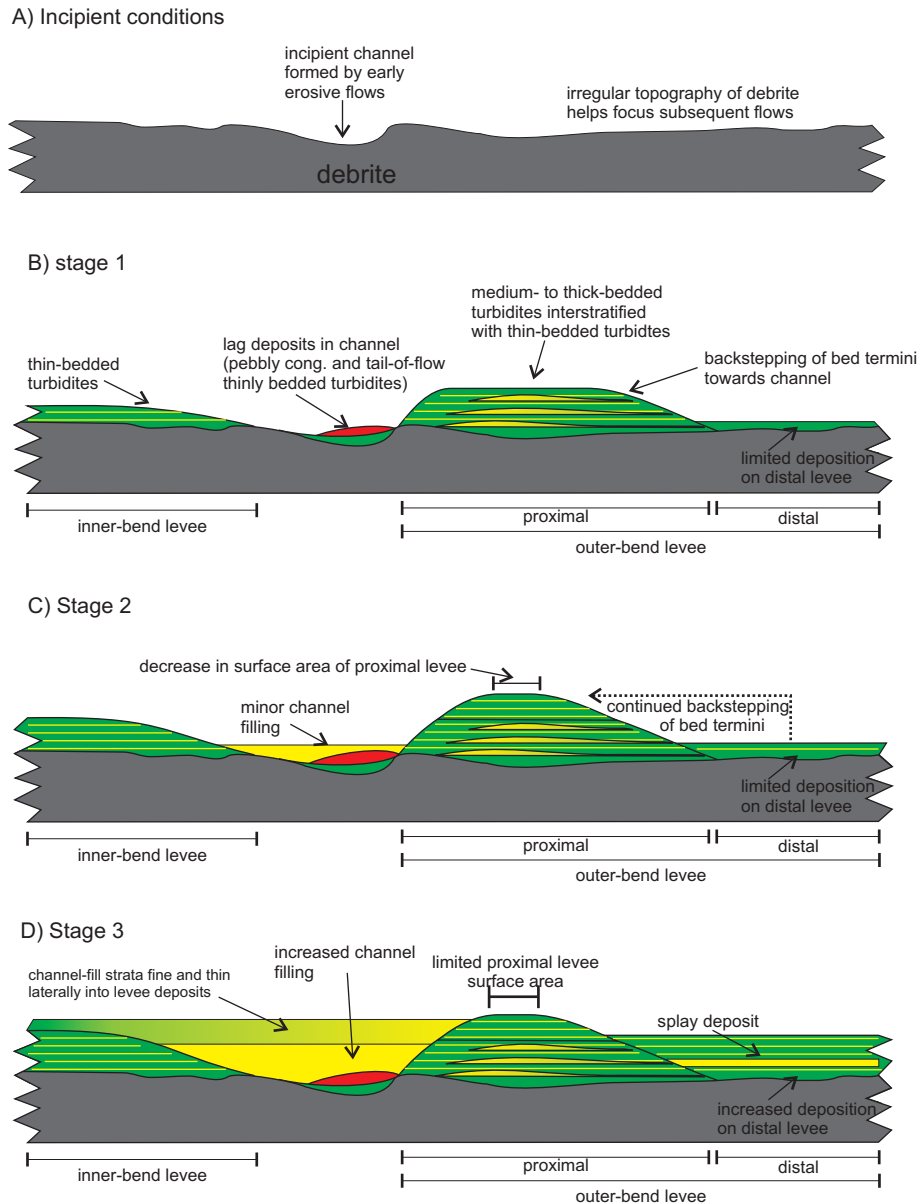


Fig. 7.1. Schematic illustrating the evolution of a sinuous, deep-marine channel-levee system. A) Topography along the top of a mass transport deposit initially focused turbidity currents downslope, which progressively eroded the underlying substrate and formed the incipient sinuous channel. B) Stage 1 was dominated by channel erosion and bypass coincident with levee growth. Low levee relief during this stage allowed the lower, coarser-grained part of some flows to overspill and deposit medium- to thick-bedded turbidites. Moreover, thicker bedded deposits were preferentially emplaced on the outer-bend levee due to inertial forces (i.e. superelevation and run-up). Importantly, as levee relief increased the average magnitude of overspilling flows decreased causing beds to have gradually shorter run-out lengths. As a result bed termini progressively backstepped toward the channel. C) Stage 2 was characterized by minor channel filling and progressively more dilute overspilling, which lead to the deposition of mainly thin- to very thin-bedded turbidites. D) The last stage, Stage 3 was marked by a period of channel filling that outpaced levee growth. Overspilling flows along the outer-bend margin typically bypassed the proximal levee during this stage and deposited on the distal levee. Deposition along the inner-bend, in contrast, continued to aggrade the proximal levee. Some thick-bedded deposits grade laterally into the inner-bend levee suggesting that at times levee relief along the inner-bend side of the channel was sufficiently subdued.

7.3 Insight from experimental investigation and theory

Climbing ripple cross-stratification is commonly reported from proximal levees in both modern and ancient deep-water systems and is typically interpreted to represent “high” rates of bed aggradation. In the Castle Creek study area, however, climbing ripple bed forms are rare, whereas multiple-set thick, non-climbing ripple cross-stratified beds are common. The objective of Chapter 6 was to determine the threshold conditions at which ripples begin to climb and therefore give a quantitative measure of “high” rates of bed aggradation. Flume experiments suggest that an average climb angle of less than 4° would not be readily discerned in the field, especially in exposures oriented obliquely to a section cut perpendicular to paleoflow. Based on laboratory and theoretical considerations, it was determined that for ripples of 0.25 mm sand (lower medium sand) formed just below the threshold of dune development, a bed aggradation rate of >0.015 cm/s is required to form readily identifiable climbing ripples, and it is this rate that necessarily equates to the qualitative term “high”. At lower flow speeds and therefore lower ripple migration rates, however, climbing ripples require lower rates of sediment fallout. This illustrates that fallout rate alone does not necessarily determine whether a ripple will climb or not. Moreover, it is proposed here that well-sorted flows are more likely to form climbing-ripple bed forms compared to flows composed of a range of grain sizes.

7.4. Summary of thesis study contributions

1. Provides a novel model of producing levee topography wherein strata are deposited sub-parallel to regional sedimentary layering. This contrasts starkly with the

commonly held opinion that levee topography results from the vertical stacking of dipping strata.

2. Presents a synthetic seismic model that produces dipping reflections from sub-parallel stratigraphy. Importantly, the dipping reflections in this model are an artifact of lateral facies changes and not stratal dip.
3. Documents and contrasts the stratal attributes and evolution of inner- and outer-bend levees in an ancient large-scale sinuous channel-levee complex. To the author's knowledge, no such study has been reported in the literature and, therefore, makes a significant contribution to deep-water scholarship.
4. Presents a new equation that can be used to compact or decompact thin successions (<100 m) of sedimentary rocks:

$$T_{solid} = (z_2 - z_1) \left(1 - \phi_0 e^{-c \left(\frac{z_2 - z_1}{2} \right)} \right) \quad (5.9)$$

5. Discusses the effects of compaction on stratal dip using previously published lithologically sensitive compaction curves. Although compaction models are common in the literature to the author's knowledge none have been used to assess stratal dip in levees.
6. Determinines the threshold depositional rate at which climbing-ripple cross-stratification necessarily forms in 0.25 mm quartz sand (0.015 cm/s).

7.5. Areas for future research

7.5.1. Chapter 4

- Grain-size trends presented in this study are based on visual estimates in the field and therefore a detailed petrographic analysis of the two-end member types of levee deposits (i.e. medium- and thick-bedded T_{a-d} turbidites, and very thin and thin-bedded T_{cd} turbidites) would be useful. Finer-grained units commonly contain abundant chlorite and are extensively tectonically cleaved and therefore do not preserve well the original depositional textures. The original fabric and texture of coarser-grained units are better preserved and a detailed petrographic analysis of these rocks may lend some insight into flow characteristics at the time of deposition.
- Investigate Isaac Channel Complex 3 further to the southeast. Although outcrop quality quickly decreases beyond the current study area, there are some good isolated exposures and exploring these may give insight into the make-up of the distal part of the inner-bend levee. This additional information would be useful to better understand the depositional processes along the inner-bend margin of large-scale sinuous channels.
- A scintillometer was used to record a gamma-ray profile for many of the measured stratigraphic sections measured in Isaac Channel Complex 3. Merging these two data sets may help determine if the different levee architectures, namely thin-bedded deposits, thin-bedded interstratified with thick-bedded deposits, mainly thick-bedded deposits, and amalgamated thick-bedded deposits, exhibit unique and easily identifiable gamma-ray signatures. This would help interpret down-hole gamma-ray

profiles.

7.5.2. Chapter 5

- Detailed correlation of distal levee strata to confirm or refute lack of stratal dip. This correlation would also help determine the actual percentage of thin-bedded deposits that pinch out (at least their sandy basal part) over 100's of meters laterally. Although this correlation would be beneficial it requires tracing thin-bedded strata laterally, which becomes exceedingly difficult where locally covered.
- Expand the synthetic seismic levee model by extrapolating the decrease in sandstone proportion measured in ICC3 over about 3–5 km. At present the scale of the model is comparable to the crest region of large-scale levees, and therefore expanding the model laterally would make it more comparable with the full spatial scale of natural deep-marine levees.
- Convert the gamma-ray profiles of each measured lithostratigraphic section into relative sonic velocity profiles and then enter that data into a forward-modeling seismic program, such as Log-M by Geographix/Landmark. The model presented in Chapter 5 significantly simplifies the stratigraphy observed in the levees. The proposed forward-modeling technique uses measured stratigraphic logs and, therefore, provides a more accurate image of how levee strata of Isaac Channel Complex 3 would appear compared to using conventional seismic techniques.

7.5.3. Chapter 6

- More flume runs to better constrain the threshold limit between non-climbing and

climbing cross-stratification, thereby relying less on theoretical considerations.

- Update the sediment feed system on the flume so that it can be easily and accurately recalibrated between runs. This would add much needed consistency to the experiments.
- Use a wider range of grain sizes and grain sorting.
- Measure ripple migration rate in the center of the flume to avoid error induced by side-wall effects.
- Develop a more accurate method of measuring sediment concentration in the flow.

For example a siphon rake (e.g. Choux et al., 2005) could be placed in the flow to simultaneously measure sample concentration at multiple heights above the bed. This would help differentiate the true suspension particle concentration in the middle of the flow from those nearer the bed, which can be elevated by grain-to-grain collision and flow separation, which, in turn, causes local eddy suspension.

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Appendix A

Stratal attributes and evolution of asymmetric inner- and outer-bend levee deposits associated with an ancient deep-water channel-levee complex within the Isaac Formation, southern Canada (Marine and Petroleum Geology)



Stratal attributes and evolution of asymmetric inner- and outer-bend levee deposits associated with an ancient deep-water channel-levee complex within the Isaac Formation, southern Canada

Zishann A. Khan*, R.W.C. Arnott

Department of Earth Science and Ottawa/Carleton Geosciences Centre, University of Ottawa, Ottawa, ON K1N 6N5, Canada

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ABSTRACT

Isaac Channel 3 is a rare outcrop example of a perpendicular cut through a sinuous deep-water channel, and also where levee deposits formed on opposite sides of the channel are well exposed. Strata flanking the outer- and inner-bend margin of the channel show important differences in lithofacies, architecture and association with channel-fill strata. Proximal outer-bend levee deposits are sand-rich (N:G up to 0.68) and comprise medium- to thick-bedded, T_{a-d} turbidites interstratified with thinly-bedded, T_{cd} turbidites. The thicker-bedded deposits show lateral variation in grain size and thickness over hundreds of meters whereas thin-bedded strata thin and fine negligibly over similar distances. The distal outer-bend levee (up to 700 m laterally away from the channel) consists predominantly of thin-bedded turbidites interstratified with up to 5 m thick coarse-grained splay deposits. In contrast to the outer-bend, the inner-bend levee deposits are significantly more mud-rich (N:G as low as 0.15) and consist mostly of thin-bedded, T_{cd} turbidites with less common thicker-bedded, T_{a-d} turbidites. Lateral thinning and fining trends associated with these less common thicker-bedded deposits occur more rapidly than their outer-bend counterparts.

Erosion associated with lateral migration of the channel axis produced a sharp contact along the outer-bend channel margin causing coarse-grained channel-fill deposits to be in erosional contact with levee deposits. This suggests that the crest of the outer-bend levee was elevated above the channel floor and produced a channel margin upon which channel-fill strata overlapped. Positive topography is interpreted to have developed by overspilling processes that deposited abundant sand on the outer-bend levee while the majority of the flow continued through the channel bend and bypassed to areas further downslope. In contrast, some thick-bedded, amalgamated channel-fill deposits in the axial channel area grade laterally over 140 m into thinly-bedded turbidites on the inner-bend levee. The lack of channel-fill on lap relationships implies that topography along the inner bend was sufficiently subtle that at least some flows were able to expand laterally and over the overbank area without becoming separated from the main throughgoing channel flow.

Stratal relationships observed in Isaac Channel Complex 3 suggests three main episodes of channel-levee growth that were each initiated by a period of increased levee relief followed by channel filling and distal levee deposition. This consistent depositional history points to the regular variations, in both time and space, of sediment transport and deposition in a deep-marine sinuous channel-levee system.

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1. Introduction

Sinuuous channels are fundamental geomorphological features in both the continental (e.g. Leopold and Wolman, 1960; Walters and Simons, 1984; Brice, 1984; Miall, 1996) and deep-water realm (e.g. Garrison et al., 1982; Flood and Damuth, 1987; Abreu et al.,

2003; Deptuck et al., 2003, 2007; Posamentier and Kolla, 2003; Kolla et al., 2007). Channelization typically results from local erosion of the underlying substrate and/or by deposition of topographic highs along the sides of the conduit. Channel bounding depositional highs are termed levees and form mainly by three kinds of overbanking processes: overspill (Clark and Pickering, 1996; Hiscott et al., 1997), flow stripping (Piper and Normark, 1983), and inertial overspill (Hay, 1987a,b). Overspill results from over-fit flows that deposit simultaneously along both sides of the channel. Flow stripping, in contrast, preferentially takes place along

* Corresponding author. Tel.: +1 613 562 5800x6823; fax: +1 562 5192.
E-mail address: zishann@hotmail.com (Z.A. Khan).

channel bends where the upper more dilute part of flows becomes detached from the main throughgoing channel flow and deposits on the outer channel bank. This process is typically enhanced by the superelevation of flows along the outer bend due to fluid inertia (Imran et al., 1999; Straub et al., 2008). Inertial overspill, or simply run-up, occurs when the kinetic energy of sediment particles, specifically coarse sediment carried in the lower part of the flow, enables particles to run up and over the outer bank (Straub et al., 2008). Unless otherwise specified, this paper will refer to all of these processes collectively as overspilling or overbanking.

Overbanking processes are significantly enhanced in the deep-water realm compared to fluvial settings (Imran et al., 1999). This is because fluvial flows are constrained at their upper boundary by the water air interface which is marked by an abrupt and significant change in density. Consequently, overbanking occurs mostly during floods that commonly deposit a thin layer of sediment on the floodplain (Cazanaci and Smith, 1998; Brierley et al., 1997). In the deep-marine realm, on the other hand, turbidity currents are the main sediment transport agent and their upper boundary is marked by a much smaller density contrast. Accordingly, flows are able to easily transfer across the upper boundary because of buoyant support provided by the ambient fluid (i.e. seawater), allowing the flow to expand vertically, commonly exceeding the depth of the associated channel. Due to these conditions, channel bounding levee deposits make up a significant part of the sedimentary record of modern and ancient deep-water systems (e.g. Prior et al., 1983; Damuth et al., 1988; Pirmez, 1994; Hiscott et al., 1997; Deptuck et al., 2003; Beaubouef, 2004; Navarro et al., 2007), and have proven to be important hydrocarbon reservoirs (e.g. Tahoe and Ram/Powell fields in northern Gulf of Mexico; Clemenceau et al., 2000; Kendrick, 2000). Nonetheless, it is the volumetrically less significant channels and their associated fills that form the focus of most studies, particularly in the ancient

record. A likely reason for the scarcity of levee documentation is that channel bounding strata are typically heavily weathered or overgrown by vegetation, due to their relatively fine-grained nature, making it impractical to carry out detailed studies. Moreover, it is even less common to preserve an ancient channel complex that exposes strata on both sides of the channel and therefore there is little understanding of the stratal differences and evolution between these counterpart deposits.

Isaac Channel Complex 3 (ICC3) in the Castle Creek study area is an 80 m thick, and more than 1.8 km-wide seismic scale outcrop of a submarine sinuous channel-levee system (Navarro et al., 2007), in which levee deposits are well exposed along both channel margins and can be traced out continuously for hundreds of meters laterally. Also, the spatial and temporal relationships between levee strata and their association with related channel-fill deposits are easily observed (Navarro et al., 2007). Outcrops like the one described in this manuscript are important to define reservoir architecture and lithofacies associations in different sub-environments of deposition, as well as sub-seismic geometries. The aim of this study, therefore, is to 1) describe the stratal attributes of inner- and outer-bend levees, and then use those characteristics to 2) construct a generalized model describing the stratigraphic evolution, in both space and time, of an ancient channel-levee system.

2. Geological setting and Castle Creek study area

Sedimentary rocks of the Neoproterozoic Windermere Supergroup (WSG) extend the length of western North America from northwestern Mexico to the Yukon-Alaska border – a distance of more than 4000 km (Fig. 1a). The deep-water part of the WSG is well exposed in the southern Canadian Cordillera (Fig. 1b), which forms part of an orogenic belt related to plate convergence in Mid-Jurassic to Early Cenozoic time (Reid et al., 2002). WSG strata here

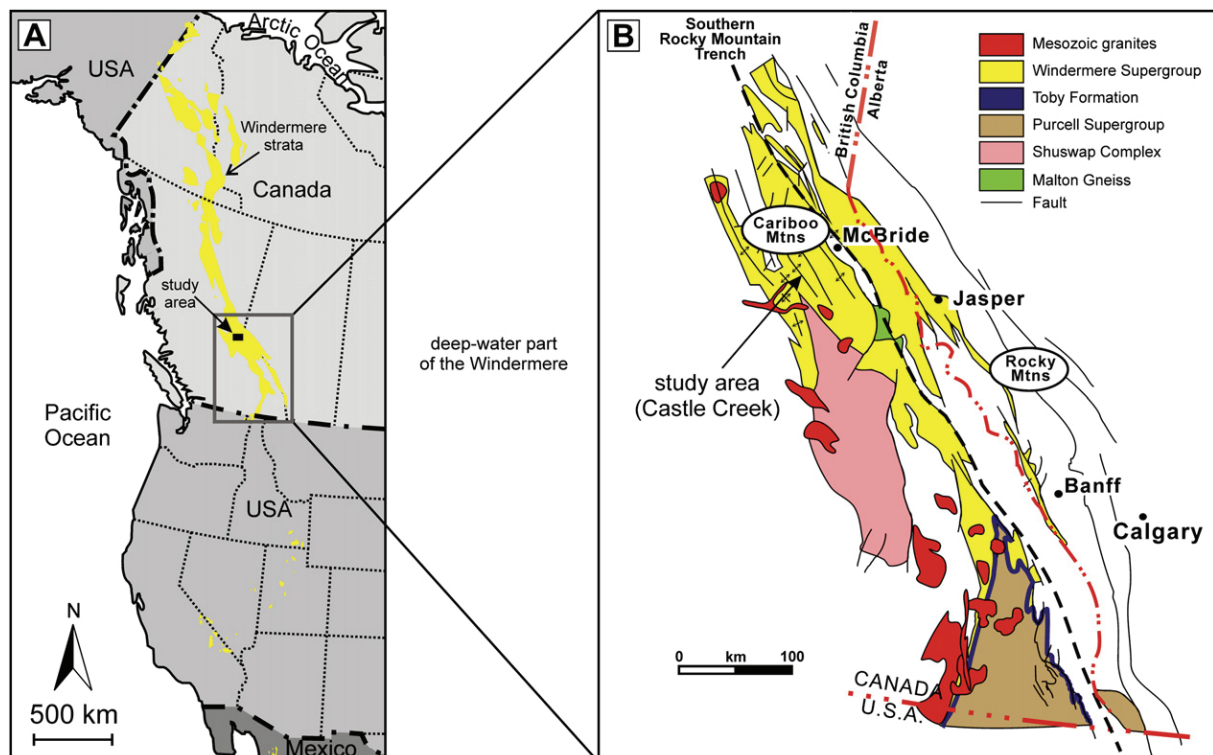


Fig. 1. Location of the study area in relation to (A) map of western North America (outcrop distribution from Ross, 1991) and (B) geological map of the deep-water part of the Windermere Supergroup and surrounding units in the southern Canadian Cordillera (modified from Ross and Murphy, 1988).

are interpreted to have been deposited along part of the Laurentian (present day North America) continental margin during break-up of the supercontinent Rodinia ca. 750 Ma (Ross, 1991; Brookfield, 1993; Ross et al., 1995). In this area the lowermost part of the WSG (i.e. Toby and Irene formations; Fig. 2) is interpreted to record deposition during active rifting and comprise conglomerate, glaciogenic diamictite, and local mafic volcanic rocks that are up to 1 km thick (Stewart, 1972; Ross, 1991). Overlying these rift related rocks, and representing the passive-margin phase of WSG deposition, is a 5–7 km thick pile of predominantly deep-water siliciclastic rocks with subordinate carbonates. Based on regionally extensive marker units (i.e. Old Fort Point Formation and two deep-water carbonate successions) the WSG deep-water basin may have been of the order of 80 000 km² or more (Ross and Arnott, 2007).

In the northern Cariboo Mountains of east-central British Columbia strata of the WSG are well exposed at the Castle Creek study area (Fig. 1). At this locality Windermere strata crop out on the overturned, nearly vertically-dipping limb of a southwest verging fold. Notwithstanding regional deformation and low-grade metamorphism in the area, effects on the preservation of primary sedimentary structures have been relatively minor, even down to mm-scale features. The exposure is about 7 km parallel to bedding and 2.4 km perpendicular to bedding. Fig. 3 shows only a 2.2 km by 2.4 km portion of the Castle Creek study area.

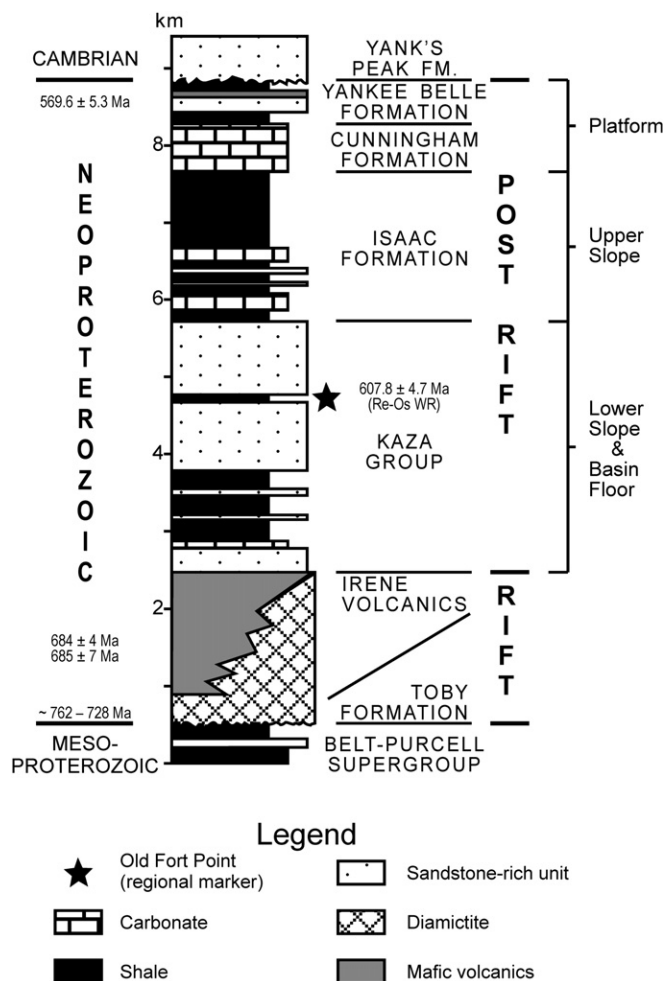


Fig. 2. General stratigraphic log of the Windermere in the southern Canadian Cordillera (Modified from Ross et al., 1995). Age dates from Ross et al. (1995), Lund et al. (2003), Kendall et al. (2004), and Colpron et al. (2002).

In Fig. 3 the lower 800 m of the section comprises the upper Kaza Group, which consists predominantly of massive to normally graded, medium- to very thick-bedded turbidites intercalated with subordinate intervals up to 40 m thick made up of very thin- to thin-bedded, fine-grained turbidites (Meyer and Ross, 2007). Thicker-bedded strata comprise a mixture of rhythmically-stacked, quasi-tabular elements interpreted to represent basin-floor distributary lobes, or similarly, terminal splays (Meyer and Ross, 2007).

The Isaac Formation conformably overlies the Kaza Group and is exposed vertically for 1.6 km in the study area. In contrast to the Kaza, coarse sandstone and conglomerate form major but laterally discontinuous stratal units up to about 100 m thick. These units exhibit channel geometries (Arnott, 2007; Navarro et al., 2007; Schwarz and Arnott, 2007) and are informally termed Isaac Channel Complexes 1–6 (ICC1–ICC6). Channel complexes are encased on all sides by predominantly thin-bedded turbidites interbedded with subordinate thicker (up to 5 m), amalgamated sandstone units. These strata are collectively interpreted to represent channel-margin levee and other extra-channel deposits (Ross and Arnott, 2007). Also observed in the Isaac Formation are thick debris flow, slump, and slide units that together indicate gravitational instability which is common in a deep-water slope setting (Arnott et al., in press).

Overall, strata exposed at Castle Creek record a kilometer-scale upward-shoaling succession from basin-floor (Kaza Group) to base-of-slope (Isaac Formation) deposits interpreted to be the result of progradation of the Laurentian continental margin in response to diminishing rates of thermally-driven subsidence (Ross, 1991; Ross and Arnott, 2007). Although this study focuses on only one channel-levee complex, Castle Creek as a whole fits well within the theme of this special issue by recording the large-scale evolution and stratal stacking patterns in an ancient basin fill. Other examples of kilometer-scale upward-shoaling ancient deep-water successions include the Miocene Marnoso-Arenacea Formation of Italy (e.g. Lucente, 2004; Tinterri et al., 2011) and the Upper Cretaceous Magallanes Basin of southern Chile (Romans et al., 2011). These examples, however, were deposited along tectonically active margins and hence contrast the passive margin of the WSG.

3. Isaac Channel Complex 3

Lateral facies trends from proximal to distal levee and the relationship between levee strata and channel-fill deposits are well developed in ICC3 (Fig. 4). This channel crops out as an almost perpendicular-to-flow cut through the bend of an 80 m thick, more than 1.8 km wide sinuous deep-water channel-levee system that is made up of three discrete channel-levee units that range from 12 to 50 m thick. ICC3 is underlain by an up to 60 m thick, mud-rich debrite.

The sinuous channel interpretation of ICC3 is based on three key observations. Firstly, channel-fill units are vertically stacked and laterally offset creating a terraced geometry along one side of the channel that sharply separates levee and channel-fill strata. This architecture suggests discrete periods of erosion associated with the systematic lateral shifting of a sinuous channel followed by the onset of channel filling (Campion et al., 2000; Samuel et al., 2003; Beaubouef, 2004; Navarro et al., 2007). Secondly, the levee deposits occurring along one side of the channel (outer bend) are significantly sandier than those on the opposite side (inner bend). This likely resulted from thicker, higher energy flows preferentially overspilling the outer-bend channel margin due to inertial effects (i.e. superelevation and inertial run-up; Imran et al., 1999; Straub et al., 2008). Thirdly, paleocurrent directions in the outer-bend levee average 28°, whereas those in the inner-bend levee average 80°, both of which are oblique to the main paleocurrent channel trend of 40°. The divergence in paleocurrent between the

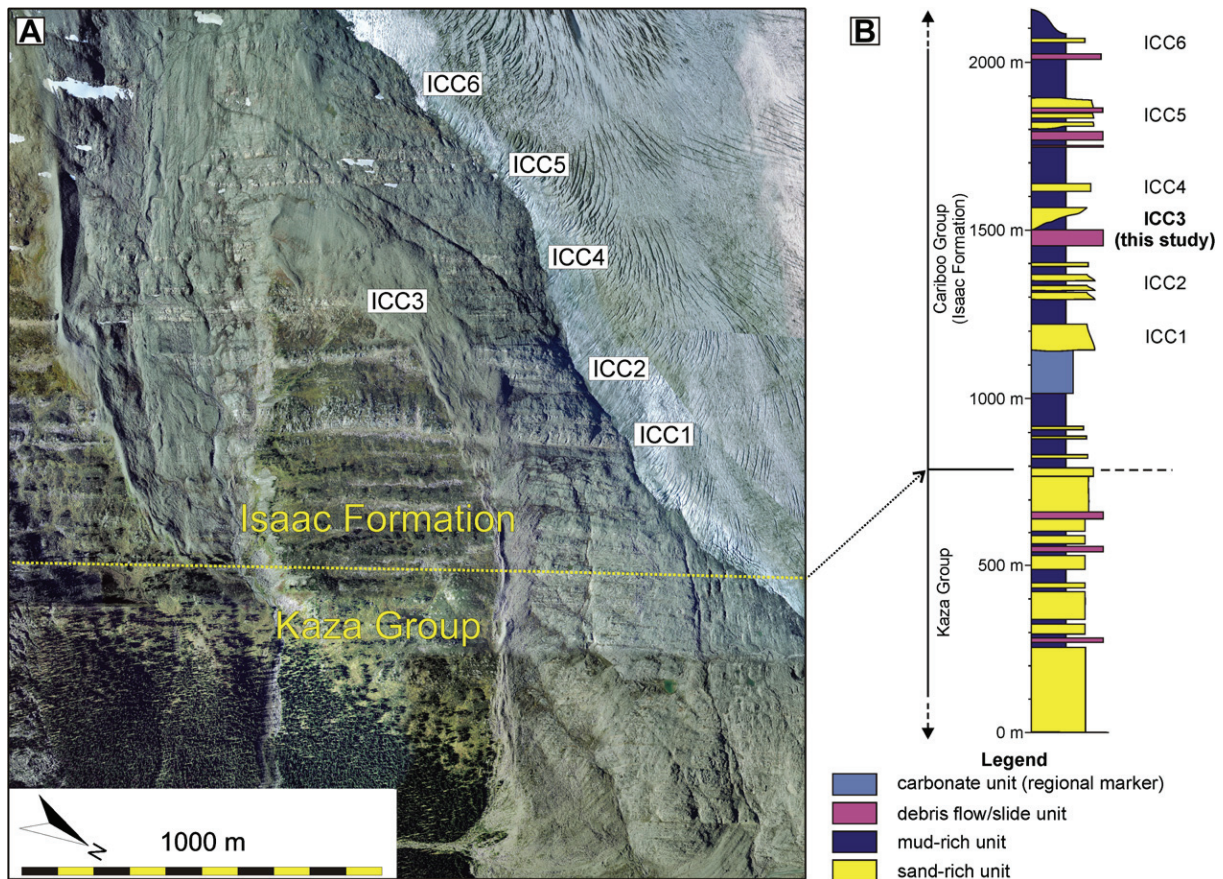


Fig. 3. A) Aerial photomosaic of the Castle Creek south study area. Beds strike SE–NW and are nearly vertical. The lower 800 m of stratigraphy comprises the upper member of the Kaza Group whereas the overlying 1.6 km of stratigraphy comprises the Isaac Formation. The location of Isaac Channel Complex Units 1–6 is shown (ICC1–ICC6). B) Generalized stratigraphic log of the Castle Creek study area (modified from Schwarz and Arnott, 2007).

channel and levees is consistent with levee formation resulting from channel flows overspilling the channel banks and continuing in a direction oblique to the main channel trend (DeVries et al., 1994; Hesse and Dalton, 1995; Hiscott et al., 1997; Hickson and Lowe, 2002; Fildani and Normark, 2004; Hubbard et al., 2008; Kane et al., 2010a). Paleocurrents were measured predominantly from the axes of 3D current ripple cross-stratification. These measurements were made where the outcrop was locally cut forming a 3D face.

4. Lithofacies of channel-fill strata in ICC3

The channel-fill portion of ICC3 is about 80 m thick and comprises four individual channel-fill units (C1–C4) that range from 7 to 50 m thick (Navarro et al., 2007). The lower two channel-fill units in outcrop are at least 1.5 km wide whereas the upper two units are less than 500 m wide (Fig. 4), however, apparent upward narrowing may be an artifact of the changing obliquity of the cut. Individual channel-fill units generally fine and thin upward and collectively form a large-scale upward-thinning and -fining trend. The coarsest strata, consisting of pebbly conglomerate, occur at the base of C1. Beds are up to 2.5 m thick and are laterally discontinuous over distances of hundreds of meters, typically because of scour by younger conglomerate beds. Upward these strata grade into beds of crudely normally graded, coarse-grained sandstone, which are up to 2.5 m thick and are laterally continuous, in some cases over several hundreds of meters. Still further upward these sandstone beds become progressively interbedded with thinly-bedded turbidites. C2, in contrast, consists of an extensive mud-

clast breccia at its base and conglomerate beds are less common. C3 and C4 are generally finer grained than C1 and C2, and at their bases consist typically of very coarse sandstone, however breccia layers are observed locally. C3 and C4 also fine- and thin-upward, but unlike C1 and C2 grade rapidly laterally over a few hundred meters from amalgamated coarse-grained sandstone into medium- to thick-bedded coarse sandstones turbidites interbedded with thin- to medium-bedded upper-division turbidites toward the channel margins.

5. Lithofacies of overbank strata in ICC3

In relation to the channel fill of ICC3, overbank strata are partitioned into 3 discrete locations: proximal outer-bend (POB), distal outer-bend (DOB), and inner-bend levee (IB) (Fig. 4). The IB levee is subdivided vertically into a lower (IB1) and upper (IB2) unit. Two main lithofacies are recognized in overbank strata of ICC3. These are 1) medium- to thick-bedded, coarse to fine sandstone T_{a-d} turbidites, and 2) very thin- to thin-bedded, fine sandstone to siltstone T_{cd} turbidites. Facies are briefly described next and are summarized in Table 1.

5.1. Medium- to thick-bedded, coarse to fine sandstone T_{a-d} turbidites (F1)

5.1.1. Description

F1 beds are most common in the POB but occur also in IB and DOB. Beds generally range from 20 to 70 cm thick although beds up to 1.5 m thick are observed. F1 beds are T_{a-d} turbidites that

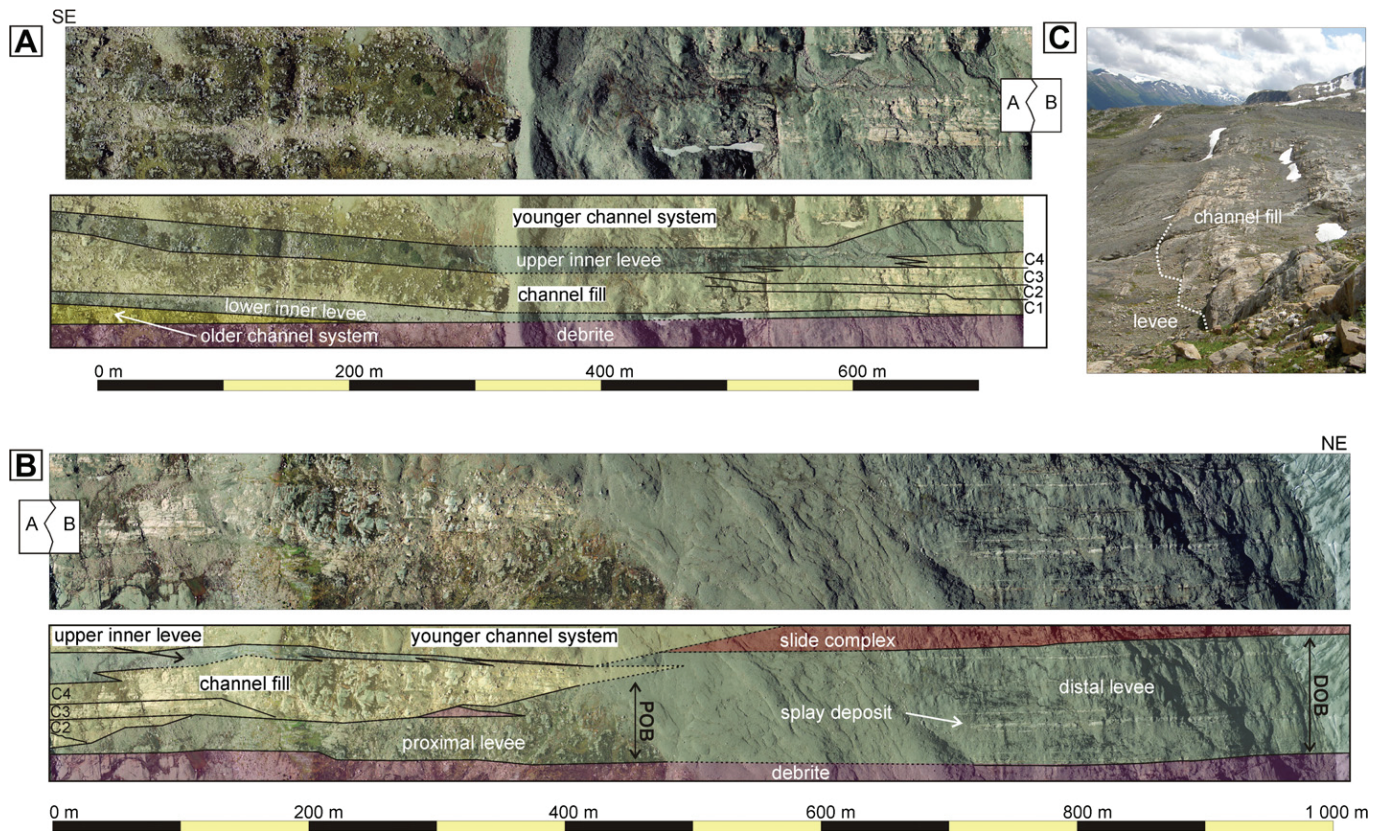


Fig. 4. Photomosaic of Isaac Channel Complex 3 showing the continuity of outcrop and exposure of levees along both sides of the channel. Channel-fill strata are divided into four units (C1–C4). Levee strata along the inner-bend (SE) side of the channel in (A) are divided into lower and upper units. Strata in the inner-bend levee consists predominantly of thin-bedded turbidites. Strata in the lower unit are truncated by the shallowly dipping channel margin whereas in the upper unit channel-strata thin and fine laterally away from the channel. In (B) levee strata in the outer-bend proximal levee range from very thin- to thick-bedded turbidites. Note the occurrence of an up to 10 m thick debrite within the outer-bend proximal levee. Collectively these deposits are sharply separated from channel-fill deposits along the terraced channel margin (C). Proximal levee strata transition laterally over hundreds of meters into principally thinner-bedded distal levee deposits that locally are intercalated with up to 5 m thick amalgamated overbank splay units. ICC3 is bounded at its base by a debrite and older channel system and at its top by a slide complex and younger channel unit.

normally grade from moderately-sorted medium to fine sandstone, however, the basal part of some beds consist of massive to subtly graded coarse-grained sandstone (Fig. 5). Beds are typically overlain abruptly by a 1–5 cm thick layer of coarse to fine siltstone marking a sharp grain size break. Less commonly beds lack an upper siltstone (T_d) cap and form amalgamated units up to 5 m thick, the thickest of which occur in the lowermost proximal levee and also along discrete stratigraphic levels in the distal levee.

5.1.2. Interpretation

Structureless T_a divisions at the base of some F1 sandstone beds were most likely deposited rapidly from suspension (Lowe, 1988; Arnott and Hand, 1989; Sumner et al., 2008). Planar laminated (T_b) divisions associated with F1 indicate deposition from flows where the rate of suspension fallout was sufficiently low that the distance of bedform migration became long enough that planar lamination became visible (Paola et al., 1989). Accordingly, the transition from T_a to T_b deposition indicates a temporal reduction in sedimentation rate. Non-climbing ripples are common in levee deposits of ICC3 and either overlie T_a or more commonly T_b divisions, or form beds up to 40 cm thick composed entirely of the T_c division. These ripple cross-stratified successions typically fine upward from medium- to fine-grained sandstone and likely represent competence-driven deposition in which the fallout rate was never so high as to create a perceptible angle of climb. The thin, markedly finer-grained (silt) T_d layer that caps most turbidites is interpreted to have been deposited during the final waning stage of

flow. The missing intervening grain sizes (i.e. lower fine and very fine sand fraction), on the other hand, are interpreted to have been maintained in suspension during the earlier, higher energy stages and transported into the distal levee.

5.2. Very thin- to thin-bedded, fine sandstone to siltstone T_{cd} turbidites (F2)

5.2.1. Description

F2 sandstones occur in both proximal (inner- and outer-bend) and distal levee deposits. Beds range from 1 to 10 cm thick and consist of T_{cd} turbidites with rare T_b bases (Fig. 6a,b). Beds commonly grade from fine sandstone to siltstone and unlike F1 commonly show no sharp break in grain size. The basal T_c portion of these beds is up to 8 cm thick and comprises single to multiple (up to four) non-climbing ripple cross-stratified sets. In the uppermost part of T_c divisions, individual ripple foresets are commonly inter-laminated with finer-grained sandstone and silt drapes (Fig. 6c). T_d divisions typically range from 2 to 8 cm thick and are locally faintly parallel laminated but most commonly appear structureless. Where present, T_d laminae are commonly characterized by interlaminated thin (<2 mm) light grey coarse siltstone or very fine-grained sandstone and dark grey finer-grained siltstone (Fig. 6d).

5.2.2. Interpretation

The deposition of F2 is interpreted to be largely from dilute, fine-grained overspilling turbidity currents. Similar to F1 the lower

Table 1
Characteristics of overbank lithofacies.

Lithofacies	Lithology	Bed thickness	Sedimentary structures	Bouma divisions*	Bounding surfaces	Lateral trend	Depositional process
F1	Coarse- to fine-grained sandstone	10–150 cm	Massive, parallel laminated, and ripple cross-stratified	Most common T_{bc} , T_b , and T_c , less common T_a , T_{ab} , and T_{abc} . Beds are typically overlain by thin massive T_d cap	Sharp bases and tops	Thicken laterally over ~120 m away from the channel and thereafter thin over the next 100's of meters	Unconfined, depletive waning flows
F2	Fine-grained sandstone to siltstone	<1–10 cm	Generally ripple cross-stratified, rare massive to planar laminated. Siltstone cap are generally massive and rarely laminated	Most common T_{cd} , less common T_{abcd} , and T_{bcd}	Sharp bases and tops	Grain size and bed thickness change negligibly laterally	Unconfined, uniform waning flows

*Follows the nomenclature and classification proposed by Bouma (1962).

ripple cross-laminated portions of beds do not climb indicating that sedimentation rates were consistently low. Also, the gradual upward change from fine sandstone to siltstone suggests a gradual reduction in flow competence as finer sediment was progressively transferred from the suspended load to the bedload layer. Common structureless T_d divisions in ICC3 levee strata are interpreted to represent the very fine-grained suspension settling toward the end of a flow event. Rare T_d lamination recorded as coarse siltstone or very-fine sandstone interlayers within finer-grained siltstone are interpreted to have been created by the presence of internal waves within the channelized turbidity current. This in turn potentially caused fluctuations (or pulses) in the hydraulic regime and grain-size make up of the overspilling flow (e.g. Hiscott et al., 1997; Gervais et al., 2001). Alternatively, Stow and Bowen (1980) suggest that the fine-scale interstratification of coarser and finer layers is the result of rhythmic changes in flocculation/deflocculation and sediment settling processes within the viscous sublayer of a low-energy turbidity current.

6. Geometries and spatial trends of overbank elements

6.1. Outer-bend proximal levee

6.1.1. Description of vertical packaging

The proximal outer-bend levee is subdivided vertically into three stratal packages (POB1–POB3 ; Fig. 7) that range from 12 to 28 m thick. Individually, each package is divided into lower and upper parts. The contact between these two units is clearly marked by an abrupt decrease in average bed thickness and sandstone proportion (Figs. 7b,c and 8a,b). Previously, Navarro et al. (2007) divided the proximal levee of ICC3 into four units. The top unit, however, lacks a well developed lower part and here makes up the majority of the upper part of POB3.

The lower part of packages range from 7 to 11 m and consist of intercalated medium- to thick-bedded and very thin- to thin-bedded turbidites of F1 and F2, respectively (Fig. 7b,c). F1 strata account for about 60% of the total thickness of the lower part of packages and individual beds average about 40 cm in thickness. The basal sandy part of intercalated F2 strata, on the other hand, generally range between 1 and 2 cm thick, but on average account for about 70% of the total number of sandstone beds recorded in the unit. Accordingly, the average sandstone bed thickness (10–20 cm) is biased by the abundance of very thin- to thin-bedded F2 strata (Fig. 8a).

The upper part of levee packages range from 5 to 18 m thick and consist of no less than 85% very thin- to thin-bedded F2 strata. The sandy part of F2 beds, like those in the lower part of packages, typically range from 1 to 2 cm thick. Although significantly less common, intercalated beds of F1 in the upper part of POB1 and POB2 average about 22 cm thick and account for about 30% of the total thickness. As a result average sandstone bed thickness is disproportionately increased to about 4.5 cm (Fig. 8a). F1 strata in the upper part of POB3, in contrast, account only for 6% of the total thickness and accordingly have little impact on average bed thickness. Moreover, the ratio of sandstone thickness to total stratal thickness, or simply N:G, progressively decreases upward (Fig. 8b) – note that sandstone is considered to consist of sediment coarser than very fine sand.

6.1.2. Description of lateral trends within packages

In the outer-bend area levee and channel-fill strata are separated by an erosive and terraced surface that locally dips up to 25° (Fig. 4). Here, F1 strata show a distinctive lateral thickening then thinning trend (Fig. 9). Specifically, individual beds typically thicken rapidly laterally over about 120 m exhibiting an average

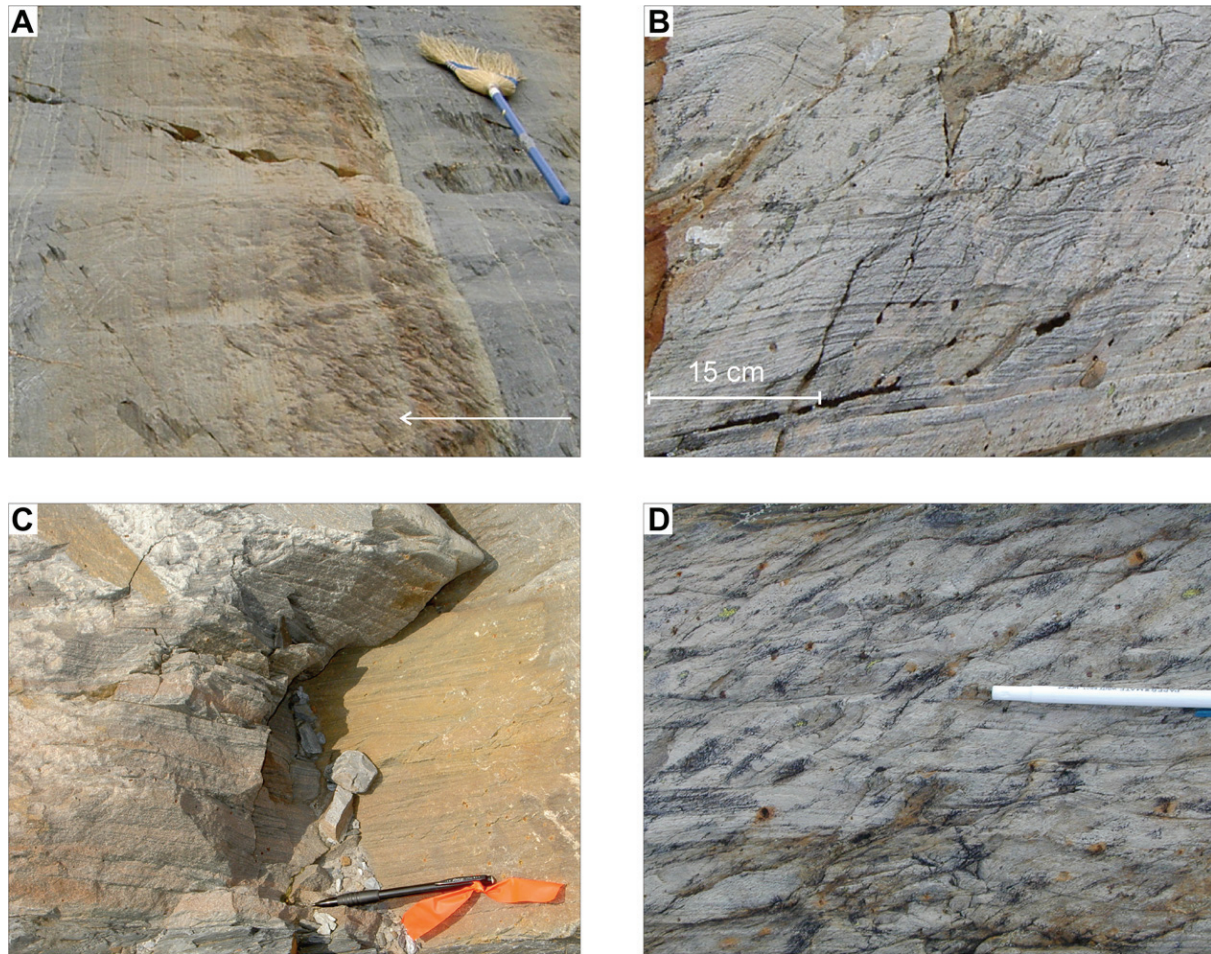


Fig. 5. Facies F1 strata (medium- to thick-bedded, coarse to fine sandstone T_{a-c} turbidites). A) Thick-bedded T_{ab} turbidite that fines upward (towards the left) from medium to fine-grained sandstone. The transition from massive to planar laminated occurs approximately at mid height of the bed. B) Thick-bedded T_{bc} turbidite that fines from medium to fine-grained sandstone. The planar laminated part of this bed occurs only in the lower 6 cm. C) Medium-bedded, medium-grained sandstone T_b turbidite with minor vertical fining. D) Thick-bedded, medium- to fine-grained T_c turbidite consisting of multiple sets of non-climbing ripples. Bed fines vertically from medium-grained to upper fine-grained sandstone.

thickening gradient of 0.3 (thickening (cm))/lateral distance (m)). After achieving their maximum thickness, beds begin to thin over the next 180 m with an average thinning ratio of 0.2. Once beds thin to about 10 cm or less, and now are considered to be F2, they show negligible lateral thinning over the remaining length of the outcrop (up to 50 m). Similarly, beds in the proximal levee that are initially less than 10 cm thick (F2) show negligible lateral thinning along their entire length. Although rare, some F1 and F2 beds are observed to pinch out entirely within the proximal levee. Moreover, lateral thinning of F1 strata generally coincides with a lateral fining trend wherein grain size typically decreases from medium- (up to coarse-) to fine-grained sandstone. In contrast grain size in F2 strata changes little laterally.

The lateral trends in bed thickness directly control lateral changes in N:G. For example, Fig. 10a shows that the sandier lower part of levee packages, which in terms of overall thickness are composed mostly of F1 sandstones, are on average at their maximum N:G 120 m laterally away from the channel reaching values ranging from 0.6 to 0.8. Subsequently, N:G decreases rapidly over the next 180 m away from the channel to values as low as 0.46. In the upper part of levee packages, strata consist mostly of lower N:G F2 beds that show negligible lateral change in thickness or grain size, and therefore N:G. Nevertheless, it is the thicker but much less common F1 beds that control the lateral N:G trend. For instance, in

the upper part of POB1 and POB2 N:G begins to decrease appreciably only at 300 m and 150 m (Fig. 10b), respectively, which coincides with the location where F1 strata rapidly transition laterally into thinly-bedded F2 strata (see Fig. 7b). The upper part of POB3, in contrast to underlying packages, consists of relatively few F1 beds and therefore N:G changes negligibly over 100 m (Fig. 10b). A single levee package (lower + upper part) yields maximum N:G values up to 0.68 close to the channel margin and as low as 0.34 over the next few hundred meters laterally away from the channel margin (Fig. 10c).

6.2. Outer-bend distal levee

6.2.1. Description of distal levee

The outer-bend distal levee is subdivided into a lower (DOB1) and upper package (DOB2; Fig. 11). DOB1 is about 80 m thick and consists predominantly of fine-grained F2 strata interbedded with less common coarser grained F1 beds. Locally these strata are interbedded with up to 3 m-thick units made up of contorted F2 strata (i.e. slump deposits). The sandy base of F2 beds in DOB1 range from <1 cm to 8 cm thick, but typically range from 1 to 2 cm. Initially beds of F2 appear to be continuous across the width of the outcrop (300 m wide) and show negligible lateral thinning. However based upon five closely-spaced (~40 m apart), 2 m thick

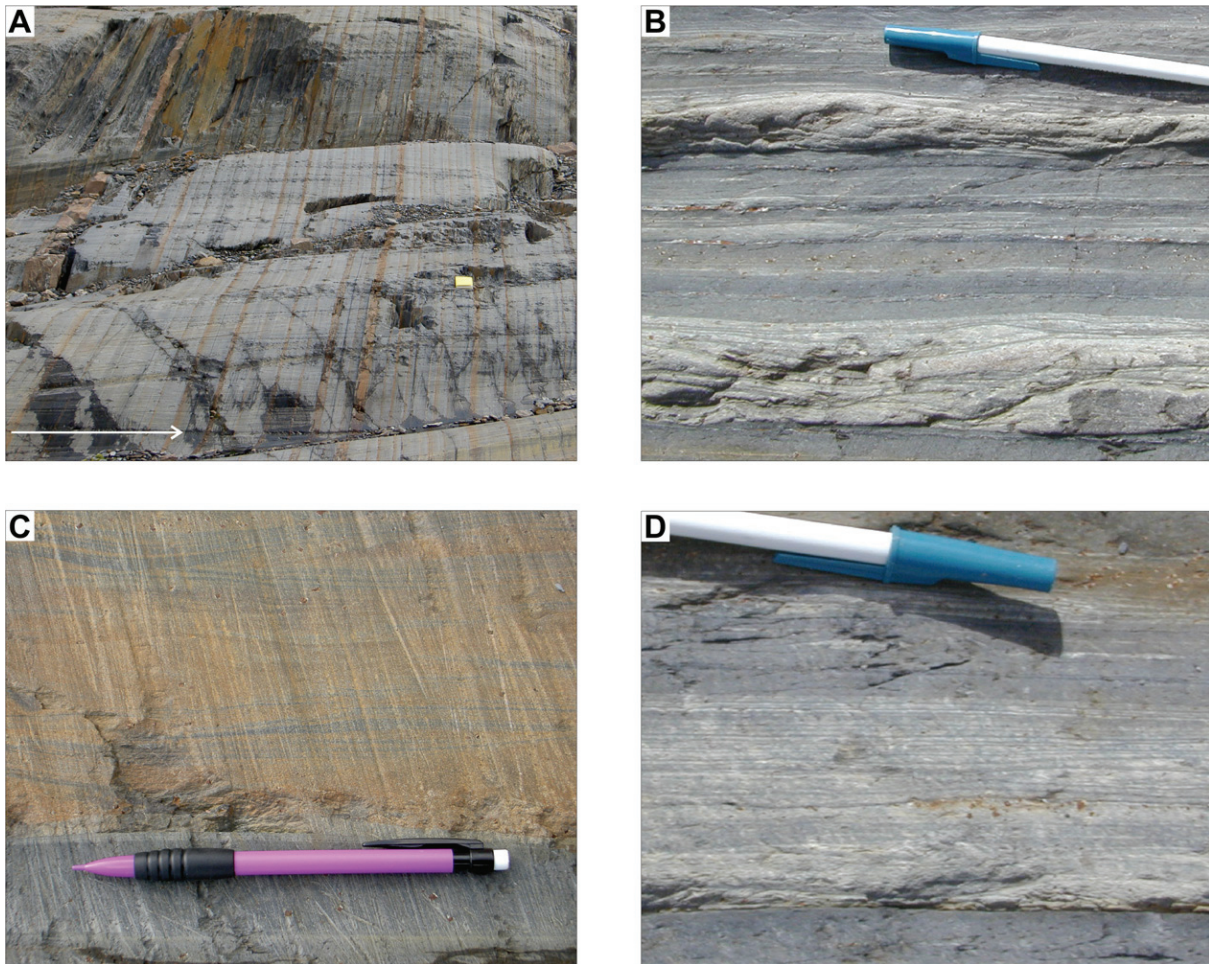


Fig. 6. Facies F2 strata (very thin- to thin-bedded, fine sandstone to siltstone T_{cd} turbidites). A) Predominantly thin-bedded, fine sandstone to siltstone T_{cd} turbidites (yellow notebook for scale). B) Close up of very thin- to thin-bedded T_{cd} turbidites. C) Fine-grained T_c turbidite with interlaminated finer-grained drapes. D) Interlaminated thin light grey very fine-grained sandstone and dark grey siltstone.

stratigraphic logs (see Fig. 11) it is noted that the number of beds varies between logs (specifically bed number between logs 1 and 2 is 24, increases to 29 in log 3, decreases to 26 in log 4, but then increases to 31 in log 5), suggesting that within this small window some number of beds pinch out whereas others become initiated. Interbeds of F1 strata range from 10 to 100 cm thick and average about 26 cm. Generally these beds thin away from the channel (i.e. toward glacier), although some thin and even pinch out toward the channel. There is no obvious upward thinning or fining trend within the lower 62 m of DOB1, although F1 beds become noticeably less common above about 45 m. Specifically, F1 strata account for 27% of the total number of event bed in the lower half but only 12% in the upper half. The uppermost 18 m of stratigraphy in DOB1 exhibits a well developed thinning-upward trend wherein average sandstone thickness decreases from 4 to less than 1 cm (Fig. 12). Collectively, bed thickness trends are recorded in the N:G, which in the lower half is about 0.44 and in the upper half averages about 0.29 (Fig. 13). The thinning-upward trend in the uppermost 18 m is mirrored by a gradual upward decrease in N:G from about 0.4 to 0.08. Furthermore, F1 and F2 strata are intercalated with 2–5 m thick amalgamated units made up of distinctively thicker, coarser amalgamated F1 strata that lack intercalated siltstone layers. These units occur along four horizons: at the base of the levee succession and at about 28 m, 40 m, and 65 m above the base. Amalgamated units are laterally continuous along their entire exposed length,

which is up to 300 m. Although the units are laterally extensive they generally thin appreciably towards the glacier. Internally beds are typically laterally discontinuous, commonly being eroded by an overlying bed in less than 100 m. The uppermost package (DOB2) is up to 12 m thick and is overlain by an erosion surface that marks the base of a thick slide deposit. This levee package consists of intercalated F2 and F1 strata with an average N:G of 0.48. In addition, a 2–3 m thick amalgamated unit occurs 2 m above the base.

As previously stated the majority of F1 beds in the proximal levee thin and fine to beds of F2 over less than 300 m laterally thus implying that the majority of F1 strata in the distal levee are not continuous beneath the moraine. In contrast, some if not many F2 beds in the proximal levee probably correlate across the moraine to F2 beds in the distal levee. Nevertheless, the fact that some F2 beds pinch out towards the moraine in both the proximal and distal levees suggests that at least some of these beds are physically discontinuous beneath the moraine. Moreover, the uppermost distal levee package (DOB2) may represent more proximal levee deposition associated with the lateral migration of ICC3. This would explain the relatively higher N:G in comparison with the upper half of the immediately underlying package. Alternatively, DOB2 may represent the proximal levee associated with the younger overlying channel system (see Fig. 4). Due to the local moraine cover neither scenario can be substantiated and therefore this unit is not considered in the evolution of ICC3.

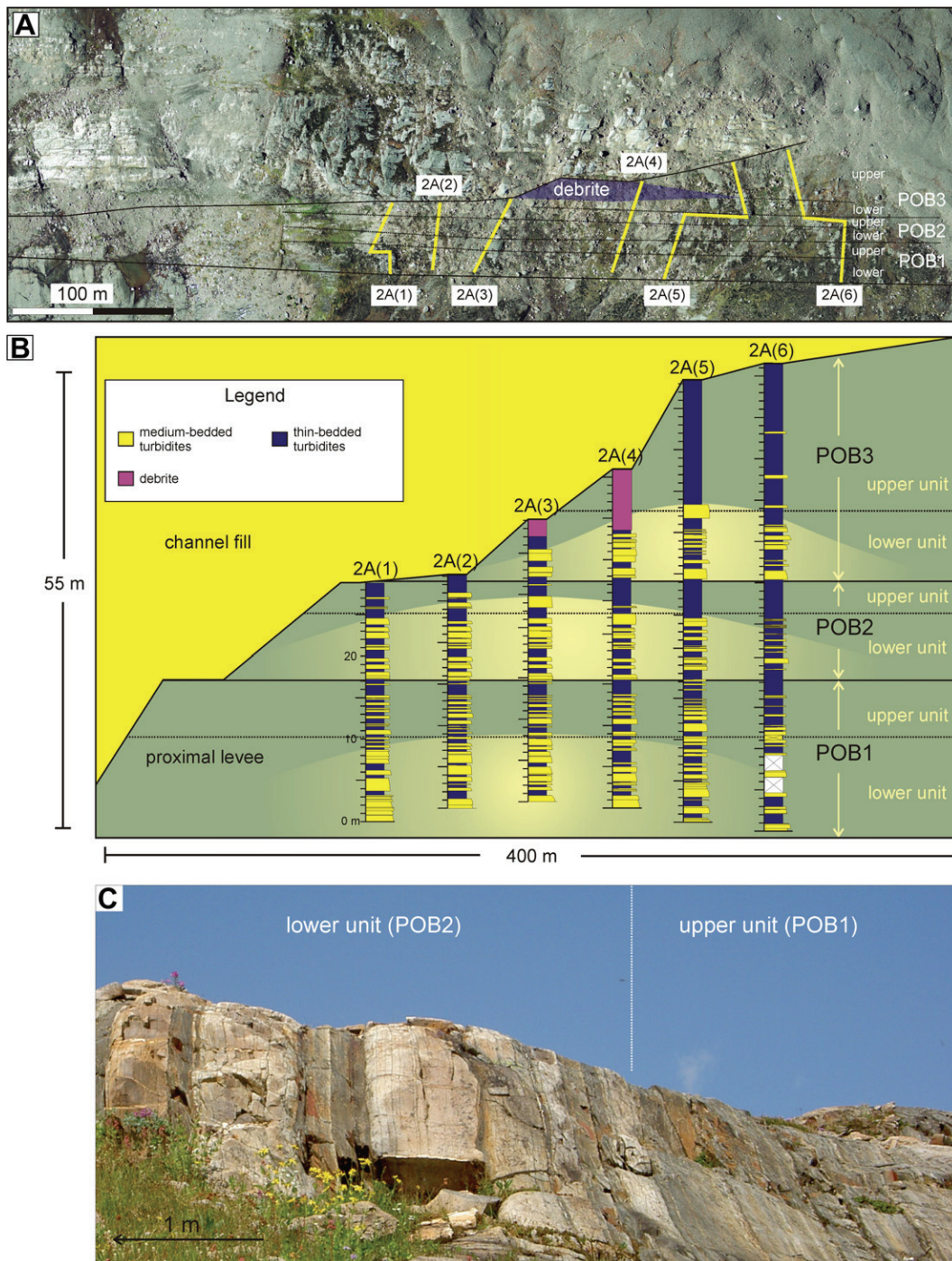


Fig. 7. Levee deposits in the proximal outer-bend levee. A) Aerial photo showing location of measured stratigraphic logs. B) Correlation panel showing the detailed architecture of proximal levee deposits. Proximal levee deposits comprise three distinct packages that are divided into a lower part that comprises interbedded thick-bedded (F1) and thinner-bedded (F2) turbidites and a more siltstone-rich, thin-bedded (F2) upper part. Beds in the sandier lowermost parts of packages typically thicken then thin laterally. As a result, lower packages as a whole typically increase and then decrease in N:G laterally away from the channel margin (represented by the lateral shade grading). Note that levee packages POB1 to POB3 show a consistent lateral offset that is especially noticeable when comparing POB2 with POB3 – this pattern mirrors the direction of lateral migration of the channel. C) Photograph showing the sharp contact between the upper part of POB1 and the lower part of POB2.

6.3. Inner-bend levee

6.3.1. Description of inner-bend levee

Deposits along the inner-bend levee of ICC3 are divided into lower and upper units, respectively about 17 m and 40 m thick (Fig. 4). Strata of the lower inner-bend levee (IB1) are observed to be truncated by channel-fill deposits. Unlike in the outer-bend, the

inner-bend channel margin is not terraced but instead shows a more consistent dip towards the channel axis of about 2–4° (Fig. 14a). IB1 consists largely of F2 strata that, like those on the outer bend, show negligible lateral thinning or fining. Although the majority of beds are thin-bedded turbidites, less common coarser, thicker F1 beds are observed in the lower 5 m of the section most proximal to the channel (Fig. 14b). In contrast to the proximal outer-

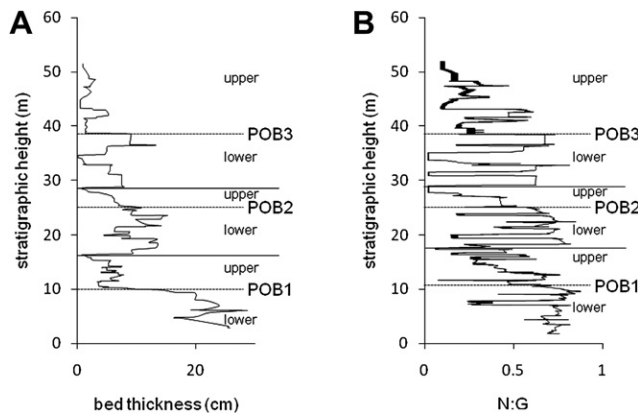


Fig. 8. A) Data displayed from 474 sandstone beds (not including siltstone tops) showing how bed thickness varies upward. Data for POB1 and POB2 is from Log A2(1) whereas data for POB3 is from Log A2(5.5). Log A2(5.5) is not shown in this paper but was measured between logs A2(5) and A2(6) (see Fig. 7b). B) Data displayed from 524 beds (including siltstone beds) measured in Log A2(5.5) showing upward trend in N:G. A decrease in N:G indicates an upward fining trend. These plots use a 15-bed moving average.

bend levee where there is a relatively abrupt vertical transition from F1 to F2 strata, here on the inner-bend levee a more gradual upward-thinning trend is recorded (Fig. 14b). Moreover, F1 beds at the base of the section thin away from the channel margin over a distance of less than 100 m from 23 cm to less than 10 cm thick before onlapping the surface of an older channel unit (Fig. 14a). As a result of this lateral thinning N:G decreases away from the channel from 0.40 to 0.15 in which the majority of the decrease takes place within the first 50 m (Fig. 14c).

Similar to IB1 the upper inner-bend levee (IB2) is composed mostly of F2 strata. The channel geometry associated with IB2, however, differs markedly from that of IB1. Here some thick-bedded, coarse-grained, amalgamated channel-fill deposits in the axial channel grade laterally over about 140 m into fine, thinly-bedded turbidites that become progressively interbedded with other F2 inner-bend levee strata (Fig. 15). More specifically, amalgamated channel-fill strata up to 2 m thick thin laterally into either a single or multiple F2 beds, forming a distinctive interfingering pattern (Fig. 15a–e). N:G values reflect these lateral trends decreasing from 0.9 immediately adjacent where channel-fill strata begin to split, to 0.15 140 m laterally after which N:G remains constant (Fig. 15f).

7. Levee evolution

This section assembles the data presented earlier to illustrate the stratigraphic evolution of ICC3 and highlight the asymmetry of outer- and inner-bend levee deposits. The inception of ICC3 is interpreted to have been controlled by topography along the top of

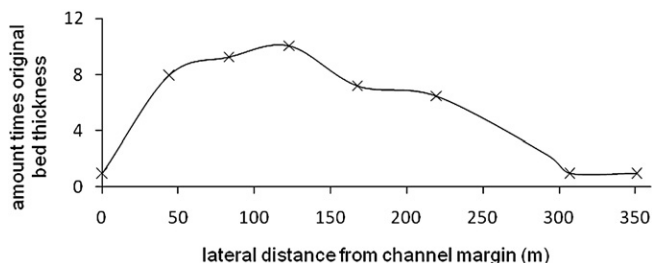


Fig. 9. Average thickening then thinning trend of medium-bedded sandstone in the lower part of levee packages. This trend is the average of 25 thick- and medium-bedded sandstones that display a well developed thickening then thinning trend in the lower part of levee packages.

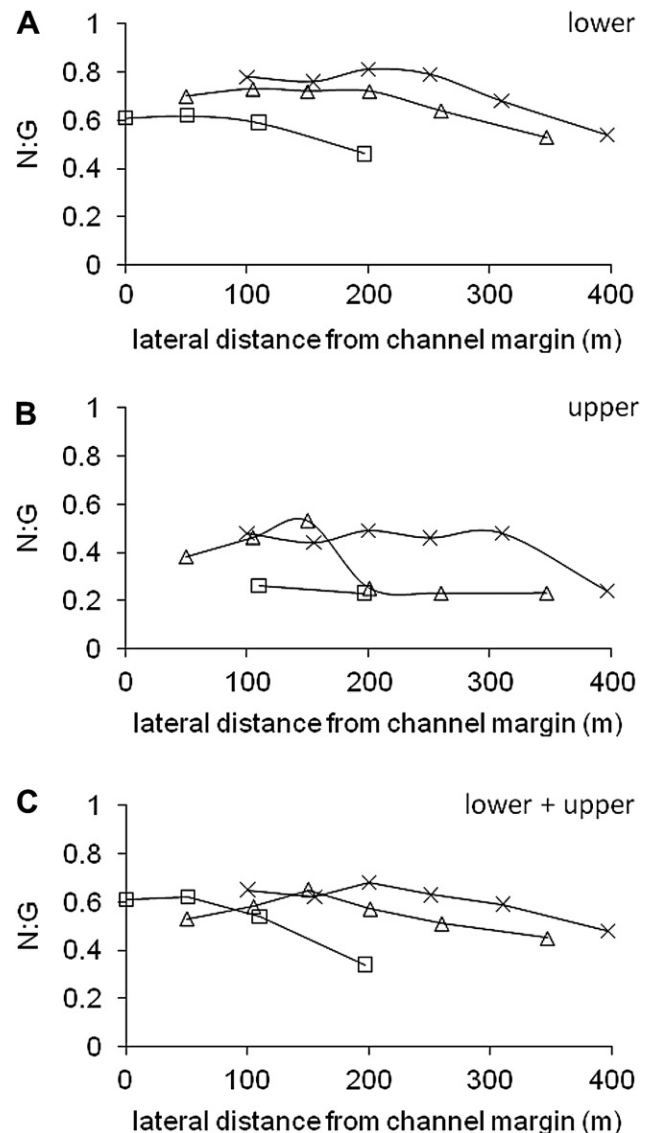


Fig. 10. Lateral variation of N:G in the proximal outer-bend levee (X = POB1; triangle = POB2; square = POB3). A) N:G trends in the lower part of packages. In places strata immediately adjacent the channel margin are not well exposed and as a result N:G could not be measured there. Nevertheless the existing database suggests that F1 beds thin toward the channel, which in turn decreases N:G. B) N:G trends in the upper part of packages. The paucity of F1 strata in the upper part of POB3, in comparison to POB1 and POB2, is reflected by a negligible lateral change in N:G. C) N:G trends of complete levee packages (lower + upper parts).

a thick debrite that caps an areally extensive mass transport deposit (Fig. 16a). Rugosity along the top of debrites and localization of overlying channel-levee systems has been reported from other deep-marine systems, including the deep-water part of the Niger Delta (Samuel et al., 2003), Indus Fan (Deptuck et al., 2003), and the ancient Tres Pasos Formation (Armitage et al., 2009). Following inception, the evolution of ICC3 consisted of three major episodes of channel-levee growth (Channel–Levee Units 1–3), which are all recorded in the outer bank proximal levee packages POB1 to POB3 (Fig. 7). Only the correlative inner-bend levee deposits of packages POB1 and POB3 are exposed, which respectively are IB1 and IB2. As for the channel, C1 is genetically related to POB1 and IB1, C2 to POB2, and C3 and C4 to POB3 and IB2. Individual episodes of channel-levee growth are characterized by three successive stages of levee growth and channel fill.

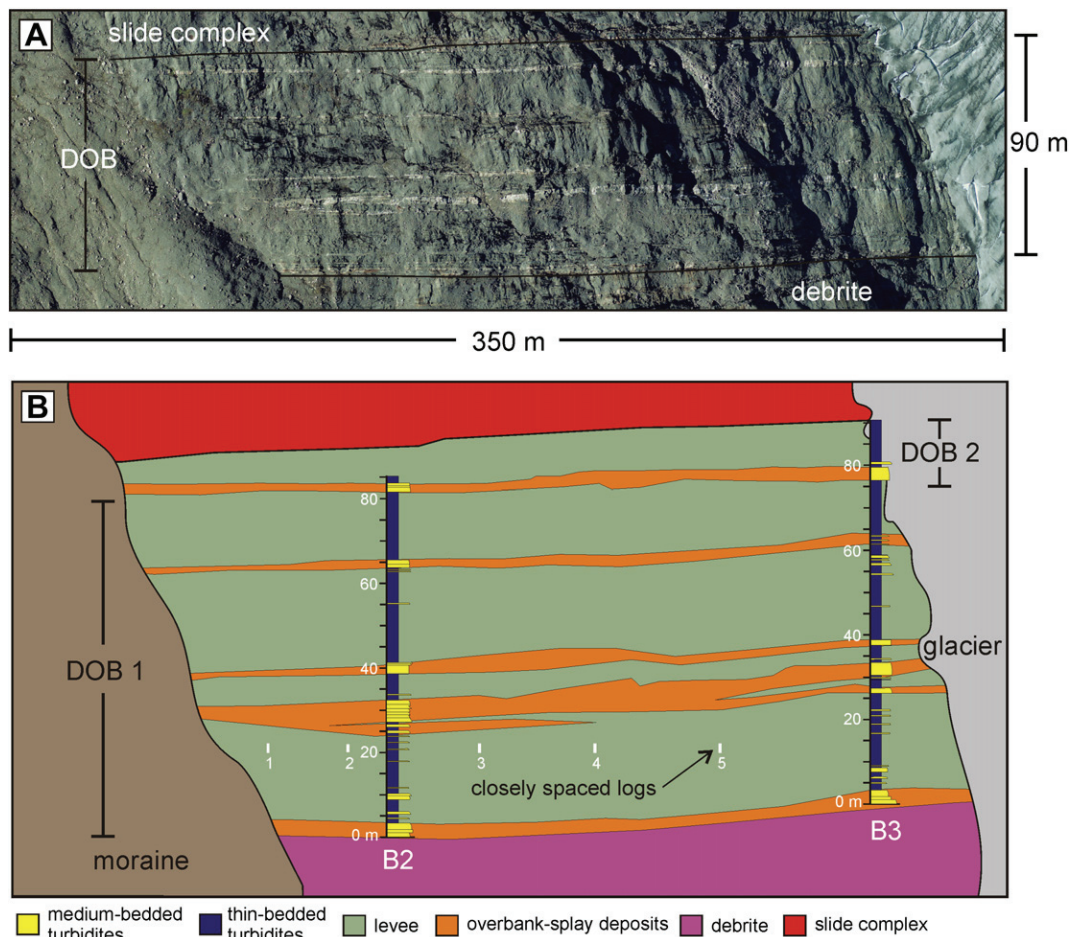


Fig. 11. Strata in distal outer-bend levee (DOB). A) Aerial photomosaic of DOB showing the contacts between the underlying debrite and overlying slide complex. B) Interpreted panel showing the distribution of overbank splay deposits and the contact between DOB1 and DOB2. Also included are two measured litho-stratigraphic logs and the location of the closely-spaced logs used for lateral correlation (see Section 6.2.1).

7.1. Development of Channel-Levee Unit 1 (CLU1)

7.1.1. Low levee relief (Stage 1)

Stage 1 of the first channel-levee unit (CLU1) began immediately after the inception of ICC3, at which point the channel is estimated to have been a few hundred meters wide and confined by topographic highs along the debrite surface no more than a few meters thick. This weakly confined system allowed the lower, coarse-grained part of relatively large magnitude turbidity currents to overspill resulting in the deposition of medium- to thick-bedded,

coarse- to fine-grained turbidites (F1) along both sides of the channel (Fig. 16b). These thicker-bedded deposits are recorded in the lower part of IB1 and POB1. The greater abundance and thickness of F1 strata on the outer-bend levee suggests that fluid and sediment inertia effects (i.e. superelevation and run-up) enhanced deposition there.

Baas et al. (2004) showed experimentally that the maximum rate of deposition associated with denser turbidity currents occurred further from the point of unconfinement compared to slower, less dense ones. This explains the initial thickening trend

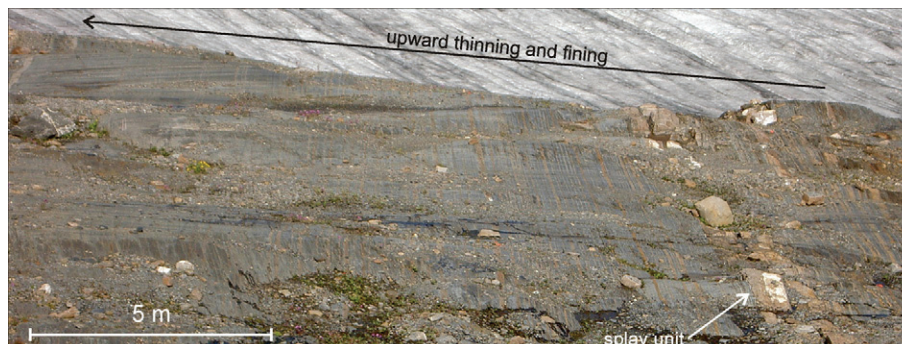


Fig. 12. Photograph showing the thinning- and fining-upward trend observed in the upper 18 m of the distal outer-bend levee (DOB1).

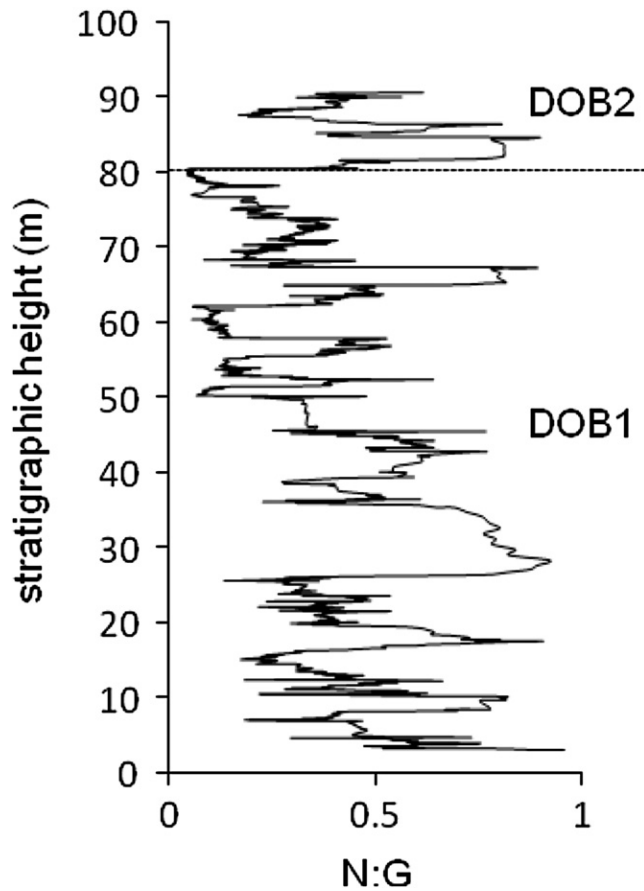


Fig. 13. Data displayed from 531 beds (including siltstone beds) measured in Log B2 showing upward trends in N:G. There is an abrupt decrease in average N:G above about 45 m which coincides with a decrease in the frequency of F1 strata. In the upper 18 m of DOB1 N:G progressively decreases upward, which is reflected by a well developed thinning-upward trend (see Fig. 12). Note that a splay unit at about 65 m (stratigraphic height) causes a local spike in N:G. This plot uses a 15-bed moving average.

observed in many F1 sandstones in the outer-bend levee where inertia associated with the overspilling of high energy, highly concentrated flows displaced the zone of maximum sedimentation about 120 m away from the channel (Fig. 9). Beyond this zone of maximum deposition, F1 beds began to fine and thin laterally over the next 180 m, suggesting a more progressive loss in flow competence and capacity. Eventually these sandy flows responsible for the deposition of F1 strata became depleted laterally in their coarser grained fraction to the point where the remaining particles in suspension would be characterized by sufficiently low settling that turbulent mixing created an almost uniformly distributed sediment dispersion (Kneller and McCaffrey, 2003; Baas et al., 2004). Deposition under such conditions resulted in the deposition of F2 strata wherein negligible lateral thinning or fining occurs and therefore, explains the common lateral transition from F1 to F2 strata.

Interstratified finer, thinner strata of F2 within the lower part of the levee were similarly deposited during conditions of low channel confinement, but by channel flows of significantly smaller magnitude in which the coarser basal part of the flow remained confined entirely within the channel. The upper part of the flow that did overspill was sufficiently fine-grained and well sorted that sediment distribution was more or less uniform resulting in the deposition of F2 strata. The intercalation of F1 and F2 strata in the lower parts of levee packages suggests that flow magnitude within

the channel commonly fluctuated. In the lower part of levee packages about 70% of the beds are F2 implying that most overspilling flows were comparatively small.

The abundance of generally thicker-bedded strata along the outer-bend channel margin created an asymmetry in levee height. If it is assumed that each turbidite on the outer bend has an inner-bend equivalent, deposited contemporaneously during a single overspilling event, then it is possible to quantify the difference in levee height at any given time. For example, the lower part of POB1, interpreted to have accumulated entirely by the end of Stage 1, consists of 64 beds yielding a total thickness of 10.5 m. Accumulation of the first 64 beds in the lower inner-bend levee (IB1), in contrast, is 5 m thick. This implies that after the first 64 overspilling events of ICC3 the outer-bend levee was about two times higher than its inner-bend counterpart.

During this early stage of levee growth, flows in the channel are interpreted to have bypassed most of their sediment load basinward leaving behind coarse-grained lag deposits and thin-bedded, fine-grained tail-of-flow turbidites (Navarro et al., 2007). Bypassed flows continuously scoured outer-bend levee deposits as the channel thalweg shifted laterally. This modified the channel margin facing into the channel, which locally dips up to 25°. As levee height increases, so too does the slope along the levee backside (Straub and Mohrig, 2008; Kane et al., 2010b). Under certain conditions, this increase in gradient is interpreted to cause overspill flows to become supercritical leading to the formation of sediment waves (Migeon et al., 2001, 2004; Lee et al., 2002; Normark et al., 2002) or large-scale scour features (Fildani et al., 2006). There is no clear evidence of such large-scale features in ICC3, however, it is interpreted that slope along the backside of the levee promoted bypass, albeit limited, and deposition on the distal levee. At the same time, but on the inner-bend levee, diminished flow height and strength resulted in reduced erosion forming a shallow dipping (up to 4°) ramp towards the channel axis.

7.1.2. High levee relief (Stage 2)

Increased levee relief during Stage 1 confined the then approximately 500 m wide channel by 10.5 m high outer- and 5 m high inner-bend levees. This added relief along the outer-bend margin is interpreted to represent a threshold height where after only the upper, more dilute part of relatively large magnitude throughgoing channel flows overspilled (i.e. flow stripping). Accordingly, the lower part of POB1 is abruptly overlain by predominantly thin-bedded, fine-grained sandstone turbidites (F2) whose deposition marked the beginning of Stage 2 (Fig. 16c). At the same time, deposition along the inner bend, which during Stage 1 had experienced comparatively minor levee aggradation, was marked by even thinner and finer-grained F2 strata emplaced by increasingly more dilute overspill flows. The gradual thinning-upward trend observed in the lower inner-bend levee (IB1) demonstrates that effects of levee aggradation on overspilling flow dynamics were not as dramatic compared to the outer-bend levee where an abrupt facies transition occurs between Stage 1 and Stage 2 deposition.

Importantly, it is likely that during this period of high levee relief that small magnitude flows passing through the channel were confined completely within the channel or that only the very dilute uppermost part of the flow would be stripped and enter into the levee area. Such dilute overspilling flows, however, would lack sandy sediment resulting in the inability to form the sandy ripple cross-stratified basal part of F2 beds and consequently would not be differentiated in the field as individual events but instead considered part of an earlier T_d layer. If a similar ratio of small to large magnitude flows calculated from the lower part of levee packages is assumed for the upper part, then it is likely that on average 70% of throughgoing channel flows during periods of high

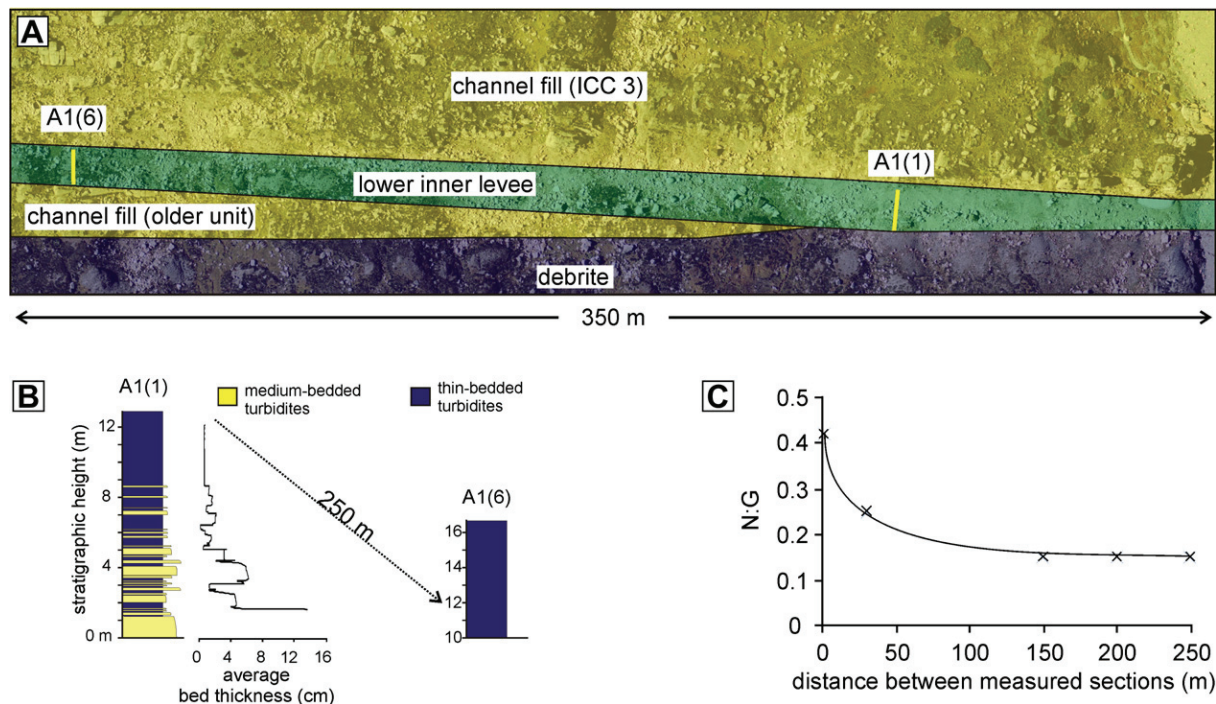


Fig. 14. Strata in the lower inner-bend levee (IB1). A) Aerial photomosaic of the lower inner-bend area of ICC3 showing locations of measured litho-stratigraphic logs in (B). C) Trend in N:G laterally away from the channel margin based on five measured logs (only the first and last logs are shown in A).

channel confinement did not overspill enough sandy sediment to be recorded within the levee as an individual event.

Minor channel filling likely occurred throughout Stage 2, however, it is assumed that flow in the channel was largely in a state of bypass and for the most part continued to erode and modify the channel margins. Growth of the outer-bend levee is also interpreted to have continued to outpace its inner-bend counterpart throughout Stage 2 sustaining limited deposition on the outer-bend distal levee. Differences in levee relief along opposing channel margins is estimated by assuming that the upper part of levee package POB1, which consists of 90 beds and is about 7 m thick, accumulated largely by the latter part of Stage 2. Accordingly, the total thickness of levee package POB1 (upper + lower parts), which consists of 154 beds and a thickness of 17.5 m, estimates levee relief that existed along the outer bend towards the end of Stage 2. The first 154 beds of IB1 (including the 64 beds deposited during Stage 1), in contrast, have a cumulative thickness of 9.5 m. This implies that the outer-bend levee was slightly less than two times higher than its inner-bend counterpart by the end of Stage 2 suggesting that levee asymmetry changed little during the first two stages of channel-levee growth. Also, erosion along the outer and inner bend approximately doubled the width of the channel, however, due to increased levee relief the width to height ratio of the channel did not change significantly during Stage 2.

7.1.3. Onset of channel-filling and distal levee deposition (Stage 3)

Stage 3 was characterized by the onset of channel deposition that significantly outpaced levee aggradation and progressively reduced channel confinement (Fig. 16d). Also, as a result of increased levee relief developed during Stages 1 and 2, the slope along the backside of the outer-bend levee was highest during the early part of Stage 3. The occurrence of slump deposits in the distal levee supports the existence of a steep slope along the backside of the levee (Hickson and Lowe, 2002; Hubbard et al., 2008). Therefore, it became increasingly common for overbanking flows to bypass the proximal

outer-bend levee and deposit on the distal levee. Deposition from these bypassing flows is recorded in DOB1 as F1 and F2 strata. Amalgamated F1 sandstone units represent deposition on the distal levee from successive large magnitude and sandy flows interpreted to have occurred due to both typical overbank-bypassing processes (i.e. overbank splays) and breaches in the levee (i.e. crevasse channel fill and splays). Furthermore, implied by these bypassing conditions is that the 17.5 m of levee thickness at the end of Stage 2 remained relatively constant throughout Stage 3 (i.e. negligible deposition on the proximal outer-bend levee). Deposition on the inner-bend levee during Stage 3, in contrast, is interpreted to have continued unabated because levee relief and the style of deposition remained insufficient to develop a steep slope along the backside of the levee and change transport and depositional conditions. Continued deposition on the inner-bend levee, which coincided with mainly bypassing flows on the proximal outer-bend levee, progressively reduced the difference in levee height along opposing channel banks. The thickness of the inner-bend levee near the end of Stage 3 is estimated by the total thickness of IB1, which is about 17 m. This value compared to the 17.5 m thick levee along the outer bend illustrates how drastically the difference in levee height along opposite sides of the channel decreased during Stage 3.

The interpretation of flows bypassing the outer-bend proximal levee is supported by the number of event beds that is recorded along opposite sides of the channel. For instance, IB1 consists of 367 event beds whereas only 159 occur in POB1 thus implying that more than half of the flows bypassed the proximal levee and deposited only on the distal levee. Moreover, it is interpreted that preferential deposition on the distal levee gradually reduced relief between the distal and proximal levees, and in doing so reduced the slope along the levee backside. Such negative feedback relationships have been reported along the backside of levees in modern fluvial systems (e.g. Cazanagli and Smith, 1998). In ICC3, preferential deposition on the distal levee eventually created a “new” near horizontal levee surface area.

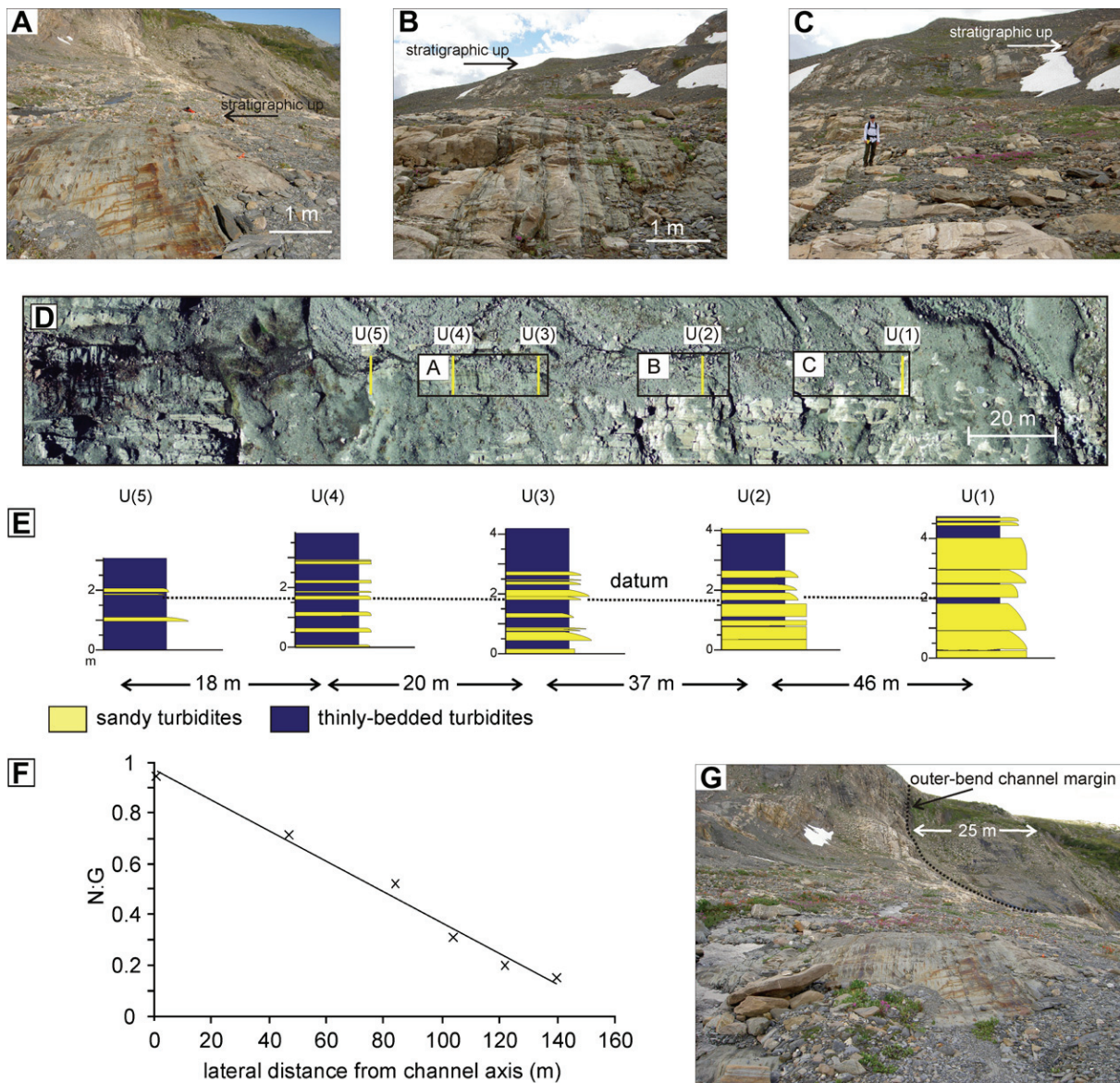


Fig. 15. Strata in the upper inner-bend levee (IB2). Photographs illustrate the lateral progression from amalgamated channel fill (A) into interbedded levee strata (B and C). D) Shows location of (A, B, and C) in addition to measured litho-stratigraphic logs in (E). F) Lateral trend in N:G from channel through inner-bend levee. G) Photograph from upper inner-bend levee towards outer-bend channel margin. Note levee strata in foreground and channel-fill deposits in background along the same stratigraphic horizon.

7.2. Development of Channel–Levee Unit 2 (CLU2)

During the latter stage of CLU1 it is interpreted that flows within the channel transitioned from bypassing to depositional, possibly related to allocyclic controls that temporally diminished flow magnitude and/or increased the ratio of sand. If correct then it follows that some time later allocyclic controls would again progressively increase the magnitude of throughgoing channel flows. As a consequence, flows within the channel that previously were depositional would again become non-depositional (bypass) to erosive. These higher magnitude flows reoccupied the once again weakly confined channel of ICC3 marking the onset of Stage 1 of CLU2 (Fig. 16e). In addition these energetic flows, which locally eroded up to 12 m deep, scoured the outer bend and caused it to migrate laterally about 50 m forming a “flat” along the channel margin. Overall, Stages 1–3 of CLU2 (Fig. 16e–g) are much alike that of CLU1 where throughgoing channel flows constantly eroded and

readjusted the channel-levee margin as levee relief increased locally forming surfaces that are inclined up to 25° toward the channel axis. The result is a “flat and step” (*sensu* Eschard et al., 2003) morphology as observed along the outer channel bank. CLU2 culminated in the generation of a “new” near horizontal proximal levee surface for the onset of the next channel-levee unit. Differences in levee heights that existed on opposite sides of the channel throughout the development of CLU2 could not be documented due to insufficient exposure on the inner bend.

7.3. Development of Channel–Levee Unit 3 (CLU3)

Stages 1 and 2 of CLU3 were similar to those in the underlying unit (CLU2) where early on throughgoing channel flows eroded up to 10 m deep and added another flat about 100 m wide to the “flat and step” morphology of the channel margin (Fig. 16h–i). The channel-filling stage of CLU3 (Stage 3), however, differs significantly

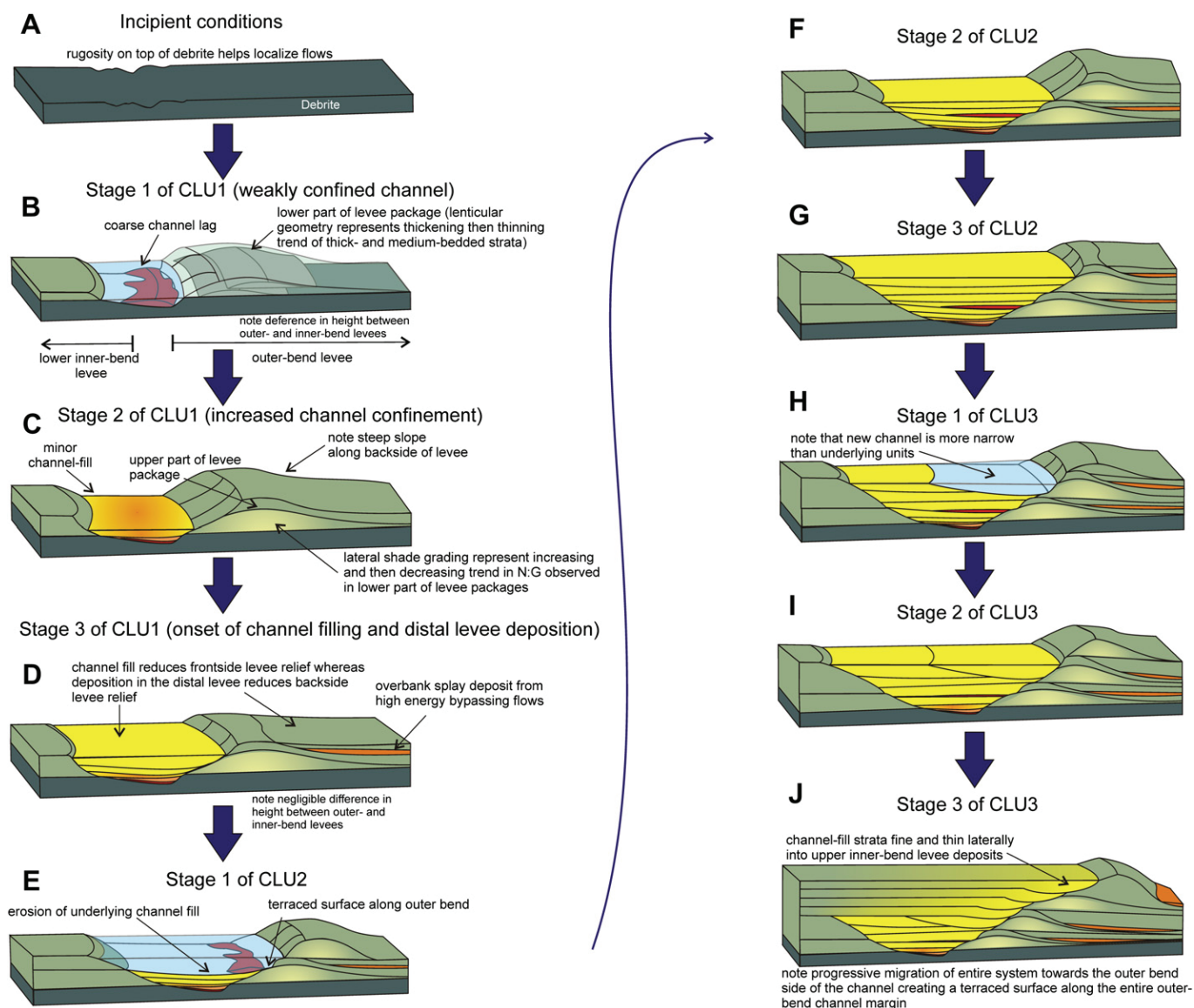


Fig. 16. Depositional model of ICC3 consists of three major episodes of channel-levee growth (CLU1 – CLU3), each of which are characterized by three similar growth stages (model not to scale). A) Rugosity along the top of this unit helped focus subsequent flows that ultimately localized and initiated ICC3. B) Stage 1 of CLU1 – low channel confinement and superelevation of turbidity currents caused preferential deposition of thicker-bedded and coarser-grained strata along the outer bend resulting in the outer-bend levee being twice the height of its inner-bend counterpart. At this time the channel was largely in a state of bypass. C) Stage 2 of CLU1 – increased channel confinement enabled only the upper, more dilute fraction of flows to overspill, resulting in the deposition of thinly-bedded turbidites over both inner- and outer-bend levees. D) Stage 3 of CLU1 – is marked by the onset of significant channel filling with flow bypass on the outer-bend proximal levee and preferential deposition on the outer-bend distal levee. In addition, continued deposition on the inner-bend levee reduced the height difference between the outer- and inner-bend levees. A consequence of significant channel filling was a reduction in the amount of levee relief enabling subsequent high energy flows to once again spill out of the channel and begin the development of CLU2 (E). Note that bypass flows during Stage 1 of CLU2 eroded underlying stratigraphy and also along the outer bend creating a terraced surface. Stage 2 (F) and Stage 3 (G) of CLU2, as well as Stage 1 (H) and Stage 2 (I) of CLU3, are similar to those of CLU1. Stage 3 (J) of CLU3, however, differs markedly from that of both CLU1 and CLU2. Here subdued topography along the inner bend allowed channelized flows to spill out and deposit on the inner-bend levee without becoming detached from the main channelized flow resulting in an “interfingering” pattern.

from previous channel-levee units. Here, channel-fill strata onlap the outer-bend channel margin whereas relief along the inner bend of the channel was reduced to the point where it allowed larger than average channel flows to expand laterally over the levee without becoming detached from the main channel flow. These flows deposited thick-bedded, coarse-grained sandstone turbidites that grade laterally into thinly-bedded F2 strata along the inner-bend channel margin and onto the levee (see Section 6.3.1) forming a distinctive interfingering pattern (Fig 16j).

The lowermost part of the inner-bend levee is not exposed and therefore a common datum could not be established to allow direct comparison of inner- versus outer-bend levee height in CLU3.

Nonetheless, the fact that channel-fill strata onlap the outer-bend margin but are interstratified along the inner-bend levee suggests that the outer-bend levee was at least somewhat higher than its inner-bend counterpart.

8. Discussion

8.1. Comparison with other deep-water channel systems

Although several studies provide criteria for recognizing levee deposits in the rock record (e.g. Kane and Hodgson, 2011; Campion et al., 2011), it is typically difficult to

unequivocally access the relationship between levee and channel-fill deposits in the field. This is well exemplified by the numerous interpretations for channel systems in the Cerro Toro Formation, which include: 1) channel-levee growth where fine-grained deposits are genetically linked with juxtaposed coarse-grained channel fills (Winn and Dott, 1979; Beaubouef, 2004; Hubbard et al., 2008), 2) cut and fill where channel flows erode into underlying, unrelated fine-grained deposits (Coleman, 2000), and 3) more complex evolution wherein the fine-grained interchannel and coarse-grained channel sediments were likely deposited concurrently by different flows or by the same flow that was decoupled upstream (Crane and Lowe, 2008). Similarly, alternative interpretations might fit the observations from ICC3. This next section will compare the interpretation presented in this paper concerning the stratigraphic evolution of ICC3 with other common channel-levee models.

8.1.1. Cut and fill

The erosive outer-bend channel margin of ICC3 superficially supports the cut and fill model (Coleman, 2000; Gardner and Borer, 2000), wherein flows responsible for finer-grained deposition are not associated with flows responsible for channel incision. However, in Castle Creek the terraced profile of one of the channel margins (outer-bend), the in-step lateral migration of channel fill and levee facies (Fig. 7b), and divergent paleocurrent directions on either side of the channel supports more strongly a genetic channel-levee relationship. Furthermore, on the other side of the channel (inner-bend), channel-fill strata in the upper part of ICC3 grade laterally into and interfinger with finer-grained levee deposits, clearly indicating their co-eval nature (Fig. 15).

8.1.2. Large-scale build and fill

The complete outer-bend levee of ICC3 has a thickness of about 80 m, making it comparable in thickness to large-scale channels such as those on the Amazon fan (Hiscott et al., 1997; Pirmez et al., 1997; Piper and Normark, 2001). Amazon fan levees commonly display an thinning- and fining-upward trend interpreted to be linked with increasing levee relief that causes flows to progressively overspill their more dilute upper part. The outer-bend proximal levee in ICC3, in contrast, is not characterized by a single large-scale thinning- or fining-upward trend but instead is subdivided into three distinct levee packages. Moreover, fine sand rarely overspills onto levees higher than 25 m on the Amazon fan (Hiscott et al., 1997). In contrast, thick-bedded, up to coarse sandstone beds are common as much as 40 m above the base of the levee in ICC3. This makes it highly unlikely that ICC3 represents a channel-levee system in which the levee built to appreciable height before channel filling began.

8.1.3. Multiple episodes of channel-levee growth

This model involves multiple episodes of levee growth followed by channel fill on the decameter scale (e.g. Beaubouef, 2004), and is supported here by the consistent stacking of ICC3 levee packages in a direction that mirrors the lateral migration of the channel fill (Fig. 7b). Theoretically, channel filling is succeeded by abandonment and deposition of fine-grained strata (e.g. Eschard et al., 2003; Samuel et al., 2003). The lack of fine-grained deposits in channel-fill strata of ICC3 may be because: 1) the channel system was never abandoned but instead alternated between periods of bypass and coarse sediment deposition; or 2) intense erosion between the late and early stages of two channel-levee units erased any evidence of these strata. The latter hypothesis is supported by the presence of deeply scoured surfaces (up to 12 m deep) commonly overlain by thick mudstone breccia layers at the base of individual channel fills.

8.2. Sinuosity and lateral migration

The interpreted sinuous nature of ICC3 is supported by the greater levee height recorded along one side of the channel, which typically occurs along the outer bend of sinuous channels (e.g. Deptuck et al., 2003; Posamentier, 2003). This does occur in straight channels if flows are significantly influenced by the Coriolis effect, however, in ICC3 the preferential emplacement of generally coarser grained sediment along one side of the channel suggest an association with the inertial effect of sediment-charged flows as they round a channel bend (Straub et al., 2008). Also, the step and flat geometry along the channel margin suggests a consistent directional shifting of the thalweg, which is a phenomenon commonly ascribed to meander loop expansion (sensu Posamentier, 2003). It is especially difficult to estimate how the channel bend evolved in time, however the consistent asymmetry in the height of the levees on opposite sides of the channel, in addition to their contrasting internal stratal make up and architecture, suggests that the channel bend did not migrate so far downstream that the straight part of the channel cropped out in the study area.

8.3. Levee shape

In the literature levee beds are commonly reported to exhibit a well developed lateral thinning trend, especially thicker-bedded deposits (DeVries et al., 1994; Beaubouef, 2004; Kane and Hodgson, *in this issue*). Thickening- then thinning-trends are less commonly reported, but have, for example, been described from levees deposits in the Rosario Formation, Mexico (Kane et al., 2007). It is commonly assumed that the vertical stacking of laterally thinning levee strata give rise to the typical “gull-wing” levee morphology (Kane et al., 2007, 2010b; Straub and Mohrig, 2008).

Based on the analysis of modern deep-marine levees, Skene et al. (2002) showed that levee thickness decayed exponentially away from the channel. Later Kane et al. (2007, 2010b) suggested that (paleo)geographical position was the primary control on levee shape, wherein slope channels show power law decay but exponential decay in more basinward systems. Based on theoretical arguments, Birman et al. (2009) proposed that levees showing power law profiles are associated with flows that have constant rates of fluid entrainment or sediment detrainment, whereas levees associated with flows that experience negligible entrainment thin exponentially laterally. Lateral trends in ICC3, however, show both exponential and power law decay, in addition to logarithmic and linear decay. For example, the thinning trend (excluding initial thickening) of individual medium- and thick-bedded (F1) strata in the proximal outer-bend levee of ICC3 are equally well described by linear ($R^2 = 0.96$) and logarithmic ($R^2 = 0.95$) decay profiles. This lateral change (reduction) in bed thickness is reflected in the lateral decrease in N:G, which is equally well described by logarithmic ($R^2 = 0.97$) and power law ($R^2 = 0.97$) decay functions. Moreover, the lateral trend in N:G of the lower inner-bend levee is best described by logarithmic decay ($R^2 = 0.99$), but a power law ($R^2 = 0.97$) relationship fits the data almost as well. In the upper inner-bend levee, on the other hand, N:G change is best described by a linear decay ($R^2 = 0.99$), only moderately by exponential decay ($R^2 = 0.92$), and poorly by logarithmic ($R^2 = 0.73$) and power law ($R^2 = 0.55$). Accordingly, it would appear that lateral trends in channel margin levees are not necessarily reliable indicators of (paleo)geographic location.

9. Conclusion

This paper documents the lithofacies and architecture of outer- and inner-bend levees associated with an up to 1.8 km wide by

80 m thick passive-margin, deep-water slope channel complex. The principal observations and interpretations are:

- ICC3 levees consist of two main lithofacies: 1) thin- to thick-bedded, coarse to fine sandstone T_{a-d} turbidites (F1); and 2) very thin- to thin-bedded, fine sandstone to siltstone T_{cd} turbidites (F2). F1 strata initially thicken (sometimes more than ten times their original thickness) over about 120 m and then thin laterally into F2 over the next 180 m away from the channel. Thereafter beds show little lateral change in thickness or grain size. F2 strata not associated with the thinning of F1 strata also show little change in grain size or bed thickness laterally.
- Along the outer-bend channel margin, levee and channel-fill strata are erosively juxtaposed. The proximal outer-bend levee is subdivided vertically into three stratal packages, which then are subdivided into lower (intercalated F1 and F2 strata) and upper (predominantly F2 strata) parts. The contact between the parts is clearly marked by an abrupt decrease in average bed thickness and sandstone content. N:G is up to 0.68 in the proximal levee and decreases over several hundreds of meters to about 0.35 in the distal levee, an area dominated by mudstone-rich F2 strata.
- The lower part of the channel-fill onlaps the predominantly fine-grained, inner-bend levee wherein N:G decreases over about 100 m from 0.4 to 0.15. Strata in the upper part of the channel, in contrast, grade laterally over 140 m into fine, thinly-bedded turbidites on the inner-bend levee. These strata become interbedded and are indistinguishable from other inner-bend levee strata. As a result, N:G decreases from 0.9, immediately adjacent where channel-fill strata begin to split, to 0.15 over 140 m laterally away from the channel. The upward change in channel margin geometry along the inner bend suggests that levee relief became more subdued allowing channelized flows to overspill onto the levee without becoming detached from the main channel flow.
- The outer-bend channel margin forms a composite terraced surface formed by three discrete episodes of channel inception followed by channel-levee growth. During each episode the channel thalweg migrated progressively toward the outer bend. Moreover, each episode of channel-levee growth began with a period of mostly levee deposition and channel bypass. Enhanced flow along the outer bend of the channel, related to inertial and buoyancy forces, enabled the outer-bend levee to become significantly higher and coarser grained compared to its inner-bend counterpart. In some cases the asymmetry in levee height was interpreted to be at least 2:1. This was then followed by significant channel filling that caused a reduction in levee relief. These conditions “reset” the channel system and allowed the next episode of channel-levee growth to repeat the pattern of channel-levee inception, growth and fill.

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Appendix B

Chapter six supplemental material

Method to calculate the theoretical sediment concentration from the rate of bed aggradation (using Run 3 as an example)

- Back out volume of sediment added to flume based on the rate of bed aggradation
 - Density of wet loose sand
 - 1.922 g/cm^3 (from www.simetric.co.uk)
 - This value assumes a porosity of about 0.44 (i.e. 0.56 solid material)
 - Width of flume is 61 cm
 - Aggradation rate of Run 3 is 0.0010 cm/sec
 - Solution
 - $\text{percent of solid material} \cdot (\text{width} \times \text{length} \times \text{aggradation rate}) = \text{volume of sediment added to flume}$
 - $0.56 \cdot (61 \text{ cm} \times 100 \text{ cm} \times 0.0010 \text{ cm/sec}) = 3.42 \text{ cm}^3/\text{sec}$
 - 100 cm is an arbitrary length chosen for this example
- Volume of sediment in suspension is estimated based on volume of sediment added to flume and the particle settling velocity
 - Settling velocity for 0.25 mm (diameter) sand is 2.97 cm/sec (by Heywood tables)
 - This value is from www.filtration-and-separation.com/settling/sediment, and assumes a particle density of 2.65 g/cm^3 , fluid density of 1 g/cm^3 , and a fluid viscosity of $0.0012 \text{ Pa} \cdot \text{sec}$
 - With a flow 18 cm deep (average flow depth of Run 3) a particle at the top of the flow would theoretically take 6.06 sec to reach the bed. Therefore, the estimated volume of sediment in suspension is
$$6.06 \text{ sec} \times 3.42 \text{ cm}^3/\text{sec} = 20.73 \text{ cm}^3$$
 - Volume percent of sediment in suspension over length of 100 cm and depth of 18 cm is

$$\begin{aligned} &\text{Volume of sediment/volume of water} \\ &20.73 \text{ cm}^3 / (61 \text{ cm} \times 100 \text{ cm} \times 18 \text{ cm}) = \mathbf{0.0002\%} \end{aligned}$$

- Note that this value is the same as the measured value (see Table 6.2)

How to create sediment peels

Following each experiment the flume was slowly drained and then allowed to naturally dry. After about 3-5 days the ripple cross-stratification and individual (ripple) set boundaries became more visible, but mostly in the upper part of the deposit (Fig. 1). Stratal detail in the lower part of the deposit remained poorly defined. If left to dry for a longer period of time, the upper part of the deposit became too dry, obscuring the cross-stratification. In addition, photographs taken through the flume's plexiglass walls obscured the fine-scale detail of the stratification. Attempts were then made to fabricate peels from the deposit, which could then be variously sectioned and easily photographed. Initial attempts to lithify the samples using a variety of synthetic epoxy hardeners proved unsuccessful. A simple and much less expensive procedure using diluted carpenter's glue had good results and is described next:

1. Drain flume and allow sediment to dry for 3 to 5 days.
2. Use a shovel to remove some sediment from center of flume and then use a thin metal sheet to create a smooth cross-section cut (see Figure 1).
3. Cut a thin 2 to 3 cm thick section of sediment using a thin metal sheet and sandwich the other side of the sediment with a stiff piece of cardboard (or plastic) and then carefully remove sediment from flume. It is recommended to cover stiff cardboard in a thin sheet of newspaper before removing sediment. Once the sediment is removed place on a table and carefully remove the cardboard from under the newspaper leaving sediment peel intact. To improve drying time it is useful to place peel on an apparatus like a cookie cooling tray. This will air to circulate around both the top and bottom of the peel.
4. Once removed, spray dilute carpenter's glue on sediment peel using a 1:10 glue to water mix. Use approximately 50 to 100 ml of dilute glue per 20 cm × 8 cm × 3 cm peel.
5. Let glue dry for about 3 to 5 days. To indurate the peel completely (e.g. for grain fabric analysis) apply the diluted glue mixture daily for several days.
6. Remove thin layer (e.g. 2 to 5 mm) from top of peel. A serrated hacksaw blade works well (see Figure 2).
7. Spray peel with dilute food colouring or ink (India ink works well) to highlight stratification (see Figure 3). Three to five drops of food colouring or ink per 100 to 200 ml of water is sufficient.



Fig. 1. Example of a cross-section created using a thin metal sheet.

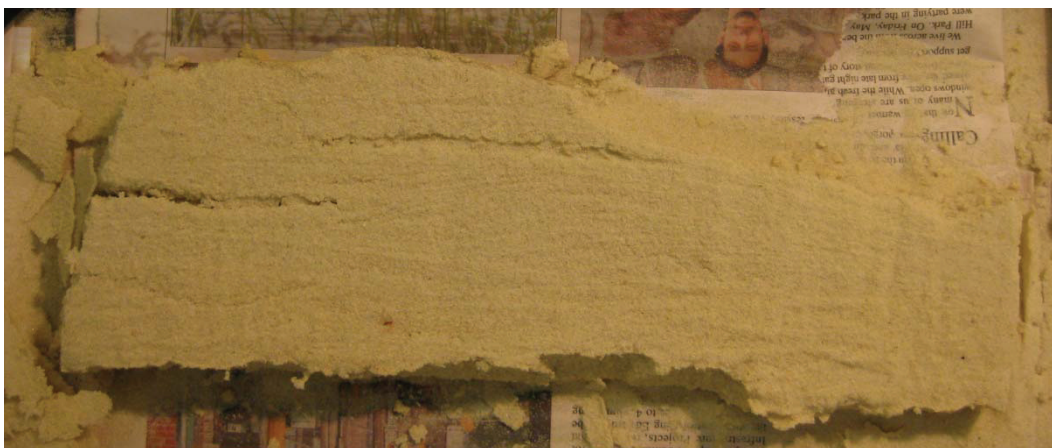


Fig. 2. Peel after removing thin layer from top using a serrated hacksaw blade.



Fig. 3. Peel after applying dilute India ink.

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The Journal of Sedimentary Research



May 27, 2011

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