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Differentiating submarine channel-related thin-bedded turbidite facies: Outcrop examples from the Rosario Formation, Mexico

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Abstract

Thin-bedded turbidites deposited by sediment gravity flows that spill from submarine channels often contain significant volumes of sand in laterally continuous beds. These can make up over 50% of the channel-belt fill volume, and can thus form commercially important hydrocarbon reservoirs. Thin-bedded turbidites can be deposited in environments that include levees and depositional terraces, which are distinguished on the basis of their external morphology and internal architecture. Levees have a distinctive wedge shaped morphology, thinning away from the channel, and confine both channels (internal levees) and channel-

belts (external levees). Terraces are flat-lying features that are elevated above the active channel within a broad channel-belt. Despite the ubiquity of terraces and levees in modern submarine channel systems, the recognition of these environments in outcrop and in the subsurface is challenging. In this outcrop study of the Upper Cretaceous Rosario Formation (Baja California, Mexico), lateral transects based on multiple logged sections of thin-bedded turbidites reveal systematic differences in sandstone layer thicknesses, sandstone proportion, palaeocurrents, sedimentary structures and ichnology between channel-belt and external levee thin-bedded turbidites. Depositional terrace deposits have a larger standard deviation in sandstone layer thicknesses than external levees because they are topographically lower, and experience a wider range of turbidity current sizes overspilling from different parts of the channel-belt. The thickness of sandstone layers within external levees decreases away from the channel-belt while those in depositional terraces are less laterally variable. Depositional terrace environments of the channel-belt are characterised by high bioturbation intensities, and contain distinctive trace fossil assemblages, often dominated by ichnofabrics of the echinoid trace fossil *Scolicia*. These assemblages contrast with the lower bioturbation intensities that are recorded from external levee environments where *Scolicia* is typically absent. Multiple blocks of external levee material are observed in the depositional terrace area where the proximal part of the external levee has collapsed into the channel-belt; their presence characterizes the channel-belt boundary zone. The development of recognition criteria for different types of channel-related thin-bedded turbidites is critical for the interpretation of sedimentary environments both at outcrop and in the subsurface, which can reduce uncertainty during hydrocarbon field appraisal and development.

Keywords: Thin-bedded turbidites; submarine channels; overbank; terraces; levees

1. Introduction

Submarine channels are major conduits for transfer of sediment from the continental shelf to the basin floor. They are commonly associated with thin-bedded turbidites deposited by overbank flow; thin-bedded turbidites can occur in sub-environments that include levees (internal and external; Kane and Hodgson, 2011) and depositional terraces (*sensu* Hansen et al., 2015, i.e., the portion of terraces that are due to overbank deposition from the channel). These two sub-environments are distinguished on the basis of their external morphology and internal architecture (Hansen et al., 2015). Depositional terraces and internal levees both occur in the channel-belt but depositional terraces form flat deposits while internal levees form wedge shaped deposits (Hansen et al., 2015). Submarine channels and their associated overbank environments have been studied using data from the modern seafloor such as shallow cores, side scan sonar, and swath bathymetry (e.g., Hübscher et al., 1997; Piper and Deptuck, 1997; Von Rad and Tahir, 1997; Deptuck et al., 2003; Migeon et al., 2006; Babonneau et al., 2010; Paull et al., 2013; Gamberi et al., 2015; Jobe et al., 2017), outcrops (e.g., Mutti, 1977; Beaubouef et al., 2007; Bouma et al., 2007; Cronin et al., 2007; Dykstra and Kneller, 2007; Kane et al., 2007; Di Celma et al., 2011; Hodgson et al., 2011; Kane and Hodgson, 2011; Khan and Arnott, 2011; Gamberi et al., 2013; Hubbard et al., 2014; Morris et al., 2014) and subsurface data (e.g., Leach et al., 1999; Clemenceau et al., 2000; Kendrick, 2000; Mayall et al., 2006; McHargue et al., 2010; Janocko et al., 2013; Hansen et al., 2017) which have all shown the architectural complexity of these sedimentary environments. Outcrop examples of ancient systems provide the opportunity to better study the vertical facies distribution within channel-levee systems but usually lack unequivocal time lines, full three-dimensional control and an unambiguous context.

The Upper Cretaceous Rosario Formation in Canyon San Fernando in Baja California, Mexico, provides an extensively studied and near complete cross-sectional exposure of a submarine slope channel-levee system (Morris and Busby-Spera, 1988, 1990; Dykstra and Kneller, 2007; Kane et al., 2007, 2009; Callow et al., 2013; McArthur et al., 2016). The outcrops offer the opportunity to study various thin-bedded turbidite environments both within the channel-belt and the external levee. It should be noted that thin-bedded turbidites can also occur in the margins of individual channels, which are not deposited by overbank processes but rather by the fine-grained tails of largely bypassing currents that deposited coarser grained sediment only in the most axial part of the channel (Hubbard et al., 2014; Li et al., 2016). The aim of this work is to identify criteria to differentiate successions of thin-bedded turbidites that are deposited within the channel-belt from those deposited as external levees but does not include thin-bedded turbidites that occur at the base or margins of individual channels. This work combines data from numerous sedimentary logs as well as geological mapping, and builds on the external levee dataset presented in Kane et al. (2007) as well as the ichnological and sedimentological study presented in Callow et al. (2013). This study also highlights the complexity of the boundary zone between the channel-belt and the external levee, which in this case is significantly modified by gravitational collapse. Studying the variety of thin-bedded turbidites within and adjacent to submarine channels can significantly improve our understanding of flow processes (bypassing vs depositional flows) and the evolution of the fill of channel systems. Together with characteristics that help distinguish different thin-bedded turbidites, this can reduce uncertainty in determining sand distribution and connectivity in hydrocarbon reservoirs.

2. Geological setting

The Rosario Formation comprises the upper part of a belt of Upper Cretaceous sedimentary rocks that make up the Peninsular Ranges fore-arc basin complex, which is essentially the continental slope of the active margin of the North American plate (Morris and Busby-Spera, 1990; Dykstra and Kneller, 2007). This basin complex is discontinuously exposed along the Pacific coast of southern California and Baja California for a distance of almost 500 km (Morris and Busby-Spera, 1990, and references therein) (Fig. 1A). The Rosario Formation unconformably overlies the continental El Gallo Formation and records a marine transgression from a shallow to deep-marine environment; in turn it is unconformably overlain by conglomerates of the non-marine to shallow marine Paleocene Sepultura Formation (Morris and Busby-Spera, 1990) (Fig. 1B).

The exposures in the area of Canyon San Fernando are about 35 km south of the town of El Rosario (Fig. 1A) and comprise a ~ 1000 m-thick succession of the upper (deep-water) Rosario Formation. Several coarse-grained mid-slope channel systems are exposed, including the ~ 6 km wide and ~ 400 m thick Canyon San Fernando channel-levee system, the geology and stratigraphy of which have been extensively documented and described elsewhere (Morris and Busby-Spera, 1990; Dykstra and Kneller, 2007; Kane et al., 2007, 2009; Dykstra and Kneller, 2009; Thompson, 2010). As in many modern slope channel systems, this channel system was oriented oblique to the continental slope due to structural control (Dykstra and Kneller, 2007; Thompson, 2010); the channel-belt is confined to the north-west by external levee deposits, but onlaps dominantly hemipelagic slope deposits to the south-east. The field area can thus be broadly subdivided into three sub-areas: the external levee, the channel-belt, and the slope (Fig. 2).

Four different channel-complex sets (*sensu* Sprague et al., 2002, 2005; Thompson, 2010), termed CCS-A to CCS-D (Figs. 2, 3) have been interpreted in outcrop, each of which is ~ 100 m thick and is postulated to represent a fourth-order sea-level cycle (Dykstra and Kneller, 2007). Overall the system is aggradational, indicating that sedimentation rates on the north-western levee were high enough to preserve confinement of the whole system (Dykstra and Kneller, 2007). The external levee outcrop is composed of concordant thin-bedded turbidites and has an exposure of over 90 m thick and a lateral extent of over 2 km. This outcrop formed the basis of the study of the external levee shown in Kane et al. (2007), who documented gross lateral thickness decay, change in sedimentary structures and sandstone bed thickness from the proximal to distal external levee.

Between the axis of the channel-belt and the external levee (Fig. 3), the system is dominated by thin-bedded turbidites of variable bed thickness; their relation to the channel bodies has previously been ambiguous. The exposure of these thin-bedded turbidites is not continuous but numerous outcrops exist in modern valleys that penetrate this area. A number of condensed sections are present that mark the tops of channel complex sets, and can be used to correlate from the margin of the channel-belt towards its axis. Detailed sedimentary logs throughout this area were placed in a stratigraphic framework by using these condensed intervals, which made it possible to study the differences in facies and depositional processes between thin-bedded turbidite outcrops within the channel-belt and the external levee outcrops.

3. Data and methods

Sixty-six sedimentary logs of thin-bedded turbidites from the channel-belt and the external levee are included in this study (Fig. 2). The data are a combination of logs collected

specifically for this study (Logs 3 to 56) as well as those presented in Kane et al. (2007) (Logs A to J from the external levee) and the dataset from Callow et al. (2013; logs denoted MBL). The focus was on collection of data on bed thickness, palaeocurrent directions, sedimentary structures and ichnofabric. Grain size of the beds was assessed visually in the field with the use of a grain size chart and hand lens.

Due to limitations of exposure the majority of the logs are less than 10 m long, with the exception of those collected for the study presented in Kane et al. (2007) from the external levee, which are up to 92 m long; these are correlated using tuff beds combined with photomosaic interpretation and were included to account for the impact of the vertical bed thickness changes commonly reported from external levees (Walker, 1985; Beaubouef, 2004; Kane et al., 2007; Kane and Hodgson, 2011). Due to the varying log lengths, error bars were included to indicate that the longer logs collected by Kane et al. (2007) are more indicative of the overall lateral trend in the external levee than the shorter logs. The error bars were calculated by dividing the sandstone proportion of the log by the square root of the number of sandstone layers present in the log. Consequently, a log with a higher number of sandstone layers has a smaller error bar.

Field mapping demonstrates that the logs within the channel-belt are distributed throughout CCS-A, CCS-B and CCS-D (Figs. 2, 3), enabling investigation of the stratigraphic variations in bed thickness within the channel-belt. The logs from the external levee and the channel-belt are located at various distances away from the channel-belt axis (Figs. 2, 3), making it possible to assess lateral variations in bed thickness with increasing distance away from the channel axis.

A turbidite bed is typically composed of a sandstone-mudstone couplet (referred to as an event bed) with each part referred to as a layer. Little hemipelagite has been observed in the thin-bedded turbidites, and due to the difficulty of accurately distinguishing turbidite mud from hemipelagite in the field, a simple sandstone and mudstone layer division was used for data collection. The bed thickness data of each sedimentary log were used to calculate the average sandstone proportion per log (taken here as the ratio of sandstone thickness to the total thickness of the measured interval), as well as the average and standard deviation of the sandstone and mudstone layer thickness.

4. Thin-bedded turbidites of the external levee

4.1. Observations

Twenty-two logs were collected from the external levee to supplement data presented in Kane et al. (2007) (Figs. 2-4), in order to clarify how the thin-bedded turbidites of the external levee differ from those within the channel-belt fill. The observations presented here are a summary of both datasets with the main focus on variations in bed thickness and sedimentary structures from channel-proximal to channel-distal areas over a lateral distance of 2 km.

There is a distinct decrease in sandstone proportion (from ~70% to 10%), average sandstone layer thickness (from ~6 cm to 1 cm) and standard deviation of sandstone layer thickness (from ~6 cm to 1 cm), from proximal to distal areas to the channel-belt (Fig. 4A, Table 1). The thin-bedded turbidites in the proximal areas are composed of sandstone and mudstone couplets with low thickness variations (standard deviation = 4.5 cm) (Figs. 5A, 6A) and are well exposed along the northern part of Mesa San Vicente (Fig. 2). The thin-bedded turbidites in the distal areas are much muddier, hence less well exposed, but sandstone layers still have

a low thickness variation (standard deviation = 1 cm) (Figs. 5B, 6A). Sandstone layers have abundant tractional sedimentary structures with parallel lamination often occurring at the base followed by ripple or climbing ripple cross-lamination. Sinusoidal lamination, if present, occurs towards the top of sandstone layers. Massive sandstone layers are present only in the most channel-proximal setting. The sandstones are either ungraded or more commonly normally-graded, and the transition between sandstone and mudstone is commonly abrupt. Occasionally the base of a sandstone layers exhibits erosion with minimal (cm scale) down-cutting, which is most common in channel-proximal areas. No amalgamation of sandstone beds was noted.

4.2. Interpretation

The thinning and decrease in sandstone content are key factors in the recognition of levees in general (DeVries and Lindholm, 1994; Kane et al., 2007; Kane and Hodgson, 2011; Nakajima and Kneller, 2013; Morris et al., 2014; Hansen et al., 2015). The average and standard deviation of sandstone layer thickness and the sandstone proportion all decrease according to a power law for the longer levee logs included in the study by Kane et al. (2007; see also Birman et al., 2009), with which the shorter logs collected for this study are consistent (Fig. 4A). Similar lateral changes in external levee properties are also observed in other outcrops such as the Fort Brown Formation, Karoo Basin, South Africa (Kane and Hodgson, 2011; Morris et al., 2014). As the flow over spills the levee crest it becomes unconfined and rapidly decelerates (Hiscott et al., 1997); this reduces the competence of the flow, leading to sediment deposition (Kane et al., 2010a). Particles with higher settling velocities, such as sand, are deposited first as the decelerating current spreads over the levee, whilst clay remains in suspension until the flow is almost at rest (McCave, 1975; Lowe, 1988; Hiscott, 1994; Kneller, 1995; Kneller and McCaffrey, 2003). Hence, the sand fraction

of a bed generally decreases in thickness away from the channel more rapidly than finer sediment, and the clay and silt fraction of the bed may actually thicken away from the channel (Kane et al., 2007).

Massive sandstones and traction dominated sedimentary structures such as ripples, climbing ripples and sinusoidal laminations are common in the proximal external levee, while in the distal external levee only starved ripples occur, which indicates a scarcity of sand available to build bedforms. Changes in the relative abundance of massive sandstones and starved ripples indicate that suspended load fall-out decreases distally across the external levee (Kane et al., 2007) because much of the suspended load is deposited during the initial overspill across the levee crest, while little sand is transported in suspension to the more distal parts of the levee. The absence of grading observed between the sandstone and mudstone layers (refer to sedimentary logs in Fig. 6A) of a bed may either be due to absent grain sizes in the overspilling currents, or due to a period of sustained velocity of the flow following deposition of the sand fraction, allowing bypass of the missing grain size fraction (Kneller and McCaffrey, 2003; Kane et al., 2007).

5. Thin-bedded turbidites of the channel-belt

Based on outcrop and seismic studies, thin-bedded turbidites are estimated to make up over 50% of the Canyon San Fernando channel-belt fill but their characteristics vary between different areas of the channel-belt. Logs were collected from two areas (A and B) to investigate the depositional architecture, bed thicknesses and sedimentary structures at different positions within the channel-belt (Figs. 2, 3). Area A comprises thirty logs from the western part of the channel-belt, where thin-bedded turbidites are exposed in a number of modern valleys incised into the flanks of Mesa San Vicente (Fig. 2). They were collected

from CCS-A and CCS-B at progressively greater distances from the channel axis (Fig. 3). In Area B towards the north-eastern part of the channel-belt four additional logs were collected from below the condensed section of CCS-D (Fig. 3), which was used to correlate the logs. Logs in Area B are located between 400 m to 3355 m away from the interpreted position of the CCS-D channel-belt axis. We describe the log observations and discuss their interpretations separately for the two areas.

5.1. Western part of channel-belt fill (Area A)

5.1.1. Observations

The thin-bedded turbidites exposed in Area A of the channel-belt show a very large variation in sandstone layer thickness (standard deviation = 11 cm) with dominantly thin sandstone (0.5-20 cm) and mudstone (1-30 cm) layers interrupted by very thick sandstone layers (0.5-1 m) (Figs. 4B, 6B, 7A, Table 1). The thinner sandstone layers are commonly ripple cross-laminated or have a basal parallel laminated part and an upper sinusoidal laminated part (Fig. 7B, C). The thicker sandstone layers are usually massive with occasional flame structures at the base (Fig. 7B). The sandstone proportion shows minimal lateral variation with distance away from the channel-belt axis with an average proportion of around 49%, while a slight increase in average sandstone layer thickness is observed (from ~4 to 7 cm) (Fig. 4B).

It is not possible to correlate the logs from the western part of the channel-belt due to the limitations of exposure and lack of laterally continuous marker beds. However, field mapping and correlation of condensed sections allows us to assign the logs to a particular channel complex set, and relative stratigraphic position within the channel complex set (Fig. 3), allowing us to investigate vertical variation. This shows that there is no apparent vertical variation in sandstone proportion or average sandstone layer thickness in the thin-bedded

turbidites of a channel-complex set, and that the properties of the logs in CCS-A are similar to the properties of the logs in CCS-B (Fig. 8).

Turbidite bed bases are generally sharp with erosion most commonly observed beneath the coarsest sandstone layers (very coarse sandstone to pebble grade), and may cut down <50 cm into the underlying stratigraphy (Fig. 9). Sandstone layers are most often normally graded or otherwise ungraded. The tops of the beds are mostly sharp with only a few showing progressive normal grading from sandstone to mudstone. Some amalgamation of the thickest turbidite beds is also noted.

5.1.2. Interpretations

There is no indication of a decrease in sandstone proportion or average sandstone layer thickness away from the channel-belt axis (Fig. 4B), suggesting that the thin-bedded turbidites in this part of the channel-belt did not form internal levees (*sensu* Kane and Hodgson, 2011; Hansen et al., 2015). While internal levees do not appear to be present in Area A of the channel-belt, this does not rule out their presence elsewhere in the Canyon San Fernando slope channel system, as thin-bedded turbidites are abundant throughout the channel-belt fill.

Both the average and standard deviation of sandstone layer thickness, as well as the average grain size of the sandstone layers, are higher within the channel-belt than on the external levee (Table 1). This suggests that these thin-bedded turbidites were deposited at relatively small elevations above the active channel, thus capturing the coarser grained suspended load carried in the lower part of the current (Hansen et al., 2015, and references therein), which is confined to the channel-belt. The higher standard deviation of sandstone layer thickness is

interpreted to result from a greater range of turbidity current sizes depositing sediments in these areas of the channel-belt, while the height of the external levee crest limits the turbidity current size that is able to overspill. For these reasons we interpret these thin-bedded turbidites as depositional terrace deposits (*sensu* Hansen et al., 2015). Depositional terraces have been identified on the modern seafloor where they form flat bench-like areas adjacent to the channels within the channel-belt. Good examples of terraces are seen in the La Jolla Canyon in California (Paull et al., 2013), the Congo submarine fan (Babonneau et al., 2010), the Mahin, Avon, and Benin major channels offshore Nigeria (Deptuck et al., 2003; Hansen et al., 2017), and the X and Y channel from the western Niger Delta slope (Jobe et al., 2017). In outcrop they have been described from the Cambro-Ordovician Cap Enragé Formation in Quebec, Canada (Hein and Walker, 1982) but are rarely identified in the ancient record. This may partly be due to exposure limitations, making the study from the Canyon San Fernando an important addition to previously published examples.

Tractional sedimentary structures indicate bedload transport during bed aggradation, as is also observed in the proximal external levee (Kane et al., 2007). Sinusoidal lamination represents aggrading bedforms with high rates of suspension fallout from a waning current (Allen, 1963; Hunter, 1977; Jobe et al., 2012); this appears to be much more common in sandstone layers from the depositional terrace than in those from the external levee. This may be due to the formation of these structures beneath the lower parts of turbidity currents where the concentration of sediment is greater, resulting in higher suspended load fallout rates (e.g., Lowe, 1982; Kneller and Branney, 1995). The erosional contacts at the base of sandstone beds also indicate interaction by the lowermost and highest energy parts of the flow with the bed (e.g., Xu, 2010, and references therein), where many of the currents have sufficient energy to erode.

Sandstone layers with significant erosional (estimated less than 50 cm) lower contacts (Fig. 9) are interpreted as chute channels (Gamberi, 2001; Hansen et al., 2017) that formed in the depositional terrace area due to erosion by energetic overspilling currents. Coarser material (up to gravel sized) was preferentially transported within these chute channels as they are topographically lower than the surrounding terrace surface. Chute channels in overbank regions have been reported from outcrop in the Cambro-Ordovician Cap Enragé Formation in Quebec, Canada (Hein and Walker, 1982) as well as on the modern seafloor in the Tyrrhenian Sea in Italy (Gamberi and Marani, 2011; Hansen, 2016).

5.2. North eastern part of channel-belt fill (Area B)

5.2.1. Observations

Logs of thin-bedded turbidites from Area B are located above and adjacent to the channel fills of CCS-D and below the condensed section that caps the whole system (Fig. 2). The logs (logs 52 to 54, Figs. 2, 3) represent the same stratigraphic interval and allow investigation of spatial changes in thin-bedded turbidite properties. This interval is composed of thin sandstones (average = 2 cm) and thick mudstones (average = 13 cm). Beds become progressively thinner and finer with increasing height (Fig. 6C). The key characteristics of deposits from Area B are summarized in Table 1. From the channel-proximal to channel-distal areas (laterally) of the channel-belt axis of CCS-D the average sandstone layer thickness initially increases from 1.7 to 3.1 cm but then decreases slightly to 1.7 cm again (Fig. 4C). There is little variation of sandstone layer thickness with a standard deviation of 4 cm, which is similar to the average standard deviation of sandstone layer thickness in the proximal external levee (4.5 cm) (Table 1). There is also little lateral variation in sandstone

proportion, which remains between 14-18% from proximal to distal areas with respect to the CCS-D channel-belt axis (Fig. 4C).

5.2.2. Interpretations

The average sandstone layer thickness for each log increases then decreases from proximal to distal areas over almost 3 km (Fig. 4C), which may suggest that these are internal levee deposits, with the wedge shape. CCS-D is the uppermost exposed channel complex set of the Canyon San Fernando channel-levee system in which coarse-grained channel fills are exposed. The thin-bedded turbidites stratigraphically overlie the coarser grained channel fills of CCS-D, and are unlikely to have been deposited by overspill from the CCS-D channels unless the channels were laterally migrating. They may also be related to overspill from a small aggradational channel whose fine grained fill is either not exposed or has not been differentiated, either as internal levee or as deposits that are transitional between levees and depositional terraces (Hansen et al., 2015). Internal levees are most commonly described from the aggradational phase towards the top of channel-levee systems (Samuel et al., 2003; Mayall et al., 2006; Kane and Hodgson, 2011; Hansen et al., 2015; Hodgson et al., 2016) in which channel fills are typically narrow and often fine grained.

Alternatively, turbidity current activity may have declined progressively as the system shut down, generating mostly small turbidites separated by thick hemipelagic and pelagic mudstone intervals (Fig. 6C). This may suggest that these thin-bedded turbidites were deposited as the initial stage of an abandonment drape. Abandonment can occur due to channel avulsion farther upstream, relative sea-level rise, or decreased sediment supply due to changes in climate or tectonic activity in the source area (Clark and Pickering, 1996). These deposits appear towards the final stage of the channel-belt fill, interpreted to be associated

with a third order transgression and highstand, and thin upwards, culminating in a condensed section above logs 52-54. Since late-stage aggradational channels are typically associated with the waning stages of channel system evolution system (Winker, 1996; Kneller, 2003), these two interpretations are not mutually exclusive.

6. Quantitative bed thickness analysis

6.1. Observations

Quantitative analysis of bed and layer thickness distribution was undertaken to distinguish thin-bedded turbidites of the external levee from thin-bedded turbidites within the channel-belt fill. A map-based statistical analysis using the average recorded log properties at a given log site was employed. These maps were gridded in map view in ArcGIS® using a natural neighbour interpolation method which is part of the Spatial Analyst® tool. The natural neighbour interpolation method (Sibson, 1981) performs well for irregularly distributed data as it finds the closest subset of input samples to an interpolation point and applies a weighting to them based on the proportionate areas using Voronoi diagrams to interpolate a value (Gold, 1989; Sambridge et al., 1995; Ledoux and Gold, 2005).

Plotting the average properties of each log in map view shows that deposits from Area A in the channel-belt and those from the proximal external levee area have the highest proportion of sandstone (40-70%) and sandstone layer thickness (4-14 cm), while Area B and the distal external levee have much lower values (< 20% and < 4 cm respectively) (Fig. 10A, B). The standard deviation of sandstone layer thickness within Area A (average = 11 cm) is significantly greater than in Area B (average = 4 cm) and in the external levee (proximal external levee = 4.5 cm, distal external levee = 1 cm) (Table 1, Fig. 10C). There is also an

increase in the average sandstone layer thickness, sandstone proportion and standard deviation of sandstone layer thickness around log MBL4 (Fig. 10A-C).

6.2. Interpretations

The higher sandstone proportion and average sandstone layer thickness in the proximal external levee and depositional terrace area are attributed to the proximity of these areas to the active channel at the time of deposition. However, the most noticeable contrast can be seen in the standard deviation of sandstone layer thickness between thin-bedded turbidites of the depositional terrace and the external levee (Figs. 6A, B, 10C); this is attributed to the relationship between levee height, and the thickness and vertical structure of turbidity currents in the channel. Levee height, combined with the vertical grain-size distribution within the turbidity current, determines the grain size range and amount of sand that can overspill onto the levee. Smaller currents that are able to overspill from the channel but not overtop the external levee will remain confined within the channel-belt to form depositional terraces across the width of the channel-belt (Fig. 11). This also results in coarser grain sizes being deposited onto the terraces where the lower and coarser-grained parts of the turbidity current are able to deposit. In contrast, only the top parts of larger currents, which generally have a smaller grain size, are able to overspill onto the external levee crest, hence limiting the grain size range of the sandstone layers on the external levee (Hansen et al., 2015). The greater abundance and coarser grain-size of turbidites on areas that are less elevated than the levee has also been reported from the Congo turbidite channel (Babonneau et al., 2010).

Due to levee growth over time, a thinning upwards trend is common in levees (Morris et al., 2014), but Kane et al. (2007) reported an overall sandstone thickening upwards trend within the proximal part of the external levee (Fig. 10B). They implied that either the magnitude and

frequency of flows increased over time, or that the aggradation of the channel-belt floor and/or the collapse of levee material decreased the height of the levee, enabling an increasing amount of sand to overspill onto the external levee. While the amount of sediment overspill increased vertically, there is no apparent corresponding change in the standard deviation of the sandstone layer thickness (Fig. 10C), which means that the levee height is still acting as a filter that allows only currents of a certain size and vertical distribution of sediment to overspill.

7. Deformed thin-bedded turbidites

7.1. Observations

Towards the western edge of the channel-belt, several zones of deformation are observed. Deformation includes some localised folding of thin-bedded turbidite successions but is dominated by large (< 100 m across) rotated blocks of coherently bedded strata. Beds within the blocks can be inclined as steep as 80° compared with the shallower dip (<10-20°) observed for the non-deformed beds in the field area (Fig. 12). The poles of the strike and dip data from the external levee blocks indicates that they have either collapsed downslope or towards the external levee as they rotate away from the inner external levee margin (Fig. 12). The bedding within these blocks has noticeably different bed thickness statistics from the surrounding *in situ* deposits. The blocks have a lower standard deviation of sandstone layer thickness (standard deviation = 3.6 cm) than the surrounding *in situ* deposits (standard deviation = 11 cm) but have similar values to the proximal external levee deposits (standard deviation = 4.5 cm) (Fig. 10C, Table 1). The zone where these blocks are present is here defined as the channel-belt boundary zone (CBBZ).

7.2. Interpretations

The position of the rotated slide blocks within the distal part of the depositional terrace area, together with the similarity of bed thickness statistics of the thin-bedded turbidites of these blocks to the external levee deposits, suggests that they originated from the proximal regions of the external levee and have collapsed into the channel-belt. Channel-belt boundaries are highly dynamic areas where sediment is likely to collapse into the channel-belt due to steep gradients along the inner side of the levee (Skene et al., 2002; Hansen et al., 2015; Bain and Hubbard, 2016). Moreover levee crests can migrate towards or away from channel-belts over time (Kane et al., 2007), making it difficult to identify a unique boundary between the channel-belt and the external levee. Collapse of the proximal part of levees has been reported from the subsurface in the Gulf of Mexico (Clemenceau et al., 2000) and the Nile Delta (Samuel et al., 2003), and shallow seismic data from the Indus upper fan (Deptuck et al., 2003), Mississippi Canyon area (Sawyer et al., 2007, 2014) and offshore Nigeria (Hansen et al., 2017), where collapsed levee blocks can be well imaged. Even though a preserved levee crest is often visible in shallow seismic data, such as from the Indus slope (Deptuck et al., 2003), this crest has most likely been modified significantly over time when the channel was active. These observations illustrate the complexity of the CBBZ, which can be several hundred metres wide (Deptuck et al., 2003).

8. Ichnology

8.1. Observations

Trace fossils are abundant in the Rosario Formation in Canyon San Fernando and have been used to generate an ichnofabric-based model for palaeoenvironmental interpretation of submarine channel-levee systems (Callow et al., 2013).

The sedimentary logs of the thin-bedded turbidites reveal that the irregular echinoid trace fossil *Scolicia* (horizontal to oblique burrows with concave, meniscate, backfill structures 2-5 cm in diameter (Callow et al., 2013) (Fig. 13A, B) is most abundant in parts of the channel-belt where the standard deviation of the sandstone layer thickness is high (Figs. 6B, 14). Within the CBBZ, *Scolicia* is not always present and specifically is consistently absent from the external levee blocks (Fig. 14). *Scolicia* usually occurs at the junction between fine-grained sandstone and the overlying siltstone or mudstone layer and is often associated with high bioturbation intensity ichnofabrics (Callow et al., 2013).

The trace fossils *Phycosiphon* and *Nereites* are the dominant component of the ichnofabric in logged sections where the standard deviation of the sandstone layer thickness is relatively low (i.e., the external levee environment and not the channel-belt, Fig. 14). The trace fossil *Phycosiphon* is used to refer to the distinctive ‘frogspawn ichnofabric’ recognized by the presence of dark-coloured, fine-grained burrow cores together with an associated coarser grained, quartz-rich halo that is 1-5 mm in diameter (Callow et al., 2013) (Fig. 13C, D). *Nereites* refers to the larger, 0.5-1.0 cm diameter, burrows in transverse cross-section with sand-rich halos surrounding a clay-rich core (Callow et al., 2013) (Fig. 13C, D). These traces are often found in the mudstone caps to turbiditic sandstones and generally do not result in high bioturbation intensities.

8.2. Interpretations

The correspondence between ichnofabric and standard deviation of sandstone layer thickness helps to differentiate the depositional environments of depositional terrace (*Scolicia* ichnofabric) and external levee (*Nereites* and *Phycosiphon* ichnofabric) in this system. Similar patterns of trace fossil and ichnofabric distribution have been reported from a number

of other submarine channel systems from a variety of ages and settings (e.g., Shulz and Hubbard, 2005; Heard and Pickering, 2008; Monaco et al., 2010; Cummings and Hodgson, 2011; Phillips et al., 2011). Such similar patterns of distribution are interpreted as evidence that a combination of physical and chemical factors have the dominant control on the spatial distribution of trace making organisms in deepwater channel and fan systems (Callow et al., 2014). The presence of *Scolicia* in the depositional terrace deposits is here interpreted to be a consequence of increased supply of oxygenated water and organic matter by turbidity currents in the channel-belt compared to the external levee (Callow et al., 2014). Rare examples of *Phycosiphon* are the only discernible trace in the uppermost thin-bedded unit of the channel-belt in Area B (Fig. 6C). If oxygen and organic matter are transported into the deep sea by turbidity currents then it can be interpreted that turbidity current activity diminished during the abandonment phase of the channel system, making it a less suitable environment for trace-making infaunal echinoids. *Phycosiphon* and *Nereites* are common throughout various other depositional environments of Canyon San Fernando (Callow et al., 2013) as their trace makers appear to have broader environmental tolerances than those of the *Scolicia* organism. The absence of *Scolicia* in the external levee blocks indicates that these blocks have collapsed into the channel-belt from the external levee and are not *in situ* deposits. The variable presence of *Scolicia* in the CBBZ suggests that living conditions in this zone were not always ideal for *Scolicia* trace-makers to colonize; initially (CCS-A and CCS-B) *in situ* beds in the CBBZ are dominated by *Scolicia* while the *Phycosiphon/Nereites* assemblage is confined to collapsed blocks, whereas at higher stratigraphic levels the *Phycosiphon/Nereites* ichnofabric is ubiquitous in the CBBZ, though this region still retains a distinctive ichnological signature (Fig. 14).

9. Palaeocurrent data

9.1. Observations

Palaeocurrent data from Kane et al. (2007) were supplemented with new data wherever possible. The scarcity of traction structures in the distal external levee and in Area B in the channel-belt meant that few measurements were possible in these areas. In the proximal external levee and Area A of the channel-belt, palaeocurrent data were collected from 3D current ripples or sole marks on thicker sandstone layers. In Area A of the channel-belt the mean palaeoflow (213.4° , $n = 109$) is similar to the mean palaeoflow documented in the channel fill (198°) (Morris and Busby-Spera, 1990; Dykstra and Kneller, 2007) (Fig. 14). The proximal external levee shows a wide divergence of palaeocurrents (147° – 217°) both convergent to and divergent from the channel-belt. Towards the mid and distal levee palaeoflow is generally sub-parallel to the channel-belt (180° – 209°) but there is significant scatter of palaeocurrents directions. In summary, palaeocurrents from the external levee indicate a relative lack of confinement, whilst palaeocurrents from the channel-belt are more aligned and suggestive of confinement in the channel-belt.

9.2. Interpretations

Palaeocurrents of the depositional terrace deposits that are sub-parallel to or slightly divergent from the mean channel axis may suggest that the flows were still largely confined within the channel-belt, and in the overbank region flowed parallel to the external confinement of the channel-belt. Evidence of reflections of the currents from the inner external levee confinement is limited, suggesting that the currents were sufficiently large to occupy the entire width of the channel-belt (Hansen et al., 2015; Southern et al., 2015; Kane et al., 2010b) (Fig. 11A). External levees are characterized by palaeocurrent directions that exhibit greater scatter/divergence (Fig. 14). This may indicate the presence of local topographic irregularities, or that the overspilling currents flowed more directly down the

levee backslope (Kane et al., 2010b), resulting in the divergence of overspilling currents away from the regional maximum slope gradient (Kane et al., 2007). However, the lack of a more consistent trend in the mid to distal levee may also in part be due to the limited number of measurements ($n = 68$). Palaeocurrent indicators using grain-fabric data from the external levee have shown that flows went from strongly divergent to more unidirectional flows in the later stages of overspill (Kane et al., 2010b). The levees may also have been affected by other types of currents such as contour currents or internal waves and tides that are known to rework sediments in many continental slope environments (Stow and Lovell, 1979; Rebesco and Camerlenghi, 2008; Stow et al., 2013; Callow et al., 2014; Rebesco et al., 2014).

10. Discussion – Distinguishing thin-bedded turbidite environments

By combining statistical, sedimentary, and ichnological characteristics of the various thin-bedded turbidites in the Canyon San Fernando slope channel system, it is possible to infer their depositional environments and establish criteria that help distinguish them. These distinguishing characteristics consequently allow us to build a model for the evolution of the Canyon San Fernando system.

Sandstone layer thickness is an important criterion for differentiating between thin-bedded turbidite environments. Cross-plotting the standard deviation of sandstone layer thickness against sandstone proportion for each log (Fig. 15) reveals distinct fields for the depositional terrace, proximal external levee (including external levee blocks), distal external levee, and abandonment or internal levee deposits. There is a continuum of the data points of the external levee, with the higher value point-cloud indicating the proximal external levee and the lower value point cloud indicating the distal external levee (Fig. 15). The rotated external levee blocks plot within the proximal levee point-cloud, clearly indicating their difference

from the surrounding depositional terrace deposits, as well as their likely source from the proximal external levee. Whilst there is overlap between the proximal external levee and depositional terrace regions, an upper threshold limit can be defined for the expected standard deviation of external levee sandstone layer thickness, where values above this threshold imply a depositional terrace interpretation (Fig. 15). In regions of overlapping sandstone proportion/bed thickness variation, ichnological observations and the presence or absence of collapsed external levee blocks aid the depositional environment interpretation.

Only a limited size of currents is able to overspill and deposit on the external levee. This implies that more sandstone should be deposited in the depositional terrace within the channel-belt than on the external levee. However, the sandstone proportion between the proximal external levee and depositional terrace environments is quite similar. This may be due to more energetic bypassing currents eroding significant amounts of sediment from within the channel and depositional terrace area whilst depositing sediment on the external levee at the same time. Increased erosion in the depositional terrace area is indicated by the presence of amalgamated sandstone beds and the presence of chute channels, both of which are absent on the external levee. Also, due to the thickening upward trend observed in the proximal part of the external levee (Kane et al., 2007) the exposed sections may represent an anomalously sandy and higher part of the levee succession, which would be less prevalent in levees that exhibit a thinning upwards trend (e.g., Fort Brown Formation levees in South Africa (Morris et al., 2014)), where a levee section at an equivalent height above the channel would be much less sandy. Lastly, due to the proximity of the depositional terrace to the channel fills comprising the channel complex sets, some of thicker sandstone layers may represent channel margin deposits, as seen in the Cretaceous Tres Pasos Formation in Chile

(Hubbard et al., 2014). However, mapping suggests that Area A is at least 500 m to the NW of the closest channel fill deposits.

As mentioned previously, and confirmed by the cross-plot in Figure 15, the properties of the deposits of the uppermost unit of thin-bedded turbidites (Area B), below the condensed section that caps the San Fernando system, are not unambiguously indicative of a specific depositional environment. However they plot in or close to the distal external levee field.

Based on the data presented here a model for the evolution of the Canyon San Fernando channel system can be established (Fig. 16). The initial incision into the slope by the channel system was guided by a structural lineament resulting in the system running oblique to the palaeoslope (Dykstra and Kneller, 2007). The presence of slide blocks in the lower part (CCS-A and CCS-B) of the Canyon San Fernando channel system (Figs. 3, 16) indicates that the majority of the inner external levee collapse occurred early in the development of the channel system (Fig. 16 B-E). This perhaps indicates that, due to early rapid aggradation of the external levee, the relief was greatest early in the development of the channel system and/or the channels being more erosional in character at this stage (Deptuck et al., 2003; Hodgson et al., 2016). Collapse of the inner external levee during the early evolution of a channel system has also been reported from the Benin-major channel-levee system (Deptuck et al., 2003, 2007) and the Ursa channel-levee system in the Gulf of Mexico (Sawyer et al., 2014). The different channel complex sets are interpreted to migrate SE up the palaeoslope as the channel-belt becomes increasingly confined by the external levee to the NW. This also increases the width of the depositional terrace area towards the NW. As the external levee aggrades, and proximal levee blocks collapse into the channel-belt, the levee crest migrates towards or away from the channel-belt over time (Kane et al., 2007) (Fig. 16 – levee crest

positions 1 to 3), creating the CBBZ. The magnitude and frequency of the currents decreases over time as sea level rises and sediment supply is decreased (Clark and Pickering, 1996; Hansen et al., 2015; Hodgson et al., 2016), resulting in the abandonment of the system, which is marked by a regional condensed section.

11. Conclusions

During the evolution of the Canyon San Fernando channel system, large quantities of thin-bedded turbidites were deposited both within the channel-belt and on the external levee. We differentiate three different thin-bedded turbidite facies that are interpreted as: external levee, depositional terrace, and potential internal levee/abandonment deposits, with the latter two occurring within the channel-belt. Depositional terraces have a much higher standard deviation of sandstone layer thickness than external levee and internal levee/abandonment deposits, but in some cases can also be differentiated based on the presence of a distinctive *Scolicia*-dominated trace fossil assemblage. External levee deposits show very regular sandstone layer thicknesses, which is attributed to the height of the external levee crest acting as a filter for the sizes of turbidity currents able to overspill. Internal levee/abandonment deposits are very mud prone with generally very thin sandstone layers that do not show large variation in sandstone layer thickness.

Within the channel-belt a zone dominated by sediment deformation (named the CBBZ) was interpreted where a number of large external levee blocks are shown to have slid into the channel-belt, probably associated with the over-steepening of the inner levee. These can be distinguished based on their anomalous bedding dips, the general absence of characteristic terrace trace fossils and the low standard deviation of sandstone layer thickness compared to the surrounding depositional terrace deposits.

These detailed observations better constrain the interpretations of thin-bedded turbidites in the Canyon San Fernando system, and potentially provide sedimentological, stratigraphic and ichnological criteria for their recognition and interpretation in other submarine channel systems. The properties of the different thin-bedded turbidites presented here can be applied to the description of hydrocarbon reservoirs in the subsurface, where sparsity of well data and the limitations of seismic resolution necessitate the use of analogue-derived models.

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Figure 1. (A) Map showing the location of the Canyon San Fernando channel-levee system on the Pacific coast of Baja California, Mexico. (B) Stratigraphy of the Rosario embayment of the Peninsular Ranges forearc basin complex showing the context of the lower and upper members of the Rosario Formation (modified from Morris and Busby-Spera, 1990; Dykstra and Kneller, 2007, 2009; Kane et al., 2007).

Figure 2. Geological map of the Canyon San Fernando field area. The map shows the location of the channel-belt, external levee and intervening channel-belt boundary zone (CBBZ) as well as the distribution of various channel-complex sets (CCS) inside the channel-belt. CCS boundaries were locally impacted by faulting. The red circles indicate locations of all logs used for the data analysis.

Figure 3. Schematic cross-section of Canyon San Fernando slope channel system with lithologies and the four channel complex sets (CCS-A to D) indicated in the channel-belt. The depositional terrace areas (Area A), internal levee/ abandonment deposits (Area B) and external levee with indicative locations of the various logs included in this study are shown (modified from Thompson, 2010).

Figure 4. Lateral variation of sandstone layer attributes (total sandstone proportion, average sandstone layer thickness, standard deviation of sandstone layer thickness) with increasing distance from the channel-belt axis in (A) external levee deposits (longer logs included in the study by Kane et al. (2007) are indicated in red and the shorter logs included for the purpose

of this study are indicated in black), (B) channel-belt thin-bedded turbidites in depositional terrace deposits / Area A, and (C) abandonment or internal levee deposits / Area B.

Figure 5. Distinctive lithofacies from the external levee of Canyon San Fernando. (A) Proximal external levee: Regular thin-bedded (<10 cm) medium- to fine-grained turbidite sandstone alternating with mudstone (average = 7 cm). (B) Distal external levee: Regular very thinly bedded (<5 cm) fine- to very fine-grained turbidite sandstone alternating with thick mudstone beds (average = 10 cm). Orange layers are the sandstone layers.

Figure 6. Example sedimentary logs from (A) the proximal and distal external levee. (B) The channel-belt thin-bedded turbidites in Area A. (C) The channel-belt thin-bedded turbidites in Area B.

Figure 7. Distinctive lithofacies from the depositional terrace deposits of Canyon San Fernando. (A) Deposits from Area A in the channel-belt with thin (1 cm to 30 cm) turbidite sandstone alternating with mudstone, interrupted by very thick sandstone layers (0.5 to 1 m). Geological hammer in the white circle for scale. (B) Variation in sandstone layer thicknesses seen in the depositional terrace deposits in Area A. Mudstone intervals contain abundant thin ripple-cross-laminated sandstone and relatively thicker sandstone layers with basal parallel laminations overlain by sinusoidal lamination. Flame structures occur at the base of thicker sandstones. (C) Ripple cross-lamination and sinusoidal lamination in sandstone layer typical for Area A.

Figure 8. Vertical changes in (A) sandstone layer thickness and (B) sandstone proportion of logs in the channel-belt in Area A. The stratigraphic position of logs is based on their location on the geological map shown in Figure 2.

Figure 9. Log MBL4 (PH = *Phycosiphon*, Ne = *Nereites*, SC = *Scolicia*, vsfb = vertical sand filled burrows (unidentified)) from the interpreted chute channel in Area A within the channel-belt.

Figure 10. Quantitative bed thickness analysis. (A) Interpolated sandstone proportion map using available log data. (B) Interpolated average sandstone layer thickness map using available log data. (C) Interpolated map of the standard deviation of the sandstone layer thickness within each log.

Figure 11. Schematic diagram indicating how (A) smaller turbidity currents remain completely confined within the channel-belt, while (B) larger turbidity currents are able to overspill sediment onto the external levee.

Figure 12. Map indicating the distribution of strike and dip data in the CBBZ and the depositional terrace area within the channel-belt. The red symbols are measurements taken from the external levee blocks (blocks shown in green). The black symbols are measurements taken from the surrounding in situ deposits.

Figure 13. (A) *Scolicia* (S) ichnofabric of the depositional terrace environment. (B) *Scolicia* ichnofabric found on the bedding planes of thicker turbiditic sandstone layers in depositional

terrace deposits. (C-D) *Phycosiphon* and *Nereites* ichnofabric within the siltstone and mudstone layers of the external levee environment. Large horizontal *Nereites* (N) with dark, mud-rich, lenticular core and lighter concentric halo. Varying sizes of *Phycosiphon* (Ph) with some forming a hook shape with a dark, mud-rich, faecal core and lighter, sand-rich halos.

Figure 14. Palaeocurrent data for all logs. For the external levee palaeocurrent directions are shown for multiple locations while the palaeocurrents for the depositional terrace in Area are summed onto a single rose diagram. The logs are colour coded based on *Scolicia* or *Phycosiphon/Nereites* presence. The map is an interpolated map of the standard deviation of the sandstone layer thickness within each log.

Figure 15. Cross-plot of the sandstone proportion and standard deviation of sandstone layer thickness for all logs.

Figure 16. Schematic representation of the deposition of sediment in the Canyon San Fernando channel-levee system during CCS-A to CCS-D. Note the collapse of external levee blocks into the channel-belt during CCS-A and CCS-B deposition. (A) Early incision oblique to the regional slope due to a weakness created by underlying fault. (B) Early aggradation of channel fill in CCS-A. Oversteepening of inner external levee and external levee block sliding into channel-belt. (C) Early aggradation of depositional terrace on external levee block. Continued aggradation of early channel fill in CCS-A. (D) Aggradation of depositional terrace during deposition of later channel fills in CCS-A. (E) Erosion and deposition of early channel fill of CCS-B into previous deposits and deposition of later channel fill and their associated depositional terraces. Further oversteepening of inner external levee and external

levee block sliding into channel-belt. (F) Erosion and deposition of early channel fill of CCS-C into previous deposits and deposition of later channel fill and their associated depositional terraces. (G) Deposition of CCS-D channel fill and abandonment deposit or possible internal levee.

Table captions

Table 1. Sandstone layer thickness, standard deviation of sandstone layer thickness, average sandstone proportion, grain size, sedimentary structures, bed contacts and ichnology of the different thin-bedded turbidite depositional environments described from the Canyon San Fernando outcrops.

Depositional environment		Average, maximum and StDev of sandstone layer thickness	Sandstone layer grain size	Average sandstone proportion	Sedimentary structures	Bed contacts	Ichnology
External levee	Proximal (up to 3.5 km from the channel axis)	Average = 5.5 cm Max. = 20 cm StDev = 4.5 cm	Medium to very fine grained	50%	Sandstones exhibit abundant tractional sedimentary structures: parallel lamination, ripple cross-lamination and climbing ripple cross-lamination. Sandstones typically exhibit a vertical transition from parallel lamination to ripple cross-lamination. Sinusoidal lamination, when present, occurs at the top of sandstones. Massive sandstone layers are present only in most channel-proximal setting.	Sharp sandstone bases with abrupt transition between sandstone and mudstone layers. Occasionally in channel-proximal areas the base of a sandstone layer exhibits erosion with minimal down-cutting. Amalgamation of beds not observed.	<i>Phycosiphon</i> and <i>Nereites</i> dominant
	Distal (up to 4.5 km from the channel axis)	Average = 1.5 cm Max. = 5 cm StDev = 1 cm	Very fine grained	15%	Sedimentary structures are much less common and/or visible. Starved ripples are most common with some ripple cross-lamination noted.	Sandstone bases are sharp and the transition between sandstone and mudstone layers is abrupt.	<i>Phycosiphon</i> and <i>Nereites</i> dominant
Channel-belt	Area A (Depositional terrace)	Average = 5cm Range = 0.5-20 cm separated by thicker-bedded turbidites up to 1 m thick StDev = 11 cm	Fine sandstone to pebble grade	49%	Sandstone layers have abundant tractional structures such as parallel lamination, current ripple cross-lamination, climbing ripples, and starved ripples, but sinusoidal lamination is the most common. Occasional bed amalgamation.	Sandstone bases are generally sharp. Progressive normal grading is present from sandstone to mudstone. Erosional bed bases may occur below bases of the coarsest sandstone layers (very coarse to pebble grade sand) and cut down < 50 cm into the underlying stratigraphy.	<i>Scolicia</i> dominant
	Channel-belt boundary zone	Average = 4 cm StDev = 5 cm	Medium to very fine grained	44%			<i>Scolicia</i> dominant
	Area B (Abandonment or internal levee deposit)	Average = 2 cm Range = 0.5-33 cm StDev = 4 cm	Medium to very fine grained	15%	Sedimentary structures are rare in the thin sandstone layers but, if visible, parallel laminations are the most common. No bed amalgamation.	Sandstone tops are very sharp. Irregular erosional basal contacts are absent.	<i>Phycosiphon</i> and <i>Nereites</i> dominant

Table 1