FLUVIAL FAN ARCHITECTURE, FACIES, AND INTERACTION WITH LAKE: LESSONS LEARNED FROM THE SUNNYSIDE DELTA INTERVAL OF THE GREEN RIVER FORMATION, UINTA BASIN, UTAH

by

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ABSTRACT

The Early Eocene Sunnyside Delta interval of the middle Green River Formation in the Uinta Basin changes from fluvial channel and floodplain deposits to interbedded fluvial, deltaic, and lacustrine deposits across 20 km in the Nine Mile Canyon. Outcrop measured sections and photomosaics with GPS survey of excellent cliff face exposures are integrated with areal mapping of channel dimensions, channel to floodplain ratio, and sedimentary facies variability. This study identifies the Sunnyside Delta interval as a fluvial fan system, composed of variable discharge river and floodplain deposits that basinward transitions into deltaic, mixed and lake facies. Variable discharge signatures include the abundant Froude supercritical flow and high deposition rate sedimentary structures, in-channel mudstones, in-channel bioturbation and desiccation, and the low abundance of cross stratification. This work identifies upstream and downstream bar-scale (macroform) accretion styles that were not previously recognized, as variable discharge river macroforms are commonly referred to as poorly developed or even non-existent. This work identifies low angle downstream dipping accretion, steep upstream accretion, and vertical aggradation as some of the characteristic accretion styles. The fluvial fan stratigraphy is characterized by multiple scales of upward sandying and thickening successions with upward increasing channel to floodplain ratio, channel size, the degree of channel amalgamation, and the proportion of floodplain splay sandstones. The smallest scale upward sandying and thickening successions is recognized as avulsion packages that form the building blocks of the stratigraphy. Larger scale successions are likely to indicate lobe and fan progradation. The common avulsions also determines how the fan interacts with Lake Uinta, as this work documents lateral transitions between channel and floodplain, deltaic and lake facies across just a few hundred meters to a few kilometers. Furthermore, some mouth bar deposits consist of alternating carbonate grainstones and siliciclastic sandstones, and some abandoned channels are filled with dolomitic mudstones.

iii
All these transitions indicate a highly irregular shoreline, where fluvial and deltaic deposits build out locally at the active channel locations, laminated dolomitic mudstones accumulate in protected embayments or abandoned channels, and lime grainstones where lake’s wave and current energy is high. We interpret these fluvial-lacustrine interactions in Sunnyside Delta interval as a result of river avulsions and contemporaneous carbonate productions.
TABLE OF CONTENTS

ABSTRACT ........................................................................................................ iii

LIST OF FIGURES ................................................................................................. ix

LIST OF TABLES ..................................................................................................... xii

ACKNOWLEDGEMENTS ....................................................................................... xiii

DEDICATION .......................................................................................................... xv

CHAPTER 1 INTRODUCTION ............................................................................... 1

1.1 Motivation ........................................................................................................ 1

1.2 Dissertation Organization ............................................................................... 3

1.3 Reference Cited ............................................................................................. 4

CHAPTER 2 VARIABLE DISCHARGE RIVER MACROFORMS IN SUNNYSIDE
DELTA INTERVAL OF THE GREEN RIVER FORMATION,
UINTA BASIN, UTAH ....................................................................................... 6

2.1 Abstract .......................................................................................................... 6

2.2 Introduction ..................................................................................................... 7

2.3 Geological Setting and Background ............................................................ 8

2.4 Dataset and Methods ................................................................................... 9

2.5 Facies Groups .............................................................................................. 10

2.5.1 Facies Group 1: Sandstones and Conglomerates with
Long-wavelength, Low-angle Laminations with Foresets and
Backsets, and Planar Laminations .................................................................... 10

2.5.2 Facies Group 2: Sandstones with Steep Foresets .................................... 14

2.6 Channel Marcoforms .................................................................................. 15
2.6.1 Large Vertically and Laterally Amalgamated Channel Lithosomes ...... 15
2.6.2 Large Channel Lithosomes with Upstream Migrating Macroforms ...... 17
2.6.3 Channel Lithosomes with Low Angle Downstream Accretion Sets....... 19
2.6.4 Amalgamated Channel Lithosomes with High Angle Accretion Sets ..... 20
2.6.5 Heterolithic Aggradational Channel Lithosomes ............................... 21
2.6.6 Small Lenticular Channel Lithosomes ............................................. 23
2.7 Discussion .......................................................................................... 24
2.7.1 Lateral Accretion Sets................................................................. 24
2.7.2 Downstream Accretion ................................................................. 25
2.7.3 Upstream Accretion and the Hierarchy of Froude Supercritical Flow Deposits ................................................................. 26
2.8 Conclusions ...................................................................................... 27
2.9 Reference Cited .................................................................................. 44

CHAPTER 3 VERTICAL AND LATERAL FACIES VARIATIONS IN FLUVIAL
FANS: EARLY EOCENE GREEN RIVER FORMATION, UINTA
BASIN ............................................................................................... 55

3.1 Abstract ............................................................................................ 55
3.2 Introduction ....................................................................................... 56
3.3 Geological setting and background .................................................. 57
3.4 Methods and dataset ......................................................................... 58
3.5 Facies Associations in Fluvial Channel-floodplain Lithosomes .......... 59
  3.5.1 Facies Association 1: Large Vertically and Laterally Amalgamated Sandy Channel and Floodplain Lithosomes ............................... 59
  3.5.2 Facies Association 2: Laterally Amalgamated Sandy to Heterolithic Channel Accretion Set and Floodplain Lithosomes .................. 61
4.6.2.1 Facies Association 2.1: Lime Grainstone ................. 109
4.6.2.2 Facies Association 2.2: Stromatolites .................... 109
4.6.2.3 Facies Association 2.3: Gastropod Wackestone .......... 110
4.6.3 Deltaic Facies Association ...................................... 110
4.6.3.1 Facies Association 3.1: Sharp-based Tabular Sandstone ... 110
4.6.3.2 Facies Association 3.2: Interbedded Siliciclastic and Carbonate Mudstone and Sandstone .................. 111
4.6.4 Fluvial-lacustrine Mixed Facies Associations ............... 112
4.6.4.1 Facies Association 4.1: Mixed Siliciclastic-Carbonate Accretion Sets ............................................. 112
4.6.4.2 Facies Association 4.2: Sigmoidal Sandy Grainstone Macroforms ......................................................... 113
4.6.4.3 Facies Association 4.3: Mixed Siliciclastic-carbonate Lenticular Mudstone ............................................ 114
4.6.4.4 Facies Association 4.4: Laminated Dolomitic Mudstone .... 115
4.7 Lateral and Vertical Trends ........................................... 115
4.8 Discussion on Fluvial-lake Interaction .......................... 116
4.9 Conclusions ........................................................... 117
4.10 Reference Cited ....................................................... 142

CHAPTER 5 CONCLUSIONS ............................................... 150
LIST OF FIGURES

| Figure 2.1 | Stratigraphic column and base map of Uinta Basin and Nine Mile canyon. | 28 |
| Figure 2.2 | Representative examples of measured sections organized by channel lithosome types. | 29 |
| Figure 2.3 | Example outcrop photos and measured sections of Facies group 1: sandstones and conglomerates interpreted as Froude supercritical deposits. | 30 |
| Figure 2.4 | Example outcrop photos and measured sections of Facies group 2: sandstones with steep foresets interpreted as Froude subcritical flow deposits. | 31 |
| Figure 2.5 | Example photos of in-channel mudstones and bioturbation. | 32 |
| Figure 2.6 | Large vertically and laterally amalgamated channel lithosomes with low-angle accretion sets. | 33 |
| Figure 2.7 | Large channel lithosomes with upstream migrating macroforms. | 34 |
| Figure 2.8 | Channel lithosomes with low-angle accretion sets. | 35 |
| Figure 2.9 | Channel lithosomes with high-angle (up to 20°) accretion sets with preserved mudstone between sand bodies. | 36 |
| Figure 2.10 | Heterolithic aggradational channel lithosomes. | 37 |
| Figure 2.11 | Small lenticular channel lithosomes. | 38 |
| Figure 2.12 | Summary diagram of the 6 channel lithosome types in flow parallel (to the left, arrow points downstream) and perpendicular views (right). | 39 |
| Figure 3.1 | Stratigraphic column and base map of Uinta Basin and Nine Mile canyon. | 73 |
| Figure 3.2 | Outcrop examples of facies association 1. | 74 |
| Figure 3.3 | Photos of common sedimentary structures in facies association 1. | 75 |
| Figure 3.4 | Outcrop examples of facies association 2. | 76 |
Figure 3.5 Photos of common facies in facies associations 2 and 3. ...................... 77
Figure 3.6 Outcrop examples of facies association 3. ................................. 78
Figure 3.7 Photos of common facies in facies association 3. ....................... 79
Figure 3.8 Examples of characteristic vertical trends. Photos (A, B) and a measured section (C) of upward thickening and sandying trends in floodplain deposits with capping channel lithosomes. ................... 80
Figure 3.9 Vertical changes in channel (yellow) vs. floodplain (grey) proportions at each examined location (by mile marker). ................................. 82
Figure 3.10 Photos showing thin floodplain-rich fluvial intervals along Road 191 outcrops. ................................................................. 83
Figure 3.11 Example photomosaics that illustrate lateral changes from most proximal to most distal locations, respectively (A through G). See Fig. 3.1 for mile marker locations. ................................. 84
Figure 3.12 Google Earth image showing Ili River that originates from the Tianshan Mountains and builds a large fluvial fan on the southern margin of the Balkhash Lake. ............................... 88
Figure 4.1 Stratigraphic column and base map of Uinta Basin and Nine Mile canyon. ................................................................. 118
Figure 4.2 Example outcrop photos of fluvial facies association 1. .................. 119
Figure 4.3 Outcrop photos of lime grainstone (FA2.1). ............................ 122
Figure 4.4 Outcrop photos of stromatolite (FA2.2) and gastropod wackestone (FA2.3) facies associations. ........................................ 123
Figure 4.5 Outcrop photos of the deltaic facies association 3. ....................... 124
Figure 4.6 Vertical trends of facies associations. Two styles are presented in A and B with a typical measured section and an outcrop photo. .............. 127
Figure 4.7 Outcrop photos of the mixed siliciclastic-carbonate accretion sets (FA4.1). ................................................................. 128
Figure 4.8 Outcrop photos of sedimentary structures in oolitic sandstone (F26) in the mixed siliciclastic-carbonate accretion sets (FA4.1). .................. 129
<table>
<thead>
<tr>
<th>Figure</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>4.9</td>
<td>Outcrop photos of sigmoidal sandy grainstone macroform (FA4.2).</td>
<td>130</td>
</tr>
<tr>
<td>4.10</td>
<td>Outcrop photos of mixed siliciclastic-carbonate lenticular mudstone (FA4.3).</td>
<td>131</td>
</tr>
<tr>
<td>4.11</td>
<td>Outcrop photos of laminated dolomitic mudstone (FA4.4).</td>
<td>132</td>
</tr>
<tr>
<td>4.12</td>
<td>Depositional trends across the Nine Mile Canyon.</td>
<td>133</td>
</tr>
<tr>
<td>4.13</td>
<td>Outcrop examples showing lateral and vertical transitions.</td>
<td>134</td>
</tr>
<tr>
<td>4.14</td>
<td>Examples of lake-fluvial interaction in Ile River, Lake Balkhash.</td>
<td>135</td>
</tr>
</tbody>
</table>
LIST OF TABLES

Table 2.1 Description and Interpretation of Sedimentary Facies ......................... 40
Table 3.1 Description and Interpretation of Sedimentary Facies ......................... 89
Table 3.2 Description, Interpretation, and Comparison of Facies Associations .......... 93
Table 4.1 Description and Interpretation of Sedimentary Facies .......................... 136
Table 4.2 Description and Interpretation of Sedimentary Facies Association .......... 142
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CHAPTER 1
INTRODUCTION

1.1 Motivation

Large fluvial fans are significant unconfined depositional systems that are in need of comprehensive facies and stratigraphic models and a better understanding of the system dynamics, as well as a predictive reservoir model. There are currently multiple fluvial fan models among which only the fluvial megafan model (e.g. Singh et al., 1993; Shukla et al., 2001; Chakraborty and Gosh, 2010) consider avulsions rather than bifurcations as the key mechanism for fan formation. All other fan models, such as the distributive fluvial systems (DFS) (Weissmann et al., 2010, 2013), terminal fans (Kelly and Olsen, 1993), and fluvial distributary systems (Nichols and Fisher, 2007). Interestingly, the facies and stratigraphic model recently proposed for the DFS model (Davidson et al., 2013) is the same as the fluvial megafans model proposed already by Singh et al. (1993). Furthermore, occurrence of fluvial megafans has been linked to variable discharge rivers, and the control of variable precipitation, such as today in the monsoon zone and subtropics (Leier et al., 2005). In contrast DFS have been interpreted to occur in any climate types (Hartley et al., 2010).

The Sunnyside Delta interval of the middle Green River Formation in the Uinta Basin was chosen for this study due to its laterally continuous exposure across ca 20 km in the Nine Mile Canyon. The interval has been identified as a possible fluvial fan systems, and deposits of variable discharge rivers, and linked to global warming during the Early Eocene Climate Optimum (EECO) (Plink-Björklund and Birgenheier, 2012, 2013; Plink-Björklund et al., 2014; Plink-Björklund, 2015).

Outcrops of the Sunnyside Delta Interval has been previously interpreted as a terminal fan (Pusca, 2003) or deltaic deposits (Keighley et al., 2003; Schomacker et al., 2010; Moore et al., 2012; Burton et al., 2014). This discrepancy elevates our interests to study what
causes the differences in interpretations.

The fluvial megafan model was firstly well illustrated by Shukla et al. (2001) using Ganga Megafan. A systematic channel style change from braided, to anastomosing, to meandering channels due to a decreasing discharge from proximal to distal part of the fan were observed (Singh et al., 1993; Shukla et al., 2001). It is the most well developed model and has numerous modern and ancient examples. Numerous studies in early Eocene fluvial outcrops (Schmitz and Pujalte, 2007; Storey et al., 2009; Foreman et al., 2012; Plink-Björklund and Birgenheier, 2012) propose that Paleocene Eocene Thermal Maximum (PETM) promotes a dramatic change in water and sediment discharge and promotes formation of fluvial fans. Fluvial megafans in Himalayan and central Andean foreland basins began to develop as the Asian monsoon and seasonal precipitation started respectively (Leier et al., 2005). Highly seasonal discharge triggers frequent avulsion and channel migration at times, thus the lateral instability promotes fluvial fan formation (Leier et al., 2005).

This study shows the differences between channel and floodplain deposits and deltaic mouth-bar deposits in the Sunnyside Delta Interval, and suggests that the depositional intervals were in places incorrectly identified due to the unfamiliar combinations of sedimentary structures and depositional architecture, not represented in current fluvial facies models. The study confirms the Sunnyside Interval as a fluvial fans deposit, and further documents the facies as well as bar-scale architecture of variable discharge rivers. The study also documents that a considerable lateral variability in fluvial and lacustrine deposits occurs where the fluvial fan enters the lake, as a result of frequent river avulsions characteristic for fluvial fans. Considerable lateral variability also occurs in siliciclastic, mixed and carbonate facies, as the river avulsions and the resultant fluvial sediment input to the lake interact with contemporaneous carbonate production.
1.2 Dissertation Organization

The dissertation is organized into five chapters: introduction to the dissertation, three papers to be submitted/submitted to peer reviewed journals, and general conclusions. Each paper provides introduction, geological background, methodology and data, results, and conclusions on its own. 1st paper (Chapter 2) has been submitted to Sedimentology, 2nd paper (Chapter 3) has been submitted to Basin Research, and 3rd paper (Chapter 4) will be submitted to Journal of Sedimentary Research in 2018. The figures, tables, and references cited in each paper are included in the end part of each paper. A brief summary of three papers and their contributions is provided below.

Chapter 2 (1st paper) recognizes variable discharge signatures in fluvial channel lithosomes and identifies 6 types of macroforms varying from low angle dipping downstream accretion to steep obliquely lateral or upstream accretion, and vertical aggradation. This paper presents a first systematic description of such macroforms that have hitherto been referred to as poorly developed (Plink-Björklund, 2015) or even missing (Fielding et al., 2018) in variable discharge rivers.

Chapter 3 (2nd paper) identifies a fluvial megafan system characterized by an avulsion dominated stratigraphy by documenting the lateral extent, internal architecture, and lateral and vertical facies transitions of channel-floodplain deposits. This paper compares the results with a modern analogue and with the existing fluvial fan models such as megafans, distributive fluvial systems, fluvial distributary systems, and terminal fan fluvial systems.

Chapter 4 (3rd paper) investigates the fluvial-lacustrine interaction and the carbonate-siliciclastic mixing process by documenting the complex fluvial-lacustrine facies transition styles and lateral changes in mixed siliciclastic-carbonate facies. These transitions are interpreted as a result of river avulsions and contemporaneous carbonate productions at a highly irregular lake shoreline. The results caution the interpretation of transgressive-regressive cycles that link vertical transitions between fluvial and lacustrine deposits to lake-level changes.
1.3 Reference Cited


2.1 Abstract

The channel lithosomes of the Sunnyside Delta Interval of the Green River Formation in the Uinta Basin, Utah, USA display an abundance of Froude supercritical flow sedimentary structures and high deposition rate sedimentary structures, and a low abundance of dune cross strata, in-channel mud layers, and in-channel bioturbation and desiccation. Such abundance or even dominance of Froude supercritical flow as well as high deposition rate sedimentary structures has been recognized as a signature of variable discharge rivers, where sediment transport and deposition is related to high magnitude floods, and base flow discharge is characteristically low and commonly below the sediment movement threshold. While this link between the abundance of Froude supercritical flow sedimentary structures and variable river discharge is relatively well established, the bar-scale channel architecture is essentially unknown, and is commonly described as poorly developed, or even absent. Using excellent cliff face exposures, this study identifies a variety of bar-scale (macroform) accretion styles that vary from low-angle downstream accretion to steep upstream accretion, and vertical aggradation in six different types of channel lithosomes. The channel lithosomes
vary in size and lithology and range from sandstones to interbedded sandstones and mudstones. The macroforms are linked with sedimentary facies that vary from abundant Froude supercritical flow deposits to climbing ripple laminated deposits and in-channel mudstones.

2.2 Introduction

River discharge variability, as linked to a combination of seasonal and inter-annual variability, is suggested to be a first-order control on river morphodynamics (e.g., Fielding et al., 2009, 2017; Plink-Björklund, 2015; Nicholas et al., 2016). As a result, variable discharge river deposits differ from established fluvial facies models in terms of their meso- and macro-scale bedforms (see e.g., Fielding et al., 2009; 2017; Plink-Björklund, 2015; Nicholas et al., 2016). On the mesoform scale, a variable discharge river facies model is emerging that consists of an abundance of Froude supercritical and transcritical flow sedimentary structures and a reduced proportion of cross strata (e.g. Fielding et al., 2009, 2017; Plink-Björklund, 2015 and references therein). Considerably less data is available regarding the macroform (bar) scale, and bars are considered to be significantly modified (Nicholas et al., 2016; Ghinassi et al., 2017), poorly developed (see Plink-Björklund, 2015 and references therein) or even absent (Fielding et al., 2018).

This paper documents outcrop examples of barforms of variable discharge river deposits of the Eocene Green River Formation in Uinta Basin, Utah, US. These deposits are recognized as variable discharge rivers based on the abundance of Froude supercritical flow and high deposition rate sedimentary structures, in-channel mud deposits, flood event beds, in-channel bioturbation, and mud-clast conglomerates (see Plink-Björklund, 2015 and references therein). A variety of barforms are documented that are distinct from the models for braid bars (Ashley, 1990; Lunt et al., 2004; Lunt and Bridge, 2004; Sambrook Smith et al., 2009) and point bars (Allen, 1965, 1970; Ethridge and Schumm, 1978; Tyler and Ethridge, 1983; Miall, 1985, 1988; Labrecque et al., 2011; Durkin et al., 2015), and provide documentation on their facies. These barforms range from low-angle tabular sets of flood-event
beds, to steep obliquely accreting upward coarsening sets, to large-scale backsets, to vertical aggradation sets. This paper thus show that the macroforms are not absent, but rather differ from the established facies models, and have thus not been readily recognized.

2.3 Geological Setting and Background

The Uinta basin is an intra-foreland basin formed to the east of the Sevier thrust front (Fig. 2.1) in response to the Laramide uplift of the Uinta Mountains (Dickinson et al., 1988; DeCelles, 2004). The basin is asymmetrical with a gentle southern margin and a steep northern margin. The early to middle Eocene Green River Formation consists of interbedded fluvial, deltaic and lacustrine deposits (Ryder et al., 1976) that are 0.6 to 1.8 km thick (Morgan and Bereskin, 2003). A large fluvial-deltaic system on the southeastern basin margin was fed by the California Paleoriver, a ca 750 km long river that drained from the Mojave region, rather than from local Laramide uplifts (Davis et al., 2010; Dickinson et al., 2012). Flowing northward, the far-traveled California paleoriver is interpreted to form a large fluvial fan system, including the Sunnyside Delta Interval (Plink-Björklund and Birgenheier, 2013). Previously documented paleo-flow directions in Colton/Wasatch and Green River Formations indicate a general northward direction (Fouch et al., 1976; Dickinson et al., 1986; Remy, 1992; Schomacker et al., 2010; Ford et al., 2016; Gall et al., 2017; Jones, 2017). Two carbonate successions, the Uteland Butte Limestone and the Carbonate Marker Unit are the first wide-spread lacustrine carbonates of the Green River Formation (Ryder et al., 1976; Pitman et al., 1982; Burton et al., 2014) that are stratigraphically interbedded with the fluvial deposits. The overlying Sunnyside Delta Interval of this study (Fig. 2.1) is a ca 150 meters thick fluvial-deltaic and lacustrine deposit stratigraphically bound by younger wide-spread carbonate intervals, the D Marker below and the C Marker above (Remy, 1992; Keighley et al., 2003; Morgan, 2003). The age of the Sunnyside Delta Interval is constrained by an absolute age date from the C Marker of 49.6 Ma (Smith et al., 2015), as well as the 49 Ma and 48.4 Ma dates below and above the Mahogany interval higher in stratigraphy.
(Smith et al., 2010) (Fig. 2.1). The Sunnyside Delta Interval is below loosely constrained by an older 54 Ma age in the Carbonate Marker Unit (Remy, 1992), and Paleocene fauna in the lower part of the Colton/Wasatch Formation (Fouch et al., 1987) (Fig. 2.1).

The Sunnyside Delta Interval thus coincides with part of the Early Eocene Climatic Optimum (EECO) (e.g. Littler et al., 2014). Changes in precipitation patterns in response to the EECO global warming have been interpreted to result in highly seasonal and flashy discharge (Plink-Björklund and Birgenheier, 2013; Plink-Björklund et al., 2014; Plink-Björklund, 2015; Rosenberg et al., 2015; Gall et al., 2017).

Previous work (Jacob, 1969; Remy, 1992; Morgan, 2003) has documented the volumetric abundance of fluvial deposits in the Sunnyside Delta Interval in the Nine Mile Canyon. Pusca (2003) interpreted the extensive fluvial deposits as a terminal fan system (sensu Kelly and Olsen, 1993; Legarreta and Uliana, 1998) and linked the fluvial deposits to wet-dry climate cycles in a semi-arid conditions. Pusca (2003) was also the first to observe that many of the sedimentary structures in Sunnyside Delta Interval indicate high deposition rates. Recent work has mostly focused on the effects of variable discharge (Plink-Björklund and Birgenheier, 2013; Plink-Björklund et al., 2014; Plink-Björklund, 2015; Rosenberg et al., 2015; Gall et al., 2017), and on deltaic deposits (Schomacker et al., 2010; Moore et al., 2012).

2.4 Dataset and Methods

In this outcrop study, about 550 m of stratigraphic sections at 5-10 cm resolution were measured at 8 locations in Nine Mile Canyon (Fig. 2.1). Proportions of sandstone sedimentary facies were calculated based on the measured-section dataset. A total of 111 paleocurrent directions measurements were gained from parting lineations and the dip directions of cross strata and laminae, or projected by the dip directions measured from the steepest surfaces of scours related to cross strata (Fig. 2.1). Since the Sunnyside Delta Interval has an abundance of Froude supercritical flow sedimentary structures that may dip down- or upstream and are
commonly gentle, the measurements are few, therefore we also use paleocurrent directions measured throughout the Wasatch/Colton and Green River Formations and documented in previous papers (Fouch et al., 1976; Dickinson et al., 1986; Remy, 1992; Schomacker et al., 2010; Ford et al., 2016; Gall et al., 2017; Jones, 2017). These combined measurements indicate a consistently northward direction with variability from northwest to northeast (Fig. 2.1B). Dip directions of accretion sets were measured on exposed bedding surfaces and plotted on stereonets. Sixteen photomosaics were used to address lateral sedimentary facies variations and internal architecture in 2D and 3D exposures. A laser range finder (TRUPULSE™ 360B) was used for channel thickness and width measurements. Channel widths were measured in outcrops as close to perpendicular to paleocurrent directions as possible. This paper focuses on channel deposits only, and associated floodplain and distal deltaic deposits are not documented here.

2.5 Facies Groups

Twelve sedimentary facies were defined using field observations on grain size, textures, sedimentary structures, bar architectures and trace fossils (Table 2.1, Fig. 2.2). The facies were then divided into two facies groups based on the sedimentary structures.

2.5.1 Facies Group 1: Sandstones and Conglomerates with Long-wavelength, Low-angle Laminations with Foresets and Backsets, and Planar Lamina-

This facies group is the most abundant, consisting of about 70% of facies in sandstone and conglomerates of channel lithosomes. The facies consists of conglomerates (<2%, F 1.1) and mostly very fine to lower medium grained sandstones with scour and fill structures (30%, F 1.2), ripple-filled scour and fill structures (<1%, F 1.3), convex-up low angle laminae (8%, F 1.4), gradational planar laminae (14%, F 1.5), distinct planar laminae (20%, F 1.6), structureless (1%, F1.7), and soft sediment deformation (2%, F1.8) (Table 2.1, Figs. 2.2 and
All sedimentary structures in group 1 are of long wavelength (>1m-10s of m) and have gently dipping laminae (mostly <15°). The thickness of lamina sets varies from a few dm to a few m. The internal laminations are on mm to cm scale.

Facies 1.1 consists of erosionally bound conglomerates with sandy matrix (Figs. 2.3A and B). Conglomerates contain mostly pebbles, but up to cobble sized limestone clasts and bioclasts, and/or mudstone clasts occur. The clasts are subrounded (Fig. 2.3B) to angular (Fig. 2.3A) and poorly sorted. Some clasts are elongate in shape and mostly a few cm to 2 dm wide. The clast composition is, in most cases, the same as the underlying siliciclastic or carbonate deposits. The limestone intraclasts are mostly unorganized rather than stratified and occur as layers a few cm thick in the upper portion of sandstone bodies. The mixed limestone and mudstone clasts form a few cm thick conglomerates at the base of sandstone bodies. In places, they are preserved as scour conformable laminae that flatten upward and transition into planar or convex-up planar laminae with a wavelength of several meters, which is essentially a hummock-like configuration. Clast imbrication is also common, and detrital plant fragments are common in places at the base of basal scour surfaces. The conglomerate facies occur throughout the Sunnyside Delta Interval.

Facies 1.2 consists of upper-very-fine to lower-medium grained, moderately to poorly sorted sandstone with scour and fill structures with upward flattening laminae (Fig. 2.3C). The laminae fill the scours asymmetrically in most cases with laminae dipping down- and upstream. Some scours are filled symmetrically. Dip angle of the laminae varies from 5-25° but is in most cases gentler than 15°. This facies occurs at the bases of sandstone bodies above basal erosion surfaces, or within the sandstone bodies alternating with other facies. Scour and fill structures have a wavelength from a few cm to 10s of m and thickness on cm to m scale. In places, sets of scour and fill structures occur throughout the beds with thicknesses from a few dm to a few m. Thicker, meter scale beds with coarser grained sandstones tend to have larger scour and fill structures. In places, detrital organic matter is distributed along the laminations.
Facies 1.3 consists of upper-very-fine to lower-fine grained moderately sorted sandstone with scour and fill structures similar to Facies 1.2, but asymmetric ripple sets fill the upward flattening laminae, or in places occur only at the lowest parts of the scour and fill structures (Fig. 2.3D). In places, entire beds that are a few dm to a couple of m thick consist of this facies only (Fig. 2.2), though not as commonly as Facies 1.2.

Facies 1.4 consists of sandstone with convex-up low angle laminae bounded by low-angle scours (Fig. 2.3E). Laminae dip very gently (< ca 10°) in down- and upstream directions. This facies is a few dm to 1m thick and the low-angle laminae have a wavelength of a few meters. These structures are commonly of longer wavelength and lower amplitude as compared to Facies 1.2 and 1.3. This facies commonly occurs together with scour and fill structures (F1.2; F1.3) in lower parts of channel lithosomes.

Facies 1.5 consists of very-fine to fine grained, poorly to moderately sorted sandstone with gradational planar laminae (Fig. 2.3F). In places this facies contains more than 50% carbonate grains, such as ooids and ostracods. Internally, each lamina is normally or reversely graded with gradational boundaries. Lamina thickness varies from 2 mm to 2 cm in a few dm to a few m thick sets. In places, the laminations become undulating laterally across a few meters. The carbonate grains are evenly distributed or occur as layers. On lamina scale, both normal and reverse grading trends are common consisting of larger sized carbonate grains and smaller siliciclastic grains. On a bed scale, in places coarsening upward trends occur, where gradational planar laminae grade upward into a carbonate grainstone composed of 100% carbonate grains.

Facies 1.6 contains very-fine to lower-medium grained moderately sorted sandstone with distinct planar laminae (Fig. 2.3G). Boundaries between laminae are distinct rather than gradational. Laminae are on mm scale and bed thickness is on a few dm to a few m scale. Parting lineations occur in some places.

Facies 1.7 is fine to very coarse grained, poorly to moderately sorted structureless sandstone with no visible sedimentary structures. Thickness varies from a few cm to a few dm.
This facies is only associated with facies group 1.

Facies 1.8 is soft sediment deformed sandstone occurs in facies group 1 as convolute bedding with mostly vertical limbs and overturned limbs in some cases (Fig. 2.3H). Deformed zones are a few dm thick and a couple of m wide.

In many places group 1 facies transition vertically (measured sections in Figs. 2.2 and 2.3) or laterally into each other. For instance, sandstone with scour and fills (F1.2; Figs. 2.3C and D) transition laterally and vertically into convex-up low angle lamina (F1.4; Fig. 2.3E), and planar lamina (F1.5 and 1.6; Figs. 2.3F and G). Planar laminations (F1.5 and 1.6) become convex-up low angle laterally in many places. Vertically these facies also grade into group 2 facies. Contact between different facies can be erosive to gradual. In some cases, gradational planar laminations (F1.5) are interbedded with climbing ripple laminations (F2.3). Sandstone with distinct planar lamina (F1.6) is commonly overlain by sandstone with climbing ripple laminations (F2.3) with a gradual contact.

Interpretation: By comparison to sedimentary structures produced by experiments, the facies group 1 sedimentary structures are the product of Froude supercritical flow (see Alexander et al., 2001; Cartigny et al., 2014). Supercritical flow bedforms are mutually transitional as linked to increasing flow velocities, but primarily to increase in maximum Froude number (Cartigny et al., 2014). Low-angle concave- and convex-up laminations (F1.1 to F1.4) are likely to be produced by antidunes and steeper scour and fill structures by chute-and-pool formation or cyclic steps (Alexander et al., 2001; Cartigny et al., 2014). Cyclic steps are a more likely interpretation where steep scour and fill structures are amalgamated, rather than transition into other facies laterally (Cartigny et al., 2014). Stable antidunes form due to passage of internal waves from entirely in supercritical flow conditions, whereas unstable antidunes, chutes and pools and cyclic steps form at hydraulic jumps where flow reverses to subcritical conditions (Alexander et al., 2001; Cartigny et al., 2014). Foresets occur where bedforms migrate downstream, whereas backssets indicate upstream migration (Cartigny et al., 2014). Planar laminations (F1.5 and 1.6) that laterally transition to low-angle...
lamination (F1.1 to F1.4) have been produced by experiments lateral to stable antidunes or where antidune aggradation rates are low (Cartigny et al., 2014). Planar laminations (F1.5 and 1.6) that do not transition laterally to low-angle laminations may form in subcritical, transcritical or supercritical flow conditions (e.g. Paola, 1989; Cheel, 1990). The transition from planar lamination to low-angle lamination is especially likely at high deposition rates (Leclair and Arnott, 2005; Duller et al., 2008). Gradational laminations (F1.5) indicate high deposition rates (Duller et al., 2008). Experiments further show that high aggradation rates commonly occur in supercritical flow conditions, especially at hydraulic jumps where flow abruptly transitions from supercritical to subcritical with a large energy loss (Alexander et al., 2001; Cartigny et al., 2014), promoting deposition of structureless sandstones (F1.7). Convolute bedding (F1.8) indicates rapid porewater escape and may be related to rapid loading by overlain sediments, or to porewater expulsion during rapid lowering of water levels, such as during waning phase of floods (e.g. Williams, 1971; Stear, 1985; Singh and Bhardwaj, 1991). The accumulation and preservation of these facies requires high deposition rates during peak discharge and rapid flow deceleration prevents sediments from reworking by subcritical flow (Alexander and Fielding, 1997).

2.5.2 Facies Group 2: Sandstones with Steep Foresets

Description: This facies group contains very fine to lower medium grained, moderately sorted sandstone with cross stratification (F2.1), climbing cross stratification (F2.2), ripple lamination (F2.3), and climbing ripple lamination (F2.4). The latter two facies have finer grained sandstone than the former (Table 2.1; Fig. 2.4). Facies of this group form ca 30% of the sandstone volume. Climbing ripple laminated sandstone (F2.4) is the most common facies in this group. This facies group is characterized by steep foresets (ca 30°). Set thickness is 6 cm to 20 cm in cross strata and 8 cm to dm scale in sandstone with climbing cross strata (Figs. 2.4A and B). Ripple lamina set thickness is on cm scale and less than 5 cm (Figs. 2.4C and D). Set boundaries in climbing cross strata (F2.2) and climbing ripple laminae
(F2.4) dip upstream and opposite to the cross strata and laminae dip direction. Dip angle of set boundaries is mostly less than 15° (Figs. 2.4B and D).

Interpretation: Cross strata form by migration of dunes and cross laminae form by migration of ripples (e.g. Simons et al., 1965). These are characteristic bedforms in subcritical flow conditions, where dunes occur at higher flow velocities and in some cases in coarser grain sizes than ripples (e.g. Ashley, 1990). Subcritical flow bedforms can only migrate downstream and thus preserve only foresets as bedload sediment is eroded from the stoss side and deposited on the lee side (Simons et al., 1965; Ashley, 1990; Allen, 1986; Miall, 2014). Climbing bedforms suggest high sediment accumulation rates (Allen, 1970), with sediment fallout rate exceeding the bedform migration rate.

2.6 Channel Marcoforms

Sunnyside Delta Interval channel lithosomes refers to lithosomes that are recognized as deposits of active channels and bars within a channel belt. Channel deposits contain 70% Froude super- and transcritical, and 30% subcritical flow sandstone sedimentary facies, of which ca 80% indicate high deposition rates. The channel lithosomes also contain some in-channel mudstone layers (Figs. 2.5A to C), which vary in color from purple to brown to greenish grey (F3.1 to 3.4 in Table 2.1). In places vertical or horizontal bioturbations occurs (Figs. 2.5E to G). We document six different channel lithosome styles based on channel dimensions, macroform geometries, sedimentary facies and the degree of lateral and vertical amalgamation.

2.6.1 Large Vertically and Laterally Amalgamated Channel Lithosomes

Description: The large vertically and laterally amalgamated channel lithosomes are 10-35 m thick, and 100s of m to km wide in a flow oblique view. Some such lithosomes are massive in appearance and form low relief and sharp-based “sandstone walls” along Nine Mile Canyon (Fig. 2.6A). Others are lenticular in shape in flow perpendicular view, have a smaller
width with composite erosional base and a flat top, commonly with a few meters of local truncation at the base (Fig. 2.6B). They commonly display multiple internal erosion surfaces in various degrees of continuity outlining lenticular and tabular sandstones with different shapes (Figs. 2.2A and 2.6). Some internal erosion surfaces cut through the entire lithosome (Fig. 2.6B), whereas others bound low-angle downstream accretion sets (Fig. 2.6A). Some channel lithosomes are bounded by a flat top and a concave up basal surface (Fig. 2.6B), in others both bounding surfaces are concave up or irregular (Fig. 2.6). Erosion surfaces are commonly aligned with mudstone clasts (Figs. 2.2 and 2.3A), and in a few places wood fragments exist along the basal erosion surface. The lithosomes are dominated by moderately to poorly sorted, fine grained sandstones (Figs. 2.2A and 2.3) with Froude supercritical flow sedimentary structures (85%) and/or high deposition rate sedimentary structures (78%), and characteristically lack in-channel mudstone layers. Typical sedimentary structures observed in this type of channel lithosome (Figs. 2.2A and 2.3) include scour and fill structures (38%, F1.2), gradational planar laminations (21%, F1.5), convex-up low-angle laminations (8%, F1.4), and soft sediment deformation (4%, F1.8). There is a minor proportion of Froude subcritical flow sedimentary structures (9%), such as cross strata (2%, F2.1) and climbing ripple laminae (7%, F2.4). Some lithosomes have a few dm thick conglomerate at the base (Figs. 2.2A, 2.3A and B), which rarely stratified. No in-channel vegetation or bioturbation occurs.

Interpretation: Accretion sets abound by erosion surfaces are interpreted as bar clinoforms (sensu Mohrig et al., 2000), and the erosionally bound lenticular sandstones as individual channels. Laterally extensive sandstones with internal erosion surfaces indicate channel amalgamation. The sedimentary structures indicate dominant deposition from Froude super- and trans-critical flow conditions and high deposition rates (see Alexander et al., 2001; Cartigny et al., 2014). Abundance or dominance of Froude super- and trans-critical flow sedimentary structures is characteristic for variable discharge rivers where deposition dominantly occurs during high magnitude floods (Fielding, 2006; Fielding et al., 2009; Plink-Björklund,
Erosionally-bound, low-angle downstream accretion sets have been recognized as flood event beds (sensu flood units of Plink-Björklund, 2015) deposited from individual floods or flood waves (Tunbridge, 1981; Sneh, 1983; Turner, 1986; Abdullatif, 1989; Deluca and Eriksson, 1989; Nichols and Hirst, 1998; Shukla et al., 2001; Hinds et al., 2004; Hampton and Horton, 2007; Chakraborty and Ghosh, 2010; Chakraborty et al., 2010; Mader and Redfern, 2011; Donselaar et al., 2013; Plink-Björklund, 2015). This channel lithosome type is similar to that in the proposed facies model of Plink-Björklund (2015) for extremely flashy rivers where most deposition occurs during high-magnitude floods as a result of short intense rainfall events.

2.6.2 Large Channel Lithosomes with Upstream Migrating Macroforms

Description: The large channel lithosomes with upstream macroform migration are 10-15 m thick, a few 100s m wide in a flow parallel view, and close to a 100 m wide in flow perpendicular view. They form a complex geometry with flat or undulating tops and undulating erosional bases (Fig. 2.7). The local erosional relief on the compound basal surface is a few m with a lateral spacing of 10s of m, whereas the accretion-set bounding erosion surfaces are steep (5-20°) and 10-15 m deep, and upstream-facing (towards south) (Fig. 2.7A). These erosion surfaces bound lenticular sandstones that are individually 10-15 m thick and at least 150 m wide in flow parallel view. The erosionally-bound sandstone lenses display steep, upstream facing geometries along the erosion surfaces that flatten laterally upstream towards the next erosion surface (Fig. 2.7A). The sandstones are dominantly fine grained with no apparent vertical grain size trends, other than along basal scours, where grain size is up to very coarse grained sand and mixed with gravel grains. These coarse basal layers are structureless (F1.7). Wood pieces and mud clasts occur above the basal scours in places. These sandstones consist of 40% Froude supercritical flow and/or 100% high deposition rate sedimentary structures. Soft sediment deformed structures are common (40%) and occur in the lower and upstream part of the lenticular sets (Fig. 2.8B). Soft sediment deformation (F
1.8) occurs as convoluted beds with overturned limbs. They gradually change upward and laterally into climbing cross strata (20%, F2.2), gradational planar laminations (10%, F1.5), and scour and fill structures (30%, F1.3). The cross-strata foresets (F2.2) dip northeast as compared with the southward dip of the accretion surfaces. Scour and fill structures (F1.3) consist of upward flattening laminae that fill meter wide scours, and dip in both in north and south directions at approximately 5-15°. They commonly pass into planar and convex-up low angle laminations (F1.6 and 1.4) laterally and vertically.

Interpretation: The upstream-dipping accretion sets are backsets (sensus Kostic et al., 2010; Cartigny et al., 2014) characteristic for Froude supercritical flow deposition in chutes and pools or cyclic steps. The systematic backsets with regularly spaced scours also suggest deposition in cyclic steps (Kostic et al., 2010; Cartigny et al., 2014). Cyclic steps migrate upstream as erosion occurs in Froude supercritical flow conditions on the upstream side of a hydraulic jump, where flow returns to subcritical conditions and causes rapid deposition at the hydraulic jump (Kostic and Parker, 2006; Cartigny et al., 2014) (Fig. 2.7C). The strata tend to be flatter downstream of the hydraulic jump, where the current starts accelerating and gradually reaches supercritical conditions in the upstream part of the next hydraulic jump. Highest deposition rates occur at the hydraulic jump where there is a large energy loss (Figs. 2.8B and C). High deposition rate deposits are suggested to be common in hydraulic jumps (Postma et al., 2009; Postma, 2014; Postma and Cartigny, 2014), such as the structureless (F1.7) and soft-sediment deformed (F1.8) deposits in this dataset (Fig. 2.7B). Upstream migration of cyclic steps results in erosionally bounded asymmetrical lenses that have characteristic vertical trends, with rapidly deposited structureless or graded hydraulic jump deposits above the basal erosion surface, overlain by sub- to supercritical flow deposits with internal sedimentary structures (Postma, 2014). Channel-scale or bar-scale cyclic steps have hitherto not been documented in river deposits, but cyclic steps on similar scales have been documented in deltas (Massari, 1996; Ventra et al., 2015) and in slope channels (e.g. Ono and Plink-Björklund, 2017).
2.6.3 Channel Lithosomes with Low Angle Downstream Accretion Sets

Description: The channel lithosomes with low-angle accretion sets are 10-15 m thick and at least 100s of m wide in a strike view. These channel lithosomes consist of tabular or wedge shaped accretion sets, 0.5-2 m thick that dip at 4-10° NNE (downstream) (Fig. 2.8). Accretion set bounding surfaces are sharp and draped by mm to cm thick mud layers. The sandstones are moderately to well sorted, and very fine to fine grained. There are no obvious vertical grain size trends (Fig. 2.2C). These channel lithosomes have a high proportion of Froude supercritical flow sedimentary structures (73%) and/or high deposition rate sedimentary structures (80%). The sedimentary structures are ripple-filled scour and fill structures (28%, F1.3), distinct or gradational planar lamina (45%, F1.5 or F1.6) and climbing ripple laminations (27%, F2.4). Some lithosomes consist completely of climbing ripples (Fig. 2.2C). Sets of climbing ripples with different climb angles are bounded by erosion surfaces (Fig. 2.4E). A few dm thick couplets with planar lamina (F1.5 or F1.6) and climbing ripples (F2.4) are common as well (Fig. 2.4F). The transition from planar lamina (F1.5 or F1.6) to climbing ripples (F2.4) is mostly distinct (Fig. 2.2) and in places erosional. Foresets of the climbing ripples dip NNE similar to the low-angle accretion sets.

Interpretation: The low angle downstream dipping accretion sets are interpreted as downstream accreting macroforms. Dip angle of the accretion sets is much lower than bar slip-face cross strata commonly described in sandy channel lithosomes (e.g. Bridge and Lunt, 2006), and have been suggested as common macroform geometries in variable discharge rivers (Plink-Björklund, 2015; see also Tunbridge, 1981; Stear, 1985; Olsen, 1989; North and Taylor, 1996; Zaleha, 1997a; Thomas et al., 2002; Kumar et al., 2003; Allen et al., 2013, 2014; Plink-Björklund and Birgenheier, 2013).

Transitions between Froude supercritical and subcritical flow sedimentary structures (climbing ripples) are suggested to be linked to a rapid flow deceleration, common in high magnitude floods, where deposition rate remains high during the falling flood limb (e.g., Chakraborty and Ghosh, 2010). In-channel mud deposits, and especially lack of ripples be-
between the transitions from supercritical flow structures to overlying mud deposits indicates extremely rapid decline in flow strength, with no low-stage reworking of the flood deposit (e.g., Jones, 1977; Tunbridge, 1981). The sandstone and mudstone couplets form flood event deposits, interpreted as deposits of single floods or flood waves (Tunbridge, 1981; Sneh, 1983; Turner, 1986; Abdullatif, 1989; Deluca and Eriksson, 1989; Nichols and Hirst, 1998; Shukla et al., 2001; Hinds et al., 2004; Hampton and Horton, 2007; Chakraborty and Ghosh, 2010; Chakraborty et al., 2010; Mader and Redfern, 2011; Donselaar et al., 2013; Plink-Björklund, 2015). Such newly formed mud layers have been documented after floods in modern variable discharge rivers (e.g. Stear, 1985; Abdullatif, 1989; Singh et al., 1993; Billi, 2007).

2.6.4 Amalgamated Channel Lithosomes with High Angle Accretion Sets

Description: The channel lithosomes with high-angle (most 10-15°, but up to 20°) accretion sets consist of interbedded sandstones and mudstones and are up to 14 m thick, and 10s of m wide in a flow perpendicular view (Fig. 2.9). The accretion sets are a few dm to a meter thick, and some thin towards the base (Fig. 2.9A and some in Fig. 2.9B), whereas others thin towards the top of the sets (some in Figs. 2.9B and C). Mudstone layers are cm to dm thick (Fig. 2.2D) and in places thicken towards the base of the accretion sets (Fig. 2.9A). The interbedded sandstone and mudstone accretion sets dip at a high angle (up to 20°, stereonet plots in Fig. 2.9), and in places display a sigmoidal shape in a flow parallel view and a tabular or wedge shape in a flow perpendicular view (Figs. 2.9C and D). Dip direction of accretion set varies from obliquely upstream to lateral (Fig. 2.9A), to obliquely downstream facing (Figs. 2.9B and C). In places, cosets in accretion sets are bound by erosion surfaces (Fig. 2.9B). Commonly they lack grain size trends, but in places mud proportion increases towards the toes of the accretion sets, forming a general upward coarsening trend (Figs. 2.2D and 2.9A). The accretion sets consist of 35% Froude supercritical flow sedimentary structures (scour and fill, F1.2; distinct planar lamina, F1.6; gradational planar lamina, F1.5), 35% high deposition rate sedimentary structures, and mostly ripple
cross laminations (F2.4). Desiccation cracks (Fig. 2.5D), plant material (Figs. 2.5E and 2.6) and trace fossils (Fig. 2.5G) are commonly present at accretion set boundaries (Fig. 2.2D). Isolated trace fossils are mm to a dm long.

Interpretation: The accretion direction indicates obliquely upstream to lateral and downstream accretion directions. This is in contrast to the established facies models where steep accretion sets are assigned exclusively to lateral accretion (Miall, 2014). Upward coarsening trends in accretion sets have been linked to deposition during floods (Ghinassi et al., 2016, 2017), when helical flow pattern is destroyed by the shift of the zone of maximum boundary shear stress from channel thalweg to bar top (Hooke, 1975; Dietrich et al., 1979; Dietrich and Smith, 1983). The erosionally bound downstream accretion sets have geometries similar to those described from deltas (Massari, 1996; Ventra et al., 2015) and in slope channels (e.g. Ono and Plink-Björklund, 2017) and are linked to aggradational infilling of hydraulic jump scours.

This channel lithosome type has a lower proportion of Froude supercritical flow deposits and a higher proportion of climbing ripples. The latter are interpreted to be linked to the rapid deceleration of flood waters with high sediment concentration (e.g. Mckee et al., 1967; Picard and High, 1973; Croke et al., 1998). The preservation of mud is unique relative to the previous lithosomes. Thick mud layers further suggest a rapid deceleration with accompanying mud deposition during waning stages of floods (e.g., Jones, 1977; Tunbridge, 1981). The in-channel desiccation cracks indicate seasonally dry channels, common in arid to sub-humid climates (Sneh, 1983; Nichols and Fisher, 2007). However, the water table level was relatively high as the vertical burrows are less than 1 dm long (Hasiotis, 2004).

2.6.5 Heterolithic Aggradational Channel Lithosomes

Description: Heterolithic aggradational channel lithosomes consist of interbedded dm to m thick sandstones and dm thick mudstones that fill scours up to 10 m deep and 100 m wide. The sandstone layers commonly thin laterally (Fig. 2.10). In places the aggradational sets
are conformable with the basal erosion surface and flatten upward (Fig. 2.10A). In other places the sets are flat lying and terminate laterally onto the scour surface (Fig. 2.10B). Some set boundary contacts are erosional, especially towards the top of the channel lithosomes, and overlain by sandstones or mudstones (Fig. 2.10A). Some channel lithosomes extend outside the visible channel-bounding erosion surface towards the top and form “wings” or fill other scours laterally (Fig. 2.10). In places they are erosionally truncated above (Fig. 2.10B). The sandstones are moderately sorted, very fine grained and mostly moderately to highly bioturbated. A considerably smaller amount of Froude supercritical flow sedimentary structures (20%, planar laminations, F1.5 and 1.6) and/or high deposition rate sedimentary structures (50%, climbing ripple laminations, F2.4; gradational planar laminations, F1.5) are present. Most common sedimentary structures are climbing ripple cross laminations (40%, F2.4). In some places, sandstones are heavily bioturbated throughout or just at the boundaries of the accretion sets (Fig. 2.2E). Roots occur mostly in growth position on top of the accretion sets. Mudstone layers in sand-to-mud couplets are composed of greenish grey or purple mudstone (F3.2 or 3.1; Figs. 2.2E, 2.5A to C) with some plant fragments and brownish oxidization bands.

Interpretation: The sandstone-mudstone couplets are interpreted as deposits of individual floods or flood event beds (sensu Plink-Björklund, 2015). Similar to channel type 4 a large proportion of sandstones contain climbing ripples (F2.4) and mudstone layers are relatively thick, indicating dominant deposition from late-stage waning floods (Chakraborty and Ghosh, 2010). Trace fossils and pedogenic modification at the accretion set boundaries suggest that channels were dry between the episodic flood events (Nichols and Hirst, 1998; Shukla et al., 2001; Donselaar et al., 2013; Plink-Björklund, 2015). The lateral onlapping relationship and the upward flattening geometry suggest aggradational channel lithosomes. The dominant climbing ripples and gradational planar laminations indicate high deposition rates during the floods and lack of reworking and thus efficient base flow (Fisher et al., 2008; Chakraborty et al., 2010; Plink-Björklund, 2015). During the dry season, soil moisture in
spots can be high enough for terrestrial plants to grow and creatures to live within the channels, which is responsible for the preserved root traces in growth position at accretion set boundaries (Fielding et al., 2009; Allen et al., 2014; Bashforth et al., 2014). Aggradation or vertical accretion has also been suggested as a common characteristic for variable discharge rivers and is formed due to vertical channel bed aggradation (e.g. Tunbridge, 1981; Stear, 1985; Olsen, 1989; Zaleha, 1997; Thomas et al., 2002; Kumar et al., 2003; Allen et al., 2013, 2014; Plink-Björklund, 2015).

2.6.6 Small Lenticular Channel Lithosomes

Description: The small lenticular channel lithosomes have lenticular shape with flat tops and convex-up erosional bases (Fig. 2.11). They are commonly encased in purple and greenish mudstones. The channel lithosomes are ca 2 m thick, and 10-10s of m wide in a flow-perpendicular view (Fig. 2.11). Some are laterally truncated and partially amalgamated with some thin mudstones in between (Figs. 2.11A and C). Others thin in both directions (Fig. 2.11B). Sandstones are very fine grained and contain climbing ripple cross laminations (80%, F2.4), climbing cross strata (10%, F2.2), and soft sediment deformation (10%, F1.8), which sums up to 10% Froude supercritical flow and/or 100% high deposition rate sedimentary structures. Some sandstones completely consist of climbing ripple laminae (Fig. 2.2, F2.4). There are no obvious grain size trends.

Interpretation: The lower proportion of Froude supercritical flow sedimentary structures and a higher proportion of climbing ripples and mud layers indicates relatively low flow velocity compared to the other types of channel types (Mckee et al., 1967; Croke et al., 1998; Chakraborty and Ghosh, 2010), even though the deposition rate is high as seen by the 100% high deposition rate sedimentary structures. This channel lithosome type is somewhat similar to type 5, but much smaller in dimension, and less amalgamated.
2.7 Discussion

The channel deposits in Sunnyside Delta Interval are significantly different from those in the widely used fluvial facies models in that (1) Froude super- and trans-critical flow bedforms are abundant; (2) abundant downstream accretion, upstream accretion, and vertical aggradation exist with only a minor and oblique lateral accretion, in places with upward-coarsening trends; (3) heterolithic channel lithosomes with in-channel muds, in-channel bioturbations, and mud-clast conglomerates are common; (4) erosionally bound flood units are abundant. Some of these characteristics have been linked to variable discharge river deposits (see reviews in Fielding, 2006; Fielding et al., 2009; Plink-Björklund, 2015). Although some of these characteristics are now well established, relatively little is known about depositional dynamics of such rivers. We follow with a comparison to the existing understanding of channels and bars both in persistent and variable discharge rivers with the aim to improve our understanding of variable discharge river dynamics (Fig. 2.12).

2.7.1 Lateral Accretion Sets

Lateral accretion sets occur as minor components in channel types 3 and 4, where they are low and high angle accretions sets respectively (Figs. 2.7 and 2.8). The latter accretion sets can be considered morphologically similar to point-bar accretions sets (Allen, 1965, 1970; Miall, 1985). The differences are the strong oblique downstream (channel type 3) or upstream migration component (channel type 4), and the low accretion angle in type 3 channels, and the upward coarsening trend in channel type 4. Furthermore, a mud plug or counter point bar deposits (Smith et al., 2011), rather than the distinct sandstone-mudstone couplets are expected to be associated with traditional point bars as opposed to the ones in this study.

Point bars are formed by lateral accretion in a direction that is approximately perpendicular to the flow (Bathurst et al., 1977; Dietrich and Smith, 1983; Miall, 1994; Labrecque et al., 2011). Point bars are the result of helical flow dynamics, as the flow gradually loses mo-
mentum from channel base up to bar surface (Bathurst et al., 1977; Ethridge and Schumm, 1978). A fining upward succession is typically developed with a basal erosion surface, coarse lag, followed by cross stratified sandstone that grades into ripple laminated sandstone, and is capped by planar laminated mudstone or siltstone (Tyler and Ethridge, 1983; Miall, 1988; Donselaar and Overeem, 2008; Labrecque et al., 2011). In cross section, point bars occur in characteristic steep accretion sets in flow perpendicular view (Miall, 1994), and relatively flat to convex-upward lobate shape in flow parallel view (Durkin et al., 2015).

In summary, the key differences of the documented lateral accretion sets are their partial downstream or upstream accretion, lower-than-expected angle and in places upward coarsening, together with the sand-mud couplets or flood event beds and the dominance of either Froude supercritical flow deposits or climbing ripples rather than cross strata.

2.7.2 Downstream Accretion

Downstream accretion occurs in channel types 1, 3, and 4, where the former two types are amalgamated sandstones and the latter consist of sandstone-mudstone couplets (Figs. 2.12A to H). In addition to being suggested to be common in variable discharge rivers (Plink-Björklund, 2015 and references therein), downstream accretion has been documented in some braid bars, as well as in mouth bars.

Braid bars are fundamentally compound bars built by unit bars superimposed by dunes (Lunt and Bridge, 2004; Bridge, 2006). Transverse bars are those attached to river bank and grow in direction perpendicular to the flow. Longitudinal bars are normally developed in the middle of channel, have a bar head and bar tail geometry, and mainly accrete downstream at the angle of repose or less (Sambrook Smith et al., 2006, 2009; Bridge and Lunt, 2006). Internally braid bar deposits consist of sets of cross stratified sandstones on different scales (Ashley, 1990; Lunt et al., 2004). No obvious vertical grain size trend occurs in braid bar deposits, though a downstream decreasing grain size is common in both bar scale and river reach scale (Rice and Church, 2010). In a cross section, complex undulated 3rd order
bounding surfaces separate the stacked compound bars (Miall, 1985). The here documented downstream accretion sets (channel types 1, 3, and 4) differ from mid-channel bars in that the accretion sets are of low angle or concave-up (erosional) where steep, and in lack of downstream migrating cross strata and presence of scour and fill structures and climbing ripple laminae.

Due to the common downstream accretion features, the barforms in the Sunnyside Delta Interval has been previously interpreted as mouth bar deposits (Schomacker et al., 2010). Mouth bars occur as sharp or gradationally based lithosomes associated with distributary channels and delta-front deposits, and typically show coarsening and thickening upward trends as they form by progradation of deltaic deposits (Gilbert, 1885; Scruton, 1960; Bhattacharya, 2006). Mouth bars may contain both sediment gravity flow and traction flow deposits (Plink-Björklund and Steel, 2005). Cross strata, ripple lamination, climbing ripple laminations are common sedimentary structures documented in some ancient examples of mouth bars (Olariu and Bhattacharya, 2006), whereas others display gradational planar laminations (Plink-Björklund and Steel, 2004).

The here documented downstream accretion sets (channel types 1, 3, and 4) differ from mouth bars in that they lack the link to delta-front or prodelta deposits, and that the accretion set boundaries are commonly concave up (Fig. 2.9).

### 2.7.3 Upstream Accretion and the Hierarchy of Froude Supercritical Flow Deposits

The channel lithosomes with upstream accretion sets (channel types 2 and 4; Figs. 2.12I to K) have a similar geometry to the cyclic steps produced by experiments (Winterwerp et al., 1992; Parker, 1996; Taki and Parker, 2005; Spinewine et al., 2009), as they display the characteristic asymmetrical geometry of hydraulic jumps, with a steep upstream and gentle downstream margin (Cartigny et al., 2014) and a fill that is structureless and deformed in the lower part and stratified towards the top (Postma et al., 2009, 2014). The cyclic
steps deposits in channel type 2 are 10-15 m thick, but consist internally of dm to m scale Froude transcritical and supercritical flow sedimentary structures. A similar occurrence is described in experiments, where smaller dm to cm scale cyclic steps and antidunes occurred within a large m-scale hydraulic jump (Spinewine et al., 2009). Thus, at least two levels of hierarchy potentially occurred in cyclic step dynamics. On the modern seafloor, various scales of scours (100s to 1000s meter long and 10s to 100s meter deep) and sediment waves (100s or 1000s of meter wavelength and a few meters or 100s meter wave heights) have been recently identified as formed by cyclic steps (Symons et al., 2016; Carvajal et al., 2017). Outcrop work further indicates that the thickness and wavelength of Froude supercritical flow deposits are extremely variable from centimeter to hundreds of meters scale (Postma et al., 2009; Postma and Cartigny, 2014; Ono and Plink-Björklund, 2017).

2.8 Conclusions

Documentation of the Sunnyside Delta Interval fluvial deposits reveals deposition in variable discharge rivers by the abundance of Froude super- and transcritical flow sedimentary structures, in-channel mud deposits, trace fossils, and desiccation cracks. This paper documents six macroform geometries, and shows that variable discharge river macroforms are considerably different from the established point bars or braid bars. This paper presents a first systematic description of such macroforms that have hitherto been referred to as poorly developed (Plink-Björklund, 2015) or even missing (Fielding et al., 2018) in variable discharge rivers. This paper document down- and upstream migrating and obliquely lateral accretion sets, and a variety of accretion angles. In many cases the accretion sets are erosionally bound and resemble geometries produced by hydraulic jumps in experiments.
Figure 2.1 Stratigraphic column and base map of Uinta Basin and Nine Mile canyon. (A) The study area (Nine Mile Canyon) in the Uinta Basin (Utah, U.S.A.). The extent of Colton / Wasatch and Green River Formations are in maroon and orange respectively. Location of Fig. 1B is highlighted in a small box with dashed lines. Modified from Dickinson et al. (2012), Schomacker et al. (2010), Sato and Chan (2015), and Jones (2017). (B) Measured stratigraphic sections (M1.1, 1.2; 2; 3; 4; 5; 6.1, 6.2 in Fig. 2.2) and figure numbers are marked in dark grey star shapes. Grey numbers are mile marks from 37 to 49 miles along Nine Mile Canyon Road. Paleocurrent directions are compiled from several measured sections and plotted in rose diagrams to show the variation at different locations from west to east. Black color in rose diagrams represent all paleocurrent measurements, and the grey ones are taken from true cross stratifications. (C) Stratigraphic framework in this study. Sunnyside Delta Interval (outlined in dashed lines) is the focus of this paper, located in the middle part of Green River Formation, and bounded by D and C markers. Modified from Fouch et al. (1987), Remy (1992), and Smith et al. (2010, 2015).
Figure 2.2 Representative examples of measured sections organized by channel lithosome types.
Figure 2.3 Example outcrop photos and measured sections of Facies group 1: sandstones and conglomerates interpreted as Froude supercritical deposits. (A) Stratified and imbricated conglomerates (F1.1). (B) Disorganized conglomerates (F1.1). (C) Sandstone with scour and fill structure (F1.2). (D) Sandstone with ripple filled scour and fill structure (F1.3). (E) Sandstone with convex-up low angle lamina (F1.4). (F) Sandstone with gradational planar lamina (F1.5). (G) Sandstone with distinct planar lamina (F1.6). (H) Soft sediment deformed sandstone (F1.8). Stars mark the position of respective facies in the measured sections.
Figure 2.4 Example outcrop photos and measured sections of Facies group 2: sandstones with steep foresets interpreted as Froude subcritical flow deposits. (A) Cross stratified sandstone (F2.1). (B) Sandstone with climbing cross strata (F2.2). (C) Rippled sandstone (F2.3). (D) Sandstone with climbing ripple lamina (F2.4). Stars mark the position of respective facies in the measured sections. (E) Vertical transition from climbing ripples with high angle climb angles to low climb angles. (F) Couplets of planar lamina and climbing ripples.
Figure 2.5 Example photos of in-channel mudstones and bioturbation. (A) Purple mudstone with bioturbation (F3.1). (B) Purple greenish grey mixed color siltstone (F3.5). (C) Fine laminated purple to greenish grey siltstone (F3.1; F3.2). (D) Desiccation cracks preserved at accretion set boundaries. Stars mark the position of respective facies in the measure sections. (E-F) Root traces in structureless sandstones. (G) Ancorichnus in ripple laminated sandstones.
Figure 2.6 Large vertically and laterally amalgamated channel lithosomes with low-angle accretion sets. (A) Example of an amalgamated channel lithosome with low relief concave top and bottom surfaces. (B) Examples of a lenticular amalgamated channel lithosomes. Bar clinoforms dipping to the left indicate downstream accretion.
Figure 2.7 Large channel lithosomes with upstream migrating macroforms. (A) Channel macroforms consist of upstream dipping accretion sets, bound by erosion surfaces. (B) is Part of (A). Example of a single erosionally bound set with characteristic distribution of facies, where structureless or soft-sediment deformed sandstones occur close to the basal scour and more laminated facies towards the top in downstream direction. (C) Sketch of a hydraulic jump illustrating the supercritical upstream edge, the hydraulic jump where deposition rates are highest, and the gentler downstream side with subcritical to transcritical flow conditions. (D) A flow perpendicular view of the same channel lithosome. (E) The paleocurrent data plotted in rose diagram to the left was collected from the cross stratified sandstone layer in Cottonwood Canyon. The rose diagram to the right is the dip measurements from macroforms dipping upstream.
Figure 2.8 Channel lithosomes with low-angle accretion sets. (A, C) Downstream accreting accretion sets or bar clinoforms are clearly visible in flow-parallel view. (B) Obliquely perpendicular view shows the same channel lithosomes with an obliquely lateral component. (D) Plots the paleocurrents and dip angle of the accretion sets in rose diagram and stereonet. White arrows show average paleoflow direction in each outcrop photos (A to C) and the plan view map (D).
Figure 2.9 Channel lithosomes with high-angle (up to 20°) accretion sets with preserved mudstone between sand bodies. Paleocurrent directions are from ripple cross lamination. Together with the dip angle of accretion sets are plotted to the right. (A) Obliquely upstream and lateral accretion sets with sandstone layers thin and mud layers that thicken towards the base, forming a coarsening upward trend. (B) Amalgamated erosionally bound downstream accretion sets. (C) Obliquely downstream and lateral accretion sets. (D) A more flow perpendicular view of the same channel lithosome as in C. White arrows are average paleoflow direction in each example.
Figure 2.10 Heterolithic aggradational channel lithosomes. (A) Upward flattening aggradational sandstone layers interbedded with thin mud layers, truncated on top by an erosion surface. (B) Interbedded sandstone and thick mudstone layers onlap onto the erosive scour surface.
Figure 2.11 Small lenticular channel lithosomes. (A) Laterally amalgamated lenticular sandstones. (B) Vertically amalgamated lenticular sandstones. (C) Laterally and vertically amalgamated lenticular sandstones.
Figure 2.12 Summary diagram of the 6 channel lithosome types in flow parallel (to the left, arrow points downstream) and perpendicular views (right). Red lines are erosion surfaces. Black lines are bedding surfaces and sedimentary structures. Yellow and green/purple represent sandstone and mudstone respectively. Examples of A, C, E, F, and G are downstream accretion styles. B, D, and H are the corresponding styles in flow perpendicular view. Upstream accretion styles are shown in I to K. Examples L and M look aggradational in both views.
Table 2.1: Description and Interpretation of Sedimentary Facies

<table>
<thead>
<tr>
<th>Facies</th>
<th>%</th>
<th>Textures</th>
<th>Structures</th>
<th>Width/Thickness</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.1 Stratified and imbricated conglomerates</td>
<td>&lt;1%</td>
<td>Conglomerates (coarse sand to pebble sized carbonate clast/bioclast and/or mud clast, subrounded to angular, poorly sorted, varied in shape) in sandy matrix, clast or matrix supported</td>
<td>Imbricated, planar stratified, low angle stratified, scour and fill structures,</td>
<td>Thickness: a few cm per conglomerate layer; Total thickness: a few dm - m</td>
<td>Intrabasinal clasts of floodplain mudstones or lake carbonates deposited with sand matrix by traction flow (subcritical to supercritical)</td>
</tr>
<tr>
<td>1.1 Disorganized conglomerates and diamictites</td>
<td>&lt;1%</td>
<td>Conglomerates (coarse sand to pebble sized carbonate clast/bioclast and/or mud clast, rounded to angular, poorly sorted, varied in shape) in sandy or muddy matrix, in places with terrestrial organic matter (woody fragments)</td>
<td>disorganized</td>
<td>Thickness: a few dm - 1m</td>
<td>Intrabasinal clasts of floodplain mudstones or lake carbonate deposited by collapse and fluidization of channel banks during falling stage of flow</td>
</tr>
<tr>
<td>1.2 Sandstone with scour and fill structures</td>
<td>30%</td>
<td>Upper very fine to lower medium grained sandstone, moderately to poorly sorted, in places mixed with carbonate grains</td>
<td>Scours filled with upward flattening lamina</td>
<td>Thickness: a few dm - a few m; Width: a few dm - 10s of m</td>
<td>Supercritical flow, high deposition rates, suspension deposition, large antidune or chute and pool formation</td>
</tr>
<tr>
<td>1.3 Sandstone with ripple-filled scour and fill structures</td>
<td>&lt;1%</td>
<td>Upper very fine to lower fine grained sandstone, moderately sorted</td>
<td>Ripples fill the toe part of upward flattening scour and fill lamina</td>
<td>Thickness: a few dm - 1m; Width: a few dm - a few m</td>
<td>Supercritical to subcritical flow, high deposition rates, suspension and bedload deposition</td>
</tr>
<tr>
<td>Facies</td>
<td>%</td>
<td>Textures</td>
<td>Structures</td>
<td>Width/Thickness</td>
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<tr>
<td>1.4 Sandstone with convex-up low angle lamina</td>
<td>8%</td>
<td>Very fine to fine grained sandstone, moderately to poorly sorted, in places mixed with carbonate grains</td>
<td>Low angle convex-up laminations, very low angle hummocky-like cross strata with upward increasing steepness</td>
<td>Thickness: a few dm - 1m; Width: &gt;1m</td>
<td>Supercritical flow, high deposition rates, suspension deposition, antidune formation</td>
</tr>
<tr>
<td>1.5 Sandstone with gradational planar lamina</td>
<td>14%</td>
<td>Very fine to fine grained sandstone, moderately to poorly sorted</td>
<td>Diffused, gradational planar laminations, each lamina is internally reversely to normally graded</td>
<td>Thickness (lamina): 2mm - 2cm; Thickness (bed): a few dm - a few m</td>
<td>Subcritical to supercritical flow, high deposition rates, suspension deposition, plane bed formation</td>
</tr>
<tr>
<td>1.6 Sandstone with distinct planar lamina</td>
<td>20%</td>
<td>Very fine to lower medium grained sandstone, moderately sorted</td>
<td>Planar laminations with distinct lamina boundaries</td>
<td>Thickness (lamina): mm scale; Thickness (bed): a few dm - a few m</td>
<td>Supercritical or transcritical flow, normal deposition rates, bedload deposition, plane bed formation</td>
</tr>
<tr>
<td>1.7 Structureless sandstone</td>
<td>1%</td>
<td>fine to very coarse grained sandstone, moderately to poorly sorted</td>
<td>No apparent sedimentary structures</td>
<td>Thickness: a few cm - a few dm</td>
<td>high deposition rates, suspension deposition</td>
</tr>
<tr>
<td>Facies</td>
<td>%</td>
<td>Textures</td>
<td>Structures</td>
<td>Width/Thickness</td>
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<tr>
<td>1.8 Soft sediment deformed sandstone</td>
<td>2%</td>
<td>Very fine to lower medium grained sandstone, moderately to poorly sorted</td>
<td>Flame structures, overturned folds, dish structures, ball and pillow structures.</td>
<td>Thickness: a few dm</td>
<td>Water escape or local collapse</td>
</tr>
<tr>
<td>2.1 Cross stratified sandstone</td>
<td>2%</td>
<td>Fine to lower medium grained sandstone, moderately sorted</td>
<td>Planar and trough cross stratified</td>
<td>Thickness (cross set): 5cm -20cm; Thickness (co-set): a few dm</td>
<td>Subcritical flow, normal deposition rates, bedload deposition, dune migration</td>
</tr>
<tr>
<td>2.2 Sandstone with climbing cross strata</td>
<td>&lt;1%</td>
<td>Fine to lower medium grained sandstone, moderately to poorly sorted</td>
<td>Planar and trough cross strata with climbing set boundaries</td>
<td>Thickness (cross set): 8cm - dm scale; Thickness (co-set): a few dm</td>
<td>Subcritical flow, high deposition rates, suspension and bedload deposition, dune migration and aggradation</td>
</tr>
<tr>
<td>2.3 Cross laminated sandstone</td>
<td>10%</td>
<td>Very fine to lower fine grained sandstone, moderately sorted</td>
<td>Asymmetrical cross-lamination</td>
<td>Thickness (cross set): cm scale &lt; 5cm; Thickness (co-set): a few dm</td>
<td>Subcritical flow, normal deposition rates, bedload deposition, ripple migration</td>
</tr>
<tr>
<td>Facies</td>
<td>%</td>
<td>Textures</td>
<td>Structures</td>
<td>Width/Thickness</td>
<td>Interpretation</td>
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</tr>
<tr>
<td>2.4 Sandstone with climbing ripple lamina</td>
<td>13%</td>
<td>Very fine to fine grained sandstone, moderately sorted</td>
<td>Asymmetrical cross lamination with climbing set boundaries</td>
<td>Thickness (cross set): cm scale &lt;5cm; Thickness (co-set): a few dm - a few m</td>
<td>Subcritical flow, high deposition rates, suspension and bedload deposition, ripple migration and aggradation</td>
</tr>
<tr>
<td>2.5 Bioturbated sandstone</td>
<td>-</td>
<td>Very fine to lower medium grained sandstone, moderately to poorly sorted</td>
<td>Vertical, horizontal and 3D trace fossils</td>
<td>Thickness: a few dm - a few m</td>
<td>Trace fossils formed by insects, dwelling, resting, crawling traces</td>
</tr>
<tr>
<td>3.1 Purpled red mudstone</td>
<td>-</td>
<td>Clay to silt</td>
<td>Laminated or blocky or crumbly, in places rooted, bioturbated, or deformed</td>
<td>Thickness: a few cm - a few dm (blocky); a few mm (each lamina)</td>
<td>Oxidized, well drained in-channel mudstone</td>
</tr>
<tr>
<td>3.2 Greenish grey mudstone</td>
<td>-</td>
<td>Clay to silt</td>
<td>Laminated or blocky or crumbly</td>
<td>Thickness: a few cm - a few dm (blocky); a few mm (each lamina)</td>
<td>Reduced to poorly drained in-channel mudstone</td>
</tr>
<tr>
<td>3.3 Brown to dark grey mudstone</td>
<td>-</td>
<td>Clay to silt</td>
<td>laminated</td>
<td>Thickness: a few cm - a few dm</td>
<td>Oxidized poorly drained in-channel mudstone</td>
</tr>
<tr>
<td>3.4 Mixed color mudstone</td>
<td>-</td>
<td>Clay to silt,</td>
<td>Blocky greenish grey to purple</td>
<td>Thickness: a few dm</td>
<td>Variable Red-ox in-channel mudstone</td>
</tr>
</tbody>
</table>
2.9 Reference Cited


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53

CHAPTER 3
VERTICAL AND LATERAL FACIES VARIATIONS IN FLUVIAL FANS: EARLY EOCENE GREEN RIVER FORMATION, UINTA BASIN

A paper submitted to Basin Research

Jianqiao Wang*1 and Piret Plink-Björklund1

3.1 Abstract

In this study, outcrop measured sections and photomosaics with a GPS survey are integrated with areal mapping of channel dimensions, channel to floodplain ratio, and sedimentary facies variability to study the channel and floodplain deposits in the Sunnyside Delta Interval of the Early Eocene Green River Formation in the Uinta Basin. The documented lateral extent, internal architecture, and lateral and vertical facies transitions indicate a fluvial (mega)fan. Sandying and thickening upward successions occur as an increase in channel to floodplain ratio, channel size, and the degree of channel amalgamation. Similar trends are also observed laterally as channel fill facies become more heterolithic, channel amalgamation degree decreases and the proportion of floodplain deposits increases downstream. There are multiple scales of upward sandying and thickening packages. The smallest scale packages are avulsion packages that form the building blocks of the stratigraphy. The larger-scale upward thickening and sandying successions are interpreted as whole fan and lobe progradation. High avulsion rates and channel return frequency are interpreted to control the high degree of channel amalgamation in the proximal fans.

The avulsion stratigraphy indicates that the succession is comparable to fluvial megafans rather than to distributive fluvial systems (DFS), fluvial distributary systems or terminal

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fans that all invoke bifurcations rather than avulsions. Furthermore, the abundant Froude supercritical flow and high deposition rate facies, in-channel mud deposits and in-channel bioturbation and desiccation indicate variable discharge rivers, similar to modern fluvial megafans.

3.2 Introduction

Large fluvial fan systems or fluvial megafans can cover areas of 1,000 to 100,000 km$^2$ (Leier et al., 2005), and have a significant hydrocarbon reservoir potential. While previous studies on Pleistocene to modern fluvial fans (e.g. Shukla et al., 2001; Assine, 2005; Wilkinson et al., 2006; Chakraborty et al., 2010; Latrubesse et al., 2012; Rossetti et al., 2012; Assine et al., 2014; Pupim et al., 2017) have assessed their geomorphology, self-organization processes, as well as their link to foreland basins and seasonal climate (Leier et al., 2005; Fontana et al., 2008, 2014; Assine et al., 2014), documentation on how fluvial facies vary laterally is rare (Singh et al., 1993; Shukla et al., 2001; Chakraborty et al., 2010). Ancient datasets provide data on facies and their variability, but some systems are interpreted as fluvial megafans (e.g. DeCelles and Cavazza, 1999; Horton and Decelles, 2001; Saez et al., 2007; Trendell et al., 2013), others as distributive fluvial systems (DFS) (Leleu and Hartley, 2010; Leleu et al., 2010; Weismann et al., 2014; Owen et al., 2015, 2017), fluvial distributary systems (Nichols and Hirst, 1998; Nichols and Fisher, 2007; Fisher and Nichols, 2013), or terminal fans (Kelly and Olsen, 1993; Pusca, 2003). In contrast to the widely used sequence stratigraphy models that link amalgamated and isolated channels zones to lake or sea-level cycles (e.g. Keighley et al., 2003), degree of channel amalgamation has been shown to vary from proximal to distal, and thus progradation and backstepping of fluvial fans builds a predictable stratigraphy (Singh et al., 1993; Shukla et al., 2001; Chakraborty and Ghosh, 2010; Weismann et al., 2014; Owen et al., 2017). Such proximal-distal relationships in facies trends (Trendell et al., 2013; Weismann et al., 2014; Owen et al., 2015, 2017), and paleosol moisture variability (Bhattacharyya et al., 2011; Hartley et al., 2013) have also been
recognized in ancient systems.

This paper focuses on documentation of vertical and lateral facies relationships that goes beyond the simple proximal-distal and vertical trends and links channel lithosome types and floodplain sand content to their position in the fan. We compare to modern fan systems, and discuss the role of self-organization in building the fan stratigraphy, and how it relates to the different fan models. The specifics of channel architecture and the interaction of fluvial and lacustrine deposits are not discussed in this paper, but addressed in companion studies.

3.3 Geological setting and background

The 25,000 km$^2$ Uinta Basin was formed to the east of the Sevier Thrust Front in response to the Laramide uplift of the Uinta Mountains to the north (Cashion, 1967; Dickinson et al., 1988; DeCelles, 2004). The basin is approximately 210 km from west to east and 275 km from north to south (Cashion, 1967). Fluvial facies in the Green River Fm are extensively distributed across the southern margin of the Uinta Basin (Ryder et al., 1976). The Green River Fm is subdivided into three informal units: lower, middle, and upper, based on lacustrine pulses including the Uteland Butte Member, the Carbonate Marker Unit, and the Mahogany Oil Shale Member (Fig. 3.1) (Fouch, 1975; Pitman et al., 1982; Fouch et al., 1987). The three major siliciclastic intervals are the Colton Tongue, the Renegade Tongue, and the Sunnyside Delta Interval. D and C Marker beds mark the bottom and top of the Sunnyside Delta Interval (Remy, 1992; Morgan, 2003). The C marker is equivalent to the C1 marker of Jacob (1969). The timing of the Sunnyside Delta Interval deposition (Fig. 3.1) is constrained by an absolute age date from the C marker of 49.6 Ma (Smith et al., 2015) and 54 Ma from the Carbonate Marker Unit (Remy, 1992).

Previous work has mainly focused on the fluvial-deltaic deposition and sequence stratigraphy of the Sunnyside Delta Interval (Remy, 1992; Keighley et al., 2003; Taylor and Ritts, 2004; Schomacker et al., 2010; Moore et al., 2012), but some recent studies have interpreted that the Sunnyside Delta Interval is one of the fluvial fans (Plink-Björklund and Birgenheier,
2013; Plink-Björklund et al., 2014) developed by the California paleoriver that originated from the Mojave segment of the Cordilleran magmatic arc (Davis et al., 2010; Dickinson et al., 2012). The Sunnyside Delta Interval has also been previously interpreted as a terminal fan deposit (Pusca, 2003). The Wasatch/Colton age fan apex is estimated to have prograded from about 250 km south of the Uinta Basin into the southern part of the basin (Jones, 2017). Over a thousand paleocurrent measurements from Colton/Wasatch Formation were collected along the south basin margin (Jones, 2017) to reconstruct the apex location. Paleocurrent measurements from Jones (2017) and other previous studies indicate consistent northward (ranging from northwest to northeast) fluvial sediment transport in the southern Uinta Basin margin in Colton/Wasatch as well as Green River Formations (Fouch et al., 1976; Dickinson et al., 1986; Remy, 1992; Schomacker et al., 2010; Ford et al., 2016; Gall et al., 2017; Jones, 2017).

### 3.4 Methods and dataset

Field data from the Sunnyside Delta Interval (Fig. 3.1) was collected from the northwest-to-southeast oriented Nine Mile Canyon and several side canyons in north-to-south directions (Fig. 3.1). Observations cover an area from the 32- to 45-mile marker across ca 20 km in the Nine Mile Canyon, and continue to the mile marker 49 to the south in the Cottonwood Canyon. A total of about 550 m of stratigraphic sections were measured at 10 cm resolution. Percentage of sedimentary facies was calculated using measured sections. Lateral relationships of channel-floodplain lithosomes were examined on cliff-face exposures and mapped by walking out the stratigraphic intervals in continuous outcrops or using the bounding carbonate marker units (D and C markers) as correlation guides in discontinuous outcrops. Where possible, channel lithosome bounding surfaces were physically followed to where they pinch out. Photomosaics were used to assist with analyzing channel architecture, and calculating channel-floodplain proportions. Architectural dimensions (e.g. channel width and thickness) were obtained by using laser range finder (TRUPULSE™ 360B) in field.
3.5 Facies Associations in Fluvial Channel-floodplain Lithosomes

Fluvial channel-floodplain lithosomes are organized into facies associations according to their relative geographic location and lateral relationships, and defined by lithology, channel size, channel lithosome architecture, degree of channel amalgamation, channel-floodplain deposit ratio, and trace fossils. Facies are summarized in Table 3.1.

3.5.1 Facies Association 1: Large Vertically and Laterally Amalgamated Sandy Channel and Floodplain Lithosomes

Facies association 1 consists of amalgamated sandstones 10-35 m thick and 100s of m to km wide, and interbedded mudstones and sandstones with variegated colors (Fig. 3.2). Basal surfaces are either concave-up or undulating. These sandstones have the appearance of a sandstone wall in outcrops, as they form 85% of the facies association. They commonly display multiple internal irregular concave-up erosion surfaces (Figs. 3.2B and C) with mudstone clasts along some of these surfaces. The sandstones consist of erosionally bound, downstream- (Fig. 3.2B) and upstream-oriented accretion sets (Fig. 3.2C). Local erosional relief on the compound basal surface is a few meters, whereas the accretion set boundaries that are also erosional are 10-15 m tall (Fig. 3.2B). The thick sandstones of facies association 1 are characterized by a high proportion of climbing and aggradational bedforms (78-100%, F1.2, 1.4, 1.5, and 1.8 in Table 3.1) low-angle convex- and concave-up sedimentary structures (40-85%, F1.2, 1.4, 1.5, and 1.6 in Table 3.1). The dominant sedimentary structures are scour and fill structures (F1.2, Figs. 3.3A and B, Table 3.1), convex-up low-angle laminations (F1.4, Fig. 3.3C, Table 3.1), gradational and distinct planar laminations (F1.5 and 1.6, Figs. 3.3D and E, Table 3.1), and soft sediment deformation (F1.8, Fig. 3.3F, Table 3.1). In the upstream migrating bedforms, a typical vertical transition from soft-sediment deformed sandstone above the basal scour surfaces to sandstones with gradational planar lamina, climbing cross stratification, and scour and fill structures is common (Fig. 3.2C).
Interbedded mudstones and sandstones form less than 15% of facies association 1 (Fig. 3.2, Table 3.2). In places they only occur as lenses below the irregular erosion surfaces of the thick sandstones (Fig. 3.2D). The sandstone beds consist of up to meter-thick very-fine grained sandstone with planar laminations (F1.6, Table 3.1, Fig 3.3E) and ripple cross lamina (F2.3, Table 3.1, Figs. 3.2D and 3.3G). Sandstones form 70% of the interbedded sandstone and mudstone component of facies association 1 (Table 3.2). Mudstones vary in color from purple to greenish grey, and are commonly soft sediment deformed (Figs. 3.3G and H, Table 3.1). Some mudstones are purple (F3.1) or brown (F3.3) throughout and others transition from purple to greenish grey (F3.4), or have irregular brownish or grey patches (Table 3.1). In places, the thick sandstones directly overlay mudstones without the sandstone interbeds (Figs. 3.3G and H). The mudstone facies (Table 3.1) are the same as in facies associations 2 and 3, where they occur in greater thickness and lateral extent, and are better preserved.

Interpretation: The sandstone wall outcrop appearance is due to the high amalgamation degree of sandstones. The accretion sets are interpreted as bar clinoforms (sensu Mohrig et al., 2000), and the thick sandstones are interpreted as amalgamated channels. The dominant sedimentary structures are indicative of Froude supercritical flow, and high deposition rates (Alexander et al., 2001; Cartigny et al., 2014), which are characteristic for variable discharge rivers where deposition occurs from high magnitude floods (Fielding, 2006; Fielding et al., 2009; Plink-Björklund, 2015). The internal erosion surfaces outline individual flood event beds (sensu flood units in Plink-Björklund, 2015) (Figs. 3.2B to D). The upstream migrating macroforms are interpreted as deposition in cyclic steps because the systematic backsets are similar to cyclic step deposits produced by experiments, where they occur as bounded by the upstream facing erosion surfaces and separated by regularly spaced scours (Cartigny and Postma, 2010; Kostic et al., 2010; Cartigny et al., 2014). The resultant vertical trend in sedimentary structures in the backsets is characteristic for upstream migration of cyclic steps, indicating rapidly deposited hydraulic jump deposits above the basal erosion surface, overlain by sub- to super-critical flow deposits deposited further downstream (Postma et al.,
2014). The interbedded mudstones and sandstones are interpreted as floodplain deposits with interbedded floodplain mudstones and crevasse splay or sheetflow deposits. Mottled coloring is common in paleosols (Retallack, 1988; Leckie et al., 1989; Wright, 1992), and is a result of local redox changes due to fluctuating drainage conditions (Wright, 1992). Mottled color and soft sediment deformation are characteristic for paleosol with variable drainage conditions (Retallack, 1988; Leckie et al., 1989; Wright, 1992). The direct juxtaposition of channel lithosome and floodplain mudstone indicates abrupt avulsions (Jones and Hajek, 2007). The high degree of vertical and lateral channel amalgamation indicates high avulsion rates and a high channel return frequency (Hajek and Wolinsky, 2012; Hajek and Edmonds, 2014), and is the reason for low preservation of floodplain facies.

3.5.2 Facies Association 2: Laterally Amalgamated Sandy to Heterolithic Channel Accretion Set and Floodplain Lithosomes

Facies association 2 is dominated by 3-15 m thick laterally amalgamated sandstones with low (4-10°) or high angle (up to 20°) downstream and upstream accretion sets (Fig. 3.4). The degree of sandstone amalgamation is lower than in facies association 1, and accretion set thickness is 57-70% of the thickness in facies association 1 (Table 3.2). The overall sandstone and heterolithic accretion set lithosome percentage is 40-70% (Table 3.2). The low angle accretion sets consist of tabular or wedge-shaped sandstones, 0.5-2 m thick, and form at least 100s of m wide lithosomes in strike view (Fig. 3.4A). The high angle sets are 10s of m wide in a flow perpendicular view, and composed of sandstone layers that are a few dm to one meter thick and thin towards either the base or the top of the sets. Facies association 2 has a relatively lower proportion of climbing and aggradational bedforms (35-80%, F1.2, 1.3, 1.5, 1.6, and 2.4 in Table 3.1) and low-angle convex- and concave-up sedimentary structures (35-73%, F1.2, 1.3, 1.5, and 1.6 in Table 3.1). The most common facies are scour and fill structures (15%, F1.2), ripple-filled scour and fill structures (28%, F1.3), gradational and distinct planar laminations (30%, F1.5 and 1.6), and climbing ripple
laminations (27-35%, F2.4) (Table 3.1, Fig. 3.5). Vertical transition from low-angle convex- and concave-up sedimentary structures (F1.2, 1.3, 1.5, and 1.6) to climbing ripples (F2.4) is common (Fig. 3.5A). Some lithosomes are entirely composed of climbing ripple laminations (Fig. 3.5B). Some accretion sets fine from sandstones to mm-dm thick mudstones (F3.1 to 3.4) laterally within individual accretion sets. In places flute casts occur below the accretion sets (Fig. 3.5C). Mm to dm thick in-channel mud layers (Fig. 3.5D), with desiccation cracks (Fig. 3.5E) and terrestrial trace fossils up to 1 dm long (Figs. 3.5G and H) are common at accretion set boundaries.

The sandstone and heterolithic accretion set lithosomes are laterally and vertically associated with interbedded mudstones and sandstones with variegated color. The mudstones are dm thick and purple (F3.1, Fig. 3.5I) or brown (F3.3) throughout, transition from purple to greenish grey (F3.4), or have irregular brownish or grey patches (F3.4, Table 3.1). Sandstones are commonly dm thick, but thicken upward to m-thick beds in up to 10 m thick intervals that have a lateral extent of 100s of meters (Figs. 3.4A and C). The vertical transition from interbedded sandstone and mudstone with upward thickening sandstones to sandstone and heterolithic accretion set lithosomes is common (Figs. 3.4A and C). In a few places there is an abrupt transition from mudstones to sandstone (Fig. 3.5I) and heterolithic accretion set lithosomes. The sandstone proportion in the interbedded mudstone and sandstone lithosomes decreases from 70% to around 50% compared to facies association 1 (Table 3.2). Bioturbation is common in the purple mudstone. In one area, possible mammal tracks occur in a 3 dm thick heavily bioturbated purple mudstone (Fig. 3.5H). Calcite filled cracks and calcite nodules are common in all four mudstone facies.

Interpretation: The accretion sets are interpreted as bar clinoforms (sensu Mohrig et al., 2000), and the sandstones lithosomes are interpreted as laterally amalgamated channels. The dominant sedimentary structures are indicative of Froude supercritical flow, and high deposition rates (Alexander et al., 2001; Cartigny et al., 2014), and are characteristic for variable discharge rivers where deposition occurs from high magnitude floods (Field-
ing, 2006; Fielding et al., 2009; Plink-Björklund, 2015). The common transition between Froude supercritical flow structures and climbing ripples is suggestive of a rapid decline in flow velocity, a characteristic of flood events (e.g. Mckee et al., 1967; Picard and High, 1973; Croke et al., 1998; Plink- Björklund, 2015). In-channel mud layers are interpreted to be deposited from even more rapidly waning floods (e.g., Jones, 1977; Tunbridge, 1981; Plink- Björklund, 2015). The presence of trace fossils and desiccation cracks in channels at accretion set boundaries indicate that the channels were dry and the beds exposed for sustained periods between floods (Sneh, 1983; Nichols and Fisher, 2007; Plink- Björklund, 2015). Mottled color, roots, and nodules are common in paleosols (Retallack, 1988; Leckie et al., 1989; Wright, 1992). Burrowing activity and roots are indicators for subaerial exposure and surface stability (Retallack, 1988; Leckie et al., 1989). The calcite filled cracks are possible calcified root structures (Leckie et al., 1989). Mottling is a result of local redox changes due to fluctuating drainage conditions (Wright, 1992). Red mudstones indicate well drained floodplain, whereas purple mudstones had higher soil moisture, and brown mudstones indicate sustained soil moisture (Kraus and Aslan, 1996; Alonso-Zarza, 2003). The thickening upward trends have been interpreted as transitional river avulsion deposits (Kraus and Wells, 1999; Jones and Hajek, 2007) that indicate successively larger amounts of sand transported to floodplain, followed by a channel avulsion. In contrast, direct transition from floodplain mudstone to channel lithosome is interpreted to be associated with abrupt avulsion, where channel avulsion is not preceded by gradual splay buildup (Jones and Hajek, 2007). While the latter is likely to indicate abrupt avulsions, it is unclear whether the former is a strong indicator for gradual avulsions, as there is no robust way to test the genetic link between the channel and the floodplain deposits. Furthermore, the floodplain sandstones may originate from sheetfloods rather than distinct splays.
3.5.3 Facies Association 3: Small Isolated Channel and Floodplain Lithosomes

Facies association 3 consists of isolated or partially amalgamated lenticular lithosomes with heterolithic aggradational accretion sets (Fig. 3.6B) that form ca 10-30% of the facies association, and interbedded sandstones and mudstones. The lenticular lithosome thickness decreases by ca 30-50% compared to the accretion set thickness in facies association 2 (Table 3.2). The heterolithic aggradation sets consist of interbedded dm to m thick sandstones and dm thick mudstone layers in 2-10 m deep and 10s to a 100 m wide erosion surfaces. The sandstone layers commonly thin towards channel margins (Figs. 3.6A and B). Some terminate onto the basal erosion surface and others are truncated on top. Only 10-20% low-angle convex- and concave-up sedimentary structures (F1.5, 1.6, and 1.8 in Table 3.1) including gradational and distinct planar laminations (F1.5 and 1.6) and soft sediment deformation (F1.8) occur in these lenticular lithosomes. Climbing ripple laminations are dominant with 40-80% (F2.4, Fig. 3.7A, Table 3.1). Mudstone layers are greenish grey or purple (Figs. 3.5E and F), and contain some plant fragments and brownish oxidation bands (Fig. 3.7B). Desiccation cracks occur at some accretion set boundaries. Sandstones are commonly burrowed throughout or at the accretion set boundaries (Fig. 3.7D).

The interbedded mudstones and sandstone lithosomes form 70-90% of facies association 3 and consist of dm-thick mudstone (F3.1-3.4) and cm to dm thick ripple laminated (F2.3) or structureless (F1.7) very-fine grained sandstone. Interbedded mudstones and sandstones form up to 10s of meter thick lithosomes that are laterally continuous across 100s of m to kms and show upward thickening sandstone trends (Figs. 3.6B and 3.7E), or are laterally eroded by lenticular lithosomes. Commonly, the interbedded mudstone and sandstone beds are extensively rooted or bioturbated and sedimentary structures are not preserved (Fig. 3.7C). In places the mudstones coarsen upward into ripple laminated sandstones on dm scale (Fig. 3.7F).

Interpretation: The erosionally based lenticular lithosomes are interpreted as isolated channels encased in floodplain deposits. There is a considerably smaller amount of Froude
supercritical flow sedimentary structures, and the abundance of climbing ripples and gradational planar laminations indicates high deposition rates. Desiccation cracks, trace fossils, and pedogenically modified in-channel mudstones bound flood event beds indicate that channels were dry for sustained periods. Such desiccated in-channel mudstones, and climbing-ripple filled channels are shown to occur in variable discharge rivers (Sneh, 1983; Nichols and Hirst, 1998; Shukla et al., 2001; Nichols and Fisher, 2007; Allen et al., 2011; Donselaar et al., 2013; Plink-Björklund, 2015). The interbedded mudstone and sandstone beds are interpreted as crevasse splays and floodplain mudstones, due to their lateral extent, presence of roots, and the internal coarsening upward trend from structureless mudstone to ripple laminated sandstone (Mohrig et al., 2000; Jones and Hajek, 2007; Sendziak, 2012).

3.6 Vertical Trends

Upward sandying and thickening trends on multiple scales are common in all the facies associations. The smallest scale of such trends consist of upward thickening floodplain sandstones, interbedded with mudstones and capped with channels, mostly 2-5 m thick (Figs. 3.8A-C; see also Figs. 3.2D, 3.4A, C, and 3.6). These upward sandying packages are laterally continuous for at least 100s of m to kms (Figs. 3.4A and 3.6A). Statistical analysis shows that such packages can be recognized on 87% of the beds in measured sections as bed thickening as well as grain-size-coarsening trends (Figs. 3.8C-E). A total number of 82 such packages are documented in the numerical dataset and their thickness ranges from 1.5-20 m with a median of 8 m, and 25th and 75th percentiles at 2-12 m (Fig. 3.8D). The thick outliers are likely to contain multiple packages that are not recognized by the analyses as they lack the capping channel deposits.

Intermediate scale upward thickening and sandying packages are most commonly 23-40 m thick and consist of upward transitions from facies association FA3 to FA2, from FA2 to FA1, or from FA3 to FA1 (Figs. 3.2A, 3.4A, and 3.6A). These larger-scale packages extend laterally for 10s of kilometers as many of such packages can be followed across the Nine Mile
Canyon study area (Fig. 3.9). A total number of 15 such packages are documented in the numerical dataset and their thickness ranges from 22-55 m with a median of 29 m, and 25th and 75th percentiles at 23-40 m (Fig. 3.8D).

The largest scale upward thickening and sandying trends are most commonly 59-103 m thick (Fig. 3.8D) and consist of multiple upward facies association transitions, such as for example the double stack from FA3 to FA2 in Fig. 3.6A. Lateral extent of these largest scale packages is at least 20 km as they extend across the Nine-Mile Canyon study area, but it is yet unclear if they extend across the whole basin. A total number of 4 such packages are documented in the numerical dataset and their thickness ranges from 56-139 m with a median of 78 m, and 25th and 75th percentiles at 59-103 m (Fig. 3.8D).

Interpretation: The smallest-scale packages are interpreted as avulsion packages as they indicate splay or sheetflood buildout, followed by channel avulsions (Jones and Hajek, 2007). Their lateral extent is dependent on the alluvial ridge width as determined by the distance of overbank sediment transport during floods. These avulsion packages are the basic building blocks of the studied fluvial stratigraphy. The origin of the intermediate scale packages is less obvious and will be discussed later in this paper.

### 3.7 Lateral Trends

The Sunnyside Delta Interval has been mapped across the more than 150 km wide southern basin margin. This mapping shows considerable variability in the distribution of facies associations 1-3 within the fluvial succession, as well as in the proportions of the deltaic and lake facies. In general, the proportion of FA3 increases considerably within the channel-floodplain deposits, while channel size and the total proportion of channel-floodplain deposits decrease towards the northwestern basin margin (Road 191 outcrops, Fig. 3.10). The proportion of FA1 is largest in the central part of the southern basin margin, such as in the easternmost part of the Nine Mile Canyon and along the Green River (Fig. 3.1).
In addition to these basin-scale trends there is considerable lateral facies variability across the ca 20 km wide continuous outcrops within the Nine Mile Canyon. A general trend is that the degree of channel amalgamation, proportion of sandstone, channel size, and thus the proportion of FA1 decrease from the southeasternmost outcrops in the Cottonwood Canyon towards the northwest along the Nine Mile Canyon. The channel to floodplain ratio decreases from ca 80% in the easternmost Nine Mile Canyon (mile markers 46-45) to ca 50% across ca 20 km towards the northwest (mile marker 32) (Fig. 3.11). In addition to these general trends there is further internal variability. For example, within certain stratigraphic intervals the proportion of FA1 decreases from 90% to 30% as it transitions into FA3 over only a few kilometers distance (Fig. 3.9). Whereas in other intervals a similar transition occurs more gradually through transitioning into FA2 first and then into FA3 across ca 5 km (Fig. 3.9). Yet in other intervals FA2 persists across the whole study area. At mile marker 46 and 45 (Figs. 3.11A and B), the upper part of the exposed succession is a “sandstone wall” with amalgamated channels of FA1, and the lower part FA2 with a large heterolithic channel is at marker 46. At mile 44.5, ca 800 m downstream, the uppermost “sandstone wall” (FA1) has transitioned into an upward sanding package of FA3 and FA2 (Fig. 3.11C). The lower part displays a lateral change from FA1 to FA2 and FA3 across only a few 100 m (Fig. 3.11C). The exposure at mile marker 44 shows a larger proportion of the Sunnyside stratigraphy (Fig. 3.11D), where the upper upward sanding packages contain FA3 and FA2, but the stratigraphically lower section consists of FA1. At mile 41.2 (Fig. 3.11E) the upper part of the stratigraphy is sandstone dominated as the uppermost upward sanding packages have transitioned back to FA1 across ca 4.5 km. There are also some lake deposits that appear at some of the upward sanding package boundaries (orange layers in Fig. 3.11E). At mile marker 37, the whole succession consists of transitions between FA3 and FA2, and majority of the channel fills are heterolithic (Fig. 3.11F). At mile marker 32 the whole Sunnyside Interval is exposed, and much of the succession consists of deltaic and lake deposits. The channel-floodplain intervals are relatively sandy and channel prone and consist of FA1-3 (Fig.
3.11G).

3.8 Discussion on Vertical and Lateral Trends

The large extent of the Sunnyside Interval river and floodplain deposits, together with the downstream decrease of channel size, the degree of channel amalgamation, and increase in floodplain deposit proportions suggest deposition in a fluvial (mega)fan system (sensu Singh et al., 1993; Shukla et al., 2001). Fluvial fans are built by avulsions (Shukla et al., 2001; Leier et al., 2005; Chakraborty et al., 2010; Latrubesse et al., 2012; Sinha et al., 2012) because only avulsions as opposed to bifurcations (cf. Weissmann et al., 2010; see also discussion in North and Warwick, 2007) can generate amalgamated channel belts. The degree of channel amalgamation is highest in the most proximal fan where the area is smallest and the channel return frequency is the highest (Chakraborty and Ghosh, 2010; see also Hajek and Wolinsky, 2012). The decrease in channel size is linked to water loss to floodplain and infiltration into the fan deposits (Singh et al., 1993; Shukla et al., 2001). The increase in floodplain deposits is related to the increase in outward fan surficial area and the decrease in channel size and degree of amalgamation. Comparison to modern fluvial fans shows that channels are highly amalgamated, such as in FA1, in proximal fans within 5-25 km of the fan apex in the Kosi fan (Singh et al., 1993; Chakraborty et al., 2010), and within 20 km in the Ganga fan (Shukla et al., 2001). The 40-70% channel lithosome proportion of FA2 is similar to the medial fan zone in the Kosi fan where the floodplain mud volume increases to 45% from medial to distal fan within 25-130 km from fan apex (Singh et al., 1993). In the medial zone of the Ganga fan, sandy to heterolithic low-angle downstream and upstream accretion sets, and vertical aggradation sets are documented in the braided zone (Shukla et al., 2001). Within the 50-80 km from fan apex in the Ganga fan, shallow and narrow channels are interbedded with floodplain muds and thin sandstones (Shukla et al., 2001), similar to FA2. The distal Ganga fan, 80-100 km from fan apex is characterized by meandering channels with broad floodplains (Shukla et al., 2001), similar to the laterally extensive floodplain deposits in FA3.
Upward thickening and coarsening trends have been recognized in modern fans as indicative of fan progradation (Chakraborty and Ghosh, 2010; Sinha et al., 2014). Such fan-scale progradation trends have also been suggested in ancient systems (DeCelles and Cavazza, 1999; Uba et al., 2005; Weissmann et al., 2013; Owen et al., 2015, 2017). Yet this dataset shows that not all upward thickening and coarsening trends occur across the entire fan, or even across 20 km (Fig. 3.11). Furthermore, the here documented multiple scales of upward thickening and coarsening packages (Fig. 3.8E), and the recognition of individual avulsion packages (Figs. 3.8A-C) suggest that there are multiple mechanisms for generating such trends. The avulsion scale packages are likely to have the smallest lateral extent, as determined by the width of overbank sediment transport. These channel-floodplain processes on alluvial ridge-scale are likely to determine the shortest scale lateral changes, such as transitions from FA1 to FA2 and FA3 across merely a few 100 m to kms (Fig. 3.11C), where we observe a lateral transition from a locally amalgamated channel belt to a floodplain-dominated succession. While the largest scale vertical upward thickening trends (Fig. 3.8E) are likely to correspond to the entire fan progradation, it remains unclear what controls the intermediate scale packages. A possible hypothesis is lobe-scale progradation, as modern fans are documented to consist of 3-4 lobes (Assine and Silva, 2009; Chakraborty et al., 2010; Chakraborty and Ghosh, 2010; Assine et al., 2014). Such lobes form because channel avulsions cluster within lobes (Chakraborty et al., 2010).

### 3.9 Controls on Fluvial Fan Formation

Fluvial fan formation has been linked to a sufficient aggradation rate and discharge, if a river has a variable (seasonally fluctuating) discharge (Leier et al., 2005). Channel bed aggradation and the resultant channel superelevation are the main trigger of avulsions (Bryant et al., 1995; Mohrig et al., 2000; Makaske, 2001; Stouthamer and Berendsen, 2001; Jerolmack and Mohrig, 2007; Sinha and Sarkar, 2009; Hajek and Edmonds, 2014). The high deposition rate sedimentary structures (F1.1-1.8, F2.2, and F2.4, Table 3.1) indicate that high
deposition rates were common in the Sunnyside system. The latter, together with abundant Froude supercritical flow sedimentary structures (F1.1-1.8, Table 3.1), in-channel mudstones (F3.1-3.4, Table 3.1), and in-channel desiccation and pedogenic modification indicate deposition in variable discharge rivers (Plink-Björklund, 2015). Large discharge is evidenced by the large bar clinoform height, as well as suggested by the size of the California paleoriver (Dickinson et al., 2012).

Modern fluvial fans have high avulsion rates. For example, the Kosi river experienced avulsions on average every 7 years between 1760 and 1960 (Chakraborty et al., 2010). The Kosi river is currently confined by artificial levees, but still experienced avulsions in 2008 and 2009 (Chakraborty et al., 2010). The 2008 avulsion was an abrupt avulsion where the Kosi River shifted more than 100 km eastward across the fan (Sinha, 2009). This avulsion occurred during the annual monsoon flood (Sinha, 2009; Sinha et al., 2013). The avulsed channel reoccupied one of its former paleochannels and 80–85% of the river flow was diverted into the new course. The avulsion occurred as a result of channel bed aggradation and superelevation and a resultant artificial levee break (Sinha, 2009).

3.10 Modern Analogue

The Ili River fluvial fan is deposited on the southeastern margin of Lake Balkash in Khazakstan, ca 600 km from the Tian Shan mountain front (Fig. 3.12). It is thus similar to the setting of the Wasatch/Colton and Green River Formation fluvial systems, in terms of the fan size and shape, the lake size, as well as the fan position away from the mountain front (see also Jones, 2017). Lake Balkhash is a closed lake (Birkett, 1995) with a variable surface area dependent on seasonal fluctuations and water input from the fluvial megafan (Kezer and Matsuyama, 2006). From Google Earth measurements, the fan area is ca 25, 000 km² with an apex-distal fan distance of 250 km and a fan width of 200 km (Fig. 3.12), similar to the estimated apex location of the Wasatch fluvial fan, which was 250 km south of the Uinta Basin during deposition of the distal fan in the Uinta Basin (Jones, 2017). The
distance to Ili River fan from the Tian Shan Mountains is also comparable to that of the California Paleoriver that drained from the Mojave region ca 750 km away from the Uinta Basin (Davis et al., 2010; Dickinson et al., 2012).

3.11 Comparison to Fluvial Fan Models

The abundance of avulsion packages, and the channel amalgamation clearly demonstrate that avulsions are the key process of lateral river mobility in the Sunnyside Delta Interval. This is in agreement with our understanding of fluvial megafans (Shukla et al., 2001; Leier et al., 2005; Chakraborty et al., 2010; Latrubesse et al., 2012; Sinha et al., 2012), but contradicts the distributive fluvial systems (DFS) (Weissmann et al., 2010), fluvial distributary systems (Nichols and Fisher, 2007) and terminal fan (Kelly and Olsen, 1993) models that all invoke bifurcations rather than avulsions as a means of lateral channel mobility. For this reason all these latter models have an inherent discrepancy, as their plan-view models show bifurcations, but the cross-sectional views are based on data that indicate amalgamated to isolated channel and floodplain deposits.

Pusca (2003) interpreted the Sunnyside Delta Interval as a terminal fan system (sensu Kelly and Olsen, 1993; Legarreta and Uliana, 1998) and suggested wet-dry climate cycles in the semi-arid climate zone. The terminal fan is interpreted to prograde and retrograde in respond to the increasing and decreasing precipitation during wet and dry periods (Pusca, 2003). Pusca (2003) was also the first to observe that many of the sedimentary structures in Sunnyside Delta Interval indicate high deposition rates. However, the term “terminal” that implies the fan never reached to the lake (sensu Kelly and Olsen, 1993), is misleading, since the sediments prograded into the basin and formed mouth bars and deltas as the fan met the lake (Schomacker et al., 2010).
3.12 Conclusion

This outcrop study recognizes the Sunnyside Delta Interval channel and floodplain deposits as a fluvial fan system. Detailed study across 20 km in the Nine Mile Canyon links channel and floodplain lithosome types to their position in the fan, and recognizes that self-organization by avulsions is the key component of fan stratigraphy. Upward thickening and sandying trends are recognized on multiple scales, and their origin ranges from avulsions to lobe and fan progradation. Laterally, channel-floodplain lithosomes show a decreasing degree of channel amalgamation, sand to mud ratio, and channel size from proximal to distal fan. In addition to these general lateral trends, the degree of cannal amalgamation and the channel to floodplain proportions vary laterally on multiple scales. We identify that as flu- vial megafans rather than distributive fluvial system (DFS), fluvial distributary system, or terminal fan models, as the latter invoke bifurcations rather than avulsions as the mechanism for lateral channel mobility. A modern analogy of this type system was also provided.
Figure 3.1 Stratigraphic column and base map of Uinta Basin and Nine Mile Canyon. (A) Geological map of the Wasatch/Colton (maroon) and Green River (orange) Formations in the Uinta Basin, with study area locations at Nine Mile Canyon, Road 191 and Hay Canyon (modified from Dickinson et al., 2012; Schomacker et al., 2010; Sato and Chan, 2015; Jones, 2017). (B) Study area at Nine Mile Canyon (see location in A) with figure locations and mile markers (grey numbers). Rose diagrams show paleocurrent directions measured in this study. (C) Stratigraphic table with available age constraints (modified from Fouch et al., 1987; Remy, 1992; Smith et al., 2010, 2015).
Figure 3.2 Outcrop examples of facies association 1. Each has an interpreted pair of photos. (A) Cliff-view at 46-mile marker in Nine Mile Canyon. (B) Vertically and laterally amalgamated channel lithosomes with no floodplain deposits preserved. Internal erosion surfaces bound flood event beds in downstream accretion sets. (C) Upstream migrating macroforms with limited mudstone preserved below local erosional relief. (D) A lens of preserved floodplain deposits (purple) with thick splay sandstones below a channel lithosome (yellow). Facies association and facies components are marked in figures. Large white arrows point in the general paleoflow direction.
Figure 3.3 Photos of common sedimentary structures in facies association 1. (A) Sandstone with scour and fill structure (F1.2). (B) Sandstone with ripple filled scour and fill structure (F1.3). (C) Sandstone with convex-up low angle lamina (F1.4). (D) Sandstone with gradational planar lamina and scour and fill (F1.5). (E) Sandstone with distinct planar lamina (F1.6). (F) Soft sediment deformed sandstone (F1.8). (G) Meter thick soft sediment deformed purple mudstone (F3.1) interbedded with bioturbated sandstone (F2.5) below a channel. (H) Soft sediment deformed greenish grey mudstone (F3.2) below a channel.
Figure 3.4 Outcrop examples of facies association 2. (A) Cliff-view at 39-mile marker in Nine Mile Canyon. (B) Low angle downstream accretion sets. (C) High angle accretion sets dip obliquely upstream. (D) Downstream accretion sets. (E) Channel (yellow) and floodplain (purple) deposits with dm to m thick splay sandstones. Large white arrows point into the general paleoflow direction.
Figure 3.5 Photos of common facies in facies associations 2 and 3. (A) Transition from climbing ripples (F2.4) to planar lamination (F1.5). (B) Sandstone with climbing ripple laminations (F2.4). (C) Flute casts under a channel sandstone. (D) Thin layer of mixed color mudstone (F3.4) preserved between two channel sandstones. (E) Brownish mudstone with calcite nodules. (F) Desiccation cracks preserved at accretion set boundary. (G) Roots in sandstone (F2.5). (H) Vertical burrows in ripple laminated sandstone. (I) Possible mammal tracks preserved at the base of channel sandstone.
Figure 3.6 Outcrop examples of facies association 3. Each has an interpreted pair of photos. (A) Cliff-view of interbedded facies associations 2 and 3 in two upward thickening and sanding units. (B) Isolated channel lithosomes (yellow) with in-channel mudstone layers (gray) encased in floodplain deposits (purple) with dm-thick splay deposits. Large white arrows point in the general paleoflow direction.
Figure 3.7 Photos of common facies in facies association 3. (A) Climbing ripple laminated sandstone (F2.4). (B) Nodules in floodplain mudstone (F3.1). (C) Heavily rooted sandstone surface. (D) 10 cm long root/burrow at accretion set boundary. (E) Coarsening upward trend from thinly laminated mudstone to ripple laminated sandstone (F2.3) in crevasse splay deposits. (F) Close up of the ripple laminated sandstone (F2.3) in (E).
Figure 3.8 Examples of characteristic vertical trends. Photos (A, B) and a measured section (C) of upward thickening and sanding trends in floodplain deposits with capping channel lithosomes. These upward sanding and thickening packages correspond to the smallest, avulsion scale packages. Statistical analysis reveal multiple scales of upward thickening trends (D, E). The blue lines in (D) trace the smallest-scale upward thickening trends. The box and whisker diagrams in (E) show three different scales in upward thickening trends. (F) Comparison of upward thickening trends to trends in grain size and sedimentary structures shows that thickening trends are linked to sanding trends, but sedimentary structures do not follow these trends.
Figure 3.9 Vertical changes in channel (yellow) vs. floodplain (grey) proportions at each examined location (by mile marker). Percentages are estimated from combining measured section and photomosaics. Note the top interval gradually transitions from FA1 to FA2 then to FA3 from mile marker 46 to 43 (across ca 5km). Some other transitions from FA1 to FA3 are more abrupt across over only a few km. Other intervals show FA2 persists across the whole study area.
Figure 3.10 Photos showing thin floodplain-rich fluvial intervals along Road 191 outcrops. (A) Overview of fluvial (channels in yellow and floodplain in red) and lake facies (not colored) proportions. (B) Example of a thin fluvial interval with a heterolithic lenticular channel deposit (yellow) and floodplain deposit with thin splay sands.
Figure 3.11 Example photomosaics that illustrate lateral changes from most proximal to most distal locations, respectively (A through G). See Fig. 3.1 for mile marker locations.
Figure 3.12 Google Earth image showing Ili River that originates from the Tianshan Mountains and builds a large fluvial fan on the southern margin of the Balkhash Lake. (A) The Balkhash Lake. (B) Close-up of the fan. White arrows show river’s path. Note that the fan (B) is not built at the foot of the mountains, but rather, the river transits for a few 100km before it builds the fan.
Table 3.1: Description and Interpretation of Sedimentary Facies

<table>
<thead>
<tr>
<th>Facies</th>
<th>%</th>
<th>Textures</th>
<th>Structures</th>
<th>Width/Thickness</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.1 Stratified and imbricated</td>
<td>&lt;1%</td>
<td>Conglomerates (coarse sand to pebble sized</td>
<td>Imbricated, planar</td>
<td>Thickness: a few cm per</td>
<td>Intrabasinal clasts of floodplain mudstones or lake carbonates deposited with</td>
</tr>
<tr>
<td>conglomerates</td>
<td></td>
<td>carbonate clast/bioclast and/or mud clast,</td>
<td>stratified, low angle stratified,</td>
<td>conglomerate layer; Total</td>
<td>sand matrix by traction flow (subcritical to supercritical)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>subrounded to angular, poorly sorted, varied</td>
<td>scour and fill structures,</td>
<td>thickness: a few dm - m</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>in shape) in sandy matrix, clast or matrix</td>
<td></td>
<td></td>
<td></td>
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<tr>
<td></td>
<td></td>
<td>supported</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1.1 Disorganized conglomerates and</td>
<td>&lt;1%</td>
<td>Conglomerates (coarse sand to pebble sized</td>
<td>Disorganized</td>
<td>Thickness: a few dm - 1m</td>
<td>Intrabasinal clasts of floodplain mudstones or lake carbonate deposited by</td>
</tr>
<tr>
<td>diamictites</td>
<td></td>
<td>carbonate clast/bioclast and/or mud clast,</td>
<td></td>
<td></td>
<td>collapse and fluidization of channel banks during falling stage of flow</td>
</tr>
<tr>
<td></td>
<td></td>
<td>rounded to angular, poorly sorted, varied in</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>shape) in sandy or muddy matrix, in places</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>with terrestrial organic matter (woody</td>
<td></td>
<td></td>
<td></td>
</tr>
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<td></td>
<td></td>
<td>fragments)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1.2 Sandstone with scour and fill</td>
<td>30%</td>
<td>Upper very fine to lower medium grained</td>
<td>Scours filled with upward</td>
<td>Thickness: a few dm - a few m;</td>
<td>Supercritical flow, high deposition rates, suspension deposition, large</td>
</tr>
<tr>
<td>structures</td>
<td></td>
<td>sandstone, moderately to poorly sorted, in</td>
<td>flattening lamina</td>
<td>Width: a few dm - 10s of m</td>
<td>antidune or chute and pool formation</td>
</tr>
<tr>
<td></td>
<td></td>
<td>places mixed with carbonate grains</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1.3 Sandstone with ripple-filled</td>
<td>&lt;1%</td>
<td>Upper very fine to lower fine grained</td>
<td>Ripples fill the toe part of</td>
<td>Thickness: a few dm - 1m; Width:</td>
<td>Supercritical to subcritical flow, high deposition rates, suspension and</td>
</tr>
<tr>
<td>scour and fill structures</td>
<td></td>
<td>sandstone, moderately sorted</td>
<td>upward flattening scour and fill</td>
<td>a few dm - a few m</td>
<td>bedload deposition</td>
</tr>
<tr>
<td></td>
<td></td>
<td>lamina</td>
<td>lamina</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Facies</td>
<td>%</td>
<td>Textures</td>
<td>Structures</td>
<td>Width/Thickness</td>
<td>Interpretation</td>
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<tr>
<td>---------------------------------------------</td>
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<td>----------------------------------------------</td>
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<td>-----------------------------------------------------</td>
</tr>
<tr>
<td>1.4 Sandstone with convex-up low angle lamina</td>
<td>8%</td>
<td>Very fine to fine grained sandstone, moderately to poorly sorted, in places mixed with carbonate grains</td>
<td>Low angle convex-up laminations, very low angle hummocky-like cross strata with upward increasing steepness</td>
<td>Thickness: a few dm - 1m; Width: &gt;1m</td>
<td>Supercritical flow, high deposition rates, suspension deposition, antidune formation</td>
</tr>
<tr>
<td>1.5 Sandstone with gradational planar lamina</td>
<td>14%</td>
<td>Very fine to fine grained sandstone, moderately to poorly sorted</td>
<td>Diffused, gradational planar laminations, each lamina is internally reversely to normally graded</td>
<td>Thickness (lamina): 2mm - 2cm; Thickness (bed): a few dm - a few m</td>
<td>Subcritical to supercritical flow, high deposition rates, suspension deposition, plane bed formation</td>
</tr>
<tr>
<td>1.6 Sandstone with distinct planar lamina</td>
<td>20%</td>
<td>Very fine to lower medium grained sandstone, moderately sorted</td>
<td>Planar laminations with distinct lamina boundaries</td>
<td>Thickness (lamina): mm scale; Thickness (bed): a few dm - a few m</td>
<td>Supercritical or transcritical flow, normal deposition rates, bedload deposition, plane bed formation</td>
</tr>
<tr>
<td>1.7 Structureless sandstone</td>
<td>1%</td>
<td>Very fine to very coarse grained sandstone, moderately to poorly sorted</td>
<td>No apparent sedimentary structures</td>
<td>Thickness: a few cm - a few dm</td>
<td>high deposition rates, suspension deposition</td>
</tr>
<tr>
<td>Facies</td>
<td>%</td>
<td>Textures</td>
<td>Structures</td>
<td>Width/Thickness</td>
<td>Interpretation</td>
</tr>
<tr>
<td>---------------------------------</td>
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<td>---------------------------------------------------------------------------</td>
<td>------------------------------------------</td>
<td>-----------------------------------------------------</td>
</tr>
<tr>
<td>1.8 Soft sediment deformed sandstone</td>
<td>2%</td>
<td>Very fine to lower medium grained sandstone, moderately to poorly sorted</td>
<td>Flame structures, overturned folds, dish structures, ball and pillow structures.</td>
<td>Thickness: a few dm</td>
<td>Water escape or local collapse</td>
</tr>
<tr>
<td>2.1 Cross stratified sandstone</td>
<td>2%</td>
<td>Fine to lower medium grained sandstone, moderately sorted</td>
<td>Planar and trough cross stratified</td>
<td>Thickness (cross set): 5cm -20cm; Thickness (co-set): a few dm</td>
<td>Subcritical flow, normal deposition rates, bedload deposition, dune migration</td>
</tr>
<tr>
<td>2.2 Sandstone with climbing cross strata</td>
<td>&lt;1%</td>
<td>Fine to lower medium grained sandstone, moderately to poorly sorted</td>
<td>Planar and trough cross strata with climbing set boundaries</td>
<td>Thickness (cross set): 8cm - dm scale; Thickness (co-set): a few dm</td>
<td>Subcritical flow, high deposition rates, suspension and bedload deposition, dune migration and aggradation</td>
</tr>
<tr>
<td>2.3 Cross laminated sandstone</td>
<td>10%</td>
<td>Very fine to lower fine grained sandstone, moderately sorted</td>
<td>Asymmetrical cross-lamination</td>
<td>Thickness (cross set): cm scale &lt; 5cm; Thickness (co-set): a few dm</td>
<td>Subcritical flow, normal deposition rates, bedload deposition, ripple migration</td>
</tr>
<tr>
<td>Facies</td>
<td>%</td>
<td>Textures</td>
<td>Structures</td>
<td>Width/Thickness</td>
<td>Interpretation</td>
</tr>
<tr>
<td>------------------------------------------------</td>
<td>-------</td>
<td>--------------------------------</td>
<td>-----------------------------------------------------------------------------------------------</td>
<td>---------------------------------------------------------------------------------------------------</td>
<td>-------------------------------------------------------------------------------------------------------</td>
</tr>
<tr>
<td>2.4 Sandstone with climbing ripple lamina</td>
<td>13%</td>
<td>Very fine to fine grained sandstone, moderately sorted</td>
<td>Asymmetrical cross lamination with climbing set boundaries</td>
<td>Thickness (cross set): cm scale &lt;5cm; Thickness (co-set): a few dm - a few m</td>
<td>Subcritical flow, high deposition rates, suspension and bedload deposition, ripple migration and aggradation</td>
</tr>
<tr>
<td>2.5 Bioturbated sandstone</td>
<td>-</td>
<td>Very fine to lower medium grained sandstone, moderately to poorly sorted</td>
<td>Vertical, horizontal and 3D trace fossils</td>
<td>Thickness: a few dm - a few m</td>
<td>Trace fossils formed by insects, dwelling, resting, crawling traces</td>
</tr>
<tr>
<td>3.1 Purpled red mudstone</td>
<td>-</td>
<td>Clay to silt</td>
<td>Laminated or blocky or crumbly, in places rooted, bioturbated, or deformed</td>
<td>Thickness: a few cm - a few dm (blocky); a few mm (each lamina)</td>
<td>Oxidized, well drained floodplain and in-channel mudstone</td>
</tr>
<tr>
<td>3.2 Greenish grey mudstone</td>
<td>-</td>
<td>Clay to silt</td>
<td>Laminated or blocky or crumbly</td>
<td>Thickness: a few cm - a few dm (blocky); a few mm (each lamina)</td>
<td>Reduced to poorly drained floodplain and in-channel mudstone</td>
</tr>
<tr>
<td>3.3 Brown to dark grey mudstone</td>
<td>-</td>
<td>Clay to silt</td>
<td>Laminated</td>
<td>Thickness: a few cm - a few dm</td>
<td>Oxidized poorly drained floodplain and in-channel mudstone</td>
</tr>
<tr>
<td>3.4 Mixed color mudstone</td>
<td>-</td>
<td>Clay to silt</td>
<td>Blocky greenish grey to purple</td>
<td>Thickness: a few dm</td>
<td>Red-ox variability on floodplain and in-channel mudstone</td>
</tr>
</tbody>
</table>
Table 3.2: Description, Interpretation, and Comparison of Facies Associations.

<table>
<thead>
<tr>
<th>Facies Association</th>
<th>Environment</th>
<th>Channel %</th>
<th>Sandstone% of Floodplain</th>
<th>Channel Thickness</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 Large vertically and laterally amalgamated sandy channel-floodplain</td>
<td>Proximal fluvial fan</td>
<td>85-100%</td>
<td>&gt;70%</td>
<td>10-35m</td>
</tr>
<tr>
<td>lithosomes</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>2 Laterally amalgamated channel-floodplain lithosomes with sandy to</td>
<td>Medial fluvial fan</td>
<td>40-70%</td>
<td>~50%</td>
<td>3-15m (57-70%</td>
</tr>
<tr>
<td>heterolithic accretion sets</td>
<td></td>
<td></td>
<td></td>
<td>decrease from FA1</td>
</tr>
<tr>
<td>3 Small isolated channel-floodplain lithosomes</td>
<td>Distal fluvial fan</td>
<td>10-30%</td>
<td>~30%</td>
<td>2-10 (30% decrease</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>from FA2)</td>
</tr>
</tbody>
</table>
3.13 Reference Cited


4.1 Abstract

Eocene-aged fluvial and lake deposits in the Sunnyside Delta Interval in the Nine Mile Canyon show lateral fluvial-lake interactions. River and floodplain deposits transition into an interbedded succession of river and floodplain deposits, deltaic deposits, and lake deposits across the 20-km wide study area in Nine Mile canyon. However, this transition is not gradual, but rather, river channel and floodplain facies laterally alternate with carbonate or mixed carbonate-siliciclastic deposits across distances of only a few hundred meters to a few kilometers. Furthermore, some mouth bar deposits consist of alternating carbonate grainstones and siliciclastic sandstones, and some abandoned channels are filled with dolomitic mudstones. These numerous transitions between depositional environments indicate a highly irregular shoreline, where fluvial and deltaic deposits build out locally at the active channel locations, and laminated dolomitic mudstones accumulate in protected embayments or abandoned channels, and lime grainstones where lake’s wave and current energy is high. We interpret these fluvial-lake interactions in Sunnyside Delta interval as a result of river avulsions and contemporaneous carbonate productions.

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4.2 Introduction

Vertical transitions between fluvial and lacustrine deposits are commonly interpreted as transgressive-regressive cycles linked to lake level changes (Astin, 1990; Abdul Aziz et al., 2003; Keighley et al., 2003; Olsen, 2009; Zhao et al., 2015), where fluvial deposits are linked to lowstands and lake carbonates to highstands in a reciprocal sedimentation relationship (Thrana and Talbot, 2006; Ford et al., 2016; Rahiminejad et al., 2018). Such interpretation is driven by the sequence stratigraphy concept and considers the shut-off or dilution of carbonate production when or where siliciclastic input is high (Handford and Loucks, 1993). Although this assumption has been demonstrated in outcrop studies including the well-known Capitan margin of the Permian Delaware Basin (Harris and Saller, 1999; Kerans and Tinker, 1999), it does not apply in many modern coastal and shallow marine environments where siliciclastic and carbonate sediments are coevally mixed, resulting in lateral transitions between carbonate and fluvial sediments (Chiarella et al., 2017). The carbonate factory in a closed lake is even less sensitive to siliciclastic input, as chemical precipitation occurs when the lake water is supersaturated (James and Jones, 2016).

Outcrops of the Eocene Green River Formation Sunnyside Delta Interval in the Uinta Basin (Utah) display lateral fluvial and lake (carbonate) transitions across only a few hundreds of meters. For example, thinly laminated dolomitic mudstone-filled channels change laterally into fluvial channel sandstones, and/or interbedded siltstone and sandstones with paleosols. There are also examples of mouth-bar sandstones interbedded with carbonate grainstones and mudstones, and sandy carbonate grainstones that laterally pass into mouth bar sandstones. All of these lateral transitions caution the interpretation that vertical lake and fluvial transitions necessarily involve lake level changes as they are shown to have Walthorian relationships.

While growing amount of research has been conducted in modern and ancient systems (Halfar et al., 2004; McNeill et al., 2004; Quesne et al., 2009; Tanavsuu-Milkeviiciene et al., 2009; Zeller et al., 2015), since Mount (1984) first discussed the mixing processes in
modern shallow marine environment, not much data are available on the compositional and stratal mixing processes (Chiarella et al., 2017) in ramp settings of lacustrine basins (Song et al., 2017). This study aims to investigate the fluvial-lacustrine interaction and the mixing process by documenting the fluvial-lacustrine facies transition styles and lateral changes in mixed siliciclastic-carbonate facies. We discuss the fluvial avulsion processes as a significant control on the mixing processes.

4.3 Geological Setting

Uinta Basin formed in response to the Laramide uplift of the Uinta Mountains (Cashion, 1967; Dickinson et al., 1988; DeCelles, 2004). During the interfingering fluvial-lacustrine deposition of Green River Formation in early to middle Eocene between ca 53 to 46 Ma, Lake Uinta has been recognized as an underfilled lake (Picard, 1955; Fouch, 1975; Ryder et al., 1976; Castle, 1990; Fouch et al., 1994; Smith et al., 2003, 2008; Davis et al., 2010). It is featured a ramp style southern lake margin (Carroll and Bohacs, 1999) with shallow water depth for long distances. Mixed siliciclastic and carbonate deposits of Green River Formation developed as Lake Uinta experienced from fresh to transitional, rising to highlake, and closing lake stages, defined by organic richness (Cashion and Donnell, 1972, 1974; Johnson et al., 2010; Tanavsuu-Milkeviciene and Sarg, 2012; Tanavsuu-Milkeviciene and Sarg, 2017). Recent work on the siliciclastic part of Green River Formation in the Uinta Basin has mostly focused on the effects of variable discharge on rivers (Plink-Björklund and Birgenheier, 2013; Plink-Bjorklund et al., 2014; Plink-Björklund, 2015; Rosenberg et al., 2015; Gall et al., 2017), as well as on deltaic deposits (Jacob, 1969; Schomacker et al., 2010; Moore et al., 2012). On the carbonate part, lacustrine profundal and shoal deposits have been documented (Pitman et al., 1982; Smith et al., 2008; Burton et al., 2014). Mixed littoral to sublittoral carbonates and siliciclastics were recently documented in the eastern basin but the main focus was on vertical trend and climate (Gall et al., 2017) as well as lake level control (Tanavsuu-Milkeviciene and Sarg, 2017).
4.4 Stratigraphic Framework

The Eocene Green River Formation is subdivided into three informal units: lower, middle and upper, based on regionally extensive lacustrine intervals including the Uteland Butte Member, the Carbonate Marker Unit, and the Mahogany Oil Shale Member (Fig. 4.1) (Fouch, 1975; Pitman et al., 1982; Fouch et al., 1987). For the Nine Mile canyon area, which is located in the southwestern Uinta Basin, the three major siliciclastic intervals are the Colton Tongue, the Renegade Tongue, and the Sunnyside Delta Interval. D and C Marker beds mark the bottom and top of the Sunnyside Delta interval and can be followed across the study area in the Nine Mile Canyon (Remy, 1992; Morgan, 2003). D marker is located ca 160 m above the Carbonate Marker Unit (Picard and High, 1973; Ryder et al., 1976; Remy, 1992). C marker is equivalent to the C1 marker of Jacob (1969) and correlated to ‘R4’ zone of the eastern Uinta Basin, thus, Sunnyside Delta Interval is equivalent to Douglas Creek Member in the eastern Uinta Basin (Roberts, 1964; Tanavsuu-Milkeviciene and Sarg, 2017). The age of the Sunnyside Delta interval deposition is constrained by an absolute date range from the C marker of 49.6 Ma (Smith et al., 2015) and 54 Ma from the Carbonate Marker Unit (Remy, 1992) (Fig. 4.1).

4.5 Dataset and Methods

Field data from Sunnyside Delta interval (Fig. 4.1) was collected from the west-to-east oriented Nine Mile Canyon and several north-to-south oriented side canyons (Fig. 4.1). Observations start from mile marker 32 to 45 in Nine Mile Canyon, and continue to mile marker 49 to the south in Cottonwood Canyon. A total of about 550 m of measured section data were collected at 10 cm resolution. Acid testing was used to determine dolomite vs. limestone in the field. The Dunham classification modified by Embry and Klovan (1971) is applied in this study to include coarse grained carbonates. Thin sections were made in the Thin Section Laboratory (Department of Geology and Geological Engineering at Colorado School of Mine) to further observe mixed siliciclastic and carbonate facies. Lateral
relationships of sedimentary facies were examined on cliff-face exposures and mapped by walking out the stratigraphic intervals in continuous outcrops. Photomosaics were used to assist with analyzing channel architecture. Architectural dimensions (e.g. channel width and thickness) were obtained by using laser range finder (TRUPULSE™ 360B) in field.

4.6 Fluvial and Lacustrine Facies Associations

4.6.1 Fluvial Facies Association

4.6.1.1 Facies Association 1.1: Sandstone to Heterolithic Lenticular Lithosomes with Accretion Sets

Facies association 1 includes lenticular and laterally amalgamated sandstone to heterolithic erosionally based lithosomes, 10-35 m thick (Figs. 4.2A-C). Thicker lithosomes contain downstream and upstream facing accretion sets 10-15 m thick (Figs. 4.2A-C), whereas thinner lithosomes display vertical aggradation, a few to 10 m thick (Fig. 4.2D). Some lithosomes are sandy throughout, whereas others contain up to dm-thick mud layers in sand-mud couplets (Fig. 4.2D). This facies association consists primarily of very-fine to fine grained, moderately to poorly sorted sandstone with up to 10% of mudstone and up to 2% conglomerate. The latter occurs along the erosional base of channel lithosomes. Sedimentary structures are dominated by (over 70%) low-angle convex- and concave up sedimentary structures (F3, 5, 6, and 7; Figs. 4.2E-H; Table 4.1). Cross strata (F10-13) are the minor with less than 26% and commonly occur in climbing sets (Figs. 4.2I and J). Sandstone with scour and fill structures (F3; 30%), distinct planar lamina (F7; 20%), gradational planar lamina (F6; 14%), climbing ripple lamina (F13; 13%), ripple lamina (F12; 10%), and convex-up low angle lamina (F5; 8%) are the dominant facies (Fig. 4.2; Table 4.1). Most lithosomes do not display obvious vertical grain size trends, but some upstream accretion sets coarsen upward in places because mudstone layers in accretion sets thicken downward. Basal and internal erosion surfaces are concave up and flat or undulating, and in places aligned with
mudstone clasts. Accretion sets are commonly erosionally bound (Figs. 4.2A-D). Wood fragments are associated with erosion surfaces in places. In places bioturbation occurs at accretion set boundaries, including cm to dm long roots and vertical burrows (Figs. 4.2K and L). Desiccation cracks occur in places at the mud-to-sand contact within the accretion sets (Fig. 4.2M). This facies association is commonly associated with interbedded mudstones and sandstones of FA1.2 (Figs. 4.2A-D), but in places directly overlies micritic or carbonate cemented siltstones of FA3.2.

Interpretation: The erosionally bound lenticular accretion sets are interpreted as channels with a variable degree of lateral and vertical amalgamation. Sandstone facies indicate dominant Froude supercritical flow, as seen by comparison of the sedimentary structures with experimentally produced supercritical flow deposits (e.g. Alexander et al., 2001; Cartigny et al., 2014). In channel bioturbation and desiccation indicates that channels were seasonally or inter-annually dry, such as is common in arid to sub-humid climates (Sneh, 1983; Stear, 1985). Abundance or dominance of Froude supercritical flow sedimentary structures is characteristic for variable discharge rivers, where deposition dominantly occurs during high magnitude floods (Fielding, 2006; Fielding et al., 2009; Plink-Björklund, 2015). In channel mud layers suggest a rapid deceleration of floods with mud deposition during waning stages (e.g., Jones, 1977; Tunbridge, 1981). The erosionally bound sandy accretion sets and the sandstone-mudstone couplets are interpreted as flood event beds (Tunbridge, 1981; Sneh, 1983; Turner, 1986; Abdullatif, 1989; Deluca and Eriksson, 1989; Nichols and Hirst, 1998; Shukla et al., 2001; Hinds et al., 2004; Hampton and Horton, 2007; Chakraborty and Ghosh, 2010; Chakraborty et al., 2010; Mader and Redfern, 2011; Donselaar et al., 2013; Plink-Björklund, 2015). Despite the seasonal or inter-annual desiccation, the water table was relatively high as the vertical burrows are less than 1 dm long (Hasiotis, 2004).
4.6.1.2 Facies Association 1.2: Interbedded Mudstones and Sandstones with Variegated Color

Facies association 1.2 consists of interbedded mudstones and sandstones and is purple (F15), greenish grey (F16), brown (F17), or shows color transitions from purple to greenish grey (F18) on cm to dm scale. In places irregular brownish or grey patches occur. Mudstones are dm-thick (F15-18) and interbedded with ripple laminated (F12) or structureless (F8) very-fine grained sandstones, cm to a m thick. In places mudstones coarsen upward into sandstones (Fig. 4.2N). The interbedded mudstones and sandstones commonly form upward sandying and thickening packages, erosionally overlain and laterally truncated by lenticular or amalgamated accretion sets of FA 1.1 (Figs. 4.2A-D). The lateral extent of interbedded sandstones and mudstones varies from only 10s of meters where severely erosionally truncated to 100s of m or kms. Bioturbation and soft sediment deformation are common in the purple mudstone (F15; Fig. 4.2O). Possible mammal tracks occur in a 3 dm thick heavily bioturbated purple mudstone layer (Fig. 4.2O). Calcite filled cracks (Fig. 4.2P) and calcite nodules are common in all mudstone facies (F15-18).

Interpretation: The mottled color, burrowing and tracks, and carbonate nodules are characteristic for paleosols and subaerial exposure (e.g. Retallack, 1988; Leckie et al., 1989; Wright, 1992). Some calcite filled cracks are possible calcified root structures. Mottling is a result of local redox changes due to fluctuating drainage conditions and is also reflected in the purple to brown colors (Wright, 1992). The paleosols and the association with channels of FA 1.1 indicate that the interbedded mudstones and sandstones are floodplain deposits. The sandstone interlayers are interpreted as crevasse splays, and the upward coarsening and thickening trends with capping channel sandstones as avulsion packages (Mohrig et al., 2000; Jones and Hajek, 2007; Sendziak, 2012). The upward coarsening and sandstone thickening trends overlain by channels indicate transitional avulsions (Jones and Hajek, 2007). The splay rich floodplain indicates frequent avulsions (Shukla et al., 2001; Hajek and Wolinsky, 2012).
4.6.2 Lacustrine Facies Associations

4.6.2.1 Facies Association 2.1: Lime Grainstone

Ooids and ostracods are the dominant grains (F23) in the lime grainstone and they are commonly well sorted (Fig. 4.3A). Fragments of bivalve (Fig. 4.3C), gastropod (Fig. 4.3C), and fish scales (Fig. 4.3D) are present only locally in the lower part of the 10-50 cm thick grainstone layers (Fig. 4.3H) and form a floatstone or rudstone (F24). Grainstone layers are laterally consistent and commonly continuous for 10s to 100s of m (Fig. 4.3I), but vary in allochem content and layer thickness. In places ripple laminations with round and symmetrical crests occur (Figs. 4.3F and G). In places, cm long desiccation cracks occur at the top of the beds (Fig. 4.3G). In outcrop, this facies association is white-tan to yellowish orange in color and forms erosion-resistant layers.

Interpretation: Strong waves or currents roll grains in shallow littoral zone near the lake shoreline to form ooids (Ryder et al., 1976). Chemical carbonate precipitation allows the accumulation of coated grains, such as ooids, oncolites and pisoliths (Cole and Picard, 1978). Fragmented bivalve, gastropod, and fish scales indicate that they are redistributed by storm waves or currents (Milroy and Wright, 2002). The abundance of ooids and ostracods and lack of carbonate mud suggest an agitated environment where wave energy is high (Platt and Wright, 1991). The well sorted grains inherently prohibit the formation of sedimentary structures and explain the poorly developed ripple laminations.

4.6.2.2 Facies Association 2.2: Stromatolites

Stromatolite boundstone (F27) has hemispherical heads, cm to m tall (Figs. 4.4C and D), with internal mm thick convex-up lamination (Fig. 4.4E). Stromatolites (F27) are commonly yellowish in color and change laterally into oolite and pisolith/oncolite grainstone (F23; Fig. 4.3B), and stromatolite rudstone (F25). This transition exclusively occurs in the D marker (Fig. 4.1C) that is mapped across the Nine Mile Canyon. Inverse grading from ooids to
pisolites and oncolites is common in dm thick grainstones (F23; Figs. 4.3E and H).

Interpretation: The internally laminated stromatolites are interpreted to form in the lower littoral to sublittoral zone (Platt and Wright, 1991; Bridge and Demicco, 2008). The yellowish color has been shown to indicate dolomite content (Symcox, 2015). The lateral transition between stromatolite (F26) and stromatolite rudstone (F25) and oolite and pisolite/oncolite grainstone (F23) is indicative of high hydrodynamic energy including storm events in littoral and sublittoral zone of lake (Krylov, 1982; Cohen and Thouin, 1987; Suriamin, 2010; Sarg et al., 2013). The inverse grading from ooids to pisolites and oncolites indicates a shoaling upward succession as wave energy increases shoreward (Platt and Wright, 1991).

4.6.2.3 Facies Association 2.3: Gastropod Wackestone

Gastropods-bearing fine grained structureless micrite (F22) occur as up to 2 dm thick, laterally continuous, sharply bound, dark resistant beds (Figs. 4.4A and B). These wackestones are commonly vertically associated with oolitic or ostracodal grainstone (F23) and oolitic or ostracodal sandstone (F26) (Fig. 4.4A).

Interpretation: The occurrence of mudstone matrix in gastropod wackestone suggests deposition in a lower littoral to sublittoral zone or lagoon where the environment is calm (Renaut and Gierlowski-Kordesch, 2010).

4.6.3 Deltaic Facies Association

4.6.3.1 Facies Association 3.1: Sharp-based Tabular Sandstone

The sharp-based tabular sandstone is 5 to 10 m thick, and 100s of meter wide (Figs. 4.5A and B). It consists of fine to upper-fine grained moderately sorted sandstone dominated by gradational planar lamina (F6; 33%), scour and fill structures (F3; 31%), cross strata (F10; 12%), soft sediment deformation (F9; 9%), convex-up low angle lamina (F5; 8%), and climbing ripple cross lamina (F13; 7%). Convex- and concave-up laminated sandstones (F3, 5, and 6) form 72% of the succession and climbing and aggradational bedforms occur (F3, 5,
6, 9, and 13) in 88% of the deposits. The tabular sandstones do not display grain size trends (Fig. 4.5), but wood and carbonate fragments occur in places. Some sandstones display low-angle downstream accretion sets. The characteristic sharp but flat basal surface changes in places laterally into a concave-up erosion surface that bounds a lenticular sandstone (Figs. 4.5D and E). Multiple yellow layers of mm thick silicilastic mudstone form dewatering flame structures. Vertically this facies association overlies interbedded mudstone and sandstone of FA3.2 (Figs. 4.5A and B) or lacustrine carbonate or mixed facies associations (FA2.1-2.4; FA4.1-4.3) (Fig. 4.5D). This facies association transitions laterally into fluvial (FA1.1-1.2), and lacustrine and mixed facies associations (FA2.1-2.4; FA4.1-4.3).

Interpretation: The tabular geometry together with downstream accretion, high deposition rate sedimentary structures, and the association with lacustrine deposits suggests deposition in mouth bars (Scruton, 1960; Plink-Björklund and Steel, 2005; Olariu and Bhattacharya, 2006). The lateral transition of the sharp based tabular sandstone to lenticular shaped sandstone with concave-up erosion surface suggests mouth bar transition into its adjacent distributary channel (Olariu and Bhattacharya, 2006). Sedimentary structures indicate high deposition rates and abundant Froude supercritical flow.

4.6.3.2 Facies Association 3.2: Interbedded Siliciclastic and Carbonate Mudstone and Sandstone

This facies association consists of cm to 1 m thick very-fine grained sandstones (F12) and greenish grey or dark grey, and in places limy mudstones (F19) (Figs. 4.5A-C). Individual beds commonly fine upward from ripple laminated sandstone (F12) to thinly laminated mudstone (F19) (Fig. 4.5C). Dewatering structures and Scoyenia trace fossil are preserved at mudstone to sandstone boundaries. The interbedded sandstones and mudstones form upward coarsening packages where sandstone thickness in successive beds increases upward (Fig. 4.5C) or the mudstones occur as a dm to m thick layer sharply overlain by FA 3.1, and in places interbedded with carbonate facies associations (FA2.1 and 2.3) (Figs. 4.5A,
B). Vertically, interbedded mudstone and sandstone is overlain by sharp-based tabular sandstone (FA3.1), or amalgamated to lenticular lithosomes of FA1.1, or variegated interbedded mudstone and sandstone of FA1.2 (Fig. 4.5A).

Interpretation: The Scoyenia traces, the gray color, and the lack of pedogenic modification indicate a subaqueous lake environment (Hasiotis, 2004). The upward coarsening deposits of interbedded mudstone and sandstone beds are interpreted as delta front deposits. The fining upward trend of individual beds is characteristic for turbidites that are likely of hyperpycnal flow origin (Plink-Björklund and Steel, 2004; Bhattacharya and MacEachern, 2009) where mudstones are truncated.

4.6.4 Fluvial-lacustrine Mixed Facies Associations

4.6.4.1 Facies Association 4.1: Mixed Siliciclastic-Carbonate Accretion Sets

This facies association appears orange brown in outcrop and occurs as sharp-based accretion sets, dm to m thick and 10s of meter wide (Figs. 4.7 and 4.8). It is composed of oolitic sandstones (F26), oolitic and ostracod grainstones (F23), siliciclastic sandstones (F3, 6, 9, and 10), and limy mudstones (F19) (Figs. 4.7A and B). Oolitic sandstones consist of poorly sorted lower fine to lower-medium grained siliciclastic sands (30-40%) and similar sized carbonate grains (ooids and/or ostracods) (F26; Fig. 4.8). Sedimentary structures in oolitic sandstones (F26; Fig. 4.8) include cross stratification (30%), gradational planar lamina (30%), scour and fill structures (20%), symmetrical ripples with sharp crests (10%), unidirectional asymmetric ripple laminations (8%), and soft sediment deformation (2%). Carbonate grains are either evenly distributed through the tabular mixed unit (Fig. 4.7B) or align with lamina (Fig. 4.8). Siliciclastic sandstones commonly display scour and fill structures (F3), gradational planar lamina (F6), structureless (F9), and soft sediment deformation (F10). Dewatering and loading structures are characteristic at the boundary with siliciclastic sandstones (Figs. 4.7E and 4.8D). The limy mudstone layer (F19) between the siliciclastic sandstone (F3, 6, 9, and 10) and oolitic sandstone is thinly laminated (F26) (Fig. 4.7). Cross
strata foresets strata dip NNE similar to the accretion sets. A typical vertical transition (Fig. 4.7B) starts with a couple of dm thick carbonate grainstone (FA2.1) or stromatolite (FA2.2) interbedded with calcareous mudstone (FA3.2), then top truncated by sharp-based tabular sandstone (FA3.1) or mixed siliciclastic-carbonate accretion sets (FA4.1).

Interpretation: The mixed accretion sets are interpreted as a mixed carbonate-siliciclastic mouth bar complex, based on the occurrence as basinward-dipping accretion sets, the similarity of siliciclastic sandstones to mouth bar sandstones on FA 3.1, and the similar vertical transitions of this FA to FA 3.1 (Fig. 4.7A). The poor sorting suggests the source of ooids comes from a contemporaneous carbonate production nearshore and in-situ mixing with sand (Mount, 1984; McNeill et al., 2004). Fluvial-derived siliciclastic sands may have been mixed with carbonate grainstones due to lateral mixing, as the allochems were transported along the lake shoreline by waves or currents. The dewatering and loading features are indicative of high deposition rate (Lowe and LoPiccolo, 1996). The presence of symmetrical wave ripples suggests wave reworking (Leckie, 2003).

4.6.4.2 Facies Association 4.2: Sigmoidal Sandy Grainstone Macroforms

Sigmoidal sandy grainstone macroforms consists of light brown moderately sorted upper fine to lower medium grained siliciclastic sands and slightly larger sized carbonate grains (ooids and/or ostracods; >85%). The deposit coarsens upward with larger size carbonate grains on top and more siliciclastic sand at the bottom. The macroform is 2-3m thick and exhibits tabular shape with flat and sharp lower and upper boundaries (Fig. 4.9). Lateral extent of the macroform is ca 400 m, and it consists of sigmoidally or erosionally bound cross strata 1.5-2 m thick with topset, foreset, and bottomset geometries in flow parallel view (Fig. 4.9). Foresets dip obliquely perpendicular (SW230°) as compared to fluvial sediment transport direction at an angle of 15-25°. In flow perpendicular view, the bedding looks planar. The macroform is capped by a 20 cm thick grainstone layer (FA2.1), and it lies on top of a meter thick greenish grey laminated dolomitic carbonate mudstone bed (FA2.3).
Interpretation: The sigmoidal sandy grainstone macroform is interpreted as formed by punctuated mixing of large amount of carbonate grains with detrital sand (Mount, 1984). The migration direction suggests shore-parallel transport and storm-generated currents, which may occur during periodic high energy storms (Mount, 1984; Halfar et al., 2004).

4.6.4.3 Facies Association 4.3: Mixed Siliciclastic-carbonate Lenticular Mudstone

The lenticular mudstones occur as lenticular erosionally based deposits (Fig. 4.10) that consists of limy mudstone (F19), argillaceous dolomitic mudstone (F21), and in places also interbeds of oolitic grainstone (F23), oolitic sandstone (F26), and limy mudstones (F19) (Fig. 4.10C). The argellaceous dolomitic mudstone has cm-scale laminations and contains detrital clay as well as dolomite (F21). It is light/dark grey altered to dark brown (Fig. 4.11). The limy and dolomitic mudstones (F19 and 21) fill dm-scale scours or multi-m scale channels (Fig. 4.10), and display internal scour surfaces. In some places wood fragments occur at these scour surfaces. This facies association is laterally and vertically associated with mouth bars of FA 3.1 and oolitic sandstone accretion sets (FA4.1).

Interpretation: The occurrence of dolomitic mudstones, interbedded mudstones, siliciclastic sandstones, and carbonate grainstones in channels, the lateral association with mouth bars and distributary channel deposits, and the presence of wood fragments suggest deposition in abandoned channels. The alternation of color and the laminations are associated with variation of organic matter content, which is possibly associated with algal blooms (Renaut and Gierlowski-Kordesch, 2010). Such dolomitic mudstones and organic rich marl deposits (Platt and Wright, 1991) tend to occur in nearshore protected lagoons or embayments (Teller et al., 2000), and could thus also accumulate in channels abandoned by avulsions, as such channels would form areas protected from waves. Internal erosion surfaces, woody fragments, occasional siliciclastic sandstone and argillaceous layers suggest that the abandoned channels transferred discharge and minor amount of sediment at times. The grainstone interbeds
indicate transient current or wave action.

### 4.6.4.4 Facies Association 4.4: Laminated Dolomitic Mudstone

Dolomitic mudstone forms thinly laminated beds, mm to cm thick (Fig. 4.11). It is internally black but weathers into bluish grey color (F20; Figs. 4.11A and B). In places, there are sand-filled burrows on top of the beds. The laminated dolomitic mudstone layer extends laterally for kilometers.

Interpretation: The black mm thick dolomitic mudstone is interpreted as organic rich marl deposits (Platt and Wright, 1991). The presence of bioturbation and the km scale extent suggests deposition in a nearshore protected lagoon or embayment (Teller et al., 2000).

### 4.7 Lateral and Vertical Trends

In general, the ca 200 m thick Sunnyside Delta Interval across the 20 km of Nine Mile Canyon exposures displays a distinct change from southeast to northwest, as the southeasternmost outcrops only contain fluvial deposits (FA1.1. and 1.2), and the proportion of lacustrine facies increases gradually towards northwest (Fig. 4.12). First thin lime grainstone (FA2.1) and dolomitic mudstone (FA4.4) intervals alternate with the thick packages of fluvial deposits (Figs. 4.12B and C), but eventually the lacustrine deposits and deltaic deposits form a significant part of the stratigraphy (Fig. 4.11D).

This transition from river to lake deposits is not a simple gradual change, but rather, this transition is complex, as lateral transitions between fluvial (FA1), lacustrine (FA2), deltaic (FA3), and mixed fluvial-lacustrine (FA4) deposits occur in 100s of m to km scale (Figs. 4.12-4.14). For example, at mile marker 45 ca 35 m thick amalgamated sandy erosionally based fluvial channel deposit (FA1.1) overlying a dm thick gastropod wackestone (FA2.3) transitions westward into a mouth bar complex with dolomitic mudstone filled scours at the base (FA4.3), overlain by a mixed siliciclastic-carbonate mouth bar (FA4.1), floodplain
deposits (FA1.2), a lens of limey mudstone (FA3.2), and then a siliciclastic mouth bar (FA3.1) across 300 m (Fig. 4.13). The siliciclastic mouth bar is associated with underlying interbedded mudstone and sandstone (FA3.2) and mudstone with variegated color (FA1.2). Further northwest, this whole complex pinches out into a thick floodplain succession across 200 m (Fig. 4.13). In another example at mile marker 40.8, a mixed siliciclastic-carbonate mouth bar (FA4.1) that overlies oolitic or ostrocodal grainstone (FA2.1), transitions southeastward into floodplain deposits (FA1.2), where it is truncated by a river channel (FA1.1) towards northwest across 430 m (Fig. 4.13).

4.8 Discussion on Fluvial-lake Interaction

These numerous transitions between depositional environments indicate a highly irregular shoreline, where fluvial and deltaic deposits build out locally at the active channel locations, laminated dolomitic mudstones accumulate in protected embayments or abandoned channels, and lime grainstones, where lake wave and current energy is high. In a modern example, similar lateral transitions occur where Ile River fan enters Lake Balkhash (Fig. 4.14). Ili River is highly avulsive as it builds a large fluvial fan. Where the active river channel is, multiple deltaic lobes develop and prograde into the lake (Fig. 4.14A). At paleochannel locations (Figs. 4.14B and C), abandoned channels and different types of lagoons occur, as well as wave or current built bars or shoals. Thus the observed lateral variability in the Sunnyside Delta interval is interpreted to be a function of river avulsions and the contemporaneous carbonate grain and mud production. This mixing process along an irregular lake shoreline is also responsible for the mixed siliciclastic-carbonate mouth bars, sandy grainstone bars, and dolomitic mudstone filled channels.

This lateral variability further has a large impact on the nature of transgressions and regressions, as river and deltaic successions can locally prograde as seen by the lakeward perturbations, where the active channels are (Fig. 4.14D), whereas coeval transgression may occur laterally in areas without fluvial sediment input.
4.9 Conclusions

The Eocene Sunnyside Delta Interval transitions from river deposits to interbedded river, deltaic, and lake deposits across 20 km in the Nine Mile Canyon. This transition is not gradual but rather highly variable as river and floodplain deposits transition laterally into lake carbonates or deltaic deposits across a few hundred meters or a few kilometers. Some mouth bar deposits are mixed siliciclastic-carbonate and indicate coeval carbonate grain production and fluvial sediment input. Some abandoned channels are filled with dolomitic mudstones. River and delta lobe avulsions and the contemporaneous carbonate production are interpreted to control carbonate-siliciclastic mixing along a highly irregular shoreline.
Figure 4.1 Stratigraphic column and base map of Uinta Basin and Nine Mile canyon. (A) Geological map of the Wasatch/Colton (maroon) and Green River (orange) Formations in the Uinta Basin, with study area locations at Nine Mile Canyon, Road 191 and Hay Canyon (modified from Dickinson et al., 2012; Schomacker et al., 2010; Sato and Chan, 2015; Jones, 2017). (B) Study area at Nine Mile Canyon (see location in A) with figure locations and mile markers (grey numbers). Rose diagrams show paleocurrent directions measured in this study. (C) Stratigraphic table with available age constraints (modified from Fouch et al., 1987; Remy, 1992; Smith et al., 2010, 2015).
Figure 4.2 Example outcrop photos of fluvial facies association 1. (A-D) Vertical transitions between channel (FA1.1) and floodplain (FA1.2) facies associations. (E) Sandstone with scour and fill structure (F3). (F) Sandstone with convex-up low angle lamina (F5). (G) Sandstone with gradational planar lamina and scour and fill structures (F6 and F3). (H) Sandstone with distinct planar lamina (F7). (I) Sandstone with ripple cross lamina (F12). (J) Sandstone with climbing ripple cross lamina (F13). (K) Roots in sandstone. (L) Vertical burrows in ripple laminated sandstone. (M) Desiccation cracks at the bottom of sandstone. (N) Upward coarsening floodplain mudstone (F15-18) to sandstone (F12) (FA1.2). (O) Mammal tracks preserved at the base of a sandstone overlying a purple mudstone. (P) Possible calcite filled cracks in purple mudstone (F15).
Figure 4.3 Outcrop photos of lime grainstone (FA2.1). (A) Well sorted oolitic grainstone (F23). (B) Pisoid and oncoid grainstone (F23). (C) Gastropods and bivalve shells and fragments in non-stromatolite bioclastic floatstone (F24). (D) Fish scale fragments in non-stromatolite bioclastic floatstone (F24). (E) Inverse grading from ooids to pisolites and oncolites is common in dm thick grainstone beds (F23). (F-G) Sharp crested symmetrical ripple lamina. Note the cm long desiccation cracks at the upper boundary of rippled oolitic grainstone (F23) in G. (H) dm thick oolitic grainstone (F23) overlain by mixed siliciclastic-carbonate mouth bars (FA4.1).
Figure 4.4 Outcrop photos of stromatolite (FA2.2) and gastropod wackestone (FA2.3) facies associations. (A) A 20 cm thick gastropod bearing wackestone bed (F22) overlying ostracodal grainstone (F23). (B) A gastropod in FA2.3 or F22. (C) D marker with laterally linked stromatolites. (D) A few dm tall stromatolite layer in D marker. (E) Internal laminations in stromatolites in D marker. (F) Fragments of mm scale stromatolites in stromatolite rudstone (F25).
Figure 4.5 Outcrop photos of the deltaic facies association 3. (A, B) Characteristic vertical transition from delta front (FA3.2) interbedded sandstones and mudstones to mouth bar and distributary complex (FA3.1) amalgamated sandstones. (C) Upward fining succession (marked by white triangle) from sandstone (F12) to mudstone beds (F19) in delta front (FA3.2). (D-F) Photos and measured sections of sharp based mouth bar and distributary complex (FA3.1). Photos in D and E are 30 m apart.
Figure 4.6 Vertical trends of facies associations. Two styles are presented in A and B with a typical measured section and an outcrop photo. (A) Fluvial channel and floodplain (FA1.1 and 1.2) deposits are on top of deltaic (FA3.1 and 3.2) and lacustrine (FA2.2) deposits. The measured section is obtained from the outcrop in the figure. (B) Floodplain deposits (FA1.2) vertically transition into lacustrine (FA2.1) and deltaic (FA3.2) deposits, which are overlain by mixed siliciclastic accretion sets (FA4.1). The measured section is not from the outcrop in the figure.
Figure 4.7 Outcrop photos of the mixed siliciclastic-carbonate accretion sets (FA4.1). (A) Photo and line drawing of accretion set architecture. (B) Evenly distributed carbonate grains. (C) Greenish grey mudstone (F19) below cross stratified oolitic sandstone (F26). (D) Cross stratified grainstone (F26). (E) Loading structure at the base of a sandstone. (F) Cross stratified and soft sediment deformed oolitic sandstone (F26).
Figure 4.8 Outcrop photos of sedimentary structures in oolitic sandstone (F26) in the mixed siliciclastic-carbonate accretion sets (FA4.1). (A) Symmetrical ripples with sharp crests. (B) Gradational planar lamination. (C) Scour and fill structures. (D) Soft sediment deformation.
Figure 4.9 Outcrop photos of sigmoidal sandy grainstone macroform (FA4.2). The outcrop is continuous in west-to-east orientation. (A) is located to the east of (B). Note the composing bedforms and red lines represent their boundaries. An overview of the outcrop is provided in (C). This deposits laterally transition into mouth bar deposits across ca 200m to northwest.
Figure 4.10 Outcrop photos of mixed siliciclastic-carbonate lenticular mudstone (FA4.3). (A, B) Erosionally based lenticular mudstones with internal erosion surfaces. (C, D) Erosionally based lenticular interbedded mudstones, siliciclastic sandstones and carbonate grainstones.
Figure 4.11 Outcrop photos of laminated dolomitic mudstone (FA4.4). (A, B) The characteristic bluish gray color of dolomitic mudstone (F20). (B) Sand-filled horizontal burrows on top of a bed. (C) Dark brown color altered with light grey color in argillaceous dolomitic mudstone (F21).
Figure 4.12 Depositional trends across the Nine Mile Canyon. (A) Channel and floodplain deposits. (B, C) Channel and floodplain deposits with thin lake interbeds. (D) Interbedded and laterally variable lake and fluvial deposits.
Figure 4.13 Outcrop examples showing lateral and vertical transitions. (A) Lateral and vertical trends at mile marker 44.5 to 45 outcrop oriented in west (left) to east (right). (B) Lateral trends at mile marker 40.8 outcrop oriented in NW330° to the left. A mixed siliclastic-carbonate mouth bar (FA4.1) that overlies oolitic or ostrocodal grainstone (FA 2.1), transitions southeastward into floodplain deposits (FA 1.2) and is truncated by a river channel (FA1.1) towards northwest across 430 m.
Figure 4.14 Examples of lake-fluvial interaction in Ile River, Lake Balkhash. (A) Actively prograding delta lobes with an interlobe lagoon. (B) Multiple styles of lagoons, abandoned channels, and wave or current reworking of sediment. (C) Abandoned with standing water. Note how river and floodplain deposits are laterally associated with the lake. (D) Delta perturbations along the Lake Balkash shoreline.
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<thead>
<tr>
<th>Facies</th>
<th>Textures</th>
<th>Structures</th>
<th>Width/Thickness</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 Stratified and imbricated</td>
<td>Conglomerates (coarse sand to pebble sized carbonate clast/bioclast and/or mud clast, subrounded to angular, poorly sorted, varied in shape) in sandy matrix, clast or matrix supported</td>
<td>Imbricated, planar stratified, low angle stratified, scour and fill structures,</td>
<td>Thickness: a few cm per conglomerate layer; Total thickness: a few dm - m</td>
<td>Intrabasinal clasts of floodplain mudstones or lake carbonates deposited with sand matrix by traction flow (subcritical to supercritical)</td>
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<td>conglomerates</td>
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<td>2 Disorganized</td>
<td>Conglomerates (coarse sand to pebble sized carbonate clast/bioclast and/or mud clast, rounded to angular, poorly sorted, varied in shape) in sandy or muddy matrix, in places with terrestrial organic matter (woody fragments)</td>
<td>Disorganized</td>
<td>Thickness: a few dm - 1m</td>
<td>Intrabasinal clasts of floodplain mudstones or lake carbonate deposited by collapse and fluidization of channel banks during falling stage of flow</td>
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<td>3 Sandstone with scour and</td>
<td>Upper very fine to lower medium grained sandstone, moderately to poorly sorted, in places mixed with carbonate grains</td>
<td>Scours filled with upward flattening lamina</td>
<td>Thickness: a few dm - a few m; Width: a few dm - 10s of m</td>
<td>Supercritical flow, high deposition rates, suspension deposition, large antidune or chute and pool formation</td>
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<td>fill structures</td>
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<td>4 Sandstone with ripple</td>
<td>Upper very fine to lower fine-grained sandstone, moderately sorted</td>
<td>Ripples fill the toe part of upward flattening scour and fill lamina</td>
<td>Thickness: A few dm - 1m; Width: A few dm - a few m</td>
<td>Supercritical to subcritical flow, high deposition rates, suspension and bedload deposition</td>
</tr>
<tr>
<td>filled scour and fill</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>structures</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
### Table 4.1: Continued.

<table>
<thead>
<tr>
<th>Facies</th>
<th>Textures</th>
<th>Structures</th>
<th>Width/Thickness</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>5</td>
<td>Sandstone with convex-up low angle lamina</td>
<td>Very fine to fine grained sandstone, moderately to poorly sorted, in places mixed with carbonate grains</td>
<td>Low angle convex-up lamina, very low angle hummocky-like cross strata with upward increasing steepness</td>
<td>Thickness: A few dm - 1m; Width: &gt;1m</td>
</tr>
<tr>
<td>6</td>
<td>Sandstone with gradational planar lamina</td>
<td>Very fine to fine grained sandstone, moderately to poorly sorted</td>
<td>Diffused, gradational planar lamina, each lamina is internally reversely to normally graded</td>
<td>Thickness (lamina): 2mm - 2cm; Thickness (bed): a few dm - a few m</td>
</tr>
<tr>
<td>7</td>
<td>Sandstone with distinct planar lamina</td>
<td>Very fine to lower medium grained sandstone, moderately sorted</td>
<td>Planar laminations with distinct lamina boundaries</td>
<td>Thickness (lamina): mm scale; Thickness (bed): a few dm - a few m</td>
</tr>
<tr>
<td>8</td>
<td>Structureless sandstone</td>
<td>fine to lower medium grained sandstone, moderately to poorly sorted</td>
<td>No apparent sedimentary structures</td>
<td>Thickness: a few cm - a few dm</td>
</tr>
<tr>
<td>Facies</td>
<td>Textures</td>
<td>Structures</td>
<td>Width/Thickness</td>
<td>Interpretation</td>
</tr>
<tr>
<td>---------------------------------------</td>
<td>-----------------------------------------------</td>
<td>-------------------------------------------------</td>
<td>-----------------</td>
<td>----------------------------------------------------</td>
</tr>
<tr>
<td>9  Soft sediment deformed sandstone</td>
<td>Very fine to lower medium grained sandstone, moderately to poorly sorted</td>
<td>Flame structures, overturned folds, dish structures, ball and pillow structures.</td>
<td>Thickness: a few dm</td>
<td>Water escape or local collapse</td>
</tr>
<tr>
<td>10 Cross stratified sandstone</td>
<td>Fine to lower medium grained sandstone, moderately sorted</td>
<td>Planar and trough cross stratified</td>
<td>Thickness (cross set): 5cm - 20cm; Thickness (co-set): a few dm</td>
<td>Subcritical flow, normal deposition rates, bedload deposition, dune migration</td>
</tr>
<tr>
<td>11 Sandstone with climbing cross strata</td>
<td>Fine to lower medium grained sandstone, moderately to poorly sorted</td>
<td>Planar and trough cross strata with climbing set boundaries</td>
<td>Thickness (cross set): 8cm - dm scale; Thickness (co-set): a few dm</td>
<td>Subcritical flow, high deposition rates, suspension and bedload deposition, dune migration and aggradation</td>
</tr>
<tr>
<td>12 Cross laminated sandstone</td>
<td>Very fine to lower fine grained sandstone, moderately sorted</td>
<td>Asymmetrical cross-lamination</td>
<td>Thickness (cross set): cm scale &lt; 5cm; Thickness (co-set): a few dm</td>
<td>Subcritical flow, normal deposition rates, bedload deposition, ripple migration</td>
</tr>
<tr>
<td>13 Sandstone with climbing ripple lamina</td>
<td>Very fine to fine grained sandstone, moderately sorted</td>
<td>Asymmetrical cross lamination with climbing set boundaries</td>
<td>Thickness (cross set): cm scale &lt; 5cm; Thickness (co-set): a few dm</td>
<td>Subcritical flow, high deposition rates, suspension and bedload deposition, ripple migration and aggradation</td>
</tr>
</tbody>
</table>
Table 4.1: Continued.

<table>
<thead>
<tr>
<th>Facies</th>
<th>Textures</th>
<th>Structures</th>
<th>Width/Thickness</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>14 Bioturbated sandstone</td>
<td>Very fine to lower medium grained sandstone, moderately to poorly sorted</td>
<td>Vertical, horizontal and 3D trace fossils</td>
<td>Thickness: a few dm - a few m</td>
<td>Trace fossils formed by insects, dwelling, resting, crawling traces</td>
</tr>
<tr>
<td>15 Purpled red mudstone</td>
<td>Clay to silt</td>
<td>Laminated or blocky or crumbly, in places rooted, bioturbated, or deformed</td>
<td>Thickness: a few cm - a few dm (blocky); a few mm (each lamina)</td>
<td>Oxidized, well drained floodplain and in-channel mudstone</td>
</tr>
<tr>
<td>16 Greenish grey mudstone</td>
<td>Clay to silt</td>
<td>Laminated or blocky or crumbly</td>
<td>Thickness: a few cm - a few dm (blocky); a few mm (each lamina)</td>
<td>Reduced to poorly drained floodplain and in-channel mudstone</td>
</tr>
<tr>
<td>17 Brown to dark grey mudstone</td>
<td>Clay to silt</td>
<td>Laminated</td>
<td></td>
<td>Oxidized poorly drained floodplain and in-channel mudstone</td>
</tr>
<tr>
<td>18 Varigated color mudstone</td>
<td>Clay to silt, Blocky greenish grey to purple</td>
<td></td>
<td>Thickness: a few dm</td>
<td>Red-ox variability on floodplain</td>
</tr>
<tr>
<td>Facies</td>
<td>Textures</td>
<td>Structures</td>
<td>Width/Thickness</td>
<td>Interpretation</td>
</tr>
<tr>
<td>---------------------------------------</td>
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<td>-----------------</td>
<td>---------------------------------------</td>
</tr>
<tr>
<td>19 Limy mudstone</td>
<td>Detrital clay to silt, lime mud</td>
<td>Structureless, some faint lamina, commonly interbedded with ripple laminated sandstone (F12), occur in tabular, laterally extensive beds.</td>
<td>Thickness: a few dm - 1m</td>
<td>Siliciclastic derived, deposited in lake</td>
</tr>
<tr>
<td>20 Oraganic rich dolomitic mudstone</td>
<td>Lime mud, organic rich</td>
<td>Structureless, thinly laminated</td>
<td>Thickness: a few mm</td>
<td>Nearshore protected lagoon or embayment, low energy</td>
</tr>
<tr>
<td>21 Argillaceous dolomitic mudstone</td>
<td>Dolomitic mudstone with detrital clay</td>
<td>Planar laminated</td>
<td>Thickness (lamina): a few cm; Thickness (bed): a few dm</td>
<td>Sublittoral</td>
</tr>
<tr>
<td>22 Gastropod wackestone</td>
<td>Lime mud, gastropods</td>
<td>Structureless</td>
<td>Thickness: a few cm - 2dm</td>
<td>Lower littoral to sublittoral</td>
</tr>
<tr>
<td>23 Oolitic or ostracodal grainstone</td>
<td>Ooids and ostracods dominated, pisoids and oncolites also present in D marker</td>
<td>Symmetrical ripple lamina, structureless, planar lamina, inverse grading in places,</td>
<td>Thickness: 10-50 cm; Width: laterally continuous for 10s - 100s of m</td>
<td>Shallow littoral, high energy</td>
</tr>
<tr>
<td>Facies</td>
<td>Textures</td>
<td>Structures</td>
<td>Width/Thickness</td>
<td>Interpretation</td>
</tr>
<tr>
<td>---------------------------------------</td>
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<td>-------------------------------------</td>
</tr>
<tr>
<td>24 Non-stromatolite bioclastic floatstone or rudstone</td>
<td>Ca 30% fragmented bivalves, gastropods, fish scales, muddy intraclasts, size of the clasts are larger than 2mm</td>
<td>Structureless</td>
<td>Thickness: a few cm - 1dm</td>
<td>Lower littoral to sublittoral, high energy</td>
</tr>
<tr>
<td>25 Stromatolite rudstone</td>
<td>Stromatolite fragments, muddy intraclasts, elongate,</td>
<td>structureless</td>
<td>Thickness: dm</td>
<td>Restricted lake</td>
</tr>
<tr>
<td>26 Oolitic or ostracodal sandstone</td>
<td>Poorly sorted, ooids or ostracods, upper fine to lower medium grained siliciclastic sand</td>
<td>Symmetrical ripple lamina, asymmetrical ripple lamina, gradational planar lamina, scour and fill structures, soft sediment deformation</td>
<td>Total thickness: up to 1.5m; Thickness per layer: ca 50cm</td>
<td>Littoral, high energy</td>
</tr>
<tr>
<td>27 Stromatolite</td>
<td>Hemispheroidal microbialites</td>
<td>Internal laminated</td>
<td>Thickness: cm - m; Thickness (lamina): mm</td>
<td>Littoral, biogenic</td>
</tr>
</tbody>
</table>
Table 4.2: Description and Interpretation of Sedimentary Facies Association

<table>
<thead>
<tr>
<th>Environment</th>
<th>Facies Association</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fluvial</td>
<td>1.1 Amalgamated and isolated channel lithosomes</td>
<td>Fluvial channel</td>
</tr>
<tr>
<td></td>
<td>1.2 Purple, greenish grey, brownish mudstone</td>
<td>Floodplain</td>
</tr>
<tr>
<td>Lacustrine</td>
<td>2.1 Lime grainstone</td>
<td>Littoral, high energy</td>
</tr>
<tr>
<td></td>
<td>2.2 Stromatolite</td>
<td>Lower littoral to sublittoral (restricted lake)</td>
</tr>
<tr>
<td></td>
<td>2.3 Gastropod wackestone</td>
<td>Sublittoral, lagoon, low energy</td>
</tr>
<tr>
<td>Delta</td>
<td>3.1 Amalgamated</td>
<td>Distributary channel-mouth bar complex</td>
</tr>
<tr>
<td></td>
<td>3.2 Thinly interbedded mudstone and sandstone</td>
<td>Delta front</td>
</tr>
<tr>
<td>Mixed fluvial-lacustrine</td>
<td>4.1 Mixed siliciclastic-carbonate accretion sets</td>
<td>Carbonate shoal mixed with siliciclastic grains; wave reworked mouth bars</td>
</tr>
<tr>
<td></td>
<td>4.2 Sigmoidal sandy grainstone macroform</td>
<td>Stormed induced carbonate shoal</td>
</tr>
<tr>
<td></td>
<td>4.3 Mixed siliciclastic-carbonate lenticular mudstone</td>
<td>Siliciclastic derived mudstone deposited in calm lake environment</td>
</tr>
<tr>
<td></td>
<td>4.4 Laminated dolomitic mudstone</td>
<td>Nearshore lagoon</td>
</tr>
</tbody>
</table>

4.10 Reference Cited


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CHAPTER 5
CONCLUSIONS

This dissertation uses high resolution outcrop data and documents channel architecture in a fluvial fan system, proximal to distal changes within the fan, and the fan transition to the lake. This field-based study provides systematic descriptions of bar-scale macroforms, lateral and vertical facies transitions, to challenge the current understanding of fluvial facies models. The macroforms presented in Chapter 2 (1st paper) raise the awareness that the well-developed macroforms in variable discharge rivers are different from the ones in the classical fluvial models. In Chapter 3 (2nd paper), the detailed lateral and vertical transitions documented in a fluvial fan system significantly add to the simple proximal-distal and vertical trends published so far, and link channel lithosome types and floodplain sandyness to their position in the fan. This work challenges previous interpretations of the Sunnyside Delta interval of the middle Green River Formation and links this succession to variable discharge rivers and their ability to build fluvials fans. The results and the comparison to modern fans advance our understanding the large fluvial fan systems. Chapter 4 (3rd paper) documents the effects of avulsions and contemporaneous carbonate production, such as a highly irregular lake shoreline due to lateral transitions between fluvial progradation and lacustrine deposition. The findings caution the transgressive-regressive interpretations applied to vertical transitions between fluvial and lake deposits, and propose that such lateral variability has a large impact on the nature of transgression and regression.