PROGRADATIONAL SLOPE ARCHITECTURE AND SEDIMENT PARTITIONING IN THE
MIXED SILICICLASTIC-CARBONATE BONE SPRING FORMATION,
PERMIAN BASIN, WEST TEXAS

by

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Slope-building processes and sediment partitioning in mixed carbonate-siliciclastic sediment routing systems are poorly understood but are important constraints on the spatial and temporal distribution of reservoir-forming elements. The Bone Spring Formation, Delaware Basin, west Texas is a mixed carbonate-siliciclastic system that consists of cyclic slope-to-basin hemipelagites, turbidites, and debrites that were sourced from the Victorio Peak Formation carbonate shelf margin and Bone Spring Formation slope during Leonardian time (~275 Ma). Much research has focused on the basinal deposits of the Bone Spring Fm., but there has been little research on the proximal, upper slope segment of the Bone Spring sediment routing system. In this study, we constrain the stratigraphic architecture of the Bone Spring Fm. that outcrops in Guadalupe Mountains National Park in order to delineate the slope clinothem geometry and the dynamics of carbonate and siliciclastic sediment delivery to the basin. We record the outcropping Bone Spring Fm. upper-slope as composed predominantly (~90% of the study area) of fine-grained carbonate hemipelagites and sediment gravity flows containing a high biogenic silica content (i.e. chert). Interbedded within the carbonate slope facies at various scales are detrital terrigenous hemipelagic and sediment gravity flow deposits, carbonate mass-transport deposits, and carbonate submarine channel deposits. We identify ten slope-building clinothems that vary from siliciclastic-rich to carbonate-rich and show significant variability in slope propagation direction. Clinothems are truncated by slope detachment surfaces that record large-scale mass-wasting of the shelf margin and upper slope. X-ray fluorescence (XRF) data indicates that slope detachment surfaces contain a higher-than-normal proportion of terrigenous siliciclastic sediment, suggesting failure is triggered by accommodation or sediment supply changes at the
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CHAPTER 1

INTRODUCTION

The dynamics of continental margin evolution and sediment partitioning impact the spatial and temporal distribution of reservoir-forming elements (Saller et al., 1989; Bull et al., 2009; Playton et al., 2010; Janson et al., 2011; Stevenson et al., 2015; Hurd et al., 2016; Playton and Kerans, 2018) and record autogenic and allogenic processes acting on the system (Shanley and McCabe, 1994; Covault et al., 2006; Romans et al., 2008; Burgess, 2016; Madof et al., 2016). The importance of stratigraphic architecture and sediment partitioning in the evolution of continental margins has been documented in both siliciclastic (Kertznus and Kneller, 2009; Sylvester et al., 2012; Salazar et al., 2015; Stevenson et al., 2015; Prather et al., 2017) and carbonate margins (Bosellini, 1984; Sonnenfeld, 1991; Kerans et al., 1993; Ross et al., 1994; Sarg et al., 1999; Mulder et al., 2012). In particular, clinothems (surface-bounded packages of sediment) in both siliciclastic and carbonate systems record slope evolution and the variable distribution of lithologies (Rich, 1951; Mitchum et al., 1977; Vail, 1987; Sonnenfeld, 1991; Ross et al., 1994; Sarg et al., 1999; Playton et al., 2010; Salazar et al., 2015). While most studies of clinothems have focused on siliciclastic margins (Mitchum et al., 1977; Vail, 1987; Bull et al., 2009; Kertznus and Kneller, 2009; Sylvester et al., 2012; Salazar et al., 2015; Stevenson et al., 2015; Prather et al., 2017) or steep, reef-rimmed carbonate margins (Bosellini, 1984; Katz et al., 2010; Harman, 2011; Mulder et al., 2012; Jo et al., 2015; Principaud et al., 2015; Playton and Kerans, 2018), studies of low-relief, mixed siliciclastic-carbonate margins are less well
documented (Saller et al., 1989; James et al., 1991; Sonnenfeld, 1991; Fitchen, 1997; Grosheny et al., 2015; Tassy et al., 2015).

In the Permian-aged Delaware Basin of west Texas, the Victorio Peak (shelfal facies) and Bone Spring (slope, basinal facies) formations represent a low-relief, mixed siliciclastic-carbonate depositional system that has garnered interest in recent years as a prolific hydrocarbon system (Allen et al., 2013; Driskill et al., 2018; Schwartz et al., 2018). Studies in the Bone Spring Fm. have focused primarily on the basinal deposits, which record heterogeneity between siliciclastic and carbonate lithologies and a mixture of turbidites, mass-transport deposits, and hemipelagic-pelagic deposits (Saller et al., 1989; Montgomery, 1997a, 1997b; Asmus and Grammer, 2013; Nance and Rowe, 2015; Driskill et al., 2018). A few studies (Kirkby, 1982; Fitchen, 1997) have focused on the shelfal (Victorio Peak) deposits, documenting cyclical deposition of platform carbonates and bypassing siliciclastic sands. While both the proximal and distal portions of the sediment routing system have been documented, the upper slope segment of the Bone Spring Formation has only been partially documented (King, 1948; McDaniel and Pray, 1967; Kirkby, 1982; Fitchen, 1997).

This study constrains the progradational slope architecture and the sediment partitioning of the upper-slope Bone Spring deposits that are superbly exposed in Guadalupe Mountains National Park, west Texas. The objectives of this study are to document (1) slope-building clinothems of variable and mixed lithology, (2) slope detachment surfaces bounding clinothems, and (3) abundant mass-wasting deposits on the mixed slope. We use these results to speculate on slope evolution in a mixed-lithology margin, the role of mass-wasting and terrigenous sediment supply in shaping the margin and delivering sediment to the basin, and how the evolution of the
upper slope affects the depositional processes and stacking patterns of carbonate and siliciclastic sediment in the distal Delaware Basin.
CHAPTER 2

GEOLOGIC AND STRATIGRAPHIC SETTING

2.1 Geologic Setting

The Bone Spring Formation was deposited in the Delaware Basin, a sub-basin of the larger Permian Basin of west Texas during Leonardian time (middle Permian, ~275-280 Ma; Figure 2.1 inset). Preceding the Delaware Basin was the ancestral Tobosa Basin, a Cambrian to Mississippian sag basin (Hill, 1996; Asmus and Grammer, 2013). During the late-Mississippian assembly of the supercontinent Pangea (~326 Ma), the Permian Basin began to take shape as a foreland basin north of the Marathon-Ouachita-Sonora orogeny (Poole et al., 2005; Figure 2.1 inset). Compression reactivated Precambrian areas of weakness and uplifted the Central Basin Platform, creating two sub-basins of the Permian Basin, the Delaware and Midland Basins (Figure 2.1 inset; Hills, 1984; Hill, 1996; Amerman, 2009; Nance and Rowe, 2015). Tectonic activity occurred until at least the middle Wolfcampian (~295 Ma; Hills, 1984; Amerman, 2009) while the Leonardian period (~285 to 275 Ma) was generally a quiescent tectonic environment (Hills, 1984; Amerman, 2009). Subsidence related to sediment loading as well as isostatic adjustment on the basin margins created a deep (~450 meters, Hills 1984) basin where up to 2,500 meters of sediment accumulated in the Delaware Basin during the Permian (Hills, 1984). During Permian time, the Delaware Basin was bounded to the west and north by the Diablo Platform and Northwest Shelf, to the south by the Marathon-Ouachita-Sonora fold belt and
Hovey Channel, and to the east by the Central Basin Platform and San Simon and Sheffield Channels (Figure 2.1A inset; Asmus and Grammer, 2013).

Figure 2.1: Overview map of the Permian outcrops in and around Guadalupe Mountains National Park (GMNP), west Texas (modified from King, 1948). A) Geologic map of GMNP. Black box denotes Figure 2.1B location. White line A-A’ indicates location of cross-section in Figure 2.2. Inset map shows Permian Basin paleogeography with GMNP denoted as a red box along the western margin of the Delaware basin, blue line indicates Figure 8.4 seismic line; HV = Hovey Channel, SS = San Simon Channel, SH = Sheffield Channel. B) Study area focusing on Leonardian-aged outcrops. Red dashed line indicates 3D model shown in Figure 3.1, and red solid lines highlight interpreted outcrop exposures shown in Figures 5.1, 5.3, 5.5, and 5.7. Dotted tan line marks Shumard Trail.

Sediment routing into the Delaware Basin originated from the north and east (Soreghan and Soreghan, 2013) with some sediment input from the Marathon-Ouachita-Sonora region to the south (Hu et al., 2018; Soto-Kerans et al., 2018), where aeolian and fluvial processes delivered sediment to the shelf and shelf margin (Presley, 1987; Fisher and Sarnthein, 1988). During Leonardian time, especially during low sea-level conditions, the entrance to the open Panthalassa Ocean to the west was restricted by a sill in the Hovey Channel (Fitchen, 1997). This sill
restricted water circulation in the basin, leading to euxinic basin conditions (McDaniel and Pray, 1967), minimal bioturbation, and the preservation of organic-rich sediment (Hills, 1984).

2.2 Shelf-to-Basin Stratigraphy

The evolution of the shelf-margin and basinal strata that comprise the Delaware Basin are well documented by King (1948), Sarg and Lehman (1986), Kerans et al. (1993), Sarg et al. (1999), and Kerans and Kempter (2002). Figure 2.2 shows the chronostratigraphic and lithostratigraphic units within the Delaware Basin that have been correlated from shelf to basin. Representing the lower-most Permian-aged rock is the Wolfcamp Fm., a mixed carbonate-siliciclastic prograding shelf-to-basin system (Silver and Todd 1969; Kvale and Rahman, 2016) that does not outcrop in the study area (Figure 2.1A). Overlying the Wolfcamp Fm. are Leonardian-aged prograding carbonate banks to rimmed platforms with slopes of 5-20 degrees (Harris, 2000) that transition into a deep basin assemblage (Figure 2.2; Fitchen, 1997; Asmus and Grammer, 2013; Hurd et al., 2018). The Leonardian system is composed of the proximal Yeso Formation that represents a restricted shelf environment with aeolian red beds and evaporitic deposits (Stanesco 1991; Fitchen, 1997). The Yeso Formation transitions to the open platform Victorio Peak Formation, a carbonate grain-margin (Kirkby, 1982), that transitions to the Bone Spring Formation carbonate and siliciclastic slope and basin deposits (Figure 2.2; Saller et al., 1989; Montgomery, 1997a; Fitchen, 1997; see Figure 2.2). Fitchen (1997) described six third-order sequences within the Victorio Peak-Bone Spring margin (L1-L6; Figure 2.2), where sequence boundaries represent subaerial exposure on the shelf and coeval sand deposition in the basin. Within sequences are lowstand (sand-rich) and highstand to transgressive (carbonate-rich)
members thought to be representations of cyclicity in sea-level and basinal subsidence (Figure 2.2; Silver and Todd, 1969; Saller et al., 1989; Fitchen, 1997; Nance and Rowe, 2015). Within each of the sand-rich and carbonate-rich members, higher-order lithologic cyclicity exists in the Bone Spring basinal deposits (Montgomery, 1997a; Nance and Rowe, 2015; Driskill et al., 2018) and has been interpreted to represent allogenic high frequency sequence development (Nance and Rowe, 2015). A significant erosional surface separates the Victorio Peak and Bone Spring Formations from the overlying Cutoff Formation (Figure 2.2). This surface is known as the top of LD10 (Sarg et al., 1999) or top of L6 (Fitchen, 1997; Hurd et al., 2018). The Cutoff Formation began as a lowstand system that eroded parts of the Victorio Peak/Bone Spring margin before reaching a maximum transgression (L8/G1) that has been biostratigraphically correlated with the Leonardian to Guadalupian boundary (Hurd, 2016). Above the Cutoff Formation is the Guadalupian-aged Delaware Mountain Group, including the Brushy Canyon Formation (G5-G7), which consists of a submarine channel-fan system (Zelt and Rossen, 1995;1995; Gardner and Sonnenfeld, 1996; Gardner et al., 2008; Figure 2.2). Capping the succession are Guadalupian-aged reef-rimmed carbonate platforms (Capitan Formation) and their coeval basal deposits (Figure 2.2; Kerans et al., 1993; Harman, 2011).

Constraining the outcropping Victorio Peak and Bone Spring Formations in Guadalupe Mountains National Park (Figure 2.1) into this stratigraphic context is difficult because of paleo-erosional features (e.g. Cutoff Fm.) and poor-resolution biostratigraphy. Lithostratigraphic correlations from Fitchen (1997) suggest the Victorio Peak-Bone Spring outcrops represent the L5 and L6 shelf margin to upper slope sequences; this interpretation is supported by recent biostratigraphic and lithostratigraphic correlations in the Cutoff Fm. (Hurd et al., 2016). To complicate matters, correlating the Bone Spring Fm. outcrops to the basin can be difficult as
many industry naming schemes are purely lithostratigraphic (e.g., 1st Bone Spring Sand, 1st Bone Spring Carbonate, Avalon) and biostratigraphic age control is poor in the subsurface basinal deposits (Figure 2.2; Driskill et al., 2018; Hurd et al., 2018). Hurd et al. (2018) correlates the base of the outcropping Cutoff Fm. (base L7) to the base of the Upper Avalon Shale in the basin (Figure 2.2). Therefore, the Bone Spring Fm. outcrops in our study area likely correlate to basinal rocks referred to as the Middle Avalon Carbonate, Lower Avalon Shale (boundary between L5/L6; Sarg, 1988), and some portion of the 1st Bone Spring Carbonate (Figure 2.2).

Figure 2.2: Stratigraphic section (A-A’) of the west face of Guadalupe Mountains National Park (modified from Kerans and Kempter, 2002). This study focuses on the Bone Spring (upper slope) and Victorio Peak (outer shelf) L5 and L6 sequences (red box). Outcrop-defined sequences shown in the stratigraphic column to the left compiled from Fitchen 1997, Sarg et al., 1999, Kerans and Kempter, 2002, and Hurd et al., 2016. The stratigraphic section at right defines the basin terminology with inferred chronostratigraphic correlations to outcrops. Note the Bone Spring Fm. outcrops are interpreted to correlate to the basinal rocks referred to as the Middle Avalon Carbonate, Lower Avalon Shale, and some portion of the 1st Bone Spring Carbonate.
CHAPTER 3

STUDY AREA AND OUTCROP MAPPING

3.1 Study Area

The study area is focused along the ‘western escarpment’ of Guadalupe Mountains National Park (Figure 2.1B), a northward-trending footwall fault block that was created during Cenozoic extensional tectonism (Hills, 1984; Hill, 1997) and exposes Leonardian and Guadalupian-aged carbonate and siliciclastic shelf-margin stratigraphy (Figure 2.2; King, 1948; Hills, 1984; Harris, 1987). Post-depositional loading (Hills, 1984), Late-Cretaceous transpression in the Trans-Pecos region to the west (Montgomery, 1997a), and the Cenozoic-reactivated Huapache Monocline (Hayes, 1964; Resor and Flodin, 2010) contribute to a 2-4° eastward dip of Permian rocks along the western escarpment. King (1948) extensively mapped the National Park and the surrounding area, including the Bone Spring Formation that is well-exposed in a system of west-east trending canyons (Figure 2.1B). This study focuses primarily on the outcrops in Shumard and Bone Canyons, as well as the west-facing exposures between the canyons (Figure 3.1). Located at the entrance to Bone Canyon is the historic Williams Ranch House (Figure 2.1B), a blue-painted homestead built in 1908.
3.2 3D Outcrop Model

A 3-dimensional (3D) digital outcrop model was built using Agisoft software and over 2,000 drone-collected photographs (Figure 3.1). Using the existing stratigraphic framework (King, 1948; Kirkby, 1982; Sarg et al., 1999; Hurd et al., 2018) the study area was constrained below the Cutoff Fm. and down-dip of the lithostratigraphic boundary with the Lower Victorio Peak (Figure 3.1). Field observations from bedding-attitude transects (N=16 transects, n=593 bedding measurements), nine measured sections (Appendix A), and six photopanel interpretations (Figures 5.1 5.3, 5.5, 5.7, Appendix B) were incorporated into the 3D model to capture depositional elements, facies relationships, and prominent stratigraphic surfaces (Figure 3.1).

Figure 3.1: Stratigraphic Architecture of the outcropping Bone Spring Fm. A) Plan view of 3D model with dip data. Lithostratigraphy of Brushy Canyon, Cutoff, Upper Victorio Peak (UVP), and Lower Victorio Peak (LVP) Formations shown. B) 3D digital outcrop model of the stratigraphic architecture of the Bone Spring Fm. Depositional elements, lithology variability, stratigraphic surfaces, and dip direction displayed. Ten clinothems (orange numbers) are bounded by nine slope detachment surfaces (black lines and blue numbers)
Figure 3.1: continued
3.3 Stratigraphic Surface Nomenclature and Mapping

Prominent stratigraphic surfaces of various scales can be mapped throughout the outcrop (Figure 3.1). Surfaces are identified using bedding attitude changes and truncation/onlap relationships (Figure 3.1). Large-scale surfaces are defined as those with more than 20 m of truncation/onlap that can be mapped along the outcrop extent (kilometer-scale) before disappearing into the subsurface, coalescing with another surface, or transitioning northwestward into the Victorio Peak shelfal facies (black surfaces, blue numbers in Figure 3.1B). These large-scale surfaces are interpreted as slope detachment surfaces (SDS) that may be associated with larger scale clinoform geometries. Some SDS show roll-over in the study area (SDS 4, 7, 8, 9; cf. Rich, 1951), suggesting these SDS have a clinoforming shape, much like the clinoforms mapped in the Leonardian shelf margin by Sarg (1988) and Fitchen (1997). However, because of the limited exposure of the full geometries of many of the surfaces, we refer to them herein as SDS. We identify nine SDS and ten intervening clinothems (defined as the strata bounded by SDS) within the study area (orange numbers, Figure 3.1B). Smaller-scale surfaces are defined as those with less than 20 m of truncation/onlap (red lines in Figure 3.1B). We interpret these smaller-scale surfaces as mass-wasting scars, and they occur within every clinothem (Figure 3.1B).
CHAPTER 4

SEDIMENTARY FACIES AND DEPOSITIONAL ENVIRONMENTS

4.1 Facies

4.1.1 Naming Schemes

Facies naming schemes can be difficult in mixed carbonate-siliciclastic systems because of confusion between textural/compositional facies schemes (e.g., Dunham, 1962; Folk, 1980) and schemes that define facies based on depositional process (e.g., Bouma, 1962; Lowe, 1982; Hubbard et al., 2008). In recent years, efforts have been made to name mixed sediments focusing on the mudstone dominated environments (Milliken, 2014; Lazar et al. 2015; Driskill et al., 2018; Thompson et al., 2018). These naming conventions are useful in the basin setting but break down as one moves up-dip along the sediment routing system into the classical carbonate realm (e.g., a transition from calcareous siltstone deposited by sediment gravity flow to a lime wackestone deposited on a carbonate platform). For the purposes of this paper, we developed a system-scale facies scheme that is valid all along the sediment routing system (i.e. from shelf to basin) and can move along a continuum from carbonate-rich to siliciclastic-rich deposits (Figure 4.1). In our scheme, we use the historical naming convention with the highest constituent component. That is, if carbonate makes up $>50\%$ of the sediment, we use the Dunham classification scheme (Dunham, 1962), and if siliciclastic and argillaceous sediment make up $>50\%$ of the sediment we use the Folk classification scheme (Folk, 1954; Folk, 1980). To further
clarify the composition of the facies, we add a modifier if a secondary constituent makes up greater than 10% of the sediment (Chiarella and Longhitano, 2012; Lazar et al., 2015). We also include a siliciclastic-carbonate ratio (s/c in Table 4.1) to quantify compositional variability (Chiarella and Longhitano, 2012). Because most facies within the Bone Spring Formation have primary sedimentary structures, we also add a modifier (e.g., ‘laminated’) to the facies name to differentiate depositional process.

4.1.2 Facies Descriptions

Eight facies were identified based on composition, grain size, depositional process, bed thickness, sedimentary structures, and fossil content. In addition to field-scale observations, facies were constrained by thin section analysis, scanning electron microscope (SEM) analysis, and X-ray fluorescence (XRF) data (Figures 4.1 and 4.2, Table 4.1, Appendix C). The eight facies are listed below, with the dominant interpreted depositional process in parentheses based on observations outlined in Table 4.1: 1) thin-bedded laminated lime mudstone (hemipelagite/sediment gravity flow deposit); 2) thin to thick-bedded deformed lime mudstone (mass-transport deposit); 3) thick-bedded bioclastic lime wackestone to packstone (hemipelagite/sediment gravity flow deposit); 4) interbedded lime mudstone and bioclastic packstone (interbedded hemipelagites and turbidites); 5) thick-bedded normally-graded bioclastic lime packstone to grainstone (turbidites); 6) thin-bedded laminated bioclast quartz siltstone (hemipelagites and turbidites); 7) thin-bedded laminated quartz lime mudstone (hemipelagites and turbidites); and 8) thick-bedded bioclastic lime packstone to grainstone (in-place shallow-water carbonate platform deposits to reworked platform deposits).
Facies in the Bone Spring outcrops show a high degree of mixing of siliciclastic and carbonate sediment (Figure 4.1B). Biogenic silica (i.e. chert) is abundant throughout facies and is differentiated from detrital siliciclastic by a lack of clay content (cf. Driskill et al., 2018). Depositional processes also vary considerably between facies (Figure 4.1A). The primary facies on the upper slope are carbonate-dominant mudstone to wackestone facies (Facies 1, 2, 3, 4, Figure 4.1A) interpreted as hemipelagic and sediment gravity flow processes on the slope. Facies 1 and 2 represent a continuum of deformation on the slope ranging from undeformed (Facies 1) to highly deformed (Facies 2). Carbonate slope deposits are interbedded in places with coarser-grained sediment gravity flow deposits (Facies 4 and 5) interpreted as calciturbidites. The carbonate-dominant facies (e.g., Facies 1) occur along a facies continuum with siliciclastic-dominant Facies 6, with Facies 7 representing a medial position on the carbonate-siliciclastic continuum (Figure 4.1C). This continuum represents variable carbonate and siliciclastic compositional mixing on the shelf and slope during transport (Chiarella et al., 2017). Facies 8 represents the Lower Victorio Peak of King (1948) and Kirkby (1982).

4.2 Depositional Environments

Four facies associations in the outcropping Bone Spring Formation are interpreted to represent sub-environments within the mixed-lithology, shelf-slope depositional system. (Figures 4.3, 4.4).

4.2.1 Facies Association 1 Description

Facies Association 1 (FA1), includes F1, F3, and F8, with a predictable stacking pattern shown in Figure 4.3A. F1, F3, and F8 always occur in stratigraphic sequence, with F1 at the
Figure 4.1: Facies analysis of Bone Spring Fm. deposits. A) Facies diagram displaying eight facies with generalized XRF readings. Facies are grouped based on composition and depositional process. B) Ternary diagram displaying XRF data color-coded by carbonate- mixed- or siliciclastic- dominant facies (blue, orange, yellow, respectively). C) Schematic of naming scheme used in this study with facies projected.
Figure 4.2: Facies pictures from outcrop (upper photo) and thin section (lower photo). A) Facies 1 thin-bedded laminated lime mudstone. Pencil is marking ripples. Thin section of Facies 1 is predominantly lime mudstone, but detrital quartz grains are present. B) Facies 2, thin to thick-bedded deformed lime mudstone with lines indicating deformation. Thin section of Facies 2 with deformation-induced calcite-cemented fractures with background facies identical to Facies 1. C) Facies 3, thick-bedded bioclastic lime wackestone to packstone. Thin section of Facies 3 shows an increase in mud content to the top interpreted as possible turbidity current. D) Facies 4, interbedded lime mudstone and bioclastic packstone with interbedded packstone indicated. Thin section shows interbedded packstone beds with calcite cementation and lenticular to continuous nature. E) Facies 5, thick-bedded normal-graded bioclastic lime packstone to grainstone. Normal grading shown with finger placed on basal coarse-grain deposit. Thin section shows bryozoan (by), brachiopods (ba), and undifferentiated carbonate allochems with chert cement. F) Facies 6, thin-bedded laminated bioclastic quartz siltstone. Note different color and weathering pattern to Facies 1. Thin section shows noticeably higher detrital quartz present in comparison to Facies 1. G) Facies 7, thin-bedded laminated quartz lime mudstone. Interbedded with Facies 6 showing different weathering pattern. Note brown color in comparison to Facies 1. Thin section of Facies 7 with detrital quartz content less than Facies 6 but greater than Facies 1. H) Facies 8, thick-bedded bioclastic lime packstone to grainstone. Thin section of Facies 8 reveals bryozoan (by), sponge spicules (sp), rugose corals (co), brachiopods (ba), and unidentified carbonate material.
Table 4.1. Summary of descriptions and interpretations of lithofacies

<table>
<thead>
<tr>
<th>Facies</th>
<th>Basin Naming Scheme (Lazar et al., 2015)</th>
<th>Mud Content silt/clay type</th>
<th>Coarse-grain % size type</th>
<th>Si/Ca ratio (s/c)</th>
<th>Sedimentary Structures</th>
<th>Diagenetic Features</th>
<th>Depositional Process</th>
<th>Depositional Environment</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>F1</strong> thin-bedded laminated limy mudstone (Figure 5A)</td>
<td>laminated calcareous siltstone</td>
<td>75-80%</td>
<td>20-25%</td>
<td>08/92 s/c&lt;1</td>
<td>&gt;99% planar laminations (millimetre beds)</td>
<td>chert beds mostly planar (5-10 cm thick) or nodular, occurring every 10-20 cm</td>
<td>Hemipelagic and sediment gravity flow</td>
<td>Hemipelagic deposits on the upper-konormal slope with sediment gravity flow common</td>
</tr>
<tr>
<td></td>
<td>60/40 silt/clay</td>
<td></td>
<td></td>
<td></td>
<td>some evidence of soft-sediment deformation</td>
<td>minor dolomitization</td>
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<td></td>
<td>carbonate clay, argillaceous clay,</td>
<td></td>
<td></td>
<td></td>
<td>some bioturbation</td>
<td>pyrite formation</td>
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<td></td>
<td>carbonate grains, gravel, quartz,</td>
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<tr>
<td></td>
<td>organic matter, pyrite</td>
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<tr>
<td><strong>F2</strong> thin to thick-bedded deformed limy mudstone (Figure 5B)</td>
<td>deformed calcareous siltstone</td>
<td>75-80%</td>
<td>20-25%</td>
<td>08/92 s/c&lt;1</td>
<td>Planar laminations</td>
<td>chert beds mimic bedding structure and can be folded, deformed, nodular, or planar</td>
<td>Hemipelagic deposition with soft sediment deformation, slope creep, sediment gravity flow</td>
<td>Hemipelagic and sediment gravity flow slope deposits that have undergone deformation from high slope angles or high sediment supply</td>
</tr>
<tr>
<td></td>
<td>60/40 silt/clay</td>
<td></td>
<td></td>
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<td>soft-sediment deformation: folding, fractures, fluid-escape, decollement surfaces</td>
<td>Calcite-filled fractures</td>
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<tr>
<td></td>
<td>carbonate clay, argillaceous clay,</td>
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<td></td>
<td>carbonate grains, gravel, quartz,</td>
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<tr>
<td><strong>F3</strong> thick-bedded bioclastic lime wackestone to packstone (Figure 5C)</td>
<td>thick-bedded bioclastic lime siltstone to very fine sandstone</td>
<td>30%</td>
<td>70%</td>
<td>5/95 s/c&lt;1</td>
<td>Grading</td>
<td>chert beds mostly nodular (5-10 cm thick), occurring every 30-40 cm</td>
<td>Hemipelagic and sediment gravity flow</td>
<td>Hemipelagic slope deposits that are proximal to the shelf margin relative to F1 and F2 and have experienced reworking from tidal, storm, or current forces</td>
</tr>
<tr>
<td></td>
<td>70/30 silt/clay</td>
<td></td>
<td></td>
<td></td>
<td>Laminations</td>
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<tr>
<td></td>
<td>carb. and arg. clay, carbonate grains,</td>
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<td></td>
<td>Some in-place production</td>
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<td></td>
<td>quartz, organic matter, pyrite</td>
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<tr>
<td></td>
<td>50% mud</td>
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<tr>
<td></td>
<td>50% sparite</td>
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Table 4.1 continued.

<table>
<thead>
<tr>
<th>Facies</th>
<th>Basin Naming Scheme (Lazar et al., 2015)</th>
<th>Mud Content</th>
<th>Coarse-grain % size type</th>
<th>Si/Ca ratio (s/c)</th>
<th>Sedimentary Structures</th>
<th>Diagenetic Features</th>
<th>Depositional Process</th>
<th>Depositional Environment</th>
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</thead>
<tbody>
<tr>
<td>F4</td>
<td>interbedded lime mudstone and bioclastic packstone (Figure 5D)</td>
<td>interbedded calcareous siltstone and bioclastic very fine sandstone</td>
<td>70%</td>
<td>Same as Facies 1</td>
<td>30% vfs</td>
<td>08/92 s/c&lt;&lt;1</td>
<td>Packstone beds occur every 1-2 cm and are planar or continuous or lenticular</td>
<td>Significant dolomitization and calcite cement in packstone beds</td>
</tr>
<tr>
<td>F5</td>
<td>thick-bedded normal-graded bioclastic lime packstone to grainstone (Figure 5E)</td>
<td>normal-graded bioclastic calcareous sandstone</td>
<td>15% 0% carbonate mud 100% sparite silt-sized carbonate grains (shell frg, spicules) sparite can reach 30% in some cases</td>
<td>85% coarse (0.5-1mm in diameter) carbonate and biogenic grains, crinoids, peloids, spicules, brach., bryo., bivalves, shell fragments</td>
<td>0/100 s/c&lt;&lt;1</td>
<td>Some normal grading and grain size segregation</td>
<td>Siliceous cement (chert)</td>
<td>Sediment gravity flow (turbidity currents)</td>
</tr>
<tr>
<td>F6</td>
<td>thin-bedded laminated bioclastic quartz siltstone (Figure 5F)</td>
<td>laminated bioclastic-rich siliceous siltstone</td>
<td>75-85% 70/30 silt/clay argillaceous clay (Al- and K-rich), qtz, crinoids, peloids, shell frg. Minimal sparite</td>
<td>15-25% vfs quartz, carbonate grains</td>
<td>75/25 s/c&gt;1</td>
<td>Planar laminations (&quot;flaggy&quot; bedding) Ripples Scouring perpendicular to bedding</td>
<td>Iron oxidation and/or calcification of carbonate grains Minor dolomitization Chert absent</td>
<td>Hemipelagic and sediment gravity flow (turbidity currents)</td>
</tr>
<tr>
<td>Facies</td>
<td>Basin Naming Scheme (Lazar et al., 2015)</td>
<td>Mud Content silt/clay type</td>
<td>Coarse-grain % size type</td>
<td>Si/Ca ratio (s/c)</td>
<td>Sedimentary Structures</td>
<td>Diagenetic Features</td>
<td>Depositional Process</td>
<td>Depositional Environment</td>
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<tr>
<td>F7</td>
<td>thin-bedded laminated lime mudstone (Figure 5G)</td>
<td>laminated quartz-rich calcareous siltstone</td>
<td>60-70% 50% mud and silt 50% sparite</td>
<td>30-40% vfs some fs mostly qtz, some shell frg</td>
<td>45/55 s/c&lt;&lt;1</td>
<td>Planar laminations Scouring perpendicular to bedding Minor soft-sediment deformation</td>
<td>Chert beds mostly planar (5-10 cm thick) or nodular, occurring every 10-20 cm Minor dolomitization</td>
<td>Hemipelagic and sediment gravity flow (turbidity currents) Possibly aeolian settling</td>
</tr>
<tr>
<td>F8</td>
<td>thick-bedded bioclastic lime packstone to grainstone (Figure 5H)</td>
<td>bioclastic lime very fine sandstone to sandstone</td>
<td>20-30% 0% mud 100% sparite</td>
<td>70-80% coarse range: vfs-pebble crinoids, peloids, spicules, brach., bryozoan, sponges, bivalves, carbonate grains, shell fragments, sparite grains</td>
<td>0/100 s/c&lt;1</td>
<td>None observed (in-place) No siliciclastic observed</td>
<td>Chert beds mostly nodular (5-10 cm thick), occurring every 0.5-1 meter Dolomitization</td>
<td>Carbonate platform in situ growth; some tidal,storm, or current reworking</td>
</tr>
</tbody>
</table>
base, F3 in the middle, and F8 at the top (Figure 4.3A). Contacts between facies are gradational and can transition over several meters (Figure 4.3A).

4.2.2 Facies Association 1 Interpretation

FA1 represents an upward-shoaling carbonate slope environment, with increasing grain size, bed thickness, sparite and fossil content, and decreasing chert content from Facies 1 to Facies 8, representing a shoaling succession from slope mudstones and turbidites to platform carbonates (McDaniel and Pray, 1969). The upward-shoaling character of FA1 suggests the Leonardian carbonate margin was predominantly progradational in the study area.

4.2.3 Facies Association 2 Description

The primary facies association, Facies Association 2 (FA2), makes up roughly 90% of the study area and includes some mixture of F1, F2, and F4 (Figures 4.3B, 4.4A, B). The F1, F2, and F4 facies are interbedded and can be found transitioning laterally in a single bed-set (Figure 4.3B). Contacts between facies are predominantly gradational but can be sharp, with truncation below and onlap above the surface, particularly between F1 and F2 (Figure 4.3B, Figure 4.4B).

4.2.4 Facies Association 2 Interpretation

FA2 represents a carbonate slope environment with abundant mass failure. The sharp erosional surfaces found within FA2 are interpreted as slope failure scarps and/or erosional bypass surfaces (Figure 4.4B). The lack of coarse-grained material directly mantling these surfaces (Figure 4.4B) supports a failure scarp interpretation. These failure surfaces are often filled with a wedge architectural pattern that is interpreted to be the filling of local topography. The chaotic bedding and folded features found in FA2 are interpreted to be mass-
failure deposits on the slope (Figure 4.2B, 4.4B). This “failure-and-fill” architecture has been well-documented in other carbonate slopes (Bosellini, 1984; Ross et al., 1994; Katz et al., 2010; Playton et al., 2010; Mulder et al., 2012; Playton and Kerans, 2018) as a mechanism for slopes prograding and aggrading over its failed deposits.

4.2.5 Facies Association 3 Description

Facies Association 3 (FA3) makes up a minor proportion of the study area (less than 5%) and is composed of F1, F3, F5, and F8 (Figure 4.3C, 4.4C, D). The type locale of FA3 occurs on the south wall of Shumard Canyon (Figure 3.1), where a sharp surface with 10 m relief truncates F3, with F5 onlapping the surface (Figure 4.4C). The F5 deposit is a 100 m wide and 10 m thick lenticular deposit with a concave base and a flat top. Over a 10 m interval, F5 gradually transitions into F8 (Figure 4.3C). Other instances of FA3 (Figure 3.1, 4.3C, 4.4D) show similar architecture, but smaller dimensions (e.g., a 10 m wide and 0.5 m thick, lenticular F5 lying above a surface that truncates F1; Figure 4.4D). In some instances, Facies 5 beds offset stack, with fine-grained F1 draping previous deposits (Figure 4.4D).

4.2.6 Facies Association 3 Interpretation

FA3 is interpreted as submarine channel deposits developed in a carbonate slope setting. The erosional truncation of fine-grained facies (F1, F3) and overlying coarse-grained channel fill with normally graded beds (F5) indicates erosion and deposition by turbidity currents (Figure 4.1E, 4.4C, D; Talling et al., 2012; Janocko et al., 2013). The presence of amalgamation surfaces (Figure 4.4C) indicate multiple erosive events, suggesting that the channels were long-lived conduits for sediment to the deeper basin. The presence of F8 (Lower Victorio Peak) overlying the channel fill on the Shumard south wall outcrop suggests that this submarine channel was
located very near the shelf edge (Figure 4.3C). The smaller channel deposits in association with F1 are interpreted to lie in a mid-slope position and may represent slope gully fill (Figure 4.3C; Shumaker et al., 2016).

**4.2.7 Facies Association 4 Description**

Facies Association 4 (FA4) makes up 5 to 10% of the outcrop, mostly constrained to one area (Figure 3.1) and is composed of gradational interbedding of F6 and F7 (Figure 4.3D, 4.4E). The typical thickness of these interbeds of siliciclastic (F6) and mixed-lithology (F7) facies are ~10 cm (Figure 4.4E), but on the west wall of Shumard Canyon the thickness of F6 can reach 10 meters (Figure 3.1). Contacts between the F6 and F7 components of FA4 are typically sharp and undulatory (Figure 4.4E). Like FA1, FA4 hosts truncation surfaces of variable azimuth, with deformation of the units above the surface.

**4.2.8 Facies Association 4 Interpretation**

FA4 is interpreted as periods where more siliciclastic material was delivered to the outer carbonate bank and upper slope. We interpret that this terrigenous silt-rich sediment was deposited by hemipelagic and sediment gravity flow processes. The interbedded nature of F6 and F7 (Figure 4.4E) suggests a high-frequency cyclicity in siliciclastic and carbonate deposition. Deformed intervals and truncation surfaces suggest an unstable slope setting dominated by failure and bypass, similar to FA2 carbonate deposits. The increased detrital siliciclastic material differentiates FA4 from FA1, suggesting a change in sediment supply that affected the primary slope-building facies (FA1).
Figure 4.3: Facies Associations of the outcropping Bone Spring Fm. A) Facies Association 1: Upward-shoaling carbonate margin. Transition from Bone Spring Fm. (BS) to Victorio Peak Fm. (VP) facies indicated. B) Facies Association 2: Carbonate slope deposits with mass wasting. C) Facies Association 3: Submarine carbonate channel deposits. Mid-slope and shelf-edge settings shown. D) Facies Association 4: Upper-slope siliciclastic-dominant hemipelagic and sediment gravity flow deposits.
Figure 4.4: Photos of Facies Associations from the outcrop. A) Facies Association 1 (FA1): undeformed F1 prograding slope with planar chert beds (dark colored rock). B) Discordant surface within FA2; note truncation of F1, with F2 overlying the surface. C) Facies Association 3 (FA3): shelf-edge submarine channel deposit cutting into slope deposits (F1). Erosional surfaces shown in yellow. D) FA3: mid-slope submarine gully deposits show offset stacking and axis-to-margin fining. E) Facies Association 4 (FA4): Interbedding of siliciclastic-rich Facies 6 (fissile, grey, recessive) and mixed-composition Facies 7 (tan colored, more resistant) on the upper slope.
CHAPTER 5

STRATIGRAPHIC ARCHITECTURE

Six photopanels demonstrate the stratigraphic architecture of the Bone Spring Formation (Figure 2.1B, Appendix B.1). We discuss four photopanels below in detail (Shumard Canyon north, Shumard Canyon south, Bone Canyon north, and Bone Canyon south). The intervening areas (west wall Shumard, west wall Bone) were used to correlate between Shumard and Bone Canyons and provide additional stratigraphic context and are included in Appendix B.

5.1 Shumard Canyon

5.1.1 North Wall of Shumard Canyon

The north wall of Shumard Canyon represents the best-exposed transition between the Victorio Peak and Bone Spring Formations (Figure 5.1). Eastward and southward dipping Bone Spring outcrops (F1, F2, F3) comprise most of the north wall, with flat-lying Victorio Peak Formation (F8) making up the uppermost cliffs (Figure 5.1). Dip data show the Bone Spring Fm. slope built out predominantly in an easterly direction but varies in orientation from 060° to 180° (Figure 5.1, Figure 3.1A). Several areas of interest from the north wall of Shumard canyon are highlighted in Figure 5.2. In area A, SDS 3 (Figure 5.2A) spans the entire height of the outcrop (~40m relief). Bedding orientation changes significantly across the surface, shifting from 18/090 (dip magnitude/dip azimuth) below to 23/045 above. Above SDS 3, Clinothem 4 is characterized
by FA4, with siliciclastic content (F6) increasing up-section (Figure 5.2A). Siliciclastic-
dominant beds are truncated by SDS 4, a prominent truncation surface (Figure 5.2A) which has
~80 meters of visible relief and shows a bedding orientation change from 23/045 to 14/100.

Above SDS 4, F7 gradually transitions to F1, and FA2 characterizes Clinothem 5 and 6. Bedding
orientation also changes across SDS 6 (Figure 5.2B), with a 5-10 m thick MTD sitting directly
above the surface in Clinothem 7 (Figure 5.2B). Above SDS 7, Clinothem 8 is characterized by
FA2 but lacks a basal MTD (Figure 5.2C). However, Clinothem 8 contains many discordant
surfaces (red surfaces, Figure 5.1), one with 10-20 m of overlying F1 with a wedge geometry
(Figure 5.2D). A prominent dip azimuth shift from due east to due south also occurs in
Clinothem 8, and where this change occurs, there are several FA3 channel deposits (Figures
4.4D, 5.1).

5.1.2 South Wall of Shumard Canyon

SDS 3-8 and clinothem packages 4-9 can be traced from the north wall of Shumard
Canyon across the canyon floor to the south wall (Figure 5.3). The FA4-dominated Clinothem 4
continues across the canyon (Figure 5.4C), but sand beds are thinner (cm-scale) and more
interbedded with Facies 7 than the deposits on the north wall. Moving up-section, Clinothems 5-
7 are poorly exposed but the MTD in the basal part of Clinothem 7 on the north wall of Shumard
Canyon (Figure 5.2B) can be correlated across the canyon to the south wall (Figure 5.3). At this
locale, the MTD displays decollement surfaces and compressional deformation features (Figure
5.4B). Above SDS 7, the basal Clinothem 8 contains a ~ 10 m thick MTD (Figure 5.3, 5.4A),
which is not present on the north wall (Figure 5.1), suggesting significant lateral variability. FA2
is most common on the south wall, but this locale also contains the largest submarine channel
deposit (FA3) in the study area (Figure 5.3, Figure 4.4C). This channel deposit is located just below Victorio Peak facies in Clinothem 8.

5.2 Bone Canyon

5.2.1 North Wall of Bone Canyon

SDS 6-9 were correlated from Shumard Canyon to Bone Canyon (Figure 5.5). The clinothems on the north wall of Bone Canyon display fewer bedding orientation changes and mass wasting features (F2) than in Shumard Canyon. At the entrance to the canyon in area A (Figure 5.5) the FA4-dominant Clinothem 4 is present, but the sand-rich F6 becomes progressively more discontinuous from Shumard to Bone canyon (Figure 3.1, Figure 5.6A1, Appendix B.3). At the mouth of Bone Canyon, FA4 deposits in Clinothem 4 are offset by numerous normal faults (Figure 5.6A2) that are likely related to the primary Cenozoic escarpment-bounding fault (Figure 2.1). In Clinothem 8, a small-scale discordant surface truncates F1 beds and is traceable for only ~ 10 m laterally, with minimal dip attitude change across the surface (Figure 5.6C). Moving up stratigraphically, SDS 8 shows the same architectural elements as the SDS in Shumard Canyon, displaying dip attitude changes, truncation, and carbonate facies both above and below with MTDs dispersed on top (Figure 5.6B). In area D, SDS 9 is overlain by a 20-30 m wide by 1 m thick submarine channel deposit (FA3; Figure 5.6D).
5.2.2 South Wall of Bone Canyon

The south wall of Bone Canyon displays where the Cutoff Formation has eroded into the upper Bone Spring Formation (Figure 5.7; Hurd et al., 2016), with debrites (Hurd et al., 2016) of Victorio Peak Fm. lying on the contact. Area A highlights a large (~100 m wide x 20 m thick) wavy, deformed FA2 interval with localized thrust faults (Figure 5.7, 5.8A). Individual F1 beds can be traced through the entire feature and dip at greater than 40° in some places with some minor (~10 cm) faults. We interpret this unit of FA2 to be a slope failure deposit (Figure 5.7, 5.8A). SDS 8 can be traced across Bone Canyon from the north wall just beneath this MTD (Figure 5.7, 5.8B), where a 40° bedding orientation change occurs across the surface (Figure 5.8B). In area C, SDS 9 and several overlying smaller-scale discordant surfaces are identified on both canyon walls (Figure 5.8C). Facies 2 MTDs overlie the SDS 9 surface, and numerous bedding orientation changes indicate a failure-prone Clinothem 10 (Figure 5.8C).
Figure 5.1: Stratigraphic Architecture of the north wall of Shumard Canyon. SDS and clinothems labeled by blue and orange circles, respectively. Note the prominent dip-azimuth change in Clinothem 8, coincident with a concentration of mass wasting deposits and FA3 channel deposits. Numbered inset boxes correspond to Figure 5.2, where siliciclastic-dominant intervals (A), mass transport deposits (B), and discordant truncation surfaces (C, D) are highlighted. Arrow symbols represent dip direction (where North is up). Compensationally stacked channels shown in Figure 4.4D indicated in black. Location of XRF transects 1-3 shown in blue.
Figure 5.2: Architectural features visible on Shumard north wall. A) Prominent SDS (3 and 4) with dip attitude and lithology changes across surfaces. Above SDS 3, siliciclastic beds dominate and are truncated by SDS 4. Figure 6.2A and 6.1D indicated in black boxes. Arrow and numerical value represent dip azimuth and magnitude. B) SDS 6 with a 5 m thick MTD sitting directly above the surface. Note debrites (db) and packstone beds (pb) with some preserved strata internally in the MTD (white lines) with the healing phase topography above the MTD. Geologist for scale. C) SDS 7 with truncation and dip attitude change in Facies 1. D) Discordant surface (red) within Clinothem 8, with ~ 10-20 m of overlying Facies 1 with a wedge geometry. Note the geologist for scale.
Figure 5.3: Stratigraphic Architecture of the south wall of Shumard Canyon. SDS and clinothems labeled by blue and orange circles, respectively. The large submarine channel deposit is shown in blue (see Figure 4.4C for details). Inset boxes correspond to Figure 5.4, where a large MTD (A), a compressional fold domain of an MTD (B), and interbedded siliciclastic (F6) and carbonate (F7) deposits (C) are highlighted. Note that the Shumard trail passes directly through many of these architectural features.
Figure 5.4: Architectural features visible on Shumard south wall. A) Large MTD overlying SDS 7. Figure 6.1C indicated in black box. B) MTD overlying SDS 6 shows multiple detachment surfaces (red) separating folded and faulted Facies 1 deposits. Note geologist for scale. C) FA4 in Clinothem 4 consists of interbedded siliciclastic (F6) and carbonate (F7) deposits.
Figure 5.5: Stratigraphic Architecture of the north wall of Bone Canyon. The Cutoff Fm. here has been eroded from the overlying Brushy Canyon Fm. channel. SDS and clinothems labeled by blue and orange circles, respectively. Numbered inset boxes correspond to Figure 5.6, where sand beds and normal faults (A), SDS 8 (B), a 5-10-meter discordant surface (C), and a calciturbidite deposit (FA3) sitting on top of SDS 9 are highlighted. Note vertical fractures in the eastern side of the outcrop.
Figure 5.6: Architectural features visible on Bone north wall. A1) Interbedding of siliciclastic and carbonate beds (FA4) in Clinothem 4. A2) Post-depositional faulting. Faulting shown here is small and mostly antithetic to the primary escarpment-bounding fault to the west of the outcrop (Figure 2.1 in Study Area). Geologist in circle. B) SDS 8 showing dip attitude change, truncation, and similar facies on either side of the surface. Location of XRF transect 4 (Figure 8.2D) shown. Symbols represent XRF readings below (circle), along (square), and above (triangle) surface. C) Example of small-scale discordant surface in red. Surface is on the meter-scale, cuts through only 1-2 beds, with minimal dip change across the surface. D) SDS 9 with a calciturbidite (FA3) sitting above the surface and a dip attitude shift from below to above. Location of XRF transect 5 (Figure 8.2E) shown. Backpack indicated by circle.
Figure 5.7: Stratigraphic Architecture of the south wall of Bone Canyon. SDS and clinothems labeled by blue and orange circles, respectively. Numbered inset boxes correspond to Figure 5.8, where a large mass transport deposit (A), bedding orientation change across SDS 8 (B), and SDS 9 with related mass wasting features (C) are highlighted. Figure 6.2B indicated. Note the Cutoff Fm. eroding into the Bone Spring Fm. (Hurd et al., 2016) and the overlying Brushy Canyon Fm. (Gardner et al., 2008).
Figure 5.8: Architectural features visible on Bone south wall. A) Large mass transport deposit that shows minimal internal deformation other than minor folding and a soft-sediment thrust fault shown in green. B) A 40° dip azimuth change occurs across SDS 8, with FA2 both above and below the surface. The surface itself dips at 29/050. C) SDS 9 surface with overlying deformed Facies 2 MTDs, and other small-scale discordant surfaces (red surfaces) with variable F1 and F2. Note the Cutoff Fm. contact just above this surface, and the geologist for scale.
CHAPTER 6

FAILURE AND DEFORMATION IN THE BONE SPRING FM.

The Bone Spring Fm. outcrops provide an opportunity to observe failure and deformation in a mixed siliciclastic-carbonate slope environment, which has been understudied in comparison to purely siliciclastic margins (cf. Moscardelli and Wood, 2016). Carbonate MTDs can act as barriers, baffles, source, and reservoirs in the Delaware Basin and elsewhere (Saller et al., 1989; Allen et al. 2013; Asmus and Grammer, 2013; Thompson et al., 2017; Bhatnager et al., 2018; Driskill et al., 2018). Therefore, an improved understanding of failure and deformation from the outcrop can lead to better identification in core and well-logs, resulting in better reservoir characterization parameters such as scale, heterogeneity, reservoir and mechanical properties, and potential compartmentalization.

6.1 Scale of Failure and Deformation

Failure and intrastratal deformation occur at many scales on the Bone Spring Fm. upper slope. Most commonly, intrastratal deformation occurs on the micro-scale, typically acting on individual lamina (< 1 cm) within individual 5-20 cm thick beds (Figure 6.1A1, A2). Micro-scale failure and intrastratal deformation is common within carbonate mudstone facies (F1, F2), including slumping, water-escape, folding, imbricate stacking (cf. Auchter et al., 2016), convolute bedding, micro-faults, and detachment surfaces (Figure 6.1 A1, A2). Soft-sediment deformation of similar geometry is also found on the meso-scale (1-20 meters, Figure 6.1B, C, D) and macro-scale (>20 meters, Figure 5.8A). At the meso-scale, failure surfaces (red surfaces
in Figure 5.2D, Figure 5.6C, Figure 3.1) likely represent detachments for slope-attached failures (Moscardelli and Wood, 2008). Meso-scale MTDs are common on the slope and consist of meter-scale carbonate F1 and F2 slump and debris flow deposits (Figures 6.1B, C, 5.4B) and siliciclastic slope facies (F6 and F7; Figure 6.1D).

At the macro-scale, slope detachment surfaces (Figure 3.1) can be correlated the length of the study area (>1 km) and display minimum visible relief of 20-100 meters. These surfaces are marked by truncation, bedding orientation changes across the surface, and a lack of karsting or other evidence of subaerial exposure. We interpret these surfaces to represent the basal detachment for subaqueous mass-failures of the shelf-margin and slope that create zones of sediment evacuation (Bosellini, 1984; Bull et al., 2009; Mazzanti and De Blasio, 2010; Janson et al., 2011; Mulder et al., 2012; Principaud et al., 2015). Headwall scarps from sediment evacuation on carbonate slopes often have steep angles (Mulder et al., 2012; Jo et al., 2015; Principaud et al., 2015), and slope detachment surfaces in the study area have structurally-restored dips of 15-25 degrees. The lack of macro-scale MTDs in the study area suggests that most MTDs were sourced from this steep (~15 degrees) upper slope locale and deposited more distally; indeed, MTDs have been documented in the Delaware Basin at the toe-of-slope and in the deep basin (Saller et al., 1989; Montgomery, 1997a; Nance and Rowe, 2015; Schwartz et al., 2018). This MTD slope segmentation, with large-scale MTDs sourced in the upper slope, bypassing the slope, and being deposited at the toe-of-slope or further into the basin, has been documented both in the Permian Basin (Allen et al., 2013; Bhatnager et al., 2018) and in other carbonate slope systems (Blasio et al., 2005; Moscardelli and Wood, 2008; Mazzanti and De Blasio, 2010; Janson et al., 2011; Mulder et al., 2012; Dakin et al., 2013; Principaud et al., 2015; Cardona et al., 2016; Moscardelli and Wood, 2016).
Figure 6.1: Examples of different scales of syn-sedimentary, intrastratal deformation observed on the Bone Spring outcrops. A) Deformed lime mudstone facies (F2) with micro-scale deformation. A2) Line drawing of figure A1. B) Meso-scale deformation. Debrite (F2) highlighted in red eroding into underlying strata. Note deformed chert beds within debrite. White lines indicate undeformed bedding below and above debrite. Location indicated in Figure 5.7. C) Meso-scale deformation. MTD sitting above SDS 7 erodes into underlying carbonate mudstone facies (F1). Some deformed strata and chert beds indicated by white lines. See location in Figure 5.4A. D) Meso-scale deformation within siliciclastic-dominant facies (F6). Deformed bedding in red with overlying undeformed bedding highlighted in white. Yellow lines indicate unrelated Cenozoic normal faults. See location in Figure 5.2A.
6.2 Style and Characteristics of Deformation

Different styles of deformation can be identified on the outcrop that provide insight into process, material strength and rheology, basin orientation, and failure conditions (Figure 6.2; Dott, 1963; Fisher, 1983; Stow, 1986; Elverhoi et al., 2000; Eyles and Eyles, 2000; Strachan, 2002; De Blasio et al., 2006; Moscardelli and Wood, 2008; Tripsanas et al., 2008; Haughton et al., 2009; Mazzanti and De Blasio, 2010; Talling et al., 2012; Auchter et al., 2016; Jablonska et al., 2018). Styles of intrastratal deformation range from creep to slide to slump to debris flow, but often deposits reflect a continuum between these styles (Figure 6.2; Dott, 1963; Nemec, 1990; Strachan, 2002; Tripsanas et al., 2008; Haughton et al., 2009; Talling et al., 2012). Creep deposits (sensu Auchter et al., 2016) are observed at many scales on the outcrop (Figure 6.2A, Figure 6.1A1, A2, Figure 5.8A) and are composed of carbonate mudstone facies (F1, F2, F3). Characteristics of creep deposits are mostly preserved strata with minimal plastic deformation (folds, boudinage) and minimal brittle failure (Figure 6.2A, Dott, 1963; Stow, 1986; Moscardelli and Wood, 2008; Auchter et al., 2016). A macro-scale example of creep is documented from Bone Canyon south (Figure 5.8A), where deformed beds reach dips of 40° with minimal intra-bed disturbance, indicating high strength and coherency of the failing rock material (Dott, 1963; Stow, 1986; Elverhoi et al., 2000; Tripsanas et al., 2008; Talling et al., 2012). The elastic and coherent nature of creep deposits indicates high sediment yield strength and/or low strain rates (Dott, 1963; Elverhoi et al., 2000), suggesting that sedimentation and/or gravitational forces created by steep slope angles may have caused failure. The prevalence of micro-scale creep within the carbonate slope facies (Figure 6.1A1, A2) suggest that the Bone Spring slope was almost always over-steepened and prone to failure (Stow, 1986).
Slide and slump deposits in the study area are composed of carbonate mudstone facies (F1, F2, F3) where bedding is generally preserved (Figure 5.4A), but plastically deformed (folds, boudinage, disrupted bedding) with minor brittle deformation (faulting, breccia; Figure 6.2B1, B2, B3). Slump deposits often sit on basal detachment surfaces that show brecciation and fracturing at the base (Figure 6.2B1, B2, B3; see basal shear zone, Cardona et al., n.d.). Slumping is differentiated from creep by plastic folding structures and evidence of brittle failure and high basal shear, suggesting detachment and transportation along the slope (Stow, 1986; Eyles and Eyles, 2000; Strachan, 2002; Moscardelli and Wood, 2008). Deposition of slump deposits on the steep Bone Spring slope indicate high strength of the slumped material that prevented subsequent failure and/or flow transformation to the basin in these instances (Dott, 1963; Fisher, 1983; Elverhoi et al., 2000; De Blasio et al., 2006; Tripsanas et al., 2008).

Lastly, debris flow deposits (i.e., debrites) are composed of carbonate mudstone facies (F1, F2, F3) with minimal strata preserved (Figure 6.1B, C, Figure 5.2B), a chaotic fabric with matrix supported clasts (Figure 6.2 C, D), brittle deformation features (breccia, fractures; Figure 6.2C, D), and erosional bases (Figure 6.2C, D), features common to debrites (Dott, 1963; Fisher, 1983; Stow, 1984; Moscardelli and Wood, 2008; Tripsanas et al., 2008; Talling et al., 2012). The chaotic fabric with brittle and erosional basal deformation suggests laminar flow with high basal shear stresses during transport (Dott, 1963; Stow 1984; Haughton et al., 2009., Talling et al., 2012). Flow transformation (Fisher, 1983; Haughton et al., 2009; Talling et al., 2012) of these debrite deposits along the sediment routing system may reflect the high occurrence of documented hybrid event beds in the distal Delaware basin (Driskill et al., 2018).
Figure 6.2: Examples of the various styles and characteristics of deformation deposits found on the Bone Spring outcrop. A) Creep deposit. Individual lamina set highlighted by white arrows shows micro-scale detachment and deformation but no failure at the bed-set (~ 1 m) scale. Note chert nodules mimic the primary bedding. Location indicated in Figure 5.3A. Pencil circled for scale. B1) Slump deposit. Two large (meso-scale) folds separated by decollement surfaces in red. Folded bedding in white. Compressional thrust faults in yellow below first fold. Location for B2 and B3 in boxed areas. Geologist for scale. B2) Breccia at the base of lower fold. Arrow and white lines indicate brecciated basal zone. B3) Thin section image near base of lower fold. Fractures in white are calcite-filled. Matrix is a F1 lime mudstone. Location of fold indicated in Appendix B.2. C) Debris flow deposit (debrite). Debrite truncates underlying undeformed strata in white. Note chert and mudstone clasts within debrite. Location on Shumard Cyn. south wall. D) Debris flow deposit displaying chaotic nature of chert and carbonate mudstone matrix. Geologist pointing at large chert nodule. Location on Bone Cyn. south wall.
CHAPTER 7

PARTITIONING OF CARBONATE, BIOGENIC SILICA, AND TERRIGENOUS SILICICLASTIC SEDIMENT

The presence and partitioning of mixed sediment on the Bone Spring slope is constrained by handheld XRF measurements (Figure 7.1). Plots of Silicon vs. Calcium establish carbonate-rich and siliciclastic-rich facies end-members (Figure 7.1A). These domains are corroborated by thin section, SEM, and hand sample analysis. However, biogenic silica is abundant on the Bone Spring slope due to sponge spicules and radiolaria that have been diagenetically altered to chert nodules and beds (Figure 4.4A; McDaniel and Pray, 1967). Chert beds are high in silicon and can cause confusion for evaluating a terrigenous source of silica; to mitigate this, we plot Calcium against detrital indicators Aluminum and Titanium (Tribovillard et al., 2006; Figure 7.1B). Some samples have high Si but low Ti+Al (e.g., two Facies 5 samples in Figure 7.1A, B), and thin section analysis (Figure 4.2E) reveals that these samples (1) are cemented by siliceous chert that is not derived from a terrigenous source and (2) little to no detrital siliciclastic sediment present. Other XRF-based methods to distinguish biogenic silica from detrital silica (e.g., Si/Al, Zr/Al, and Zr/Cr ratios) have also been useful in the Bone Spring Fm. (Driskill et al., 2018).

The most common facies in the study area are carbonate mudstones (F1, F2, F4), and these facies plot in the carbonate domain but with variable terrigenous input (Figure 7.1A). Thin sections reveal that a small volume (<10%) of well-rounded, silt-sized, terrigenous siliciclastic
sediment is present (Figure 4.1A, B), causing the observed scatter in the XRF-based terrigenous proxy (Figure 7.1B). The siliciclastic sediment is interpreted to be aeolian-derived dust that was transported from onshore aeolian fields (Presley, 1987; Fisher and Sarnthein, 1988; Cecil et al., 2018) during high relative sea levels and high carbonate production. The siliciclastic siltstone facies (F6) plot within the siliciclastic-domain (Figure 7.1A) and are interpreted as hemipelagic and sediment gravity flow processes connected to higher detrital siliciclastic sediment supply. The mixed-facies (F7) plot along a continuum between the carbonate- and siliciclastic-domain and represent a range of compositional mixing between F1 and F7 (Chiarella et al., 2017).

Figure 7.1: Facies-based XRF results. A) Calcium vs Silicon plot shows a carbonate- mixed- and siliciclastic-domain that represent facies end-members in the Bone Spring Fm. Highlighted in blue, two hand samples identified as Facies 5 (i.e. calciturbidites) plot in the siliciclastic-domain. Highlighted in red are carbonate-dominant slope facies (F1, F2) with background detrital siliciclastic sediment present. B) Plot of calcium vs detrital indicators titanium and aluminum (Tribovillard et al., 2006). Facies 5 samples plot along the y-axis, indicating diagenetic chert present, while the slope facies (F1, F2) plot off the y-axis in most cases, indicating some background detrital sediment present. These results are confirmed in thin section (Figure 4.2A, E).

XRF transects taken across SDS 4, 6, 7, 8, and 9 (Figure 5.6, Figure 5.1) demonstrate an enrichment of terrigenous sediment associated with slope detachment surface development.
Each transect, with the exception of transect 1, begins within the mixed- or carbonate-domain (circle symbols, Figure 7.2) and shifts toward the siliciclastic-domain at the surface (square symbols). Following this shift, the transects move back into the mixed- or carbonate-domain (triangle symbols). This shift occurs in every transect, but each transect is positioned in different parts of the calcium/silicon spectrum. Most transects (Figure 7.2B-E) occur within the mixed-domain with a shift toward the siliciclastic-domain at the surface, while the SDS 4 transect (Figure 7.2A) occurs predominantly within the siliciclastic-domain. This latter transect is associated with the FA4 deposits in Clinothem 4 (Figure 3.1, Figure 5.2A).

Figure 7.2: XRF transects through Shumard and Bone Canyons. Results demonstrate terrigenous sediment is correlated with slope detachment surfaces. The ‘x’ marks the first measurement with the stratigraphic path of the transects indicated by arrows. Turquoise circles mark readings of clinothems below surfaces, surface readings indicated by turquoise squares, and triangles represent clinothem readings above surfaces. Gray symbols correspond to facies in Figure 7.1. A) transect 1 through SDS 4. This transect is located predominantly within the siliciclastic-domain. B) transect 2 through SDS 6. Transect located within the mixed- to carbonate-domain and shifts toward the siliciclastic-domain at the surface. C-E) transects 3-5 and SDS 7-9, respectively. Surface transects show same trend as transect 2 with a shift from the mixed-domain toward the siliciclastic-domain at the slope detachment surface.
We interpret these results to record failure of the margin as the result of an influx of terrigenous siliciclastic sediment from the shoreline to the outer shelf margin from an interplay of accommodation and sediment supply (hereafter referred to as A/S, cf. Shanley and McCabe, 1994). We find the A/S ratio from Shanley and McCabe (1994) useful in a mixed siliciclastic-carbonate system to capture the interplay of accommodation and sediment supply. ‘A’ refers to accommodation that directly impacts both the position of terrigenous sediment (shoreward vs basinward) and the production of carbonate (e.g. high production during high accommodation). Variable progradation and aggradation ratios (P/A) of carbonate sediment are captured in the ‘A’ term (e.g. high aggradation of carbonate decreases accommodation). ‘S’ refers to sediment supply of terrigenous sediment. Because terrigenous siliciclastic sediment is associated with development of slope detachment surfaces (i.e., failures), we interpret several possible mechanisms for large-scale failure: (1) an increase in terrigenous sediment supply (Sultan et al., 2004; Vanneste et al., 2014), (2) weakened substrate from siliciclastic material (Kenter and Schlager, 1989; Kenter, 1990), (3) steep relict slopes created by the carbonate-dominant environment (Schlager, 1986; Schlager and Camber, 1986; Ross et al., 1994). Likely a combination of all three mechanisms initiated large-scale slope failure. The steep Bone Spring slope locally (10-25°) surpasses the predicted stability spectrum for carbonate mudstone margins (Kenter, 1990), so the introduction of weaker siliciclastic sediment onto an over-steepened carbonate slope is a likely failure mechanism. The position of four of the surfaces within the mixed- to-carbonate-domain (Figure 7.2B-E) suggest that only a slight increase in terrigenous sediment is necessary to trigger large-scale failure.

Storage of siliciclastic sediment in thin draping beds (<5 cm, Figures 5.2B, C, 5.5B, D) or meter-thick beds (Figure 5.2A) correlate to location within the carbonate/siliciclastic spectrum.
Transects associated with thin draping surfaces occur primarily within the mixed-domain (Figure 7.2B-E), while the transect associated with meter-scale siliciclastic beds is located predominantly within the siliciclastic-domain (Figure 7.2A). These results suggest local variability in A/S; for example, the meter-scale siliciclastic beds may record a relatively large decrease in A/S, while thin beds record only a minor decrease in A/S. In either case, allogenic or autogenic changes can produce these A/S changes, and the result is slope failure and bypass of siliciclastic sediment into the basin.
8.1 Evolution of the Victorio Peak-Bone Spring Mixed Margin

Outcrop observations of SDS and clinothem characteristics coupled with facies distributions and XRF results aid in reconstruction of the Leonardian shelf-slope profile in the study area. We use the evolution of the 10 clinothem packages described above to generalize slope-building processes and sediment delivery/partitioning in a mixed-lithology margin, including a 3D reconstruction of the study area (Figure 8.1A-D) and the resulting shelf-to-basin cross-section (Figure 8.1E). Four possible evolutionary steps are detailed (A, B, C, D), and the route the system takes through these steps may vary both laterally and temporally.

In time step A (Figure 8.1A) A/S is high (i.e. A/S>1), promoting high carbonate production with minimal detrital siliciclastic sediment input. Carbonate-rich hemipelagic and sediment gravity flow facies (Facies 1, 2, 3, 4, 5, 8) deposit on the slope and basin with minor siliciclastic input present as aeolian dust transport (Figure 7.1C; Cecil et al., 2018). The slope builds out with variable progradation/aggradation ratios (P/A), accounting for temporal changes in carbonate production and along-strike variability in slope morphology (Saller et al., 1989). The dominance of carbonate facies creates a relatively stable, albeit steep, environment, with minor micro-scale intrastratal deformation and meso-scale slope-attached MTDs (Figure 8.1A;
Moscardelli and Wood, 2008). Clinothem packages 1-3, 5-10 (Figure 3.1) represent the stratigraphic record of time step A.

In time step B (Figure 8.1B) siliciclastic and argillaceous sediment supply increases, decreasing A/S (e.g., A/S approaching 1), destabilizing the shelf-margin and upper slope, creating macro-scale, shelf-attached failures that develop into a slope detachment surfaces with associated MTDs (Facies 2). These SDS may be part of a larger clinoform surface. Siliciclastic sediment draping surfaces indicates bypass into the basin (Facies 6 and 7, Figure 8.1E; Armitage et al., 2009; Amerman et al., 2011; Grosheny et al., 2012). SDS 1, 2, 5-9 (Figures 3.1, 5.2B, C, 5.5B, D) and Clinothems 1-3, 5-10 are representative of time step B (Figure 3.1).

In time step C (Figure 8.1C) further decrease of A/S (e.g. A/S approaching 0) introduces larger volumes (relative to time step B) of siliciclastic and argillaceous material to the shelf edge and slope. A clinothem is built by siliciclastic material (Facies 6), with the amount of carbonate facies (Facies 7) dependent on the local carbonate production and flux of siliciclastic sediment. The steep, inherited slope also promotes bypass of siliciclastic sediment into the basin (Figure 8.1E). SDS 3 and 4 and Clinothem 4 represent the outcrop expression of time step C (Figure 3.1).

In time step D (Figure 8.1D) A/S returns to time step A conditions. Carbonate production again dominates, and the slope begins to prograde and aggrade over its failed deposits. Dip attitude changes across SDS in the study area suggest a complex, 3D slope morphology as the slope builds over its relict topography, perhaps with a strike-oriented lobate clinothem shape (Figure 8.3C1, C2, Figure 5.1, 5.3, 5.5, 5.7). This lobate style of progradation and aggradation on carbonate slopes has been documented as a mechanism for slope building in the Bone Spring
Fm. (Saller et al., 1989) and in other carbonate clinoform systems (Sonnenfeld, 1991; Gomez-
Perez et al., 1999; Katz et al., 2010; Playton et al., 2010; Playton and Kerans, 2018). Carbonate
packstone and MTD facies (Facies 2, 4, 5) are common at the base of clinolothems, as the relict
scarp surfaces attract coarse-grained sediment bypass (Eggenhuisen et al., 2010; Janson et al.,
2011; Stevenson et al., 2015). Toward the top of clinothem fill, undeformed lime mudstone
facies (Facies 1) dominate as the slope finds local equilibrium.

The implications of a lobate clinothem architecture and coarse-grained bypass is
demonstrated on the north wall of Shumard Canyon (Figure 8.2). In Clinolothems 1-7, bedding
orientations show an easterly slope progradation direction (90°, Figure 8.2B). At or near SDS 7
(Figure 8.2A), bedding orientations shift to a primarily southward progradation direction (180°,
Figure 8.2B). We interpret this rotation to record a slope inflection point, where a local re-entrant
may have locally focused deposition (Figure 8.2C). A high density of slope failure surfaces and
MTDs at the inflection point may be related to focusing of deposition (Figure 8.2A, C).
Additionally, four submarine channel deposits are aligned with this inflection point (Figure 3.1,
Figure 8.2A), suggesting that topographic lows created from failures may have acted as conduits
for coarse-grained sediment gravity flows (Figure 8.2C). These observations suggest that the
Shumard Canyon area may have been an entry point for coarse-grained sediment for a portion of
the northwestern Delaware Basin (Figure 2.1). Documentation of a younger Brushy Canyon Fm.
channel entry point at this location (Gardner et al., 2008) is further evidence for a prolonged
basin entry point at this location.
Figure 8.1: Interpretive schematics of Leonardian margin associated with the Guadalupe Mountains National Park outcrops. A) Time step A. High A/S with high carbonate production and minimal siliciclastic input. Slope progrades and aggrades at different rates laterally (P/A ratios). B) Time step B. Detrital siliciclastic and argillaceous sediment introduced to the outer margin from a decrease in A/S. The increase in siliciclastic material weakens the slope and creates large-scale, shelf-attached failure. Large-scale failure creates slope detachment surfaces that are likely part of a larger clinoform surface (magenta surface). Surfaces are coeval with MTDs at the toe-of-slope and in the basin. C) Time step C. Further A/S decrease introduces large volumes of siliciclastic sediment to the outer margin and upper slope. Continued surface development as siliciclastic and argillaceous sediment bypass to the basin. D) Time step D. Return to high A/S with the slope prograding and aggrading over its relict topography creating a new clinothem. E) Schematic shelf-to-basin cross-section based on the slope reconstructions representing an ABDABCD time sequence. Red surfaces represent slope detachment surfaces and corresponding time-lines similar to those documented on the outcrop. The transitioning facies of Facies 5, 6, and 7 represent expected transition from proximal to distal deposits. Boxed area represents outcrop-constrained portion of the schematic.
Figure 8.2: Slope inflection points may act as conduits for coarse-grained sediment to the basin. A) Line drawing of Shumard north wall (Figure 5.1) with dip azimuth readings (north is up). Orange arrows represent dip readings within clinothems 1-7, while blue arrows represent readings in clinothem 8. Note the calciturbidites and red discordant surfaces become more common near SDS 7. The large FA3 outcrop on Shumard south wall (Figure 5.3) also aligns with this region. B) Dip data taken along Shumard north wall. Colors correspond to location on Figure A. Average dip azimuth shifts 90 degrees after SDS 7. C) Schematic of Shumard north wall with a local inflection point in the slope. This inflection point creates instability from over-sedimentation and the resultant failure scarps act as conduits for coarse-grained turbidites to the basin.

In the study area, 7 of the 9 surfaces (SDS 1, 2, 5-9) likely followed an ABD path, skipping time step C and only storing siliciclastic sediment as thin bypass surfaces (Figure 7.2B-E, Figure 5.2B, C, Figure 5.6B, D). From SDS 3 to 4, the system likely followed an ABCD path, with a high magnitude decrease in A/S accounting for thicker siliciclastic beds on the slope (i.e. time step C; Figure 7.2A, Figure 5.2A). A prolonged decrease in A/S (e.g. A/S ~ 0), for example
the Bone Spring 1\textsuperscript{st}, 2\textsuperscript{nd}, and 3\textsuperscript{rd} Sands, would follow a similar path (i.e. ABCD), with time step C representing relatively large geologic time periods and large volumes of siliciclastic sediment bypass to the basin (Stevenson et al., 2015). A schematic cross-section of this time sequence (i.e. ABDABCD) is illustrated in Figure 8.1E.

8.2 Implications for Sequence Stratigraphic Concepts

Sequence stratigraphic concepts are commonly used for predicting facies types from seismic-scale geometries (Mitchum et al., 1977; Vail, 1987). Seismic-scale geometries are believed to record changes in allogenic (e. g. RSL) and/or autogenic (e. g. sediment supply) conditions (Vail, 1987). These concepts are important and useful for prediction, but often allogenic fluctuations (i.e. RSL changes) are relied upon without considering sediment supply and along-strike variability (see discussion in Burgess, 2016), resulting in over-simplified stratigraphic ‘pancake’ models (e. g., a sand body locally interpreted to represent a correlable lowstand all across a basin; Figure 8.3A; Saller et al., 1989; Montgomery, 1997b; Crosby et al., 2017; Bhatnager et al., 2018; Schwartz et al., 2018). Many studies have shown that sediment supply, accommodation, along-strike variability, and other factors affect the regional and local development of both low-order and higher-order systems tracts and sequences (Covault et al., 2006; Burgess, 2016; Madof et al., 2016; Harris et al., 2018; Trower et al., 2018).

Results from this study demonstrate carbonate and siliciclastic partitioning in Bone Spring Fm. clinothems that can be created by many forcing mechanisms. From an allogenic perspective, the siliciclastic beds associated with SDS (Figure 3.1, Figure 7.2) could record RSL fluctuations of different magnitude; in this case, we would expect similar processes occurring regionally, resulting in a relatively correlable basin stratigraphy (Li et al., 2015; Nance and
From an autogenic perspective, variable progradation and aggradation rates result in a rugose margin (Saller et al., 1989) that may provide conduits for siliciclastic sediment supply sources (i.e. channels) to reach the margin and bypass sediment into the basin without changing sea level and even during ‘highstand’ conditions (cf. Covault et al., 2006). As the margin compensationally builds by growth and failure (Figure 8.1; Saller et al., 1989; Playton et al., 2010), along-strike variability (cf. Madof et al., 2016) may result in localized differences in sediment input, clinothem composition and architecture, and a highly heterogeneous basin stratigraphy with carbonate and siliciclastic sediment deposited contemporaneously.

The sand-rich clinothem documented on the outcrop (Clinothem 4, Figure 3.1) may provide insight into this question. Based on the disconnected architecture and the interbedded (F6 and F7) basal contact of the sand-bodies (Figures 3.1, 4.4E, 5.6A1), we do not interpret this clinothem to represent the slope onlap of the 1st Bone Spring Sand, but instead favor an interpretation that this FA4 represents a sand-body within the larger-scale prograding carbonate package. Well-log correlations in the nearby Cutoff Fm. (Hurd et al., 2018) and Brushy Canyon Fm. (Gardner et al., 2008) also support this sand-rich clinothem being located within the L5 carbonate unit. Further work correlating this clinothem to the basin via biostratigraphy and well-log correlations would provide further context to this hypothesis. In either case, the interbedding with carbonate facies (Figures 4.4E, 5.4C, 5.6A1) suggests that autogenic processes are perhaps superimposed onto an allogenic signal, but deconvolving those signals would be very difficult. We advise to consider that both autogenic and allogenic processes contemporaneously act to build stratigraphy, and this complexity should be considered when making local and regional well correlations in the Delaware Basin and in similar mixed sediment routing systems (Figure 8.3B; Hampson, 2016; Madoff et al., 2016; Romans et al., 2016).
8.3 Sub-seismic-scale Predictions from Seismic-scale Architectural Elements

The spatial and temporal distribution of facies and depositional elements associated with the outcropping clinothems and SDS show how seismic-scale architectural elements can be used for prediction of sub-seismic-scale elements. Along strike, SDS 1-9 can be correlated the length of the study area (> 1 km) and dip and oblique view (Figures 8.4C1, C2, 5.1, 5.3, 5.5, 5.7) show that SDS are seismic-scale, with a minimum relief of 20-100 meters with a frequency (i.e. thickness of clinothems) of roughly every 20-40 meters of stratigraphic thickness.

Subsurface features of similar scale and architecture are imaged in seismic reflection data from the Leonardian margin along the Northwest Shelf (Figure 8.4; Sarg, 1988, Sarg et al., 1999). A seismic-scale lowstand siliciclastic wedge is interpreted (labeled Lower Avalon, Figure 8.4B) with a carbonate highstand package prograding over the top of the sand (labeled Victorio...
Peak and Bone Spring Carbonate, Figure 8.4B). Within the prograding package, clinoform geometries are identified (orange lines, Figure 8.4B). Outcrops of the Bone Spring Fm. from this study are shown at the same scale as the seismic data (Figure 8.4C1, C2), reinforcing the outcrop as an analog for the subsurface, particularly for predicting sub-seismic-scale facies distributions. From the results in this study, we expect MTDs and siliciclastic facies to onlap clinoforming slope detachment surfaces at the toe-of-slope and in the basin and become progressively more carbonate-rich moving stratigraphically towards the top of the clinothem (Figure 8.1E).
Figure 8.4: Predicting sub-seismic facies types from seismic-scale architecture. A) Uninterpreted seismic line of the Delaware Basin shelf margin from Sarg (1988) and Sarg et al. (1999). Location shown in Figure 2.1 inset. Red box indicates location of part B. B) Interpreted seismic section of Leonardian and Guadalupian shelf-to-basin stratigraphy. The unit labeled Bone Spring Carbonate would roughly correlate to the upper section (L6) of the Bone Spring outcrops in the study area. Orange lines highlight clinoform geometries within the prograding carbonate package. The Lower Avalon represents a basin-floor siliciclastic fan between L5 and L6 (Figure 2.2). C1) Uninterpreted oblique view of Shumard Canyon north. C2) Line drawing of C1 highlighting the similarity of scale and geometry of the clinoforming SDS on the outcrop to those seen in seismic. Results in this study suggest that an increase in MTDs and siliciclastic-argillaceous material would be associated with clinoforming surfaces in the basin.
CHAPTER 9

CONCLUSIONS

The stratigraphic architecture of the outcropping Bone Spring Fm. of Guadalupe Mountains National Park provides an opportunity to investigate slope-building processes and sediment delivery in a mixed siliciclastic-carbonate margin. A 3-D digital outcrop model constrains the stratigraphic architecture and sediment partitioning on the outcrop and reveals slope-building clinothems of mixed lithology. Bounding clinothems are slope detachment surfaces that are the result of macro-scale (km-scale) subaqueous failure of the outer margin related to terrigenous sediment influx. Handheld XRF results demonstrate a direct correlation between slope detachment failure surfaces and terrigenous sediment suggesting that terrigenous sediment weakened an over-steepened carbonate margin creating failure and subsequent terrigenous sediment bypass. At the base of clinothems, carbonate mass-transport deposits (MTDs) and coarse-grained carbonate allochem facies are common as the slope fills in its failed topography. At the top of clinothems, undeformed carbonate mudstone facies dominate as the slope finds local equilibrium. Bedding attitude data show dip azimuth changes from clinothem to clinothem, suggesting that the primary mechanism for slope evolution was through compensationally-stacked lobate slope-building packages. The distribution of calciturbidites within slope orientation inflection points suggests that coarse-grained entry points to the basin are controlled by slope morphology. We suggest that slope-building processes documented on the Bone Spring Fm. elucidate how mixed margins evolve and act as a primary control on
compositional stacking patterns and depositional styles in the basin. Insight from this study can be used to reconstruct local margins and aid in predicting reservoir-forming facies types in the basin.

Furthermore, siliciclastic deposits of the Bone Spring Fm. slope provide insight into sequence stratigraphic concepts in a mixed environment. The nature of a siliciclastic-rich clinothem documented in the study area (i.e. thickness, discontinuity, interbedding with carbonate sediment) is interpreted as a sand-body encased in a prograding carbonate package, suggesting that autogenic and allogenic processes contemporaneously act to build stratigraphy and better explain the documented heterogeneity in the Delaware Basin. Therefore, we suggest that both autogenic and allogenic processes be considered concurrently when making well-to-well correlations in the Delaware Basin. Lastly, this study provides a more robust stratigraphic context for the Bone Spring Fm. of the ‘western escarpment’ in Guadalupe Mountains National Park, an important field area for studying and observing shelf-to-basin stratigraphy and mixed-margin evolution.
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APPENDIX A

MEASURED SECTIONS

Figure A.1: Measured Section 1.

Figure A.2: Measured Section 2.
Figure A.3: Measured Section 3.

Figure A.4: Measured Section 4.
Figure A.5: Measured Section 5.

Figure A.6: Measured Section 6.
Figure A.7: Measured Section 7.

Figure A.8: Measured Section 8.
Figure A.9: Channel measured Section 1.
APPENDIX B

ADDITIONAL PHOTOMOSAICS

Figure B.1: Location of six photopanels compiled along the outcrop shown here overlaying the 3D textured model.
Figure B.2: West wall of Shumard Canyon photopanel. Location of figures in text indicated.
Figure B.3: West wall of Bone Canyon photopanel. Location of figures in text indicated.
APPENDIX C

EXTENDED FACIES DESCRIPTIONS

**Facies 1: thin-bedded laminated lime mudstone (hemipelagite/sediment gravity flow deposit)**

Facies 1 is the primary facies present in the Bone Spring outcrop (Figure 4.1). The grains in this facies include carbonate allochems, detrital quartz, and pyrite. These grains are surrounded by a matrix composed of carbonate clay and minor argillaceous clay. The detrital siliciclastic to carbonate percentage is approximately 08/92 percent. However, the s/c ratio can show variability representing increasing or decreasing compositional mixing. Planar laminations can be seen on the outcrop and thin section scale (Figure 4.2A). The planar beds alternate between dark black to dark brown in the outcrop, and thin sections show that the dark black portion is composed of higher proportion of clay (Figure 4.2A). The alternation of clay and silt layers may represent segregation of grain size indicating some degree of turbulence. Some evidence of soft-sediment deformation (SSD) can be found in this facies, but Facies 1 is a lower-end member of a continuum with Facies 2 representing decreasing to increasing deformation on the slope (Figure 4.1A). Laterally continuous chert beds are ubiquitous in Facies 1. These chert beds appear to be cyclical and are typically 5-10 cm thick, occur approximately every 10-20 cm, and follow bedding planes. Facies 1 is interpreted to be hemipelagic and sediment gravity flow processes deposited on the upper-to-middle slope.
**Facies 2: thin to thick-bedded deformed lime mudstone (mass-transport deposit)**

Facies 2 is identical to Facies 1 in composition, grain size, and siliciclastic to carbonate content, but differs in sedimentary structures. Facies 2 represents the high-end member of the deformation continuum (Figure 4.1A) with Facies 1. In this facies, SSD sedimentary structures such as water escape, convolute bedding, microfracturing, and folded strata (recumbent folds, imbricate folds) can be identified, but in many instances, there is no distinguishable bedding (i.e. chaotic bedding, Figure 4.2B). The prevalence of identifiable sedimentary structures and bedding distinguishes the intensity of deformation experienced, with more visible and coherent bedding moving toward the low-end continuum member. Chert beds typically mimic the character of the bedding. In highly deformed hand samples, thin sections show a high degree of fracturing with fractures filling with carbonate cement (Figure 4.2B). Facies 2 is interpreted to be hemipelagic and sediment gravity flow slope deposits that have experienced syn- and post-deposition deformation on the upper-to-middle slope.

**Facies 3: thick-bedded bioclastic lime wackestone to packstone (shallow-water, reworked-carbonate platform deposit)**

Facies 3 is similar to Facies 1 and 2 in composition but differs in higher content of coarse-grained bioclastic material and sparite. In this facies, crinoids, peloids, shell fragments, and sponge spicules can be easily identified in outcrop (Figure 4.2C). Additionally, Facies 3 differs from Facies 1 and 2 with less-visible sedimentary structures, lighter color, and lower frequency of chert beds (approximately every 30-40 cm). The lower frequency of chert beds likely indicates overall thickening of beds. Chert beds may be continuous, like found in Facies 1 and 2, but are more often nodular (Figure 4.2C). Thin sections show the coarse-grained fraction to be made up
very fine sand carbonate and biogenic grains (crinoids, bryozoan, brachiopods, peloids, shell fragments, spicules) with minimal, if any, detrital siliciclastic grains observed (s/c ratio <<1; 5C). Facies 3 is interpreted to be hemipelagic deposits on the upper-slope. The high presence of coarse-grained bioclastic content indicates proximity to the shelf margin relative to Facies 1 and 2.

_Facies 4: interbedded lime mudstone and bioclastic packstone (interbedded hemipelagites and turbidites)_

Facies 4 is composed of two elements: a carbonate mudstone element and a bioclastic packstone element (Figure 4.1A). The carbonate mudstone is identical to Facies 1. The bioclastic packstone element is composed of coarse-grained (very fine to coarse sand) carbonate grains that are mostly grain supported. Carbonate grains are composed of similar material found in Facies 1-3 (crinoids, brachiopods, bryozoan, spicules; Figure 4.2D). In the bioclastic packstone beds there is evidence of cementation from calcite and dolomite (Figure 4.2D). The packstone beds are typically on the order of cm- to mm-scale and have a frequency within the lime mudstone approximately every 1-2 centimeters and can be continuous, lenticular, or starved ripple beds (Figure 4.2D). Facies 4 is interpreted to be hemipelagic slope deposits with occurrences of sediment gravity flows, likely distal or low-density turbidity flows. These flows are pulses of shelfal material being swept off the shelf margin onto the slope, likely from wave, storm, current, or tidal forces.

_Facies 5: thick-bedded normally-graded bioclastic lime packstone to grainstone (turbidites)_

Facies 5 is made up of approximately 85% coarse-grained sediment, typically medium to coarse grain (0.5 to 1 mm in diameter) with some grains reaching pebble size. Grains are made up
entirely of carbonate or biogenic grains (siliciclastic to carbonate content, 0:100) that are
distinguished in hand sample and thin section as crinoids, peloids, brachiopods, bryozoan,
mollusks, sparite grains, sponges, and sponge spicules (Figure 4.2E). Thin sections reveal that
there is high occurrence of sparite and siliceous chert cement (Figure 4.2E). The source of the
chert cement is from biogenic siliceous material present on the upper-slope (sponge spicules and
radiolarians). In some occurrences of Facies 5, chert cement has entirely replaced beds.
Additionally, some samples show higher degrees of sparite, ranging from 15-30% on the
outcrop. Fabric indicates some normal grading, but also show poorly-sorted, “patchy” beds in
many places (Figure 4.2E). Other sedimentary structures observed are low-angle scours,
amalgamation surfaces, stylolites, and continuous red-brown colored beds. The presence of
grading, amalgamation surfaces, and depositional hiatuses (red surfaces) suggest these are
multiple carbonate sediment gravity flows, most likely turbidity current deposits (calciturbidites).

**Facies 6: thin-bedded laminated bioclast quartz siltstone (hemipelagites and turbidites)**

Facies 6 is similar to Facies 1 except for a higher detrital siliciclastic silt fraction (s/c>1, Figure
4.2F). This facies is made up of very fine detrital quartz and carbonate allochems set in a silt and
clay matrix. The matrix is dominated by siliciclastic grains and argillaceous mud (Figure 4.2F).
The ratio of siliciclastic to carbonate can vary and represents a siliciclastic-rich end member in
continuum with Facies 1 (Figure 4.1A). The increasing siliciclastic and argillaceous content
contributes to a noticeable lighter-brown color and different weathering pattern (Figure 4.2F).
Sedimentary structures are planar laminations to mostly a homogeneous, structureless face both
in outcrop and in thin section (Figure 4.2F). Chert is noticeably absent in this facies. Facies 6 is
interpreted to be siliciclastic-rich hemipelagic deposits on the upper-slope and possibly direct
settling of aeolian sediment blowing offshore. Facies 6 represents a clear change in accommodation or sediment supply in comparison to the other hemipelagic facies (F1, F2, F3).

**Facies 7: thin-bedded laminated quartz lime mudstone (hemipelagites and turbidites)**

Facies 7 is on a siliciclastic-to-carbonate continuum with Facies 1 and 6, representing approximately a medial position between the two facies (Figure 4.1A). The siliciclastic-carbonate ratio here is approximately 1, with about 45% siliciclastic and 55% carbonate material (Figure 4.2G). The facies is composed of very fine detrital quartz with minimal allochems present set in a carbonate mud and sparite matrix (Figure 4.2G). Like Facies 6, detrital quartz grains are well rounded (Figure 4.2G) and likely represent aeolian sediment. In outcrop, minimal sedimentary structures are observed but show some lamination with chert beds 5-10 cm thick and occur every 10-20 cm. This facies is slightly browner in color than Facies 1 but darker than Facies 6 (Figure 4.2G). Facies 7 is interpreted to be hemipelagic material with an increase in detrital influence, either from aeolian settling or hemipelagic processes.

**Facies 8: thick-bedded bioclastic lime packstone to grainstone (shallow-water carbonate platform deposits)**

Facies 8 is similar to Facies 1, 2, and 3 but shows a higher coarse-grained bioclastic content, lighter color, less sedimentary structures, and thicker bedding (Figure 4.1A). Thin sections show Facies 8 is grain supported with crinoids, bryozoan, brachiopods, peloids, sparite grains, bivalves, sponges, and sponge spicules visible (Figure 4.2H). The fine-grained fraction is entirely sparite, with no carbonate mud present. Chert beds are continuous to nodular and occur approximately every 0.5 to 1 meter, which is interpreted to represent higher bed thicknesses than Facies 1-3. Sedimentary structures are rarely observed and have not been documented, likely
indicating proximity to production centers. Facies 8 is interpreted to be carbonate platform in situ deposits. The higher content of bioclastic material indicates a more proximal location to the outer shelf than Facies 3 so represents the outer shelf margin environment rich in sponges, crinoids, and brachiopods. Deposits were likely interacting with tidal, storm, and/or current processes. This facies has been previously identified as Lower Victorio Peak by Kirkby (1982) and will be considered Lower Victorio Peak for the remainder of this paper.
APPENDIX D

CALCIUM VS SILICON XRF PLOTS

Figure D.1: XRF transects of clinoforms 4, 6-9 plotted as calcium vs silicon