VARIATIONS IN THE ARCHITECTURE OF FLUVIAL DEPOSITS WITHIN A MARGINAL MARINE SETTING, EOCENE SOBRARBE AND ESCANILLA FORMATIONS, SPAIN

by

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ABSTRACT

Fluvial systems located near marginal-marine settings are prolific hydrocarbon reservoirs around the world. Outcrops that contain strata that can be correlated from fluvial deposits to coevally deposited marginal marine deposits are important because they provide a rare opportunity to relate fluvial architecture to relative changes in sea level (accommodation to sediment supply changes (A/S)). This information is important because it can be used to aid in reservoir characterization and prediction.

This dissertation comprises four outcrop studies of a marginal marine and fluvial deposits of the Eocene Sobrarbe and Escanilla Formations, Ainsa Basin, Spain, focused at three different scales of observation: third- (Chapter 2), fourth- (Chapters 2, 3, 4, and 5), and fifth-order cyclicity (Chapters 4 and 5). This dissertation advances our scientific knowledge about the deposition of fluvial systems in high-accommodation, high-sediment supply settings. The continuity between the axis and margin of the basin and fluvial and coevally deposited marginal marine strata allow for the quantitative documentation of: (1) structure-stratigraphic interactions of the Sobrarbe and Escanilla Formations (Chapter 2); (2) the relationship between fluvial architecture and A/S (Chapter 3); (3) axis-to-margin variations of transgressive fluvial deposits (Chapter 4); and (4) relationships between fluvial architecture and shoreline trajectory (Chapter 5).

Key contributions of this dissertation are the following. First, this dissertation provides a more comprehensive knowledge about the structure-stratigraphic evolution of the Sobrarbe and Escanilla Formations than what was previously known (Chapter 2). Second, this dissertation
provides knowledge of how large-scale stratigraphic stacking patterns can be used as a predictor of small- (reservoir) scale characteristics and how subdividing populations on the basis of geological distinctions can have important implications when building reservoir models (Chapter 3). Third, this dissertation provides a better understanding of the lateral and vertical distributions of stratigraphic architecture and static connectivity of fluvial strata within a transgressive fluvial system (Chapter 4). Fourth, this dissertation provides a better understanding of the role of autogenic and allogenic processes on stratigraphic architecture of transgressive fluvial deposits and shoreline trajectory at both fourth-order and fifth-order scales of cyclicity (Chapter 5).
TABLE OF CONTENTS

ABSTRACT ................................................................................................................................... iii
LIST OF FIGURES ........................................................................................................................ x
LIST OF TABLES ....................................................................................................................... xiii
ACKNOWLEDGMENTS ........................................................................................................... xiv

CHAPTER 1  INTRODUCTION AND DISSERTATION FORMAT ........................................... 1
  1.1 Introduction to Shallow Marine Settings ................................................................. 1
  1.2 Dissertation Format ................................................................................................. 4
  1.3 References ............................................................................................................... 6

CHAPTER 2  GEOLOGIC EVOLUTION OF THE SOBRARBE AND ESCANILLA
  FORMATIONS, SPAIN ......................................................................................................... 14
  2.1 Introduction ........................................................................................................... 14
  2.2 Geology of the Ainsa Basin .................................................................................. 15
    2.2.1 Stratigraphy ............................................................................................. 16
      2.2.1.1 Sobrarbe Formation .................................................................... 16
      2.2.1.2 Escanilla Formation .................................................................... 18
    2.2.2 Structure .................................................................................................. 19
      2.2.2.1 Mediano Anticline ...................................................................... 21
      2.2.2.2 Boltaña Anticline ........................................................................ 22
      2.2.2.3 Olson Anticline ........................................................................... 23
      2.2.2.4 Arcusa Anticline ......................................................................... 23
      2.2.2.5 Buil Syncline ............................................................................... 24
      2.2.2.6 Arcusa Syncline .......................................................................... 24
  2.3 Data and Methods ................................................................................................. 24
3.5.1 Cycle 2 .............................................................. 74
3.5.2 Cycle 3 .............................................................. 74
3.5.3 Cycle 4 .............................................................. 75
3.5.4 Cycle 5 .............................................................. 75

3.6 Axis to Margin Variations in Stratigraphic Architecture ........................................... 76
3.6.1 Cycle 2 .............................................................. 76
3.6.2 Cycle 3 .............................................................. 77
3.6.3 Cycle 4 .............................................................. 78
3.6.4 Cycle 5 .............................................................. 78
3.6.5 Summary of Axis to Margin Variations in Stratigraphic Architecture ................... 79

3.7 A/S Variations in Relation to Stratigraphic Architecture ........................................... 80

3.8 Discussion ..................................................................................................................... 82

3.9 Conclusions .................................................................................................................. 84

3.10 References .................................................................................................................. 85

CHAPTER 4 QUANTITATIVE OUTCROP CHARACTERIZATION OF AXIS TO MARGIN CHANGES IN STRATIGRAPHIC ARCHITECTURE OF TRANSGRESSIVE FLUVIAL DEPOSITS: ESCANILLA FORMATION, SPAIN .................... 103

4.1 Abstract ....................................................................................................................... 103

4.2 Introduction .................................................................................................................. 104

4.3 Geologic Setting ......................................................................................................... 107

4.4 Dataset and Methods ................................................................................................. 111
  4.4.1 Lithofacies ............................................................................................................. 111
  4.4.2 Fluvial Hierarchy of Architectural Elements ....................................................... 111
    4.4.2.1 Story ............................................................................................................... 112
# LIST OF FIGURES

<table>
<thead>
<tr>
<th>Figure</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Figure 1.1</td>
<td>Marginal marine depositional settings</td>
<td>10</td>
</tr>
<tr>
<td>Figure 1.2</td>
<td>Diagram documenting interplay between A/S</td>
<td>11</td>
</tr>
<tr>
<td>Figure 1.3</td>
<td>Fluvial sequence stratigraphy model</td>
<td>12</td>
</tr>
<tr>
<td>Figure 1.4</td>
<td>Paleogeographic map of the Tremp-Ainsa-Jaca Basin</td>
<td>13</td>
</tr>
<tr>
<td>Figure 2.1</td>
<td>Ainsa Basin history</td>
<td>43</td>
</tr>
<tr>
<td>Figure 2.2</td>
<td>Geologic map of the Ainsa Basin</td>
<td>44</td>
</tr>
<tr>
<td>Figure 2.3</td>
<td>Structural cross section of Ainsa Basin</td>
<td>45</td>
</tr>
<tr>
<td>Figure 2.4</td>
<td>Geologic map of the study area</td>
<td>46</td>
</tr>
<tr>
<td>Figure 2.5</td>
<td>Photographs of western half of study area</td>
<td>47</td>
</tr>
<tr>
<td>Figure 2.6</td>
<td>A longitudinal stratigraphic cross section of the western study area</td>
<td>48</td>
</tr>
<tr>
<td>Figure 2.7</td>
<td>Photographs of eastern half of study area</td>
<td>48</td>
</tr>
<tr>
<td>Figure 2.8</td>
<td>A longitudinal stratigraphic cross section of the eastern study area</td>
<td>49</td>
</tr>
<tr>
<td>Figure 2.9</td>
<td>Photographic examples of depositional environments</td>
<td>50</td>
</tr>
<tr>
<td>Figure 2.10</td>
<td>Cross-plots of cycle vs. channel thickness and net-sand content</td>
<td>51</td>
</tr>
<tr>
<td>Figure 2.11</td>
<td>Geologic map of Cycle 1</td>
<td>52</td>
</tr>
<tr>
<td>Figure 2.12</td>
<td>Geologic map of Cycle 2</td>
<td>53</td>
</tr>
<tr>
<td>Figure 2.13</td>
<td>Geologic map of Cycle 3</td>
<td>54</td>
</tr>
<tr>
<td>Figure 2.14</td>
<td>Geologic map of Cycle 4</td>
<td>55</td>
</tr>
<tr>
<td>Figure 2.15</td>
<td>Geologic map of Cycle 5</td>
<td>56</td>
</tr>
<tr>
<td>Figure 2.16</td>
<td>Geologic map of Cycle 6</td>
<td>57</td>
</tr>
<tr>
<td>Figure 2.17</td>
<td>Summary diagram</td>
<td>58</td>
</tr>
<tr>
<td>Figure 3.1</td>
<td>Schematic diagram of fluvial components and depositional settings</td>
<td>90</td>
</tr>
</tbody>
</table>
Figure 3.2 Ainsa Basin history ................................................................. 91
Figure 3.3 Geologic map of study area ............................................... 92
Figure 3.4 Photographs of study area.................................................. 93
Figure 3.5 A longitudinal stratigraphic cross section of the study area ... 94
Figure 3.6 Illustrations of architectural elements .............................. 94
Figure 3.7 Photographic examples and data of architectural elements .... 95
Figure 3.8 Diagram defining shelf edge trajectory .............................. 96
Figure 3.9 Paleogeographic maps of Cycles 2-5 ................................. 97
Figure 3.10 Data documenting axis to margin changes of fluvial system .. 98
Figure 3.11 Data documenting changes in fluvial system in relation S.E. trajectory .. 99
Figure 3.12 Subdividing data of fluvial deposits ................................. 100
Figure 3.13 Summary diagram for fluvial deposits ............................. 101
Figure 4.1 Schematic diagrams of fluvial settings ............................... 129
Figure 4.2 Ainsa Basin history ............................................................. 130
Figure 4.3 Geologic map of study area .............................................. 131
Figure 4.4 Photographic examples of stories ....................................... 132
Figure 4.5 Illustrations of architectural elements ............................... 133
Figure 4.6 Diagram defining static connectivity ................................. 134
Figure 4.7 Photopanel of Mondot field area ...................................... 135
Figure 4.8 Photopanel of La Susia field area ...................................... 136
Figure 4.9 Photopanel of Peñalebrera field area ................................. 137
Figure 4.10 Lateral stratigraphic cross section of transgressive interval of Cycle 2 .. 138
Figure 4.11 Quantitative data of axis to margin changes in stratigraphy .... 139
Figure 4.12  Charts documenting variations in static connectivity ............................. 140
Figure 5.1   Schematic diagrams of fluvial settings ................................................... 173
Figure 5.2   Ainsa Basin history ............................................................................. 174
Figure 5.3   Geologic map of field area.................................................................... 175
Figure 5.4   Illustrations of architectural elements.................................................. 176
Figure 5.5   Cross section and photopanel documenting shoreline trajectory.......... 177
Figure 5.6   Photopanels of field area...................................................................... 178
Figure 5.7   Lateral stratigraphic cross section of the axial part of Cycle 2 .......... 179
Figure 5.8   Data of fluvial architecture in relation to shoreline trajectory .......... 180
Figure 5.9   Summary diagram of conclusions......................................................... 181
LIST OF TABLES

Table 2.1  Descriptions of depositional environments identified in this study .......... 59
Table 3.1  Descriptions of stories identified in this study ........................................ 102
Table 4.1  Descriptions of lithofacies identified in this study ................................. 141
Table 4.2  Descriptions of stories identified in this study ........................................ 142
Table 4.3  Descriptions of architectural elements identified in this study ............... 143
Table 5.1  Descriptions of stories identified in this study ........................................ 182
Table 5.2  Descriptions of architectural elements identified in this study ............... 183
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CHAPTER 1
INTRODUCTION AND DISSERTATION FORMAT

1.1 Introduction to Marginal Marine Settings

The confluence of non-marine (i.e. fluvial and eolian) and shallow marine (i.e. delta, estuary, strandplain) depositional systems (Fig. 1.1) occurs within a zone, termed the marginal marine setting. This setting extends from fluvial and coastal plain environments across the shoreline, to shallow-marine environments. The stratigraphic architecture of deposits in marginal-marine settings are controlled by the dynamic interaction between river processes which are controlled by tectonics and climate and ocean processes, which are controlled by eustasy, subsidence, waves, and tides (Curray, 1964; Galloway, 1975; Wright, 1977; Boyd et al., 1992; Fig. 1.1).

Modern and ancient marginal marine deposits are scientifically important as they record relative changes in sea level, evolution of landforms, tectonic events on the continental margin, and sediment transfer from the continent, across the continental shelf and into deep marine environments. Marginal marine deposits are economically important because they are prolific hydrocarbon reservoirs around the world. For example, the Gorgon Field located in northwest Australia contains approximately 35.3 trillion cubic feet (TCF) of natural gas (JPT, 2010); and the Columbus Basin located in Trinidad, West Indies contains approximately 20 TCF of gas reserves and more than one billion barrels (BBL) in oil reserves (Sydow et al., 2003).

Marginal marine oil and gas reservoirs are commonly investigated using seismic data that resolve stratal units on the order of 10s of meters in thickness at reservoir depths; and well logs
and cores which record reservoir properties such as grain-size, porosity, and permeability at millimeter scale in vertical orientation, and no control on lateral and longitudinal scales. As such, subsurface investigations lack the resolution that can be obtained from well-exposed outcrops. Therefore, outcrop studies aid in reservoir modeling and reservoir development.

Sequence stratigraphy models place genetically related facies within a framework of chronostratigraphically significant surfaces such as sequence boundaries and flooding surfaces in context of changes in relative sea-level (Sloss et al., 1949; Sloss, 1963; Mitchum et al., 1977; Vail et al., 1977; Posamentier and Vail, 1988, Posamentier et al., 1988; Van Wagoner et al., 1988). Sequence stratigraphy therefore places marginal marine, marine, and fluvial deposits in a classification defined by regressive (shoreline migrating basinward) and transgressive (shoreline migrating landward) shoreline trajectories, which are caused by relative changes in sea level. The primary factors that control shoreline trajectory are the ratios between the rate of change in accommodation and sediment supply (Galloway, 1987; Posamentier and Vail, 1988; Posamentier et al., 1988; Jervey, 1988; Muto and Steel, 1992; Shanley and McCabe, 1994; Muto and Steel, 1997). Figure 1.2 documents upward changes in stratigraphy related to different accommodation and sediment supply regimes.

Following the development of sequence stratigraphy in marginal marine settings, geoscientists applied sequence stratigraphic principles and concepts to fluvial stratigraphy (Fig. 1.3; Posamentier and Vail, 1988; Shanley and McCabe, 1991; Miall, 1991; Schumm, 1993; Wright and Marriott, 1993; Shanley and McCabe, 1994; Holbrook et al., 2006). Fluvial sequence stratigraphic models primarily focus on low accommodation successions in which the lower bounding surface of a sequence, a sequence boundary, is an unconformity that is manifested as an incised valley formed during a drop in relative sea-level (Shanley and McCabe, 1991;
Alexander, 1992; Wright and Marriott, 1993; Schumm, 1993; Shanley and McCabe, 1994;
Aitkin and Flint, 1995; Olsen et al., 1995; Martinsen et al., 1999; Plint et al., 2001; Posamentier,
2001; Arnot et al., 2002; Holbrook et al., 2006). Few outcrop studies have focused on the fluvial
strata in a high-accommodation, high sediment supply settings in fluvial (Burns et al., 1997;
Fanti and Catuneanu, 2010) and deltaic (Plink-Bjorklund et al., 2001; Pyles and Slatt, 2007)
settings in which no incised valleys are formed.

The Eocene Sobrarbe and Escanilla Formations, located in the Ainsa Basin, Spain (Fig.
1.4), were coevally deposited and contain fluvial and deltaic strata, respectively, that were
deposited within a high accommodation, marginal-marine setting when the rate of aggradation
and sedimentation were high (Bentham et al., 1992; Dreyer et al., 1999; Moss-Russell, 2009;
Mochales, 2012; Pyles et al., in review, Appendix A). The well-exposed and continuous outcrops
of the Sobrarbe and Escanilla Formations contain multiple condensed-section (organically-rich
shale layers) bounded regressive-transgressive (R-T) cycles that build basinward through time
(Dreyer et al., 1999; Moss-Russell, 2009). The upper and lower surfaces of each R-T cycle can
be correlated from marginal-marine deposits to coevally deposited fluvial strata.

The goal of this dissertation is to use well exposed outcrops to quantitatively document
upward, lateral, and longitudinal trends in stratigraphic architecture and lithofacies in relation to
shoreline trajectory for fluvial deposits deposited in a high accommodation, high sediment
supply setting, to address the following questions:

• How does the stratigraphic architecture of fluvial deposits vary in relation to changes in the
  rate of accommodation and sediment supply at a regional scale (Chapters 2 and 3)?
• How does the stratigraphic architecture, grain-size, net-sand content, and static connectivity
  of fluvial channel and floodplain deposits relate to variations in shoreline trajectory within a
single fourth-order transgressive succession (Chapters 4 and 5)?

1.2 Dissertation Format

The following chapters of this dissertation are outlined below. Chapter 2 is a regional geologic study of the Sobrarbe and Escanilla Formations, which provides context to Chapters 3-5, which have been submitted to peer-reviewed journals and adhere to the format of the journal to which it is submitted. Chapters 3-5 include their own abstract, introduction, geologic setting, data and methods, discussion, conclusion, and reference sections. The key goals and contributions of each chapter are summarized below.

- **Chapter 2**- Chapter 2 is a study of the entire Sobrarbe and coevally deposited Escanilla Formations. Chapter 2 presents new data in order to provide a geologic overview of the entire Sobrarbe Formation and coevally deposited Mondot Member of the Escanilla Formation and the structure-stratigraphic interactions between the following syndepositionally active structures: Mediano Anticline, Boltaña Anticline, Olson Anticline, Arcusa Anticline, Buil Syncline, and Arcusa Syncline. Chapter 2 gives greater context to Chapters 3, 4, and 5 of the dissertation, which are studies of the coevally deposited fluvial deposits of the Escanilla Formation.

- **Chapter 3**- Fluvial systems are important hydrocarbon reservoirs located around the world. One of the problems facing the development strategies of fluvial reservoirs is our lack of understanding of axis-to-margin variations in fluvial systems and how varying rates of accommodation and sediment supply during deposition relates to spatial and temporal variations in stratigraphic architecture of fluvial systems. Chapter 3 quantitatively documents spatial and temporal patterns in stratigraphic architecture of fluvial channel and floodplain
deposits for four regressive-transgressive (R-T) cycles of the Sobrarbe and Escanilla Formations. This information is used to document how fluvial architecture: (1) changes along an axis-to-margin transect through the system and (2) relates to A/S ratio. Results are used to emphasize the importance of subdividing stratigraphic data based on geologically meaningful distinctions when building reservoir models. This study will be submitted to JSR for publication. David Pyles is a coauthor for this article.

- **Chapter 4**- Transgressive fluvial strata are deposited during an overall landward migration of the shoreline. Few studies have decidedly focused on transgressive fluvial strata, especially those deposited in high-accommodation settings. The goals of this article are to quantitatively document, for the first time, spatial patterns in stratigraphic architecture, net-sand content, the size and modal grain size of channel-belt elements, and static connectivity in order to evaluate how stratigraphic architecture varies laterally and vertically within a transgressive fluvial system deposited in a high-accommodation setting. Concepts and data generated in this study can be used to aid in the interpretation of subsurface data and quantitatively constrain geologic models, thereby reducing uncertainty in the development of reservoirs. This chapter has been submitted to the AAPG Bulletin for publication. David Pyles is a coauthor for this article.

- **Chapter 5**- Non-marine sequence stratigraphic models are developed from low-accommodation sequences, whereas, little is known sequence stratigraphy in high-accommodation settings. This study uses outcrop data to document how stratigraphic architecture, net-sand content, grainsize, and static connectivity of fluvial channels relate to changes in shoreline trajectory within a high-accommodation transgressive unit at fourth- and fifth-order scales of cyclicity. This information is used to evaluate: (1) differences between
fluvial strata deposited when the shoreline trajectory was moving basinward versus strata deposited when the shoreline trajectory was moving landward for fifth-order R-T cycles within the transgressive unit of the fourth-order R-T cycle, (2) differences in fluvial strata deposited from the base to the top of the fourth-order transgressive unit, and (3) the role of autogenic and allogenic processes on the stratigraphic architecture of fluvial deposits and shoreline trajectory at both fourth-order and fifth-order scales. This chapter will be submitted to Sedimentology for publication. David Pyles is a coauthor for this article.

- **Chapter 6** - This concluding chapter discusses how Chapters 2-5 have expanded our scientific knowledge of the overarching topic being studied.

### 1.3 References

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Boyd, R., Darlymple, R., and Zaitlin, B.A., 1992, Classification of clastic depositional environments: Sedimentary Geology, 80, 139-150.


Plint, A.G., McCarthy, P.J., and Faccini, U.F., 2001, Nonmarine sequence stratigraphy: Updip expression of sequence boundaries and systems tracts in a high resolution framework,


Figure 1.1 (A) The primary marginal marine depositional settings are barrier islands, beaches, deltas, estuaries, lagoons, and strand plains. From Boyd et al. (1992). These settings are controlled by the interaction between continental and marine processes. For example, the plan view geometry of a delta is controlled by the dynamic interaction between tidal, river, and wave energy. Aerial photographs of tidal, river, and wave-dominated deltas are shown in B, C, and D respectively.
Figure 1.2 Diagram documenting how rate of accommodation and rate of sediment supply control the direction of shoreline migration through time, as well as the stratigraphic architecture of a sequence. Modified from Shanley and McCabe (1999).
Figure 1.3 Fluvial sequence stratigraphy model developed by Shanley and McCabe (1994). This diagram illustrates the relationships between the stratigraphic architecture of shoreface and fluvial strata as a function of changes in base-level where sediment supply being held constant. (A) Slow rates of base-level fall. (B) Reduced rates of base-level fall to slowly rising base level. (C) High rates of base-level rise. (D) Low rates of base-level rise.
Figure 1.4 Paleogeographic map of the Tremp-Ainsa-Jaca basin (modified from Michael et al., 2014).
CHAPTER 2
GEOLOGIC EVOLUTION OF THE SOBRARBE AND ESCANILLA FORMATIONS,
AINSA BASIN, SPAIN

2.1 Introduction

The Ainsa basin formed during the Pyrenean orogeny during the late Cretaceous through Miocene periods, as a result of the collision between the Iberian micro-plate and the Eurasian plate (Puigdefabregas et al., 1986, Seguret et al., 1984; Remacha et al., 1988; Fernandez et al., 2004). The Ainsa Basin is bounded by five syndepositionally active structures (Poblet et al., 1998; Fernandez et al., 2004; Hoffman, 2009) (Figs. 2.1, 2.2, 2.3): (1) the Boltaña Anticline to the west, (2) the Mediano Anticline to the east, (3) the Añisclo Anticline to the north, (4) the Cotiella thrust to the northwest, and (5) the Montsec thrust to the south. The Ainsa Basin also contains several syndepositionally active intrabasinal structures including the Arcusa and Olson Anticlines and the Arcusa and Buil Synclines (Figs. 2.1, 2.2).

Several studies have documented the structure-stratigraphic interactions within the Ainsa Basin. Garrido-Megias (1968), Garrido-Megias (1973), Poblet et al. (1998), Fernandez et al. (2004), Fernandez et al. (2012), Munoz et al. (2013) conducted studies of the structure-stratigraphic interactions within the Ainsa Basin and the surrounding region the third order scale- (sensu Mitchum and Van Wagoner, 1991). Dreyer et al. (1999), documented structure-stratigraphic interactions within the Sobrarbe Formation at the fourth-order scale (sensu Mitchum and Van Wagoner, 1991) by conducting a 2-D study of the outcrop along the western part of the Ainsa Basin that analyzed the influence of the Arcusa and Olson Anticlines on the
stratigraphic architecture of strata within the Sobrarbe Formation. Bentham et al. (1992) and Bentham and Burbank (1996) used magnetostratigraphy to document the timing of the growth of the Mediano Anticline during deposition of the Escanilla Formation. To date, no study has combined stratigraphic data from both the Sobrarbe and Escanilla Formations to better understand the structure-stratigraphic interactions within the Ainsa Basin during deposition of these formations.

This chapter builds upon these previous works by using outcrop data from the Sobrarbe Formation and coevally deposited Mondot member of the Escanilla Formation at a fourth-order scale to document variations in the nature of geologic contacts (i.e. conformable and unconformable), paleocurrents, cycle thicknesses, channel and mouth bar thicknesses, and net-sand content. This information is used to evaluate the structure-stratigraphic interactions between this strata and the following syndepositionally active structures: Mediano Anticline, Boltaña Anticline, Olson Anticline, Arcusa Anticline, Olson Syncline, and Arcusa Syncline. This chapter gives greater context to Chapters 3, 4 and 5 of the dissertation.

2.2 Geology of the Ainsa Basin

The Ainsa Basin is located 280 km northwest of Barcelona, Spain (Fig. 2.1), is ~ 1000 km² in area, and extends ~ 25 km in the east-west direction and ~ 40 km in the north-south direction (Fig. 2.2). This section describes the stratigraphy and syndepositional basin bounding and intrabasinal structures of the Ainsa Basin.
2.2.1 Stratigraphy

The Ainsa Basin fill succession is divided into two groups: (1) the Hecho Group, which is a ~4 km thick package of deepwater strata that is overlain by (2) the Campodarbe Group, which is a ~2 km thick package of marine, deltaic, and fluvial strata (Fig. 2.1; Das Gupta and Pickering, 2008). These sediments unconformably overlie mixed carbonate and siliciclastic Paleocene strata and Ypresian Alveolina Limestone that predate the Ainsa Basin (Figs. 2.1, 2.3; Fernandez et al., 2004).

The Hecho Group is ~4 km thick and is subdivided into seven smaller units, termed turbidite systems (Mutti et al., 1989; Pickering and Corregidor, 2005; Pickering and Bayliss, 2009) and formations by Moody et al. (2012). From oldest to youngest they are: (1) Fosado, (2) Arro-Charo, (3) Gerbe, (4) Banaston, (5) Ainsa, (6) Morillo, and (7) Guaso (Figs. 2.1, 2.2). Each is a third-order stratigraphic unit, meaning they record ~1-2 million years of deposition each (sensu Mitchum and Van Wagoner, 1991).

The focus of this study is on the Campodarbe Group, which is ~2 km thick and is divided into the Sobrarbe and Escanilla Formations (Fig. 2.1). These formations record the final filling of the Ainsa Basin and the progradation of a linked fluvial-deltaic system over the area (Bentham et al., 1992; Dryer et al., 1999; Pickering and Bayliss, 2009; Moss-Russell, 2009; Silalahi, 2009; pds et al., in review, Appendix A).

2.2.1.1 Sobrarbe Formation

The Sobrarbe Formation is the basal formation of the Campodarbe Group and represents the youngest marine strata in the Ainsa basin-fill succession (Fig. 2.1). Based on biostratigraphic and magnetostratigraphic data, the Sobrarbe Formation was deposited over a duration of ~3
million years in the Middle-Late Leutian (Dreyer et al., 1999; Mochales et al., 2012) and is ~1000 m thick (Dreyer et al., 1999). Based on biostratigraphic and magnetostratigraphic data, rates of sediment accumulation (undecomposed) are ~32 cm k/yr (Dreyer et al., 1999; Mochales et al., 2012).

The Sobrarbe Formation conformably overlies marine deposits of the Guaso Formation and conformably underlies the fluvial deposits of the Escanilla Formation. The base of the Sobrarbe Formation is an organic-rich black shale interpreted as a condensed section (Silalahi, 2009; Moss-Russell, 2009) that represents a long period of non-deposition in the basin. This condensed section can be correlated across the basin (Dreyer et al., 1999; Silalahi, 2009; Moss-Russell, 2009; Hoffman, 2009). The Sobrarbe Formation is sourced from the Pyrenean massif through the Tremp-Graus Basin in the east (Dreyer et al., 1999).

The Sobrarbe Formation contains cyclic alternations between mudstone-dominated delta plain deposits, carbonates, delta front sandstones, collapse complexes, muddy delta slope deposits, and turbidite sandstone (Dreyer et al., 1999). Dreyer et al. (1999) divided the Sobrarbe Formation into two tectonostratigraphic packages and four composite sequences (CS): the Comaron CS, Las Gorgas CS, Barranco el Solano CS, and Buil CS. Each composite sequence contains prograding clinoforms and slumps increasing in size and abundance from one composite sequence to the next. However, Moss-Russell (2009) and Pyles et al. (in review, Appendix A) divided the Sobrarbe Formation into six condensed section bound regressive-transgressive (R-T) cycles (Figs. 2.4, 2.5). The condensed section boundaries for each R-T cycle can be mapped through out the basin (Fig. 2.4). Each R-T Cycle is approximately fourth-order in duration meaning they record approximately 0.1 to 0.5 m.y. of deposition each (sensu Mitchum and van Wagoner, 1991). Each cycle has an identifiable shelf-edge. The location of successive shelf-edges record northward progradation and aggradation of the depositional system (Fig. 2.5A).
Moss-Russell’s (2009) division of cycles are used in this study.

2.2.1.2 Esanilla Formation

The Esanilla Formation was first described by Garrido-Megías (1968). It is part of the Campodarbe Group and represents non-marine deposition within the south-central Pyrenees. Based on biostratigraphic and magnetostratigraphic data, the Esanilla Formation was deposited between the late Lutetian-early Bartonian through the late Preabonian ages (41-34 Ma) (Cuevas Gozalo, 1990; Bentham and Burbank, 1996, Mochales et al., 2012). The Esanilla Formation conformably overlies the deltaic and shallow marine deposits of the Sobrarbe Formation, is up to 1.1 km thick (Bentham et al., 1992), and is unconformably overlain by the Oligocene Collegats Formation, a conglomeratic alluvial fan deposit (Garrido-Megías, 1973 in Bentham et al., 1992). The Esanilla Formation is sourced from the Pyrenean massif through the Tremp-Graus Basin in the southeast (Vincent, 2001).

Bentham, (1992) divided the Esanilla Formation into lower, middle, and upper members, whereas Dreyer et al. (1993) divide the Esanilla Formation into a Mondot and Olson member based on depositional environments, with the Mondot member being a transitional unit between the underlying deltaic Sobrarbe Formation and the fully fluvial Olson member (Figs. 2.1, 2.2, 2.3). The focus of this study is on the Mondot member of the Esanilla Formation.

The Mondot Member of the Esanilla Formation varies in thickness laterally across the Ainsa Basin. The unit is ~ 400 meters thick in the axis of the basin and thins to ~ 200 meters thick toward the margins of the basin (Bentham et al., 1992). Paleocurrents collected from the Mondot member are to the NW-NNW, consistent with the coevally deposited Sobrarbe Formation to the north (Bentham et al., 1992, Dreyer et al., 1999; Moss-Russell, 2009; Pyles et
The base of the Mondot Member consists of regressive and transgressive delta-plain strata with shallow marine fauna, which transitions upwardly to low-sinuosity fluvial channels, fine-grained overbank deposits, and red-matrix, quartz-pebble conglomerates (Bentham et al., 1992). Fine-grained overbank strata were deposited adjacent to channel deposits. Overbank strata are highly burrowed and range in color from purple to orange to green. Based on biostratigraphic and magnetostratigraphic data, rates of sediment accumulation (undecomposed) vary upwardly (Bentham et al., 1992; Mochales et al., 2012) and laterally across the basin from ~ 12-57 cm k/yr, with the slower rates at the margins of the basin and faster rates at the axis of the basin (Bentham et al., 1992).

### 2.2.2 Structure

From the late Cretaceous to Miocene periods, the Iberian micro-plate and the Eurasian plate converged, forming an asymmetric, doubly-vergent fold and thrust belt that strikes east-west and is associated with foreland basins to the north and south of its axis, the Northern and South Pyrenean Central Thrust Systems respectively (Fig. 2.1; Seguret et al., 1984; Puigdefabregas et al., 1986; Remacha and Fernandez, 2003; Munoz et al., 1992; Fernandez et al., 2004; Pickering & Corregidor, 2005). The Ainsa Basin is located in the South Pyrenean Central Thrust System (Fig. 2.1). From upper Maastrichtian to Paleocene, basin inversion became the primary mode of deformation in South Pyrenean Central Thrust System, followed by thin-skinned thrusting in the Early Eocene (Puigdefabregas et al., 1992; Fernandez et al., 2004). The South Pyrenean Central Thrust System is subdivided into upper, middle, and lower thrust systems (Puigdefabregas et al., 1992) that form a piggyback sequence, with the upper thrust system being displaced southward by the lower thrust systems. The Ainsa Basin is located in the
upper thrust system and is part of a larger Tremp-Ainsa-Jaca piggyback basin (Fig. 2.1).

The Tremp-Ainsa-Jaca basin is an elongate foreland basin in which the paleoshoreline opened to the west toward the Atlantic Ocean. The main sediment transport direction was from east to west. The Jaca-Ainsa-Tremp basin formed during the Lower Eocene, due to the advance of the Montsec thrust sheet (Fig. 2.1). In the Middle Eocene, the Tremp-Ainsa-Jaca basin was integrated into the hanging wall of the Gavarnie-Sierras-Exteriores thrust sheet and became a piggyback sub-basin (Figs. 2.1, 2.3; Fernandez et al., 2004). The Tremp-Ainsa-Jaca basin was then divided into three smaller basins due to deformation from an oblique fold and thrust system, the Garvanie-Sierra Marginales Thrust Sheet (Fig. 2.1; Fernandez, 2004).

The Ainsa Basin is bounded by five syndepositionally active structures (Figs. 2.1, 2.2; Dreyer et al., 1999; Fernandez et al., 2004; Farrel et al., 1987; Hoffman, 2009): (1) the Boltaña Anticline to the west, (2) the Mediano Anticline to the east, (3) the Añisclo Anticline to the north, (4) the Cotiella thrust to the northwest, and (5) the Montsec thrust to the south. The Ainsa Basin also contains several syndepositionally active intrabasinal structures including the Arcusa and Olson Anticlines, which are located in the southern half of the basin (Figs. 2.1, 2.2; 2.3; Garrido-Megias, 1968; Garrido-Megias, 1973; Bentham and Burbank, 1996; Dreyer et al., 1999; Fernandez et al., 2012). In the northern half of the Ainsa Basin, the axis of the basin is the Buil syncline (Figs. 2.2, 2.3, 2.4; Garrido-Megias, 1968; Garrido-Megias, 1973; Fernandez et al., 2004). The Olson Anticline subdivides the southern half of the Ainsa Basin into two synclines (Fig. 2.2; Garrido-Megias, 1968; Garrido-Megias, 1973; Fernandez et al., 2004; Fernandez et al., 2012): (1) the Buil Syncline is located in the eastern half of the basin between the Olson and Mediano Anticlines; and (2) the Arcusa Syncline is located in the western half of the basin between the Olson and Boltaña Anticlines.
This chapter is focused on the structures that were active during deposition of the Campodarbe Group. These structures include (Figs. 2.2, 2.3, 2.4; Dreyer et al., 1999; Fernandez et al., 2004; Farrel et al., 1987; Hoffman, 2009; Fernandez et al., 2012): (1) the Mediano Anticline, (2) the Boltaña Anticline, (3) the Olson Anticline, (4) the Arcusa Anticline, the (5) Olson Anticline, (6) the Arcusa Syncline, and (7) the Buil Syncline. A brief description of these structures and timing of growth during deposition of the Hecho Group is given below. Evidence for growth during deposition of the Campodarbe Group are discussed in the later sections.

2.2.2.1 Mediano Anticline

The Mediano Anticline is located along the eastern side of Ainsa Basin (Figs. 2.2, 2.3, 2.4). The fold is ~ 15 km long and its axis trends roughly north-south and plunges 7° to the north (Poblet et al., 1998; Hoffman, 2009). The southern region of the fold, near the town of Simitier (Figs. 2.3, 2.4), exhibits the greatest structural relief (~ 3000 m) and shortening (Poblet et al., 1998; Fernandez et al., 2012). The fold is tight and asymmetric with a minimum interlimb angle of 45°. The eastern fold limb is overturned, dips approximately 60° to the northwest, and is cut by a steep northeast-southwest oriented, northwest dipping thrust fault (Fig. 2.3). The Mediano Anticline is interpreted as a detachment fold (Poblet et al. 1998).

The Ainsa formation is the oldest unit that thins onto the Mediano Anticline and has an unconformable lower contact has an angular relationship with subjacent units as well as pregrowth strata in the axis of the Mediano Anticline (Poblet et al., 1998; Hoffman, 2009). Therefore, growth of the Mediano Anticline is interpreted to have initiated during deposition of the Ainsa formation (Poblet et al., 1998; Hoffman, 2009). Evidence of sustained activity of the Mediano Anticline is present in the Morillo Formation (Moody et al., 2012). The lower bounding
surface of the Morillo Formation is an angular unconformity that locally forms a canyon at Semitier, indicating growth of the Mediano Anticline during deposition of the Morillo Formation (Setiawan, 2009; Hoffman, 2009).

2.2.2.2 Boltaña Anticline

The Boltaña Anticline is located along the western side of the Ainsa Basin (Figs. 2.2, 2.3, 2.4). The fold is ~ 25 km long and its axis trends north-south. The Boltaña Anticline is characterized by a near horizontal fold axis except at its southern termination where it plunges 8° to the south (Fernandez et al., 2004; Hoffman, 2009). The eastern limb of the Boltaña Anticline dips more shallowly (20°-30°) than the western limb, which is nearly vertical (70°-80°) and has > 2 km of structural relief (Fernandez et al., 2004). The Boltaña Anticline is interpreted as a fault-propagation fold from a north-south oriented, west dipping blind thrust fault (Fig. 2.3; Fernandez, 2004).

The Banaston and Ainsa formations onlap lower Eocene carbonates of the backlimb of the Boltaña Anticline (Hoffman, 2009) indicating that a paleo-high was present at this location (Hoffman, 2009). The convergence and northward deflection of turbidite channels of the Morillo, and Guaso formation in proximity to the Boltaña Anticline indicate that channel architecture and stacking is directly affected by Boltaña structural growth (Setiawan, 2009; Hoffman, 2009; Moody et al., 2012; Gordon, 2013). Additionally, mass transport deposits present within the Morillo and Guaso formations along the western basin margin formed due to slope instabilities caused possibly by increased gradient associated with growth on the Boltaña Anticline (Hoffman, 2009; Moody et al., 2012; Gordon, 2013).
2.2.2.3 Olson Anticline

The Olson Anticline is a north-south oriented structure located in the center of the southern half of the Ainsa Basin (Fig. 2.3). The fold is ~ 10 km long and terminates abruptly at its northern and southern limits. The fold is generally symmetrical with both limbs dipping ~ 20°-25°, and has ~ 1 km of structural relief (Fernandez et al., 2012). The Olson Anticline is interpreted as a detachment fold from a north-south oriented, west dipping blind thrust fault (Fernandez et al., 2012).

At this location, the lower Hecho group is part of the pregrowth strata (Fig. 2.3; Munoz et al., 2013) whereas the upper Hecho Group onlaps the eastern and western flanks of the Olson Anticline indicating growth of the Olson Anticline (Fig. 2.3; Munoz et al., 2013). The surface expression of the Olson anticline is covered by strata of the Sobrarbe and Escanilla Formations. However, the anticline is clearly imaged on seismic data (Fig. 2.3; Munoz et al., 2013).

2.2.2.4 Arcusa Anticline

The Arcusa Anticline is a north-south oriented structure located in the southwestern part of the Ainsa Basin on the eastern flank of the Arcusa Syncline and is ~ 4 km long (Figs. 2.2, 2.3, 2.4). The fold plunges to the north-northwest at 8° (Hoffman, 2009). The eastern limb dips more steeply (~ 20°-25°) relative to the western limb (~10°) (Hoffman, 2009). The Arcusa Anticline is interpreted as a fault-propagation fold from a north-south oriented, west dipping blind thrust fault (Soto and Casas, 2001).

Lower R-T cycles of the Sobrarbe Formation exhibit no thickness changes relative to proximity of the Arcusa Anticline (Moss-Russell, 2009; Silalahi, 2009). However, Dreyer et al. (1999) recognized wedge geometries in the upper R-T cycles of the Sobrarbe Formation that thin
westward towards the axis of the Arcusa Anticline and interpreted them to be deposited coeval with the onset of growth of the Arcusa Anticline. The Arcusa Anticline was only active in the Late Lutetian to Early Bartonian during deposition of the Sobrarbe Formation (Fig. 2.1; Soto and Casas, 2001).

2.2.2.5 Buil Syncline

The Buil Syncline is a north-south oriented structure located in the center of the Ainsa Basin and plunges 10° to the south (Figs. 2.2, 2.3, 2.4). To the north, the Buil Syncline splays around the Añisclo Anticline into the Buerba and San Vicente synclines (Fig. 2.2). To the south, the axial trace of the Buil Syncline is displaced to the east due to growth of the Olson anticline (Figs. 2.2, 2.4; Fernandez et al., 2012). The western limb of the Buil Syncline dips 25°-30°, and the eastern limb dips 20°-25° (Fernandez et al., 2012).

2.2.2.6 Arcusa Syncline

The Arcusa Syncline is a north-south oriented structure located on the western flank of the southern part of the Ainsa Basin between the Olson and Boltaña Anticlines (Figs. 2.2, 2.3, 2.4; Fernandez et al., 2012; Munoz et al., 2013). The surface expression of the Arcusa Syncline is covered by strata of the Sobrarbe and Escanilla Formations. However, the syncline is clearly imaged on seismic data (Fig. 2.3; Munoz et al., 2013).

2.3 Data and Methods

The goals of this study are to evaluate the structure-stratigraphic interactions between the Sobrarbe Formation and coevally deposited Mondot Member of the Escanilla Formation and the
following structures: Mediano Anticline, Boltaña Anticline, Olson Anticline, Arcusa Anticline, Olson Syncline, and Arcusa Syncline. Data used in this study include: (1) a geologic map that documents the aerial distribution and character (i.e. conformable or unconformable) of the boundaries of formations and R-T cycles in the Sobrarbe and Esanilla Formations, and paleocurrent measurements (Fig. 2.4); (2) 27 detailed stratigraphic columns totaling > 6.5 km in thickness that document lithology, grain-size, physical sedimentary structures, and stratal boundaries at centimeter-scale resolution (Fig. 2.4; Appendix B); and (3) interpreted photopanels used to document key strata surfaces and the distribution of fluvial, deltaic, and marine strata and the location of stratal boundaries (Figs. 2.5, 2.7). These data were used to construct regional stratigraphic cross sections of the west (B-B'; Fig. 2.6) and east (C-C'; Fig. 2.8) sides of the basin. These cross sections document the distribution of fluvial, coastal plain/deltaic, and marine strata in the Ainsa Basin and surfaces that were used to correlate time-stratigraphic units across the basin. The northern part of the western cross section (B-B'; Fig. 2.6) is oriented parallel to the average sediment transport direction and is therefore a depositional-dip oriented profile, whereas the southern part of the cross section is oriented nearly perpendicular to the average sediment transport direction and is therefore a depositional-strike oriented profile (Fig. 2.6).

The eastern cross section (C-C'; Fig. 2.8) is oriented parallel to the average sediment transport direction and is therefore a depositional-dip oriented profile. The key stratigraphic surfaces depicted on the cross sections and geologic map are the Guaso-Sobrarbe contact and the boundaries between the six R-T cycles of the Sobrarbe and Esanilla Formations (Figs. 2.6, 2.8). Multiple datums were used in the construction of the cross sections and each R-T cycle contains its own datum based on the following criteria. The datum for the northern part of the cross sections are surfaces that separate sand-rich deltaic deposits from marine mudstones (flooding
surfaces) on the shelf, whereas the datum for the southern part of the cross sections is a stratigraphic surface that records the interface between the progradational and retrogradational units of each R-T cycle. The lowermost surface on the cross sections is a regionally continuous, black, organic-rich shale horizon that is interpreted as a condensed section and is the genetic boundary between the underlying Guaso Formation and the overlying Sobrarbe Formation (Silalahi, 2009, Moss-Russell, 2009; Hoffman, 2009). This bed was not used as a datum because it is interpreted to reflect the inherited shelf-to-basin profile created during deposition of the Guaso Formation and is therefore not a geometrically flat time surface (Pyles et al., in review, Appendix A). The uppermost surface of the cross section is the top of Cycle 6, which is the top of the Mondot Member of the Escanilla Formation. This boundary represents a shift in paleocurrent directions from predominantly north to predominantly west and ultimately to the south due to regional tectonic uplift of the Cotiella thrust sheet to the northeast (Bentham et al., 1992, Dreyer et al., 1999).

Strata within the Sobrarbe and Escanilla Formations is divided into three depositional settings on the basis of lithological characteristics (Fig. 2.9): (1) fluvial, (2) deltaic, and (3) marine. For brevity, photographic examples and descriptive characteristics of depositional settings are presented in Figure 2.9 and Table 2.1, respectively. The stratigraphic architecture of these settings is discussed in greater detail in Chapters 3, 4, and 5 of this dissertation.

Within each R-T cycle, variations in the thickness of each cycle, thickness of channel and mouthbar deposits, and net-sand content are documented. A relationship between cycle thickness, size of channels and mouth bars, and net-sand content is evident. Fluvial and deltaic strata associated with the thickest part of the cycle contain characteristically thicker channels and mouth bars and higher net-sand content compared to those units associated with thinner parts of
the cycle (Fig. 2.10). The locations where the cycles are thickest are interpreted as the axes of the system. The locations where the cycles are thinnest are interpreted as the margins of the system.

2.4 Structure-Stratigraphic Interactions in R-T Cycles

This study documents the stratigraphic architecture of the six R-T cycles of the Sobrarbe and Escanilla Formations in order to evaluate the structure-stratigraphic interactions during deposition of the Sobrarbe and Escanilla Formations. Data used to address this goal are derived from the geologic map, photopanels, and from the stratigraphic columns. This section presents both longitudinal (proximal to distal; parallel to mean sediment transport direction) and lateral (axis to margin of basin; perpendicular to mean sediment transport direction) data for each individual R-T cycle. Data for only the fluvial and deltaic depositional settings of the cycles are presented below.

2.4.1 Cycle 1

Deltaic strata of Cycle 1 are exposed on the western and eastern flanks of the Ainsa Basin (Fig. 2.11). Marine strata are exposed north of the deltaic strata around the northern perimeter of the basin (Fig. 2.11). In the western part of the Ainsa Basin, deltaic strata onlap pregrowth strata of the Boltaña Anticline. Near the town of Eripol (Fig. 2.11), paleocurrent directions within deltaic strata are to the northwest (vector mean=319°). At this location, the thickness of Cycle 1 is 105 m, the average thickness of delta mouth bars is 2.40 m, and the average net-sand content is 0.38 (Fig. 2.11). The shelf edge, the interface between shelf and slope strata, is exposed northwest of the town of Eripol. In the eastern part of the Ainsa Basin, northeast of the town of La Mata (Fig. 2.11), paleocurrent directions within deltaic strata are to the northwest (vector
mean=305°; Fig. 2.11). At this location, the thickness of Cycle 1 is 60 m, the average thickness of delta mouth bars is 2.20 m, and the average net-sand content is 0.28 (Fig. 2.11). The shelf edge in this area is exposed north of the town of La Mata. East of the town of La Mata, the base of Cycle 1 is unconformable with older marine mudstone strata of the Guaso Formation (Fig. 2.11). The base of Cycle 1 is conformable with the underlying Guaso Formation in all other locations except east of the town of La Mata. The average strike of the shelf edge across the basin for Cycle 1 is 074°.

Data presented above are used to make the following interpretations of Cycle 1. Sediment deposited in Cycle 1 is sourced from the southeast. Cycle 1 onlaps pregrowth strata of the Boltaña Anticline indicating that the structure was a topographic high during deposition. The thickest part of the Cycle 1, as well as the thickest mouth bar deposits and highest net-sand content is located in the axis of the Arcusa Syncline near the town of Eripol (Fig. 2.11) and is interpreted to indicate that the axis of the system was focused by this structure. The shelf edge is located north of the town of Eripol (Fig. 2.11). Cycle 1 remains uniform in thickness near the present day location of the Arcusa Anticline indicating that the structure was not yet active. Cycle 2 is the lowest cycle exposed in the center of the basin however, using seismic data, Munoz et al. (2013) document a structural high, the Olson Anticline, located at this location (Figs. 2.3, 2.11). The thickest part of the Cycle 1, as well as the thickest mouth bar deposits and highest net-sand content is located on the east side of the Olson Anticline, in the axis of the Buil Syncline by the town of La Mata (Fig. 2.11) and is interpreted to indicate that the axis of the system was focused by this structure. The shelf edge is located north of the town of La Mata due west of the town of Simitier (Fig. 2.11). An angular unconformity located at the western flank of the Mediano Anticline indicates growth of the structure during deposition of the early phase of
Cycle 1. In summary, during deposition of Cycle 1, the Boltaña, Olson, and Mediano Anticlines were active and the axes of the fluvial system were located in the Arcusa and Buil Synclines, where the cycle and mouth bars are thickest and the net-sand content is highest.

2.4.2 Cycle 2

Cycle 2 is exposed around the perimeter of the basin (Fig. 2.12). Fluvial strata of Cycle 2 are locally exposed in the southern part of the basin (Fig. 2.12). Deltaic strata of Cycle 2 are exposed to the north of the fluvial strata across the entire Ainsa Basin and marine strata are exposed to the north of the deltaic strata (Fig. 2.12). On the western flank of the Arcusa Syncline between the towns of Almazorre and Eripol (Fig. 2.12), paleocurrent directions are to the north-northwest (vector mean=344°). At this location, the thickness of Cycle 2 is 96 m, fluvial channels are on average 2.00 m thick, and the average net-sand content is 0.15 (Fig. 2.12). Cycle 2 thickens to the north to 110 m, where the average thickness of the coevally deposited delta mouth bars is 2.10 m, and the average net-sand content is 0.48 (Fig. 2.12). Farther east, near the town of Mondot (Fig. 2.12), paleocurrent directions are to the north-northwest (vector mean=347°). At this location, Cycle 2 thickens to 110 m, fluvial channels are on average 4.10 m thick, and the average net-sand content is 0.44 (Fig. 2.12). North of these fluvial channels, the cycle is thickest (124 m), the average thickness of the coevally deposited delta mouth bars is 3.10 m, and the average net-sand content is 0.69 (Fig. 2.12). The shelf edge for Cycle 2 is exposed north of the town of Mondot (Fig. 2.12). Further east on the eastern flank of the Arcusa Syncline, Cycle 2 thins to 95 m, fluvial channels are on average 2.41 m thick, and the average net-sand content is 0.28, (Fig. 2.12). Due to modern erosion, no coevally deposited deltaic outcrops are located north of these fluvial channels. Further east, within the Buil Syncline, north of the town of
La Mata (Fig. 2.12), paleocurrent directions are to the northwest (vector mean=326°). At this location, the thickness of Cycle 2 is 100 m, fluvial channels are on average 2.50 m thick, and the average net-sand content is 0.43 (Fig. 2.12). North of these fluvial channels, Cycle 2 thins to 56 m, the average thickness of the coevally deposited delta mouth bars is 2.50 m, and the average net-sand content is 0.43 (Fig. 2.12). In this area, the shelf edge of Cycle 2 is exposed north of the town of Castejon de Sobrarbe (Fig. 2.12). The average strike of the shelf edge across the basin for Cycle 2 is 070°.

Data presented above are used to make the following interpretations of Cycle 2. Sediment deposited in Cycle 2 is sourced from the southeast. The thickest part of the Cycle 2 on the west side of the basin, as well as the thickest channel and mouth bar deposits and highest net-sand content is located near the town of Mondot (Fig. 2.12) and is interpreted to indicate that the axis of the system was focused by subsidence associated this structure. The axis of Cycle 2 in the Arcusa Syncline is located 3 km east of the axis of Cycle 1, which is interpreted to indicate growth of the Boltaña Anticline (Fig. 2.12). The shelf edge of Cycle 2 is located north of the town of Mondot (Fig. 2.12), which is 4 km north of the shelf edge of Cycle 1 (Figs. 2.6, 2.12) indicating sediment supply exceeded the rate that accommodation was generated. Cycle 2 thickens near the present day location of the Arcusa Anticline indicating that the structure was not yet active. Cycle 2 thins over the Olson Anticline and thickens again within the Buil Syncline indicating growth of the Olson Anticline. The thickest part of the Cycle 2 on the east side of the basin, as well as the thickest channel and mouth bar deposits and highest net-sand content is located in the axis of the Buil Syncline west of the town of La Mata (Fig. 2.12) and is interpreted to indicate that the axis of the system was focused by subsidence associated with this structure. The axis, or the thickest part of Cycle 2 in the Buil Syncline, is located 2 km west of the axis of
Cycle 1, possibly indicating growth of the Mediano Anticline (Fig. 2.12). The shelf edge of Cycle 2 is located north of the town of La Mata (Fig. 2.12), which is 3.5 km north of the shelf edge of Cycle 1 (Figs. 2.8, 2.12) indicating sediment supply exceeded the rate that accommodation was generated. In summary, during deposition of Cycle 2, the Boltaña, Olson, and Mediano Anticlines were active and the axes of the fluvial system were located in the Arcusa and Buil synclines, where the cycle, channels, and mouth bars are thickest and the net-sand content is highest.

2.4.3 Cycle 3

Cycle 3 is exposed in multiple areas across the basin (Fig. 2.13). Fluvial strata of Cycle 3 are exposed across the entire southern part of the Ainsa Basin (Fig. 2.13). Deltaic strata of Cycle 3 are exposed in several discontinuous outcrops in the middle of the basin and marine strata are exposed in the northern part of the basin (Fig. 2.13). On the western flank of the Arcusa Syncline, near the town of Eripol (Fig. 2.13), paleocurrent directions are to the north-northwest (vector mean=346°). At this location, the thickness of Cycle 3 is 58 m, fluvial channels are on average 1.90 m thick, and the average net-sand content is 0.39 (Fig. 2.13). Due to modern day erosion, no coevally deposited deltaic outcrops are located north of these fluvial channels. East of the town of Mondot, paleocurrent directions are to the north-northwest (vector mean=331°). At this location, Cycle 3 thickens to 68 m, fluvial channels are 3.93 m, and the average net-sand content is 0.47 (Fig. 2.13). North of these fluvial channels, Cycle 3 thins to 49 m, the average thickness of the coevally deposited delta mouth bars is 2.20 m, and the average net-sand content is 0.59 (Fig. 2.13). The shelf edge of Cycle 3 is exposed north of the town of Arcusa (Fig. 2.13). Farther east, in the Buil Syncline (Figs. 2.4, 2.13), near the town of Castejon de Sobrarbe (Fig,
2.13), paleocurrent directions are to the northwest (vector mean=319°). At this location, Cycle 3 is thickest (129 m), fluvial channels are on average 3.00 m thick, and the average net-sand content is 0.45 (Fig. 2.13). North of these fluvial channels, the thickness of Cycle 3 is 130 m, the average thickness of the coevally deposited delta mouth bars is 3.10 m, and the average net-sand content is 0.54, (Fig. 2.13). In this area, the shelf edge of Cycle 3 is exposed north of the town of Camporotuno (Fig. 2.13). The average strike of the shelf edge across the basin for Cycle 3 is 065°. East of the town of La Mata, the base of Cycle 3 is unconformable with fluvial strata of Cycle 2 (Fig. 2.13). The base of Cycle 3 is conformable with the underlying Cycle 2 in all other locations except east of the town of La Mata.

Data presented above are used to make the following interpretations of Cycle 3. Sediment deposited in Cycle 3 is sourced from the southeast. The thickest part of the Cycle 3 on the west side of the basin, as well as the thickest channel and mouth bar deposits and highest net-sand content is located in the axis of the Arcusa Syncline east of the town of Mondot (Fig. 2.13) and is interpreted to indicate that the axis of the system was focused by subsidence associated with this structure. The axis of Cycle 3 in the Arcusa Syncline is located 1 km east of the axis of Cycle 2 indicating growth of the Boltaña Anticline and/or initiation of growth of the Arcusa Anticline (Fig. 2.13). The shelf edge of Cycle 3 is located north of the town of Arcusa (Fig. 2.13), which is 4 km north of the shelf edge of Cycle 2 (Figs. 2.6, 2.13) indicating sediment supply exceeded the rate that accommodation was generated. Cycle 3 thins near the present day location of the Arcusa Anticline indicating that the structure was active. Cycle 3 is 48 m thinner in the axis of the Arcusa Syncline than Cycle 2 indicating a decrease in accommodation. This decrease in accommodation is attributed to filling of the Arcusa Syncline with sediment and growth of the Arcusa Anticline. Cycle 3 thins over the Olson Anticline and thickens again within the Buil
Syncline indicating growth of the Olson Anticline. The thickest part of the Cycle 3 on the east side of the basin, as well as the thickest channel and mouth bar deposits and highest net-sand content is located in the axis of the Buil Syncline near the town of Castejon de Sobrarbe (Fig. 2.13) and is interpreted to indicate that the axis of the system was focused by subsidence associated with this structure. The axis, of Cycle 3 in the Buil Syncline, is located 1 km west of the axis of Cycle 2 indicating growth of the Mediano Anticline (Fig. 2.13). The shelf edge of Cycle 3 is located north of the town of Castejon de Sobrarbe (Fig. 2.13), which is 1.5 km north of the shelf edge of Cycle 2 (Figs. 2.8, 2.13) indicating sediment supply exceeded the rate that accommodation was generated. An angular unconformity located at the eastern flank of the Buil Syncline indicates growth of the Mediano Anticline during deposition of either the later phases of deposition of Cycle 2 or early phases of deposition of Cycle 3. In summary, during deposition of Cycle 3, the Boltaña, Arcusa, Olson, and Mediano Anticlines were active and the axes of the fluvial system were located in the Arcusa and Buil synclines, where the cycle, channels, and mouth bars are thickest and the net-sand content is highest.

2.4.4 Cycle 4

Cycle 4 is exposed in multiple areas across the basin (Fig. 2.14). Fluvial strata of Cycle 4 are exposed across the entire southern part of the Ainsa Basin (Fig. 2.14). Deltaic strata of Cycle 4 are exposed in several discontinuous outcrops to the north of the towns of Arcusa and Castejon de Sobrarbe within the middle of the basin and marine strata are exposed in the northern part of the basin (Fig. 2.14). On the western flank of the Ainsa Basin near the town of Eripol (Fig. 2.14), paleocurrent directions are to the north-northwest (vector mean=342°). At this location, the thickness of Cycle 4 is 65 m, fluvial channels are on average 2.65 m thick, and the average net-
sand content is 0.31 (Fig. 2.14). Due to modern day erosion, no coevally deposited deltaic outcrops are located north of these fluvial channels. In the center of the basin (Fig. 2.14), Cycle 4 thickens to 104 m, fluvial channels are 3.30 m thick, and the net-sand content is 0.34 (Fig. 2.14). North of these fluvial channels, paleocurrent directions are to the northwest (vector mean=322°). At this location, Cycle 4 thins to 100 m, the average thickness of the coevally deposited delta mouth bars is 2.29 m, and the average net-sand content is 0.45 (Fig. 2.14). The shelf edge of Cycle 4 in this area is exposed northeast of the town of Castellazo (Fig. 2.14). Northeast of the town of Arcusa, the base of Cycle 4 is unconformable with deltaic strata of Cycle 3 (Fig. 2.14). The base of Cycle 4 is conformable with the underlying Cycle 3 in all other locations. On the east side of the basin north of the town of La Mata (Fig. 2.14), paleocurrent directions are to the north-northwest (vector mean=348°). At this location, the Cycle 4 is thickest (125 m), fluvial channels are on average 3.44 m thick, and the average net-sand content is 0.49 (Fig. 2.14). North of these fluvial channels, the thickness of Cycle 4 is 104 m, the average thickness of the coevally deposited delta mouth bars is 2.60 m, and the average net-sand content is 0.43 (Fig. 2.14). In this area, the shelf edge of Cycle 4 is exposed south of the town of Santa Maria de Buil (Fig. 2.14). The average strike of the shelf edge across the basin for Cycle 4 is 064°.

Data presented above are used to make the following interpretations of Cycle 4. Sediment deposited in Cycle 4 is sourced from the southeast. On the west side of the basin, the cycle is thinnest. The thickness of Cycle 4 increases to the east. The thickest part of the Cycle 4, as well as the thickest channel and mouth bar deposits and highest net-sand content is located in the axis of the Buil Syncline west of the town of La Mata (Fig. 2.14) and is interpreted to indicate that the axis of the system was focused by subsidence associated with this structure. An angular unconformity located northeast of the town of Arcusa indicates growth of the Arcusa Anticline.
during deposition of either the later phases of deposition of Cycle 3 or early phases of deposition of Cycle 4. The axis of Cycle 4 is located 3 km east of the western axis of Cycle 3 also indicates growth of the Boltaña and Arcusa Anticlines (Fig. 2.14). No changes in thickness at the location of the Olson Anticline indicates the Olson Anticline is not active during deposition of Cycle 4. The thickening of Cycle 4 to the east as well as the increase in channel and mouth bar thickness and increase in net-sand content indicate the Arcusa Syncline is filled with sediment and only one axis of the system is present within Cycle 4, the Buil Syncline. The axis of Cycle 4 in the Buil Syncline is located in the same location as the axis of Cycle 3 indicating no growth of the Mediano Anticline (Fig. 2.14). The shelf edge is located south of the town of Santa Maria de Buil (Fig. 2.14), which is 1.5 km north of the shelf edge of Cycle 3 (Figs. 2.6, 2.14) indicating sediment supply exceeded the rate that accommodation was generated. In summary, during deposition of Cycle 4, the Boltaña, and Arcusa Anticlines were active and the axis of the fluvial system was located in the Buil syncline, where the cycle, channels, and mouth bars are thickest and the net-sand content is highest.

2.4.5 Cycle 5

Cycle 5 is exposed in multiple areas across the basin (Fig. 2.15). Fluvial strata of Cycle 5 are exposed across the entire southern part of the Ainsa Basin (Fig. 2.15). Deltaic strata of Cycle 5 are exposed only in the northern part of the field area around the town of Santa Maria de Buil (Fig. 2.15). Due to modern erosion, no marine strata for Cycle 5 are located. On the western flank of the Ainsa Basin, east of the town of Eripol (Fig. 2.15), paleocurrent directions are to the north (vector mean=002°). At this location, the thickness of Cycle 5 is 98 m, fluvial channels are on average 3.23 m thick, and the average net-sand content is 0.11 (Fig. 2.15). No coeally
deposited deltaic outcrops are located north of these fluvial channels. East of the town of Mondot the thickness of Cycle 5 thickens to 105 m, fluvial channels are on average 6.11 m thick, and the average net-sand content is 0.45 (Fig. 2.15). To the north of these fluvial channels, on the west side of the town of Santa Maria de Buil (Fig. 2.15), paleocurrent directions are to the northwest (vector mean=307°). At this location, Cycle 5 is thickest (124 m), the average thickness of the coevally deposited delta mouth bars is 5.50 m, and the average net-sand content is 0.45 (Fig. 2.15). On the eastern flank of the Ainsa Basin, northwest of the town of La Mata, paleocurrent directions are to the north-northwest (vector mean=349°). At this location, Cycle 5 is 120 m thick, fluvial channels are on average 5.20 m thick, and the average net-sand content is 0.41 (Fig. 2.15). To the north of these fluvial channels, Cycle 5 thins to 118 m, the average thickness of the coevally deposited delta mouth bars is 4.32 m, and the average net-sand content is 0.73 (Fig. 2.15). In this area, the shelf edge of Cycle 5 is exposed northwest of the town of Santa Maria de Buil (Fig. 2.15). The average strike of the shelf edge across the basin for Cycle 5 is 045°. The shelf-edge of Cycle 5 is composed of multiple, large-scale slides due to shelf-edge failure. Callot et al. (2008) documented that these slides affected and removed up to 15% of the delta front strata. The average trend for the slide scars is 55° with northwest dips of 20°-40° (Callot et al., 2008).

Data presented above are used to make the following interpretations of Cycle 5. Sediment deposited in Cycle 5 is sourced from the southeast. On the west side of the basin, the cycle is thinnest. The thickness of Cycle 5 increases to the east. The thickest part of the Cycle 5, as well as the thickest channel and mouth bar deposits and highest net-sand content is located in the axis of the Buil Syncline northwest of the town of La Mata (Fig. 2.15) and is interpreted to indicate that the axis of the system was focused by subsidence associated with this structure. The axis of
Cycle 5 is located 1 km west of the axis of Cycle 4 indicating growth of the Mediano Anticline (Fig. 2.15). The shelf edge is located north of the town of Santa Maria de Buil (Fig. 2.15), which is 5 km north of the shelf edge of Cycle 4 (Figs. 2.6, 2.15) indicating sediment supply exceeded the rate that accommodation was generated. The change in paleocurrent directions between the fluvial (predominantly north) and deltaic strata (predominantly northwest) and the change in the strike of the shelf edge between Cycles 4 and 5 are attributed to deflection around the Arcusa Anticline. There is an increase in the thickness of Cycle 5 relative to Cycle 4 as well as an increase in the sizes of channels and mouth bars are interpreted to indicate an increase in sediment supply and discharge in the basin. The large slides located at the shelf edge are attributed to high rates of sediment loading and are triggered by growth of the different structures in the basin. In summary, during deposition of Cycle 5, the Boltaña, Arcusa, and Mediano Anticlines were active and the axis of the fluvial system was located in the Buil syncline, where the cycle, channels, and mouth bars are thickest and the net-sand content is highest.

2.4.6 Cycle 6

Cycle 6 is composed of only fluvial strata and is exposed in multiple areas across the basin (Fig. 2.16). Fluvial strata of Cycle 6 are exposed across the entire southern part of the Ainsa Basin, near the town of Santa Maria de Buil, and in the northwestern part of the Ainsa Basin on top of the Boltaña Anticline (Fig. 2.16). On the western flank of the Ainsa Basin north of the town of Hospital (Fig. 2.16), paleocurrent directions are to the north (vector mean=011°). At this location, the thickness of Cycle 6 is 129 m, fluvial channels are on average 5.90 m thick, and the average net-sand content is 0.33 (Fig. 2.16). Within the middle of the Ainsa Basin, east of the town of Olson (Fig. 2.16), Cycle 6 is thickest (140 m), fluvial channels are on
average 8.20 m thick, and the average net-sand content is 0.39 (Fig. 2.16). On the eastern flank of the Ainsa Basin, south of the town of La Mata (Fig. 2.16), paleocurrent directions are to the northwest (vector mean=328°). At this location, Cycle 6 thins to 131 m, fluvial channels are on average 5.60 m thick, and the average net-sand content is 0.40 (Fig. 2.16). North, of the town of Santa Maria de Buil (Fig. 2.16), paleocurrent directions are to the west-northwest (vector mean=299°). At this location, Cycle 6 thins to 125 m, fluvial channels are on average 7.90 m thick, and the average net-sand content is 0.40 (Fig. 2.16). West-northwest of the town of Santa Maria de Buil, the base of Cycle 6 is unconformable with pregrowth carbonate strata of the Boltaña Anticline (Fig. 2.16).

Data presented above are used to make the following interpretations of Cycle 6. Sediment deposited in Cycle 6 is sourced from the southeast and is composed of only fluvial strata. On the west side of the basin, the cycle is thinnest. The thickness of Cycle 5 increases to the east. The thickest part of the Cycle 6, as well as the thickest channel and mouth bar deposits and highest net-sand content is located in the axis of the Buil Syncline west of the town of La Mata (Fig. 2.16) and is interpreted to indicate that the axis of the system was focused by subsidence associated with this structure. The thickness of Cycle 6 decreases to the east. The axis of Cycle 6 is located 1.5 km west of the axis of Cycle 5 indicating growth of the Mediano Anticline (Fig. 2.16). The lack of outcrop does not allow evaluation of the Arcusa Anticline. The shelf edge is no longer located within the Ainsa Basin, but has prograded west towards its present day location. Fluvial strata erode into the top of the Boltaña Anticline to the northwest of the town of Santa Maria de Buil indicating growth of the Boltaña Anticline during deposition of Cycle 6 (Fig. 2.16). In summary, during deposition of Cycle 6, the Boltaña and Mediano Anticlines were active and the axis of the fluvial system was located in the Buil syncline, where the cycle,
channels, and mouth bars are thickest and the net-sand content is highest.

2.5 Conclusions

This study documents for the first time variations in the nature of geologic contacts, paleocurrents, cycle thickness, channel thicknesses, and net-sand content of the Sobrarbe Formation and coevally deposited Mondot Member of the Escanilla Formation. This information is used to evaluate the structure-stratigraphic interactions between strata and the following syndepositionally active structures: Mediano Anticline, Boltaña Anticline, Olson Anticline, Arcusa Anticline, Olson Syncline, and Arcusa Syncline. The purpose of this chapter is to give greater context to Chapters 3, 4 and 5 of this dissertation.

The Boltaña Anticline was active during deposition of Cycles 1-6 (Fig. 2.17). The Arcusa Anticline was active during deposition of Cycles 2-5 (Fig. 2.17). The Olson Anticline was active during deposition of Cycles 1-3 (Fig. 2.17). The Mediano Anticline was active during deposition of Cycles 1-4, and 6 (Fig. 2.17).

During deposition of Cycles 1-3, the southern part of the Ainsa Basin, as well as the axes of the system were divided between the Arcusa and Buil Synclines. Deposition of Cycle 1-3 filled the Arcusa Syncline, while at the same time the axis of these three cycles shifted east predominantly due to growth of the Boltaña and Arcusa Anticlines. The axis of deposition for Cycles 4-6 is located within the Buil Syncline.

The northward progradation and aggradation of the shelf edge from one cycle to the next is attributed to sediment supply exceeding the rate that accommodation was generated. Changes in paleocurrent and the average strike of the shelf edge from one cycle to the next is attributed to growth of the Arcusa Anticline.
2.6 References


Gordon, G.S., Pyles, D.R., Clark, J., and Hoffman, M., in review, Stratigraphic evolution, architectural analysis, and geometric model of a ponded submarine fan: The Guaso 1
Figure 2.1 (A) Paleogeographic map documenting the location of the different thrust sheets (TS) of the Pyrenean Orogeny as well as the location of the Tremp-Ainsa-Jaca Basin (modified from Michael et al., 2014). (B) Generalized stratigraphy of the Ainsa-Jaca Basin (modified from Pickering and Bayliss, 2009). (C) Chronostratigraphic chart of the Hecho and Campodarbe Groups illustrating the different lithostratigraphic units located in the Ainsa Basin, the eustatic curves (Haq et al., 1988), and chart documenting the periods of structural growth for the different basin bounding and intrabasinal structures of the Ainsa Basin (Poblet et al., 1998; Dreyer et al., 1999; Soto and Casas, 2001; Fernandez et al., 2004; Fernandez et al., 2012). This study is focused on the Sobrarbe and Escanilla Formations.
Figure 2.2 Geologic map of the Ainsa Basin documenting the contacts between the lithostratigraphic units and bounding and intrabasinal structures (map modified from Hoffman, 2009).
Figure 2.3 Cross section through the southern part of the Ainsa Basin based on seismic data (modified from Munoz et al., 2013). See Fig. 2.2 for location of cross section.
Figure 2.4 Detailed geologic map of the field area emphasizing details of the Sobrarbe and Escanilla Formation. Contacts between the different lithostratigraphic units, bounding structures, and the Buil Syncline are labeled (map modified from Hoffman, 2009). This study is focused on the Sobrarbe and Escanilla Formations, which are formally divided into six regressive/transgressive cycles (Cycles 1-6) and are labeled on the map.
Figure 2.5 Interpreted photopanels of the Sobrarbe and Escanilla Formations on the west side of the Ainsa Basin documenting the boundaries between the six interpreted fourth-order regressive/transgressive stratigraphic cycles as defined by Moss-Russell (2009) and Pyles et al. (in review, Appendix A) as well as the location of the stratigraphic columns. (A) Photopanel taken from the town of Castellazo looking east. Photopanels taken from the town of Eripol looking north (B), east (C), and south (D). Location of photopanels shown in Fig. 2.4.
Figure 2.6 Stratigraphic cross section of Sobrarbe and Escanilla Formations constrained by geologic mapping and stratigraphic columns documenting the contact between the Guaso and Sobrarbe Formations the boundaries between the six R-T cycles defined by Moss-Russell (2009) and Pyles et al. (in review, Appendix A), and the location of successive shelf edge deltas. The southern part of the cross section is oriented along depositional strike and the northern part is oriented parallel to depositional dip. The location of cross section is shown in Fig. 2.4. Detailed measured sections are located in Appendix B.

Figure 2.7: Interpreted photopanels of the Sobrarbe and Escanilla Formations on the east side of the Ainsa Basin documenting the boundaries between the six interpreted fourth-order regressive/transgressive stratigraphic cycles as defined by Moss-Russell (2009) and Pyles et al. (in review, Appendix A) as well as the location of the stratigraphic columns. (A) Photopanel taken from above the town of Simitier looking west. (B) Photopanel taken from the town of Coscojuela de Sobrarbe looking west. (C) Photopanel taken from northwest of the town of Morillo de Tou looking south. Location of photopanels shown in Fig. 2.4.
Figure 2.8 Stratigraphic cross section of Sobrarbe and Escanilla Formations constrained by geologic mapping and stratigraphic columns documenting the contact between the Guaso and Sobrarbe Formations, the boundaries between the six R-T cycles defined by Moss-Russell (2009), the location of successive shelf edge deltas, and the shelf-edge trajectory for Cycles 1-5. The cross section is oriented parallel to depositional dip. The location of cross section is shown in Fig. 2.4. Detailed measured sections are located in Appendix A.
Figure 2.9 Photographic examples of depositional environments identified in this study. (A) Fluvial strata consist of channel, crevasse splay, and floodplain fine deposits. Deltaic strata consist of distributary channel and mouth bar deposits. (C) Marine strata consist of submarine channel, submarine lobe, marine mudstone sheet, and mass transport deposits. Characteristics are described in Table 2.1.
Figure 2.10 Cross-plots of the thickness of the R-T cycles versus thickness of channel-belts and mouth bars (A) and net-sand content (B). Both cross-plots document a positive, although weak correlation between the thickness of cycles and the thickness of channel-belts, mouth bars, and net-sand content.
Figure 2.11 Geologic map of Cycle 1 depicting some of its paleogeographic aspects including the axes of the system and location of the shelf edge. The map is constrained by the geologic map, stratigraphic columns, and paleocurrents.
Figure 2.12 Geologic map of Cycle 2 depicting some of its paleogeographic aspects including the axes of the system and location of the shelf edge. The map is constrained by the geologic map, stratigraphic columns, and paleocurrents.
Figure 2.13 Geologic map of Cycle 3 depicting some of its paleogeographic aspects including the axes of the system and location of the shelf edge. The map is constrained by the geologic map, stratigraphic columns, and paleocurrents.
Figure 2.14 Geologic map of Cycle 4 depicting some of its paleogeographic aspects including the axes of the system and location of the shelf edge. The map is constrained by the geologic map, stratigraphic columns, and paleocurrents.
Figure 2.15 Geologic map of Cycle 5 depicting some of its paleogeographic aspects including the axes of the system and location of the shelf edge. The map is constrained by the geologic map, stratigraphic columns, and paleocurrents.
Figure 2.16 Geologic map of Cycle 6 depicting some of its paleogeographic aspects including the axis of the system. The map is constrained by the geologic map, stratigraphic columns, and paleocurrents.
Figure 2.17 Chronostratigraphic chart of the Hecho and Campodarbe Groups and chart documenting the periods of structural growth for the different basin bounding and intrabasinal structures in reference to the six regressive-transgressive cycles of the Sobrarbe Formation and coevally deposited Mondot Member of the Escanilla Formation (Poblet et al., 1998; Dreyer et al., 1999; Soto and Casas, 2001; Fernandez et al., 2004; Fernandez et al., 2012).
Table 2.1 Depositional environments described in this study.

<table>
<thead>
<tr>
<th>Depositional Setting</th>
<th>Cross sectional shape in strike view</th>
<th>Common lithofacies</th>
<th>Physiographic position</th>
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<tr>
<td><strong>Fluvial</strong></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>Fluvial channel</td>
<td>concave upward base, planar top, aspect ratio ~12-52</td>
<td>cross-stratified sandstone and conglomerate</td>
<td>non-marine to proximal shelf</td>
</tr>
<tr>
<td>Crevasse splay</td>
<td>planar base, convex-upward top</td>
<td>bioturbated to structureless sandstone interbedded with bioturbated and rooted siltstone and mudstone</td>
<td>non-marine to proximal shelf</td>
</tr>
<tr>
<td>Floodplain fines</td>
<td>planar base and top</td>
<td>bioturbated and rooted varicolored mudstone</td>
<td>non-marine to proximal shelf</td>
</tr>
<tr>
<td><strong>Deltaic</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Delta mouth bar</td>
<td>planar base, convex-upward top</td>
<td>bioturbated, planar-laminated or structureless sandstone inclined in the direction of progradation</td>
<td>shelf</td>
</tr>
<tr>
<td>Distributary channel</td>
<td>concave upward base, planar top, aspect ratio ~19</td>
<td>structureless to trough and planar cross-stratified sandstone</td>
<td>proximal to distal shelf</td>
</tr>
<tr>
<td><strong>Marine</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Submarine channel</td>
<td>concave upward base, planar top</td>
<td>interbedded structureless to rippled sandstone and siltstone</td>
<td>slope</td>
</tr>
<tr>
<td>Submarine lobe</td>
<td>planar base, convex-upward top</td>
<td>structureless sandstone interbedded with planar and ripple-laminated sandstone and siltstone</td>
<td>basinfloor</td>
</tr>
<tr>
<td>Marine mudstone sheet</td>
<td>planar base and top</td>
<td>bioturbated to laminated gray mudstone</td>
<td>shelf, slope and basinfloor</td>
</tr>
<tr>
<td>Mass transport deposit</td>
<td>planar to concave upward base and irregular to planar top</td>
<td>contorted mudstone and sandstone</td>
<td>slope</td>
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3.1 Abstract

There is a paucity of data that document axis-to-margin and A/S variations in reservoir architecture of fluvial systems. The Escanilla Formation of the Ainsa Basin, Spain contains world-class outcrops of fluvial strata, which can be correlated to coevally deposited shallow marine deposits of the Sobrarbe Formation. Measurements of four consecutive regressive-transgressive cycles, with varying shelf-edge trajectories document the following. Axial strata contain a higher percentage of channel-belt elements and splay stories, thicker splay beds, thicker channel-belt elements, and a higher net-sand content than their counterparts deposited in the margin of the system. Fluvial strata associated with high A/S (high shelf-edge trajectories) contain thicker and a higher percentage of floodplain-belt elements, smaller channel-belt elements, a higher net-sand content, and channel-belt elements are thinner in relation to their genetically related floodplain-belt elements associated with their low A/S (low shelf-edge
trajectory) counterparts. Therefore, large scale stacking patterns of a depositional system can be used as a predictor of meso- (reservoir) scale characteristics. Also, subdividing populations on the basis of geological distinctions reveal important differences that have implications to reservoir models.

3.2 Introduction

Fluvial systems are important hydrocarbon reservoirs around the world (e.g. Kern River, USA; Mungaroo Formation, Australia). Fluvial stratigraphy can be divided into channel and floodplain deposits (Fig. 3.1A), each having unique reservoir properties. Channel deposits can be further divided into bars and channel fill (Fig. 3.1A). Due to the high percentage of sand, bars are the most important hydrocarbon reservoir units in subsurface reservoirs. In contrast, fluvial floodplain deposits accumulate when floodwater carrying suspended sediment overtops a river bank and spills into adjacent floodplains and their deposits can be divided into muddy (e.g. floodplain fines) and sandy deposits (e.g. crevasse splays) (Fig. 3.1A). Due to the discontinuous nature of fluvial bar deposits, the presence of crevasse-splay deposits within a fluvial reservoir can impact production strategies as these deposits can be hydrocarbon reservoirs and they can be pathways for fluid migration between channel sand bodies (e.g. bars) that are otherwise not connected (Tye, 2004; Anderson, 2005).

Both fluvial channel and floodplain deposits evolve in response to many different controls such as sediment supply, climate, discharge, tectonics, changes in base level and physiographic location where deposition occurs (Figs. 3.1B and 3.1C; e.g. Vail et al., 1977; Leeder, 1978; Bridge and Leeder, 1979; Retallack, 1986; Autin et al., 1991; Heller and Paola, 1992; Bridge and Mackey, 1993; Leeder, 1993; Wright and Mariott, 1993; Shanley and McCabe, 1994; Willis and
Behrensmeyer, 1994; Mackey and Bridge, 1995; Aslan and Blum, 1999; Kraus and Aslan, 1999; Plint et al., 2001; Weissmann et al., 2010). An understanding of these controls permits an analysis of a fluvial system’s depositional history. For example, previous work documents longitudinal changes in fluvial architecture. Proximal fluvial architecture is characterized by pebbly and sandy, multistory, amalgamated channels with little preservation of floodplain strata, whereas distal fluvial architecture is characterized by relatively smaller, finer grained channels that have little static connectivity with well preserved more heterolithic floodplain strata (e.g. Legarreta and Uliana, 1998; Holbrook et al., 2006; Nichols and Fisher, 2007; Karssenberg and Bridge, 2008; Hartley et al., 2010; Weissmann et al., 2010). Also, previous work has related floodplain architecture to accommodation. Mud-rich floodplain deposits containing mature paleosols are interpreted as being deposited in low accommodation floodplains (Fig. 3.1C), whereas more lithologically heterogeneous and sand-rich floodplain successions are interpreted as being deposited in high accommodation floodplains (Fig. 3.1C; Kraus and Aslan, 1993; Wright and Marriott, 1993; Kraus, 1997; McCarthy et al., 1997; Aslan and Autin, 1999; McCarthy et al., 1999).

Some relationships that have yet to be fully made in fluvial stratigraphy are the following. First, a better understanding of how meso- (reservoir) scale characteristics of fluvial deposits vary within a sequence stratigraphic framework (i.e. Stear, 1983; Obrien and Wells, 1986; Mjos et al., 1993; Wright and Marriott, 1993; Gibling and Bird, 1994; Shanley and McCabe, 1994; Smith and Perez-Arlucea, 1994; Tandon and Gibling, 1994; Bristow et al, 1999; Ethridge et al., 1999; Tooth, 2005). Second, there is a paucity of data that document how axis-to-margin and A/S variations relate to reservoir architecture of the complete fluvial system. Third, in general, fluvial deposits, especially floodplain sandstones, are difficult to image in seismic
images due to their small size and homogeneous grain-size. Therefore, outcrop studies that relate meso- (reservoir) scale characteristics (i.e. proportion and size of channel and floodplain deposits) to larger scale stacking patterns (i.e. shelf-edge trajectories), which are resolvable in seismic images are valuable.

This study uses exceptionally well-exposed outcrops of the Eocene Sobrarbe and Escanilla Formations in the Ainsa Basin, Spain to address some aspects of these relationships. Well-exposed, progradational deltaic deposits of the Sobrarbe Formations can be correlated to coevally deposited fluvial strata in the Escanilla Formation. As such, it is possible to quantitatively document how fluvial architecture relates to coevally deposited deltaic strata, and how variations in stratigraphic architecture vary from the axis of the fluvial system to its margin.

The goals of this study are to quantitatively document stratigraphic architecture of fluvial channel and floodplain deposits for four regressive-transgressive (R-T) cycles of the Sobrarbe and Escanilla Formations. This information will be used to evaluate: (1) axis-to-margin change in fluvial architecture at the system scale; and (2) relate stratigraphic architecture to accommodation and sediment supply (A/S). These results are used to develop two concepts: (1) the importance of subdividing fluvial stratigraphic data based on geological distinctions and (2) regional stacking patterns can be used as predictors to some meso- (reservoir) scale characteristics of fluvial systems. Both of these concepts have implications when building reservoir models.

3.3 Geologic Setting

The Ainsa Basin, Spain (Fig. 3.2A), a sub-basin of the larger Tremp-Ainsa-Jaca Basin, developed from a foreland basin to a thrust-top (piggy-back) basin south of the axial zone of the
South Pyrenean Central Thrust System (Mutti, 1977; Puigdefabregas, 1992; Munoz et al., 1994; Fernandez, 2004). The Ainsa basin extends ~ 40 km in the north-south direction and ~ 25 km in the east-west direction. It is bounded by four syndepositionally active structures (Poblet et al., 1998; Dreyer et al., 1999, and Fernandez et al., 2004; Hoffman, 2009): (1) the Boltaña Anticline to the west, and (2) the Mediano Anticline to the east; (3) the Ãnisclo Anticline to the north, and (4) the Cotiella Thrust to the northwest. The Ainsa Basin contains several syndepositionally active intrabasinal structures including the Arcusa and Olson Anticlines, located in the southern half of the basin (Figs. 3.3; Garrido-Megias, 1968; Garrido-Megias, 1973; Bentham and Burbank, 1996; Dreyer et al., 1999; Fernandez et al., 2012). The Olson Anticline subdivides the southern half of the Ainsa Basin into two synclines (Fig. 3.3; Fernandez et al., 2012): (1) the Buil Syncline is located in the eastern half of the basin between the Olson and Mediano Anticlines; and (2) the Arcusa Syncline is located in the western half of the basin between the Olson and Boltaña Anticlines.

The basin-fill succession is divided into the Hecho and Campodarbe Groups, both of which overlie mixed carbonate and siliciclastic pre-growth strata (Figs. 3.2B, 3.2C; Poblet et al., 1998; Fernandez et al., 2004). The Hecho Group is ~4 km thick and is subdivided into seven smaller units, termed turbidite systems (Mutti et al., 1989; Pickering and Corregidor, 2005; Pickering and Bayliss, 2009) and formations by Moody et al. (2012). From oldest to youngest they are: (1) Fosado, (2) Arro-Charo, (3) Gerbe, (4) Banaston, (5) Ainsa, (6) Morillo, and (7) Guaso (Figs. 3.2B, 3.2C). Each is a third-order stratigraphic unit, meaning they record ~1-2 million years of deposition each (sensu Mitchum and Van Wagoner, 1991).

The focus of this study is on the overlying Campodarbe Group which is ~2 km thick and is divided into the Sobrarbe and Escanilla Formations (Figs. 3.2B, 3.2C, 3.3). These formations
record the final filling of the Ainsa Basin and the progradation of a linked fluvial-deltaic system over the area (Bentham et al., 1992; Dryer et al., 1999; Pickering and Bayliss, 2009; Moss-Russell, 2009; Silalahi, 2009; Pyles et al., in review). The Sobrarbe and Escanilla Formations are both sourced from the Pyrenean massif through the Tremp-Graus Basin to the east (Fig. 3.2A; Garrido-Mégiاس 1968; Vincent, 2001; Michael et al., 2014).

The Sobrarbe Formation is the basal formation of the Campodarbe Group and represents the youngest marine strata in the Ainsa basin-fill succession. Based on biostratigraphic and magnetostratigraphic data, the Sobrarbe Formation was deposited over a period of approximately 3 million years in the Late Leutian (Fig. 3.2C; Dreyer et al., 1999; Mochales et al., 2012), is ~1000 m thick (Dreyer et al., 1999), and has calculated rates of sediment accumulation (undecompacted) being ~32 cm/kyr (Dreyer et al., 1999; Mochales et al., 2012).

The Sobrarbe Formation displays cyclic alternations between mudstone-dominated delta plain deposits, carbonates, delta front sandstones, collapse complexes, muddy delta slope deposits, and turbidite sandstone. Dreyer et al. (1999) divided the Sobrarbe Formation into four composite sequences. Each composite sequence is composed of multiple smaller-scale regressive-transgressive successions. Work by Moss-Russell (2009) divided the Sobrarbe Formation into six condensed section bound R-T cycles (Figs. 3.3, 3.4, 3.5). Each R-T cycle has an identifiable shelf-edge, a shelf-margin delta, and large-scale shelf-to-basin clinoforms. Each shelf edge from one R-T cycle to the next progrades basinward and aggrades vertically (Figs. 3.3, 3.4, 3.5). Each R-T cycle is approximately fourth-order in duration, meaning they record approximately 0.1 to 0.5 m.y. of deposition each (sensu Mitchum and van Wagoner, 1991). Moss-Russell’s (2009) divisions are used in for this study.

The Escanilla Formation interfingers and conformably overlies the deltaic and shallow
marine deposits of the Sobrarbe Formation, is up to ~1.1 km thick (Bentham et al., 1992), and unconformably underlies the Oligocene Collegats Formation, a conglomeratic alluvial fan deposit (Figs. 3.2C, 3.3; Garrido-Méñigas, 1973; Bentham et al., 1992). Dryer et al. (1993) divided the Escanilla Formation into a Mondot and Olson Member. The older Mondot Member was coevally deposited with the Sobrarbe Formation and is a transitional unit between the deltaic Sobrarbe Formation and the fully fluvial Olson Member (Figs. 3.2C, 3.3). The Mondot Member is composed of low-sinuosity fluvial channels and fine-grained floodplain deposits (Bentham et al., 1992). Paleocurrents collected from the Mondot Member are to the NW\NNW, consistent with the coevally deposited Sobrarbe Formation to the north (Fig. 3.3; Bentham et al., 1992, Dreyer et al., 1999; Moss-Russell, 2009). This study divides the Mondot Member of the Escanilla Formation into six R-T cycles that are equivalent to the six R-T cycles of the Sobrarbe Formation (Figs. 3.3, 3.4, 3.5). The focus of this study is on Cycles 2, 3, 4, and 5.

3.4 Data and Methods

The study area is located on the western limb of the Buil syncline (Fig. 3.3). Data used to address the questions of the study include: (1) a geologic map that documents the aerial distribution of the boundaries of formations, R-T cycles, and paleocurrent measurements (Fig. 3.3); (2) 18 detailed stratigraphic columns totaling > 4 km in thickness that document lithology, grain-size, physical sedimentary structures, and stratal boundaries at centimeter-scale resolution (Figs. 3.3, 3.4, 3.5); and (3) interpreted photopanels used to document the distribution of architectural elements and the location of stratal boundaries (Fig. 3.4).

These data were used to construct a stratigraphic cross section that documents the distribution of fluvial, coastal plain/deltaic and marine strata in the study area and surfaces that
were used to correlate time-stratigraphic units across the study area (Fig. 3.5). The cross section was in turn used to quantitatively document shelf-edge trajectories for Cycles 2-5. The northern part of the cross section is oriented parallel to the average sediment transport direction and is therefore a depositional-dip oriented profile, whereas the southern part of cross section is oriented nearly perpendicular to the average sediment transport direction and is therefore a depositional-strike oriented profile (Fig. 3.5). The key stratigraphic surfaces depicted on the cross section and geologic map are the Guaso-Sobrarbe contact and the boundaries between the six R-T cycles of the Sobrarbe and Escanilla Formations (Figs. 3.3, 3.4, 3.5). Multiple datums were used in the construction of the cross section and each R-T cycle contains its own datums based on the following criteria. The datum for the northern part of the cross section are surfaces that separate sand-rich deltaic deposits from marine mudstones (flooding surfaces) on the shelf, whereas the datum for the southern part of the cross section is a stratigraphic surface that records the interface between the progradational and retrogradational units of each R-T cycle. The lowermost surface on the cross section is a regionally continuous, black, organic-rich shale horizon that is interpreted as a condensed section and is the genetic boundary between the underlying Guaso Formation and the overlying Sobrarbe Formation. This bed was not used as a datum because it is interpreted to reflect the inherited shelf-to-basin profile created during deposition of the Guaso Formation and is therefore not a geometrically flat time surface (Pyles et al., in review). The uppermost surface of the cross section is the top of Cycle 6, which is the top of the Mondot Member of the Escanilla Formation. This boundary represents a shift in paleocurrent directions from being predominantly north to predominantly west and ultimately to the south due to regional tectonic uplift to the northeast (Bentham et al., 1992, Dreyer et al., 1999).
3.4.1 Fluvial Hierarchy of Architectural Elements

The identification and analysis of architectural elements is critical in the interpretation of deposits and their depositional history. The concept of architectural elements was first introduced by Miall (1985) to describe non-marine deposits. An architectural element is defined as “a mesoscale lithosome characterized by its external shape in depositional strike view that acts as a fundamental building block for larger stratigraphic units” (Pyles, 2007, p. 5). The architectural elements of the Escanilla Formation are grouped into a three-level hierarchy based in part on the methodology proposed by Ford and Pyles (2014). From smallest to largest, the three levels are: story, element, and archetype (Fig. 3.6). Each hierarchical level is composed of different combinations of components that account for the variability in sedimentation styles observed in the Escanilla Formation. Each hierarchical level is constrained by stratal surfaces, lithofacies, external shape in depositional strike view of units, and cross-cutting relationships documented in the study area. This study is focused on the story and element levels (Figs. 3.6, 3.7).

3.4.1.1 Story

A story is “a meso-scale volume of strata formed from genetically related beds or bedsets produced by the migration, fill or overbank discharge of a single fluvial system” (Ford and Pyles, 2014, pg. 1281). The thickness of each story scales to bank-full discharge and flood-stage water depth. Stories are the fundamental building blocks for larger stratigraphic units: elements and archetypes (Figs. 3.6, 3.7).

Eight different types of stories were identified in the study area and are divided into channel fill components and floodplain fill components (Fig. 3.6). Channel-fill components are: lateral accreting, downstream accreting, fine-grained fill associated with lateral accretion, and
erosionally based fine-grained fill (Fig. 3.6). Floodplain-fill components are: splay, crevasse channel with heterolithic fill, crevasse channel with erosionally based fine-grained fill, and floodplain fines (Fig. 3.6). Each story is distinctive in terms of cross-sectional shape in depositional strike view, lithofacies, modal grain size, and sediment transport directions in relation to stratal geometry. For brevity, descriptive characteristics are described in Table 3.1.

3.4.1.2 Element

An element is defined “as a macroscale lithosome produced from the migration and overbank discharge of a single fluvial channel” (Ford and Pyles, 2014, pg. 1294). An element is separated from stratigraphically adjacent elements by floodplain fines or an erosional surface when eroded into by a younger element. An element is composed of one or more stories (Figs. 3.6, 3.7). Multistory elements are defined as an element that contains more than one story that stack laterally and/or vertically within the element (Feofilova, 1954; Ford and Pyles, 2014). Two types of elements were recognized: (1) channel-belt elements, and (2) floodplain-belt elements (Figs. 3.6, 3.7).

3.4.1.2.1 Channel-Belt Element

A channel-belt element is composed of multiple channel fill stories (Figs. 3.6, 3.7A). Channel-belt elements contain a combination of lateral and downstream accreting stories and either fine-grained fill associated with lateral accretion story or erosionally based fine-grained fill story (Fig. 3.7A). In the study area, channel-belt elements have an average thickness of ~2.99 m, an average width of ~152 m and an average aspect ratio of ~51. The lower bounding surface is weakly erosional. The initial fill of channel-belt elements is composed of either laterally
accreting or downstream accreting stories which transition vertically and laterally into
downstream accreting stories and fine-grained fill associated with lateral accretion or erosionally
based fine-grained fill stories near the top of the element. The amount of erosion ranges from 0.5
m to 7 m. The upper bounding surface is conformable except where younger channel fill stories
erode into it (Fig. 3.7A). The lateral margins are sharp and erode into adjacent strata (Fig. 3.7A).
In depositional strike view, channel-belt elements are thickest in the axis and thin either abruptly
or gradually toward its lateral margins (Fig. 3.7A). Channel-belt elements are interpreted as
braided channel belts.

3.4.1.2.2 Floodplain-Belt Element

A floodplain-belt element is composed of a combination of multiple floodplain fill
components (Figs. 3.6, 3.7B): splay stories, crevasse channel with heterolithic fill stories,
crevasse channel erosionally based fine-grained fill stories, and floodplain fine stories. In the
study area, floodplain-belt elements have an average thickness of ~2.36 m, an average width of
~450 m and an average aspect ratio of ~191. The lower bounding surface of a floodplain-belt
element is conformable to erosional depending on the floodplain fine story that is at the base of
the element. The upper surface is conformable to undulatory expect when eroded into by
younger strata. Laterally, a floodplain-belt element is thickest adjacent to its genetically related
channel belt and thins towards its margins except when eroded into by a channel-belt element.
Two types of floodplain-belt elements are documented: (1) associated non-coeval floodplain-belt
elements and (2) unassociated floodplain-belt elements (Fig. 3.7B).

Associated non-coeval floodplain-belt elements account for 62% of the floodplain-belt
elements. From base to top, they are composed of: (1) amalgamated floodplain fine stories,
which can be overlain by thin beds of carbonaceous mudstone interpreted to be small evaporite ponds within the floodplain (Fig. 3.8B) (2) distal splay beds, composed of bioturbated and/or rotted structureless to rippled sandstone intercalated with siltstone (Fig. 3.7C); (3) progressively more proximal splay beds, composed of structureless, planar laminated, and rippled very fine- to fine-grained sandstone intercalated with siltstone (Fig. 3.7C) and (4) either a crevasse channel with heterolithic fill stories or crevasse channel with erosional based fine-grained fill stories.

Splay stories are composed of multiple splay beds (Fig. 3.7C). Splay beds are composed of structureless to planar cross-stratified, and rippled very fine- to fine-grained sandstone, which is overlain by bioturbated and/or rooted structureless to rippled silty sandstone to mudstone. Splay beds are often completely bioturbated. The average thickness of a splay bed is 0.36 m, an average width of 300 m, and aspect ratio of 833. The average splay story contains ~6 splay beds per story. Even though individual splay beds fine upwards, the overall succession of a splay story both coarsens and thickens upward from one bed to the next.

Associated non-coeval floodplain-belt elements are always in direct contact with an overlying channel-belt element (Fig. 3.7B). The axis of the channel-belt element erodes into the axis of the floodplain-belt element, interpreted to indicate a genetic linkage between the two. Associated non-coeval floodplain-belt elements are interpreted to represent the progradation of a crevasse splay complex into a floodplain and the full avulsion of a channel-belt element (Smith et al., 1989; Aslan and Blum, 1999; Stouthamer, 2001; Slingerland and Smith, 2004).

Unassociated floodplain-belt elements account for 38% of the floodplain-belt elements (Fig. 3.7B). From base to top, they are composed of: (1) amalgamated floodplain fine stories at its base, which can be overlain by thin beds of carbonaceous mudstone interpreted to be small evaporite ponds within the floodplain; (2) distal splay beds, composed of bioturbated and/or
rotted structureless to rippled sandstone intercalated with siltstone; (3) progressively more proximal splay beds, composed of structureless, planar laminated, and rippled very fine- to fine-grained sandstone intercalated with siltstone; and (4) individual splay beds within the splay story that thin from one to the next in an upward succession and capped with amalgamated floodplain fine stories. Unassociated floodplain-belt elements coarsen upwards from one bed to the next and then fine upward from one bed to the next. Unassociated floodplain-belt elements are interpreted to represent the progradation of a crevasse splay complex into a floodplain and a failed avulsion of a channel-belt element (Smith et al., 1989; Aslan and Blum, 1999; Stouthamer, 2001; Slingerland and Smith, 2004).

3.4.2 Shelf-edge Trajectory

To document the rate of accommodation in relation to sediment supply, examination of shelf-edge trajectories was conducted (sensu Helland-Hansen and Martinsen, 2009; Pyles et al., 2011). Trajectory of the shelf edge ($T_{se}$) is defined as: $T_{se} = \frac{dy_{se}}{dx_{se}}$, were $dy_{se}$ is the vertical aggradation from one shelf edge to the next and $dx_{se}$ is the longitudinal translation from one shelf edge to the next (Pyles et al., 2011; Fig. 3.8A). The shelf edge—the interface between the shelf and slope, within each R-T cycle is defined as the point of maximum progradation of the shoreline during regression and is overlain by marine mud of the transgressive component of the R-T cycle. The trajectory of the shelf edge can be used as a proxy for the ratio between accommodation and sediment supply (A/S) where $dy_{se}$ is an indicator of accommodation and $dx_{se}$ is an indicator of how sediment supply relates to accommodation. If both $dy_{se}$ and $dx_{se}$ are positive, such as in the Sobrarbe Formation (Fig. 3.8B; Pyles et al., in review), a range of shoreline trajectories is possible (Fig. 3.8C). For example, if $dy_{se}$ is held constant from one shelf
edge to the next, and $dx_{se}$ increases from one shelf edge to the next, $T_{se}$ decreases (Fig. 3.8C). In this scenario sediment supply was higher than accommodation and is proportional to $T_{se}$. The shelf-edge trajectories mapped within the deltaic deposits of the Sobrarbe Formation were physically correlated to time-equivalent fluvial deposits of the Escanilla Formation (Fig. 3.5), which facilitates a quantitative study of how variations in fluvial architecture relate to variations in shelf-edge trajectory and therefore A/S. The shelf-edge trajectory of Cycle 2 is $(dy_{se}/dx_{se}; 52 \text{ m} / 2848 \text{ m})$ 0.02 or 1.05°, rising basinward (northward) (Fig. 3.5). The shelf-edge trajectory of Cycle 3 is $(dy_{se}/dx_{se}; 66 \text{ m} / 1372 \text{ m})$ 0.05 or 2.77°, rising basinward (northward) (Fig. 3.5). The shelf-edge trajectory of Cycle 4 is $(dy_{se}/dx_{se}; 41 \text{ m} / 1503 \text{ m})$ 0.03 or 1.56°, rising basinward (northward) (Fig. 3.5). The shelf-edge trajectory of Cycle 5 is $(dy_{se}/dx_{se}; 45 \text{ m} / 3821 \text{ m})$ 0.01 or 0.67°, rising basinward (northward) (Fig. 3.5). The cycles with the highest $T_{se}$ and therefore higher A/S, are Cycles 3 and 4, whereas, the cycles with the lowest $T_{se}$ and therefore lower A/S, are Cycles 2 and 5.

### 3.5 Geology of Fluvial Strata within the R-T Cycles

This study quantitatively documents the stratigraphic architecture of four R-T cycles of the Escanilla Formation, Cycles 2, 3, 4, and 5 in order to evaluate the importance of dividing quantitative data into different references based on geologic distinctions. Data used to facilitate a quantitative analysis is derived from the geologic map, photopanels, and from multiple stratigraphic columns (columns 13, 14, 15, 16, 17, and 18) (Figs. 3.3, 3.4, 3.5). This section presents undifferentiated (average over the entire cycle) data for each individual R-T cycle. The following two sections document fluvial architecture based on (1) axis-to-margin location in a fluvial system and (2) shelf-edge trajectory (A/S).
3.5.1 Cycle 2

The map, photopanel, and stratigraphic columns are used to document the following undifferentiated (average over the entire cycle) characteristics of fluvial strata within Cycle 2 (Figs. 3.3, 3.9, 3.10A). First, paleocurrent are to the north-northwest (vector mean of 347°) (Fig. 3.9A). Second, Cycle 2 contains: channel belt stories (8%), splay stories (8%), and floodplain fine stories (84%) (Fig. 3.10B). The average channel-belt element thickness is 2.10 m, average floodplain-belt element thickness is 1.49 m, average splay bed thickness is 0.37 m, and the average number of splay beds within Cycle 2 is 21 (Fig. 3.10A). Third, the ratio of the area of sandstone to total area being evaluated (net-sand content) is 0.30 with 55% of the sand located within channel-belt stories and 45% located within splay and crevasse channel stories (Fig. 3.10A).

3.5.2 Cycle 3

The map, photopanel, and stratigraphic columns are used to document the following undifferentiated (average over the entire cycle) characteristics of fluvial strata within Cycle 3 (Figs. 3.3, 3.9, 3.10A). First, paleocurrent are to the north-northwest (vector mean = 346°) (Fig. 3.9B). Second, Cycle 3 contains (Fig. 3.10B): channel belt stories (15%), splay stories (18%), and floodplain fine stories (67%). The average channel-belt element thickness is 2.92 m, average floodplain-belt element thickness is 4.17 m, average splay bed thickness is 0.34 m, and the average number of splay beds within Cycle 3 is 43 (Fig. 3.10A). Third, the net-sand content is 0.39 with 39% of the sand located within channel-belt stories and 61% located within splay and crevasse channel stories (Fig. 3.10A).
3.5.3 Cycle 4

The map, photopanel, and stratigraphic columns are used to document the following undifferentiated (average over the entire cycle) characteristics of fluvial strata within Cycle 4 (Figs. 3.3, 3.9, 3.10A). First, paleocurrent are to the north-northwest (vector mean = 342°) (Fig. 3.9C). Second, Cycle 4 contains (Fig. 3.10B): channel belt stories (16%), splay stories (8%), and floodplain fine stories (76%). The average channel-belt element thickness is 3.05 m, average floodplain-belt element thickness is 1.79 m, average splay bed thickness is 0.27 m, and the average number of splay beds within Cycle 4 is 34 (Fig. 3.10A). Third, the net-sand content is 0.44 with 56% of the sand located within channel-belt stories and 44% located within splay and crevasse channel stories (Fig. 3.10A).

3.5.4 Cycle 5

The map, photopanel, and stratigraphic columns are used to document the following undifferentiated (average over the entire cycle) characteristics of fluvial strata within Cycle 5 (Figs. 3.3, 3.9, 3.10A). First, paleocurrent are to the north (vector mean = 002°) (Fig. 3.9D). Second, Cycle 5 contains (Fig. 3.10B): channel belt stories (30%), splay stories (6%), and floodplain fine stories (64%). The average channel-belt element thickness is 4.67 m, average floodplain-belt element thickness is 1.97 m, average splay bed thickness is 0.38 m, and the average number of splay beds within Cycle 5 is 17 (Fig. 3.10A). Third, the net-sand content is 0.28 with 74% of the sand located within channel-belt stories and 26% located within splay and crevasse channel stories (Fig. 3.10A).
3.6 Axis to Margin Variations of Stratigraphic Architecture of the System

Below we discuss lateral variations in outcrop characteristics along the strike-oriented part of the cross section which is oriented NE-SW (Figs. 3.3, 3.5, 3.10). Stratigraphic characteristics evaluated in this analysis are: (1) proportions of elements (2) thickness of channel-belt and floodplain-belt elements, (3) net-sand content, and (4) thickness of R-T cycles. Variations summarized below are interpreted to reflect axis-to-margin variations in the stratigraphy of fluvial strata within each R-T cycle.

3.6.1 Cycle 2

From the northeastern outcrops of the study area to the southwestern outcrops, variations in proportions of stories, element thickness, net-sand content and R-T cycle thickness are documented within Cycle 2. The northeastern outcrop near the town of Mondot contains (Figs. 3.3, 3.9A, 3.10B): channel belt stories (16%), splay stories (11%), and floodplain fine stories (73%), whereas to the southwest between the towns of Eripol and Almazorre, the outcrop contains (Figs. 3.3, 3.9A, 3.10B): channel belt stories (4%), splay stories (5%), and floodplain fine stories (91%) (Fig. 3.10B). The average channel-belt element thickness is 4.10 m (northeast) versus 2.00 m (southwest) (p-value at 95% CI= 0.0048; t=4.3583), average floodplain-belt element thickness is 1.23 m (northeast) versus 1.71 m (southwest) (p-value at 95% CI= 0.9165; t=0.1102), average splay bed thickness is 0.40 m (northeast) versus 0.30 m (southwest) (p-value at 95% CI= 0.1227; t=1.5779), and the average number of splay beds within Cycle 2 is 25 (northeast) versus 18 (southwest) (Fig. 3.10A). The net-sand content in the northeastern outcrop is 0.44 with 61% of the sand located within channel-belt stories and 39% located within splay and crevasse channel stories, whereas in the southwestern outcrop, the net-sand content is 0.15
with 48% of the sand located within channel-belt stories and 52% located within splay and crevasse channel stories (Fig. 3.10C). The thickness of Cycle 2 is 110 m in the northeastern outcrops and thins to 96 m to the southwest (Fig. 3.9A).

3.6.2 Cycle 3

From the northeastern outcrops of the study area to the southwestern outcrops, variations in proportions of stories, element thickness, net-sand content and R-T cycle thickness are documented within Cycle 3. The northeastern outcrop near the town of Mondot contains (Figs. 3.3, 3.9B, 3.10B): channel belt stories (24%), splay stories (24%), and floodplain fine stories (52%), whereas to the southwest between the towns of Eripol and Almazorre, the outcrop contains (Fig. 3.3): channel belt stories (8%), splay stories (15%), and floodplain fine stories (77%) (Fig. 3.10B). The average channel-belt element thickness is 3.93 m (northeast) versus 1.90 m (southwest) (p-value at 95% CI= 0.0863; t=3.1806), average floodplain-belt element thickness is 5.60 m (northeast) versus 2.73 m (southwest) (p-value at 95%CI= 0.0858; t=2.1351) , average splay bed thickness is 0.36 m (northeast) versus 0.31 m (southwest) (p-value at 95%CI= 0.0965; t=1.7591), and the average number of splay beds within Cycle 3 is 55 (northeast) versus 25 (southwest) (Fig. 3.10C). The net-sand content in the northeastern outcrop is 0.47 with 50% of the sand located within channel-belt stories, whereas in the southwestern outcrop, the net-sand content is 0.39 with 27% of the sand located within channel-belt stories and 73% located within splay and crevasse channel stories (Fig. 3.10A). The thickness of Cycle 3 is 68 m in the northeastern outcrops and thins to 58 m to the southwest (Fig. 3.9B).
3.6.3 Cycle 4

From the northeastern outcrops of the study area to the southwestern outcrops, variations in proportions of stories, element thickness, net-sand content and R-T cycle thickness are documented within Cycle 4. The northeastern outcrop between the towns of Mondot and Olson contains (Figs. 3.3, 3.9C, 3.10B): channel belt stories (24%), splay stories (13%), and floodplain fine stories (63%), whereas to the southwest between the towns of Eripol and Almazorre, the outcrop contains (Fig. 3.3): channel belt stories (3%), splay stories (9%), and floodplain fine stories (88%) (Fig. 3.10B). The average channel-belt element thickness is 3.44 m (northeast) versus 2.65 m (southwest) (p-value at 95%CI= 0.2015; t=1.4341), average floodplain-belt element thickness is 2.40 m (northeast) versus 1.20 m (southwest) (p-value at 95%CI= 0.1210; t=2.1476), average splay bed thickness is 0.31 m (northeast) versus 0.16 m (southwest) (p-value at 95%CI= 0.6372; t=0.4780), and the average number of splay beds within Cycle 4 is 45 (northeast) versus 23 (southwest) (Fig. 3.10C). The net-sand content in the northeastern outcrop is 0.49 with 64% of the sand located within channel-belt stories and 36% located within splay and crevasse channel stories, whereas in the southwestern outcrop, the net-sand content is 0.31 with 48% of the sand located within channel-belt stories and 52% located within splay and crevasse channel stories (Fig. 3.10A). The thickness of Cycle 4 is 125 m in the northeastern outcrops and thins to 64 m to the southwest (Fig. 3.9C).

3.6.4 Cycle 5

From the northeastern outcrops of the study area to the southwestern outcrops, variations in proportions of stories, element thickness, net-sand content and R-T cycle thickness are documented within Cycle 5. The northeastern outcrop near the town of Olson contains (Figs. 3.3,
3.9D, 3.10B): channel belt stories (39%), splay stories (5%), and floodplain fine stories (56%), whereas to the southwest between the towns of Eripol and Almazorre, the outcrop contains (Fig. 3.3): channel belt stories (13%), splay stories (7%), and floodplain fine stories (80%) (Fig. 3.10B). The average channel-belt element thickness is 6.11 m (northeast) versus 3.23 m (southwest) (p-value at 95%CI= 0.0865; t=2.1288), average floodplain-belt element thickness is 2.60 m (north) versus 1.33 m (south) (p-value at 95%CI= 0.0035; t=8.4719), average splay bed thickness is 0.42 m (northeast) versus 0.33 m (southwest) (p-value at 95%CI= 0.0710; t=1.9806), and the average number of splay beds within Cycle 5 is 18 (northeast) versus 16 (southwest) (Fig. 3.10C). The net-sand content in the northeastern outcrop is 0.45 with 85% of the sand located within channel-belt stories and 15% located within splay and crevasse channel stories, whereas in the southwestern outcrop, the net-sand content is 0.11 with 62% of the sand located within channel-belt stories and 38% located within splay and crevasse channel stories (Fig. 3.10C). The thickness of Cycle 5 is 105 m in the northeastern outcrops and thins to 98 m to the southwest (Fig. 3.9D).

### 3.6.5 Summary and Interpretations

In summary, paleocurrents averaged over all the R-T cycles are to the north-northwest (Figs. 3.3, 3.9). The outcrop orientation is generally perpendicular but slightly oblique to the average paleocurrent direction. The average strike distance from the northeast to the southwest parts of the field area is ~ 4 km (Fig. 3.3) for each cycle. Within each individual R-T cycle, the majority of strata are floodplain-belt elements, followed by channel-belt elements irrespective of physiographic location of deposition (e.g. northeast or southwest) (Fig. 3.10). However, there are differences in several characteristics from the northeast and southwest parts of the field area (Fig.
3.10). The northeastern part of the field area contains larger, and a higher proportion, of channel-belt elements, a higher proportion of splay and crevasse channel stories, and a higher net-sand content than their counterparts deposited in the southwestern part of the field area (Fig. 3.10). Also, the thickest part of each R-T cycle is located in the northeastern part of the field area (Fig. 3.9).

Based on these observations the northeastern part of the field area is interpreted to be the axis of the system for each R-T cycle and the southwestern part of the field area is interpreted to be the margin of the system for each R-T cycle (Figs. 3.9, 3.10). Focusing of the fluvial system is hypothesized to be due to structural focusing of the system. The bounding structures of the Ainsa basin (Boltaña and Mediano Anticlines) were actively growing during deposition of Hecho and Campodarbe Groups (Mutti et al., 1989; Puigdefabregas et al., 1992; Dreyer et al., 1999; Anastasio and Holl, 2001; Hoffman, 2009; Mochales et al., 2012; Pyles et al., in review).

Additionally, the field area for this study is located within the Arcusa Syncline (Fig. 3.3). The northeast part of the field area is located within the axis of the Arcusa Syncline, while the southwest part of the field area is located on the western flank of the Arcusa Syncline. The systematic eastward shift in the axis of each R-T cycle (axis of the Arcusa Syncline) during deposition of Cycles 2-6 is attributed to the growth of the Arcusa Anticline (Fig. 3.9; Dreyer et al., 1999; Chapter 2).

### 3.7 Stratigraphic Architecture in Relation to A/S

This study quantitatively documents how the stratigraphic architecture in four R-T cycles relates to A/S ratio (shelf-edge trajectory) (Fig. 3.5, 3.8, 3.11). For comparison, we categorize shelf-edge trajectories $\geq 0.03$ as high shelf-edge trajectories (high A/S) (e.g. Cycles 3 (T$_{se} =$
0.05) and 4 (T_{se} = 0.03)) and shelf-edge trajectories ≤ 0.02 as low shelf-edge trajectories (low A/S) (e.g. Cycles 2 (0.02) and 5 (0.01)). Stratigraphic characteristics evaluated in this analysis are: (1) proportions of elements (2) thickness of channel-belt and floodplain-belt elements, (3) net-sand content, and (4) relationship between genetically related floodplain-belt and channel-belt elements.

The proportions of elements within R-T cycles associated with low A/S ratios (Cycles 2 and 5) are different than those associated with high A/S ratios (Cycles 3 and 4). First, channel-belt elements associated with low A/S ratios are higher in proportion than those associated with high A/S ratios: 19% vs. 15%, respectively. Second, channel-belt elements associated with low A/S ratios have an average thickness of ~ 3.58 m but ~ 2.56 m for high A/S ratios (Fig. 3.11A) (p-value at 95%CI= 0.1466; t=1.4965). Floodplain-belt elements associated with low A/S ratios have an average thickness of ~ 2.54 m but ~ 2.83 m for high A/S ratios (Fig. 3.11A) (p-value at 95%CI= 0.9724; t=0.0350). Third, the net-sand content for strata associated with low A/S ratios is 0.29 but 0.40 for high A/S ratios (Fig. 3.11A). When the A/S ratio is low, 64% of the sandstone is located within channel fill stories and 36% within splay and crevasse channel stories, whereas when the A/S ratio is high, 46% of the sandstone is located within channel fill stories and 54% within splay and crevasse channel stories (Fig. 3.11A). Fourth, R-T cycles associated with a low A/S ratio contain thicker channel-belt elements in relation to their genetically related floodplain-belt elements than their high A/S counterparts (Fig. 3.11B).

The variations of fluvial architecture from one cycle to the next, as documented above may also be due to varying distances of the measured strata from their contemporaneous shelf-edge. Fluvial data for subsequent cycles was collected at increasing distances from their coevally deposited shelf edge: Cycle 2 = 1-2 km; Cycle 3 = 2-3 km; Cycle 4 = 3-4 km; and Cycle 5 = 6-7 km.
km (Fig. 3.10B). Previous work of different fluvial systems has documented a systematic increase in channel amalgamation and size the further landward the channel is located (e.g. Legarreta and Uliana, 1998; Holbrook et al., 2006; Nichols and Fisher, 2007; Karssenberg and Bridge, 2008; Hartley et al., 2010; Weismann et al., 2010). This study also documents this trend of increasing channel size the further the strata is from its coeval shelf edge when using only the undifferentiated (average over the entire cycle) data (Fig. 3.10A). However, data differentiated by axis versus margin and A/S ratio document a different pattern. For example, when only using data from the axis of the system, Cycle 2 contains channel-belt elements that are thicker than those deposited within Cycles 3 and 4, even though Cycle 2 is located closer to its contemporaneous shelf-edge than Cycles 3 and 4 (Figs. 3.10, 3.11). Other trends such as the ratio of channel-belt element thickness to floodplain-belt element thickness (Fig. 3.11B), the number and size of channel-belt elements, net-sand content, and the number and size of splay elements correlate to variations in A/S ratio and not the distance the fluvial strata is from their contemporaneous shelf-edge (Figs. 2.10, 2.11).

3.8 Discussion

This study relates large-scale variations in fluvial architecture to position in an axis to margin transect and A/S ratio. These intrasystem patterns are important to consider when building reservoir models. For example, the average thickness for floodplain-belt elements in Cycles 2-5 is 2.36 m with an average of six splay beds per element (Fig. 3.12A). However if the data is divided on the basis of axis to margin position a different pattern is evident. The average thickness for floodplain-belt elements in the axis of the system is 2.95 m with an average of seven splay beds per element and 1.74 m with an average of five splay beds per element in the
margin (Fig. 3.12B). If one further divides data from the axis position by A/S ratio, the average thickness for floodplain-belt element strata associated with high A/S ratios is 3.48 m with an average of eight splay beds per element and 2.57 m with an average of five splay beds per element for low A/S ratios (Fig. 3.12C). The values for floodplain-belt element thicknesses are 47% thicker in the axis of the system associated with high A/S ratios than the average value of the entire system (Cycles 2-5; Fig. 3.12). There is also an average of two more splay beds per floodplain-belt element in the axis of the system associated with high A/S ratios than the average value of the entire system (Cycles 2-5; Fig. 3.12). An increase in the number of splays can make an impact on static connectivity between channel and splay deposits by increasing the static connectivity (see Chapter 4). Therefore, the stratigraphic data collected from the axis of a system are different than those of the margin and the average is therefore not a good representation of local aspect of the system.

Channel-belt element thicknesses follow similar trends. The average thickness for channel-belt elements in Cycles 2-5 is 2.99 m (Fig. 3.12A). However, if the data is divided on the basis axis to margin position a different pattern is evident. The average thickness for channel-belt elements in the axis of the system is 3.92 m and 2.93 m in the margin (Fig. 3.12B). If one further divides data from the axis position by A/S ratios, the average thickness for channel-belt element strata associated with high A/S ratios is 3.66 m but 4.07 m for low A/S ratios (Fig. 3.12C). The axial data within a high accommodation sequence is different than the average data for the entire system (Cycles 2-5; Fig. 3.12). The values for channel-belt element thicknesses are 36% thicker in the axis of the system associated with high A/S ratios than the average value of the entire system (Cycles 2-5; Fig. 3.12). Therefore, large-scale stacking patterns such as shelf-edge trajectories are related to small- (reservoir) scale characteristics of a fluvial system.
Therefore, when building reservoir models and developing exploration and production strategies it is important to think about what kind of data is going into the model and to critically consider subdividing populations of data on the basis of geological distinctions such as axis versus margin of the system as well as A/S ratios. This is important because there are differences in the volume of sand and potential for static connectivity between the axis and margin of the system as well as strata deposited under different A/S ratios. If one were to use only the average population data, they may under estimate the volume of hydrocarbon deposits in the axis, and over estimate in the margin. This study also relates meso-scale characteristics of the fluvial system that are not resolvable in seismic data to larger-scale stacking patterns that are observable in seismic data. This allows for the ability to predict meso-scale attributes such as proportions of elements, thickness of channel-belt and floodplain-belt elements, net-sand content, and the relationship between genetically related floodplain-belt and channel-belt elements.

3.9 Conclusion

The Eocene Sobrarbe and Escanilla Formations in the Ainsa Basin, Spain provide an opportunity to axis to margin and A/S variations in stratigraphic architecture of a fluvial system. It is therefore important to subdivide data sets on the basis of these and other geologic distinctions.

Fluvial strata deposited within the axis of the system contain a higher percentage of channel-belt elements and splay stories, thicker splay beds, thicker channel-belt elements, and a higher net-sand content than their counterparts deposited in the margin (Fig. 3.13A).

Fluvial strata associated with lower A/S ratios (shelf-edge trajectories) contain a higher percentage of channel-belt elements, larger channel-belt elements, and a lower net-sand content.
Channel-belt elements are thicker in relation to their genetically related floodplain-belt elements (Fig. 3.13B).

Fluvial strata associated with higher A/S ratios (shelf-edge trajectories) contain thicker and a higher percentage of floodplain-belt elements, smaller channel-belt elements, and a higher net-sand content. Channel-belt elements are thinner in relation to their genetically related floodplain-belt elements (Fig. 3.13B). Hence, subdividing populations on the basis of geological distinctions is important when developing reservoir models.

3.10 References


Karssenberg, D. and Bridge, J.S., 2008, A three-dimensional numerical model of sediment transport, erosion and deposition within a network of channel belts, floodplain and hillslope: extrinsic and intrinsic controls on floodplain dynamics and alluvial architecture: Sedimentology, 55, p. 1717-1745.


Figure 3.1 (A) Schematic diagram documenting the different elements found within fluvial systems and the variation in grain sizes of different floodplain deposits (modified from Allen, 1965). (B) Schematic diagram of a distributive fluvial system (sensu Weissmann et al., 2010) deposited in a structural basin illustrating lateral variations in fluvial stratigraphic architecture from the axis and the margin of the system. The axis of the system is depicted to have larger channels than the margin. (C) Schematic diagrams relating accommodation to stratigraphic architecture. Low accommodation systems are associated with mud-rich floodplains, whereas high accommodation setting are associated with more heterolithic, sand-rich floodplains. Modified from Aslan and Blum, 1999.
Figure 3.2 (A) Paleogeographic map of the Tremp-Ainsa-Jaca basin (modified from Fernandez et al., 2004). (B) Generalized stratigraphy of the Ainsa-Jaca Basin (modified from Pickering and Bayliss, 2009). (C) Chronostratigraphic chart of the Hecho and Campodarbe Groups showing the different lithostratigraphic units located in the Ainsa Basin. This study is focused on the Sobrarbe and Escanilla Formations.
Figure 3.3 Geologic map of the Ainsa Basin documenting the contacts between the different lithostratigraphic units, bounding structures (Boltaña Anticline and Mediano Anticline), smaller anticlines (Olson and Arcusa anticlines), and the Buil Syncline (map modified from Hoffman, 2009). This study is focused on the Sobrarbe and Escanilla Formations, which are formally divided into six regressive/transgressive cycles (Cycles 1-6) and are labeled on the map.
Figure 3.4 Interpreted photopanels of the Sobrarbe and Escanilla Formations documenting the boundaries between the six interpreted fourth-order regressive/transgressive stratigraphic cycles as defined by Moss-Russell (2009) and Pyles et al. (in review) as well as the location of the stratigraphic columns. (A) Photopanel taken from the town of Castellazo looking east. Photopanels taken from the town of Eripol looking north (B), east (C), and south (D).
Figure 3.5 Stratigraphic cross section of Sobrarbe and Escanilla Formations constrained by
geologic mapping and stratigraphic columns documenting: the contact between the Guaso and
Sobrarbe Formations, the boundaries between the six R-T cycles defined by Moss-Russell
(2009), the location of successive shelf-edge deltas, and the shelf-edge trajectory for Cycles 2-5.
The southern part of the cross section is oriented along depositional strike and the northern part
is oriented parallel to depositional dip. The location of cross section is shown in Fig. 3.3.

Figure 3.6 Schematic diagram of methodology developed by Ford and Pyles (in press) for fluvial
hierarchy of architectural elements. Time span of deposition, crosscutting relationships, and
superposition increase in an upward transect through the hierarchical levels. Components are not
drawn to scale.
Figure 3.7 Photographic examples, representative stratigraphic columns, and histograms of fluvial elements identified in this study. Elements are subdivided into channel-belt (A) and floodplain-belt (B) elements. Floodplain-belt elements are composed of splay beds and floodplain fines (C).
Figure 3.8 (A) Trajectory of the shelf edge is defined as $T_{se} = \frac{dy_{se}}{dx_{se}}$. (B) Matrix showing a range of trajectory styles. The trajectories measured in this study fall in the upper left window which has a positive $dy_{se}$ and $dx_{se}$. (C) The trajectory of the shelf edge can be used as a proxy for the ratio between accommodation and sediment input where $dy_{se}$ is an indicator of accommodation and $dx_{se}$ is an indicator of how sediment input scales to accommodation. Low trajectories correspond to low A/S whereas high trajectories are associated with high A/S.
Figure 3.9 Geologic maps depicting some paleogeographic aspects of Cycle 2 (A), Cycle 3 (B), Cycle 4 (C), and Cycle 5 (D). The maps are constrained by the geologic map, stratigraphic columns, and paleocurrents. The colored parts of the maps correspond to outcrop of the cycles being studied.
Figure 3.10 Quantitative data documenting axis-to-margin changes in the stratigraphy of the fluvial system for Cycle 2-5. (A) Data tables for undifferentiated data by cycle, axis of the system, and margin of the system. (B) Pie charts documenting proportions of channel, splay, and floodplain fine stories. Pie chart data derived from stratigraphic columns 14, 15, 17, and 18. The center of each pie chart is positioned at the distance from the shelf-edge the data was collected. (C) Stratigraphic columns from the axis (Section 14) and margin (Section 18) of fluvial system (see Fig. 3.3 for location of stratigraphic columns).
Figure 3.11 (A) Quantitative data documenting stratigraphic architecture in relation to variations in shelf-edge trajectories. (B) Cross plot of channel-belt element thicknesses versus floodplain-belt thicknesses from channels located in the axis of the system. Channels associated with low A/S (low shelf-edge trajectories) plot in a different domain than those associated with high A/S (high shelf-edge trajectories).
Figure 3.12 (A) Undifferentiated statistical data for all channel-belt and floodplain-belt element thicknesses within Cycles 2-5. (B) Statistical data for all channel-belt and floodplain-belt element thicknesses differentiated by axis and margin. (C) Statistical data for all axial data differentiated on the basis of high and low A/S ratios.
Figure 3.13 Summary diagram of variations in fluvial stratigraphic architecture in relation to (A) axis to margin positions in a fluvial system, and (B) A/S (shelf-edge trajectories).
Table 3.1 Descriptions of the different stories identified in this study.

<table>
<thead>
<tr>
<th>Story</th>
<th>Bounding Surfaces</th>
<th>Depositional Strike View</th>
<th>Lithofacies Composition (in order of decreasing abundance)</th>
<th>Interpretation</th>
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<td>medium-grained sandstone with pebbles; (3) pebble to cobble</td>
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<td>conglomerate with medium- to coarse-grained sandstone lenses</td>
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<td>Lateral accreting</td>
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<td>conformable except when eroded into by younger strata</td>
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<td>toward its lateral margins</td>
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<td>downstream accreting mid-</td>
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<td>toward its lateral margins</td>
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<td>grained sandstone with pebbles; (2) trough cross-bedded medium-</td>
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<td>to coarse-grained sandstone with pebbles; (3) trough cross-bedded</td>
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<td>conglomerate with medium- to coarse-grained sandstone lenses</td>
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<td>and is erosional on opposite margin</td>
<td>channel erosion</td>
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<td>siltstone; (2) structureless, planar laminated, and rippled</td>
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<td>wedge shaped, thickest in axis, and thins gradually</td>
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<td>fine-grained sandstone intercalated with siltstone</td>
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<td>fine-grained sandstone intercalated with siltstone</td>
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<td>(1) bioturbated and/or rooted structureless to rippled very</td>
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<td>fine-grained sandstone intercalated with siltstone</td>
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<td>Floodplain fines</td>
<td>conformable</td>
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<td>rectangular to wedge shaped, thickest in the axis and thins</td>
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<td>(1) varicolored siltstone to mudstone; (2) varicolored mudstone</td>
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<td>to fine-grained sandstone with slickensides; (3) carbonaceous</td>
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CHAPTER 4
QUANTITATIVE OUTCROP CHARACTERIZATION OF AXIS TO MARGIN
CHANGES IN STRATIGRAPHIC ARCHITECTURE OF TRANSGRESSIVE FLUVIAL
DEPOSITS: ESCANILLA FORMATION, SPAIN

A paper that has been submitted to American Association of Petroleum Geologists Bulletin.

Jeremiah D. Moody and David R. Pyles

4.1 Abstract

Transgressive fluvial strata are deposited during an overall landward migration of the shoreline. Few studies have focused on transgressive fluvial strata, especially those deposited in high-accommodation settings. The Escanilla Formation of the Ainsa Basin contains world-class outcrops of fluvial strata deposited during transgression in a high-accommodation setting. This study uses outcrop data to document, for the first time, vertical and lateral variations in stratigraphic architecture, net-sand content, grainsize, and static connectivity in this setting. Key axis-to-margin patterns in the fluvial system are an increase in the proportion of channel-fill and splay stories, and channel-belt elements at the expense of floodplain fine stories, and an increase in net-sand content, channel-belt element size, modal grain size, and static connectivity from the margin to the axis of the system. The axis of the system contains the best reservoir quality strata and potential for static connectivity. Significant vertical changes are an upward increase in channel-belt element size, net-sand content, modal grainsize within channel-belt elements, and
static connectivity. Therefore, the upper half of the system contains the best reservoir quality strata and potential for static connectivity. Data provided herein provide insight into high accommodation, transgressive fluvial deposits and can be used to reduce uncertainty in the interpretation of subsurface data, provide input to constrain rules-based forward stratigraphic models, and provide input to constrain reservoir models in transgressive fluvial systems.

4.2 Introduction

Transgressive fluvial deposits are deposited during an overall landward migration of the shoreline (Fig. 4.1; Vail et al., 1977; Posamentier and Vail, 1988; Shanley and McCabe, 1991, 1994; Wright and Marriott, 1993). If sediment input is held constant, temporal variations in subsidence and eustatic sea level create a continuum of sequences in which transgressive fluvial strata can be deposited: low-accommodation sequences and high-accommodation sequences (Fig. 4.1B; Vail et al., 1977; Posamentier and Vail, 1988; Mitchum and Van Wagoner, 1991).

Low-accommodation sequences form when the rate and magnitude of subsidence are less than the rate and magnitude of eustatic sea level change (Fig. 4.1B; a Type 1 sequence sensu Posamentier and Vail, 1988). They are composed of lowstand, transgressive, and highstand systems tracts (Fig. 4.1; Vail et al., 1977) and are characterized by incised valleys on the shelf that form from incision and sediment bypass during base-level fall (i.e. a drop in relative sea level). Transgressive fluvial deposits are initially confined within the incised valley during the subsequent rise in base level (i.e. a rise in relative sea level), become widespread once the incised valley is filled, and finally are either partially or completely eroded away by either the transgressive surface of erosion and/or the regressive surface of marine erosion during the subsequent drop in base level (Fig. 4.1B; e.g. Shanley and McCabe, 1991, 1994; Wright and
Marriott, 1993; Gardner et al., 2004; Hampson et al., 1997; Legarreta and Uliana, 1998; Varney, 2000; Bowen and Weimer, 2003; Holbrook et al., 2006; Kirschbaum and Shenk, 2011).

High-accommodation sequences form when the rate and magnitude of subsidence is greater than the rate and magnitude of eustatic sea level change (Fig. 4.1B; a Type 2 sequence sensu Posamentier and Vail, 1988). As a result, they are composed of only transgressive and highstand systems tracts and no incised valley is formed on the shelf (Fig. 4.1B). The transgressive systems tract consists of well-preserved vertically and laterally isolated fluvial channel-belt elements encased in floodplain deposits (Fig. 4.1B; e.g. Wright and Marriott, 1993; Burns et al., 1997; Rygel and Gibling, 2006; Fanti and Cantuneanu, 2010). This article is focused on fluvial deposits in transgressive systems tracts deposited in high-accommodation settings.

In the last few decades, several studies applied sequence stratigraphic principles and concepts to non-marine stratigraphy (Posamentier and Vail, 1988; Shanley and McCabe, 1991, 1994; Miall, 1991; Schumm, 1993; Wright and Marriott, 1993; Lagarreta and Uliana, 1998; Holbrook et al., 2006). Current non-marine sequence stratigraphic models however have primarily focused on low-accommodation sequences and the relationship between fluvial architecture and changes in relative sea level (e.g. Shanley and McCabe, 1991, 1994; Alexander, 1992; Wright and Marriott, 1993; Schumm, 1993; Aitkin and Flint, 1995; Olsen et al., 1995; Martinsen et al., 1999; Plint et al., 2001; Posamentier, 2001; Arnot et al., 2002; Gardner et al., 2004; Holbrook et al., 2006). Therefore, there is an opportunity to expand our knowledge of fluvial systems by working in high-accommodation settings.

It has been documented that avulsion of channel belts can be related to changes in base level (Coleman, 1969; Bridge and Leeder, 1979; Mackey and Bridge, 1995; Berendsen and Stouthamer, 2000; Schumm et al., 2000; Stouthamer and Berendsen, 2000; Stouthamer and
Berendsen, 2007). The frequency of avulsion increases with increased subsidence and decreases with increased uplift. As such, a high rate of base-level rise (i.e. subsidence) leads to a high avulsion frequency, probably due to a rapidly decreasing longitudinal gradient (Tornqvist, 1994; Bridge, 2003; Stouthamer and Berendsen, 2007). From base to top, an avulsion belt complex consists of floodplain fines, distal crevasse splay deposits, proximal crevasse splay deposits, crevasse channel deposits, and channel-belt element deposits (e.g. Smith et al., 1989; Tornqvist, 1994; Jones and Schumm, 1999; Stouthamer and Berendsen, 2000; Tornqvist and Bridge, 2002).

If base level rises, increased accommodation provides space for floodplain deposits and increased probability to preserve avulsion-belt complexes. Due to the discontinuous nature of fluvial channel sand bodies (i.e. bars), and that transgressive channel-belt elements are commonly isolated from one another, the presence of crevasse-splay deposits in a reservoir can impact production strategies as these deposits can be hydrocarbon reservoirs and can act as pathways for fluid migration and connectivity between different channel sand bodies that are otherwise not connected (Anderson, 2005). Furthermore, in general floodplain sandstone bodies are difficult to image in seismic investigations due to the relatively homogeneous grain-size of floodplain deposits and average bed thicknesses being below seismic resolution. As such, outcrop analogs are critical to better understand the role crevasse splay deposits contribute to the static connectivity of channel sand bodies (i.e. reservoirs).

This article is focused on the transgressive unit of a high-accommodation sequence in a well-exposed fourth-order regressive-transgressive (R-T) stratigraphic cycle (sensu Frazier, 1974; Galloway, 1989; Mitchum and Van Wagoner, 1991) of the Eocene Escanilla Formation, Spain. This unit is unique as transgressive fluvial deposits can be confidently correlated both longitudinally from the fluvial system to the coevally deposited shallow-marine strata (Moss-
Russell, 2009; Pyles et al., in review, Appendix A) and laterally from the axis to the margin of the fluvial system for a distance of over two kilometers. The goals of this article are to quantitatively document, for the first time, spatial patterns in stratigraphic architecture, net-sand content, the size and modal grain size of channel-belt elements, and static connectivity in order to evaluate how stratigraphic architecture varies laterally and vertically within a transgressive fluvial system deposited in a high-accommodation sequence. Concepts and data generated in this study can be used to aid in the interpretation of subsurface data and quantitatively constrain geologic models, thereby reducing uncertainty in the development of reservoirs.

4.3 Geological Setting

The Escanilla Formation is located in the southern part of the Ainsa Basin, Spain (Figs. 4.2A, 4.2B), a sub-basin of the larger Tremp-Ainsa-Jaca Basin, which developed from a foreland basin to a thrust-top (piggy-back) basin south of the Pyrenean axial zone of the South Pyrenean Central Thrust System (Fig. 4.2A; Mutti, 1977; Puigdefabregas et al., 1992; Munoz et al., 1994; Fernandez, 2004). The Ainsa basin extends ~ 40 km (25 mi) in the north-south direction and ~ 25 km (15 mi) in the east-west direction. The Ainsa Basin is located within the Buil Syncline and is bounded by four syndepositionally active structures (Poblet et al., 1998; Dreyer et al., 1999, and Fernandez et al., 2004; Hoffman, 2009): (1) the Boltaña Anticline to the west, and (2) the Mediano Anticline to the east; (3) the Ànisclo Anticline to the north, and (4) the Cotiella Thrust to the northwest (Fig. 4.2D).

The basin-fill succession is divided into the Hecho and Campodarbe Groups, both of which overlie mixed carbonate and siliciclastic pre-growth strata (Fig. 4.2B) (Poblet et al., 1998; Fernandez et al., 2004). The focus of this study is on the Campodarbe Group (Figs. 4.2B, 4.2C),
which is ~2 km (1.2 mi) thick and is divided into the Sobrarbe and Esanilla Formations. These formations record the final filling of the Ainsa Basin and the progradation of a linked shelf-to-basin system over the area (Bentham et al., 1992; Dryer et al., 1999; Pickering and Bayliss, 2009; Moss-Russell, 2009; Silalahi, 2009; Pyles et al., in review; Appendix A).

The Sobrarbe Formation is the basal formation of the Campodarbe Group. Based on biostratigraphic and magnetostratigraphic data, the Sobrarbe Formation was deposited over a period of approximately 3 million years in the Late Leutian (Fig. 4.2C; Dreyer et al., 1999; Mochales et al., 2012), is ~1 km (0.6 mi) thick (Dreyer et al., 1999), with rates of sediment accumulation (undecompressed) being ~32 cm/kyr (1 ft/kyr) (Dreyer et al., 1999; Mochales et al., 2012). During deposition of the Sobrarbe Formation, the basin bounding structures, the Mediano and Boltaña Anticlines were actively growing while the Buil Syncline was subsiding (Mutti et al., 1989; Puigdefabregas et al., 1992; Dreyer et al., 1999; Anastasio and Holl, 2001; Hoffman, 2009; Mochales et al., 2012).

The Sobrarbe Formation contains cyclic alternations between mudstone-dominated delta plain deposits, carbonates, delta front sandstones, collapse complexes, muddy delta slope deposits, and turbidite sandstone. Moss-Russell (2009) and Pyles et al. (in review; Appendix A) divided the Sobrarbe Formation into six condensed section bounded regressive-transgressive (R-T) cycles that roughly correspond to the composite sequences of Dreyer et al. (1999). Each condensed-section bounded R-T cycle is approximately fourth-order in duration, meaning they record approximately 0.1 to 0.5 m.y. of deposition each (sensu Mitchum and van Wagoner, 1991). Each R-T cycle forms a shelf-slope-basin clinothem whereby the location of the shelf edge is located in sequentially basinward (northward) and aggradational (upward) positions from one to the next.
This study is focused on Cycle 2 (Figs. 4.2E, 4.3; Moss-Russell, 2009; Pyles et al., in review, Appendix A), which is interpreted to correspond to Dreyer et al.’s (1999) Camaron Composite Sequence. The lower unit of this cycle contains, from base to top: marine mudstone sheets, delta mouth bars, and distributary channel belts that have both a progradational and aggradational stacking pattern, and are interpreted as the regressive unit of Cycle 2 (Fig. 4.2E; Moss-Russell, 2009; Pyles et al., in review, Appendix A). The upper unit contains, from base to top: tidal channel belts, thin but longitudinally continuous mouth bars, and marine mudstone sheets which have both a retrogradational and aggradational stacking pattern, and are interpreted as the transgressive unit of Cycle 2 (Fig. 4.2E; Moss-Russell, 2009; Pyles et al., in review, Appendix A).

Shoreline trajectory is a measure of temporal change in the location of the paleoshoreline and is quantified as (tan $\theta = dy/dx$) (Posamentier and Vail, 1988; Helland-Hansen and Martinsen, 1996, Pyles et al., 2011). The average shoreline trajectory for the transgressive unit of Cycle 2 is ($dy/dx; 50$ m (164 ft) / $2300$ m (7546 ft)) 0.02 or 1.14°, rising landward (southward) (Moss-Russell, 2009; Pyles et al., in review, Appendix A). The positive shoreline trajectory indicates that the shoreline advanced landward during a relative rise in sea level and therefore the rate at which accommodation was created exceeded sediment input (A>S) (Pyles et al., in review, Appendix A). The lack of an incised valley on the shelf and aggradational trajectory of the shoreline demonstrate that the rate of accommodation was high during the transgressive unit of Cycle 2 and is therefore classified as a high-accommodation sequence (Moss-Russell, 2009; Pyles et al., in review, Appendix A).

Moss-Russell (2009) documented the location of the upper and lower boundaries of Cycle 2 and the boundary between the regressive and transgressive units of Cycle 2 in the study.
area (Fig. 4.3). The outcrop is sufficiently well exposed so that each of these boundaries can be correlated (i.e. walked) directly from the deltaic strata into the coevally deposited fluvial strata of the Escanilla Formation (Fig. 4.3).

The Escanilla Formation interfingers with and conformably overlies the deltaic and shallow marine deposits of the Sobrarbe Formation, is ~ 1.1 km (0.7 mi) thick (Bentham et al., 1992), and unconformably underlies the Oligocene Collegats Formation, a conglomeratic alluvial fan deposit (Fig. 4.2C; Garrido-Meiggs, 1973; Bentham et al., 1992). The Escanilla Formation contains non-marine deposits and is sourced from the Pyrenean massif through the Tremp-Graus Basin to the east (Fig. 4.2A; Garrido-Meiggs 1968; Vincent, 2001, Michael et al., 2014).

Paleocurrents collected from Cycle 2 of the Escanilla Formation are to the northwest/north-northwest, consistent with the coevally deposited Sobrarbe Formation to the north (Fig. 4.3; Bentham et al., 1992, Dreyer et al., 1999; Moss-Russell, 2009). The lower regressive unit of Cycle 2 of the Escanilla Formation contains low aspect ratio (width/thickness) fluvial channel belts that have both a progradational and aggradational stacking pattern (Moss-Russell, 2009; Pyles et al., in review, Appendix A). The focus of this study is on the upper transgressive unit that contains fluvial channel belts interbedded with splay and non-marine floodplain fines that have both a retrogradational and aggradational stacking pattern and is overlain by a thin, gray marine mudstone that demarcates the maximum transgression of the shoreline and the upper most boundary of Cycle 2 (Pyles et al., in review, Appendix A). These fluvial deposits within the study area are located less than one kilometer from the coevally deposited shoreline and can be laterally correlated over several kilometers (Moss-Russell, 2009; Pyles et al., in review, Appendix A).
4.4 Dataset and Methods

Data used to address the goals of the study include: (1) a geologic map that documents the aerial distribution of the boundaries of formations and cycles, strike and dips of bedding surfaces, and paleocurrent measurements (Fig. 4.3); (2) 14 detailed stratigraphic columns totaling 577 m in thickness that document lithology, grain-size, physical sedimentary structures, and stratal boundaries at centimeter-scale resolution; and (3) interpreted photo panels that were used to document the spatial distribution of architectural elements and the location of stratal boundaries. These data were in turn used to quantify proportions of stories, elements, net-sand content, modal grain size of channel-belt elements, and static connectivity between channel-belt elements.

4.4.1 Lithofacies

Eleven lithofacies were identified in this study (Table 4.1). Each is distinctive in terms of grain-size, sedimentary structures, and bed thickness. Percentages and spatial trends of lithofacies are documented for fluvial architectural elements. The eleven facies are grouped into three classes on the basis of lithology and interpreted reservoir quality: (1) seal (F1-F4), (2) baffle (F5-F7), and (3) reservoir (F8-F11). For brevity, descriptive characteristics of lithofacies are presented in Table 4.1.

4.4.2 Fluvial Hierarchy of Architecture Elements

The implementation of a hierarchical scheme is critical in order to describe and quantitatively document the spatial and temporal changes within the stratigraphy of fluvial systems (Miall, 1985). The architectural elements of the Escanilla Formation are grouped into a
three-level hierarchy based on the methodology proposed by Ford and Pyles (2014). From smallest to largest, the three levels are: story, element, and archetype. Each hierarchical level is composed of different combinations of components that account for the variability in sedimentation styles observed in the Escanilla Formation. Each hierarchical level is constrained by stratal surfaces, lithofacies, external shape in depositional strike view of units, and cross-cutting relationships documented in the study area. For this study, quantitative analysis of fluvial architecture is only done at the story and element levels (Figs. 4.4, 4.5).

4.4.2.1 Story

A story is “a meso-scale volume of strata formed from genetically related beds or bedsets produced by the migration, fill or overbank discharge of a single fluvial system” (Ford and Pyles, 2014, pg. 1281). The thickness of each story scales to bank-full discharge and flood-stage water depth. Stories are the fundamental building blocks for larger stratigraphic units: elements and archetypes (Figs. 4.4, 4.5).

Eight different types of stories were identified in the study area and are divided into channel fill components and floodplain fill components (Fig. 4.4). Channel fill components are: lateral accreting, downstream accreting, fine-grained fill associated with lateral accretion, and erosionally based fine-grained fill (Fig. 4.4A). Lateral and downstream accreting stories are interpreted to be the main reservoir for hydrocarbons, whereas the channel fill components are interpreted as baffles. Floodplain fill components are: splay, crevasse channel with heterolithic fill, crevasse channel with erosionally based fine-grained fill, and floodplain fines (Fig. 4.4B). Splays stories are interpreted as a reservoir and/or a conduit for flow of hydrocarbons, potentially increasing the static connectivity of adjacent channel-belt elements. Crevasse channel stories are
interpreted as baffles while floodplain fine stories are interpreted as a seal. Each story is distinctive in terms of cross-sectional shape in depositional strike view, lithofacies, modal grainsize, and sediment transport directions in relation to stratal geometry (Table 4.2). For brevity, descriptive characteristics and photographic examples of stories are presented in Table 4.2 and Fig. 4.4 respectively.

### 4.4.2.2 Element

An element is defined “as a macroscale lithosome produced from the migration and overbank discharge of a single fluvial channel” (Ford and Pyles, 2014, pg. 1294). An element is separated from stratigraphically adjacent elements by floodplain fines or an erosional surface when eroded into by a younger element. An element is composed of one or more stories (Fig. 4.5). Multistory elements are defined as an element that contains more than one story that stack laterally and/or vertically within the element (Ford and Pyles, 2014).

Two types of elements were recognized within the transgressive unit of Cycle 2 (Fig. 4.5): (1) channel-belt elements, and (2) floodplain-belt elements. A channel-belt element is composed of multiple channel fill stories and constitutes ~ 11% of the strata within the transgressive unit. Channel-belt elements contain the best main reservoir quality strata. Three types of channel-belt elements were documented within the transgressive unit. In order of increasing reservoir quality they are: (1) low aspect ratio channel-belt elements, (2) intermediate aspect ratio channel-belt elements, and (3) high aspect ratio channel-belt elements (Fig. 4.5A). Each channel-belt element is unique in terms combinations of channel fill stories, aspect ratio, proportion of lithofacies, bounding surfaces, amount of erosion, and shape in depositional strike view (Table 4.3).
Proportions of lithofacies in channel-belt elements are calculated by dividing the area of each facies type by the area of the channel elements being evaluated. Lithofacies 8 to 11 (Table 4.1) are considered reservoir facies because they have high net sand content. On average (Fig. 4.5A): 85% of the lithofacies in low aspect ratio channel-belt elements are reservoir facies; 93% of the lithofacies in intermediate aspect ratio channel-belt elements are reservoir facies; and 93% of the lithofacies in high aspect ratio channel-belt elements are reservoir facies.

A floodplain-belt element is composed of a combination of multiple floodplain fine stories and constitutes ~ 89% of the strata within the transgressive unit. Floodplain-belt elements are predominantly baffles and seals, however larger splay bodies can be reservoirs. Two types of floodplain-belt elements were documented in the transgressive unit. In order of increasing reservoir quality they are: (1) unassociated splay elements and (2) associated non-coeval splay elements (Fig. 4.5B). For brevity, descriptive characteristics and diagrammatic examples of elements are presented in Table 4.3 and Fig. 4.5B respectively.

4.4.3 Static Connectivity

Funk et al. (2012) quantitatively described static connectivity (C) as the length of sand-on-sand contacts normalized by the total length of the interface: \( C = \frac{\sum l_{si}}{l_{tot}} \), where \( l_{si} \) is the length of individual sand-on-sand contacts between stratigraphically adjacent elements and \( l_{tot} \) is the length of the shared contact between them (Fig. 4.6A). The measure of static connectivity is dimensionless and ranges from 0-to-1 where 0 = no connectivity and 1 = fully connected.

This study quantitatively defines static connectivity as the presence of sand-on-sand contacts between: (1) a channel to splay (C-S) which applies to stratigraphically adjacent channel-belt elements and crevasse splay stories that have sand-on-sand contact between them.
(Fig. 6B) and (2) channel to channel (C-C) which applies to stratigraphically adjacent channel-belt elements that have sand-on-sand contact between them (Fig. 4.6C). The definition assumes that sand-on-sand contacts between adjacent channel-belt elements facilitate fluid migration. All of the channel-belt elements identified in this study have a high propensity for both C-S and C-C static connectivity as each have sandstone juxtaposed to their margins and coarse-grained lags at their base (Fig. 4.4B).

4.5 Geology of the Study Areas

This study quantitatively documents the stratigraphic architecture, net-sand content, the size and modal grain size of channel-belt elements, and static connectivity of the transgressive unit of Cycle 2, a fourth-order, R-T cycle of the Escanilla Formation at three different field areas: (1) the Mondot field area (Figs. 4.3, 4.7); (2) the La Susia field area (Figs. 4.3, 4.8); and the Peñalebrera field area (Fig. 4.3, 4.9). To facilitate a quantitative analysis, a cross section is constructed by projecting all stratigraphic data from these three field areas onto a plane that is orientated normal to the mean paleocurrent direction (Fig. 4.10). The cross section is oriented so that the viewer is looking in the up-current direction. The cross section documents the location of stratigraphic columns; location, size, and shape of architectural elements and stories; rose diagrams of paleocurrent directions; and hierarchical boundaries.

4.5.1 Mondot Field Area

The Mondot field area is a ~ 0.6 km (0.4 mi) wide outcrop located in the western part of the study area below the town of Mondot (Figs. 4.3, 4.7, 4.10). The map, photopanel, and cross section are used to document the following characteristics (Figs. 4.3, 4.7, 4.10 respectively).
First, paleocurrent measurements indicate that sediment exited the outcrop belt in the Mondot field area (vector mean = 008°; circular variance = 0.4). Second, in order of increasing area, this field area contains (Fig. 4.11A): crevasse channel with erosionally based fine-grained fill stories (1%), crevasse channel with heterolithic fill stories (1%), erosionally based fine-grained fill stories (1%), fine-grained fill associated with lateral accretion stories (1%), lateral accreting stories (4%), downstream accreting stories (11%), splay stories (20%), and floodplain fine stories (61%). Third, in order of increasing area, this field area contains (Fig. 4.11A): low aspect ratio channel-belt elements (1%), intermediate aspect ratio channel-belt elements (2%), high aspect ratio channel-belt elements (16%), unassociated splay elements (27%), and associate non-coeval splay elements (54%). Fourth, net-sand content, the ratio of the area of sandstone to total area being evaluated, ranges from 0.61 in the western part of the Mondot field area to 0.47 in the eastern part of the Mondot field area (Figs. 4.10, 4.11A).

Channel-belt elements have an average width of 161 m (528 ft), an average thickness of 4 m (13 ft), and an average aspect ratio of 40 (Fig. 4.11A). The overall sizes of channels increases from one to the next in an upward transect with an average thickness of ~ 1.8 m (6 ft) at the base to ~ 4.5 m (15 ft) at the top of the outcrop. Modal grain size varies from medium-grained sandstone in the lower channels of the outcrop to upper medium-grained sandstone in the upper channels of the outcrop. Individual channel-belt elements have large vertical and lateral offsets between one another resulting in 33% of channel–belt elements having C-C static connectivity. However, due to the high abundance of splay stories, 100% of channel-belt elements have C-S static connectivity (Fig. 4.12). C-C and C-S static connectivity increases from the base to the top of the transgressive unit of Cycle 2 (Fig. 4.12).
4.5.2 La Susia Field Area

The La Susia field area is a ~ 1.1 km (0.7 mi) wide outcrop located in the central part of the study area just north of Rio Susia (Figs. 4.3, 4.8, 4.10) where only the upper half of the transgressive unit of Cycle 2 crops out. Data presented below and in Figs. 4.8, 4.10, and 4.11A represent the exposed part of the transgressive unit of Cycle 2. The map, photopanel, and cross section are used to document the following characteristics (Figs. 4.3, 4.8, and 4.10 respectively). First, paleocurrent measurements indicate that sediment exited the outcrop belt in the La Susia field area (vector mean = 348°; circular variance = 0.3). Second, in order of increasing area, the field area contains (Fig. 4.11A): crevasse channel with heterolithic fill stories (<1%), crevasse channel with erosionally based fine-grained fill stories (<1%), fine-grained fill associated with lateral accretion stories (<1%), lateral accreting stories (1%), downstream accreting stories (6%), splay stories (13%), and floodplain fine stories (79%). Third, in order of increasing area, the field area contains (Fig. 4.11A): high aspect ratio channel-belt elements (1%), intermediate aspect ratio channel-belt elements (6%), associate non-coeval splay elements (40%), and unassociated splay elements (53%). Fourth, the La Susia field area has a net-sand content that ranges from 0.52 in the western part of the La Susia field area to 0.35 in the eastern part of the La Susia field area (Fig 4.10).

Channel-belt elements have an average width of 107 m (351 ft), an average thickness of 4 m (13 ft), and an average aspect ratio of 27 (Fig. 4.11A). The overall sizes of channels decreases from one to the next in an upward transect with an average thickness of ~ 5 m (16 ft) at the base to ~ 2.3 m (8 ft) at the top of the outcrop. Modal grainsize varies from upper medium-grained sandstone in the lower channels of the outcrop to medium-grained sandstone in the upper channels of the outcrop. Individual channel-belt elements have high vertical and lateral offsets
between one another resulting in 27% of channel–belt elements having C-C static connectivity (Fig. 4.12). Due to the moderate abundance of splay stories, 55% of channel-belt elements have C-S static connectivity (Fig. 4.12).

4.5.3 Peñalebrera Field Area

The Peñalebrera field area is a ~ 0.8 km (0.5 mi) wide with continuous outcrop located in the eastern part of the study area (Figs. 4.3, 4.9, 4.10). The transgressive unit of Cycle 2 is completely exposed in the Peñalebrera field area. The map, photopanel and cross section are used to document the following characteristics of this field area (Figs. 4.3, 4.9, and 4.10 respectively). First, paleocurrent measurements are parallel with the outcrop belt in the Peñalebrera field area (vector mean = 350°; circular variance = 0.3). Second, in order of increasing area, the field area contains (Fig. 4.11A): crevasse channel with heterolithic fill stories (1%), lateral accreting stories (1%), downstream accreting stories (2%), splay stories (12%), and floodplain fine stories (84%). Third, in order of increasing area, the field area contains (Fig. 4.11A): intermediate aspect ratio channel-belt elements (3%), associate non-coeval splay elements (37%), and unassociated splay elements (60%). Fourth, the Peñalebrera field area has a net-sand content that ranges from 0.35 in the western part of the Peñalebrera field area to 0.28 in the eastern part of the Peñalebrera field area (Fig 4.10).

Channel-belt elements in this field area have an average width of 39 m (128 ft), an average thickness of 2 m (7 ft), and an average aspect ratio of 19 (Fig. 4.11A). Relative channel size does not vary upwards. Modal grainsize within channels is fine-grained sandstone and does not vary upwards. Individual channel-belt elements have large vertical and lateral offsets between one another resulting in no C-C or C-S static connectivity (Fig. 4.12).
4.6 Discussion

The goals of this article are to quantitatively document spatial patterns in stratigraphic architecture, net-sand content, size and modal grain size of channel-belt elements, and static connectivity in order to evaluate how reservoir characteristics vary laterally and vertically within a transgressive fluvial system deposited in a high-accommodation sequence.

4.6.1 Axis to Margin Changes in Reservoir Characteristics

Below we discuss lateral variations in geological characteristics between the three different field areas (Fig. 4.10; 4.11A). Variations summarized below are interpreted to reflect axis-to-margin variations in the stratigraphy of fluvial strata in the transgressive unit of Cycle 2.

The vast majority of strata in the transgressive unit of Cycle 2 are floodplain-fill elements, followed by channel-belt elements. However, there is a decrease in the percentage of channel-belt elements from west to east across the study area, with 19% of the strata in the Mondot field area being channel-belt elements in comparison to only 3% in the Peñalebrera field area. The largest channel-belt elements are located in the Mondot field area which are ~100% thicker and >400% wider than those located in the Peñalebrera field area. Splay stories are also higher in proportion in the Mondot field area (20%) than the Peñalebrera field area (12%). The Mondot field area has the highest average net-sand content (0.52), with 44% of the sandstone located within channel-fill stories and 56% within splay and crevasse channel stories. In contrast, the net-sand content in the Peñalebrera field area is 0.31, with 19% of the sandstone located within channel-fill stories and 81% within splay and crevasse channel stories. Channel-belt elements located in the Mondot field area have a modal grainsize of medium-grained sandstone, whereas channel-belt elements located in the Peñalebrera field area have a modal grainsize of
fine-grained sandstone. C-C and C-S static connectivity decreases from the Mondot field area to the La Susia field area. Based on these observations the Mondot field area is interpreted to be the axis of the system, the La Susia field area is interpreted to be the off-axis part of the system, and the Peñalebrera field area is interpreted to be the margin of the system. The best reservoir bodies with the best potential for static connectivity are located within the axis of the system (i.e. the Mondot field area).

Focusing of the fluvial system is hypothesized to be due to either structural focusing of the system or by an incised valley. The former hypothesis is preferred. The bounding structures of the Ainsa basin, namely the Boltaña and Mediano Anticlines, were actively growing during deposition of Hecho and Campodarbe Groups (Mutti et al., 1989; Puigdefabregas et al., 1992; Dreyer et al., 1999; Anastasio and Holl, 2001; Hoffman, 2009; Mochales et al., 2012). The location of the deepest part of the basin (paleo-axis of the Buil Syncline) moved systematically southwestward during deposition of the older Hecho Group (Hoffman, 2009; Moody et al., 2012; Gordon et al., in review). Hoffman (2009) interprets this shift to result from foreland flexure related to structural growth of the basin bounding structures and continued movement on a lower detachment fault that underlies the basin. Critically, this location is located ~ 7 km (4.3 mi) southwest of the deepest part of the basin during deposition of the Guaso Formation (Gordon et al., in review). The systematically southwestward migration of the deepest part of the basin therefore continued at least through the deposition of Cycle 2. The valley hypothesis is not favored because Cycle 2 can be correlated across the Ainsa Basin and was found to be laterally continuous and only pinches out near the basin bounding structures, the Mediano and Boltaña Anticlines.
4.6.2 Vertical Changes in Reservoir Characteristics

Below we discuss how stratigraphic architecture in the transgressive unit of Cycle 2 varies vertically. As the axis of the system contains the best potential reservoir properties, data for this analysis is taken only from the Mondot field area.

The overall fourth-order transgressive unit of Cycle 2 is divided into three components: I, II, and III (Fig. 4.10; Chapter 5). The shoreline trajectory is highly retrogradational in component I (-0.01), becomes more aggradational in component II (-0.04), and is highly aggradational, although slightly retrogradational in component III (-0.28) (See Chapter 5 for discussion). The proportions of stories within each component are different (Fig. 4.11B). First, channel fill stories are lowest in proportion in component I (20%), intermediate in component II (22%), and highest in component III (46%). Second, component II contains the highest proportion of floodplain fine stories (60%) whereas; component III contains the lowest proportion of floodplain fine stories (47%).

There is an increase in the number of stories within each channel-belt element and channel-belt element size from component I to component III. Component I contains single story channel-belt elements that have an average thickness of ~ 2.5 m (8 ft), an average width of ~ 112 m (368 ft), and an average aspect ratio of ~ 45. Component II contains single story channel-belt elements that have an average thickness of ~ 3.9 m (13 ft), an average width of ~ 93 m (305), and an average aspect ratio of ~ 24. Component III contains multistory channel-belt elements that have an average thickness of ~ 4.9 m (16 ft), an average width of ~ 251 m (824 ft), and an average aspect ratio of ~ 51.

The net-sand content increases upwards from component I (0.35), to component II (0.38) and finally component III (0.46). In component I, 43% of the sandstone is located within channel
fill stories and 57% within splay and crevasse channel stories. In component II, 55% of the sandstone is located within channel-fill stories and 45% within splay and crevasse channel stories. In component III, 86% of the sandstone is located within channel fill stories and 14% within splay and crevasse channel stories. There is no measured difference in the modal grain size within channel-belt elements deposited within components I and II, that being lower medium-grained sandstone. However there is an increase in the modal grain size within channel-belt elements deposited within component III to medium-grained sandstone.

Component I has no C-C static connectivity however 100% of the channels have C-S static connectivity (Fig. 4.12). Half of the channel-belt elements within Component II have C-C static connectivity and 100% of the channel-belt elements have C-S static connectivity (Fig. 4.12). The amount of C-S static connectivity increases from component I to component II (Fig. 4.12). A third of the channel-belt elements within Component III have C-C static connectivity and 100% of the channel-belt elements have C-S static connectivity. The amount of C-C static connectivity decreases from component II to component III, however the amount of C-S static connectivity increases from component II to component III (Fig. 4.12).

In summary, fluvial strata deposited from one component to the next within the axis of the transgressive component of Cycle 2 are distinctive in terms of shoreline trajectory, proportions of stories, channel-belt element size, net-sand content, modal grain size within channel-belt elements, and static connectivity (Fig. 4.11B). From base to top of the transgressive unit of Cycle 2 there is an overall increase in the percentage of channel-belt elements, multi-story channel-belt elements, channel-belt element size, net-sand content, modal grain-size within channel-belt elements, and static connectivity (Fig. 4.12). The best reservoir bodies with the greatest potential for static connectivity are located at the upper most channel-belt elements of
the transgressive unit (i.e. component III). The observed variations in stratigraphic architecture are attributed to an increase in sediment input in relation to accommodation.

4.7 Conclusions

The transgressive unit of Cycle 2 is an excellent outcrop analog for hydrocarbon reservoirs in transgressive fluvial systems deposited in a high-accommodation setting. This article quantitatively documents for the first time, spatial patterns in stratigraphic architecture, net-sand content, thickness, modal grain size of channel-belt elements, and static connectivity within a transgressive unit of fluvial strata in order to evaluate lateral changes in reservoir characteristics of transgressive fluvial strata.

Significant lateral changes are an increase in the proportion of channel-fill and splay stories, and channel-belt elements at the expense of floodplain fine stories, and an increase in net-sand content, channel-belt element size, modal grain size, and static connectivity from the margin to the axis of the system. As such, the axis of the system contains the best reservoir sand bodies and potential for static connectivity. Significant vertical changes are an upward increase in channel-belt element size, net-sand content, modal grain size within channel-belt elements, and static connectivity. As such, the upper third of the system contains the best reservoir sand bodies and potential for static connectivity.

Data provided herein can be used to reduce uncertainty in the interpretation of subsurface data, provide input to constrain rules-based forward stratigraphic models (i.e., Pyrcz et al. 2005), and provide input to constrain reservoir models in transgressive fluvial systems.
4.8 References

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Figure 4.1 (A) Diagrams of three main systems tracts defined within sequence stratigraphy methods. (B) Diagrams documenting systems tracts and controls on relative sea level, longitudinal cross sections of systems tracts, lateral cross sections of systems tracts, and published non-marine sequence stratigraphic studies for low- and high-accommodation settings.
Figure 4.2 (A) Paleogeographic map of the Tremp-Ainsa-Jaca basin (modified from Michael et al., 2014). (B) Generalized stratigraphy of the Ainsa-Jaca Basin highlighting the Escanilla and Sobrarbe Formations (modified from Pickering and Bayliss, 2009). (C) Chronostratigraphic chart of the Hecho and Campodarbe Groups showing the different lithostratigraphic units located in the Ainsa Basin. (D) Geologic map of the Ainsa Basin that documents the contacts between the different lithostratigraphic units, bounding structures (Boltaña Anticline, Mediano Anticline, and Anisclo Anticline) and the location of the Buil Syncline, which forms the axis of the basin (map modified from Hoffman, 2009). (E) A longitudinal stratigraphic cross section of Cycle 2 of the Sobrarbe Formation documenting: the contact between the Guaso and Sobrarbe Formations, the lower and upper boundaries of Cycle 2, and the regressive/transgressive boundary within Cycle 2; lithology and internal correlations; and the shoreline trajectory at both fifth-order (blue line) and fourth-order (red line) cycles. The location of cross section is shown in Fig. 4.3 (A-A’). From Pyles et al. (in review, Appendix A).
Figure 4.3 Geologic map of the field area documenting: the location of the contacts between the Guaso, Sobrarbe, and Escanilla Formations; the location of the boundaries between Cycle 1-3 (see Fig. 4.2E) as well as the regressive/transgressive boundary within Cycle 2; strike and dips of bedding surfaces; and paleocurrent measurements. The location of the map is shown in Fig. 4.2D.
Figure 4.4 (A) Photographic examples of channel fill stories and (B) floodplain fill stories identified in this study.
Figure 4.5 (A) Diagrams of channel fill stories and elements identified in this study. Elements are subdivided by aspect ratio: high, intermediate, and low. Each element falls within a unique domain of width vs. thickness (see inset table). Pie charts document lithofacies proportions documented for each channel element (see Table 4.1 for lithofacies descriptions). (B) Diagrams of floodplain fine stories and elements identified in this study. (C) Generic illustration of an archetype documented in this study.
Figure 4.6 (A) Static connectivity (C) is defined as $C = (\Sigma l_i)/l_{tot}$ (Funk et al., 2012). This study quantitatively defines static connectivity as the presence of sand-on-sand contacts between: (1) a channel to splay (C-S) and (2) channel to channel (C-C). (B) Interpreted photopanel of a channel-belt element and a splay story with C-S static connectivity. (C) Interpreted photopanel of two channel-belt elements with C-C static connectivity.
Figure 4.7 Uninterpreted (top) and interpreted (bottom) photopanels of the Mondot field area documenting: the upper boundary of Cycle 2; the regressive/transgressive boundary within Cycle 2, and archetype, element, story, and stratal boundaries; the distribution of stories; and the location of stratigraphic columns. The location of the Mondot field area is shown in Fig. 4.3.
Figure 4.8 Uninterpreted (top) and interpreted (bottom) photopanels of the La Susia field area documenting: the upper boundary of Cycle 2; the regressive/transgressive boundary within Cycle 2; and archetype, element, story, and stratal boundaries; the distribution of architectural elements; and the location of stratigraphic columns. The location of the La Susia field area is shown in Fig. 4.3.
Figure 4.9 Uninterpreted (top) and interpreted (bottom) photopanels of the Peñalebrera field area documenting: the upper boundary of Cycle 2; the regressive/transgressive boundary within Cycle 2, and archetype, element, story, and stratal boundaries; the distribution of architectural elements; and the location of stratigraphic columns. The location of the Peñalebrera field area is shown in Fig. 4.3.
Figure 4.10 Lateral stratigraphic cross section of transgressive unit of Cycle 2 that was constructed by projecting all stratigraphic data from the three field areas onto a plane that is orientated normal to the mean paleocurrent direction. The cross section is oriented so that the viewer is looking in the up-current direction. This is a north-facing outcrop and therefore, east is on the left hand side of the cross section, and west is on the right. The cross section documents the upper boundary of Cycle 2, the regressive/transgressive boundary within Cycle 2; the distribution of architectural elements; paleocurrent measurements; the location of stratigraphic columns; and documentation of lateral changes in stories, net-sand content. The location of the photopanel is shown in Fig. 4.3.
Figure 4.11 (A) Quantitative data documenting axis-to-margin changes in the stratigraphy of the fourth-order transgressive unit of Cycle 2. See Fig. 4.3 and 4.10 for the locations of the three areas. (B) Quantitative comparison of vertical changes in stratigraphic architecture for components I, II, and III.
Figure 4.12 Charts documenting variations in static connectivity for channel to splay (C-S) and channel to channel (C-C) static connectivity for (A) each field area and (B) for components I, II, and III within the axis of the system. (C) Total static connectivity for the entire field area as well as a summary diagram of lateral and vertical variations in static connectivity.
Table 4.1 Descriptions of the 11 lithofacies identified in this study.

<table>
<thead>
<tr>
<th>Lithofacies</th>
<th>Name</th>
<th>Description</th>
<th>Interpreted Hydrodynamic Process</th>
<th>Reservoir Property</th>
</tr>
</thead>
<tbody>
<tr>
<td>F1</td>
<td>Carbonaceous mudstone</td>
<td>Single bed of massive carbonate mudstone; sharp base and top; 5-10 cm thick and 10-50 m wide.</td>
<td>Suspension fall-out; low energy</td>
<td>Seal</td>
</tr>
<tr>
<td>F2</td>
<td>Gray siltstone to mudstone</td>
<td>Thinly laminated to structureless mudstone to siltstone with burrows and rare symmetrical ripples; Sharp base and gradational to erosional top; 1-3 m thick and laterally extensive (100s m).</td>
<td>Lower flow-regime; tractive to suspension fall-out; low energy</td>
<td>Seal</td>
</tr>
<tr>
<td>F3</td>
<td>Varicolored siltstone to mudstone</td>
<td>Massive bedded tan, burgundy, purple, and orange colored mudstone to siltstone with minor to abundant burrows, root casts, and/or organic matter; gradational to sharp base and gradational to erosional top; 0.3-10 m thick and laterally extensive (km’s).</td>
<td>Lower flow-regime; tractive to suspension fall-out; low energy</td>
<td>Seal</td>
</tr>
<tr>
<td>F4</td>
<td>Varicolored mudstone to fine-grained sandstone with slickensides</td>
<td>Massive bedded tan, burgundy, purple, and orange colored mudstone to siltstone with minor to abundant burrows, root casts, and/or organic matter; decimeter to meter tall gypsum filled slickensides throughout facies and cuts across bedding surfaces; gradational to sharp base and gradational to erosional top; 0.5-7 m thick and laterally extensive (100s m).</td>
<td>Lower flow-regime; tractive to suspension fall-out; low energy</td>
<td>Baffle</td>
</tr>
<tr>
<td>F5</td>
<td>Bioturbated and/or rooted structureless to rippled very fine-grained sandstone intercalated with siltstone</td>
<td>Thin bedded structureless or rippled, well sorted, very fine-grained sandstone intercalated with siltstone and contains abundant burrows and/or root casts and is often mottled; sharp base and sharp to gradational top; 0.05-2 m thick and 10’s-100’s of m wide.</td>
<td>Lower flow-regime; tractive to suspension fall-out; low energy</td>
<td>Baffle</td>
</tr>
<tr>
<td>F6</td>
<td>Structureless, planar laminated, and rippled very fine-grained sandstone intercalated with siltstone</td>
<td>Thin bedded structureless, planar laminated, or rippled, moderately sorted very fine- to fine-grained sandstone intercalated with siltstone; Facies coarsens upward; Contains moderate burrows and rare root casts with centimeter to decimeter scale internal erosion and reactivation surfaces; Sandstone beds may also be amalgamated; sharp to erosional base and gradational to erosional top; 0.3-3 m thick and 10’s to 100’s m wide.</td>
<td>Lower flow-regime; tractive deposition; moderate energy</td>
<td>Baffle</td>
</tr>
<tr>
<td>F7</td>
<td>Rippled very fine-grained sandstone intercalated with siltstone</td>
<td>Thin bedded rippled, well sorted, very fine- to fine-grained sandstone intercalated with siltstone and contains moderate burrows and some root casts at the upper contact; Facies fines upward; erosional to onlapping base and gradational, sharp, or erosional top; decimeters to meters in thickness and 10’s of meters wide.</td>
<td>Lower flow-regime; tractive to suspension fall-out deposition; low energy</td>
<td>Baffle</td>
</tr>
<tr>
<td>F8</td>
<td>Trough cross-bedded fine- to medium-grained sandstone</td>
<td>Thin bedded trough cross-bedded, poorly sorted, fine- to medium-grained sandstone with moderate burrowing at the top of beds; Facies fines upwards; gradational to erosional base and gradational to sharp top; decimeters thick and 10’s of meters wide.</td>
<td>Lower flow-regime; tractive deposition; high energy</td>
<td>Reservoir</td>
</tr>
<tr>
<td>F9</td>
<td>Structureless and planar cross-bedded medium- to coarse-grained sandstone with pebbles</td>
<td>Thick bedded structureless and planar cross-bedded, poorly sorted, medium- to coarse-grained sandstone with imbricated pebble sized clasts that line the base of cross-beds; cross-beds are parallel to accretionary surfaces. In structureless sandstone, clasts line accretionary surfaces; facies fines upward; erosional base and gradational to erosional top; decimeters to meters in thickness and 10’s of meters wide.</td>
<td>Lower flow-regime; tractive deposition; high energy</td>
<td>Reservoir</td>
</tr>
<tr>
<td>F10</td>
<td>Trough cross-bedded medium- to coarse-grained sandstone with pebbles</td>
<td>Thick bedded trough cross-bedded, poorly sorted, medium- to coarse-grained sandstone with imbricated pebble sized clasts; Clasts are generally one clast thick stringers which line the base of trough cross-beds; erosional scours common within facies; facies fines upward, erosional base and gradational to erosional top; centimeter's to meters in thickness and 10’s m wide.</td>
<td>Lower flow-regime; tractive deposition; high energy</td>
<td>Reservoir</td>
</tr>
<tr>
<td>F11</td>
<td>Pebble to cobble conglomerate with medium- to coarse-grained sandstone lenses</td>
<td>Thick bedded clast supported conglomerate with imbricated pebble to cobble sized clasts in a poorly sorted, coarse- to very coarse-grained sandstone matrix; Trough cross-bedded coarse- to very coarse-grained sandstone lenses that are centimeters in thickness and decimeters in width are common; Erosional base with up to 3 m of erosion into underlying strata with mudstone clasts above erosional surfaces; gradational to erosional top; 0.5-2 m in thickness; thickness and grainsize decreases laterally and upward along accretionary surfaces; decimeter-10’s of meters in width.</td>
<td>Lower flow-regime; tractive deposition; very high energy</td>
<td>Reservoir</td>
</tr>
</tbody>
</table>
Table 4.2 Descriptions of the different stories identified in this study.

<table>
<thead>
<tr>
<th>Channel Fill Components</th>
<th>Story</th>
<th>Average: Thickness; Width, Aspect Ratio</th>
<th>Bounding Surfaces</th>
<th>Depositional Strike View</th>
<th>Lithofacies Composition</th>
<th>Modal Grain Size</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Lateral accreting</td>
<td>4.3m; 30 m; 7 1%</td>
<td>Lower: erosional (2.5-4 m)</td>
<td>conformable except when eroded into by younger strata</td>
<td>thickest in the axis and thins either abruptly or gradually toward its lateral margins</td>
<td>F9 (72%); F8 (18%); F11 (10%)</td>
<td>upper fine-grained sandstone</td>
</tr>
<tr>
<td></td>
<td>Downstream accreting</td>
<td>4 m; 160 m; 35 9%</td>
<td>erosional (0.5-7 m)</td>
<td>conformable except when eroded into by younger strata</td>
<td>thickest in the axis and thins either abruptly or gradually toward its lateral margins</td>
<td>F9 (54%); F10 (27%); F8 (10%); F11 (9%)</td>
<td>upper medium-grained sandstone</td>
</tr>
<tr>
<td></td>
<td>Fine-grained fill associated with lateral accretion</td>
<td>2.1 m; 50 m; 24 &lt;1%</td>
<td>convex upward, conformable</td>
<td>conformable</td>
<td>bowl shaped, asymmetrical, thickest in the axis and thins toward lateral margins; onlaps the margin adjacent to bars and is erosional on opposite margin</td>
<td>F7 (76%); F6 (24%)</td>
<td>very fine-grained sandstone</td>
</tr>
<tr>
<td></td>
<td>Erosionally based fine-grained fill</td>
<td>1.3 m; 58 m; 45 &lt;1%</td>
<td>convex upward and highly erosional (0.5-1.8 m)</td>
<td>conformable</td>
<td>bowl shaped, symmetrical, thickest in the axis and thins toward lateral margins</td>
<td>F7 (79%); F2 (21%)</td>
<td>lower very fine-grained sandstone</td>
</tr>
<tr>
<td></td>
<td>Splay</td>
<td>3.4 m; 325 m; 96 15%</td>
<td>conformable to locally erosional (0.01 -1 m)</td>
<td>convex up and conformable except when eroded into by younger strata</td>
<td>wedge shaped, thickest and coarsest in axis, and thins and fines gradually toward lateral margins</td>
<td>F6 (66%); F5 (34%)</td>
<td>lower very fine-grained sandstone</td>
</tr>
<tr>
<td></td>
<td>Crevasse channel with heterolithic fill</td>
<td>1.5 m; 30 m; 20 &lt;1%</td>
<td>erosional (0.2-3.2 m)</td>
<td>conformable to undulatory</td>
<td>bowl shaped, symmetrical, thickest in the axis and thins toward lateral margins</td>
<td>F6 (100%)</td>
<td>fine-grained sandstone</td>
</tr>
<tr>
<td></td>
<td>Crevasse channel with erosionally based fine-grained fill</td>
<td>1.2 m; 15 m; 12 &lt;1%</td>
<td>erosional (0.5-1.8 m)</td>
<td>conformable except when eroded into by younger strata</td>
<td>bowl shaped, symmetrical, thickest in the axis and thins toward lateral margins</td>
<td>F5 (100%)</td>
<td>lower very fine-grained sandstone</td>
</tr>
<tr>
<td></td>
<td>Floodplain Fill Components</td>
<td>4m; &lt;10 km; unknown 73%</td>
<td>conformable</td>
<td>conformable to undulatory except when eroded into by younger strata</td>
<td>rectangular to wedge shaped, thickest in the axis and thins toward lateral margins</td>
<td>F3 (94.7%); F4 (5%); F1 siltstone (0.3%)</td>
<td>fine-grained strata deposited during waning flood-stage flow due to suspension fallout and tractive flow</td>
</tr>
</tbody>
</table>
Table 4.3 Descriptions of the different architectural elements identified in this study.

<table>
<thead>
<tr>
<th>Element</th>
<th>Average: Thickness; Width, Aspect Ratio</th>
<th>% of Element</th>
<th>Bounding Surfaces</th>
<th>Depositional Strike View</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Low aspect ratio channel-belt</td>
<td>4.3 m; 50 m; 12</td>
<td>9%</td>
<td>erosional (2.5-4 m)</td>
<td>conformable except when eroded into by younger strata</td>
<td>thickest in the axis and thins either abruptly or gradually toward its lateral margins</td>
</tr>
<tr>
<td>Intermediate aspect ratio channel-belt</td>
<td>3.4 m; 77 m; 23</td>
<td>64%</td>
<td>erosional (0.5-5 m)</td>
<td>conformable except when eroded into by younger strata</td>
<td>thickest in the axis and thins either abruptly or gradually toward its lateral margins</td>
</tr>
<tr>
<td>High aspect ratio channel-belt</td>
<td>4.5 m; 250 m; 52</td>
<td>27%</td>
<td>erosional (0.5-7 m)</td>
<td>conformable except when eroded into by younger strata</td>
<td>thickest in the axis and thins either abruptly or gradually toward its lateral margins</td>
</tr>
<tr>
<td>Unassociated splay</td>
<td>3.4 m; 325 m; 96</td>
<td>42%</td>
<td>conformable</td>
<td>conformable</td>
<td>thickest in the axis and thins gradually toward its lateral margins</td>
</tr>
<tr>
<td>Associated non-coeval splay</td>
<td>3.6 m; 325 m; 90</td>
<td>58%</td>
<td>conformable</td>
<td>Conformable except when eroded into by younger strata</td>
<td>thickest in the axis and thins gradually toward its lateral margins</td>
</tr>
</tbody>
</table>
5.1 Abstract

Non-marine sequence stratigraphic models have primarily focused on low-accommodation sequences, whereas, few studies have focused on fluvial strata deposited in a high-accommodation settings. The Escanilla and Sobrarbe Formations of the Ainsa Basin contain world-class outcrops of coevally deposited fluvial-deltaic-deepwater strata deposited in a high-accommodation setting. This study uses outcrop data to document how stratigraphic architecture, net-sand content, grainsize, and static connectivity of fluvial channels relate to changes in shoreline trajectory within a high-accommodation transgressive unit at fourth- and fifth-order scales of cyclicity. At the smaller fifth-order cycle scale, fluvial strata associated with basinward shoreline trajectories contain higher percentages of channel-belt elements and associated non-coeval splay elements, a higher net-sand content, larger channel-belt elements, and larger modal grain sizes within channel-belt elements than fluvial strata associated with landward shoreline trajectories. Fluvial strata associated with landward trajectories contain higher percentages of
floodplain fine and splay stories and a lower modal grain size within channel-belt elements than
fluvial strata associated with basinward shoreline trajectories. Changes in shoreline trajectory
and fluvial architecture at the fifth-order cycle scale are interpreted to be due to the autogenic
process of avulsion. Within the larger fourth-order transgressive unit, there is an upward increase
in the steepness of the shoreline trajectory which is associated with: an increase in channel-belt
element size, net-sand content, and modal grain size within channel-belt elements; an upward
change from isolated single-story channel-belt elements to clustered multi-story channel-belt
elements; and an upward decrease in the number of splay stories. These changes are interpreted
to be due to the allogenic processes of increasing sediment input in relation to accommodation.

5.2 Introduction

“Sequence stratigraphy is the study of rock relationships within a chronostratigraphic
framework of repetitive, genetically related strata bounded by surfaces of erosion or
nondeposition, or their correlative conformity” (Van Wagoner et al., 1988, p. 39). Early work in
sequence stratigraphy focused primarily on shallow marine and marine strata (Sloss, 1963;
Frazier, 1974; Mitchum et al., 1977; Vail et al., 1977). In these environments, sequences form in
response to the dynamic interaction between the rate of the creation of accommodation space,
which is controlled by eustasy and subsidence, and the rate of sediment input, which are
controlled by climate and tectonics (Sloss, 1962; Curay, 1964; Curtis, 1970; Swift, 1975; Vail et
al., 1977; Muto and Steel, 1997). Early work in marine sequence stratigraphic models have
placed strong emphasis on global changes in sea level to account for changes in shoreline
trajectory and the subsequent architecture of the resultant stratigraphy (Sloss, 1962; Curay,
1964; Curtis, 1970; Swift, 1975; Vail et al., 1977; Posamentier and Vail, 1988).
If sediment input is held constant, variations in the rate of subsidence and eustatic sea level create a continuum of sequences: low-accommodation sequences, intermediate-accommodation sequences, and high-accommodation sequences (Fig. 5.1; Vail et al., 1977; Posamentier and Vail, 1988; Mitchum and Van Wagoner, 1991). At one end of the continuum are low-accommodation sequences, which form when the rate and magnitude of subsidence is much less than the rate and magnitude of eustatic sea level change (Fig. 5.1A; a Type 1 sequence sensu Posamentier and Vail, 1988). They are composed of lowstand, transgressive, and highstand systems tracts (Fig. 5.1; Vail et al., 1977). Low-accommodation sequences are characterized by incised valleys on the shelf that form from incision and sediment bypass during base-level fall (i.e. a drop in relative sea level). Incised valleys are then filled during the subsequent rise in base level (i.e. a rise in relative sea level) (Fig. 5.1B; 5.1C). The valley-fill succession consists of amalgamated fluvial deposits that are overlain by tidally influenced strata (Hampson et al., 1997; Varney, 2000; Bowen and Weimer, 2003; Kirschbaum and Shenk, 2011). Due to the low amount of subsidence, only enough accommodation is available to fill the incised valley before base level begins to fall again (Fig. 5.1C; Hampson et al., 1997; Varney, 2000; Zaitlin et al., 2002; Bowen and Weimer, 2003; Kirschbaum and Shenk, 2011).

At the other end of the continuum are high-accommodation sequences which form when the rate and magnitude of subsidence is greater than the rate and magnitude of eustatic sea level change (Fig. 5.1A; a Type 2 sequence sensu Posamentier and Vail, 1988). As a result, they are composed of only transgressive and highstand systems tracts and no incised valley is formed on the shelf (Figs. 5.1B, 5.1C). The transgressive systems tract consists of widespread isolated fluvial deposits encased in floodplain deposits (Fig. 5.1C; Wright and Marriott, 1993; Burns et al., 1997; Rygel and Gibling, 2006; Fanti and Cantuneanu, 2010). During early highstand,
accommodation is reduced, floodplains are reworked, and the abundance of channel bodies is higher (Fig. 5.1C; Burns et al., 1997; Rygel and Gibling, 2006; Fanti and Cantuneanu, 2010).

One of the unique aspects of high-accommodation sequences is the preservation potential of transgressive fluvial deposits. In low and intermediate sequences, transgressive fluvial deposits are either partially or completely eroded away by either the transgressive surface of erosion and/or the regressive surface of marine erosion (Fig. 5.1). Intermediate-accommodation sequences are part of the continuum of sequences that fall between low- and high-accommodation sequences (e.g. Shanley and McCabe, 1991, 1994; Wright and Marriott, 1993; Legarreta and Uliana, 1998; Gardner et al., 2004; Holbrook et al., 2006).

In the last few decades, several studies applied sequence stratigraphic principles and concepts to non-marine stratigraphy (Posamentier and Vail, 1988; Shanley and McCabe, 1991, 1994; Miall, 1991; Schumm, 1993; Wright and Marriott, 1993; Legarreta and Uliana, 1998; Holbrook et al., 2006). Current non-marine sequence stratigraphic models, however, have primarily focused on low- and intermediate-accommodation sequences and the relationship between fluvial architecture and changes in relative sea level (e.g. Shanley and McCabe, 1991, 1994; Alexander, 1992; Wright and Marriott, 1993; Schumm, 1993; Aitkin and Flint, 1995; Olsen et al., 1995; Martinsen et al., 1999; Plint et al., 2001; Posamentier, 2001; Arnot et al., 2002; Gardner et al., 2004; Holbrook et al., 2006). Therefore, there is an opportunity to expand our knowledge of fluvial systems by working in high-accommodation settings.

Some limitations and complications in existing fluvial sequence stratigraphic models include the following. First, the majority of non-marine sequence stratigraphic models are based on observations made 10’s to 100’s of kilometers from the coeval shoreline making direct correlation of stratal units difficult to impossible (i.e. Boyd et al., 1989; Shanley and McCabe,
Hampson et al., 1997; Aslan and Autin, 1999; Plint et al., 2001; Arnot et al., 2002; Holbrook et al., 2006; Cleveland et al., 2007; Fanti and Catuneanu, 2010). Second, it is difficult to objectively discriminate between allogetic processes such as climate and tectonics, and autogenic processes such as channel avulsion (i.e. Cross, 1988; Boyd et al., 1989; Gensous et al., 1993; Shanley and McCabe, 1994; Valasek, 1995; Van Wagoner, 1995; Blum and Tornqvist, 2000; Muto and Steel, 2001; Stouthamer and Berendsen, 2007). Third, few studies focus on fluvial strata deposited in high-accommodation settings (i.e. Burns et al., 1997; Fanti and Catuneanu, 2010).

This article is focused on the transgressive unit of a high-accommodation sequence in a well exposed fourth-order regressive-transgressive (R-T) stratigraphic cycle (sensu Frazier, 1974; Galloway, 1989; Mitchum and Van Wagoner, 1991) of the Eocene Sobrarbe and Escanilla Formations, Spain. This unit is ideal to study for the following reasons. First, fluvial deposits can be confidently correlated from the fluvial system to the coevally deposited shallow-marine strata (Moss-Russell, 2009; Pyles et al., in review, Appendix A). Second, fluvial deposits can be studied at two different orders of R-T cycles: fourth and fifth order. The goals of this article are to quantitatively document spatial patterns in stratigraphic architecture, net-sand content, thickness and modal grain size of channel-belt elements, and static connectivity between channel-belt elements of fluvial strata within a transgressive unit of a high-accommodation fourth-order R-T stratigraphic cycle. This information is used to evaluate: (1) differences between fluvial strata deposited when the shoreline trajectory was moving basinward versus strata deposited when the shoreline trajectory was moving landward for fifth-order R-T cycles within the transgressive unit of the fourth-order R-T cycle, (2) differences in fluvial strata deposited from the base to the top of the fourth-order transgressive unit, and (3) the role of
autogenic and allogenic processes on the stratigraphic architecture of fluvial deposits and shoreline trajectory at both fourth-order and fifth-order scales.

5.3 Geological Setting

The Sobrarbe and Escanilla Formations crop out in the southern part of the Ainsa Basin, Spain (Figs. 5.2A, 5.2B), a sub-basin of the larger Tremp-Ainsa-Jaca Basin, which developed from a foreland basin to a thrust-top (piggy-back) basin south of the axial zone of the South Pyrenean Central Thrust System (Mutti, 1977; Puigdefabregas et al., 1992; Munoz et al., 1994; Fernandez, 2004). The Ainsa basin extends ~ 40 km in the north-south direction and ~ 25 km in the east-west direction. The Ainsa Basin is located within the Buil Syncline and is bounded by four syndepositionally active structures (Poblet et al., 1998; Dreyer et al., 1999, and Fernandez et al., 2004; Hoffman, 2009): (1) the Boltaña Anticline to the west, and (2) the Mediano Anticline to the east; (3) the Ñisclo Anticline to the north, and (4) the Cotiella Thrust to the northwest (Fig. 5.3).

The basin-fill succession is divided into the Hecho and Campodarbe Groups, both of which overlie mixed carbonate and siliciclastic pre-growth strata (Figs. 5.2B, 5.2C) (Poblet et al., 1998; Fernandez et al., 2004). The focus of this study is on the Campodarbe Group (Fig. 5.2), which is ~ 2 km thick and is divided into the Sobrarbe and Escanilla Formations. These formations record the final filling of the Ainsa Basin and the progradation of a linked shelf-to-basin system over the area (Bentham et al., 1992; Dryer et al., 1999; Pickering and Bayliss, 2009; Moss-Russell, 2009; Silalahi, 2009; Pyles et al., in review, Appendix A).

The Sobrarbe Formation is the basal formation of the Campodarbe Group and represents the youngest marine strata in the Ainsa basin-fill succession. Based on biostratigraphic and
magnetostratigraphic data, the Sobrarbe Formation was deposited over a period of approximately 3 million years in the Late Leutian (Fig. 5.2C; Dreyer et al., 1999; Mochales et al., 2012), is ~ 1 km thick (Dreyer et al., 1999), with rates of sediment accumulation (undecompressed) being ~ 32 cm/kyr (Dreyer et al., 1999; Mochales et al., 2012). During deposition of the Sobrarbe Formation, the basin bounding structures, the Mediano and Boltaña Anticlines were actively growing while the Buil Syncline was subsiding (Fig. 5.2C; Mutti et al., 1989; Puigdefabregas et al., 1992; Dreyer et al., 1999; Anastasio and Holl, 2001; Hoffman, 2009; Mochales et al., 2012).

The Sobrarbe Formation contains cyclic alternations between mudstone-dominated delta plain deposits, carbonates, delta front sandstones, collapse complexes, muddy delta slope deposits, and turbidite sandstone. Moss-Russell (2009) and Pyles et al. (in review, Appendix A) divided the Sobrarbe Formation into six condensed section bounded regressive-transgressive (R-T) cycles (Fig. 5.2D) that roughly correspond to the composite sequences of Dreyer et al. (1999). Each condensed-section bounded R-T cycle is approximately fourth-order in duration, meaning they record approximately 0.1 to 0.5 m.y. of deposition each (sensu Mitchum and van Wagoner, 1991). Each R-T cycle forms a shelf-slope-basin clinothem whereby the location of the shelf edge is located in sequentially basinward (northward) and aggradational (upward) positions from one to the next (Fig. 5.2D).

This study is focused on Cycle 2 (Figs. 5.2D, 5.3; Moss-Russell, 2009; Pyles et al., in review, Appendix A), which is interpreted to correspond to Dreyer et al.’s (1999) Camaron Composite Sequence. The lower unit of this cycle contains, from base to top: marine mudstone sheets, delta mouth bars, and distributary channel belts that have both a progradational and aggradational stacking pattern, and are interpreted as the regressive unit of Cycle 2 (Moss-Russell, 2009; Pyles et al., in review, Appendix A). The upper unit contains, from base to top:
tidal channel belts, thin but longitudinally continuous mouth bars, and marine mudstone sheets which have both a retrogradational and aggradational stacking pattern, and are interpreted as the transgressive unit of Cycle 2 (Moss-Russell, 2009; Pyles et al., in review, Appendix A). Moss-Russell (2009) documented the location of the upper and lower boundaries of Cycle 2 and the boundary between the regressive and transgressive units of Cycle 2 in the study area (Fig. 5.3). The outcrop is sufficiently well exposed so that each of these boundaries can be correlated (i.e. walked) directly from the deltaic strata into the coevally deposited fluvial strata of the Mondot Member of the Escanilla Formation (Fig. 5.3).

The Escanilla Formation interfingers with and conformably overlies the deltaic and shallow-marine deposits of the Sobrarbe Formation, is ~ 1.1 km thick (Bentham et al., 1992), and unconformably underlies the Oligocene Collegats Formation, a conglomeratic alluvial fan deposit (Fig. 5.2; Garrido-Mégiás, 1973; Bentham et al., 1992). The Escanilla Formation contains non-marine deposits and is sourced from the Pyrenean massif through the Tremp-Graus Basin to the east (Fig. 5.2A; Garrido-Mégiás 1968; Vincent, 2001, Michael et al., 2014).

Dryer et al. (1993) divided the Escanilla Formation into the lower Mondot and upper Olson Members. The Mondot Member is a transitional unit between the deltaic Sobrarbe Formation and the fully fluvial Olson member (Fig. 5.2). Paleocurrents collected from the Mondot Member are to the northwest/north-northwest, consistent with the coevally deposited Sobrarbe Formation to the north (Fig. 5.3; Bentham et al., 1992, Dryer et al., 1999; Moss-Russell, 2009).

The lower regressive unit of Cycle 2 of the Escanilla Formation contains low aspect ratio (width/thickness) fluvial channel belts that have both a progradational and aggradational stacking pattern (Moss-Russell, 2009; Pyles et al., in review, Appendix A). The focus of this study is on
the upper transgressive unit that contains fluvial channel belts interbedded with splay and non-marine floodplain fines and is overlain by a thin, gray marine mudstone that demarcates the maximum transgression of the shoreline and the upper most boundary of Cycle 2 (Pyles et al., in review, Appendix A). These fluvial deposits are located less than one kilometer from the coevally deposited shoreline (Moss-Russell, 2009; Pyles et al., in review, Appendix A).

5.4 Dataset and Methods

Data used to address the goals of the study include: (1) a geologic map that documents the aerial distribution of the boundaries of formations and cycles, strike and dips of bedding surfaces, and paleocurrent measurements (Fig. 5.3); (2) 11 detailed stratigraphic columns totaling 433 m in thickness that document lithology, grain-size, physical sedimentary structures, and stratal boundaries at centimeter-scale resolution; and (3) interpreted photo panels that were used to document the spatial distribution of architectural elements and the location of stratal boundaries. These data were in turn used to quantify proportions of stories, elements, net-sand content, modal grain size of channel-belt elements, and static connectivity between channel-belt elements.

5.4.1 Fluvial Hierarchy of Architecture Elements

The implementation of a hierarchical scheme is critical in order to describe and quantitatively document the spatial and temporal changes within the stratigraphy of fluvial systems. The architectural elements of the Escanilla Formation are grouped into a three-level hierarchy based on the methodology proposed by Ford and Pyles (2014). From smallest to largest, the three levels are: story, element, and archetype. Each hierarchical level is composed of
different combinations of components that account for the variability in sedimentation styles observed in the Escanilla Formation. Each hierarchical level is constrained by stratal surfaces, lithofacies, external shape in depositional strike view of units, and cross-cutting relationships documented in the study area. For this study, quantitative analysis of fluvial architecture is only conducted at the story and element levels (Figs. 5.4A, 5.4B).

**5.4.1.1 Story**

A story is “a meso-scale volume of strata formed from genetically related beds or bedsets produced by the migration, fill or overbank discharge of a single fluvial system” (Ford and Pyles, 2014, pg. 1281). The thickness of each story scales to bank-full discharge and flood-stage water depth. Stories are the fundamental building blocks for larger stratigraphic units: elements and archetypes (Fig. 5.4A).

Eight different types of stories were identified in the study area and are divided into channel fill components and floodplain fill components (Fig. 5.4A). Channel fill components are: lateral accreting, downstream accreting, fine-grained fill associated with lateral accretion, and erosionally based fine-grained fill (Fig. 5.4A). Floodplain fill components are: splay, crevasse channel with heterolithic fill, crevasse channel with erosionally based fine-grained fill, and floodplain fines (Fig. 5.4A). Each story is distinctive in terms of cross-sectional shape in depositional strike view, lithofacies, modal grainsize, and sediment transport directions in relation to stratal geometry. For brevity, descriptive characteristics and photographic examples of stories are presented in Table 5.1 and Fig. 5.4A respectively.
5.4.1.2 Element

An element is defined “as a macroscale lithosome produced from the migration and overbank discharge of a single fluvial channel” (Ford and Pyles, 2014, pg. 1294). An element is separated from stratigraphically adjacent elements by floodplain fines or an erosional surface when eroded into by a younger element. An element is composed of one or more stories (Fig. 5.4B). Multistory elements are defined as an element that contains more than one story that stack laterally and/or vertically within the element (Ford and Pyles, 2014).

Two types of elements were recognized within the transgressive unit of Cycle 2 (Fig. 5.4B): (1) channel-belt elements, and (2) floodplain-belt elements. A channel-belt element is composed of multiple channel fill stories and constitutes ~ 16% of the strata within the transgressive unit. Three types of channel-belt elements were documented within the transgressive unit: (1) low aspect ratio channel-belt elements, (2) intermediate aspect ratio channel-belt elements, and (3) high aspect ratio channel-belt elements (Fig. 5.4B). Each channel-belt element is unique in terms combinations of channel fill stories, aspect ratio, bounding surfaces, amount of erosion, and shape in depositional strike view. A floodplain-belt element is composed of a combination of multiple floodplain fine stories and constitutes ~ 84% of the strata within the transgressive unit. Two types of floodplain-belt elements were documented in the transgressive unit: (1) associated non-coeval splay elements and (2) unassociated splay elements (Fig. 5.4B). Associated non-coeval floodplain-belt elements are always associated with a channel-belt element that erodes into the underlying splay and crevasse channel stories and are interpreted to represent the progradation of a crevasse splay complex into a floodplain and the full avulsion of a channel-belt element, whereas, unassociated floodplain-belt elements are not associated with a channel-belt element and are interpreted to represent the progradation of a
crevasse splay complex into a floodplain and a failed avulsion of a channel-belt element. For brevity, descriptive characteristics and diagrammatic examples of elements are presented in Table 5.2 and Fig. 5.4B respectively.

### 5.4.2 Shoreline Trajectory

Shoreline trajectory is a measure of temporal change in the location of the paleoshoreline and is quantified as \( \tan \theta = \frac{dy}{dx} \) (Fig. 5.5A; Posamentier and Vail, 1988; Helland-Hansen and Martinsen, 1996, Pyles et al., 2011). This study focuses on the shoreline trajectory only within the transgressive unit of Cycle 2, at two different scales: fourth- and fifth-order cycles.

The shoreline trajectories mapped within the deltaic deposits of the Sobrarbe Formation were physically correlated to time-equivalent fluvial deposits of the Escanilla Formation (Figs. 5.5B, 5.5C), which facilitates a quantitative study of how changes in fluvial architecture relate to changes in shoreline trajectory. This study relates the shoreline trajectory for fifth-order R-T cycles and for the three different components of the fourth-order transgressive unit of Cycle 2 to the corresponding architecture in the fluvial system to test how changes in the shoreline trajectory relate to fluvial architecture.

The Sobrarbe Formation was deposited over a period of approximately 3 million years (Fig. 5.2C; Dreyer et al., 1999; Mochales et al., 2012), and is composed of six R-T cycles (Pyles et al., in review, Appendix A). Assuming each cycle was deposited roughly over the same amount of time, each R-T cycle would record ~ 0.5 m.y. and would be considered a fourth-order cycle (sensu Mitchum and van Wagoner, 1991). Therefore, Cycle 2 is defined as a fourth-order cycle. The average shoreline trajectory for the fourth-order transgressive unit of Cycle 2 is \( \frac{dy}{dx}; 50 \text{ m} / 2300 \text{ m} \) 0.02 or 1.14°, rising landward (southward). The positive shoreline trajectory relates to fluvial architecture.
trajectory indicates that the shoreline advanced landward during a relative rise in sea level and therefore the rate at which accommodation was created exceeded sediment input (A\(\geq\)S) (Pyles et al., *in review*, Appendix A). The lack of an incised valley and aggradational trajectory of the shoreline demonstrate that the rate of accommodation was high during the transgressive unit of Cycle 2 and is therefore classified as a high-accommodation sequence (Fig. 5.1).

To document changes in the overall transgressive unit of the Cycle 2, the transgressive unit was divided into three components: a lower, middle, and upper component (Components I, II, and III respectively; Fig. 5.5B). Each component is composed of multiple fifth-order R-T cycle legs (Fig. 5.5B) and is on average \(\sim\) 19 m thick. The trajectory for each component is a resultant vector documenting three distinct styles of stacking patterns of the fifth-order cycles. Component I contains fifth-order cycles that stack in a predominantly transgressive style, Component III contains fifth-order cycles that stack in a predominantly aggradational style, and Component II is transitional between the two. The red line in Fig. 5.5B documents the shoreline trajectory of the interpreted interface between the delta plain and open marine strata for each component.

The smallest-scale cycles are fifth-order R-T cycles and may record approximately 0.01-0.05 m.y. of deposition each (sensu Mitchum and van Wagoner, 1991). The fourth-order transgressive unit of Cycle 2 is subdivided into multiple fifth-order R-T cycles on the basis of shoreline trajectory. Fifth-order R-T cycles are similar in scale to parasequences, although the transgressive units of these cycles are well preserved. Each fifth-order R-T cycle contains a basinward (b) and landward (l) leg. There are a total of nine legs and therefore, four-and-a-half fifth-order R-T cycles (cycles A-E; Fig. 5.5B). Legs B\(_b\), C\(_b\), D\(_b\), and E\(_b\) have a basinward trajectory with an average thickness of 5.1 m, whereas, legs A\(_l\), B\(_l\), C\(_l\), D\(_l\), and E\(_l\) have a landward
trajectory with an average thickness of 7.3 m (Fig. 5.5B; inset table). Each fifth-order R-T cycle is on average ~ 12.4 m thick. The blue line in Fig. 5.5B documents the trajectory of the interpreted interface between the delta plain and open marine strata, which is interpreted as the shoreline trajectory for the fifth-order cycles (Fig. 5.5A).

5.5 Geology of the Study Area

This study quantitatively documents the stratigraphic architecture and net-sand content of the transgressive unit of Cycle 2 (fourth-order, R-T stratigraphic cycle) of the Escanilla Formation. A cross section is constructed by projecting all stratigraphic data from the field area (Fig. 5.6) onto a plane that is orientated normal to the mean paleocurrent direction (Fig. 5.7). The cross section is oriented so that the viewer is looking in the up-current direction. The cross section documents the location of stratigraphic columns; location, size, and shape of architectural elements and stories; rose diagrams of paleocurrent directions; and hierarchical boundaries.

The base of the transgressive interval is defined as a discrete stratigraphic surface that was correlated across the length of the cross section (A-A`; Fig. 5.5B) by walking the surface out in the field. At the distal shelf, this stratigraphic surface is underlain by mouth bar deposits of the regressive unit of Cycle 2 and overlain by silty mudstone sheets, tidal channels, and elongate mouth bars of the transgressive unit (Moss-Russell, 2009). At the proximal shelf, the stratigraphic surface is both underlain and overlain by fluvial strata. However, there is an increase in the proportion of splay stories within the transgressive unit of Cycle 2 (18%) relative to the regressive unit of Cycle 2 (4%; Fig. 5.7) as well as a change from predominantly low aspect ratio channel-belt elements within the regressive unit of Cycle 2 to predominantly intermediate to high aspect ratio channel-belt elements in the transgressive unit.
The upper surface of the transgressive interval is defined as a discrete stratigraphic surface that was correlated across the length of the cross section (A-A`; Fig. 5.5B) by walking the surface out in the field. At the distal shelf, this stratigraphic surface overlies a thin, grey marine mudstone that demarcates the maximum transgression of the shoreline and is overlain by proximal delta plain strata of the regressive unit of Cycle 3. At the proximal shelf, the stratigraphic surface is both underlain and overlain by fluvial strata. However, there is a decrease in the proportion of splay stories within the regressive unit of Cycle 3 (4%) relative to the transgressive unit of Cycle 2 (18%; Fig. 5.7) and an increase in modal grain size of sediment within channel-belt elements from the transgressive (medium-grained sandstone) to the regressive unit (coarse-grained sandstone).

The field area is a ~ 0.8 km wide outcrop (east to west) located below and east of the town of Mondot (Figs. 5.3, 5.6, 5.7). The map, photopanel, and cross section are used to document the following characteristics of the transgressive unit of Cycle 2 (Figs. 5.3, 5.6, 5.7 respectively). First, paleocurrent measurements indicate that sediment exited the outcrop belt (vector mean = 002°; circular variance = 0.5). Second, in order of increasing area, the transgressive unit contains (Fig. 5.7): crevasse channel with erosionally based fine-grained fill stories (<1%), crevasse channel with heterolithic fill stories (1%), erosionally based fine-grained fill stories (<1%), fine-grained fill associated with lateral accretion stories (<1%), lateral accreting stories (3%), downstream accreting stories (11%), splay stories (18%), and floodplain fine stories (67%). Third, in order of increasing area, the transgressive unit contains (Fig. 5.7): low aspect ratio channel-belt elements (1%), intermediate aspect ratio channel-belt elements (2%), high aspect ratio channel-belt elements (16%), unassociated splay elements (27%), and associate non-coeval splay elements (54%). Fourth, net-sand content, the ratio of the area of
sandstone to total area being evaluated, ranges from 0.61 in the western part of the transgressive unit to 0.47 in the eastern part of the transgressive unit with 44% of the sandstone located within channel-fill stories and 56% within splay and crevasse channel stories. Channel-belt elements have a modal grain size of medium-grained sandstone. (Fig. 5.7).

Individual channel-belt elements have large vertical and lateral offsets between one another resulting in low static connectivity (sensu, Funk et al., 2012). Channel-belt elements have an average width of 161 m, an average thickness of 4 m, and an average aspect ratio of 40 (Fig. 5.7). The overall sizes of channels increase from one to the next in an upward transect with an average thickness of ~1.8 m at the base to ~4.5 m at the top of the transgressive unit. Modal grain size varies from medium-grained sandstone in the lower channels to upper medium-grained sandstone in the upper channels of the transgressive unit.

5.6 Stratigraphic Architecture in Relation to Shoreline Trajectory

This study quantitatively documents how stratigraphic architecture in the transgressive unit of Cycle 2 relates to shoreline trajectory for fifth-order R-T cycles and for the three different components of the fourth-order transgressive unit of Cycle 2 (Fig. 5.8). Stratigraphic characteristics evaluated in this analysis are: (1) proportions of stories (2) size of channel-belt elements, (3) net-sand content, and (4) modal grain size within channel-belt elements.

5.6.1 Fifth-Order R-T Cycles

There are significant differences in the proportion of stories that were deposited when the shoreline trajectory was moving basinward to those deposited when the shoreline trajectory was moving landward at the fifth-order cycle scale. Figure 8B quantitatively documents the following
patterns. First, channel fill stories associated with basinward trajectories are higher in proportion than those associated with landward trajectories (38% vs. 27%). In contrast, floodplain fill stories are highest in proportion as the shoreline trajectory was moving landward (73% landward, 62% basinward). Second, channel-belt elements associated with basinward trajectories have an average thickness of ~ 4.1 m, an average width of ~ 175 m, and an average aspect ratio of ~ 43; whereas those associated with landward trajectories have an average thickness of ~ 3.1 m, an average width of ~ 104 m, and an average aspect ratio of ~ 34. Third, the net-sand content for strata associated with basinward trajectories is 0.41 and only 0.32 for landward trajectories.

When the shoreline trajectory moves basinward, 73% of the sandstone is located within channel fill stories and 27% within splay and crevasse channel stories. In contrast, when the shoreline trajectory moves landward, 56% of the sandstone is located within channel fill stories and 44% within splay and crevasse channel stories. Finally, the average modal grainsize within channel-belt elements, regardless of the shoreline trajectory, is medium-grained sandstone.

In summary, at the fifth-order cycle scales, fluvial strata associated with basinward trajectories are distinctive from those associated with landward trajectories in terms of proportions of stories, net-sand content, channel-belt element size, and modal grain size within channel-belt elements (Fig. 5.8B). Fluvial strata associated with basinward trajectories contain a higher percentage of channel fill stories, a higher net-sand content, and larger channels whereas fluvial strata associated with a landward trajectory contain a higher percentage of floodplain fine and splay stories, a lower net-sand content, and smaller channels. Individual channel belts are vertically and laterally isolated one from another irrespective of shoreline trajectory. Strata deposited within each leg consist of, from base to top: floodplain fine stories, distal splay stories, proximal splay stories, crevasse channel stories, and channel fill stories. This succession of strata...
is interpreted as avulsion-belt strata (e.g. Smith et al., 1989; Tornqvist, 1994; Jones and Schumm, 1999; Stouthamer and Berendsen, 2000; Tornqvist and Bridge, 2002). This interpretation is more fully developed below.

5.6.2 Fourth-Order Transgressive Unit of Cycle 2

The overall fourth-order transgressive unit of Cycle 2 is divided into three components based on patterns in the resultant vectors of fifth-order cycles (Figs. 5.5, 5.8): I, II, and III. The average shoreline trajectory (vector mean) for all three components is directed landward. The lowest (oldest) component, Component I, contains three fifth-order R-T cycle legs. The resultant vector of the shoreline trajectories is relatively flat (-0.01; Fig. 5.8A). The middle component, Component II, contains two fifth-order R-T cycle legs and the resultant vector of the shoreline trajectories is steeper than Component I (-0.04; Fig. 5.8A). The upper component, Component III, contains four fifth-order R-T cycle legs and the resultant vector of the shoreline trajectories is comparatively steep (-0.28; Fig. 5.8A).

The proportions of stories within each component are different (Fig. 5.8C). First, channel fill stories are lowest in proportion in Component I (20%), intermediate in Component II (22%), and highest in Component III (46%). Second, floodplain fine stories are lowest in proportion in Component III (47%), intermediate in Component I (53%), and highest in Component II (60%).

There is an increase in the number of stories within each channel-belt element and channel-belt element size from Component I to Component III. Component I contains single story channel-belt elements that have an average thickness of ~ 2.5 m, an average width of ~ 112 m, and an average aspect ratio of ~ 45. Component II contains both single and multistory channel-belt elements that have an average thickness of ~ 3.9 m, an average width of ~ 93 m, and
an average aspect ratio of ~ 24. Component III contains predominantly multistory channel-belt elements that have an average thickness of ~ 4.9 m, an average width of ~ 251 m, and an average aspect ratio of ~ 51.

The net-sand content increases upwards from Component I (0.35), to Component II (0.38) and finally Component III (0.46). In Component I, 43% of the sandstone is located within channel fill stories and 57% within splay and crevasse channel stories. In Component II, 55% of the sandstone is located within channel-fill stories and 45% within splay and crevasse channel stories. In Component III, 86% of the sandstone is located within channel fill stories and 14% within splay and crevasse channel stories. There is no measured difference in the modal grainsize within channel-belt elements deposited within Components I and II, that being lower medium-grained sandstone. However there is an increase in the modal grainsize within channel-belt elements deposited within Component III to medium-grained sandstone.

In summary, fluvial strata deposited from one component to the next within the transgressive component of the fourth-order Cycle 2 are distinctive in terms of shoreline trajectory, proportions of stories, channel-belt element size, net-sand content, and modal grain size within channel-belt elements (Fig. 5.8C). The shoreline trajectory is highly retrogradational in Component I, becomes more aggradational in Component II, and is highly aggradational, but still retrogradational in Component III (Fig. 5.8A). From base to top of the transgressive unit of Cycle 2 there is an overall increase in the percentage of channel-belt elements, multi-story channel-belt elements, channel-belt element size, net-sand content, and modal grain-size within channel-belt elements (Fig. 5.8C).
5.7 Autogenic Versus Allogenic Controls on Stratigraphic Architecture

The relative role of the different controls on shoreline trajectory and stratigraphic architecture has long been debated, those controls being: eustasy (Vail et al., 1977; Pitman, 1978; Jervey, 1988; Van Wagoner et al., 1988; Posamentier et al., 1988; Posamentier and Vail, 1988), tectonics (Watts, 1982; Summerhayes, 1986), and sediment input (Galloway, 1989; Thorne and Swift, 1991; Schlager, 1993). It is generally believed that during the development of the transgressive systems tract, any significant changes in shoreline trajectory such as episodic progradation during overall retrogradation are due to either: temporal changes in external or allogenic factors such as punctuated changes in relative sea level (Cross, 1988; Boyd et al., 1989; Gensous et al., 1993; Valasek, 1995) and varied rates of sediment input in relation to rate of sea-level rise (Van Wagoner, 1995); or internal or autogenic controls such as avulsion of the fluvial/deltaic system (Stouthamer and Berendsen, 2000; Muto and Steel, 2001; Stouthamer and Berendsen, 2007). This discussion attempts to address autogenic versus allogenic controls on shoreline trajectory and stratigraphic architecture of the transgressive unit of Cycle 2.

5.7.1 Autogenic Controls

It has been documented that avulsion of channel belt and delta lobes can be caused by changes in base level (Coleman, 1969; Bridge and Leeder, 1979; Mackey and Bridge, 1995; Berendsen and Stouthamer, 2000; Schumm et al., 2000; Stouthamer and Berendsen, 2000; Stouthamer and Berendsen, 2007). The probability of avulsion increases as a consequence of subsidence and decreases during uplift and incision. In general, a high rate of base level rise leads to a high avulsion frequency due to a rapidly decreasing longitudinal gradient (Tornqvist, 1994; Bridge, 2003; Stouthamer and Berendsen, 2007). Also, in an experimental study on
cohesive deltas, Martin et al. (2009), documented an increase in channel mobility (i.e. avulsion) and a reduction in channel residency time during a rise of base level.

Fluvial strata deposited within the transgressive unit of Cycle 2 contain avulsion belt deposits that scale in thickness and are stratigraphically equivalent to the fifth-order cycle legs. Fifth-order changes in stratigraphic architecture and shoreline trajectory within the transgressive unit of Cycle 2 are therefore interpreted to be due to the autogenic process of avulsion. As base level is rising during the transgressive unit of Cycle 2, channel avulsion occurs (Fig. 5.9A). When the channel avulses out of the area, the shoreline locally retrogrades (Fig. 5.9A). When the channel avulses back into the area, the shoreline locally progrades (Fig. 5.9A). One of the products of channel avulsion during transgression in this system is that laterally, fifth-order changes in shoreline trajectory will be out of phase (Fig. 5.9A).

5.7.2 Allogenic Controls

Jervey (1988) and Ross et al. (1995) concluded that the ratio between accommodation (A) and sediment input (S), which are generated by allogenic processes, is the primary control on the degree of progradational, aggradational, or retrogradational stacking patterns. The vertical component of the shoreline trajectory may be used as a proxy for accommodation at the time of deposition relative to the lateral component that can be used as a proxy for sediment input. Therefore, a progradational shoreline results when sediment input exceeds accommodation (A<S) while a retrogradational shoreline results when sediment input is less than accommodation (A>S). However, during retrogradation, if the rate of accommodation were to be held constant, an increase in the rate of sediment input, that is still less than accommodation would produce an increase in the steepness of the shoreline trajectory through time.
The fourth-order transgressive unit of Cycle 2 is overall retrogradational (Fig. 5.5b). There is an upward increase in steepness of the trajectory from Component I to Component III (Fig. 5.9B). This trajectory trend is associated with: (1) an increase in channel-belt element size, net-sand content, and grain size within channel-belt elements; (2) an upward change from single story channel-belt elements in the lower components to multi-story channel-belt elements in the upper component; (3) channels upwardly change from dispersed to clustered; and (4) there is an upward decrease in the number of splay stories (Figs. 5.8B, 5.9B). Collectively, these patterns are interpreted to reflect the overall upward pattern of increasing sediment input in relation to accommodation. Subsidence rate studies for the region document constant to increasing subsidence rates between the Lutetian, and Bartonian, followed by a gradual decrease in subsidence rates from the Late Eocene to the Miocene (Gimenez-Montsant and Salas, 1997; Fernandez et al., 2012; Huyghe et al., 2012). Sediment input has also been documented to increase during deposition of the Sobrarbe and Escanilla Formations (Mochales et al., 2012; Michael et al., 2014). Therefore this study interprets the allogenic process of increasing sediment input in relation to accommodation to be the main control on variations in fourth-order fluvial architecture.

5.8 Conclusions

Cycle 2 of the Escanilla Formation is an excellent outcrop analog for transgressive fluvial systems deposited in a high-accommodation setting. This article quantitatively documents for the first time, spatial patterns in stratigraphic architecture, net-sand content, thickness, and modal grain size of channel-belt elements within a transgressive unit of fluvial strata in order to evaluate: (1) differences between fluvial strata deposited when the shoreline trajectory was
moving basinward versus strata deposited when the shoreline trajectory was moving landward for fifth-order R-T cycles within the transgressive unit of the fourth-order R-T cycle, (2) differences in fluvial strata deposited from the base to the top of the fourth-order transgressive unit, and (3) the role of autogenic and allogenic processes on the stratigraphic architecture of fluvial deposits at both fourth-order and fifth-order scales.

Significant changes in fluvial architecture in relation to fifth-order cyclicity scale are: (1) fluvial strata associated with a basinward trajectory contain a higher percentage of channel-belt elements, a higher net-sand content, and larger channel-belt elements; and (2) fluvial strata associated with a landward trajectory contain a higher percentage of floodplain fine and splay stories, smaller channel-belt elements, and a lower net-sand content. These fifth-order cycle changes in fluvial architecture and changes in shoreline trajectory are interpreted to be due to the autogenic process of channel avulsion.

Significant changes in fluvial architecture in relation to fourth-order cycles are: (1) the fourth-order transgressive unit of Cycle 2 is overall retrogradational; (2) the upward increase in shoreline trajectory from being predominantly retrogradational to predominantly aggradational; and (3) the upward increase in channel-belt element size, the increase in net-sand content, and increase in grainsize within channel-belt elements. These fourth-order cycle changes in fluvial architecture and changes in shoreline trajectory are interpreted to be due to the allogenic process of increasing sediment input in relation to accommodation.

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Figure 5.1 Schematic diagram documenting a continuum of fluvial settings on the basis of accommodation. (A) Diagrams documenting systems tracts and controls on relative sea level. (B) Longitudinal cross sections of systems tracts. (C) Lateral cross sections of systems tracts. (D) Examples of published non-marine sequence stratigraphic studies for low-, intermediate- and high-accommodation settings.
Figure 5.2 (A) Paleogeographic map of the Tremp-Ainsa-Jaca basin (modified from Fernandez et al., 2004). (B) Generalized stratigraphy of the Ainsa-Jaca Basin (modified from Pickering and Bayliss, 2009). (C) Chronostratigraphic chart of the Hecho and Campodarbe Groups showing the different lithostratigraphic units located in the Ainsa Basin, eustatic curves (Haq et al., 1988), accommodation rates (Mochales et al., 2012), and timing of structural activity within the Ainsa Basin (Mutti et al., 1989; Puigdefabregas et al., 1992; Dreyer et al., 1999; Anastasio and Holl, 2001; Hoffman, 2009; Mochales et al., 2012). B=Boltaña Anticline; A=Anisclo Anticline; M=Mediano Anticline. (D) Interpreted photopanel of the Sobrarbe Formation taken from the town of Castellazo. The photopanel documents the boundaries between the six interpreted fourth-order regressive-transgressive stratigraphic cycles as defined by Moss-Russell (2009) and Pyles et al. (in review, Appendix A).
Figure 5.3 The geologic map on the right side of the figure is of the Ainsa Basin and documents the contacts between the different lithostratigraphic units, bounding structures (Boltaña Anticline, Mediano Anticline, and Anisclo Anticline) and the location of the Buil Syncline, which forms the axis of the basin (map modified from Hoffman, 2009 and Moody et al., 2012). Inset map is a geologic map of the field area documenting: the location of the contacts between the Guaso, Sobrarbe, and Escanilla Formations; the location of the boundaries between Cycle 1-3 as well as the regressive/transgressive boundary within Cycle 2; strike and dips of bedding surfaces; and paleocurrent measurements.
Figure 5.4 Illustrations of architectural components of the Escanilla Formation (modified from and Pyles, 2014). (A) Photographic examples of channel fill and floodplain fill stories identified in this study. (B) Channel-belt and floodplain-belt elements identified in this study. Channel-belt elements are subdivided by aspect ratio: high, intermediate, and low. Floodplain-belt elements are subdivided into associated non-coeval elements and unassociated splay elements. (C) Generic illustration of an archetype documented in this study.
Figure 5.5 (A) Trajectory of the shoreline is defined as $T_{sl} = dy_{sl}/dx_{sl}$. The trajectory of the shoreline can be used as a proxy for the ratio between accommodation and sediment input where $dy_{sl}$ is an indicator of accommodation and $dx_{sl}$ is an indicator of sediment input. Shoreline trajectory was defined as the interface between delta plain and open marine strata. (B) A longitudinal stratigraphic cross section of Cycle 2 of the Sobrarbe and Escanilla Formations documenting: the contact between the Guaso and Sobrarbe Formations, the lower and upper boundaries of Cycle 2, and the regressive/transgressive boundary within Cycle 2; lithology and internal correlations; and the shoreline trajectory at both fifth-order (blue line) and fourth-order (red line) cycles. The location of cross section is shown in Fig. 5.3 (A-A’). Cross section modified from Pyles et al. (in review, Appendix A). (C) Interpreted photopanel of the field area documenting: (1) the upper boundary of Cycle 2 and the regressive/transgressive boundary within Cycle 2; (2) the archetype boundaries within the transgressive unit of Cycle 2; and (3) the shoreline trajectory.
Figure 5.6 Uninterpreted (top) and interpreted (bottom) photopanels of the western half (A) and eastern half (B) of the field area documenting: the upper boundary of Cycle 2; the regressive/transgressive boundary within Cycle 2, and archetype, element, story, and stratal boundaries; the distribution of stories; and the location of stratigraphic columns. The location of the field area is shown in Fig. 3. The alpha-numeric classification used for channels (e.g. L19C1) are defined as follows: L, I, H = Low, Intermediate, or High aspect ratio channel-belt elements, respectively; 19 = sequential order of channel deposition; C1= leg of fifth-order cycle that the channel-belt element was deposited in.
Figure 5.7 Lateral stratigraphic cross section of the axial part of the transgressive unit of Cycle 2 within the field area that was constructed by projecting all stratigraphic data from the three field areas onto a plane that is orientated normal to the mean paleocurrent direction. The cross section is oriented so that the viewer is looking in the up-current direction. This is a north-facing outcrop and therefore, east is on the left hand side of the cross section, and west is on the right. The cross section documents: (1) the upper boundary of Cycle 2, (2) the regressive/transgressive boundary within Cycle 2, (3) the fifth-order R-T cycles shoreline trajectories (A1-E1) within the transgressive unit of Cycle 2, (4) the distribution of architectural elements; paleocurrent measurements; the location of stratigraphic columns, (5) small-scale fifth-order variations in shoreline trajectory as documented in Fig. 5.5B, (6) documentation of vertical changes in proportion of stories from one fifth-order cycle leg to the next, and (7) proportions of stories within the regressive and transgressive unit of Cycle 2 and the regressive unit of Cycle 3. The location of the cross section is shown in Fig. 5.3 (B-B`).
Figure 5.8 (A) Longitudinal stratigraphic cross section of Cycle 2 of the Sobrarbe Formation documenting the shoreline trajectory for fourth- and fifth-order cycles (see Fig. 5.5B for explanation of cross-section). (B) Quantitative comparison of stratigraphic architecture in terms of being deposited when the shoreline trajectory was moving basinward or landward at the fifth-order R-T cycle scale (legs A1-E5). (C) Quantitative comparison of vertical changes in stratigraphic architecture at the fourth-order R-T cycle scale (Component I, II, III). See Fig. 5.7 for explanation of colors and abbreviations of stories in the pie charts.
Figure 5.9 Summary diagram of the study area illustrating (A) the relationship between channel avulsion and the higher frequency, fifth-order cycle changes in shoreline trajectory; and (B) the relationship between the fourth-order shoreline trajectory within the transgressive component of Cycle 2 and fluvial stratigraphic architecture in relation to increasing sediment supply.
Table 5.1 Descriptions of the different stories identified in this study.

<table>
<thead>
<tr>
<th>Story</th>
<th>Average: Thickness; Width, Aspect Ratio</th>
<th>Bounding Surfaces</th>
<th>Depositional Strike View</th>
<th>Lithofacies Composition (in order of decreasing abundance)</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lateral accreting</td>
<td>4.3 m; 30 m; 7</td>
<td>conformable except when eroded into by younger strata</td>
<td>thickest in the axis and thins either abruptly or gradually toward its lateral margins</td>
<td>(1) structureless and planar cross-bedded medium- to coarse-grained sandstone with pebbles (79%); (2) trough cross-bedded fine- to medium-grained sandstone (18%); (3) pebble to cobble conglomerate with medium- to coarse-grained sandstone lenses (10%)</td>
<td>lateral accreting side attached bars</td>
</tr>
<tr>
<td>Downstream accreting</td>
<td>4 m; 160 m; 35</td>
<td>conformable except when eroded into by younger strata</td>
<td>thickest in the axis and thins either abruptly or gradually toward its lateral margins</td>
<td>(1) structureless and planar cross-bedded medium- to coarse-grained sandstone with pebbles (54%); (2) trough cross-bedded medium- to coarse-grained sandstone with pebbles (27%); (3) trough cross-bedded fine- to medium-grained sandstone (10%); (4) pebble to cobble conglomerate with medium- to coarse-grained sandstone lenses (9%)</td>
<td>downstream accreting mid-channel bars</td>
</tr>
<tr>
<td>Fine-grained fill associated with lateral accretion</td>
<td>2.1 m; 50 m; 24</td>
<td>convex upward, conformable</td>
<td>bowl shaped, asymmetrical, thickest in the axis and thins toward lateral margins; onlaps the margin adjacent to bars and is erosional on opposite margin</td>
<td>(1) rippled very fine- to fine-grained sandstone intercalated with siltstone (76%); (2) structureless, planar laminated, and rippled very fine- to fine-grained sandstone intercalated with siltstone (24%)</td>
<td>fine-grained strata that filled accommodation created by channel erosion</td>
</tr>
<tr>
<td>Erosionally based fine-grained fill</td>
<td>1.3 m; 58 m; 45</td>
<td>convex upward and highly erosional (0.5-1.8 m)</td>
<td>conformable</td>
<td>bowl shaped, symmetrical, thickest in the axis and thins toward lateral margins</td>
<td>(1) rippled very fine- to fine-grained sandstone intercalated with siltstone (79%); (2) gray siltstone to mudstone (21%)</td>
</tr>
<tr>
<td>Splay</td>
<td>3.4 m; 325 m; 96</td>
<td>conformable except when eroded into by younger strata</td>
<td>wedge shaped, thickest in axis, and thins gradually toward lateral margins</td>
<td>(1) structureless, planar laminated, and rippled very fine- to fine-grained sandstone intercalated with siltstone (86%); (2) bioturbated and/or rooted structureless to rippled very fine-grained sandstone intercalated with siltstone (54%)</td>
<td>crevasse splay deposits</td>
</tr>
<tr>
<td>Crevasse channel with heterolithic fill</td>
<td>1.5 m; 30 m; 20</td>
<td>erosional (0.2-3.2 m)</td>
<td>conformable to undulatory</td>
<td>bowl shaped, symmetrical, thickest in the axis and thins toward lateral margins</td>
<td>(1) structureless, planar laminated, and rippled very fine- to fine-grained sandstone intercalated with siltstone (100%)</td>
</tr>
<tr>
<td>Crevasse channel with erosionally based fine-grained fill</td>
<td>1.2 m; 15 m; 12</td>
<td>erosional (0.5-1.8 m)</td>
<td>conformable except when eroded into by younger strata</td>
<td>bowl shaped, symmetrical, thickest in the axis and thins toward lateral margins</td>
<td>(1) bioturbated and/or rooted structureless to rippled very fine-grained sandstone intercalated with siltstone (100%)</td>
</tr>
<tr>
<td>Floodplain fill components</td>
<td>4 m; &lt;10 km; unknown</td>
<td>conformable except when eroded into by younger strata</td>
<td>rectangular to wedge shaped, thickest in the axis and thins toward lateral margins</td>
<td>(1) varicolored siltstone to mudstone (94.7%); (2) varicolored mudstone to fine-grained sandstone with slickensides (5%); (3) carbonaceous mudstone (0.3%)</td>
<td>fine-grained strata deposited during waning flood-stage flow due to suspension fallout and tractive flow</td>
</tr>
</tbody>
</table>
Table 5.2 Descriptions of the different architectural elements identified in this study.

<table>
<thead>
<tr>
<th>Element</th>
<th>Average: Thickness; Width, Aspect Ratio</th>
<th>% of Element</th>
<th>Bounding Surfaces</th>
<th>Depositional Strike View</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Low aspect ratio channel-belt</td>
<td>4.3 m; 50 m; 12</td>
<td>9%</td>
<td>erosional</td>
<td>conformable</td>
<td>thickest in the axis and thins either abruptly or gradually toward its lateral margins</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>(2.5-4 m)</td>
<td>conformable except when eroded into by younger strata</td>
<td>low sinuosity fixed channel-belt</td>
</tr>
<tr>
<td>Intermediate aspect ratio</td>
<td>3.4 m; 77 m; 23</td>
<td>64%</td>
<td>erosional</td>
<td>conformable</td>
<td>thickest in the axis and thins either abruptly or gradually toward its lateral margins</td>
</tr>
<tr>
<td>Channel-belt</td>
<td></td>
<td></td>
<td>(0.5-5 m)</td>
<td>conformable except when eroded into by younger strata</td>
<td>braided channel-belt</td>
</tr>
<tr>
<td>High aspect ratio channel-belt</td>
<td>4.5 m; 250 m; 52</td>
<td>27%</td>
<td>erosional</td>
<td>conformable</td>
<td>thickest in the axis and thins either abruptly or gradually toward its lateral margins</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>(0.5-7 m)</td>
<td>conformable except when eroded into by younger strata</td>
<td>braided channel-belt</td>
</tr>
<tr>
<td>Associated non-coeval splay</td>
<td>3.6 m; 325 m; 90</td>
<td>58%</td>
<td>conformable</td>
<td>Conformable</td>
<td>thickest in the axis and thins gradually toward its lateral margins</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Conformable except when eroded into by younger strata</td>
<td>progradation of a crevasse splay complex into a floodplain and the full avulstion of a channel-belt element</td>
<td></td>
</tr>
<tr>
<td>Unassociated splay</td>
<td>3.4 m; 325 m; 96</td>
<td>42%</td>
<td>conformable</td>
<td>conformable</td>
<td>thickest in the axis and thins gradually toward its lateral margins</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Conformable</td>
<td>progradation of a crevasse splay complex into a floodplain and the failed avulstion of a channel-belt element</td>
<td></td>
</tr>
</tbody>
</table>

Floodplain-Belt Elements
6.1 Summary of Conclusions and Contributions

This dissertation is comprised of four outcrop studies of the Eocene Sobrarbe and Escanilla Formations, located in the Ainsa Basin, Spain, at three different scales of observation: third-, fourth-, and fifth-order cyclicity. The strengths of the studies summarized below lie in the high-resolution sequence stratigraphic framework developed from previous studies and the near-continuous longitudinal and lateral exposures of both the fluvial and deltaic deposits being studied. This dissertation advances our scientific knowledge about the deposition of fluvial systems deposited in high accommodation, high sediment supply settings.

6.2 Conclusions and Contributions of Chapter 2

Chapter 2 documents fourth-order scale structure-stratigraphic interactions of the Sobrarbe and coevally deposited Escanilla Formations. This study gives greater context to Chapters 3, 4 and 5 of the dissertation.

6.3 Conclusions and Contributions of Chapter 3

Chapter 3 examines four consecutive fourth-order regressive-transgressive cycles, each with different shelf-edge trajectories, in order to demonstrate the importance of subdividing populations on the basis of geological distinctions. The axis of the fluvial system contains higher percentages of channel-belt elements and splay stories, thicker splay beds, thicker channel-belt
elements, and a higher net-sand content than their counterparts deposited in the margin of the system. Fluvial strata associated with high A/S contain thicker and a higher percentage of floodplain-belt elements, smaller channel-belt elements, a higher net-sand content, and channel-belt elements are thinner in relation to their genetically related floodplain-belt element than the low A/S counterparts. Therefore, it is important to subdivide populations of data on the basis of geologic distinctions when building reservoir models.

### 6.4 Conclusions and Contributions of Chapter 4

Chapter 4 quantitatively documents for the first time vertical and lateral variations in stratigraphic architecture, net-sand content, grainsize, and static connectivity within a transgressive interval of a single fourth-order cycle. Key axis-to-margin patterns in the fluvial system are an increase in the proportion of channel-fill and splay stories, and channel-belt elements at the expense of floodplain fine stories, and an increase in net-sand content, channel-belt element size, modal grain size, and static connectivity from the margin to the axis of the system. Significant vertical changes are an upward increase in channel-belt element size, net-sand content, modal grainsize within channel-belt elements, and static connectivity. Results provided herein provide insight into high accommodation, transgressive fluvial deposits and can be used to reduce uncertainty in the interpretation of subsurface data, provide input to constrain rules-based forward stratigraphic models, and provide input to constrain reservoir models in transgressive fluvial systems.
6.5 Conclusions and Contributions of Chapter 5

Chapter 5 quantitatively documents for the first time, spatial patterns in stratigraphic architecture, net-sand content, thickness, and modal grain size of channel-belt elements in order to better understand the role of autogenic and allogenic processes on the stratigraphic architecture of transgressive fluvial deposits at both fourth-order and fifth-order scales.

At the smaller, fifth-order cycle scale, fluvial strata associated with basinward shoreline trajectories contain higher percentages of channel-belt elements and associated non-coeval splay elements, a higher net-sand content, larger channel-belt elements, and larger modal grain sizes within channel-belt elements than fluvial strata associated with landward shoreline trajectories. Fluvial strata associated with landward trajectories contain higher percentages of floodplain fine and splay stories and a lower modal grain size within channel-belt elements than fluvial strata associated with basinward shoreline trajectories. Changes in shoreline trajectory and fluvial architecture at the fifth-order cycle scale are interpreted to be due to the autogenic process of avulsion.

Within the larger fourth-order transgressive unit, there is an upward increase in the steepness of the shoreline trajectory which is associated with an increase in channel-belt element size, net-sand content, and modal grain size within channel-belt elements; an upward change from isolated single-story channel-belt elements to clustered multi-story channel-belt elements; and an upward decrease in the number of splay stories. These changes are interpreted to be due to the allogenic processes of increasing sediment input in relation to accommodation.
APPENDIX A

Pyles et al., in review—SUPPLEMENTAL ELECTRONIC MATERIAL

Appendix A includes a copy of a manuscript submitted to AAPG Bulletin. This manuscript documents the stratigraphic architecture of a single condensed-section bounded regressive-transgressive cycle (Cycle 2) of the Sobrarbe Formation. This study gives greater context to Chapters 2, 3, 4, and 5 of this dissertation.

| A-1_Pyles_etal_inreview.pdf | Manuscript submitted to AAPG Bulletin |
APPENDIX B

Stratigraphic Columns—SUPPLEMENTAL ELECTRONIC MATERIAL

Appendix B comprises a location map and stratigraphic column files of the Sobrarbe and Escanilla Formations.

<table>
<thead>
<tr>
<th>File Name</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>B-1_Location_map_of_stratigraphic_columns_Chapters_2-3.PDF</td>
<td>Map documenting the location of stratigraphic columns of Chapter 2 (stratigraphic columns 1-27) and Chapter 3 (stratigraphic columns 1-18).</td>
</tr>
<tr>
<td>B-2_Stratigraphic_columns_for_Chapters_2-3.PDF</td>
<td>Drafted stratigraphic columns of Chapter 2 (stratigraphic columns 1-27) and Chapter 3 (stratigraphic columns 1-18).</td>
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<tr>
<td>B-3_Location_map_of_stratigraphic_columns_Chapters_4-5.PDF</td>
<td>Map documenting the location of stratigraphic columns of Chapter 4 (stratigraphic columns 1-14) and Chapter 5 (stratigraphic columns 4-14).</td>
</tr>
<tr>
<td>B-4_Stratigraphic_columns_for_Chapters_4-5.PDF</td>
<td>Drafted stratigraphic columns of Chapter 4 (stratigraphic columns 1-14) and Chapter 5 (stratigraphic columns 4-14).</td>
</tr>
</tbody>
</table>
Appendix C comprises tabular data of the Sobrarbe and Escanilla Formations. These data include GPS points (UTM; Datum: European 1950 (Spain and Portugal)). These data include: strike and dip measurements of bedding planes, plunge and trend measurements of fault plains, paleocurrent measurements, and field notes.

| C-1_GPS_Waypoint_Data.PDF | GPS waypoint data for Sobrarbe and Escanilla Formations |
APPENDIX D

Permission Letters—SUPPLEMENTAL ELECTRONIC MATERIAL

Appendix D includes copies of letters from co-authors of the journal articles associated with Chapters 3, 4, and 5. In these letters, the coauthors grant permission for the use of the articles as chapters in this dissertation.

| D-1_Pyles_permission_letter.PDF | Permission letter from David Pyles. |