RICE UNIVERSITY

Fluvial Response to Base Level Change: A Case Study of the Brazos River, East Texas, U.S.A.

by

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ABSTRACT

Fluvial Response to Base Level Change: A Case Study of the Brazos Incised Valley, East Texas, U.S.A.

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A study of the Holocene avulsion history of the Brazos alluvial valley revealed that two processes, aggradation and valley tilting, were responsible for generating two styles of avulsion, avulsion-by-progradation and avulsion-by-annexation. As aggradation migrated inland, prograding avulsions tracked the locus of aggradation. Concurrently, a nodal avulsion site developed at 55 km inland, creating avulsions by-annexation. Geomorphic evidence suggests movement along a normal listric fault occurred in proximity to the nodal avulsion site.

Within the alluvium above the marine-Oxygen isotope Stage 2 onshore incised valley of the Brazos River, the pattern of stacked channels generated by avulsion was mapped to investigate the aggrading response of the Brazos River to sea level rise. The stacked channels within the valley decrease from eight, at 40 km from the coast, to four, at 65 km from the coast, which reflects the diminishing influence of eustacy inland. As aggradation decreased, while the avulsion frequency remained constant, the younger channels became more isolated, in contradiction to previous stacking models. Those models, however, neglected the influence of antecedent topography during aggradation.
Vertically, the eight stacked channels within the lower valley are organized into four stratigraphic units that are attributed to changes in the alluvial valley gradient during aggradation, as calculated from the position of backstepping (retrograding) offshore deltas (paleoshorelines) and their correlative (aggradating) onshore floodplain deposits.

The style of avulsion and the channel stacking pattern are both understood with respect to realizable subaerial accommodation. Previous subaerial accommodation models emphasized a proportional upward shift in an equilibrium profile during a sea level rise. Yet, an equilibrium profile must be anchored at both ends. The updip elevation of an alluvial valley is controlled by sediment yield and the cumulative aggradation from all earlier episodes of sea level rise, which should exert a limit on the downdip creation of subaerial accommodation. This study, therefore, quantified the differences in the long-term sediment yield of the Brazos and Trinity rivers of east Texas over past sea level cycles, and concludes that the lower sediment yield of the Trinity River has suppressed its equilibrium profile, thereby limiting the present creation of subaerial accommodation.
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CHAPTER 1

1.0 Introduction

The stratigraphic record results from climatic, eustatic, and tectonic change, and exists as the only prehistoric empirical record of these changes (Burke et al., 1990). The sedimentary fill within incised valleys represents a subset, but probably the most well preserved subset, of these global changes (Belknap and Kraft, 1981). Recent studies have sought to extract climatic and eustatic signals from the late Pleistocene and Holocene valley fill along continental margins. This extends to both the bay portion of underfilled incised valleys (Maddox, 2005; Simms et al, 2007), as well as the fluvial portion of overfilled incised valleys (Blum, 1993; Blum and Price, 1998; Tornqvist et al., 1998; Blum and Tornqvist, 2000).

The present study provides a more quantitative analysis on the issue of fluvial response to eustatic change than has previously been reported for an overfilled incised valley. Although a climatic signal is suspected of existing within the Brazos valley fill (Fraticelli, 2003), we feel the signal is more subtle than Blum (1993) documented along the adjacent lower Colorado incised valley and Sylvia and Galloway (2006) have indicated for the lower Brazos incised valley, and feel that the only way to extract this signal is through three-dimensional modeling of the valley fill through time that quantifies and accounts for the diachronous nature of valley aggradation as well the impact of antecedent topography on rates of aggradation. Therefore, this study attributes the changes in valley fill to eustatic change, or more correctly to a duration over which
eustatic change occurred, and leaves the extraction of any embedded climatic signal to a combination of greater chronostratigraphic control and future modeling of the dataset.

The thesis has been divided into three parts that deal with fluvial response to base level change at different spatial or temporal scales. The first two papers (Chapters 2 and 3) in this study concentrate on two aspects of fluvial response to base level change along the lower Brazos incised valley, those factors that control avulsions, and how avulsions control the channel stacking pattern. The third paper (Chapter 4) deals provides a quantitative comparison of the realizable subaerial accommodation generated during relative sea level rise for two fluvial systems, the Brazos and Trinity rivers, which exhibit significantly different sediment yields.
CHAPTER 2

The influence of valley aggradation and listric normal faulting on styles of river avulsion: a case study of the Brazos River, Texas, USA

2.0 Abstract

The Holocene avulsion history of the lower Brazos alluvial valley of east Texas, USA was studied using ten drill cores, 26 radiocarbon dates, aerial photos, and a digital elevation model. This study shows that two long-term processes, aggradation and localized valley tilting (along a normal listric fault), are responsible for generating two styles of avulsion. The first process precedes avulsion-by-progradation while the second process precedes avulsion-by-annexation.

As valley aggradation migrated updip over the last 7.5 ka, three regional backstepping avulsions occurred along the lower 140 km of the valley and each generated sizable deposits. A pattern emerges of landward stepping progradational avulsions tracking the locus of valley aggradation, and of valley aggradation migrating inland even after the rate of sea level rise diminishes. At the same time, several local nodal avulsions occurred between 50 and 55 km updip of the current highstand shoreline but generated no observable deposits. Geomorphic evidence indicates that, since the late Pleistocene, active movement along a previously undocumented normal listric fault has occurred at the location of the nodal avulsion. These two long-term processes do not operate mutually exclusive of each other to promote avulsions; rather they operate concurrently. Only aggradation promotes avulsions that affect floodplain alluviation, although the total volume of these deposits comprises a small portion of the valley fill.
2.1 Introduction

Avulsion occurs when a river diverts from its established course to follow a new course (see Slingerland and Smith, 2004 for a comprehensive overview). Once initiated, most likely during a major flood, the floodplain receives some or all of the river's sediment until the avulsion is finalized. The deposits are collectively termed a crevasse splay complex (Smith et al., 1989). Avulsion deposits are increasingly seen as an integral component for understanding the distribution of sand resources and the alluvial architecture within an aggrading valley. It has been suggested that these deposits comprise a significant portion of alluvial fill, up to 50% according to some case studies (Smith et al., 1989; Kraus, 1996; Kraus and Gwinn, 1997; Aslan and Blum, 1999; Slingerland and Smith, 2004). Yet, not all avulsions deliver large quantities of sediment to the floodplain. Therefore, to better understand and model alluvial architecture it is important to determine which processes promote extensive avulsion deposits and which do not. The motivation of this paper was to examine avulsions within the Brazos valley of east Texas during the latest transgression and compare the results to the avulsions of the adjacent Colorado River of Texas and the rivers of the Rhine-Meuse delta during the latest transgression.

Although short-term processes (e.g. extreme peak discharge, ice jams, vegetation, etc.) are responsible for initiating an avulsion (Smith et al., 1989; McCarthy, et al., 1992; Slingerland and Smith, 1998; Jones and Schumm, 1999; Tornqvist & Bridge, 2002), it is impossible to predict the occurrence of these processes. Instead, long-term processes
such as aggradation, differential compaction, fault movement and climate change, are responsible for promoting the likelihood of an avulsion and provide the key to potentially predicting the locations of future avulsions. These long-term processes are also responsible for controlling the style of avulsion (Tornqvist, 1994; Aslan and Blum, 1999; Morozovo and Smith, 1999; Peakall, et al., 2000; Gibling et al., 2005; Rajchl and Uličný, 2005; Maynard, 2006).

To highlight the spectrum of avulsion deposits, they have been grouped into three styles depending upon the amount of sediment delivered to the floodplain before a new channel is established. As defined by Slingerland and Smith (2004), these include avulsion-by-progradation, -annexation, or -incision. Avulsion-by-progradation delivers the greatest amount of sediment to the floodplain, avulsion-by-incision delivers less sediment, and avulsion-by-annexation delivers the least amount of sediment (Smith et al., 1989; Aslan and Blum, 1999; Morozovo and Smith, 1999). This study determines which long-term processes relate to which styles of avulsion for the Brazos River. Avulsion-by-progradation and -annexation occur in the study area, while the third (and intermediate) style of avulsion is not pertinent to this study.

This paper provides a brief background on recent related studies. Then we use our results from the Brazos valley to answer a series of questions posed by these previous studies. Does a decrease in the rate of sea level rise lead to a change in a river’s style of avulsion? Do different long-term processes have to compete with each other to promote avulsion or can these processes promote concurrent avulsions? And finally, is there
evidence for avulsion deposits comprising a significant portion of the alluvial fill within the Brazos valley?

2.2 Background

Recent studies (Leeder, 1993; Aslan and Blum, 1999; Morozovo and Smith, 1999; Stouthamer and Berendsen, 2000; Mack and Madoff, 2005, among others) have correlated the long-term processes that promote avulsion to the styles of avulsion produced. The studies conclude that long-term processes, such as varying aggradation rates or changing climatic conditions, directly impact alluvial architecture.

By using the example of the Colorado River of the Texas Gulf coast, Aslan and Blum (1999) showed that avulsion-by-progradation was common during rapid sea level rise for this high sediment yield fluvial system. In this instance, the accommodation space created during the latest sea level rise led to rapid aggradation (and numerous avulsions) of the lower Colorado valley. Based on a study of the Saskatchewan fluvial system in Manitoba (Canada), Morozovo and Smith (1999) argued that avulsion-by-progradation occurred during cooler and wetter climatic conditions. Under those conditions, floodplain lakes rose and wetlands grew, creating greater accommodation space for avulsion deposits. As the water table rose, an avulsing river was required to spill out into lakes or wetlands for longer periods, thus generating a larger crevasse splay complex before a new channel could form downstream.
At the other extreme, fluvial systems show evidence of avulsion-by-annexation either when the creation of accommodation space is minimal (e.g. at the beginning and towards the end of sea level rise), or when limited sediment supply cannot fill the available accommodation space. An example of the former can be found in the Colorado valley and examples of the latter occur along the Nueces and Trinity valleys of the Texas Gulf coast (Aslan and Blum, 1999). Avulsion-by-annexation was also favored by the Saskatchewan River during periods of warmer and drier climate, when lake levels were lower and wetlands were reduced in size (Morozovo and Smith, 1999).

In addition to sea level rise and climate change, another long-term process that promotes avulsion is movement along faults, which may be oriented parallel or perpendicular to the river valley. The movement of normal faults is often responsible for creating valleys and is therefore parallel to the valley. As a result, these faults may cause avulsion if lateral tilting of the valley floor is rapid enough. Alternatively, movement of normal listric faults generally cuts across a valley and may cause avulsion by changing the longitudinal gradient downdip of a rollover anticline (Mike, 1975; Bridge and Leeder, 1979; Ouchi, 1985; Alexander and Leeder, 1987; Singh et al., 1993; Mackey and Bridge, 1995; Gawthorpe and Leeder, 2000; Peakall et al., 2000; Mack and Madoff, 2005; Maynard, 2006). Since faults maintain the same location for extended periods, the affect on fluvial systems is concentrated and tends to induce nodal avulsions (Southamer and Berendsen, 2000; Maynard, 2006). Once a site of nodal avulsions has been established, the fluvial system can avulse by annexation. This reduces or eliminates the time spent discharging water and sediment onto the floodplain (Slingerland and Smith, 2004).
Numerous zones of normal listric faults run parallel to the Texas coast, the oldest of which date to the early Cenozoic (Barton et al., 1933; Waters et al, 1955; Bornhauser, 1958; Owen and Jackson, 1981). These faults are driven by movement of unconsolidated sediment along deep-seated décollement surfaces at depths of one to four km (Saribudak and Nieuwenhuise, 2006). This movement introduces a gradual change in the topography, often without any visible scarp (Maynard, 2006; Saribudak and Nieuwenhuise, 2006).

While avulsions have been documented in the vicinity of normal listric faults (Maynard, 2006), the extent of their associated deposits have not been documented. Maynard (2006) described the subsurface growth of a Miocene aged rollover anticline in south Texas that was associated with a normal listric fault. The evolution of the rollover anticline increased the downdip valley gradient, promoting avulsion. He used time slices from three-dimensional seismic data to illustrate the evolution of a single channel into a drainage network of several channels as nodal avulsions occurred. The results of Maynard’s (2006) case study documented the evolution of a drainage network as fault movement occurred, however, the author did not discuss the effect these avulsions had upon alluvial architecture.

Presumably, the repeated avulsions implied by the nodal network of channels would favor avulsion-by-annexation once the channel network was in place. This style of avulsion would deliver minimal amounts of sediment to the floodplain. Aslan and Blum (1999) support this hypothesis by suggesting that avulsion-by-annexation tends to occur
where rates of aggradation are low (such as downdip of a rollover anticline where accommodation space is negative). Therefore, while nodal avulsions associated with normal listric faults help document the evolution of a fluvial system, their prominent visibility in seismic data may actually over-emphasize the importance of fault-controlled avulsions upon alluvial architecture.

Finally, if different long-term processes are responsible for producing different styles of avulsion, then it is important to confirm if these processes are in competition with each other to promote avulsions. Clearly, competition between processes would affect alluvial architecture. A study by Stouthamer and Berendsen (2000) determined that for the Rhine-Meuse fluvial/deltaic system, the occurrence of avulsions driven by either sea level rise or structural control appeared to be mutually exclusive. These authors observed that sea level rise from the early to middle Holocene caused the sites of avulsion to shift progressively landward. During this time interval, the majority of avulsions were random, not nodal. They related these observations to higher rates of landward aggradation, similar to observations along the Colorado River valley by Aslan and Blum (1999). As the rate of sea level rise decreased, the rate of aggradation and the rate at which avulsion sites stepped landward also decreased. Eventually, the landward migration of avulsion sites ceased and the sites where avulsions occurred became concentrated along known fault zones. At this point in the sea level cycle, the majority of Rhine-Meuse avulsions were nodal.
The Brazos River is similar to both the Rhine-Meuse and the Colorado fluvial systems. It occurs along a passive continental margin, has a high sediment supply and a low valley gradient, and it has experienced rapid rates of valley aggradation throughout the Holocene. However, unlike the Rhine-Meuse fluvial system, which has maintained several distributaries throughout the Holocene (Tornqvist, 1994; Stouthamer and Berendsen, 2000; 2001), the Brazos fluvial system is simpler. The Brazos River has maintained only one active channel at any one time since the middle Holocene, and possibly the early Holocene (Snow, 1998; Fraticelli, 2003, Abdulah et al., 2004). Both the Brazos and the Colorado rivers have maintained a high sediment yield for the duration of the latest sea level rise. This explains why neither fluvial system has been flooded to form a bay (Aslan and Blum, 1999; Abbott, 2001). Hence, the Brazos River should respond in a similar fashion to the Colorado River during sea level rise and valley aggradation.

In addition, active movement along normal listric faults occurs in close proximity to, but outside of, the Brazos valley, both to the northeast (Saribudak and Nieuwenhuiise, 2006) and to the southwest (Schumm et al., 1984; 2000). For this reason, and because several geomorphic anomalies exist along the lower Brazos valley, the presence of fault control on valley morphology is suspected. Therefore, the potential exists for two different processes operating within the Brazos valley to promote avulsions.
2.3 Study Area and River Characteristics

The study area covers the lower Brazos River valley from the coast to 160 km inland. The width of the valley averages 10-12 km, except along the coastal plain where the width increases to ~40 km (Fig. 2-1).

The headwaters of the Brazos drainage basin start in eastern New Mexico. From here the river flows across north central Texas before emptying into the Gulf of Mexico. Presently, the Brazos River has the second highest sediment yield of all Texas rivers, and the highest yield in terms of suspended load (Milliman and Syvitski, 1992). Paine and Morton (1989) determined that on average the Brazos yields 9.9 hectometers$^3$/yr of suspended load and 0.9 hectometers$^3$/yr of bed load to the Gulf, making it a mixed load system according to Schumm (1977).

The time frame studied for this study extends from 12 ka to the present. For most of this interval of time, a high-resolution sea level curve exists, based on dating basal peats of the Mississippi delta (Tornqvist et al., 2004), and corals and peats from the western Atlantic (Toscano and McIntyre, 2003).

2.4 Methods

Different datasets were used to document the presence of avulsions, to examine long-term processes known to promote avulsion, and to determine how closely avulsions corresponded to these processes. Aerial photographs and lithologic data from cores were examined for evidence of avulsion deposits or a channel network associated with an
Figure 2-1: Location map of the study area for the lower Brazos valley. Also shown are the sites from which two cores (BV-04-03, BV-04-06) were collected for detailed sedimentological analysis, and the locations of two cross sections of cores that were used to examine the rates of valley aggradation.
avulsion node. The three datasets used to establish the long-term processes that drove avulsion were radiocarbon dates, aerial photographs, and digital elevation models (DEM). The combined data were used to quantify rates of aggradation across the valley associated with sea level rise as well as to identify geomorphic evidence indicative of structural control.

Aerial photographs of 1 by 1 m resolution, at a scale of 1:15,000, were used to identify the location of two avulsions (Figs. 2-2 & 2-3B) and to study the style of avulsion (avulsion-by-progradation or avulsion-by-annexation). Lithologic data from core BV-04-06 were used to identify a sequence of deposits associated with avulsion-by-progradation (Fig. 2-1).

Radiocarbon dates from two cores (BV-04-03 and BV-04-06) from the lower Brazos incised valley were acquired so vertical rates of valley aggradation could be calculated (Fig. 2-1). The shorter core (BV-04-03) was collected 90 km inland using a truck-mounted push-coring system. The longer core (BV-04-06) was collected 65 km inland using a combination of a truck-mounted and a tractor-mounted push-coring system.

It was important to use floodplain samples when calculating the vertical rates of valley aggradation (Allen, 1965). Dates obtained from cores BV-04-03 and BV-04-06 came from floodplain terrestrial gastropod shells. These samples were analyzed at the Woods Hole Oceanographic Institute’s NOSAMS laboratory. Previous radiocarbon dates
Figure 2-2: Aerial photographs showing the crevasse splay complex associated with the 4.0 ka Brazos River avulsion (see Avulsion #2 in Fig. 1 for location). A yellow line is used to denote the perimeter of the avulsion deposits.
Figure 2-3: Aerial photographs of A) a modern Brazos River crevasse splay (north of Avulsion #2 in Fig. 1) for comparison with B) the crevasse splay complex associated with the 1.5 ka Brazos River avulsion (see Avulsion #3 in Fig. 1 for location). A yellow line is used to denote the perimeter of the crevasse splay and the crevasse splay complex respectively.
(Bernard et al., 1970; Abbott, 2001; Sylvia and Galloway, 2006) from other cores in the valley were also incorporated into this study. All radiocarbon ages were corrected to calendar ages using Calib 5.0 software, which accounts for atmospheric variations in radiocarbon over time by comparing sample dates against a calibrated dataset (Calib 5.0).

To estimate the two-dimensional rates of valley aggradation, isochron crosssections were constructed from the radiocarbon dates. The depths and ages of the radiocarbon dates were used to interpolate isochron lines and to construct one strike- and one dip-oriented isochron section. The strike-oriented isochron section was constructed from five cores and 14 radiocarbon dates that were previously collected 40 km updip of the coast (Abbott, 2001). An aggradation curve was then derived from this cross section. The dip-oriented isochron section was constructed from the two cores and six radiocarbon dates we collected, as well as four cores and ten radiocarbon dates from previous studies (Bernard et al., 1970; Abbott, 2001; Sylvia and Galloway, 2006). The dip line isochron section extends from 30 km to 100 km inland.

For a numerical correlation between aggradation and sea level rise, the two-dimensional interpolated isochron data were used to calculate discrete volumes of valley fill. The volumes were calculated between paleo-floodplain surfaces from 8 ka to the present. The number of stacked isochron lines limited the length of the valley segment chosen to between 40 and 90 km updip of the coast. The 2-D isochrons were interpolated across the valley and subtracted from one another using GIS software (ArcMap and Erdas Imagine) to calculate the volumes.
Finally, two interrelated events, sea level rise and delta backstepping, along with a transect of the continental shelf bathymetry, were used as proxies to determine why variations exist in the valley aggradation curve (from the strike-oriented isochron section). For this study, a western Atlantic sea-level curve, derived for the Holocene from peat and coral data (Toscano and MacIntyre, 2003), and a regional Gulf of Mexico sea-level curve, based on Holocene basal peat ages (Tornqvist et al., 2004), were used. However, some discrepancy exists between these sea level curves and our valley aggradation curve. Thus, a second proxy, the timing and proximity of the retreating shoreline during delta backstepping as sea level rose, was correlated with the aggradation curve. The timing and positions of the retreating shoreline help refine the relationship between the valley aggradation curve and sea level rise (Snow, 1998; Abdullah et al., 2004), but only when integrated with a detailed map of continental shelf bathymetry.

Aerial photographs and DEMs were used to investigate suspected structural control on the valley morphology. The geomorphic analysis was applied at a location along the Brazos valley about 25 to 55 km inland, where previous structural control had not been documented (Salvador, 1991).

2.5 Results

Two long-term processes promoted the likelihood of avulsions during the Holocene within the lower Brazos valley. Each process led to a different style of avulsion, denoted by the presence or absence of avulsion deposits. The first process,
aggradation, generated substantial avulsion deposits (Figs. 2-2 & 2-3B). The second process, movement along a normal listric fault, generated no discernable avulsion deposits (Fig. 2-4).

*Aggradation as a precursor to avulsion-by-progradation*

The record of valley filling comes from two isochron cross-sections that document the spatial and temporal variations in floodplain aggradation (Fig. 2-1). The first isochron cross-section is oriented parallel to strike and located 40 km updip of the current highstand shoreline (Fig. 2-5). The rates of aggradation vary between 2 and 4 m/kyr, with upper and lower limits of 0.3 to ~7 m/kyr. The majority of floodplain aggradation occurred between 12 and 6 ka. The long-term rates of aggradation at the center and east side of the valley are complicated by two local anomalies of up to 7 m/kyr (Fig. 2-5). We believe these anomalies represent avulsion deposits, based on lithology from Abbott’s (2001) cores. For this reason, more representative long-term rates of aggradation come from the west side of the valley. From 12 to 9 ka the rates of aggradation on the west side of the valley were about 2 m/kyr. From 9 to 6 ka the rate almost doubled, to about 4 m/kyr. After 6 ka, the rates of aggradation decreased to almost 10% of the former value (0.3-0.4 m/kyr) (Fig. 2-5). The significance of this decrease after 6 ka is only understood when placed in the context of the second isochron cross-section.

The second isochron cross-section through the axis of the valley extends from 30 km to 100 km inland (Fig. 2-6). The rates of floodplain aggradation derived from this
Figure 2-4: USGS aerial photograph showing the location of the Avulsion Node, from which Dry Bayou, Middle Bayou, and Long Pond all emanate. The modern Brazos River and Jones Creek deviate to the west, following the trend of the normal listric fault, before flowing south again. Oyster Creek is unaffected by the fault, although Big Slough is affected by its trend and flows around a remnant of the Pleistocene uplands called Baileys Prairie.
Figure 2-5: Strike line isochron cross-section (A-A') of the Brazos valley at 40 km inland (data modified from Abbott, 2001). All radiocarbon ages (yrs) are corrected. This strike line crosses the dip line shown in Figure 2-6 at Core #2. For profile location see Figure 2-1. The deposits of Avulsion #1 are based on the lithology of Abbott core #2 and the expansion of the interpolated isochrons between 7 and 8 ka.
Figure 2-6: Dip line isochron cross-section (B-B') of the Brazos valley between 30 and 100 km inland. All radiocarbon ages are corrected. Cores BV-04-03 and BV-04-06 were collected for this study. Other data are from Bernard et al. (1970), Abbott (2001), and Sylvia and Galloway (2006). This dip line crosses the strike line shown in Figure 2-5 at the location of Abbott Core #2. For profile location see Figure 2-1.
section vary between 2 and 4 m/kyr, but determination of the upper and lower limits was not possible because of limited radiocarbon control. The isochrons between 9 and 6 ka are better constrained than those after 6 ka. Closer to the coast, the majority of fluvial aggradation occurred between 12 and 6 ka (Fig. 2-5). However, with time the rate of valley aggradation dramatically decreased near the coast (Fig. 2-6), while it increased further inland. This updip increase is substantiated by following the 6 ka and older isochrons updip in figure 2-5, where they are observed at progressively greater depths. The qualitative observations from these two cross-sections were then quantified by interpolating the isochrons as surfaces across a segment of the valley.

In order to gauge how the locus of aggradation migrated inland, floodplain surfaces were interpolated at 8 ka, 7 ka, and 6 ka along a segment of the valley between 40 and 90 km (Fig. 2-7). The volume of sediment sequestered on the floodplain between 8 and 7 ka was 2.1 km³, between 7 and 6 ka was 3.0 km³, and between 6 and the present was 5.0 km³. From 6 ka onward, the volume of sediment sequestered could not be subdivided in 1000 yr intervals because of poor chronostratigraphic control, but on average the rate of sequestration was 0.8 km³/kyr. These volume calculations indicate aggradation initially increased (8 ka to 6 ka) and then decreased (6 ka to present) as the locus of aggradation migrated inland, past this segment of the valley.

Broadly speaking, the mechanism that drives valley aggradation is sea level rise, as aggradation of high sediment yield fluvial systems tends to track sea level rise (Aslan and Blum, 1999; Southamer and Berendsen, 2000, Taha and Anderson, in review-B).
Figure 2-7: Two-dimensional and three-dimensional (from Taha et al., in review-C) perspectives of paleo-floodplain surfaces identified at A) 8 ka, B) 7 ka, and C) 6 ka. This segment of the valley extends from 40 km inland to 90 km inland. The data points used for the interpolation are indicated by white dots. See figure 1 for location.
Yet, the changes shown in the aggradation curve (Fig. 2-8) derived from figure 2-5 are out of phase with the rate of sea level rise. Instead, changes in rates of aggradation correlate with the timing of previously established delta backstepping events on the continental shelf (Fig. 2-8).

Prior work on the continental shelf offshore of the study area has shown that the paleo-shoreline retreated in discrete delta backstepping events on three occasions in the Holocene (Abdulah et al., 2004). The first and second backstepping events formed new shorelines located 60 and 40 km offshore of the present coast. The timing of the first two events is inferred from correlation with ~11.5 ka and ~9.5 ka backstepping Colorado deltas just to the east (Snow, 1998). The chronology for each of these events is constrained by only one radiocarbon date, so the precise timing is constrained is uncertain. At some point around or before 6 ka, the Brazos delta and shoreline backstepped to its present position (Bernard et al., 1970).

Despite the limited age constraints on offshore paleo-shorelines, a close match is seen between shoreline retreat and the fluvial aggradation curve (Fig. 2-8). The timing of the ~11.5 ka backstepping event closely corresponds to the initiation of aggradation (12 ka) at 40 km inland (Fig. 2-5). The timing of the ~9.5 ka backstepping event closely corresponds to the time when the rate of aggradation (9 ka) almost doubled 40 km inland (Fig. 2-5). The timing of the ~6 ka backstepping event closely corresponds to a
Figure 2-8: Aggradation curve for the Brazos valley at 40 km inland (see Core #1, Fig. 2-5) plotted with A) Tornqvist et al.'s (2004) sea level curve (0-8 ka), and B) Toscano and McIntyre’s (2003) western Atlantic sea level curve (0-11 ka). In both graphs the timing of the three Brazos backstepping events is indicated with arrows. Note how, even as the rate of sea level rise diminishes between 9 and 6 ka, the rate of aggradation increases, closely matching the timing of the delta backstepping event at 9.5 ka.
substantial decrease in the rate of aggradation (6 ka) at 40 km inland (Fig. 2-5). How does this relate to rates of sea level rise?

Surprisingly, even though the first rapid rate of sea level rise since the Last Glacial Maximum occurred between about 14.6 and 14.1 ka (Meltwater Pulse 1A, Weaver et al., 2003), it did not result in a corresponding backstep of the Brazos delta from previous observations of the Brazos delta history on the continental shelf (Abdulah et al., 2004). Instead, the first delta backstepping event occurred at 11.5 ka, during the second meltwater pulse event (1B), which spanned about 1000 years and is centered around 11.3 ka (Bard et al., 1996). After Meltwater Pulse 1B, the rate of sea level began to decrease, yet the 9.5 ka backstepping event occurred during this time interval. Finally, as sea level rise slowed around 6 ka, the 6 ka backstepping event occurred. So the deltas backstepped during high, moderate, and low rates of sea level rise.

Two alternatives explanations exist for what controlled the second and third delta backstepping events. The first explanation involves variations in shelf bathymetry. Theoretically, if the rate of sea level rise remains constant but the advancing shoreline encounters a decrease in gradient, then the shoreline will backstep more quickly. Conceivably, even as the rate of sea level rise decreases, shoreline retreat may indeed increase if the decrease in shelf gradient is great enough. An example of along-strike variability in shoreline retreat as a result of variations in shelf bathymetry has been demonstrated for the adjacent east Texas shelf, just north of our study area (Rodriguez et al., 2004). We believe the entire Brazos delta may have backstepped, independent of the
rate of sea level rise, as major changes in shelf gradient were encountered during the overall sea level rise. The evidence for this is shown in a transect across the continental shelf that passes over the two paleo-Brazos deltas. The transect shows a decrease in the bathymetry at ~ -34 and ~ -26 m (Fig. 2-9). The first significant decrease in gradient occurs several meters above the 11.5 ka shoreline and may have allowed the shoreline to backstep to its position at 9.5 ka (Snow, 1998; Abdulah, et al., 2004). The second decrease in gradient is more subtle and occurs just 4 or 5 meters above the 9.5 ka shoreline. This change in shelf gradient probably acted as the threshold that sea level rise had to overcome to backstep to its current position at around 6 ka (Bernard, et al., 1970). So, the second and third delta backstepping events probably occurred when a threshold in gradient was reached during flooding of the continental shelf, even though the rate of sea level rise was decreasing.

The second explanation for why deltas backstepped as the rate of sea level rise decreased involves antecedent topography of the valley. For underfilled incised valleys such as the Trinity and Nueces valleys (Simms et al., in press-A), bayhead deltas have been shown to have backstepped as terraces were flooded and the bays expanded (Rodriguez et al., 2005; Simms, 2005). The same concept should apply to higher sediment yield, overfilled incised valleys (Simms et al, in press-A). If aggradation fills the valley above a terrace and widens the floodplain towards the end of sea level rise, then antecedent topography passively controls the timing when a large volume of sediment is being sequestered onshore. This storage would reduce the sediment being delivered to the delta and potentially initiate delta backstepping.
Figure 2-9: Bathymetric transect offshore of the modern Brazos delta shoreline. The transect cuts perpendicular to the -26 m and -44 to -46 m isobaths, where the 9.5 ka and 11.5 ka shorelines existed, respectively, according to Snow (1998) and Abdulah et al. (2004). Note the decrease in gradient at -34 m and -20 m, which may have contributed to rapid shoreline retreat and delta backstepping, during sea level rise, at 9.5 ka and ~6 ka.
It is difficult to directly constrain the link between the antecedent topography of valleys and delta backstepping. The difficulty arises because trangressive ravinement has removed portions of the offshore valley, and a much larger radiocarbon database would be needed to establish the boundary between the onshore valley antecedent topography and valley fill. Nonetheless, indirect evidence for antecedent topography exerting an influence on sediment storage does exist. It was previously determined that the greatest horizontal isolation of Brazos channels within the lower valley occurred from the mid to late Holocene (7.5 ka to 1.5 ka) (Taha and Anderson, in review-B). This isolation could only occur if the valley widened, inline with a terraced valley being overfilled. The initial timing of valley widening (at 7.5 ka) came shortly before the last delta backstepping event at 6 ka. By extension, whatever controls aggradation also promotes river avulsion.

Avulsion-by-progradation

The long-term process of valley aggradation was responsible for conditioning the paleo-Brazos River for avulsion. In two of the three avulsions (Avulsions #1 and #2) (Fig. 2-1) attributed to aggradation, a positive correlation exists between the location and timing of the avulsion and the locus of valley aggradation. Although the third and youngest avulsion (~1.5 ka, Avulsion #3) (Fig. 2-1) was located 140 km inland, beyond the limit of the strike-oriented isochron (Fig. 2-6), it is highly probable that the locus had shifted this far inland by 1.5 ka based on our quantification of the landward movement of the locus of valley aggradation in figure 2-7.
The oldest of the avulsions (Avulsion #1) that was likely to have been promoted by aggradation has since been buried (Fig. 2-5), while the deposits of the two youngest avulsions (Avulsions #2 and #3) are situated close enough to the surface to be observed in aerial photographs (Figs. 2-2 & 2-3B). Avulsion #1 occurred 40 km updip of the current highstand shoreline. The extent of the deposits is seen in the strike-oriented isochron section (Fig. 2-5) and the lithology of the deposits is illustrated in Core #2 from Abbott (2001).

The reconstruction of the paleo-floodplain at 8 ka indicates that a topographic depression (in Fig. 2-5 between cores #1 and #2) almost 5 km wide existed in the center of the valley at that time. By 7 ka this depression was actually a topographic high. Core #2 (Fig. 2-5) constrained the anomalously high rate of aggradation at almost 7 m/kyr in the center of the avulsion deposits. In contrast, the average rate of aggradation across the valley between 9 ka and 6 ka was 3 to 4 m/kyr. In Core #2 a sharp break in lithology occurs about 1 m above a 7,745 year radiocarbon date obtained from floodplain deposits (Fig. 2-5). Immediately above the contact, the alluvial fill is very sandy, and remains sandy in the overlying 4 m where the section eventually grades back into floodplain clay. A second radiocarbon date from the overlying floodplain clay constrains the age of the 4 m sandy unit to be no less than 7,285 years old. Directly above this second contact (from sand back to clay) is a prominent soil horizon (Abbott, 2001).

The combined data indicate that a major avulsion occurred between 7,745 and 7,285 years ago (tentatively 7.5 ka) and it occurred by progradation, although its areal
extent could not be determined. The dip-oriented isochron section indicates that the locus of valley aggradation between 8 and 6 ka was located around 40 km updip from the present coast (Fig. 2-6). Therefore, long-term aggradation promoted avulsion. The deposits from this avulsion are of comparable thickness (~4m) to the avulsion-by-progradation deposits documented along the Colorado valley during rapid valley aggradation (Aslan and Blum, 1999). The isochron section did not allow us to document the areal extent of the oldest avulsion deposits.

Aerial photographs reveal the extent of the two younger crevasse splay complexes and suggest avulsion-by-progradation (Figs. 2-2 and 2-3B). Each crevasse splay complex is composed of coalescing crevasse splays that prograde onto the floodplain adjacent to the location of the breakout. Each crevasse splay complex was identified by the preservation of an anastomosing channel network that expands downstream. This channel network corresponds to Smith et al.’s (1989) classification of Stage II and Stage III splays, based on a case study of modern avulsion deposits along the Saskatchewan River. As their work highlights, a crevasse splay complex (multiple lobe deposits) is significantly larger than, and should not be confused with, a crevasse splay (a single lobe deposit caused by a breach of the natural levee during a flood that does not evolve into a true avulsion).

Avulsion #2 occurred 65 km to 70 km updip of the coast, where a prominent crevasse splay complex exists (Fig. 2-2). The outlines of former streams are preserved on the floodplain to the west of Oyster Creek and display the characteristic bifurcating and
coalescing pattern described above. These avulsion deposits spread in a fan shape to
cover the eastern half of the valley, although some of the geomorphic evidence has since
been masked by the Oyster Creek channel and its floodplain deposits.

Deposits from Avulsion #2 were sampled to the southwest by core BV-04-06
(Figs. 2-2 and 2-10). This core recorded a sequence consisting of sandy avulsion deposits
capped by red clay. Above the clay is a prominent soil horizon with calcite nodules. The
avulsion sequence is about 3 to 3.5 m thick and situated at 6.1 m below ground surface
(bgs) (Fig. 2-10). Between 6.1 and 4.9 m bgs, fine-grained sand rests above red
floodplain clay. A transition of light brown silt occurs from 4.9 to 4.7 m bgs. Red clay is
encountered between 4.7 m and the top of the core. Calcite nodules occur within the red
clay between 4.1 m and 3.8 m.

Given the existing annual precipitation along the east Texas coast, pedogenesis
typically generates calcite nodules about 1.5 to 2.0 m below the surface (Royer, 1999).
In the Brazos study area, the depth of the shallowest calcite nodules occur at 4.1 m bgs,
implying the active floodplain surface was 2 to 2.5 m bgs. Hence, rapid deposition is
believed to have occurred from 6.1 to 2.5 m bgs before pedogenic processes overcame
the rapid rates of aggradation associated with the avulsion.

Avulsion #2 was initiated from the Big Slough channel (Fig. 2-2) around ~ 4.0 ka
(Bernard et al., 1970), and led to the creation of Oyster Creek. The dip line isochron
Figure 2-10: A) Grain size data (volume %) and radiocarbon dates (five from the floodplain and one from lower point bar) for core BV-04-06. B) A portion of the core that preserves evidence of an avulsion sequence between 6.1 m and 2.5 m. A sharp transition from floodplain clay to fine-grained channel sand occurs at 6.1 m. For core location, see Figure 2-1.
cross-section indicates that the locus of valley aggradation between 6 and 4 ka was located between 60 km and 70 km updip from the present coast. The width of the valley where the avulsion break out occurred (65-70 km inland) is about 11 km, and the crevasse splay complex covers an area of >15 km². For comparison, figure 2-3A shows a modern crevasse splay deposit from a recent flood of the Brazos River (note the scale is different). This is the largest of over 30 crevasse splays observed along the lower Brazos River and it only covered an area of ~1/2 km². Many other crevasse splays along the modern Brazos have areas of 1/4 km² or less. So, the crevasse splay complex associated with Avulsion #2 is 30 to 60 times larger in area than a non-avulsion related crevasse splay.

The third and youngest of the progradational avulsions (Avulsion #3) occurred ~140 km updip of the current shoreline, where the outlines of former bifurcating and coalescing streams are preserved on the floodplain (Fig. 2-3B). The valley width at this location averages 5 km. The crevasse splay complex covers the western half of the valley and, although it is more elongate in shape, is of similar size (~11.5 km²) to the deposits of Avulsion #2. Avulsion #3 was initiated from the Oyster Creek channel (Fig. 2-3B) around 1.5 ka (Aten, 1971; 1983) and led to the creation of the present Brazos River. The rates of valley aggradation were not constrained this far inland by the dip-oriented isochron. However, the premise for correlating the 1.5 ka avulsion event to migrating valley aggradation was derived from the 3-D modeling of aggradation between 40 and 90 km inland, as discussed above. Between 40 km and 90 km inland, the rates of
aggradation increased from 2.1 km$^3$/kyr between 8-7 ka to 3.0 km$^3$/kyr between 7-6 ka. From 6 ka onward, only an average volume of sediment sequestered/kyr (0.8 km$^3$/kyr) could be calculated. Collectively, these localized rates of varying valley aggradation imply one of two things. Either sediment supply oscillated dramatically throughout the Holocene or the locus of valley aggradation migrated through this segment of the valley as it shifted inland. Sediment supply is believed to have remained fairly constant over the last 8 kyr (Fraticelli, 2003). Hence, the favored interpretation is that the locus of valley aggradation migrated through this section of the valley between 8 and 6 ka, as indicated by increasing volumes of the sediment sequestered. Since 6 ka, aggradation has continued further updip, eventually passing beyond the limit of our core and radiocarbon data. Presumably, the locus of aggradation shifted as far inland as 140 km by 1.5 ka, as summarized in figure 2-11.

Normal Listric Faults and avulsion-by-annexation

Along the Cenozoic coastal plain of Texas, numerous zones of normal listric faults have been identified that parallel the coast, yet none have been previously identified within the study area (Salvador, 1991). Aerial photographs (Fig. 2-4) and the DEM for the study area (Fig. 2-12) established spatial relationships between valley widening, flow direction of the Brazos/paleo-Brazos channels, preservation of Pleistocene uplands in the middle of the alluvial valley, and a localized steep valley floor gradient (the possible downdip side of a rollover anticline). These four prominent geomorphic features occur between 25 and 55 km inland and suggest the influence of a normal listric fault. First, at around 55 km inland the valley widens considerably from 8
Figure 2-10: The relationship between a) rates of valley aggradation, b) location of progradaional avulsions, and c) rates of sea level rise from Tornqvist et al. (2004).
Figure 2-12: A) Digital elevation model of the lower Brazos valley at a scale of 1:500,000. The light gray regions indicate Pleistocene or Pliocene uplands. Note the preserved segment of Pleistocene uplands in the middle of the lower valley (Baileys Prairie). The black dotted line illustrates how the Brazos River (in white) flows southwestward, almost parallel to the coast, for up to 20 km. B) Close-up view (scale of 1:250,000) illustrating the change in the course of the river immediately south of where the valley widens. C) Close-up (scale of 1:125,000) view of segment of valley interpreted to be the location of the downdip side of the rollover anticline. The pattern of the nodal avulsions can be seen in relation to the surface of this rollover anticline.
or 9 km to about 17 km (Fig. 2-12A). Second, downdip of where the valley widens, the Brazos River and paleo-Brazos channels divert their courses westward to parallel the coast before turning and flowing towards the coast (Figs. 2-4, 2-12B, and 2-12C). The river divergence is responsible for widening the valley. Third, a remnant of the Pleistocene uplands, called Baileys Prairie (Abbott, 2001), occurs within the middle of the Brazos valley between 25 and 35 km inland (Figs. 2-4 and 2-12A). This Pleistocene remnant occurs downdip of the river divergence. Thus, it is believed to have been protected from fluvial erosion by the same structure responsible for the change in channel direction downdip of the avulsion node. Fourth, there is a steep gradient along the downdip side of a suspected normal listric fault’s rollover anticline (Fig. 2-12C). All four factors strongly suggest movement has occurred along the fault during the Holocene, and that this structure has influenced river geomorphology (Fig. 2-4).

The suspected normal listric fault is further constrained by evidence of repeated localized avulsions between 50 and 55 km inland. Three channels are visible on an aerial photograph (Fig. 2-4) and emanate from an avulsion node. The aerial photograph displays no evidence for crevasse splay complexes, suggesting avulsion-by-annexation. Although the timing of local avulsions is not well constrained, valley aggradation would have obscured the geomorphic expression of inactive channels older than 6 ka (Figs. 2-5 and 2-6). Therefore, the channels shown in figure 2-4 are believed to be 6 ka or younger in age. In fact, the channel that fed the site of the avulsion node originates from Oyster Creek, so the age is believed to be <4 ka. This does not, however, preclude the possibility these channels may have been active intermittently for a much longer period.
2.6 Discussion

The Brazos River has experienced avulsions capable of delivering large volumes of sediment to the floodplain and leading to rapid valley alluviation. Yet, not all recorded avulsions deliver sediment to the floodplain. Five avulsions were recorded within the Holocene deposits of the lower Brazos valley and are divided into two groups. Each group is associated with a different long-term process (aggradation or fault movement) and a different style of avulsion (avulsion-by-progradation or avulsion-by-annexation). Stouthamer and Berendsen (2000) indicated that for the Rhine-Meuse fluvial system, these two processes appear to operate mutually exclusive of each other in controlling river avulsions. Their study showed that aggradation promoted the late Pleistocene and early Holocene avulsions, while fault movement controlled the locations of the mid to late Holocene avulsions. However, results from this study indicate that the two processes operated concurrently. Thus, the Brazos River provides an example of where these two processes need not compete with each other to promote avulsion.

Valley aggradation and associated avulsions

The first group of avulsions documented by this study was associated with aggradation. This group involved three regional random avulsions (at 7.5 ka, 4.0 ka, and 1.5 ka) whose channel breakout points (at 40 km, 70 km, and 140 km, respectively) backstepped along the lower Brazos valley while tracking aggradation. The members of this group avulsed-by-progradation.
Usually, rapid rates of sea level rise leads to the creation of subaerial accommodation for high sediment yield fluvial systems (Taha et al., in review-C). Hence, rapid sea level rise generates aggradation and avulsion-by-progradation (Tornqvist, 1994; Aslan and Blum, 1999; Stouthamer and Berendsen, 2000). Yet, in the case of the Brazos valley, aggradation continued to occur and to promote progradational avulsions, even after the rate of sea level rise decreased around 6 ka. This is a different scenario from the style of Holocene avulsion for the Colorado River (Aslan and Blum, 1999). In the case of the Colorado River, avulsion-by-progradation occurred only during rapid rates of sea level rise. The initial and final stages of sea level rise were slower, resulting in slower rates of valley aggradation, and leading to avulsion-by-annexation (Aslan and Blum, 1999). Thus, the Brazos and the Colorado valleys display different styles of avulsion through time for the same long-term process, aggradation. Perhaps this is because the locus of aggradation for the Brazos valley did not remain static, but instead migrated inland. But what drove aggradation inland?

This study was the first of its kind to quantify strike-oriented and dip-oriented rates of floodplain aggradation for the duration of the latest sea level rise. These two isochron cross-sections provide the basis for correlating both changes in sea level rise and the timing of shoreline retreat to aggradation. As the rate of sea level rise decreased after 6 ka, aggradation migrated inland to fill the subaerial accommodation previously generated by the upward shift in the equilibrium profile (Posamentier and Vail, 1988). This represents a time delay between sea level rise initially creating subaerial accommodation and sediment supply finally filling it. Secondly, the data indicate that a
strong correlation exists between changes in the rate of aggradation (Figs. 2-5 and 2-8) and the timing of delta (shoreline) retreat on the continental shelf. The first of two delta backstepping events (at 11.5 ka and 9.5 ka) coincided with increases in the rate of valley aggradation 40 km inland (Fig. 2-5). As the shoreline retreated to its present position (~6 ka) during a third delta backstepping event, a significant decrease in aggradation occurred 40 km inland. Most likely, shoreline retreat was responsible for shifting the locus of aggradation further updip, beyond 40 km inland (Fig. 2-6). We call upon a combination of these two factors (unfilled subaerial accommodation and shoreline retreat) as having led to the updip migration of aggradation. Regardless of which mechanism may have been more dominant, younger progradational avulsions tracked the landward migrating locus of valley aggradation, with the last one occurring at 1.5 ka, long after the rate of sea level rise had slowed down.

Within the study area, aggradation is shown to have proceeded diachronously, starting at the coast and moving inland. Once aggradation was initiated at any location in the valley, it is shown to have proceeded continuously, albeit at different rates, with no indication of floodplain re-incision. These results contradict Sylvia and Galloway’s (2006) observations of the history of Brazos valley fill. They indicate that aggradation and re-incision occurred during the latest sea level rise. Their observations seem to parallel Blum’s (1993) study of the adjacent Colorado valley, where both aggradation and re-incision of valley fill occurred during the latest sea level rise. The reason for the difference in valley fill history between the lower Brazos and lower Colorado valleys is not understood. However, with the available radiocarbon control, it can be confidently
concluded that the Brazos valley fill did not experience floodplain re-incision at any time during the latest sea level rise.

*Fault related avulsions*

The second group of avulsions of the Brazos River was associated with movement along a normal listric fault at 50 to 55 km inland. Between 4.0 ka and 1.5 ka, several localized nodal avulsions, derived from Oyster Creek, took place. From topographic analysis, these avulsions are believed to have taken place on the downdip side of the rollover anticline created by the fault (Fig. 2-12C). So, the subtle increase in topography was instrumental in promoting the avulsions. Normal listric faults are common along the Cenozoic Texas coastal plain, with the Frio fault zone occurring in the vicinity of the study area, between the present coast and 65 km inland (Martin, 1978; Salvador, 1991: Plate 2). Undoubtedly, it is a previously undocumented fault associated with the Frio fault zone that is affecting the fluvial geomorphology and sedimentology in the region. However, in the absence of industry seismic data to confirm our observations, we are not able to substantiate this structural control beyond the inference from our geomorphic analysis.

The channels resulting from the nodal avulsion site are not as stable and long lived as other avulsion-derived channels. All the nodal channels at this location (50-55 km inland) are currently inactive. These channels were active in the late Holocene because they rejoin either Oyster Creek or the modern Brazos River downstream. However, they were not occupied for extended periods because no associated natural
levee deposits exist. In addition, there are no associated avulsion deposits, hence they avulsed-by-annexation. But these channels were proper avulsions because they did carry the Brazos River’s discharge at one point; the wavelengths of their meanders are similar to the meander wavelengths of the modern Brazos (Figs. 2-4 and 2-12C).

*Volume of avulsion deposits and avulsion frequency*

One question remains about the Brazos avulsion deposits: does our evidence indicate that these deposits could comprise a significant portion of the transgressive Brazos alluvial fill? If aggradation is the only long-term process responsible for promoting avulsions capable of delivering large volumes of sediment to the floodplain, then a simple volume calculation of the avulsion deposits provides an answer. The interavulsion frequency for this style of avulsion is ~2.5 ka. The average surface area of a progradational avulsion deposit is between ~12 and 15 km² (Avulsion #3 and #2, respectively), while the average thickness of an avulsion sequence, using both our results and Aslan and Blum’s (1999) results, is ~3 to 4 m. Thus, an upper limit on the volume delivered per avulsion is ~1/18 km³. Taha and Anderson (in review-B) previously determined that ~38 km³ of sediment has been sequestered in the lower 85 km of the Brazos valley during the last 12 ka. If avulsion deposits comprised half of the alluvial fill, as has been suggested for the Colorado valley (Aslan and Blum, 1999), an avulsion would have had to have occurred about once every 35 years, and not once every 2,500 years. Therefore, avulsion deposits probably represent a fraction of the total volume of sediment sequestered in the Brazos valley.
2.7 Conclusions

1. Two styles of avulsion deposits were identified along the lower Brazos valley suggesting two distinct avulsion processes. The first group of avulsions was driven by the long-term process of valley aggradation. Initially, valley aggradation along the lower 40 km was controlled by rapid rates of sea level rise (12 to 6 ka). Episodic shoreline retreat across a variable shelf gradient from 11.5 ka onward, and the relatively slow sea level rise after 6 ka, both resulted in the locus of valley aggradation stepping landward. This is a different scenario from that presented by Blum (1993) from his work on the Colorado River. His work shows that aggradation occurred prior to 6 ka, when sea level was rising rapidly and that intervals of aggradation were terminated by incision events. Sylvia and Galloway (2006) claim to have observed similar episodes of aggradation and incision for the Brazos River, but we see no evidence that this is the case.

2. Random, regional, avulsions tracked the locus of valley aggradation. In each case, the style of avulsion (avulsion-by-progradation) contributed to net alluviation through the generation of sizable crevasse splay complexes. The individual avulsions occurred at 7.5 ka, 4.0 ka, and 1.5 ka at distances of 40 km, 70 km, and 140 km updip of the current highstand shoreline.

3. A second mechanism for avulsion is a localized increase in valley gradient on the downdip side of what is believed to be a rollover anticline, associated with a normal listric fault. The resulting local avulsions commence from an avulsion node, generated no observable deposits on the floodplain, and did not contribute to net valley aggradation.
This observation fits the definition of avulsion-by-annexation. A lack of natural levees indicates these avulsion channels were unstable and only briefly occupied. However short the duration may have been, these channels were produced by the flow of the Brazos River because the preserved meander wavelengths are of the same dimensions as the modern Brazos River. The stream that produced the avulsion node flowed through Oyster Creek channel. Therefore, the timing of the avulsions is believed to span 4.0 to 1.5 ka, implying the processes promoting both sets of avulsions were concurrent.
CHAPTER 3

Fluvial Response to Base Level Change: Aggradation and Channel Stacking Pattern of the lower Brazos Incised Valley, east Texas, USA

3.0 Abstract

Stacked channel-sands were mapped within the onshore oxygen isotope Stage 2 incised valley of the Brazos River. The stacking pattern illustrates how the Brazos River responded to the latest sea-level rise in its three directions of freedom: horizontally updip, horizontally along-strike, and vertically.

The number of stacked channels decreases (horizontally) updip from eight channels within the lower 40 km of the study area to four channels at 65 km inland. In addition, the depth of lowstand valley incision beneath the present alluvial plain decreases landward by 15 m. These two trends reflect the rapidly diminishing influence of eustacy on the fluvial system further inland.

Along-strike, the younger channels within the lower 40 km of the valley became more isolated towards the end of the latest sea-level rise, as aggradation decreased and avulsion frequency remained constant. This contradicts previous channel stacking models that neglected the influence of valley widening on channel stacking as backfilling of fluvial terraces occurred during sea-level rise. However, for the Brazos River, valley widening translated into greater lateral freedom following an avulsion.
Therefore, antecedent topography was responsible for controlling horizontal channel isolation during the late transgressive and early highstand (7.5-1.5 ka).

Vertically, the eight channels within the lower 40 km of valley fill are organized into four stratigraphic intervals. This stratigraphic organization is attributed to changes in valley gradient during aggradation. The gradient was calculated from the position of backstepping (retrograding) offshore deltas and their correlative (aggrading) onshore floodplain deposits. The chronology of the stratigraphic intervals more closely corresponds to each delta backstepping event than to variations in the sea-level curve. Thus, delta retreat, antecedent topography, and distance from the coast affected the Brazos River’s channel stacking pattern.

3.1 Introduction

An incised valley’s channel stacking pattern determines the interconnectedness of its fluvial sands. The following factors influence the channel stacking pattern by affecting the reoccurrence interval of the channel on the floodplain and its elevation above older channels: avulsion frequency, aggradation/subsidence, sediment supply, and valley width (Leeder, 1978; Bryant et al., 1995; Stouthammer and Berendsen, 2000). Previous modeling and case studies of stacking patterns concentrated on documenting how changes in avulsion frequency, aggradation/subsidence, and sediment supply controlled channel stacking (Allen, 1978; 1979; Leeder, 1978; Bridge and Leeder, 1979; Bridge, 1985; Shanley and McCabe, 1993; Mackey and Bridge, 1995; Heller and Paola, 1996; Paola, 2000; Peakall et al., 2000; Kumar et al., 2003; Kumar et al., 2004).
However, the full spectrum of channel stacking patterns may not have fully been explored because variations in valley width, resulting from valley aggradation above terraces, have not been accounted for in previous modeling and case studies.

Within an incised valley, the width of a bay or a floodplain can vary during an eustatic rise as marine flooding or fluvial aggradation overtops the fluvial terraces formed during the previous relative sea-level fall. This widening of the valley has been documented in underfilled incised valleys, such as the Trinity and Nueces valleys of Texas, where marine flooding led to episodic expansion of bays as the fluvial terraces were flooded (Rodriguez et al., in press; Simms, 2005). Yet, this has not been documented in overfilled incised valleys, such as the Brazos and Colorado valleys of Texas, where fluvial aggradation above terraces should widen the active floodplain surface. As a factor in controlling channel stacking, floodplain widening towards the end of an eustatic rise can counter the affects of diminishing aggradation, regardless of the avulsion frequency.

The onshore alluvial architecture of the Brazos River is investigated and integrated with results from an offshore study of transgressive Brazos deltas (Abdulah et al., 2004). We relate the channel-stacking pattern of the valley fill to sea-level rise and antecedent topography over the last ~20 ka. This study also illustrates how the influence of sea-level rise and fall on fluvial architecture of the valley diminishes further inland.
3.2 Background

The interconnectedness of fluvial sands has a direct bearing on petroleum migration in fluvial reservoirs and aquifer transmissivity in alluvial deposits (Allen, 1978; Galloway, et al., 1982). Early attempts at modeling channel stacking patterns and channel-sand interconnectedness usually assumed avulsion frequency was constant and subsidence was the driving mechanism (Allen, 1978; 1979; Leeder, 1978). The principal observation from this modeling was that channel-sand stacking density (interconnectedness) is inversely proportional to the rate of aggradation. However, one early study and several later studies incorporated variable avulsion frequency, where the rate of aggradation drives the frequency of avulsion (Bridge and Leeder, 1979; Alexander and Leeder, 1987; Bridge and Mackey, 1993; Bryant et al., 1995). By allowing avulsion frequency to vary with the rate of aggradation, a more uniform channel stacking density occurred within the valley fill (Bryant et al., 1995). To test whether avulsion frequency is constant or variable requires consulting the recent alluvial record.

Case studies by Tornqvist (1994) and Stouthamer and Berendsen (2000) support the notion that avulsion frequency is variable for recent fluvial systems. Yet, case studies of ancient fluvial systems (Bridge, 1985; Munoz et al., 1992) indicate that frequencies of avulsion often remain fairly constant. Perhaps some fluvial systems are more prone to constant rates of avulsion and others to variable rates of avulsion. If the existence of a recent fluvial system that exhibits constant avulsion frequency is documented, this would indicate that avulsion frequencies exhibit a continuum, regardless of the relative influence of sea-level rise and subsidence.
It is generally accepted that channel stacking patterns of both modern and ancient fluvial systems are strongly influenced by eustatic rise because of sea-level's effect upon aggradation (Galloway et al., 1982; Tornqvist, 1993a; 1994; Stouthamer and Berendsen, 2000, Poole et al., 2002). Eustacy is particularly relevant for Quaternary fluvial systems, for which glacially-driven, high-frequency, high-amplitude sea-level changes have occurred (Imbrie et al., 1984; Fairbanks, 1989; Bard et al., 1990; Bassinot et al., 1994; Lisiecki and Raymo, 2005). Results from the Rhine-Meuse delta show a general correlation between a decrease in the rate of sea-level rise, diminishing aggradation, and lower avulsion frequency (Tornqvist, 1994; Stouthamer and Berendsen, 2000; 2001). Therefore, the density of channel stacking should remain constant in a strike-oriented cross-section because a decrease in avulsion frequency counteracts a decrease in the rate of aggradation.

Another important factor that should control the channel stacking pattern of an aggrading fluvial system is valley width, although this has not been demonstrated in case studies. Leeder (1978) theorized that since channels are naturally less confined in wider valleys, valley width should affect the channel stacking density by affecting the time required before a channel reoccupies or stacks above a former meanderbelt. Specifically, the channel stacking density should decrease. Longitudinal variations in valley width, along the axis of a valley, were incorporated into a 3-D modeling program by Mackey and Bridge (1995), but they did not account for vertical changes in valley width. This
decrease in the interconnectedness of channel stacking would occur independently of constant or variable rates of avulsion.

Leeder (1978) viewed valley width in a static sense, comparing a narrower alluvial valley to a wider one for his analogy. But a valley with antecedent topography, such as fluvial terraces, would produce a wider valley with time as sea-level rose and aggradation occurred. Thus, the width of an aggradating alluvial plain with terraces should be thought of as dynamic. This changing width will have an influence upon channel stacking density through time. One possible outcome of a scenario such as this is that even as the rate of sea-level rise decreases, if the valley contains terraces, then greater channel isolation may still occur, resulting in a decrease of the channel stacking density. Predicting the channel stacking pattern and channel density is difficult in light of these two opposing forcing mechanisms.

3.3 Study Area and Brazos River Characteristics

The study area includes the coastal and alluvial plains of the lower Brazos valley, starting at the coastal town of Freeport, Texas and extending 65 km inland (Fig. 3-1). The Brazos drainage basin is the second largest in Texas, at 118,000 km² (Paine and Morton, 1989), and its headwaters start in eastern New Mexico. From here the river flows across north central Texas before debouching into the Gulf of Mexico along the east Texas coast. Although its sediment load may have varied through the latest Pleistocene and Holocene (Fraticelli, 2003), the Brazos today is a suspended load-dominated river (11:1 ratio of suspended load to bedload sediment) (Paine and Morton,
Figure 3-1: Location map of the onshore Brazos incised valley study area. All well data (●) are shown that were used to map channel (lower point bar) sands, as are the locations of the core transect of Abbott (2001) (●●) that was used to construct cross-section A-A'.

Core BV-04-06 and SH-wave land seismic survey
1989). Most of the suspended load derives from exposures of Permo-Triassic redbeds of north-central Texas. That input has allowed the Brazos River to deliver more sediment to the Gulf since the latest Pleistocene than any other Texas river with the exception of the Rio Grande River (Milliman and Syvitski, 1992).

3.4 Methods

Three datasets are used to document the location and timing of ancestral Brazos River channels: sediment cores, land seismic data from a shear wave survey, and descriptions of water well cuttings. Shell and wood material extracted from the cores were analyzed at the Woods Hole Oceanographic Institute’s NOSAMS laboratory for radiocarbon ages. These ages, when combined with the radiocarbon dates from previous studies (Bernard, et al., 1970; Abbott, 2001; Sylvia and Galloway, 2006), helped establish the absolute age of the channels.

Sediment cores (piston and split spoon samples) were collected using truck and tractor mounted drilling rigs with a push sampler. One of the cores (BV-04-06) was used to calibrate the cuttings descriptions from hundreds of water wells to interpret depositional environments. Sediment core BV-04-06 was collected in the center of the valley directly above the lowstand channel (Fig. 3-2), and sampled both floodplain and pointbar deposits. The water well data descriptions are detailed enough to identify lower point bar sands and basal lag (collectively described as ‘sand’) from upper point bar and floodplain deposits (collectively described as ‘clay’). Core (BV-04-06) showed that even though the upper point bar deposits may have a high sand content, water well
Figure 3-2: Grain size data (volume %) and corrected radiocarbon dates (five from the floodplain and one from lower point bar) for core BV-04-06. Note: the stippled sands represent the lower pointbar deposits used for calibrating the channel sands of the water well cores.
descriptions do not always allow distinction between channel sand and floodplain ‘clay’ because the fine- to very fine-grained upper point bar sand has a high clay content. Therefore, within the resolvable limits imposed by the water well descriptions, the eight channel sands identified within the lower 40 km of the valley were mapped using their lower point bar sands.

The channel architecture of the valley fill was determined with data from water well descriptions (TWDB-water wells; Cronin and Wilson, 1967; Wesselman, 1972; Sandeen and Wesselman, 1973; Naftel and Fleming, 1976). Of the ~400 well descriptions examined, 121 were used to define 8 stacked channels (Table 3-1). Approximately 150 of the remaining well descriptions contained only clay, which constrained where channel sands did not exist. The last ~130 well descriptions either didn’t penetrate deep enough, or were uninterpretable, or were unable to resolve the channel sands because two or more were stacked on top of each other (see Fig. 3-1 for locations).

A 2-D land seismic streamer survey was conducted to image the base of the Brazos incised valley and characterize the valley fill (location in Fig. 3-1). We utilized a design similar to that of Pugin et al. (2003, 2004) and found that this method minimized labor and maximized speed of acquisition. Horizontal shear-waves (SH-waves) instead of P-waves, were recorded to avoid imaging the ground water table as a reflector (Pugin et al., 2003). The SH-wave velocities are also much slower than P-wave velocities, providing higher resolution because of the shorter wave lengths (Suyama et al., 1988).
<table>
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*All sand depths are measured in meters below ground surface  
**Identification number comes from Texas Water Development Board's website on water wells drilled

Table 3-1: Water well data used to map the lower pointbar sands of the six oldest channels (Channel sand 1, 2, 3a, 3b, 4a, and 4b). Except for collected cores BV-04-06 and MED Hole 3, all other water well
The survey was conducted using a seismic streamer with the 12 receivers strung together, in series, with 1.5 m spacing. Each of the 12 receivers was comprised of two 14 Hz shear-wave geophones mounted horizontally on 25 cm steel sleds, with the axes of the geophone coils oriented 180 degrees to each other. This pairing of geophones augmented the received SH-wave component while minimizing the received P-wave component. Shot stations were spaced 1.5 m apart to provide the tight CMP spacing necessary for this high-resolution survey (Pugin et al., 2003). The data were recorded using a Geometrics™ 12-channel seismograph. ProMAX™ software was used to process the data, which was stacked but not migrated.

Using core BV-4-06, the average velocity for the seismic data was determined to be 156 m/s. This was close to the velocity (160 m/s) determined from the best stacked-section while processing the data. Therefore, each 100 ms (in two-way travel time) represents 8 m. The SH-wave velocity was about 1/10 the velocity of a conventional P-wave survey on unconsolidated sediment, similar to the results of Pugin et al. (2003).

All core and water well data were integrated and mapped with GIS software (ESRI ArcMAP 9.0, ERDAS Imagine 7.1, Rockworks). A total of 242 lithologic breaks (between clay and sand) were converted to X, Y, and Z control points and provided the basis for the interpolation of the three-dimensional stacked channel sand map.

3.5 Results

The Brazos fluvial response to base-level change differs vertically and horizontally. Also, the affect of sea-level rise on fluvial processes diminishes from 40 to
65 km away from the current shoreline, and is manifested by a decrease in the number of stacked channels further inland. Before presenting the channel stacking results, we derive the chronology of the stacked channels and the evolution of the valley gradient.

Channel Chronology

The age of channels was derived in two ways. First, the average depth to the top and base of each channel sand (Table 3-2) was derived from the water well data (Table 3-1) and used to assign relative ages from superposition of the channels (Fig. 3-3). Absolute ages were then assigned to six of the eight channel sands (Lowstand, 3b, 4a, 4b, 4c, and 4d) using either radiocarbon dates from lower point bar deposits or, in the case of channels 4c and 4d, a comparison of the relative maturity of natural levee deposits as a proxy for their age.

The Brazos River flowed through Channel 4c (Jones Creek) and then Channel 4d (modern Brazos River) during the last 1,500 years (Aten, 1972; 1981). The volume of natural levee deposits associated with these two channels is used to estimate their ages (Fig. 3-4). The estimated total volume of the natural levee deposits for these two channels (calculated using ArcMap) is 4.9 km$^3$, of which 3.2 km$^3$ (65%) is associated with Channel 4c. Therefore, Channel 4c is interpreted have been active for ~1,000 years, while Channel 4d was active for about the last 500 years. Excluding these two youngest channels, the interval of time between each regional avulsion during the transgression (~19-6 ka) and early highstand (~6 ka to present) was surprisingly constant, averaging approximately 2 kyr. Ages for channels 2 and 3a were estimated using the average
<table>
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<th>Systems Tract</th>
<th>Channel Sand</th>
<th>Channel Stacking</th>
<th>Alternate Name</th>
<th>Depth of sand interval below</th>
<th>Floodplain Below ground surface (bgs)</th>
<th>Age</th>
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<td>-20 ka to 11.5 ± 1.0 ka</td>
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<td>Transgressive</td>
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<td>Brazos River</td>
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<td>~0.5 ka to Present</td>
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</tbody>
</table>

* Red clay is found underlying this channel sand, often extending to depths of 150 m or more. Hence, this channel sand cannot be mistaken for previous cycles of valley fill. Along the lower 40 km of the coast, the Lowstand Channel sands average 10-11 m thick, versus 7-8 m for all other transgressive and early highstand channel sands.

** Where it is offset from the underlying Lowstand channel, Pleistocene clay exists beneath the Channel 2 sands.

*** This channel was formed after a major avulsion 40 km up dip of the coast that occurred at 7.5 ka (Taha et al., submitted). Afterwards, the Brazos River no longer followed a route paralleling the earlier channels, and by the coast was offset by more than 10 km.

† Channel sands 2 and 3a stacked successively above Channel sand 1

TT Approximately 5.5 kyr after the isolated Channel sand 3b was formed, Channel sand 4d was stacked above it

Table 3-2: The eight stacked channels within the lower 40 km of the Brazos incised valley are organized into four stratigraphic intervals. An average depth to the top and base of the channel sands, as well as the top of the correlative floodplain, are measured with respect to the ground surface.
Figure 3-3: A) Time series showing the aggradation of floodplain deposits (developed from floodplain deposition in Fig. 3-5) and channel sand stacking (developed from this study) at ~20, ~11.5, ~9.5, 7.5, 6.0, 4.0, 1.5, and 0.5 ka. The Stratigraphic Intervals I-IV sub-divide the channel sands. Depth to base of sand = 0, depth to top of sand = ▲, depth to paleo-floodplain = □. B) Two recent sea-level curves showing a strong correlation between rates of floodplain aggradation, channel stacking, and sea level rise.
Figure 3-4: Elevation map of the lower Brazos valley. Geomorphic expression of recent natural levees is evident. Quantification of the Brazos River and Jones Creek natural levee deposits above the present floodplain was used to determine their relative ages, since these channels are known to have been active between 1.5 ka and the present.
duration of the other channels. Bridge (1985) and Munoz et al. (1992) report a similar paleo-channel life span of ~1.5 to ~2.5 kyr. The time series for channel occupation, and the depths to the base of each channel sand, provides the chronostratigraphic framework shown in figure 3-3.

The second method for deriving the age of the channels utilized floodplain chronology. A time series of the depths to paleo-Brazos floodplains was established using a strike-oriented isochron section of floodplain deposits (Fig. 3-5) to see how the channel sands and paleo-floodplain surfaces correlated (Fig. 3-3). The isochron section was constructed from radiocarbon dates from cores taken across the valley by Abbott (2001) 40 km updip of the coast (Taha and Anderson, submitted-A). This cross-section provides a second independent chronostratigraphic framework of the Brazos alluvial valley fill using floodplain deposits, and complements the chronostratigraphic framework of the channel sands. A strong correlation exists between the chronostratigraphy of both the channel sands and the floodplain deposits.

**Valley Gradients**

To establish the response of the equilibrium profile to sea-level rise, a time series of valley gradients was constructed for ~20 ka, ~11.5 ka, ~9.5 ka, and the present. The gradient of the incised valley at ~20 ka was determined from a map of the Stage 2 sequence boundary (Taha and Anderson, submitted-A), while the gradient of the modern valley was determined from a digital elevation model of the lower Brazos valley. To
Figure 3-5: Isochron cross-section (A-A') of the Brazos valley at 40 km inland of coast (data modified from Abbott, 2001). All radiocarbon ages are corrected. For profile location see Figure 3-1.
determine the intermediate gradients, paleo-shorelines had to be identified and then projected updip to their correlative floodplain deposits.

The Brazos delta has a history of stepping landward with time during the most recent sea-level rise, and the locations of these deltas have been documented by Abdullah et al. (2004). The age of these deltas was constrained by Snow (1998) as being about 11.5 and 9.5 ka. This knowledge was used to determine the paleo-gradients and the effect of backstepping deltas on the channel stacking pattern updip. The floodplain gradient at 11.5 ka was derived by taking the shoreline associated with the first transgressive, backstepped Brazos delta, found 60 km offshore at 45 m water depth (Snow, 1998; Abdullah et al., 2004), and correlating it to its most landward onlapping floodplain deposit 40 km onshore and 10 m below sea-level (Fig. 3-5, see Fig. 3-1 for location). The estimated gradient at 11.5 ka was 35 cm/km. The floodplain gradient at 9.5 ka was derived by taking the shoreline associated with the second transgressive, backstepped delta of the Brazos, which occurs 40 km offshore at 26 m water depth (Snow, 1998), and correlating it to its most landward, onlapping floodplain deposit at 65 km onshore and 2 m below sea-level. This correlative onlap surface was established with a corrected (calendar year) radiocarbon date of 9,550 years, obtained from core BV-04-06 (Fig. 3-2). The estimated gradient at 9.5 ka was 24 cm/km.

Stratigraphy and Avulsion Location

The stratigraphy of the channel sands within the lower 40 km of the Brazos valley are summarized in Table 3-2. The site of avulsion that generated each new stacked
channel from its parent channel is known for channels 3b, 4b, 4c, and 4d. Channel 3b was formed after a major avulsion 40 km updip of the coast (Figs. 3-4 and 3-6D). This avulsion occurred at 7.5 ka (Taha and Anderson, submitted-A). Bernard et al. (1970) used radiocarbon dates to determine that Channel 4b (Oyster Creek channel) was the former route of the Brazos River between 4 and 1 ka. Aten (1971, 1983) argues that the river abandoned Oyster Creek to form the modern Brazos River channel around 1.5 kyr ago, based on archaeological evidence. The geomorphic expressions of the natural levees from Oyster Creek channel are still evident in the landscape (Fig. 3-4). At ~1.5 ka, the Brazos River avulsed about 140 km updip from the coast (Taha and Anderson, submitted-A). Since then the river has occupied the same meander belt between 140 and 60 km inland. Along the lower 60 km, the river initially flowed through the route generated by Channel 4c, which is presently occupied by Jones Creek (Aten, 1983). Approximately 500 years ago, an avulsion occurred around 60 km updip that led to the modern Brazos Channel (4d). The locations of the avulsions that generated channels 2, 3a, and 4b are not known.

*Vertical channel stacking*

The eight channel sands within the lower 40 km of the Brazos valley group vertically into four distinct stratigraphic intervals (Stratigraphic Intervals I, II, III, and IV) (Fig. 3-3). These stratigraphic intervals are used to relate the eight stacked channels to the changes in valley gradient noted above that occurred as the deltas stepped landward (Table 3-2 and Fig. 3-3). The lower pointbar channel sands were interpolated and mapped for each of the six oldest channels (Fig. 3-6), using the data from the TWDB
Figure 3-6: Step-wise evolution of channel sands within the Brazos valley. Water well data points used for the interpolation are shown for the six oldest channels. A) No channel sands were mapped for Jones Creek or the modern Brazos River channels, but their locations are shown, B) Channel Sand 1, C) Channel Sand 2, D) Channel Sand 3a and 3b, E) Channel Sand 4a, F) Channel Sand 4b.
water well database (Table 3-1). Although the two youngest channel sands (4c and 4d) were not mapped, their modern streams (Jones Creek and the Brazos River) are shown in figure 6A-F.

Along-strike channel stratigraphy

Figure 3-7 shows the overall stacking pattern of the channel sands and highlights the isolation of younger channels. The three oldest channels sands (1, 2, and 3a) stack almost above each other. Channel sands 3b, 4a, and 4b are more widely separated and are located a considerable distance to the east. It should be noted that channel sands 4a and 4b are the most isolated, yet were deposited when the rates of sea-level rise and aggradation had decreased considerably. The two youngest channels (4c and 4d) stack above Lowstand Channel Sand 1 and channel sand 3b, respectively, although their channel sands are not shown in either figure 3-6 or 3-7.

Updip channel stacking at 65 km inland

So far, only the affect of sea-level rise on channel stacking within the lower 40 km of the incised valley has been considered. To document the affect of sea-level rise on channel stacking farther updip of the current shoreline, the study was extended to 65 km inland. A cross section of the valley (Fig. 3-8) was constructed using Core BV-04-06 (Fig. 3-2), data from several water wells, and the shear wave seismic survey results (Fig. 3-9). An important observation is that only four channels exist in the more landward reaches of the valley (Fig. 3-8), in contrast to the eight channels found within the lower reaches of the valley (Figs. 3-3, 3-6, and 3-7).
Figure 3-7: 3-D perspective of all Brazos stacked channel sand. Channel sands were not mapped for either Jones Creek or the modern Brazos River.
Figure 3-8: Cross-section at 65 km updip of the Brazos shoreline. Only four stacked channel sands are observed here, versus eight further downdip, indicating how rapidly the influence of eustacy and aggradation diminishes inland. See Fig. 3-1 for location.
Figure 3-9: Shear-wave seismic cross-section. A) Uninterpreted seismic data. B) Interpreted seismic data. Reflector A denotes base of Brazos incised valley; Reflector Package 1 are chaotic reflectors, which correlate to channel sand deposits in core BV-04-06; Reflector Package 2 are horizontal to sub-horizontal reflectors, which correlate to floodplain deposits in core BV-04-06 (location Fig. 3-1, lithology Fig. 3-2).
Core BV-04-06 (Fig. 3-2) penetrated the deepest portion of the incised valley (Fig 3-8). Lower pointbar sands were encountered between 18 and 25 m below ground surface (bgs) and rest on dark brown Pleistocene clay. A radiocarbon age from a shell fragment at 24.1 m yielded an age of 20,340 years. This age indicates that the channel sands are Stage 2 lowstand deposits. For comparison, the base of the incised valley at this location was 25 m bgs, contrasting with the >40 m of fluvial incision bgs documented along the coast (Taha and Anderson, submitted-A). The lithologic descriptions from seven nearby water wells also help constrain the extent of the lowstand channel sand and other stacked channels (Fig. 3-8).

The base of the valley is defined by strong amplitude, seismic reflector found between 200 and 320 ms (Reflector A) (Fig. 3-9). This agrees with data from core BV-04-06 (Fig. 3-2) as it penetrated the valley base at 25 m bgs. This reflector is an irregular surface with up to 10 m of relief. Above this prominent seismic reflector are two seismic reflection packages. The first reflection package is characterized by low amplitude, chaotic reflectors and corresponds to fluvial sands. The second reflection package is characterized by medium amplitude, horizontal to sub-horizontal reflectors and correlates with floodplain deposits.

### 3.6 Discussion

*Channel stacking pattern*

The onshore Brazos valley fill is composed mostly of floodplain and stacked channel deposits. There are eight stacked channels within the lower 40 km of the valley.
Using the chronology of the eight channels and the well-established sea-level curve for the northern Gulf of Mexico (Tornqvist et al., 2004), the channels are grouped as follows. The oldest and deepest channel belongs to the lowstand systems tract (20 ka calendar years and younger) and potentially the early transgressive systems tract and hence, is referred to as the lowstand channel. The three intermediate channels (Channels 2, 3a, and 3b) belong to the transgressive systems tract. The four youngest channels (Channels 4a, 4b, 4c, and 4d) belong to the early highstand systems tract. Two trends are evident from the pattern of channel stacking. First, channel sands are vertically grouped into distinct stratigraphic units (Fig. 3-3 and Table 3-2). Second, younger channels are more horizontally isolated than the older channels (Fig. 3-7).

The lower point bars of the eight stacked channels may be vertically grouped into four, sand-prone stratigraphic units onshore (Fig. 3-3). A strong correlation exists between delta backstepping, changes in valley gradient, and the grouping of the stacked channel sands into the four stratigraphic units. The delta backstepping is driven by sea-level rise (Abdulah et al., 2004; Taha and Anderson, in review-A). Each time the Brazos delta backstepped, valley aggradation decreased. The onshore gradient during the lowstand was 43 cm/km, around 19 ka (Weaver et al., 2003). The gradient during the transgression, around 11.5 ka, was 35 cm/km and the gradient around 9.5 ka was 24 cm/km. Presently, the gradient along the lower 100 km of the alluvial floodplain is 20 cm/km.
The oldest stratigraphic interval includes the lowstand channel sand and occurs at a depth of 29 to 39 m bgs. Given our chronostratigraphic constraints, the lowstand and early transgressive channel remained stationary from >19 to \( \sim 11.5 \pm 1.0 \) ka. This is considerably longer than the duration for which the younger channels occupied their channel routes (average \( 2.0 \pm 0.5 \) kyr). The greater longevity over which channel processes operated explains why the lowstand channel is at least twice as wide as the other channels (Fig. 3-8). In addition, its lower point bar sands are 25% to 35% thicker than those of the seven younger channel sands (Fig. 3-3 and Table 3-2). Thus, the lowstand channel remained in place longer and aggraded more than the younger channels prior to avulsing, probably as a consequence of the greater channel/valley entrenchment that had to be overcome before a major avulsion could occur.

Stratigraphic Interval II (Fig. 3-3) is composed of the lower point bars sands of one channel, Channel 2, and extends from 22 to 30 m bgs. Channel 2 was formed at 11.5 \( \pm 1.0 \) ka, coinciding with the timing of the first delta backstepping event on the continental shelf (Abdulah, et al., 2004; Snow, 1998). The valley gradient at this time is estimated to have been \( \sim 35 \) cm/km. It would seem that some intrinsic threshold to the fluvial system was crossed as a result of rapid sea-level rise and this caused the delta to backstep. No floodplain deposits older than 12 ka have been observed at 40 km inland (Fig. 3-3), indicating that aggradation within the lower onshore valley occurred after this time. As the delta backstepped, the fluvial gradient appears to have re-equilibrated quickly. This event initiated the first major avulsion, which led to the formation of Channel 2, and hence the first stacked channel.
Stratigraphic Interval III includes the lower point bar sands of two channels (3a and 3b), which extend from 14 to 25 m bgs (Fig. 3-3). Channel 3a was formed at ~9.5 ± 0.5 ka, coinciding with the timing of the second delta backstepping event on the continental shelf (Abdulah et al., 2004; Snow, 1998). The gradient at this time is estimated to have been ~24 cm/km. An avulsion at ~7.5 ka created a new channel route, Channel 3b, at approximately the same stratigraphic level. The Brazos delta is known to have backstepped a third time, positioning the shoreline near its present location. However, the lower limit for the timing of this event is not constrained, so it is not known if it is old enough to have helped initiate the avulsion that formed Channel 3b, or if it came later.

The fourth stratigraphic interval includes the lower point bar sands of the four youngest channels, 4a, 4b, 4c, and 4d (Fig. 3-3). The sands of this unit extend from 9 to 18 m bgs and occur after ~6 ka, when the rate of sea-level rise had slowed considerably (Tornqvist et al., 2004). This reduction in the rate of sea-level rise corresponds to a significant reduction in the rate of aggradation along the lower 40 km of the Brazos valley (Fig. 3-5). At the same time, Channel 3b was abandoned and the next younger channel, the Big Slough channel (Fig. 3-7), was formed. The reason for these two events being coincident is not yet clear. But by no later than 6 ka, the Brazos delta was close to its present shoreline position (Bernard et al., 1970). A lack of valley aggradation from ~6 ka onwards led to the four youngest channels all occurring at about the same stratigraphic interval.
As illustrated in Figure 3-3 A and B, a strong correlation exists between the rate of sea-level rise and the pattern of channel stacking within the Brazos valley. Further observation indicates that a correlation exists between the timing of delta backstepping events, channel avulsion, and the formation of the sand prone stratigraphic units. Yet, the correlation between rates of sea-level rise/delta backstepping events and channel interconnectedness is more ambiguous. Despite rapid sea-level rise and the resultant aggradation, the three oldest channel sands of the Brazos valley (Channels 1, 2, and 3a) stacked virtually on top of each other (Fig. 3-7). Channel 3b, which was also formed when rates of aggradation were high, was the first channel to diverge from the previous channels. From where this channel diverged almost 40 km inland to its location at the coast, it had shifted eastward more than 10 km. The next four youngest channels (4a-4d) were established after the rate of sea-level rise began to decrease and aggradation had almost stopped. Relative to the lowstand channel, Channel 4a migrated laterally eastward 40 km by the time it reached the coast, and Channel 4b was about half way in between Channel 4a and the three oldest channels.

The most plausible explanation for the greater lateral separation of the younger channels, under both varying conditions of sea-level rise and varying rates of aggradation, is the influence by antecedent topography. As aggradation proceeded, antecedent topography, in the form of at least one fluvial terrace, was backfilled (Taha et al., submitted-C). This widening of the floodplain acted to increase channel isolation and reduce channel interconnectedness because the route a new river could take upon
avulsion was less confined. The process of backfilling antecedent topography requires both sea-level rise and aggradation, but backfilling proceeds as a step function from one terrace to the next, rendering it difficult to predict when the valley will widen. Also, just because the valley widens does not mean the channel will migrate across the valley right away. Potentially, the lateral movement of Channels 3b, 4a, and 4b are all related to the valley widening associated with a single terrace. The extent of the terraces was difficult to access from our data.

Our results support Leeder's (1978) hypothesis that wider valleys will result in more isolated channels, but do so within a single valley because of the effect of antecedent topography. Therefore, in this case study, antecedent topography has overcompensated for the effect of diminishing rates of aggradation and has forced the channel stacking density to decrease with time.

Ideally, for the purposes of better modeling the migration pathway of groundwater and petroleum fluids in older incised valley deposits, the present study needs to integrate the offshore and onshore stacked channels sands. Yet, this is quite difficult because as aggradation has propagated updip in response to sea-level rise and delta backstepping (Taha and Anderson, submitted-A), a shingling effect has occurred, resulting in diachronous valley fill. Thus, the preserved offshore valley fill deposits will predate much of the onshore deposits that lie above the lowstand channel sands. In addition, transgressive ravinement will have removed the youngest portion of the offshore incised valley fill and compromised reservoir interconnectedness. The four youngest channels
within the lower 40 km of the onshore incised valley are all early highstand channels and are generally more isolated than older, onshore channel sands. These four early highstand channels all terminate at the current highstand shoreline and do not connect to any channels of the offshore incised valley. Therefore, on either side of the current highstand shoreline, a strong asymmetry exists in the preservation of onshore and offshore incised valley fill and the interconnectedness of stacked channel sands.

*Influence of sea-level with distance inland*

From core BV-04-06 and the shear wave seismic survey, the depth to the base of the incised valley at 65 km inland is documented as ~25 m bgs while it is ~40 m bgs at the present coast. Therefore, the depth of the lowstand valley increases by ~15 m toward to the coast. A valley cross-section at 65 km updip of the current highstand shoreline shows that only four stacked channels are present (Fig. 3-8). At this distance inland, the channel sands associated with channels 2, 3a, and 3b, are not discernable from the lowstand channel sand (dated using the oldest radiocarbon date in Fig. 3-2). In addition, the avulsion that generated Channel 4c from Channel 4d occurred further downstream. Hence, an important distinction is drawn between the channel stacking pattern within the lower 40 km of the Brazos valley, where eight stacked channels occur (Figs. 3-3, 3-6, and 3-7), versus at 65 km from the coast, where only four stacked channels are documented (Fig. 3-8). The implication is that several of the avulsions that produced the eight stacked channels along the lower 40 km occurred south of 65 km. This provides a constraint on the diminishing influence of base-level change on channel avulsion inland from the coast.
3.7 Conclusions

The Brazos River’s response to base-level rise has three directions of freedom and we document how it has responded in each of these directions. Vertically, delta backstepping and episodic aggradation drove channel stacking, whereas horizontally along-strike, antecedent topography has controlled channel isolation. In an updip direction, the affect of sea-level rise on fluvial processes diminished rapidly from 40 to 65 km inland from the current shoreline. This diminishing influence is manifested by a decrease in the number of stacked channels further inland.

Eight individual channels occur within the lower 40 km of the onshore Brazos valley. A chronostratigraphic framework for the valley was derived from corrected radiocarbon dates and the geomorphic prominence of natural levee deposits. Where radiocarbon control was absent for two of the channels, their ages were estimated by averaging the duration of younger Brazos channel occupations. This chronostratigraphic framework was corroborated using an independent chronostratigraphic framework constructed from paleo-floodplains (Taha and Anderson, submitted-A). From this chronology, the avulsion frequency is determined to be constant.

The sands of the stacked Brazos channels were grouped into four stratigraphic intervals. The timing for the initiation of these four stratigraphic intervals corresponds to four events related to sea-level rise. The oldest and deepest stratigraphic interval consists of a single channel that was active when sea-level was at its lowest, ~19 ka (Weaver et al., 2003), and the valley was neither incising nor aggrading. The second stratigraphic
interval also contains only a single channel and was formed ~11.5 ka, corresponding to the first of three Brazos delta backstepping events. The third stratigraphic interval includes two channels that occur at approximately the same stratigraphic interval, and was initially formed ~9.5 ka. This corresponds to a second Brazos delta backstepping event. Both delta backstepping events are attributed to rapid sea-level rise and, in turn, shifted the river equilibrium profile updip, accelerating fluvial aggradation and avulsion. A third delta backstepping event placed the delta close to its present location, but the timing of this event is not well constrained and its effect on channel stacking remains unknown. The fourth stratigraphic interval encompasses the four youngest channels and was initially formed around 6.0 ka, corresponding to a significant decrease in the rate of sea-level rise. This is associated with minimal rates of aggradation along the lower 40 km of the Brazos valley since ~6 ka.

The results of this study are not consistent with those who have asserted that as sea-level rise slows, the frequency of avulsion decreases, resulting in constant channel stacking density and channel interconnectedness (e.g. Tornqvist, 1994). In the case of the Brazos fluvial system, as the rate of sea-level rise slowed, the frequency of avulsion remained constant, which should have resulted in a higher channel stacking density. Yet, the channel stacking density did not increase. In fact, increasingly greater isolation of younger channels occurred. This is believed to reflect the control of antecedent topography (i.e. terraces) on valley aggradation.
The number of channels found within the lower 40 km of the Brazos valley is reduced by one half at a distance of 65 km inland. Lowstand fluvial incision also decreased from >40 m at the coast to ~25 m, 65 km inland. Both the reduction in the number of stacked channels and depth of fluvial incision reflects the diminishing effect of base-level change on the fluvial system with distance from the coast.

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CHAPTER 4

Sediment supply, the equilibrium profile, and subaerial accommodation: Case study of the Brazos and Trinity fluvial systems, east Texas

4.0 Abstract

Sequence stratigraphic models emphasize the effect of relative sea-level rise on the equilibrium profile of a river and the creation of near shore subaerial accommodation. However, an equilibrium profile must be anchored at both ends. The upper reaches of an equilibrium profile along a prograding passive continental margin often exist beyond the reach of relative sea-level and represent the cumulative effect of sediment yield on valley aggradation from all past cycles of relative sea-level rise. This cumulative effect, in turn, will exert limits on the downdip creation of future subaerial accommodation during relative sea-level rise. This study relates differences in the long-term (multiple relative sea-level cycles) sediment yield of the Brazos and Trinity rivers of east Texas to the present elevations of their alluvial valleys and concludes that the long-term lower sediment yield and lower alluvial valley of the Trinity River limits how much subaerial accommodation was created during the most recent relative sea-level rise.

The lower 50 km of the onshore Trinity incised valley is currently flooded, forming Galveston Bay, while an equivalent segment of the Brazos incised valley maintains a shoreline delta. From the Trinity bayhead delta to ~300 km inland, the Trinity alluvial valley has 36.6 km³ less sediment beneath its surface than an equivalent segment of the Brazos alluvial valley. This difference in volume reflects the cumulative effect of the disparity in sediment yield over multiple episodes of relative sea-level rise.
The most recent lowstand and transgressive deposits of the lower (50 km) onshore Trinity incised valley includes 9 km$^3$ of bay sediment and 3 km$^3$ of basal fluvial deposits, while an equivalent segment of the Brazos incised valley is filled with 28.6 km$^3$ of mostly fluvial deposits. For ~100 km offshore, the Trinity incised valley is filled with 19.9 km$^3$ of bay and fluvial sediment, while the Brazos incised valley is filled with 16.0 km$^3$ of fluvial deposits. These offshore volumes do not account for sediment lost by transgressive ravinement. Thus, while these two fluvial systems have experienced the same recent relative sea-level rise, the long-term lower sediment yield of the Trinity River has acted to permanently suppress its equilibrium profile and thereby limit the creation of subaerial accommodation. The model developed here applies to prograding passive continental margins because it depends on successive relative sea-level cycles to lengthen an alluvial valley.

4.1 Introduction

With the development of sequence stratigraphy three decades ago, sea-level change is now viewed as more influential than sediment supply in controlling stratigraphic architecture (Vail et al., 1977; Haq et al., 1987). When sea-level change is combined with local subsidence, a relative sea-level (RSL) curve is created which is understood to control the accommodation, or potential space available, for the deposition of sediment (Jervey, 1988). Thus, sediment supply is considered to take a secondary role, and deposition is manifested as repeatable, genetically related stratigraphic packages by the oscillation of accommodation during an RSL cycle (Posamentier et al., 1988). This is true even where synthesis studies acknowledge variability between individual drainage
basins in their stratigraphic architecture because, for any given drainage basin, the style of stratal packaging on the continental shelf and slope is repeated from one RSL cycle to the next (Anderson et al., 2004). The concept of the control that RSL exerts over accommodation works well for marine environments and has similarly been extended to near shore subaerial environments. However, the control that accommodation exerts over stratal packaging is not entirely independent of sediment supply. This paper focuses on how sediment supply would effect the accommodation of fluvial environments.

To better understand subaerial accommodation, the original definition for accommodation and the generic definition for subaerial accommodation are revisited. Although the original definition for accommodation emphasizes the 'potential' space available for sediment to fill (Jervey, 1988), other studies have sought a more tangible approach. Cross (1988) argued that the concept of accommodation would be more geologically significant if it were defined as the 'realizable' instead of the 'potential' space available. In this view, sediment supply was important to the equation of accommodation because it defined how much accommodation had been created and further, because this might effect the creation of accommodation during the next cycle. More recently, Muto and Steel (2000) presented a similar argument by suggesting that if RSL rise generates accommodation that cannot be filled by available sediment supply, then this accommodation effectively does not exist. For emphasis, they used the extreme example of a shut down in sediment supply during RSL rise to ask how the creation of accommodation would be measured.
This debate would be of little significance for fluvial environments if the present creation of accommodation were independent of the past history of valley fill (i.e. realizable accommodation). However, this study proposes that the subaerial accommodation generated during the most recent RSL rise is dependent upon the cumulative affect of realizable accommodation from all previous cycles of valley aggradation because stratigraphic base level must be anchored at both ends. This dependence is readily apparent when the longitudinal profiles of two fluvial systems with different long-term sediment yields are contrasted. In addition, we show that all the tributaries of a drainage basin respond to the long-term fill conditions of the main alluvial valley, with the counterintuitive result that an underfilled incised valley (Simms et al, in press) actually excavates more sediment from its drainage basin through time than its overfilled counterpart.

The creation of subaerial accommodation is generalized as an upward shift in the equilibrium profile of a river (Posamentier and Vail, 1988). Two end member mechanisms exist for how this upward shift would occur. Along the lower reaches of a fluvial valley, the first end-member mechanism is driven by RSL rise, in much the same manner as marine accommodation (Posamentier and Allen, 1999). This mechanism has sometimes been taken to suggest that as RSL rise creates marine accommodation, it leads to a proportional increase in subaerial accommodation (Shanley and McCabe, 1994; Willis, 1997; Aslan and Blum, 1999). As with marine accommodation, the assumption is that the role of sediment supply to stratigraphic architecture is secondary as it passively fills the available subaerial accommodation.
The second end member mechanism deals with subaerial accommodation updip of the reach of RSL change, which Shanley and McCabe (1993) generalize as ~ 100-150 km inland of the contemporary shoreline. In this case, subaerial accommodation must be controlled by one or both of the two other extrabasinal factors, climatic change and tectonics (Blum, 1993; Martinsen et al., 1999; Holbrook et al., 2006). Although the role of sediment supply is understood to be secondary in the two previous mechanisms for creating accommodation, the impact of sediment supply is here somewhat obscured because, by definition, variations in sediment supply are controlled by the interplay between climate change and tectonics (Schumm, 1993). In both these mechanisms, the direction that the equilibrium profile moves to compensate for the changing conditions has been termed stratigraphic base level (Shanley and McCabe, 1994; Muto and Steel, 2000). We propose a definition for subaerial accommodation that combines elements of the previous two explanations, and which leads us to emphasize the cumulative effect of realizable accommodation rather than potential accommodation.

The ability of sediment supply to regulate the creation of subaerial accommodation is shown here to be independent of RSL. Instead of viewing present sediment supply as a variable necessary for 'realizing' the updip equilibrium profile by being anchored at the shoreline, as Holbrook et al. (2006) have done, we view past sediment supply as the variable that anchors equilibrium profile at the updip end. To emphasize the dependence of the longitudinal profile on long-term sediment supply, we compare two fluvial systems that have experienced the same recent RSL rise but have different long-term sediment yields (the Brazos and Trinity rivers of east Texas). This
model for understanding subaerial accommodation near the shoreline is developed for prograding passive continental margins, like the northwest Gulf of Mexico during the Cenozoic (Galloway, 1981; Galloway, et al., 1982; Galloway, 1986), because it depends on the cumulative effect of sediment supply on the longitudinal profile during successive RSL cycles. If modified, it may be applicable to other types of passive continental margins such as the southeast Brazilian coast, which has both prograded and retrograded during the Cenozoic in response to changes in the sediment source (Guardado, et al., 1989; Modica and Brush, 2004).

4.2 Study Area and Drainage Basin Characteristics

The study area includes the entire Brazos and Trinity drainage basins, but mostly concentrates on the lower ~450 km of their alluvial and incised valleys, extending from the Cenozoic coastal plain of east Texas to the adjacent mid-continental shelf (Fig. 4-1). The two valleys are subdivided into three segments for analysis: offshore, from the shoreline to the mid-continental shelf (~100 km); coastal plain, from the shoreline to ~50 km updip; and continental, from ~50 km to ~350 km inland. Figure 4-2 highlights the 3D geomorphology of the underfilled Trinity valley in relation to the overfilled Brazos valley.

The Brazos drainage basin is the second largest in Texas at 118,000 km² (Pain and Morton, 1989). Its headwaters start in east-central New Mexico and the river flows across central Texas to debouch into the Gulf of Mexico along the east Texas coast. Presently the Brazos is a suspended load dominated river delivering approximately an 11:1 ratio of suspended load to bedload sediment (Paine and Morton, 1989). The Trinity
Figure 4-1: Location map of the study area. This includes both the Stage 2 incised valleys (offshore and onshore) and the alluvial valleys along the east Texas Cenozoic coastal plain for the Brazos and Trinity rivers.
Figure 4-2: 3-D geomorphic perspective of the overfilled Brazos incised valley and the underfilled Trinity incised valley. Relative to the Brazos alluvial valley, the lower elevation of the Trinity alluvial valley updip of the coastal plain clearly reduces the total subaerial accommodation that was created during the latest sea-level rise.
drainage basin is the fifth largest in Texas at 46,000 km$^2$ (GLO website), but has a substantially lower sediment yield than the Brazos River (Table 4-1). It is also a suspended load-dominated river.

4.3 Methods

*Offshore Incised Valleys*

Abdulah et al. (2004) used high-resolution marine seismic data, platform borings, and radiocarbon dates to map the marine-Oxygen Isotope Stage 2 (MIS 2) offshore Brazos incised valley, which was excavated when the continental shelf was exposed (Fig. 4-1). This valley is part of a regional erosional surface that is part of the MIS 2 sequence boundary. The surface has up to 30 m of relief where the incised valley exists (Abdulah et al., 2004; Simms et al., in press). For this study, the original paper seismic sections of Abdulah (1995) were reinterpreted to quantify the valley dimensions and to calculate the volume of valley fill using GIS software (ArcMap 9.0).

Thomas and Anderson (1994) used high-resolution marine seismic data, platform borings, and radiocarbon dates to map the MIS 2 Trinity incised valley. The valley was contoured every 4 m (Thomas and Anderson, 1994). Simms et al. (in press) digitized the map contours. For this study, the digitized incised valley contours were interpolated and subtracted from the present bathymetry using GIS software (ArcMap 9.0) to calculate the volume of fill.
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<tr>
<td>1977-1980 (tons/day)</td>
<td>28608**</td>
<td>2717*****</td>
<td>9.5</td>
</tr>
<tr>
<td>Incised valley fill (km³)</td>
<td>10-110 km</td>
<td>0-102 km</td>
<td>9.5</td>
</tr>
<tr>
<td>Offshore (0-110 km)</td>
<td>16</td>
<td>19.9</td>
<td>-3.9 km³</td>
</tr>
<tr>
<td>Onshore (0-50 km)</td>
<td>28.6</td>
<td>12</td>
<td>16.6 km³</td>
</tr>
<tr>
<td>Vol</td>
<td>Onshore (40-350 km) see Fig. Y</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Surface geology</td>
<td>See Table 2</td>
<td>See Table 2</td>
<td></td>
</tr>
<tr>
<td>Incised valley surface area (km²)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Onshore (0-50 km)</td>
<td>1250</td>
<td>1240</td>
<td></td>
</tr>
<tr>
<td>Onshore (0-350 km)</td>
<td>3129</td>
<td>3120</td>
<td>99.7</td>
</tr>
<tr>
<td></td>
<td>177.5 km³</td>
<td>140.9 km³</td>
<td>36.6 km³</td>
</tr>
<tr>
<td>Vol. of drainage basin above sea level for the Trinity and a comparable portion of the Brazos*see Fig. X</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Trinity (at 46,220 km²)</td>
<td></td>
<td>7,348 km³</td>
<td></td>
</tr>
<tr>
<td>Brazos (at 55,392 km²)</td>
<td>12,147 km³</td>
<td>4799 km³</td>
<td>60.5</td>
</tr>
</tbody>
</table>

* at Rosharon, TX  
** at Richmond, TX  
*** at Liberty, TX  
**** at Romayor, TX  
***** at Oakwood, TX

Table 4-1: Drainage basin statistics comparing the Brazos and Trinity fluvial systems.
Onshore Incised Valley and Valley Fill

For the Brazos fluvial system, three datasets were used to document the base of the onshore, MIS 2 incised valley. These datasets include: sediment cores, descriptions of water well cuttings, and land seismic data from an experimental shear wave survey (previously described in Taha and Anderson, submitted-B). Radiocarbon dates (mostly terrestrial gastropods) were used to determine the age of the valley fill and these ages were corrected to calendar years using Calib 5.0 software (Calib 5.0).

A truck mounted hydraulic push core sampler was used to collect several cores within the Brazos valley. Additional cores (piston and split spoon samples) were collected using a tractor mounted drilling rig. The cores were used to interpret the depths to the base of the valley and the depositional environments within the incised valley. More than two hundred descriptions of water well cuttings provided information on lithologic contacts between the base of the lowstand channel sand and the underlying Pliostocene Beaumont clay, or between Holocene aged red clay and underlying blue, brown, or gray Pliostocene clay. The water well descriptions occasionally provided data on the presence of carbonate nodules (indurated horizons). Radiocarbon dates came from shell and wood material extracted from the cores and analyzed at the Woods Hole Oceanographic Institute’s NOSAMS laboratory. Additional radiocarbon dates came from previous studies (Bernard, et al., 1970; Abbott, 2001; Sylvia and Galloway, 2006).

A total of 242 data points (radiocarbon dates; lithologic breaks; the presence of indurated horizons) were converted to X,Y,Z control points and used to determine the
base of the incised valley (Fig. 4-3). Radiocarbon dated horizons were mapped using GIS software (ESRI ArcMAP 9.0; ERDAS Imagine 7.1; Rockworks) and used to calculate the volume of sediment above the MIS 2 incised valley.

Rehkemper (1969) and Smyth (1991) previously used high-resolution marine seismic data, cores, and radiocarbon dates to construct a contour map of the onshore, MIS 2 Trinity incised valley within Galveston Bay. Isopach maps of the basal fluvial and overlying bay sediments were constructed by Rodriguez et al. (2005). In this study, the incised valley contours were digitized using GIS software (ArcMap 9.0) to calculate the valley dimensions and to construct digital isopach maps of the valley fill so the volumes could be calculated.

Longitudinal Alluvial Valley Profiles-Cenozoic Coastal Plain

The difference in long-term sediment yield for both rivers were quantified using GIS software (ArcMap 9.0; ERDAS Imagine 7.1) by subtracting the digital elevation surfaces of their alluvial valleys (obtained from the USGS website) from a common datum (sea-level) to compare the difference in volume. First, a digital Texas geology map (USGS website, 2) was used to extract the outlines of the Brazos and Trinity alluvial valleys along the Cenozoic coastal plain (Fig. 4-1) and calculate their surface areas. Then, the total volume of sediment beneath each alluvial valley surface was calculated above the reference datum, modern sea-level. The volume of sediment was not calculated to determine sediment storage beneath the modern alluvial plain, but to denote the long-term difference in total sediment yield during successive RSL rises. Last, a
Figure 4-3: The 242 data points used to map the onshore Brazos incised valley from the present shoreline to 85 km inland.
longitudinal profile of both alluvial valleys was generated by averaging the elevation for each alluvial surface in 1 km strips perpendicular to the axis of the valley. The longitudinal profile is not intended as a fluvial profile, but as a one-dimensional projection of the two-dimensional surface elevation.

Surface Geology of Drainage Basins

Surface geology maps were obtained by overlaying the outline of each drainage basin (TWDB-Basins) on a digital geology map of Texas (USGS-Geo) using ESRI’s ArcMap 9. software, and then depicting the area of individual geologic formations and grouping them by their geologic period. Each geologic formation contained a reference to its primary and secondary lithologies from the digital geologic map (USGS-Geo). In this study, the primary lithology for each formation was used as its lithologic label (Table 4-2). For consistency, the surface area for each geologic formation, as summarized per geologic period, was normalized and reported as a percentage of the total Trinity drainage basin area. This explains why the percentage of surface exposure for the Brazos drainage basin sums up to 241%.

Cross-Valley Profiles

The average elevation of the longitudinal profiles of both alluvial valleys places a constraint on the realizable subaerial accommodation downstream. But, the elevation of the longitudinal profiles also represents local base level for all entering tributaries. Three methods were used to qualify or quantify the long-term drainage basin response to the elevation of the longitudinal alluvial valley profile. The first method involved
<table>
<thead>
<tr>
<th>Geologic period</th>
<th>Surface Geology</th>
<th>Surface Geology</th>
</tr>
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<tbody>
<tr>
<td></td>
<td>Brazos Drainage Basin as</td>
<td>Trinity Drainage Basin</td>
</tr>
<tr>
<td></td>
<td>% of Trinity Drainage Basin</td>
<td></td>
</tr>
<tr>
<td><strong>Cenozoic</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Coastal Plain</td>
<td>57.3</td>
<td>49.7</td>
</tr>
<tr>
<td>Holocene Sand Dunes</td>
<td>~10.0</td>
<td>0.0</td>
</tr>
<tr>
<td>Pliocene (Southern High Plains)</td>
<td>43.6</td>
<td>0.0</td>
</tr>
<tr>
<td><strong>Late Cretaceous</strong></td>
<td>17.3</td>
<td>28.8</td>
</tr>
<tr>
<td><strong>Early Cretaceous</strong></td>
<td>50.6</td>
<td>15.3</td>
</tr>
<tr>
<td><strong>Late Triassic</strong></td>
<td>6.0</td>
<td>0.0</td>
</tr>
<tr>
<td>mudstone/shale</td>
<td>= 6.0</td>
<td>mudstone/shale</td>
</tr>
<tr>
<td><strong>Permian</strong></td>
<td>40.4</td>
<td>3.5</td>
</tr>
<tr>
<td>mudstone/shale</td>
<td>= 32.1</td>
<td>mudstone/shale</td>
</tr>
<tr>
<td>sandstone</td>
<td>= 7.9</td>
<td>sandstone</td>
</tr>
<tr>
<td>limestone</td>
<td>= 0.4</td>
<td>limestone</td>
</tr>
<tr>
<td><strong>Pennsylvanian</strong></td>
<td>15.7</td>
<td>2.8</td>
</tr>
<tr>
<td>mudstone/shale</td>
<td>= 11.4</td>
<td>mudstone/shale</td>
</tr>
<tr>
<td>sandstone</td>
<td>= 3.5</td>
<td>sandstone</td>
</tr>
<tr>
<td>limestone</td>
<td>= 0.9</td>
<td>limestone</td>
</tr>
<tr>
<td><strong>Total</strong></td>
<td>240.9</td>
<td>100.0</td>
</tr>
</tbody>
</table>

Table 4-2: See reference map in Figure 6. For consistency, the surface geology for the geologic units of both drainage basins are displayed as a percentage of the Trinity drainage basin. The Brazos drainage basin supplies much more sediment than the Trinity drainage basin because of the abundant surface exposures, from east to west, of early Cretaceous, Pennsylvanian, Permian, late Triassic, and Pliocene (Ogallala Group) aged shales and sandstones.
qualitatively documenting the incision of tributaries as they extend away from the common Brazos/Trinity drainage basin divide towards their respective alluvial valleys. Differences in tributary incision and valley denudation were assessed by extracting the elevations for all tributaries from their drainage basin DEMs. A second method for quantifying drainage basin response to the longitudinal profile of the alluvial valley involved producing two cross valley profiles (at 225 km and at 325 km from the coast) to compare the denudation response of each drainage basin to its local stratigraphic base level. The cross valley profiles were also used to observe if older strata, further inland, displayed greater denudation. A third method involved quantifying the three-dimensional denudation of both drainage basins in response to the average elevation along their longitudinal profiles. The volume of sedimentary strata and bedrock was calculated between the surface of the Trinity drainage basin and a reference datum (sea-level). The resulting volume was then compared to the volume calculated between a proportional area of the lower Brazos drainage basin and sea-level.

4.4 Results

The constraints on 'realizable' subaerial accommodation during the recent RSL rise (~19 ka to present) were investigated for the Brazos and Trinity incised valleys from ~100 km offshore to ~50 km onshore. During the most recent RSL fall, these two rivers excavated incised valleys of similar dimensions as the continental shelf was exposed (Fig. 4-4, and Simms et al., in press). Yet, in response to the most recent RSL rise, they have backfilled their incised valleys with significantly different volumes of sediment, particularly fluvial sediment (Figs. 4-2, 4-5, 4-6). Over the short-term (the last RSL rise),
Figure 4-4: A) Interpolated sequence boundary map of the onshore incised valley of the Brazos River. The Brazos incised valley has similar dimensions to the B) interpolated sequence boundary map of the onshore incised valley of the Trinity River (Smyth, 1991; Rodriguez et al., 2005).
Figure 4-5: Isopach maps of the valley fill for the Brazos and Trinity offshore incised valleys. The Brazos incised valley extends from 10 to 110 km offshore and the Trinity incised valley extends from 0 to 100 km offshore.
Figure 4-6: Isopach maps of the valley fill for the Brazos and Trinity onshore incised valleys. The Brazos incised valley extends from 0 to 50 km updip of the shoreline and the Trinity incised valley extends from 0 to 50 km updip of the shoreline, which includes all of Galveston Bay.
this disparity in valley aggradation reflects the difference in their sediment yield that has been established from observing modern records of river gauge sediment yield (Table 4-1). The differences in sediment yield, from the present through the last transgression, are understood in the context of spatial analysis of the surface geology. Over the long-term, these inequalities in valley backfilling from previous cycles of RSL rise are manifested as the differences shown in the longitudinal profiles of both valleys. Ultimately, it is this difference in the longitudinal profiles, related to the long-term sediment yield, which determines the maximum possible upward shift that may occur in stratigraphic base level during any future RSL rise. Presently, both the Brazos and Trinity alluvial valleys display similar gradients but unequal elevations in their longitudinal profiles (Fig. 4-7). The inequalities do not reflect differences in the active depth of fluvial scour, but differences in the elevation of the underlying bedrock upon which the Holocene alluvium and Pleistocene terraces reside. As observed from an elevation map, the surface of the Trinity alluvial valley is lower than the Brazos alluvial valley (Fig. 4-8). Finally, the denudation of the drainage basin is related to the longitudinal profiles of their respective alluvial valleys.

*Offshore: Interpolation of incised valley and volume calculation*

The base of the offshore Brazos incised valley was established by reevaluating the marine seismic sections from Abdullah et al. (2004). Then the data were digitized as X,Y,Z control points, were contoured every 4 m, and were interpolated by using GIS software (ArcMAP). Once the interpolated map was created, ERDAS Imagine was used to subtract this interpolated surface from a digital grid map of the modern bathymetry
Figure 4-7: Longitudinal profiles of the Brazos and Trinity alluvial valleys. These profiles were generated by summing the alluvial valley elevations in 1 km strips, shown on the map. Since the alluvial valleys include Holocene alluvium and preserved Pleistocene terraces, the profiles are not as smooth as expected.
Figure 4-8: An elevation map comparing the along-stripe differences in elevation between the Brazos and Trinity alluvial valleys, with distance from the modern shoreline. For consistency, the Brazos alluvial valley begins along-stripe of the Trinity bay head delta, and both alluvial valleys extend for ~300 km inland.
(USGS-Bathy). Between these two surfaces, a volume of 16.0 km$^3$ of incised valley fill was calculated (Fig. 4-5, Table 4-1). The mapped valley extends approximately 10 - 110 km offshore and narrows along the distal segment. According to Abdullah et al. (2004), the seismic facies characterization and platform borings of the valley fill indicate the sediment is fluvial.

The offshore Trinity incised valley dimensions were interpolated by extracting the individual nodal points from the valley contour lines as $X,Y,Z$ control points and then interpolating. Once the interpolated map was created, ERDAS Imagine was used to subtract this interpolated surface from a digital map of the modern bathymetry (USGS-Bathy) to calculate a volume of 19.9 km$^3$ of incised valley fill offshore (Fig. 4-5). The mapped valley extends from the coast to $\sim$100 km. According to Thomas and Anderson (1994), platform borings indicate that the valley is mostly filled with bay sediments.

**Onshore: Interpolation of incised valley and volume calculation**

Along an 85 km segment of the lower Brazos valley, the $X,Y,Z$ control points from the water well, sediment core, and seismic data (Fig. 4-3) were digitized and contoured every 5 m. After contouring, a total of $\sim$10,500 data points were used as input to generate the interpolated map (Fig. 4-4A) of the onshore MIS 2 Brazos incised valley. The maximum depth of the Brazos valley occurs near the coast and is $\sim$40 m below sea-level. The area of the lower 50 km of the Brazos valley (Fig. 4-4A) was then subtracted from the present topography, obtained from a digital elevation model (USGS-DEM), to
generate an accurate volume of incised-valley fill (Fig. 4-6). Within this area, the volume of sediment sequestered was calculated at 28.6 km$^3$.

The lower 50 km of the onshore Trinity incised valley, landward of the current shoreline, is occupied by Galveston Bay. For comparison, the maximum depth of Trinity valley incision occurs near the shoreline and is $\sim$35 m below sea-level (Smyth, 1991). By digitizing the isopach maps from Smyth’s (1991) study, 12.0 km$^3$ of sediment storage was calculated for the MIS 2 Trinity incised valley. Of this total sediment, 3.0 km$^3$, or 25% is fluvial sediment that occurs at the base of the valley, while the remaining 9.0 km$^3$, or 75%, is fine-grained bayfill sediment that overlies the fluvial sediment.

Comparison of Lower 350 km of the Brazos and Trinity Cenozoic alluvial surface

Presently, the elevation of the Brazos longitudinal alluvial profile is higher than the Trinity longitudinal profile along the entire length of the Cenozoic coastal plain. The difference reflects the cumulative effect of inequalities in their long-term sediment yield during RSL rise when their lower incised valleys backfilled. Analysis of the fluvial geomorphology along the Cenozoic coastal plain (~ the lower 350 km of the Brazos and Trinity valleys) provides an audit of the long-term storage potential of both valleys. This audit provides time averaged analysis of the sediment yield during multiple cycles of RSL rise for these two rivers, which is more useful for understanding the long-term equilibrium of their fluvial valleys.
The surface area of both alluvial valleys is almost identical along an equal length of the Cenozoic coastal plain (~350 km) (Table 4-1). In this case, the surface area of the Brazos valley is 3129.3 km$^2$ while the surface area of the Trinity valley is 3120.4 km$^2$, a less than 0.3% difference (Fig. 4-8). The longitudinal profiles, which sum the average elevation for each alluvial valley in 1 km strips, display similar gradients (Fig. 4-8). However, an offset exists between parallel sections of each profile, with the Trinity alluvial valley typically between 12 and >20 m lower than the Brazos alluvial valley (Fig. 4-8).

The volume of sediment beneath the Brazos and Trinity alluvial valleys was calculated along the Cenozoic coastal plain by using sea-level as a datum (Fig. 4-8). For the Brazos valley this volume is 177.5 km$^3$, and for the Trinity valley this volume is 140.9 km$^3$. The importance of this calculation is not the absolute value of volume of alluvium and bedrock above sea level, but the difference in volume. The difference in volume of 36.6 km$^3$ of sediment fill expresses the cumulative effect of the long-term disparity in sediment yield from one RSL rise to the next. This difference has been preserved as a consequence of progradation of the continental margin. When this difference in volume is divided by surface area, the average difference in elevation equates to the Trinity valley surface being 11.6 m lower.

*Surface geology and sediment yield*

The surface geology of both drainage basins (Table 4-2) is composed of sediments and sedimentary rocks, and both basin exhibit almost identical exposures of
Cenozoic coastal plain strata (~50%, normalized) (Table 4-2). In addition, each drainage basin exhibits a similar pattern of geologic exposures along their axis (Fig. 4-9), with three exceptions. The Brazos drainage basin has a few limited exposures of late Triassic strata, some Holocene sand dunes along its upper reaches, and a substantial portion of Pliocene (Ogallala Group) alluvium exposed along the Brazos headwaters. The Ogallala Group strata were deposited in response to renewed uplift along the Rocky Mountains during this period and are in no way related to the Pliocene aged coastal plain deposits further down dip (Trimble, 1980). Yet, these exceptions do not explain the much higher sediment yield of the present Brazos River.

Much of the difference in sediment yield between the Brazos and Trinity rivers has to do with the widespread exposures of mostly fine-grained, siliciclastic lithologies of the Permian through the Late Pennsylvanian strata of the Brazos drainage basin. The limited exposures of late Triassic strata also contribute to the higher sediment yield of the Brazos River. The single biggest supplier of sediment for the Brazos River comes from these Permo-Triassic red beds (45.8% normalized). These exposures were observed to be responsible for the red color of the discharge of the Brazos River and for the red colored alluvium found along the lower reaches of the valley. In addition, the early to late Cretaceous exposures for both drainage basins contain an abundant amount of fine-grained siliciclastic strata. However, lower surface gradients and denser vegetation cover, suggests that little sediment is eroded from these exposures (Mulder and Syvitski, 1996).
Figure 4-9: A map of the surface geology for the Brazos and Trinity drainage basins. The Brazos drainage basin is outlined with alternating black and white bars while the Trinity drainage basin is outlined by a solid black line. The surface percentage of each geologic unit, broken down by lithology, is detailed in Table 4-2.
In terms of the Permo-Triassic through late Pennsylvanian exposures, the Brazos drainage basin has 10 times as much surface area to draw easily erodible sediment from than does the Trinity drainage basin (62.1% normalized vs. 6.2%). Even if the Cretaceous exposures and the Pliocene Ogallala Group strata are included, the Brazos drainage basin still has more than 4.3 times as much surface exposure from which to draw easily erodible sediment (156.8% normalized vs. 36.2%).

*Brazos and Trinity Drainage Basin Response to Alluvial Valleys (Cross-Valley Profiles)*

The three methods used to gauge the long-term response of the Brazos and Trinity drainage networks to the elevation of their alluvial valleys suggest that the lower the elevation of the longitudinal profile, the greater drainage basin erosion that occurs. The first method qualitatively illustrates that steeper gradients exist along the tributaries of the Trinity drainage basin than along the tributaries of the Brazos drainage basin (Fig. 4-10). The difference in tributary gradients of both rivers are most easily observed by starting at the Brazos/Trinity drainage basin divide, where both drainage basins have the same elevation, and following the change in color gradient towards their respective alluvial valleys. In every case where comparisons are made parallel to the coast, the Trinity tributary gradients are steeper. The same is true, although more difficult to visualize, on the opposite side of each drainage basin (Fig. 4-10).

A second and more quantitative assessment of drainage basin erosion relative to sea-level utilizes two cross-valley profiles (A-A’ and B-B’) (Fig. 4-11). The first profile (Fig. 4-11A) is located 325 km inland. Note that the profile of the Brazos alluvial valley
Figure 4-10: Map of the elevations of all tributaries along the lower Brazos and Trinity drainage basins. The elevations of the tributaries were extracted from a digital elevation model of the drainage basins, which included a buffer zone of 100 m on either side of each tributary to enhance visibility.
Figure 4-11: These two cross-valley profiles of the Brazos and Trinity drainage basins are located in Figure 4-10, and denote the long-term denudation response of the drainage basins relative to the elevations of their alluvial valleys. Since the Trinity alluvial valley is permanently lower than the Brazos alluvial valley along the entire length of the Cenozoic coastal plain, the Trinity tributaries have responded by denuding more of the drainage basin than the Brazos tributaries. A) The profile at 325 km inland occurs above Cretaceous and Paleocene aged strata and indicates that greater tributary incision and denudation of the drainage basin has occurred than in B) the profile at 225 km inland. This second profile occurs above (younger) Eocene aged strata, so denudation of the Trinity drainage basin has had less time to re-equilibrate to the lower elevation of its alluvial valley than the other profile further inland.
is superimposed on the Trinity alluvial valley for better comparison. This profile displays between 20 to >50 m of Trinity drainage basin erosion outside of the alluvial valley, relative to the Brazos drainage basin. The second profile (Fig. 4-11B) is located 225 km inland. Erosion along the Trinity drainage basin at this location is not as acute, but still averages up to 30 m more for most of the width of the Trinity drainage basin. Where the Trinity drainage basin is actually higher relative to the Brazos drainage basin, the Brazos drainage basin is shown to have been eroded by the Navasota tributary, a long, low-gradient, major tributary to the Brazos River (Fig. 4-10).

If the lower elevation along the Trinity alluvial valley has been continuously forcing higher rates of drainage basin denudation relative to the Brazos drainage basin, then presumably this denudation is diachronous. In other words, those portions of the Cenozoic coastal plain drainage basin adjacent to older segments of the Trinity alluvial valley should exhibit a greater difference in denudation than those portions of the Cenozoic coastal plain drainage basin adjacent to younger segments of the Trinity alluvial valley. Indeed, profile A-A' (Figs. 4-10 & 4-11) crosses older strata (late Cretaceous and Paleocene) (Fig. 4-9) and displays greater denudation of the Trinity drainage basin relative to the Brazos drainage basin (Fig. 4-11) than profile B-B' (Figs. 4-10 & 4-11), which crosses younger strata (Eocene) (Fig. 4-9).

The third method provided the most quantitative assessment of the denudational control that a lower elevation of the Trinity longitudinal alluvial profile exerts on its drainage basin, and was determined by calculating the volume of both drainage basins
above the datum of sea-level. This method utilizes the entire Trinity drainage basin and a proportional area of the lower Brazos drainage basin (Fig. 4-12). The one draw back to this method is that it cannot account for underlying structures that may exert some control on the present topography of either drainage basin.

The Trinity drainage basin has considerably less topographic relief than the Brazos drainage basin. Relative to sea-level, the Trinity drainage basin has approximately 60% of the total volume of the Brazos drainage basin. While clearly much of the volume of both drainage basins above sea-level will never be eroded, yet the greater volume and high elevation of the Brazos drainage basin implies it can respond with higher sediment yield in the event of a climate change. The Trinity drainage basin response to such an event would be fairly muted because of the lower elevation (Mulder and Syvitski, 1996).

4.5 Discussion

The original definition for the creation of subaerial accommodation emphasized an upward shift in the equilibrium profile of an alluvial valley driven, most likely, by RSL rise (Posamentier and Vail, 1988). Yet, this definition does not account for potential differences in the fluvial response of adjacent alluvial valleys to the same RSL rise, which requires emphasizing the points along a longitudinal profile to which it is anchored. At any point in time, two anchor points for the longitudinal profile will always exist: the elevation of the updip limit of the alluvial valley and the downdip shoreline or bayline. While the position of the shoreline changes during an RSL cycle, the updip limit
Figure 4-12: An elevation map showing the difference in elevation between the Trinity River drainage basin and an equivalent portion of the lower Brazos River drainage basin. These elevations were subtracted from a reference datum (sea level) to allow calculation of the total volume of the strata and crystalline bedrock of each drainage basin. This comparison of this volume serves as a first order approximation for how much drainage basin denudation has occurred as the tributaries in either drainage basin have responded to the elevation of their respective alluvial valleys.
of an alluvial valley (that extends beyond the coastal plain) should fall outside of the updip influence of RSL (Shanley and McCabe, 1993). Therefore, the updip limit of an alluvial valley will exert control on the downdip limit of subaerial accommodation by controlling the updip anchor point to which the longitudinal profile is attached.

This quantitative case study of the Brazos and Trinity fluvial systems supports the hypothesis that the long-term sediment supply over past RSL cycles sets an upper limit to the present creation of subaerial accommodation at the coast. The long-term sediment supply controls the updip elevation of the longitudinal profile as an anchor point during the most recent RSL rise. This counters the idea, expressed or implied, that the creation of marine accommodation by RSL rise leads to a proportional increase in subaerial accommodation for all adjacent fluvial valleys (Aslan and Blum, 1999; Holbrook et al., 2006; Simms et al., in press). Instead, since the longitudinal profile of an alluvial valley along a prograding margin is maintained by the long-term sediment yield, this longitudinal profile should act to limit the creation of subaerial accommodation during a future RSL rise. The impact of long-term sediment yield on the creation of subaerial accommodation becomes obvious when comparing two fluvial systems that exhibit strong differences in sediment yield. In this case, the fluvial response of the Brazos and Trinity rivers to RSL change is viewed at three physical scales (incised valley, alluvial valley, and drainage basin) and two time scales (recent RSL rise and the Cenozoic). While it is true that these two rivers excavated incised valleys of similar dimensions during RSL fall (Thomas and Anderson, 1994; Abdullah et al., 2004; Simms et al., in press; this study Fig. 4-4), it is their response to RSL rise that ultimately determines
realizable subaerial accommodation. In turn, the subaerial accommodation realized during successive RSL cycles determines the direction of drainage basin evolution.

The basis for understanding those factors that constrain the creation of subaerial accommodation is first shown at the smallest physical and temporal scale. At present, the average water discharge from gauging stations near the coast (Rosharon, TX; Liberty, TX) indicates that both the Trinity and Brazos rivers have almost equal discharge (Table 4-1). Yet, the sediment yield from gauging stations near the coast (Richmond, TX; Romayor, TX; Oakwood, TX) indicate that the sediment yield of the Brazos River exceeds that of the Trinity River by an order of magnitude (Table 4-1), a phenomenon that has probably occurred throughout the Cenozoic. This difference is a function of the larger drainage basin size and the geological exposures of Permian and Triassic aged redbeds in the upper Brazos drainage basin, which have little or no equivalents in the Trinity drainage basin (Fig. 4-9 & Table 4-2). However, as a result of trangressive ravinement, which has removed the upper part of the offshore incised valley fill (Rodriguez et al, 2001), the difference noted in the modern sediment yield of these two rivers is not reflected in the almost equal volumes of fill within their present offshore incised valleys (Fig. 4-5). In particular, erosion has removed fluvial sediment from the overfilled Brazos valley and removed mostly bay and marine sediments from the underfilled Trinity valley (Simms et al., in press). But if the present onshore incised valley fill of both rivers is any guide (Fig. 4-6), the offshore Brazos incised valley would have contained far more sediment then the offshore Trinity incised valley prior to removal by trangressive ravinement.
The disparity in sediment yield between the Brazos and Trinity rivers is better reflected in their onshore incised valley fill. The Brazos River has filled its incised valley with about two and a half times the sediment volume of the Trinity incised valley for an equivalent 50 km segment up dip of the present shoreline (Fig. 4-6). Still, this difference does not indicate that the sediment yield of the Brazos River over the last RSL cycle was an order of magnitude higher. This lower-than-expected volume of fill can be accounted for in two ways. First, when the onshore Trinity incised valley was initially flooding to form Galveston Bay between 10 and 6 ka (Smyth, 1991), the Brazos River was still delivering sediment to two different deltas on the continental shelf (Abdulah et al., 2004). Neither of these paleo-Brazos deltas has been accounted for in the current study. Second, the presence of Trinity valley terraces further inland (Aslan and Blum, 1999) limits effective valley width, impedes deposition of sediment on the floodplain, and funnels most of the Trinity sediment yield to Galveston Bay. So, almost all sediment stored within the lower Trinity incised valley is believed to be contained within Galveston Bay, although Shepard (1953) argued that some sediment bypasses the bay through the Bolivar tidal inlet. In the case of the Brazos incised valley, antecedent topography has been filled and ~ 30 km$^3$ of fluvial sediment has been stored in the onshore incised valley for a segment equal to that of the Trinity onshore incised valley (Galveston Bay).

To understand how the disparity in sediment yield over one RSL rise ultimately exerts control on subaerial accommodation, it is necessary to enlarge the physical and temporal scales of the study. On average, the coastal plain/continental shelf of Texas has prograded throughout the Cenozoic from one RSL cycle to the next (Galloway, 1981;
Galloway et al., 1982; Galloway, 1986). By repeatedly underfilling the Trinity incised valley and overfilling the Brazos incised valley (Simms et al., in press) as the margin prograded, the surface of the Trinity alluvial valley was left lower in elevation along the entire ~350 km wide Cenozoic coastal plain (Fig. 4-8). The lower elevation of the Trinity alluvial valley eventually becomes the established equilibrium profile to which each new incised valley, is cut during RSL fall, will try to aggrade to during the following RSL rise. Regardless of the degree to which the rivers fill their incised valleys, they have maintained similar gradients. Therefore, the cumulative effect of the disparity in valley fill during successive RSL cycles is reflected in the present stratigraphic base level of both alluvial valleys (Fig. 4-7).

There are two important consequences of the lower alluvial surface along the Trinity valley. The first consequence affects the creation of subaerial accommodation at the coast. The updip limit of the Trinity alluvial valley extends beyond the influence of RSL change, and therefore acts as one anchor point or buttress for the equilibrium profile. Since any point along the Trinity alluvial valley is on average lower than a parallel segment of the Brazos alluvial valley, the updip limit of the lower Trinity alluvial valley sets a limit on the amount of subaerial accommodation that can be generated downdip. The other buttress of the alluvial valley has been described as the downdip position of RSL (Holbrook et al., 2006). Even though RSL rise is often seen as the sole forcing mechanism for creating near shore subaerial accommodation in sequence stratigraphic models (Posamentier and Vail, 1988; Aslan and Blum, 1999; Holbrooke et al., 2006), we contend that the longitudinal profile of the alluvial valley exerts the upper limit on how
much subaerial accommodation may be created. So, the definition of near shore subaerial accommodation becomes a function of current RSL rise and the running average of sediment yield during all previous RSL rises. As a result, the Trinity alluvial valley can never create the same amount (volume) of subaerial accommodation as its Brazos valley counterpart during a given RSL rise. However, as a percentage of the realizable subaerial accommodation, the current sediment yield for both rivers has probably backfilled their valleys to the same degree.

The second consequence of the lower surface of the Trinity alluvial valley is to cause greater denudation of its entire drainage basin. Since an alluvial valley acts as the local base level for all its entering tributaries, the lower surface of the Trinity alluvial valley forces each tributary to excavate more sediment to establish its grade (Mackin, 1948). This has a snowball effect on other lower-order tributaries further away from the alluvial valley. The end result is somewhat counterintuitive in that the higher denudation rates for the entire Trinity drainage basin relative to the Brazos drainage basin are ultimately linked to the lower sediment yield of the Trinity River during RSL rise. But the higher denudation rates of the Trinity drainage basin are an ongoing process during both RSL rise and RSL fall, and hence the sediment contribution from the additional denudation probably does not contribute much additional sediment yield during RSL rise.

Our efforts to quantify the effect of the alluvial valley surface on drainage basin denudation are illustrated qualitatively using figures 4-10 and 4-11. At the divide of the Trinity and Brazos drainage basins, all tributaries are at the same elevation. However, as
these tributaries flow towards their respective alluvial valleys, the elevation gradient is
greater for almost any tributary of the Trinity drainage basin versus the Brazos drainage
basin (Fig. 4-10). This observation is supported by two strike-oriented profiles of the
Trinity and Brazos drainage basins at 225 km and 325 km inland (Fig. 4-11). In both
cases, the alluvial valleys are superimposed above each other so that the relative elevation
of their drainage basins on either side can be compared. The Trinity drainage basin is
lower in both examples. As expected, the tributaries have denuded more of the Trinity
drainage basin further inland (325 km) than closer to the coast (225 km) because of the
greater duration over which these tributaries have responded to a lower local base level.

The volume of sediment or rock beneath the surface of the entire Trinity drainage
basin was calculated above a datum. The same procedure was performed for an ~ equal
portion of the Brazos drainage basin above the same datum (Fig. 4-12). This datum was
taken as modern sea-level, or geomorphic base level (Shanley and McCabe, 1994). The
results are summarized in Table 4-1. If sea-level is used as a datum, the Trinity drainage
basin has ~40 % less volume of sediment and bedrock beneath its surface than a
proportional area of the lower Brazos drainage basin. Even if their surface areas are
normalized, the Trinity drainage basin is ~28% less. However, one caveat exists. This
method does not account for the extent to which underlying structure exerts control over
present differences in elevation between the two drainage basins, such as the differential
offset between the Elkhart-Mount Enterprise Fault System and the Mexia-Talco Fault
Zone (Fig. 4-9) (Owen and Jackson, 1981; Woodruff and Foley, 1985).
For the present discussion we assert that the drainage basins of the Brazos and Trinity rivers appear to evolve in a direction that maintains their sediment yield from one RSL cycle to the next. In the absence of tectonic uplift, climate change can lead to an increase in the sediment yield of a river (Schumm, 1993). Yet, for two adjacent drainage basins, the effect of climate change on sediment yield will be ameliorated for that drainage basin with an overall lower gradient. Recent studies have shown that both modern and Holocene climatic oscillations have occurred and led to higher than usual sediment yields along the Brazos River and adjacent Colorado River (Blum, 1993; Fraticelli, 2003), and can be expected to affect the sediment yield of the Trinity River as well. Therefore, the inevitable outcome of the greater denudation of the Trinity drainage basin in response to its lower alluvial valley is to limit the ability of the Trinity drainage basin to respond to climatic oscillations, which helps to maintain its lower sediment yield.

While Simms et al. (in press) captured one aspect of the disparity in sediment yield between the Brazos and Trinity during RSL rise, caution should be used when applying their use of the terms underfilled and overfilled incised valleys. One implication of the term underfilled incised valley is that this longitudinal profile has yet to attain the same degree of equilibrium as an overfilled incised valley. The results of this study indicate that the Brazos and Trinity valley fill have attained nearly similar equilibriums relative to their updip longitudinal profile anchor points, even though the Brazos incised valley has clearly experienced greater aggradation than the Trinity incised
valley (Fig. 4-2). Perhaps a better term would be a ‘geomorphically underfilled’ versus a
‘geomorphically overfilled’ incised valley.

In summary, drainage basins along prograding passive continental margins evolve
in response to the initial conditions of their sediment yield and continue to evolve over
multiple RSL cycles in a direction that helps them to maintain those initial conditions. In
the case of low sediment yield fluvial systems, such as the Trinity River, the low
sediment yield prevents the incised valley cut during lowstand from completely
backfilling during RSL rise. In turn, the drainage basin evolves in a direction that makes
it difficult to ever increase its sediment yield and, hence, ‘catch up’ to the overfilled
Brazos incised valley. Mackin (1948) came to a similar conclusion, but for a much
smaller part of the fluvial system. He stated that when a stress operates on a graded river
(i.e. a river with a longitudinal profile that is in equilibrium) it will displace its
equilibrium in such a way as to absorb the stress. For a longitudinal profile, he indicated
that when discharge or sediment yield changes, a river will locally erode or deposit its
sediment in order to maintain the gradient necessary to continue flowing smoothly.

We have detailed how an entire drainage basin responds to the conditions along
its alluvial valley, which in turn are initially set by the sediment yield during successive
RSL rises. Therefore, it appears that a process of self-similarity is at work and that what
is observed at the local level, along a segment of the river bed, actually occurs at the
drainage basin scale. Unless forcing mechanisms, such as tectonic uplift or the unroofing
of a source of easily erodible strata occurs, a drainage basin with initial low sediment
yield will evolve in a direction so as to maintain its low sediment yield in the future. Neither tectonic uplift nor major changes in surface geology is believed to have occurred for the Trinity drainage basin during the Cenozoic.

4.6 Conclusions

1. The Brazos and Trinity rivers excavated incised valleys of almost the same dimensions during the last relative sea-level (RSL) fall. As a result of differences in the extent of easily erodible sedimentary strata within their different drainage basins, the Trinity River yields significantly less sediment than the Brazos River. This difference in sediment yield led to a large discrepancy in the volume of incised valley fill during the most recent RSL rise. This pattern has been repeated during successive cycles of RSL rise throughout the Cenozoic. The result is that the Trinity alluvia valley has been left permanently lower in elevation relative to the Brazos valley. The lower elevation of the Trinity alluvial valley, in turn, exerts control over the creation of new subaerial accommodation at the coast during each new RSL rise.

2. A secondary consequence of the lower elevation along the entire 350 km length of the Trinity alluvial valley, relative to the Brazos alluvial valley, is to force greater long-term denudation of the Trinity drainage basin relative to the Brazos drainage basin. This denudation is a continuous process, and therefore, adds additional sediment during both RSL rise and RSL fall. Apparently the effect of this additional sediment during RSL rise is not enough to offset the lower sediment yield of the Trinity River relative to the Brazos River. However, over time it does
have a pronounced effect on the evolution of the elevation of the entire Trinity drainage basin.
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APPENDIX

Appendix A: Experimental Shear Wave Seismic Survey

A.0 Introduction

Conventional land seismic data acquired for academic research is cheaper to collect than sediment cores and provides a 2-D perspective of depositional environments unrivaled by the acquisition of sediment cores. Yet, the speed of acquisition, the number of people and the expertise required to operate the system, and the time-intensive processing of the data afterwards can counter the benefits of acquiring land seismic data for a small stratigraphic research program. In lieu of the extra time, labor, and processing required for a conventional high frequency, 2-D land seismic survey, where geophones are planted each time in a CDP roll-along (Pugin, et al., 2004), an experimental 2-D land seismic streamer survey was conducted to see how efficiently it could image the base of the Brazos incised valley and characterize the valley fill.

The land seismic ‘streamer’ that was developed for this project could be towed behind a truck to simulate the ease of acquisition usually reserved for marine seismic surveys, and required as few as two people with minimal expertise to operate. High frequency, horizontal shear-waves (SH-waves) were recorded instead of P-waves to avoid imaging the ground water table as a reflector (Pugin et al., 2003). The velocities of the recorded SH-waves were also much slower than P-wave velocities, providing higher resolution because of the shorter wave lengths (Suyama et al., 1988). The survey was conducted using a seismic streamer consisting of 12 channels strung together in series
with 1.5 m spacing. Each shot station was spaced 1.5 m apart to provide the tight CMP spacing necessary for an SH-wave survey (Pugin et al., 2003).

An outcrop scale, seismic section of unconsolidated alluvium was then generated with this acquisition system and was successful in imaging the base of the incised valley. This technique was also successful in identifying the lithologies of the valley fill using a combination of seismic characterization of the data (with results from Abdulah et al., 2004-offshore Brazos valley) and correlation of the seismic section with the lithologies provided by a sediment core collected in close proximity. A corrected radiocarbon age (~20 ka) from a channel sand in the sediment core, which was acquired just above the base of the incised valley, identified this channel sand as belonging to the OI Stage 2 lowstand. The sediment core also confirmed the lithology at the base of the incised valley by penetrating reddish-brown clay (Pleistocene age Beaumont Formation) just beneath the channel sand.

A.1 Objectives and Background

One of the objectives of this experiment was to determine the boundary conditions (depth of incision, valley terraces) above which the Stage 2 lowstand, transgressive, and early highstand alluvial valley fill reside. Another objective of this study was to determining if shallow fluvial sand aquifers, such as lowstand channels, could be detected within the Brazos incised valley using seismic characterization. This application of the seismic data could potentially provide a greater predictive approach to
ground water prospecting. Outcrop scale, high resolution seismic data was needed to address both these issues.

The use of seismic data is critical towards understanding the evolution of incised valley development. It can allow a study to circumvent the need for multiple cores and radiocarbon dates by utilizing fewer cores and dates to ground-truth the seismic data, and relying on seismic horizon continuity and seismic characterization to detail the shape of the incised valley, including valley terraces, and its lithologic fill.

Along the Texas and western Louisiana coasts, underfilled incised valleys (Simms et al, in press) are currently flooded to form bays, such as Galveston Bay. The bay portion of several underfilled incised valleys in Texas were studied early on, including the Trinity (Rehkemper, 1969; Smyth, 1991), Lavaca (Byrne, 1975; Wright, 1980; Maddox, 2005), and Nueces (Wright, 1980; Durbin et al., 1997; Simms, 2005) valleys. These incised valleys were easier to document than overfilled incised valleys (Simms et al., in press) because of the rapid acquisition of marine seismic data and cores. For instance, the lower 50 km of the Trinity valley, updip of the present barrier island coastline, is currently submerged to form Galveston Bay (Smyth, 1991). The ease of acquiring high resolution seismic data in this marine environment, as well as the ease of acquiring cores at desired locations without permits, allowed many of the details of the Stage 2 sequence boundary and fluvial and bayfill above this boundary to be worked out over the last several decades.
The Brazos incised valley and the adjacent Colorado incised valley are fluvially overfilled (Simms et al., 2005) with shoreline deltas (Rodriguez et al., 2000), and therefore are not currently flooded to form bays. As a result of the labor intensive efforts and slower acquisition rates for acquiring land seismic data, as well as greater restrictions on locations for collecting land cores, studies oriented at documenting the base of the Brazos and Colorado onshore incised valleys have been neglected in favor or their underfilled counterparts.

Much of this study of the Brazos valley (Chapters 2, 3, and 4) had to rely upon the use of lithostratigraphy and chronostratigraphy (core data and radiocarbon dates) to document the depth to, and aggradational fill of, the Brazos incised valley. Three marine seismic surveys were previously collected to detail the offshore portion of the Brazos incised valley. The first was in 1993 using the R/V Lone Star along the intracoastal canal, and the second and third were in December of 2002 and March of 2003 using the R/V Trinity, when marine seismic surveys were again shot along the intracoastal canal of Texas near the Brazos delta. It is believed that energy dispersion occurred due to the excavated intracoastal canal walls, which degraded the signal to noise ratio and considerably reduced the quality of the seismic data.

Therefore, a second method of seismic acquisition was devised, although it did involve land seismic acquisition, which is traditionally much more labor intensive and slower to acquire than marine seismic data. It was important to investigate if an alternative could be found to the labor intensive, traditional land seismic acquisition
system where geophones are repeatedly planted in the ground and receiver cables moved along. One of the criteria for this study was a system that required minimal setup, could acquire data rapidly, and could simplify the shot/geophone (or source/receiver) geometry so the data could be processed more quickly. In addition, because of the shallow nature of the ground water table and the near-surface depth of geologic interest, a shear wave seismic survey was necessary because shear waves are impervious to interstitial fluids, and thus the ground water table would not show up as a major reflector in the same fashion as it would in a P-wave survey. One thing this experiment was not meant to investigate were the technical geophysical details behind a SH-wave seismic survey. Instead we chose to apply a previous experimental design, tested and used by Pugin, et al. (2003, 2004) for an investigation of the subsurface extent of glacial tills in Illinois, to see if this technique could be adapted to our investigation of the Brazos incised valley.

A.2 Location of Survey

The objective of this study called for the acquisition of at least one seismic line that spanned the width of the Brazos valley, several 10's of km updip of the coastline. The location chosen was a 6 km stretch of road along FM 1462 between the Oyster Creek and the Brazos River, ~50 km southeast of downtown Houston. The location of the survey was chosen because a core collected in close proximity would allow lithostratigraphic, chronostratigraphic, and seismic velocity ground truth of the seismic data. In addition, previous results from water well cuttings available near by indicated that the study area was situated over the deepest portion of the incised valley. If successful, the seismic data could be combined with the sediment core and the
Figure A-1: Location of SH-wave seismic survey along FM 1462 near Rosharon, TX
Figure A-2: Core BV-04-06, acquired near the SH-seismic survey (see Fig. A-1 for location)
descriptions of water well cuttings to produce a cross section from valley wall to valley wall, approximately 13 km wide.

Initially it was determined that the survey would span at least half the width of the valley. But as the survey progressed and the kinks in this new acquisition system were worked out, logistics and time constraints only allowed for the acquisition of 400 meters of data ( ). However, this was enough to assess its ease of acquisition and usefulness as an effective tool for determining the depth to the base of the incised valley and the interpretation of valley fill lithologies.

A.3 Methods

The first patent on a land seismic streamer data acquisition system the author is aware of dates from 1975 (Kruppen & Bedenbender, 1975). However, it was not until the 1990’s that institutions began experimenting with and applying this new technique of seismic acquisition in earnest.

Several European institutions, as well as one Japanese company, lead the way with this technique, streamlining the process of acquiring data (Suyama et al., 1988; Eiken, et al., 1989; Steeples & Miller, 1990; Van der Veen & Green, 1998; Inazaki, 1999; Van der Veen et al., 2001). Dr. Andre Pugin, of the Illinois State Geologic Survey, processed high resolution seismic data from marine surveys on Lake Geneva, Switzerland and provided the impetus for the use of the technique (Moren, et al., 2002; Pugin, et al., 2003; 2004). Dr. Pugin and Timothy Larson, also of the Illinois State Geologic Survey,
were kind enough to provide the details for building their land seismic streamer for academic purposes (personal communication and site visit).

The seismic streamer was developed as a 12 channel system composed of 24 (14 Hz) shear wave geophones. Each geophone station was thus composed of dual shear-wave geophones connected in series, with the axes of the geophone coils oriented 180 degrees to each other, and mounted to 25 cm steel-runner sleds. As a result of the geophone coils being oriented 180 degrees to each other and being connected in series, the signal from the each geophone pair was wired so the polarity from the first geophone was reversed as it was merged with the second geophone ( ). This detail ensured that any P-wave component was almost completely cancelled and the SH-wave signal was enhanced (Larson, ISGS, personal communication). The 'streamer' then consisted of 12 of these SH-wave geophone stations that were strung together, in series, with a 1.5 m spacing between each, using non-stretch nylon rope. This land seismic 'streamer' was towed behind a truck. The data was collected every 1.5 m, so the CDP interval was spaced every 0.75 m. As stated above, this experiment was not meant to test the geophysical parameters necessary for optimizing data acquisition. Both the spacing between channels and the spacing between shots, as well as the number of channels needed for rapid but high resolution seismic data, were previously optimized by the Illinois State Geologic Survey (Pugin, et al., 2003; 2004). The only experimentation performed in this survey was to rewire the geophone couplets in a way that reduced the output wires from two sets to one set ( ), to use an aluminum instead of a steel
Figure A-3: Schematic for how to rewire two shear-wave geophones in-series. This will reduce the total number of wires for the geophone couple from four to two and provide a better and more reliable attachment to the seismic receiver cable.
strike plate for the source, and to devise a better mechanism to initiate recording once the source was generated.

The source consisted of a 3 lb hammer attached to a 4-foot handle and struck, horizontal to the ground and perpendicular to the geophone streamer, against a 4 inch by 6 inch by 5 foot wooden beam which had a welded U-shaped aluminum plate wrapped around one end and bolted through the wooden beam in two locations. This wooden beam was supported beneath the tire of the truck to brace it to the ground. The wooden beam provided good coupling of the source to the earth and the use of an aluminum plate, tested against a more sturdy steel plate, reduced the ringing from the source. Each shot point was stacked three times to reduce the signal to noise ratio.

Experimentation with the seismic acquisition system on Rice University campus determined that the aluminum plate did not require a powerful blow from the 3-lb hammer to provide the necessary acoustic impulse to image 30 to 40 meters deep. This knowledge helped to extend the life of the welded three piece U-shaped aluminum plate. However, we do not recommend using an aluminum strike plate if it is welded because ours broke easily along its welded seam by the end of the survey.

Another small modification to this system was to replace the traditional hammer switch with a conductive switch as the mechanism for synchronizing the recording of the data with impulse from the source. On an earlier test, the hammer switch proved unreliable for activating the recording system each time, and required a heavy blow of the
hammer to activate the hammer switch than was seismically necessary. In addition, the hammer switch got broken in the process, increasing the cost of the survey by ~50% above initial materials and labor cost. So instead, a simple and cheap makeshift switch was constructed to facilitate rapid hammer strokes and data recording. This was the critical modification because thousands of hammer strokes were required along the 400 meter stretch where data was collected. The makeshift technique involved two wires connected to the Smart Seis Geometrics™ recorder and attached at opposite ends to the source hammer and a separate metal rod. When the metal rod was place against the aluminum strike plate attached to the wooden beam and the hammer was struck the aluminum plate, the circuit was closed and the Smart Seis Geometrics™ seismograph recording system began recording the acoustic signal felt by the receivers at the proper time.

Corrections due to topography (elevation statics) were not deemed necessary because the ground was horizontal over the short distance of the survey. But a longer survey, where the end points of seismic acquisition are beyond eyesight, would require a topographic survey to determine if elevation statics corrections needed to be applied to the data.

One of the constraints was to build an inexpensive streamer ( ). The total cost of the materials and labor came to less than $400 (not including the broken hammer switch). To buy a manufactured system was estimated to cost several thousand dollars.
Figure A-4: SH-wave seismic streamer towed behind a vehicle for rapid acquisition. Best results came from use on an asphalt road.
A.4 Data Acquisition

The data was collected during July of 2004. Pugin et al. (2004) indicate that it was essential to maintain a short spacing between geophones and short offset between the source and receivers for the SH-wave survey, otherwise surface noise causes too much interference.

The survey was recorded with a Smart Seis Geometrics™ 12 Channel acquisition system. As indicated above, a 3 lb hammer was struck, horizontal to the ground against an aluminum capped wooden beam that was weighted down by the back tire of the truck to which the seismic streamer was attached and the Smart Seis Geometrics™ acquisition system was mounted on. The offset between the first geophone, closest to the truck, and the source was 3 m. The shot spacing was 1.5 m. Two of the aluminum capped beams were attached at the ends by a 1.5 m rope so that as one was being used for the source, the second one could be placed in front for the truck to roll over next. This enhanced the rate of acquisition and ensured accurate shot spacing. Three hammer blows were used at each shot station, and the signal was stacked to increase the signal to noise ratio.

Positioning of the beginning and end of the survey was documented using a GPS unit, although the location could just as easily be identified on an aerial-photo map. No navigation data was recorded along with the survey because, as long as the shot spacing was rigorously adhered to, the geometry file used while processing the data would
spatially orient the shot gathers correctly and allow the data to be processed accurately. The maximum fold of the data was 6 CDP.

The seismic streamer was tied to both sides of the back of the truck. This kept the streamer array taut at all times and kept it centered directly behind the truck. Since the source was generated slightly offset from the center line of the receivers, this will have introduced a slight error in the geometry, but it was believed to be within the limits of the resolution of the survey.

A.5 Data Processing

After the seismic data was recorded on the Smart Seis Geometrics™ 12 channel seismograph, it was downloaded to a personal computer. The data had been recorded in a format on the Geometrics seismograph called Seg2. These Seg2 files had to be converted into Segy files before being processed on ProMAX™ seismic processing software on a Unix station. The simplest solution was to download a free trail software package from the internet (IXSeg2Segy) called IXSeg2Segy for conversion of Seg2 to Segy format. It was important to save the Segy data as:

ASCII
Most Significant Digit First
32 bit Floating Point
Then the data was transferred to a Unix machine so it could be processed using ProMAX™ software. ProMAX™ software offers more robust processing capabilities than downloadable software packages from the internet.

All the shot gathers had to be first merged. Then a separate job flow was needed to create the geometry file. A Marine Geometry Load program was used instead of a Land Geometry Load because it greatly simplified the process. As stated above, the data was acquired by towing a streamer behind the location where the source was generated instead of generating a split spread survey as is typical with most land seismic surveys. The geometry file could therefore be treated as a marine survey with the source out front and the receivers trailing behind.

Butterworth bandpass filters with several different frequency ranges were run and then observed for their effect on the stacked data. After experimenting with 100 to 200 Hz and 50 to 100 Hz it was finally determined that the majority of the useful frequencies resided between 20 and 50 Hz when viewed on the stacked data. This agrees well with Pugin et al., (2004), who had indicated that for an SH-wave survey with our configuration of shot and receiver spacing, the useful frequency range was between 20 and 40 Hz. Therefore, using velocities between 150 and 200 meters/sec (derived from processing tests and Pugen et al., 2004) and a frequency range of 20 to 40 Hz, the \( \frac{1}{4} \lambda \) tuning frequency of the data provides a resolution of 1 and 3 meters. As a result of the low frequency range, the application of a 60Hz notch filter was not necessary to remove
electrical noise because the Butterworth bandpass filter cut out before reaching this frequency range.

The data was finally CMP stacked with an AGC and the Butterworth bandpass filter values applied. The CMP stacking allowed testing of different velocities values to observe which one would produce the most coherent seismic section. Final values of 160 m/sec were used and this produced the best stacked image, which were almost identical to the average velocity of 156 m/s, determined from the sediment core. Hence, each 100 ms (in two-way travel time) represents 8 m and the SH-wave velocity was about 1/10 the velocity of a conventional P-wave survey on unconsolidated sediment, similar to the results of Pugin et al. (2003). No migration was run on this seismic data. Approximately 3600 traces were processed in the dataset. When converted into metric units, the seismic line was about 400 meters long. The displays shown below were cropped after being produced using the 'Snapshot' tool on a standard Unix machine. An uninterpreted CMP stacked seismic section of the survey is shown in [image] and an interpreted CMP stacked seismic section is shown in [image].

A.6 Results

Mitchum et al, (1977, part 6) defined a variety of seismic facies units (SFUs) based upon both internal reflector configuration and external form. Abdullah et al. (2004) utilized these SFUs in their high-resolution marine seismic dataset to describe and interpret the Brazos fluvial/deltaic environments along the continental shelf. The descriptions of environments of deposition from SFUs were noted as being more limited
Figure A-5: Shear-wave seismic cross-section. A) Uninterpreted seismic data. B) Interpreted seismic data. Reflector A denotes base of Brazos incised valley; Reflector Package 1 are chaotic reflectors, which correlate to channel sand deposits in core BV-04-06; Reflector Package 2 are horizontal to sub-horizontal reflectors, which correlate to floodplain deposits in core BV-04-06 (location Fig. A-1, lithology Fig. A-2)
within the fluvial fill of the Brazos incised valley. In fact, Abdulah et al. (2004) defined just two SFU units for the Brazos incised valley fill, SFU 7a (correlated to a sandy lithology) and SFU 7b (correlated to a clay rich lithology). Similarly, only two seismic facies, or reflector packages, were identified within the onshore Brazos incised valley fill.

The two reflector packages of the Brazos incised valley reside above a strong amplitude seismic reflector, found between 200 and 320 ms (or 16-25 m below ground surface). The strong amplitude seismic reflector is interpreted to be the base of the Brazos incised valley (Reflector A). The greatest depth of incision agrees well with data from core BV-04-06, as this core penetrated the valley base at 25 m below ground surface and cut into Pliocene clay. The irregular surface of this strong amplitude seismic reflector strongly suggests it was cut while the valley was being excavated by fluvial incision.

Above the prominent seismic reflector are two seismic reflection packages. The first reflection package (Reflector Package 1) is characterized by low amplitude, chaotic reflectors. This seismic characterization is similar to Abdulah et al's (2004) chaotic SFU 7a (a medium to coarse sand) and corresponds to fluvial sands, based on lithologic correlation with Core BV-04-06. This reflector package can be seen on the east end of the uninterpreted seismic line. The second reflection (Reflector Package 2) package is characterized by medium amplitude, horizontal to sub-horizontal reflectors. This seismic characterization is similar to Abdulah et al’s (2004) laminated
onlapping reflectors in SFU 7b and corresponds to floodplain deposits based on lithologic correlation with Core BV-04-06. This reflector package appears to onlap the antecedent topography of the basal reflector (Reflector A).

Using the radiocarbon constraint just above Reflector A, from core BV-04-06 ( ), the depth to the incised valley was determined to be about 25 m. The seismic velocities expected for this survey demarcated a sequence boundary close to the 25 m depth on the east side of the seismic section. In fact, this is where the core is interpreted to project if it superimposed on the seismic section ( ).

As a consequence of underestimating the energy requirements needed to power the equipment, (i.e. bringing too few car batteries into the field), as well as a time constraint on the free use of the rented 12 Channel Geometrics™ recording equipment, the final length of the seismic line was considerably shorter than desired. However the 400 meters of processed data ( ) clearly achieved the intended objectives of rapid acquisition with a minimum of labor.

A.7 Conclusions and Recommendations

Using the shear wave spectrum of seismic data has great benefit for high-resolution coastal and alluvial plain investigations. For the Brazos alluvium a direct correlation has been made between the chaotic reflectors of the seismic data and lithologies of sand or gravel. The same is true for a correlation between the laminar reflectors of the seismic data and lithologies of clay-rich, mostly floodplain deposits. The
SH-wave velocities proved to be an order of magnitude slower than their P-wave counterpart for this unconsolidated alluvium. Hence, the SH-wave seismic technique provides a better high-resolution seismic section with minimal source input and minimal processing.

For our purposes, the technique was much less labor intensive than a traditional land seismic survey would require. Between two and three people were used during the data collection, with only one person required to do field labor at a time, involving moving the aluminum capped wooden beams forward 5 feet for each shot station and striking the aluminum plate moderately hard three times to produced the shot/gather stacked data. As a result of limited time for acquisition, the process was not as streamlined as it should have been. In the future, more car batteries are required to be on supply. Also, additional aluminum capped wooden source beams are required incase the welded aluminum plate cracks again along the weld seam. An even better solution would be to cut an aluminum cap from a single block of aluminum so no weld seams exist. In addition, it would significantly expedite the process to not collect the data in the middle of a humid East Texas summer. This was a serious draw back which hampered the operation more than initially perceived. We recommend starting the acquisition in mid-May for east Texas. And finally, choose roads that do not have much traffic, as seismic interference from large tractor trailers were observed disturbing the seismic equipment from as far as half a mile away.