Morphodynamic modeling of channel fill and avulsion timescales during early Holocene transgression using Trinity River, TX incised valley stratigraphy

by

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ABSTRACT

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The Trinity River and the sediments that infill its incised valley system are well-constrained in terms of time and space properties of the deposits and the resulting stratigraphy. The Trinity River is therefore an excellent natural laboratory to test fluvial morphodynamic concepts in order to examine the processes of incised valley infill. We develop a numerical model that links sediment transport and deposition to the production of stratigraphy, and evaluate the effects of Holocene transgression on the development of Trinity incised valley stratigraphy. We simulate the mechanics of channel fill and avulsions for the Trinity River, by coupling fluid flow, sediment transport, and channel response, constrained by modern and early Holocene conditions. Our results show that non-uniform flow produces loci of backstepping sediment deposits that keep pace with base level rise, which corresponds to measured data within Galveston Bay. Sediment deposits also coincide with channel avulsion locations. An important parameter is the rate of base level rise: Deposits initially prograde then retreat upstream, with ever-increasing rates of base level rise (transgression). Additionally, the analyses presented herein show that including a floodplain parameter within the model framework is necessary to calculate the time to avulsion, by changing the amount of sediment deposited within the channel. Our model is applied over century to millennial timescales and is utilized to evaluate basin scale patterns of known stratigraphic variability. Because the model is well-constrained, our results
have application for predicting stratigraphy for other fluvial-deltaic systems undergoing transgression. This is especially important for predicting the valley infill of systems that lack the robust stratigraphic constraints provided by the Trinity incised valley system.
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NOTATION

Variables are defined in alphabetical order. Units for each variable are defined where L is length, T is time, and M is mass.

A: cross-sectional channel area, $L^2$
B: bankfull flow width, L, 500 m
Bf: floodplain width, L, 7.2 km
BL: base level rise rate, L T$^{-1}$
Cf: friction coefficient, unitless
D: median grain diameter, L, 250 µm
\( \eta \): bed elevation from a fixed datum, L
Fr: Froude number, unitless
\( g \): acceleration due to gravity, L T$^{-1}$
H: bankfull flow depth, L
h: water surface elevation plus bed elevation (\( \eta \)), L
i: intermittency, unitless, 0.05
\( \lambda_p \): bed porosity, 0.4
Lb: characteristic backwater length scale, L
\( \Omega \): sinuosity, unitless, 1.86
\( q \): bed material load per unit channel width, $L^2$
Qb: sediment discharge, $L^3 T^{-1}$
Qw: water discharge, $L^3 T^{-1}$
\( \rho \): water density, ML$^{-3}$
R: submerged specific gravity of sediment, unitless
S: channel bed slope, unitless, 1.6 $\times$ 10$^{-4}$
\( \tau_b \): total boundary shear stress, ML$^{-2}$ T$^{-1}$
t: time, T
\( \tau_A \): avulsion timescale, time between avulsions, T
\( \tau_f \): thickness of the floodplain, L
U: flow velocity, LT$^{-1}$
\( V_A \): vertical aggradation rate, L T$^{-1}$
x: space, L
\( x_r \): straight, down valley distance, L.
INTRODUCTION

Morphodynamic models of fluvial systems are used to evaluate sediment transport and channel evolution processes as coupled to fluid flow conditions. Recently, morphodynamic models have modified the assumption of uniform flow conditions to explore how non-uniform, 'backwater' flow influences sediment transport rates, channel filling, and channel avulsion frequency and location for rivers nearing receiving basins [Parker, 2004; Nittrouer et al., 2012; Chatanantavet et al., 2012; Lamb et al., 2012; Ganti et al., 2014]. Many of these models have also used stable boundary conditions, including constant base level, and have focused on shorter, decadal- to century-scale responses of the fluvial systems. Additionally, these models have neglected interactions with the adjacent floodplain. For example, the Mississippi River has been a focus of prior studies that have estimated the time to avulsion for the late Holocene (i.e., the last 3,000-4,000 years). This has been a period of relatively stable eustatic sea level and so neglecting the influence of rapidly changing base level conditions may be an appropriate assumption. Furthermore, interactions with the channel floodplain are not explicitly modeled in previous studies, because the time scale associated with channel migration and base level rise is outside of the scope of interest.

Here, we develop a numerical model that considers a fluvial-deltaic system under transgression. The broad aim is to identify patterns of stratigraphy that develop over century to millennial timescales. Our working hypothesis is that because the hydrodynamics of any fluvial system near the receiving basin are sensitive to base level changes, sediment transport conditions will change over time and space for conditions of base level adjustment. We base our morphodynamic model on the Trinity River, Texas (Figure 1) because its incised valley stratigraphy is well-constrained for the latest transgression (10,000 years B.P.). Our aim is to
enhance the understanding for the mechanisms of incised valley fill by capturing the dynamic boundary conditions of the latest Holocene transgression.

The Trinity incised valley captures a nearly complete stratigraphic record of the early Holocene transgression. Without an incised valley, a limited record of fluvial-deltaic response to transgression is recorded because of transgressive ravinement [Thomas and Anderson, 1988]. Fortunately, for the Holocene Trinity River and its incised valley system, previous studies have used seismic and core data to constrain its infill in space and time and so, the measured the stratigraphy in this system provides constraints on the processes that build stratigraphy and produce incised valley infill [Anderson et al., 1992; Thomas and Anderson, 1994; Rodriguez et al., 2005; Simms et al., 2006; Anderson et al., 2008]. Additionally, the local sea level conditions during the Holocene are well-constrained [Milliken et al., 2008]. Therefore, these studies provide the data required to validate model results. In this capacity, it is possible to use a morphodynamic model to examine the responses of the Trinity system to spatiotemporally varying backwater hydrodynamics, sediment transport conditions, and avulsions, based on known Holocene conditions. We aim for our model to be exportable to other fluvial-deltaic systems using input parameters that are constrainable, so that it is possible to produce informed estimates of incised valley stratigraphy.
BACKGROUND

Trinity River and Galveston Bay

Incised valleys, such as those found in the northern Gulf of Mexico, provide prime data to constrain a morphodynamic model that describes transgressive fluvial sedimentation because the deposits within these valleys are relatively unaffected by transgressive ravinement, and therefore provide as complete of a stratigraphic record that can be expected from transgressive events [Thomas and Anderson 1988]. The Trinity River (Figure 1) incised valley system was generated during the Stage 2 sea level lowstand of approximately 22-17 ka and is approximately 30-40 m deep (Figures 2-3) [Thomas and Anderson, 1994; Anderson et al., 1996; Anderson et al., 2004]. The base of the valley is interpreted as the Stage 2 sequence boundary which extends across the Gulf of Mexico continental shelf (Figures 2-3). The Trinity system has since responded to rapid Holocene sea level rise by aggrading and infilling its valley and modern Galveston Bay, beginning approximately 17 ka [Anderson et al., 2008]. The shoreline is approximated to have reached Galveston Bay and influenced fluvial-deltaic sedimentation by ~10 ka based on sea level rise rates and the slope of the continental shelf.

The present day Trinity incised valley system is considered to be underfilled, based on the existence of Galveston Bay and the exposure of Deweyville terraces upstream of Galveston Bay (Figure 1) [Failing, 1969; Anderson et al., 1992; Thomas and Anderson, 1994; Blum et al., 1995; Simms et al., 2006]. The underfilled character of the Trinity incised valley implies that sediment accumulation is unable to keep pace with the rate of base level accommodation created by rising sea level [Simms et al., 2006]. Due to the presence of the bay, numerous data such as seismic and cores have been collected via marine seismic acquisition systems and platform drilling operations; the data have been used to interpret, with relatively high precision, the
spatiotemporal patterns of the Trinity Holocene incised valley infill [Anderson et al., 1992; Thomas and Anderson, 1994; Rodriguez et al., 2005; Simms et al., 2006; Anderson et al., 2008].

The age and distribution of accumulated sediment, along with the morphology of the terraced valley, are invaluable data for developing and constraining a numerical model that replicates system response to base level rise. The data provide excellent control of input parameters and the ability to check resulting depositional patterns produced by the model (Figures 3-4). Antecedent topography exerts an important influence on incised valley infill under conditions of steady base level rise and sediment flux [Rodriguez et al., 2005; Simms et al., 2006; Anderson et al., 2008]. Because the terraced antecedent morphology is well-constrained, it is possible to explore its influence on fluvial-deltaic stratigraphy (Figure 3).

**Base level rise**

Sea level rise and subsidence rates are well-constrained for the Northern Gulf of Mexico over the last 10,000 years and are considered cumulatively to estimate the total base level rise for the Trinity system. Milliken et al. [2008] produced a sea level curve for the Northern Gulf of Mexico derived from bayline peat and swash-zone deposits which are reliable sea level markers through time as they are readily dated (Figure 5) [Milliken et al., 2008]. A constant rate of subsidence equal to 0.07 mm yr$^{-1}$ is applied, based on Simms et al. [2013], who reported that long-term subsidence along the Texas Gulf Coast is predominantly attributed to sediment loading, compaction, and isostatic adjustments. However, subsidence rates may be as high as 3 mm yr$^{-1}$ within the bay due to sediment compaction [Simms et al., 2013].
Water discharge

Water discharge measurements for the Trinity River were collected from the USGS stream gauge station 8065350 near Crockett, TX, approximately 378 km (along stream) upstream of Galveston Bay and above Livingston Dam (Figures 1 and 6). This location was selected as it is the farthest downstream stream gauge along the Trinity River that is not influenced by tides and it is located upstream of the Livingston Dam. Daily water discharge measurements are available from 1964 to present (Figure 6).

We consider water discharge for modern bankfull conditions, determined by comparing measured stage height and the corresponding water discharge (Figure 7). Bankfull discharge is evaluated to be the discharge above which no significant change in stage is measured, which typically corresponds to the one to two year flood recurrence interval [Leopold et al., 1964]. For the Trinity River, a value of 1500 m$^3$ s$^{-1}$ best fits these criteria (Figure 7). Channel width for the Trinity is relatively constant and is approximately 500 m at bankfull conditions, based on analysis of aerial maps of the Trinity system, and measurements of channel width collected by the USGS at stream gauge stations along the river.

The principle of formative discharge assumes that flood intermittency defines sediment transport duration, typically corresponding to discharges of two to five year flood events [Wolman and Miller, 1960]. Analysis of recurrence intervals of such events using the water discharge data at Crockett, TX, indicates that the two to five year flood discharge ranges 700-1500 m$^3$ s$^{-1}$. Since 1500 m$^3$ s$^{-1}$ coincides with the five year flood discharge and the estimated bankfull discharge, we assume this value in our subsequent analyses.

The Trinity River basin spans a precipitation gradient that produces generally variable and seasonal flow fluctuations. The degree of seasonality is captured through the use of an
intermittency parameter to approximate formative discharge events. The intermittency is
determined by the number of days within the dataset at the Crockett stream gauge that occur
within the formative discharge range (700-1500 m³ s⁻¹) (Figure 6). The intermittency value is
adjusted for years within the dataset that do no exhibit formative discharges. An intermittency
value of 0.05, or approximately 20 days per year at bankfull discharge conditions, is determined
to meet this condition.

**Backwater hydrodynamics**

The backwater length scale ($L_B$) is characterized by conditions where the downstream
water surface elevation asymptotically approaches the relatively constant water surface elevation
of the receiving basin, and non-uniform flow conditions arise. $L_B$ is approximated as $H S^{-1}$,
where $H$ is flow depth and $S$ is the channel bed slope [Nittroeur et al., 2012; Chatanantavet et al.,
2012; Lamb et al., 2012]. Non-uniform flow arises because the channel bed maintains a constant
downstream slope and diverges from the water surface profile, thus producing downstream
increases to flow depth and cross sectional flow area ($A$) assuming a constant bankfull channel
width ($B = 500$ m), where $A$ is evaluated as: $A = H B$ [Parker, 2004; Nittroeur et al., 2012].
Assuming a constant water discharge ($Q_w = 1500$ m³ s⁻¹), reach-average flow velocity ($U$) may be
calculated as

**Equation 1:**

$$U = \frac{Q_w}{A}.$$  

Therefore, where cross-sectional channel area increases downstream, flow velocity decreases,
therefore producing non-uniform flow conditions.
Floodplain development

Floodplain sedimentation occurs as a result of deposition on the inside of river bends, subsequent lateral migration of the channel, and by sediment-laden overbank flows that produce fine-grain sediment deposits [Leopold et al., 1964]. The thickness of the laterally migrating sandy point bar deposits is scaled by the depth of the channel [Parker et al., 2011]. For the conditions of zero base level rise and sediment transport continuity, it is expected that floodplain deposition and erosion are in equilibrium; thus sediment flux is constant, justifying the exclusion of floodplain dynamics in previous morphodynamic models of backwater induced sediment deposition [Parker, 2004; Nittrouer et al., 2012; Lamb et al., 2012; Chatanantavet et al., 2012; Ganti et al., 2014]. However, we suppose here that floodplain dynamics are an important component of long-term system response for the condition of varying base level.

In terms of overbank sedimentation, Lauer and Parker [2008A; 2008B] identified a characteristic advection length of mud over the Clark Fork River, Montana, floodplain of approximately 200 km. Because the Trinity floodplain width \( b_f = 7.2 \) km is small compared to the estimated advection distance of mud, accumulation of washload sediment in the lateral direction on the floodplain is considered negligible. Further, washload deposits are typically reworked by lateral migration of the channel and entrained and transported downstream, thus minimizing washload as a contributor to floodplain sedimentation; therefore, floodplain deposits are primarily composed of unconsolidated channel sands (Figure 8) [Parker et al., 2008C; Parker et al., 2011].
Avulsions

A river channel avulsion is the rapid abandonment of an established flow path to an alternative location on the floodplain [Mohrig et al., 2000]. This process is shown to influence the long-term distribution of sediment deposits in fluvial-deltaic basins [Mohrig et al., 2000]. Channel avulsions are set up by sediment aggradating the river bed, which diminishes the channel depth and cross-sectional area, and are triggered by events, typically large floods, which quickly alter channel capacity [Mohrig et al., 2000; Slingerland and Smith, 2004]. An avulsion is therefore characterized by sediment aggradation and decreasing channel depth [Mohrig et al., 2000; Jerolmack and Mohrig, 2007], and the characteristic time to an avulsion ($T_A$) is estimated by

\[ T_A = \frac{H}{\dot{V}_A}, \]

where $H$ is flow depth and $\dot{V}_A$ is the vertical aggradation rate of sediment on the channel bed [Mohrig et al., 2000; Jerolmack and Mohrig, 2007]. However, it is observed that in modern and paleo-systems alike, that channels tend to avulse on timescales less than $T_A$ (Equation 2) [Mohrig et al., 2000; Jerolmack, 2009; Ganti et al., 2014]. Ganti et al. [2014] suggest that channel avulsion timescales are better approximated by using $0.3-0.6H$, based on data analysis from the Yellow River, China.

The relationship between backwater hydrodynamics and avulsion locations of fluvial-deltaic systems is well-established [Jerolmack and Swenson, 2007; Nittrouer et al., 2012; Chatanantavet et al., 2012; Lamb et al., 2012]. Avulsions occur preferentially where uniform flow transitions to non-uniform flow (i.e. within the backwater region) because fluid flow enters a regime of spatial deceleration, which reduces sediment transport capacity and produces focused
sediment deposition, thereby minimizing the time required for the system to setup an avulsion 
($T_a$) [Jerolmack and Swenson, 2007; Nittrouer et al., 2012; Chatanantavet et al., 2012].
MODELING FRAMEWORK

Our 1D model calculates the interactions of dynamic boundary shear stress, sediment transport, and response of bed topography for the Trinity River during early Holocene transgression (10-8.2 ka). These variables feedback onto one another to produce sediment deposition that ultimately sets up channel avulsions. Some parameters from the modern Trinity River are used for the model, specifically, water discharge and sediment concentration. To first order, these are reasonable assumptions because early Holocene climate conditions of the study area are estimated as comparable to the modern [Muskogee et al., 2001; Weight et al., 2011]. We therefore utilize modern data to produce some input parameters for our morphodynamic model.

A channel infill model explores how differing boundary conditions, such as changing base level rise rate and water discharge, influence the behavior of the system; particularly how sediment deposition builds incised valley stratigraphy. Topographic change is estimated via a long-term Exner equation, and related to channel aggradation, in order to determine avulsion locations and timescales.

Backwater hydrodynamics, sediment transport, and bed evolution

Our model evaluates flow depth \( (H) \) through along-stream distance \( (x) \):

\[
\frac{dH}{dx} = \frac{S - C_f Fr^2}{1 - Fr^2},
\]

where \( S \) is the channel bed slope, \( C_f \) is a constant friction coefficient equal to \( 3.6 \times 10^{-3} \) [Parker, 2004], and \( Fr \) is the Froude number, determined by \( Fr = U(gH)^{1/2} \) where \( g \) is acceleration due
to gravity (9.81 m s\(^{-2}\)) [Parker, 2004; Nittouer et al., 2012]. \(\text{Equation 3}\) is solved using a simple Euler method and predictor-corrector scheme, where the initial linear bed slope is set as \(1.6 \times 10^{-4}\) [Rehkemper et al., 1969].

The backwater length scale — the length of non-uniform flow conditions — within the system is evaluated using two inflection points of the water surface elevation profile. The upstream initiation of backwater conditions starts where \(dH \, dx^{-1} > 0\), and where flow velocity deviates from the uniform flow velocity that persists where the channel bed and water surface elevation are parallel, i.e., where \(dH \, dx^{-1} = 0\). Backwater flow extends downstream to where the water surface elevation plateaus and maintains a constant slope value of nearly zero, i.e. where \(dh \, dx^{-1} = 0\) and \(h\) is the water surface elevation plus \(\eta\), the bed elevation from a fixed datum. The distance between the points where \(dH \, dx^{-1} > 0\) and \(dh \, dx^{-1} > 0\) demarcates the extent of non-uniform hydrodynamic conditions. In our model, we impose backwater conditions where \(dH \, dx^{-1}\) (upstream) and \(dh \, dx^{-1}\) (downstream) are greater than \(5 \times 10^{-5}\) as this is deemed to be significant, yet reasonable, water surface slope deviation from normal flow conditions because \(5 \times 10^{-5} \ll 1.6 \times 10^{-4}\), the initial channel bed slope.

For each time step, the water surface elevation is kept constant, i.e., it is pinned, at the channel mouth by the relative base level elevation. Water surface elevation increases for each time step based on the known, local sea level curve (\(\text{Figure 5}\)) [Milliken et al., 2008]. Water surface elevation for the remaining channel is calculated upstream starting at the river mouth. Therefore, water surface elevation is variable in time and space, and is a function of river water discharge and the base level elevation at the river mouth.

Non-uniform flow velocity coinciding with backwater hydrodynamics influences the sediment transport capacity of the system. We use the Engelund-Hansen equation for total load
of transported bed material, i.e., the sediment found in appreciable quantity on the bed that may be transported as part of either bedload or suspended load [Engelund and Hansen, 1967]. Bed material sediment is considered the ‘formative’ sediment of lowland river systems, because this sediment conditions channel dynamics for lowland fluvial systems [Parker, 2004]. Bed material load per unit channel width ($q_t$) is calculated by:

**Equation 4:**

$$q_t = \sqrt{R_g D D} \frac{0.05}{C_f} \left( \frac{\tau_b}{\rho R_g D} \right)^{2.5},$$

where $R$ is the submerged specific gravity of the sediment and $D$ is the median grain-size [Engelund and Hansen, 1967]. A constant median grain-size of 250 μm, which represents fine sand, is applied in this model to simulate the Trinity River [Rehkmper et al., 1969]. Boundary shear stress ($\tau_b$) is the available stress to move sediment and is determined by:

**Equation 5:**

$$\tau_b = \rho C_f U^2,$$

where $\rho$ is fluid density (1000 kg m$^{-3}$).

The spatial divergence in sediment flux ($\frac{\partial q_t}{\partial x}$) is determined by spatially iterating Equation 4 using the Exner equation [Exner, 1925], thereby calculating the temporal change in local bed elevation ($\frac{\partial \eta}{\partial t}$):

**Equation 6:**

$$(1 - \lambda_p) \frac{\partial \eta}{\partial t} = -\frac{\partial q_t}{\partial x},$$

where $\lambda_p$ is bed porosity which is held at a constant value of 0.4. Equation 6 is solved numerically using a central difference and predictor-corrector scheme. The intermittency
parameter is applied to Equation 6 because the fluvial system is only morphologically active for the fraction of time determined in our study by the formative discharge frequency. In other words, each iteration of Equation 6 is multiplied by the intermittency parameter. Equation 6 captures bed aggradation within the modeled domain. Here we equate the change in bed topography over time $\left(\frac{\partial h}{\partial t}\right)$ to the vertical aggradation rate of the channel $\left(\frac{V_d}{V_f}\right)$, so to approximate avulsion timescales using Equation 2.

**Floodplain**

Here, we want to consider sediment interaction between the channel and the adjacent floodplain. To this end, we use the Parker [2004] modified version of Equation 6, which describes the ratio of bankfull channel width ($B$) multiplied by sinuosity ($\Omega$) to floodplain width ($B_f$) as a reducing factor on channel-bed aggradation, whereby the long-term Exner equation is given by:

**Equation 7**

$$ \left(1 - \lambda_p\right) \frac{\partial h}{\partial t} = -\Omega \frac{B}{B_f} \frac{\partial q_t}{\partial x}.$$

Sinuosity is approximated using modern measurements, based on the ratio of along-stream distance ($\Delta x$) and the measured down-valley length (straight distance, $\Delta x_v$). Typical sinuosity for modern meandering rivers ranges 1.5-3.0 [Parker, 2004], and sinuosity for the Trinity River is approximately 1.86. Implementation of this ratio acts as a coefficient for partitioning sediment between the channel bed and the floodplain; however, floodplain sedimentation is not explicitly modeled here. Similar to Equation 6, an intermittency parameter is applied to Equation 7.
Avulsions

An avulsion occurs in the model domain at the location where the channel bed aggrades to a designated fraction of the flow depth. We simulate avulsion thresholds that range 0.3-0.6H [Ganti et al., 2014]. Flow depth (H, Equation 3) is calculated for each time step (one-year increments) within the model, and for each iteration, the change in bed elevation (η, Equation 7) is computed and compared to flow depth throughout the model domain. Since the downstream-most boundary is dynamic in time, and thus flow depth is dynamic in time, the flow depth must be calculated iteratively at each time step within the model to identify the location where the channel aggrades to the imposed avulsion threshold (0.3-0.6H).
RESULTS

Flow velocity and backwater region

The model captures variability of flow velocity in space and time (Figure 9). The flow velocity profiles maintain a similar trend through time but flow velocity at any given spatial point, identified as river kilometer (RK) upstream of the outlet, decreases through time (Figure 10). It is possible for the flow velocity profile to decrease in time because while water discharge and channel width are held constant, cross-sectional area increases as base level rises, so that via Equation 1, flow velocity is variable in time. Additionally, flow velocity decreases downstream for each time step (Figure 10). However the location for the onset of decreasing flow velocity migrates upstream over time. For example, in Figure 9, a flow velocity of 0.8 m s\(^{-1}\) occurs at 44 RK at time 0, and after 1,821 model years, the same flow velocity occurs at 87 RK. As seen in Figure 9, flow velocity decreases rapidly and is expressed as the steeply sloping region of the plotted flow velocity through time and space. A constant, flow velocity is observed in the upstream portion of the model space and is indicative of normal, uniform flow conditions. All flow velocity profiles ultimately converge to normal flow conditions upstream, thus indicating no change of flow velocity through time in the upstream domain of the model. As the model progresses, the extent of uniform flow and the transition to non-uniform flow migrate upstream.

Flow depth changes significantly in the downstream portion of the model compared to the upstream portion of the model as a result of the imposed base level rise. Therefore channel cross-sectional area is considerably greater downstream than upstream. Following Equation 1, this relationship produces a downstream decrease in flow velocity, which is observed in Figure 9 and 10, and it follows that the backwater region is spatially mobile through time, as observed in Figure 11. That is, as base level rises, the onset and termination of backwater hydrodynamics backstep through time. Interestingly, the backwater length first develops non-linearly, and after
approximately 1,000 model years exhibits nearly linear development (Figure 11). Furthermore, the measured extent, i.e. the distance between the onset and termination of the backwater region, increases (lengthens) through time.

**Boundary stress, sediment transport, and bed evolution**

Flow velocity is used to estimate boundary shear stress (Equation 5). A downstream non-linear decreasing trend emerges (Figure 12A). Boundary shear stress is used to predict sediment transport (Equation 4), and as can be expected, sediment load decreases downstream due to decreasing sediment transport capacity. The avulsion location ($L_A = 84$ RK) coincides with the regions of greatest spatial change (i.e., steep slope) of both shear stress and sediment load (Figure 12A). As seen in Figure 12B, the region of channel bed sediment deposition coincides with the region of rapid spatial decrease in shear stress and sediment transport rates. In other words, the majority of sediment deposition to the channel bed occurs within the region of decreasing sediment transport capacity.

The morphology of the sediment deposit on the channel bed is transient and varies in accordance with rates of base level rise. Figure 13 depicts four scenarios of varying rates of base level rise to portray the possible morphologies of sediment wedges deposited before reaching an avulsion threshold of $0.3H$. Under the static boundary condition of no base level rise ($BL = 0$ mm yr$^{-1}$, Figure 13A), the sediment wedge simply progrades downstream. Similarly, the sediment wedge produced for a base level rise of 2 mm yr$^{-1}$ progrades downstream, though it reaches a greater bed elevation than the static scenario, due to the increased amount of time necessary to reach the avulsion threshold (Figure 13B). When base level rise is simulated at 4 mm yr$^{-1}$, the sediment wedge first progrades, then backsteps before reaching the avulsion
threshold (Figure 13C). For rates greater than 5 mm yr\(^{-1}\), the sediment wedge progrades, then retrogrades but never reaches the avulsion threshold (Figure 13D). As base level rate of change diverges from the static (zero) scenario, the deposited sediment wedge, qualitatively speaking, produces a more rounded morphology.

The locus of sediment deposition to the bed is tracked by determining the location of the maximum change in bed elevation between each time step \( \frac{\partial h}{\partial t} \) (Figure 14). As the maximum change in bed elevation moves downstream, a negative slope is produced and the sediment is prograding. As the maximum change in bed elevation migrates upstream, a positive slope is produced and the sediment is backstepping. At rates of base level rise of 4.27 mm yr\(^{-1}\), sediment progrades downstream at a rate of 12 m yr\(^{-1}\) and then backsteps at a rate of 26 m yr\(^{-1}\). Rates of change are considered with respect to horizontal distance. The transition between prograding and backstepping sediment deposition under input parameters that represent the early Holocene Trinity River occurs after 364 model years.

**Avulsion and floodplain influence**

Avulsions are set in the model when the sediment wedge aggrades to a specified avulsion threshold (0.3-0.6H), and where this occurs, the location \((L_A)\) is noted. *Equation 6* was first applied to determine the time for avulsion under various base level rise scenarios without the influence of a floodplain. In these modeled scenarios, all sediment input to the system is deposited to the bed. The results of these model runs indicate a nearly constant time and location of avulsion (Figure 15, black lines and squares).

Utilizing *Equation 7* and considering the flux of sediment to the floodplain to determine the time and location for avulsion under varying base level rise rates resulted in an increased
avulsion time and backstepping of the avulsion location with increasing base level rise rate 
(Figure 15, red lines and x’s). At base level rise rates less than 2 mm yr$^{-1}$, the floodplain 
inclusion results remain approximately constant. When sea level rise is less than 2 mm yr$^{-1}$, 
avulsion locations are coincident despite the inclusion or exclusion of sediment flux to the 
floodplain. Using the lowest avulsion threshold, 0.3$H$, no avulsions occur at base level rise rates 
greater than 4.5 mm yr$^{-1}$ when the floodplain is included in the calculation. Increasing the 
avulsion threshold decreases the highest rate of base level rise under which an avulsion can be 
set up. For example, in Figure 16, an avulsion threshold of 0.5$H$ is only capable of setting up an 
avulsion for base level rise rates of less than or equal to 2.5 mm yr$^{-1}$.

As seen in Figures 9 and 12, the avulsion location spatially coincides with decreasing 
flow velocity, boundary shear stress, sediment load, which therefore produces sediment 
deposition. Figures 13 and 15 reveal that the avulsion location is sensitive to changes in the rate 
of base level rise. As the rate of base level rise increases, the avulsion location migrates upstream 
(Figures 13A, 13B, and 13C). However, above a certain rate of base level rise (5 mm yr$^{-1}$ in the 
case of Figure 13D), no avulsion occurs. Additionally, as the rate of base level rise increases, the 
time to reach the avulsion threshold increases (Figures 13A, 13B, and 13C).

After imposing the 0.3-0.6$H$ avulsion threshold range to the channel fill model as 
suggested by Ganti et al., [2014], it is apparent that the time to avulsion for a given rate of base 
level rise increases with increasing avulsion threshold (Figure 16). Applying the lowest avulsion 
threshold (0.3$H$), it is possible to set up avulsions at base level rise rates of up to 4.5 mm yr$^{-1}$ 
using the water and sediment discharge characteristics of the Trinity River. The highest avulsion 
threshold (0.6$H$) is unable to set up avulsions at base level rise rates greater than 2 mm yr$^{-1}$ using 
Trinity River characteristics. Below base level rise rates of 1 mm yr$^{-1}$, avulsion times for all
thresholds are contained within a narrow window in which the avulsion time responds linearly to the base level rise. Exceeding 1 mm yr\(^{-1}\) base level rise, the time for avulsion responds exponentially to base level rise.

In addition to the variability for the time to avulsion for different thresholds, the time to avulsion also varies with water discharge (Figure 17). Higher water discharge allows avulsions to occur even under higher rates of base level rise. For example, in Figure 17, water discharge equal to 1200 m\(^3\) s\(^{-1}\) can only set up avulsions for base level rise rates of less than 2 mm yr\(^{-1}\), whereas a water discharge equal to 2500 m\(^3\) s\(^{-1}\) can set up avulsions at rates of base level rise up to 7.5 mm yr\(^{-1}\). The times to avulsion for all simulated discharges cluster within a narrow range and show a linear relationship for up to approximately 2 mm yr\(^{-1}\) of base level rise (Figure 17).
DISCUSSION

Influence of floodplain on channel avulsions

The inclusion of a floodplain within the model through the application of the long-term Exner equation suggested by Parker [2004] (Equation 7) distinguishes the model approach here from recent model studies (i.e., Nittouer et al. [2012]; Lamb et al. [2012]; Chatanantavet et al. [2012]; Ganti et al. [2014]). The consideration of the floodplain geometry means that a significant volume of sediment is partitioned to the floodplain which influences both the time and location of channel avulsions (Figure 15). For example, Equation 6 cannot be used to examine the influence of base level rise because the avulsion times and locations remain constant regardless of the rate of base level rise. This occurs because Equation 6 renders all bed material sediment input to the system deposited to the channel bed. However, this is probably unrealistic in the natural world, particularly over stratigraphically meaningful timescales, because sediment is frequently exchanged between the channel and the floodplain via channel migration and overbank flood processes [Leopold et al., 1964; Lauer and Parker, 2011]. Since using Equation 6 renders all sediment deposited to the bed (i.e., no floodplain consideration), the channel aggravates to the avulsion threshold so quickly that the influence of base level rise is irrelevant.

For discussion of the Trinity River, significant base level rise rates occur above 2 mm yr\(^{-1}\). For example, in Figure 15, even with the application of floodplain sedimentation (Equation 7), it is not until base level rise rates are greater than 2 mm yr\(^{-1}\) that the time and location of avulsion increase exponentially, rather than maintain a relatively constant value as seen for conditions of 0-2 mm yr\(^{-1}\) base level rise. The primary difference here is that the results of the model using Equation 6 imply that a channel avulsion is never affected by base level rise, whereas the results of the model using Equation 7 render channel avulsion affected by base level rise rates of a certain magnitude; in the case of the Trinity River, this threshold rate is
approximately 2 mm yr$^{-1}$. In situations of no base level rise, as has been considered by previous workers (i.e., Nittroer et al. [2012]; Lamb et al. [2012]; Chatanantavet et al. [2012]; Ganti et al. [2014]), it could be adequate to model channel avulsions without consideration of the floodplain. Nevertheless, this approach decreases the time to avulsion because all sediment transported within the model is captured to the channel bed rather than interacting with the adjacent floodplain. For the purposes of linking sediment transport processes to stratigraphy, we argue that it is necessary to account for floodplain sedimentation which characterizes fluvial-deltaic systems over long, stratigraphically meaningful, time scales (i.e., centuries to millennia).

Response of backwater hydrodynamics, avulsions, and depositional morphology to base level rise

Flow depth increases downstream where the water surface profile, characterized as an M1 curve [Chow, 1959], diverges from the channel bed. This results in non-uniform flow conditions that spatially reduce flow velocity, boundary shear stress, and sediment flux. Therefore, the backwater transition is an area of preferential sediment deposition, and so the avulsion location coincides with this region because of the enhanced vertical aggradation rate (Equation 1; Figures 9 and 12).

The model results presented herein emphasize the connection between backwater hydrodynamics and the avulsion location (Jerolmack and Swenson [2007], Nittroer et al. [2012], Chatanantavet et al. [2012], and Lamb et al. [2012]). Jerolmack and Swenson [2007] first proposed that the upstream limit of the backwater regime was the maximum upstream location at which distributary avulsions and delta sedimentation could occur, although they did not model this effect numerically. Nittroer et al. [2012] observed that for the lowermost
Mississippi River, preferential sediment deposition occurs within the upper reaches of the backwater segment and that this location coincides with the known Holocene avulsions. Our simulated avulsion location for the Trinity River ($L_d = 84$ RK) remains within the backwater region throughout the run time of the model and never exceeds the upstream extent of backwater conditions (Figure II), thereby numerically confirming the observations of Jerolmack and Swenson [2007] and Nittrouer et al. [2012]. Contrary to the models of Chatanantavet et al. [2012] and Lamb et al. [2012], channel bed scour produced by an M2 curve—whereby the water surface profile is convex and drawn down toward the receiving basin thus increasing flow velocity and sediment transport capacity downstream [Chow, 1959]—are not observed for the simulated Trinity River high flow conditions. An M2 curve is therefore not required in our case to set up avulsions in the backwater region.

Base level rise affects the upstream extent of the backwater length because the elevation of the river water surface is influenced by the elevation of the receiving basin downstream. The onset of the backwater region migrates upstream as base level rises (Figure II). It follows then that the spatial decrease in flow velocity, shear stress, and sediment flux also migrate upstream as base level increases. These conditions combine to produce an upstream migration of the avulsion location as a function of base level rise. For example, Figures 13A, 13B, and 13C demonstrate that avulsions occur in positions progressively farther upstream in accordance with the respective rates of base level rise.

The influence of base level rise is also rendered in the morphology of the sediment deposited to the channel bed. As seen in Figures 13A and 13B, up to base level rise rates of 2 mm yr$^{-1}$, a prograding sediment wedge is deposited, however base level rise rates greater than 2 mm yr$^{-1}$ show initial progradation followed by upstream backstepping (Figures 13C and 13D). It
is clear, in all scenarios depicted in Figure 13, that the slope of the water surface profile mimics the slope of the sediment wedge profile. Parker [2004] developed a model, with static boundary conditions, by which sediment is deposited within in the backwater region, so to re-grade the channel bed slope downstream and thereby reclaim normal flow conditions. Our model of the Trinity River, with dynamic boundary conditions, shows that sediment is deposited within the backwater region also by backstepping, particularly for base level rise rates greater than 2 mm yr⁻¹.

It is proposed that there is a threshold, or turn-around point, for which fluvial sediment aggradation can no longer keep pace with the changes of base level, and therefore sedimentation backsteps with increasing base level elevation. The proposed threshold of minimum base level rise is reflected in the model results for the Trinity and is equal to 2 mm yr⁻¹, whereby the time and location for an avulsion remain approximately constant until this threshold of base level rise is exceeded (Figure 15).

Collating our results with the Parker [2004] model, we conclude that at rates of base level rise less than 2 mm yr⁻¹, progradation will occur for the Trinity River so that deposits prograde and aggrade to reclaim normal flow conditions downstream. It is not presently known if this proposed base level rise threshold is applicable for other lowland fluvial systems. However, we do propose that a threshold for backstepping sedimentation likely exists for lowland fluvial-deltaic systems, even though the value of base level rise rate may differ; future analysis is needed.
Sensitivity of avulsions to varying thresholds and water discharge

A range of water discharges (1,200-2,500 m$^3$ s$^{-1}$) and avulsion thresholds (0.3-0.6H) were explored to determine the range of possible avulsion locations and timescales using Trinity River parameters. Figure 16 displays the possible range of avulsion timescales using different thresholds and base level rise rates for a range of avulsion criteria. More time is required for sediment to be deposited to reach the 0.6H threshold than for the 0.3H threshold. If the rate of base level rise exceeds the rate of sediment aggradation, the avulsion threshold is not met. For example, the last point of each line in Figure 16 represents the upper limit of the base level rise rate for the system to set up an avulsion. For lower avulsion thresholds, e.g., 0.3H, avulsions are set up for higher rates of base level rise because less sediment is required to aggrade a smaller fraction of the flow depth. This reasoning explains why the avulsion time for a 0.3H threshold at 4.5 mm yr$^{-1}$ is less than the avulsion time for a 0.5H threshold at 2.5 mm yr$^{-1}$.

The range of possible avulsion times is also sensitive to water discharge (Figure 17). Higher water discharges (2,000-2,500 m$^3$ s$^{-1}$) are able to produce avulsions at higher rates of base level rise (greater than 5 mm yr$^{-1}$) because greater sediment transport capacity produces enhanced sediment transport rates which denude the channel bed upstream and allow a greater volume of sediment to be deposited to the channel bed downstream. Therefore higher water discharges reach the avulsion threshold sooner than lower water discharges (1,200-1,500 m$^3$ s$^{-1}$) because they are able to transport and thus deposit a greater volume of sediment (Figure 17).

Exploring input parameters and boundary conditions is intended to simulate a range of possible Holocene scenarios of the Trinity River. These ranges of input parameters and boundary conditions nevertheless produce similar outcomes and thus bolster the case of utilizing the Trinity River as a type system for investigating the linkages between backwater hydrodynamics,
base level adjustments to mimic transgression, and the production of stratigraphy. We aim to capture the broader essence of the system by tracking sedimentation patterns and avulsion characteristics under variable input parameters and boundary conditions.

**Application to the Trinity River**

The fluvial-deltaic deposits of the Trinity River incised valley infill are well-constrained and represent early Holocene transgression between 10,000 and 8,200 years B.P.; during this time, sea level was rising approximately 4.2 mm yr$^{-1}$ ([Figures 3-5]) [Rodriguez et al., 2005; Anderson et al., 2008; Milliken et al., 2008]. It is known that incised valleys with terraced morphologies are characterized by backstepping facies during base level rise as terraces are flooded [Rodriguez et al., 2005]. The terraced Trinity incised valley system also exhibits backstepping facies, including fluvial, bayhead delta and basin muds, throughout Holocene transgression at horizontal rates of up to −25 m yr$^{-1}$ as measured from age dated cores below the first terrace ([Figure 4]) [Anderson et al., 2008].

Our model results demonstrate the backstepping character of in-channel sedimentation and avulsion location for base level rise rates of greater than 2 mm yr$^{-1}$, as a result of backstepping backwater hydrodynamics ([Figure 13]). For a base level rise rate of 4.27 mm yr$^{-1}$, the model calculates backstepping rates of 26 m yr$^{-1}$, which corresponds well to the backstepping rates observed from Galveston Bay core data ([Figures 4 and 14]). We propose that sediment deposits backstep upstream during transgression at a rate proportional to the ratio of base level rise to the characteristic channel bed slope. For example, the Holocene Trinity River base level rise is 4.27 mm yr$^{-1}$, and the characteristic channel slope is $1.6 \times 10^{-4}$, therefore the
estimated rate of horizontal backstepping is 26.7 m yr$^{-1}$, a number quite similar to both the measured and modeled results.

Although we only model one grain size, it is assumed that when the framework is applied to a natural fluvial system, including the Trinity River, washload sediment will be deposited to the downstream bayhead delta and bay mud depositional environments. By capturing backstepping fluvial facies within the model, we assume that the observed backstepping of both bayhead delta and bay mud facies in Galveston Bay follows the channel systematically (Figure 4).

Vertical aggradation rates of fluvial sediments in Galveston Bay are calculated as ~1.5 mm yr$^{-1}$ between the onset of Galveston Bay sedimentation (~17 ka) and the known flooding surfaces at 9.6 and 8.9 ka (Figure 4) [Rodriguez et al., 2005; Anderson et al., 2008]. Our model results produce vertical aggradation rates of channel sedimentation 0.65 mm yr$^{-1}$ for the time period 10-8.2 ka. The discrepancy in vertical aggradation rates may be due to the fact that the model only considers sedimentation within the channel, which could under-predict the total amount of sediment dispersed, or because of a time bias which arises because the exact dates of the fluvial sands are largely unknown within Galveston Bay.

Based on our model results, the rate of base level rise was sufficient to generate channel avulsions only if the avulsion threshold was 0.3H because base level rise (4.27 mm yr$^{-1}$) was quite rapid and outpaced sedimentation. If we consider a greater threshold (i.e. 0.4-0.6H), avulsions can only occur at a lower base level rise rate no greater than 3.5 mm yr$^{-1}$ (Figure 16). Therefore, if avulsions did in fact occur for the Trinity River in the early Holocene, it is concluded that an avulsion threshold of 0.3H is required for the known base level rise rate.
Using the modern bankfull discharge of 1500 m³ s⁻¹ and an avulsion threshold of 0.3H, the simulated Trinity River only avulses at rates of base level rise up to 4.5 mm yr⁻¹ (Figure 17). Higher discharges of 2,000-2,500 m³ s⁻¹ are necessary for the modeled Trinity River to avulse at a base level rise rates greater than 4.5 mm yr⁻¹. However, the rate of sea level rise does not exceed 4.2 mm yr⁻¹ in our modeled time of interest (Figure 4) [Milliken et al., 2008]. The higher range of modeled discharges (2,000-2,500 m³ s⁻¹) does occur within the modern Trinity River as reported at the USGS Crockett, TX stream gauge station, but with lower frequency than the bankfull 1,500 m³ s⁻¹, and so we do not consider this a formative discharge (Figure 6).

The model results and subsequent analyses assist in characterizing the along stream morphology of fluvial deposits, estimating timescales for avulsion and sediment distribution within the incised valley fill, and ultimately the mechanisms by which sediment fills an incised valley. At early Holocene rates of base level rise (4.2 mm yr⁻¹), sediments deposited to the channel bed display a backstepping trend, whereby the sediment transitions from progradation to retrogradation, as shown in Figure 13. The sensitivity analysis of model input parameters (Figures 16 and 17) demonstrates that sediment distribution within the incised valley during the early Holocene can be facilitated by avulsion.

The floodplain plays a significant role for distributing and perhaps exchanging sediment. We use a Rouse profile [Rouse, 1937] to estimate the flux of suspended bed material sediment, by calculating sediment concentration in the water column (Figure 18). Overbank deposition of sediment occurs during flood events when the water surface exceeds the bankfull flow depth (H = 3 m), here we assume the conservative case of 2 m overbank flow depth which produces a total flow depth of d = 5 m (Figure 18). The average suspended bedload concentration in the upper 2 m of flow is 1.8 x 10⁻⁶ based on a grain size of 250 μm; this is summed with the
measured constant volumetric concentration of washload sediment, $4.3 \times 10^{-4}$ [Rice, 1969], to produce a total overbank volumetric concentration of $4.48 \times 10^{-4}$ (Figure 18). To estimate a potential volume of sediment exported to the floodplain during an overbank event, a flood discharge of $2,000 \text{ m}^3 \text{s}^{-1}$ is selected based on analysis of stage and discharge data; this value allows $500 \text{ m}^3 \text{s}^{-1}$ of water to escape the channel and enter the floodplain (Figures 6 and 7). 500 m$^3$ s$^{-1}$ is multiplied by 1,821 years (time to avulsion under 4.27 mm yr$^{-1}$ base level rise), an intermittency of 0.02 to simulate the infrequency of overbank events, and by the total volumetric concentration of overbank sediment ($4.48 \times 10^{-4}$). This calculation estimates $0.26 \text{ km}^3$ of sediment exported to the floodplain in 1,821 years. Over the area of the floodplain in the model space (2,160 km$^2$), this volume of sediment would be approximately 0.12 m thick. This thickness, when compared to the bankfull flow depth of 3 m is a mere 4% of the summed channel and floodplain height.

Over the width of the floodplain ($B_f = 7.2 \text{ km}$), these exported sediment volumes and concentrations are quite small and concluded to be minor contributors to floodplain sedimentation. Based on the 200 km advection length of fine sediment suggested by Lauer and Parker [2008A; 2008B], it is postulated that overbank washload sediment travels basinward (downdip) and may interact with the floodplain, or be advected to the delta. This scenario therefore limits washload deposition on the floodplain during overbank events.

Minimal deposition of fine-grain sediment to the floodplain with respect to channel sedimentation offers an explanation for the abundance of stacked sands relative to the proportion of mud preserved observed by Rodriguez et al. [2005] (Figure 3). We propose that floodplain sedimentation for the Trinity River is dominated by lateral migration of the channel across the floodplain, and therefore the floodplain is primarily comprised of unconsolidated channel sand
capped by a thin veneer of fine sediment associated with overbank flood events. It is established that the distribution of floodplain sediment occurs not only by avulsion, but also by lateral migration of the channel by depositing sand and reworking fine material deposited to the floodplain (Figure 8) [Lauer and Parker 2008C; Parker et al., 2011].

The rates of lateral migration for the Trinity River prior to construction of the Livingston Dam in 1968 as calculated by Wellmeyer et al. [2005] are 3.0 m yr\(^{-1}\) between 1938 and 1958 and 6.5 m yr\(^{-1}\) between 1958 and 1964. The weighted average is equal to 3.8 m yr\(^{-1}\). Using this value, it would take the Trinity River \(\sim 1,895\) years to traverse the width of the incised valley \((B_f = 7.2\) km) [Rodriguez et al., 2005]. The time for avulsion of the Trinity River in the early Holocene, according to our model, is 1,821 years.

We find this to be an interesting coincidence, and propose that the width of the incised valley may be influential in setting the time of avulsion, or vice versa, as both times are quite similar. This is to say that the channel migrates laterally across the valley; incrementally gaining elevation via sediment deposition on the channel bed as base level rises, and over the time necessary to reach its avulsion threshold, the channel has potentially covered its entire floodplain. It is possible that the valley width and the avulsion threshold contain unilateral movement of the channel, so that avulsion carries the channel to some other topographic low on the floodplain, which under the condition of continuous channel aggradation, is likely the location that has hosted the channel in the longest time.

The relationship between time to traverse the valley width and time to avulsion is seemingly consistent for other lowland fluvial systems in the Gulf of Mexico region. For example, lateral migration rates (\(\sim 1.4-1.6\) m yr\(^{-1}\) and \(\sim 50-100\) m yr\(^{-1}\)) and floodplain widths (\(\sim 3.7-7.6\) km and \(\sim 50-125\) km) are measured at the backwater transition for the Brazos and
Mississippi rivers respectively. Because migration rates are known to vary spatially, we use lateral migration rates measured at the backwater transition, as this is our area of interest in this study. The Brazos River exhibits avulsion timescales of $\sim$600-3,500 years [Taha and Anderson, 2008], while the Mississippi River is known to have avulsion timescales of $\sim$700-1,800 years [Frazier, 1967]. These floodplain widths and lateral migration values suggest a timescale of $\sim$600-5,400 for the Brazos River and $\sim$500-2,500 years for the Mississippi River to traverse their floodplains. Therefore the times to traverse the valley are also quite similar to the observed avulsion timescales. The viability of this relationship needs further testing for additional fluvial systems.
CONCLUSIONS

The Trinity incised valley stratigraphy, well-constrained by previous core and seismic studies, provides unprecedented constraint for a morphodynamic model that describes incised valley sediment filling during the latest Holocene transgression. The stratigraphy of the Trinity incised valley system is therefore linked to sedimentological processes operating in a fluvial-deltaic system, and applied to evaluate mechanistic incised valley sediment fill. The important attributes of this morphodynamic model are the dynamic boundary condition of base level rise, the consideration of the exchange of sediment between the channel and the floodplain, and the ability to constrain findings with the measured stratigraphy of the Trinity incised valley system.

By comparing model results that consider and do not consider a floodplain, it is clear that the floodplain significantly influences channel fill and predicted avulsion timescales. Consideration of floodplain sedimentation increases the time to avulsion and allows model results to be applied over stratigraphically meaningful timescales. Our mechanistic model explains sediment distribution and why incised valley sediment infill is sandy and lacks a significant proportion of mud due to lateral channel migration. The model therefore allows for a process-based understanding of incised valley stratigraphy.

Furthermore, the application of base level rise as a dynamic downstream boundary condition affects the depositional patterns of sediment whereby sediment deposition to the channel bed backsteps at a rate proportional to the rate of base level rise. The backstepping of sediment is linked to backstepping backwater hydrodynamics which ultimately affects the location of channel avulsions and time to set up avulsions.

If sediment transport processes are to be extrapolated to the stratigraphic record, Holocene and beyond, it is critical to account for long term processes such as floodplain
dynamics and base level rise in order to better predict fluvial-deltaic stratigraphic architecture.

The objective of this model is to build a framework whereupon other fluvial-deltaic systems that lack robust spatio-temporal constraints may be considered by a few boundary conditions. This model therefore may be adapted for many other systems.
Figure 1: The Trinity River basin is approximately 44,000 km$^2$ with an average water discharge of 730 m$^3$ s$^{-1}$, an average annual sediment discharge of 6.2 x 10$^6$ tons and a gradient of 1.6 x 10$^{-4}$ [Rehkemper et al., 1969; Anderson et al., 2004]. The USGS stream gauge station at Crockett, TX is used for water discharge analysis to constrain the model. Galveston Bay, where the Trinity River outlets, is enlarged in the red box and the location of the cross sections seen in Figures 3 and 4 are denoted as A-A' and B-B', respectively.
**Figure 2:** Paleogeographic map of the Stage 2 lowstand approximately 22-17 ka. During the Stage 2 lowstand, incised valleys extended across the continental shelf and were subsequently filled as sea level began to rise. Lowstand incised valleys are labeled as: RGV = Rio Grande, CV = Colorado, BV = Brazos, T/SV = Trinity-Sabine. (Modified from Anderson et al., 2004)
Figure 3: (See Figure 1 for location) Cross-sectional interpretation of facies and valley morphology derived from seismic data within Galveston Bay. The base of the valley is represented by a Stage 2 sequence boundary that formed during the last sea level lowstand. Terraces formed as the valley incised and offer a unique antecedent topography for the Trinity system. Age dating of sediments within Galveston Bay establish the timing of valley fill. The valley began filling approximately 17 ka as the Holocene transgression began. Flooding surfaces associated with increased base level rise and backstepping coastal facies are interpreted at -14 m (8.2 ka) and -10 m (7.7 ka) [Rodriguez et al., 2005].
Figure 4: (See Figure 1 for location) Axial dip section through Galveston Bay interpreted from seismic and drill cores showing core locations, age dates, and depositional facies. The facies backstep from south to north through time. Vertical aggradation rates of the fluvial facies are estimated in red (~1.4-1.5 mm yr⁻¹). (Figure modified from Anderson et al. [2008].)
Figure 5: Northern Gulf of Mexico sea level curve modified from Milliken et al. [2008] for 10 ka to present. Sea level was derived from bayline peat and swash zone deposits. Our model simulates sea level rise between 10,000 and 8,200 years B.P. with a rate of 4.2 mm yr$^{-1}$. The sea level curve for 20-10 ka is from Bard et al., 1996 and represents the last sea level lowstand (Stage 2, 20 ka) and transgression leading into the early Holocene.
Figure 6: Daily water discharge measurements collected at USGS stream gauge station 8065350 near Crockett, TX from 1964-2014. The formative discharge range (700-1500 m$^3$ s$^{-1}$) is demarcated with blue dashed lines. 1500 m$^3$ s$^{-1}$ is also the bankfull discharge. Intermittency is determined by the number of days within the formative discharge range and adjusted for the years that do not exhibit formative discharges.
Figure 7: Water discharge and stage height are plotted to determine the bankfull discharge $1500 \text{ m}^3 \text{ s}^{-1}$, demarcated with a blue dashed line. The bankfull discharge occurs where there is no longer significant change in the stage height.
Figure 8: Schematic channel cross section depicting the mechanisms and morphology of a laterally migrating channel across its floodplain. The thickness of the floodplain ($T_f$) is considered to be comprised of cohesive, fine grained, muddy sediments and is much thinner than the underlying non-cohesive, presumably sandy sediments.
Figure 9: Flow velocity is plotted in both time and space. Example points of how to read the outputs are denoted with *'s and labeled ($U =$ flow velocity, RK = river kilometers). The flow velocities are plotted from the onset of the model (0 years) to the time for avulsion ($T_A = 1,821$ years). The avulsion location ($L_A$) is drawn in with a dashed line and occurs in model year 1,821.
Figure 10: Flow velocity through time (0-1,821 model years) is shown for various locations along stream. Flow velocity decreases downstream at each time within the model.
Figure 11: The extent of the backwater length through time is shown in colors associated with the time. Gray regions of the water surface are not within the backwater region. The backwater region migrates upstream and increases its length through time.
Figure 12:
(A): Shear stress (black line, left y-axis) and sediment load (magenta line, right y-axis) are plotted through space at the time for avulsion ($T_A = 1,821$ yrs). The avulsion location is denoted by a red dashed line ($L_A = 84$ km).
(B): Bed elevation (black lines) and water surface elevation (blue lines) are plotted through space at the time for avulsion ($T_A = 1,821$ yrs). The base level rise rate is $4.27$ mm yr$^{-1}$ and the difference in water surface elevation due to this change can be seen between the dashed and solid blue lines. Similarly, the difference in the initial and final bed elevations can be seen between the dashed and solid black lines.
Figure 13: Four model runs were selected to portray the effects of base level rise rate on sediment wedge morphology and avulsion time. Water discharge was held constant at 1500 m$^3$ s$^{-1}$ and the time for avulsion is based on a threshold of 0.3H.

(A): No base level rise, or static boundary condition scenario. The sediment wedge progrades downstream and reaches the avulsion threshold after 487 model years.

(B): 2 mm yr$^{-1}$ base level rise scenario. The sediment wedge progrades downstream and reaches the avulsion threshold after 707 model years.

(C): 4 mm yr$^{-1}$ base level rise scenario. The sediment wedge progrades then backsteps and reaches the avulsion threshold after 1,539 model years.

(D): 5 mm yr$^{-1}$ base level rise scenario. The sediment wedge progrades then backsteps and the system does not avulse within the lower 200 km after 5,000 model years.
Figure 14: The location of maximum change in bed elevation through time $\frac{\partial \eta}{\partial t}$ is plotted to track the locus of sediment deposition. Times of sediment deposition, accounting for horizontal distance, are indicated by a positive slope and dashed line. Times of sediment deposition backstepping are indicated by a positive slope and solid line.

Location of maximum sediment deposition through time $\frac{\partial \eta}{\partial t}$

**Backstepping**
26 m/yr
$(y = 0.026x + 44.837)$

**Prograding**
12 m/yr
$(y = -0.012x + 51.404)$

- $Q_w = 1500 \text{ m}^3/\text{s}$
- $Q_s = 0.12 \text{ m}^3/\text{s}$
- $S = 1.6 \times 10^{-4}$
- $BL = 4.27 \text{ mm/yr}$
- Avulsion threshold = 0.3H
- $T_H = 1821 \text{ yrs}$
**Figure 15:** Comparison of time for avulsion ($T_A$) and avulsion location ($L_A$) results from *Equation 3* (no floodplain, black squares, black line) and *Equation 4* (floodplain, red x’s, red line). No avulsions occur above 4.5 mm yr$^{-1}$ base level rise rate when the floodplain is included. No avulsions occur above 6.5 mm yr$^{-1}$ base level rise rate when the floodplain is excluded. Excluding the floodplain creates constant $T_A$ and $L_A$, regardless of base level rise rate.
Figure 16: Time for avulsion is plotted under varying base level rise rates and avulsion thresholds. The avulsion threshold is expressed as a fraction of the channel flow depth and the time for avulsion ($T_a$) is the time it takes the sediment wedge to aggrade to the avulsion threshold. The last point of each avulsion threshold line is the last occurrence of an avulsion under the given $Q_w$, $Q_s$, and base level rise rate parameters. The 0.3H line is equivalent to the 1500 m$^3$ s$^{-1}$ line of Figure 10.
Figure 17: Time for avulsion is plotted under varying base level rise rates and water discharges. The last point of each plotted line is the last occurrence of an avulsion under the given $Q_w$, $Q_s$, and base level rise rate parameters.
Figure 18: Rouse profiles [Rouse, 1937] of suspended bed material and washload sediment concentration throughout the water column. The bankfull flow depth is modeled as 3 m. The upper 2 m of flow depth are considered to be overbank flow with concentrations of $1.8 \times 10^{-3}$ bed load and $4.3 \times 10^{-3}$ washload to be exported to the floodplain.
REFERENCES


