ORIGINAL ARTICLE



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Impacts of backwater hydrodynamics on fluvial-deltaic stratigraphy

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Funding information

Geological Society of America, Grant/ Award Number: ExxonMobil/GSA Student Geoscience Grants; American Association of Petroleum Geologists, Grant/Award Number: David Worthington Named Grant and Meckel Family Named Grant ; International Association of Sedimentologists, Grant/Award Number: Post-Graduate Research Grant

Abstract

The hydrodynamics of rivers approaching a receiving basin are influenced by the onset of backwater conditions that give rise to decelerating reach-average flow velocity and decreasing boundary shear stress. These changes occur across a spatial gradient over which decreasing sediment transport capacity triggers morphodynamic responses that include sediment deposition at the transition from uniform to nonuniform flow. As a consequence, the channel width-to-depth ratio and bed sediment grain size decrease downstream. While nonuniform flow and associated morphodynamic adjustments have been investigated in modern fluvial-deltaic systems, the impacts to fluvial-deltaic stratigraphy remain relatively unexplored. This represents an important unresolved gap: there are few contributions that link morphodynamic response to nonuniform flow, impacts on sediment deposition and influence on the rock record. This study uses a numerical model to explore how variable channel hydraulics influence long-term (1000s years) patterns of sediment deposition and development of stratigraphy. The model results indicate that: (a) nonuniform flow propagates upstream beyond the backwater transition that is traditionally estimated with a basic backwater length scale relationship. (b) Base-level fluctuations, especially rising, enhance the impact of nonuniform flow. (c) Sediment deposition shows large spatio-temporal variability, which ultimately contributes to unique stacking patterns of fluvial-deltaic stratigraphy. (d) Nonuniform flow imparts spatial variation in flow depth, channel bed slope and sediment grain size over the delta, and these signatures are potentially preserved and recognizable in the rock record.

KEYWORDS

backwater morphodynamics, fluvial-deltaic stratigraphy, nonuniform flow

1 | **INTRODUCTION**

Traditional basin filling models that predict sedimentary stratigraphy do not account for the effects of nonuniform flow near the river mouth. For example, geometric models typically assume depositional profiles (i.e., fluvial gradient) to be constant and stratigraphy to manifest primarily with allogenic (external) influences, including eustatic adjustments, tectonic influences and/or climate effects (Jervey, 1988; Perlmutter, Radovich, Matthews, & Kendall, 1998; Ross, Watts, & May, 1995). Alternatively, dynamic models incorporate adjustments to the fluvial profile and consider responding variations in sediment flux, but are diffusion based and therefore do not predict how downstream variations in flow hydraulics impact sediment

Basin Research. 2019;00:1-18.

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deposition (Hutton & Syvitski, 2008; Muto & Swenson, 2005; Voller & Paola, 2010; Wu et al., 2019). In both cases, the purpose of the models is to assess how allogenic control influences stratigraphy over time scales of 10^3 – 10^7 years (Paola, 2000). Recent research, however, has documented that autogenic behaviour—characterized by internal dynamics inherent to the sediment transport and dispersal system—also drive important changes in sediment depositional patterns (Hajek & Straub, 2017), particularly over relatively short time scales (i.e., 10^0 – 10^3 years; Kim, Paola, Swenson, & Voller, 2006; Jerolmack & Paola, 2010; Wang, Straub, & Hajek, 2011).

Research efforts over the past decade have focused on disentangling autogenic and allogenic signals in stratigraphy, and advances in this cause have been achieved via detailed experimental investigations documenting how strata evolves under a variety of influences (Jerolmack & Paola, 2010; Li, Yu, & Straub, 2016). Yet, the nature of experiments renders scaling flow hydrodynamics to natural systems challenging, and so there remains a critical gap in terms of assessing how nonuniform flow-a critical autogenic process as recognized in modern fluvial-deltaic systems-manifests in the stratigraphic record. The focus of this study is to explore the impacts of variable channel hydraulics on the stratigraphic record through numerical modelling and to provide new insights into interpreting paleohydraulics within sedimentary stratigraphy. Specific questions addressed here include: (a) how does nonuniform flow influence the development of fluvial-deltaic stratigraphy? (b) What signatures of nonuniform flow are expected to be preserved in the rock record? And (c) can the signals of nonuniform flow be distinguished in the rock record?

2 | BACKGROUND: NONUNIFORM FLOW IN SEDIMENTARY DISPERSAL SYSTEMS

Nonuniform hydrodynamic flow conditions, known as "backwater flow", develop where a river approaches a standing body of water, such as a lake or ocean, and the depth-averaged water velocity decelerates in the downstream direction (Lane, 1957; Nittrouer, Shaw, Lamb, & Mohrig, 2012). This condition is characterized by a water surface profile that asymptotically approaches mean base level. As the bed slope maintains a constant value downstream, a spatial deviation between the water surface and bed profiles produces an increase in the width-normalized cross-sectional flow area, and for a uniform discharge, velocity decelerates (Figure 1). The development of backwater is important for sediment dispersal because a commensurate decrease in boundary shear stress reduces the sediment transport capacity of the flow, thus promoting sediment deposition.

Highlights

- Fluvial-deltaic stratigraphy is affected by nonuniform flow upstream of where traditional backwater scaling relationships estimate.
- Base level fluctuations enhance the impact of nonuniform flow on fluvial-deltaic morphodynamics, ultimately contributing to unique stratigraphic stacking patterns.
- Spatial variability in flow depth, channel bed slope, and sediment grain size due to nonuniform flow could potentially be preserved and recognized in the rock record.

Research pertaining to backwater influence on sediment transport initiated within the civil engineering community with theoretical developments followed by morphodynamic models (e.g., Chow, 1959; Hotchkiss & Parker, 1991; Parker, 2004). Later, field investigations of modern river systems validated these efforts. Recently, research shifted to identifying the stratigraphic signature of backwater hydrodynamics by evaluating stratigraphic patterns in rock outcrops. A brief overview of this work is provided.

Morphodynamic models provide insights into backwater impacts on sediment transport, for conditions of both static and a changing base level. Parker, Muto, Akamatsu, Dietrich, and Lauer (2008a), Parker, Muto, Akamatsu, Dietrich, and Wesley Lauer (2008b) quantified the alluvial auto-retreat process associated with relative sea-level rise since the Last Glacial Maximum. This work explored how backwater and base-level adjustments combine to influence sediment deposition patterns. Moran, Nittrouer, Perillo, Lorenzo-Trueba, and Anderson (2017) developed a morphodynamic model for the Trinity River (Texas) by considering Holocene sea-level rise. This work documents a consistent back-stepping of the backwater zone, accompanied by bed material sediment filling of the Trinity River incised valley.

Within modern systems, backwater flow was demonstrated via direct measurements, which were subsequently used to inform and validate morphodynamic models of the lowermost Mississippi River (e.g., Lamb, Nittrouer, Mohrig, & Shaw, 2012; Nittrouer, Mohrig, & Allison, 2011a; Viparelli, Nittrouer, & Parker, 2015). This work has led to an understanding of feedbacks in terms of sediment deposition patterns and impact on channel geometry. For example, Nittrouer et al. (2012) document a changing channel geometry across the backwater transition of the lowermost Mississippi River, whereby channel width-to-depth ratio and median channel sediment grain size decrease by a factor of 3.



Distance upstream of river mouth (x)

FIGURE 1 Schematic showing the transition to nonuniform (backwater) flow for a fluvial-deltaic system, as indicated by channel bed and water surface profiles. Note that the profiles are parallel in the normal flow reach, and deviate in the backwater reach, where the water surface profile asymptomatically approaches base level at the river outlet. This adjustment produces a downstream increase in flow depth that, for uniform water discharge, results in a decrease of flow velocity and boundary shear stress. These adjustments trigger morphodynamic feedbacks (modified from Lamb et al., 2012)

Flume experiments have accounted for backwater flow (Ganti, Chadwick, Hassenruck-Gudipati, Fuller, & Lamb, 2016a; Ganti, Chadwick, Hassenruck-Gudipati, & Lamb, 2016b). This work documented a preferential location for deltaic channel avulsions arising near the onset of backwater due to the co-location of maximum bed deposition. This finding is consistent with observations from the Mississippi River (Nittrouer, Mohrig, Allison, & Peyret, 2011b). Additionally, this work confirmed the notion of a "channel-lobe life cycle", whereby a channel conveys water until backwater sedimentation produces an avulsion and lobe abandonment (Nittrouer, 2013). The "life expectancy" of a channel is, to first order, related to system slope, sediment supply and flow depth; for natural systems, this time scale ranges 10^{1} – 10^{3} years (Ganti, Chu, Lamb, Nittrouer, & Parker, 2014).

Stratigraphy as influenced by backwater flow is tentatively identified in the rock record. Hints of such linkages are inferred by Petter (2010), who reports measurements of grain size and paleochannel flow depth for the Lower Castlegate Sandstone (Utah, USA), along a down-dip transect, between the upstream fluvial segment to the marine interface, spanning a 25-km segment that approximates the estimated length of the backwater reach. Petter reports a downstream fining of sediment grain size coinciding with increasing bankfull depth. This suggests a spatial partitioning of sediment due to the extraction of coarse sediment, which is consistent with the hypothesis that downstream decreasing sediment transport capacity produces selective deposition. Petter interprets these paleohydrodynamic trends as arising because of the influence of backwater flow. This interpretation is further supported by patterns of channel bed scours, which show consistent scaling relationship between autogenic scour depth and length scales as backwater hydrodynamic model predicted (Trower, Ganti, Fischer, & Lamb, 2018).

Colombera, Shiers, and Mountney (2016) qualitatively document a relationship between backwater flow and distributary channel fill architecture for an interval of the Campanian Neslen Formation (Utah, USA). They hypothesize that sandprone, aggradational ribbon channel elements represent the terminal expression of distributary channels, where hydrodynamic drawdown produces multiple, clustered scour surfaces at the base of channel fills. More recently, based on analysis of subsurface data (seismic and well log), Martin et al. (2018) demonstrate downstream decreases of channel belt width and median channel sandstone grain size in the fluvial-deltaic system of the Mungaroo Formation, Carnarvon Basin (Australia) as evidence of the influence of nonuniform flow hydrodynamics. However, recognizing downstream deepening of paleoflow depth based on bar thickness is still challenging with subsurface data (Martin et al., 2018), and observations of downstream increase in bar thickness within the backwater reach of modern fluvial-deltaic systems (e.g., the Mississippi River, Fernandes, Törnqvist, Straub, & Mohrig, 2016) still require testing in the rock record.

Despite the importance of backwater hydrodynamics in shaping channel morphology and sediment dispersal patterns in fluvial-deltaic systems, impacts on the stratigraphic record remain unclear. Importantly, there is a need to constrain how autogenic processes co-evolve with allogenic controls arising over similar time scales. The gap between backwater morphodynamics and stratigraphic signatures is in part due to the differences of time scales over which geomorphologists and stratigraphers typically account for sediment transport and accumulation. As described previously, the time scale for backwater influence, as identified by firstorder impacts to fluvial-deltaic morphodynamics, is shorter than basin filling models, which tend to focus on first-order allogenic controls such as base-level changes. To account for this discrepancy, a numerical model that integrates channel hydraulics and long-term (> 10^3 -year scale) development of stratigraphy (i.e., shoreline movement, grain size variation and channel bed aggradation or degradation) is needed. The intention is to develop a morphodynamic model, and evaluate sensitivity to variability in allogenic controls, while exploring potential impacts on stratigraphy developed under the influence of nonuniform flow. The central hypothesis is that backwater morphodynamics leave distinct signatures in terms of variation of grain size on the channel bed and depth of paleoflow within the stratigraphic record, even with a dynamic boundary condition imposed by base-level adjustments.

3 | ANALYTICAL FRAMEWORK

The 1D numerical model of this study adopted the equations of conservation for the flow and the sediment transport from the morphodynamic models of Wright and Parker (2005) and Parker et al. (2008b) to simulate stratigraphic development under nonuniform flow hydraulics with minor modifications to account for mud deposition and floodplain development (Parker, 2004). A grain size-specific sediment transport relation of Naito et al. (2019) is used to account for the nonuniformity of the bed material grain size from the bedrockto-alluvial transition to the shoreline break. The profile of deltaic foreset is modelled as a shock condition with fixed foreset gradient (Parker et al., 2008a, 2008b) so that all of the sediment transported to the shoreline break is used to build the deltaic foreset. Grain size variation in the deltaic foreset is not modelled.

3.1 | Numerical modelling of backwater morphodynamics

The balance of flow momentum is treated using a backwater formulation, which provides a means to model bankfull flow depth (H_{bf}) in the along-stream distance (*x*):

$$\frac{dH_{bf}}{dx} = \frac{S - C_f F r^2}{1 - F r^2},$$
 (1)

where *S* is the channel bed slope, C_f is friction coefficient, *Fr* is Froude number, determined by $Fr = U (gH_{bf})^{-0.5}$, with *U* the depth-averaged flow velocity and *g* gravitational acceleration. Depth-averaged flow velocity may be estimated by: $U = q_w/H_{bf}$, where q_w is the width-averaged water discharge and can be calculated by: $q_w = \sqrt{gH_{bf}^3S/C_f}$ (Parker, 2004).

A morphodynamic feedback is provided by modelling sediment transport (q_s) over the domain. Here, the bed material load equation of Ma et al. (2017) is used:

$$q_s = \left(RgD_{50}^3 \right)^{1/2} \left(\frac{0.0355}{C_f} \right) \tau_*^3, \tag{2}$$

where D_{50} is the median grain size, R is the submerged specific gravity of sediment particle and τ_* is the nondimensionalized shear stress, determined by $\tau_* = \rho C_f U^2 / Rg D_{50}$.

Channel bed elevation through time $(\partial \eta / \partial t)$ is assessed through mass conservation of bed material load using a modified Exner equation (Paola & Voller, 2005):

$$\left(1-\lambda_p\right)\frac{\partial\eta}{\partial t} = -\frac{(1+\Lambda)\,\Omega I_f}{r_B}\frac{\partial q_s}{\partial x},\tag{3}$$

where λ_p is the mean porosity of the channel-floodplain complex, Λ is the mud/sand deposition ratio, whereas mud availability is assumed to be infinite, Ω is the channel sinuosity, I_f is the flood intermittency and r_{B} is the ratio between width of flood plain and channel width. Since the 1D model is widthaveraged and width changes in the along-stream direction for the channel and floodplain is not considered, the inclusion of Λ , Ω , I_f and r_B in the model helps to account for deposition in floodplain and flood intermittency of natural rivers so that the model simulation has application to real rivers (Parker, 2004). Short of measuring these parameters from field data, standard values from modern lowland systems that scale with the ancient basins are utilized (Table 1) (Moran et al., 2017; Parker et al., 2008b). The channel bed profile evolves iteratively with each time step by solving Equations (1)–(3), over the horizontal distance (x).

The model inputs include slope, water discharge, sediment size and flow depth. Additionally, the model incorporates moving boundaries to account for shoreline migration associated with channel extension (delta progradation) and/ or base-level changes. The moving boundaries are tracked using a deforming grid with a constant number of computational nodes. Discretization and flow of the calculation of the numerical model follow the method of Parker et al. (2008b). Model run time is 5,000 years with a 1-year time step. The run time is chosen so that the influence of nonuniform flow will co-evolve with longer term allogenic controls on the system, in particular, base-level adjustments (Moran et al., 2017).

Variable	Value	Description
S	2×10^{-4}	Initial channel bed slope
S _b	4.6×10^{-4}	Slope of subaerial bed rock reach
S_{f}	1.6×10^{-4}	Slope of deltaic foreset
S _{sb}	0	Slope of subaqueous basement
g	9.81 m/s ²	Gravitational acceleration
I_{f}	0.1	Flood intermittency
R	1.65	Submerged specific gravity of sediments
λ	0.4	Bed porosity
Ω	1.7	Sinuosity
Λ	1.0	Volume unit of mud depos- ited in the channel–flood- plain complex per unit sand deposited
r _B	60.0	Ratio of channel width to flood plain width

TABLE 1Key model input parameters

Fluvial–deltaic avulsions could arise due to the morphodynamic feedbacks associated with nonuniform flow (Ganti, Chadwick, Hassenruck-Gudipati, & Lamb, 2016b; Moran et al., 2017; Nittrouer et al., 2012). A distributary channel avulsion represents the cessation of lobe deposition; typically, this gives rise to reworking of the deposit (e.g., due to the influence of waves and tides). This study does not model such conditions, and instead focuses broadly on stratigraphy developed for a deltaic system, which over time includes multiple lobe cycles (Ganti, Chadwick, Hassenruck-Gudipati, & Lamb, 2016b). Avulsion and lobe abandonment are therefore not specified in in the model. This provides the ability to focus on the impacts of nonuniform flow on depositional stratigraphy of the delta.

3.2 | Grain size model

Channel bed grain size is calculated using the "surface-based" method (Naito et al., 2019; Wright & Parker, 2005). The "surface-based" method assumes that the channel bed can be subdivided into a surface layer (active layer) and substrate. In this method, grain size-specific transport rate is a function of grain size distribution of the active layer (Naito et al., 2019):

$$q_i = F_i \frac{u_*^3}{C_f Rg} \left[A_i \left(\tau_i^* \right)^{B_i} \right], \tag{4}$$

where q_i is the bed material transport rate of sediment in the *i*th grain size range, F_i is the volume fraction content of the *i*th grain size range in the active layer, u_* is the shear velocity, τ_i^* is the Shields number for the *i*th grain size Basin Besearch Seage -WILEY 5

range, A_i and B_i are the coefficients constrained by field data (Naito et al., 2019). Grain size distribution of total bed material load is calculated based on a bedload transport relation (Equation 4) for each grain size group and updated using the following grain size-specific formulation (Naito et al., 2019):

$$f_{i} = \frac{F_{i}A_{i}\left(\tau_{i}^{*}\right)^{B_{i}}}{\sum\left[F_{i}A_{i}\left(\tau_{i}^{*}\right)^{B_{i}}\right]},$$
(5)

where f_i is the volume fraction content of the *i*th grain size range in the total bed material load. The exchange of sediments between active layer and substrate is calculated using (Naito et al., 2019):

$$(1 - \lambda_p) \left[L_a \frac{\partial F_i}{\partial t} + (F_i - F_{Ii}) \frac{\partial L_a}{\partial t} \right]$$

$$= -\frac{(1 + \Lambda) \Omega I_f}{r_B} \left(\frac{\partial (q_s f_i)}{\partial x} - F_{Ii} \frac{\partial q_s}{\partial x} \right),$$
(6)

where L_a is the active layer thickness and F_{Ii} is the volume fraction content of the *i*th grain size range at the interface between substrate and the active layer. The median grain size of total bed material load is calculated based on the grain size distribution simulated using Equation (5) and updated every time step during the iteration of Equations (1–3). Channel bed aggradation is realized through transfer of sediment from the active layer to the substrate, where the fraction of coarser grain size groups in the active layer decreases as this material is more easily deposited than the finer grain size groups (Cui, Paola, & Parker, 1996; Naito et al., 2019; Wright & Parker, 2005). As a result, the fraction of coarser grain size groups decrease so that the median grain size of both the active and substrate layer decreases.

4 | MODEL RESULTS

Four model scenarios were tested to investigate how stratigraphy patterns vary under the influence of nonuniform flow for different basin physiography (i.e., basin depth), and with and without base-level change. The model runs are referred to as "Case 1", "Case 2", "Case 3" and "Case 4". The initial shoreline break is located at 0 km. The initial channel bed surface slope is set at 2×10^{-4} with the downstream boundary (shoreline break) set at an elevation of -7 m. For simplicity, the air–water interface at the outlet is set to 0 m at the beginning of each model run, which results in an initial flow depth at shoreline break of 7 m. Tectonic forcing (e.g., differential subsidence and tectonic tilting) is not considered for simplicity, thus base level is essentially equivalent to sea level in the AS EAGE

model. Slope of the deltaic foreset is 1.6×10^{-3} , and the slope of the subaqueous basement, onto where the fluvial-deltaic system builds, is zero. Upstream flow velocity (U) is set at 2 m/s. Normal flow conditions persist at the upstream end, and so flow depth is calculated by combining the depth-slope product ($\tau_{h} = \rho gHS$) and the relation $\tau_{h} = \rho C_{f} U^{2}$. Grain size distribution through the channel bed was divided into 11 bins (31.3, 44.2, 62.5, 88.4, 125, 176.8, 250, 353.6, 500, 707.1 and 1,000 µm) and set as normally distributed with an initial median grain size D_i of 177 µm and a standard deviation of 14.4 µm. Key model input parameters are listed in Tables 1 and 2. Case 1 is a base case simulation that shows stratigraphy development with a constant base level and an initial basin depth of 30 m below sea level (Figure 2). Case 2 and Case 3 simulations share the same modelling parameters as Case 1 except for conditions when base level rises at a rate of 10 mm/year in Case 2 and falls at a rate of 3 mm/year in Case 3. These values were chosen to represent extreme rates of sea-level change documented and estimated from the geological records (Cutler et al., 2003; Lambeck, Rouby, Purcell, Sun, & Sambridge, 2014; Miller et al., 2005). Case 4 maintains the same parameters as Case 1 except for a deeper basin (depth of 150 m below sea level).

4.1 **Development of stratigraphy**

The cross-sectional view of the modelled stratigraphy is illustrated with down-dip profiles of channel bed and deltaic foresets shown for 1,000-year increments. All model cases show stratigraphic development through channel bed adjustment and shoreline movement (e.g., progradation, retrogradation, rising and falling).

The channel bed profile in Case 1 is concave upward, which is enhanced throughout the model simulation. Upstream migration of the alluvial/bed rock transition (upstream boundary of the modelled fluvial reach) is neglectable when compared to the progradation of the shoreline break. The extent of the fluvial reach was thus lengthened as the shoreline prograded. The shoreline elevation lowers slightly through the model run. The overall stratigraphy shows an upward fining of median grain size (Figure 3a and e). Median grain size of the channel bed also shows downstream fining throughout the run (Figure 3e).

Case 2 shows stratigraphic development through channel bed aggradation and shoreline movement. Progradation is rapid at the start of Case 2 and gradually aggradation occurs

TABLE 2 Specified input parameters for each model case

	Case 1	Case 2	Case 3	Case 4
Rate of base level change (mm/yr)	0	10	-3	0
Basin depth (m)	30	30	30	150

so that the location of the shoreline break tends to be fixed in space and the size of the deltaic foreset is nearly static for the last 1,000 years of the simulation. The channel bed profile is concave upward but developed some convexity at the downstream reach at the end of the model run. The overall stratigraphy shows an upward fining of median grain size (Figure 3b and f). Median grain size of the channel bed also fines towards downstream and the grain size at shoreline break reduced to less than 20% of the initial grain size (Figure 3f).

Case 3 shows stratigraphic development through channel bed aggradation and shoreline movement (Figure 2). The shoreline break prograded much farther compared to Case 1 and Case 2, while falling about 20 m. Even under a high rate of base-level fall, the channel bed still aggraded and sustained its concave upward profile. The height of the deltaic foresets diminished throughout Case 3 to nearly a value of zero by the end of the simulation. Channel bed grain size at downstream reach fined throughout the simulation (Figure 3c and g). Upward coarsening of channel bed developed around the location of the initial shoreline break (x = 0). The overall stratigraphy shows a fining-upward trend upstream of the initial shoreline break, and a coarsening-upward trend downstream of the initial shoreline break.

Both channel bed aggradation and shoreline progradation in Case 4 were much less distinct compared to Case 1-Case 3. Development of deltaic foresets was more prominent than fluvial stratigraphy in Case 4 as the depth of basin is 5 times that of Cases 1-3, while the channel bed aggrades less than 10 m. Median channel bed grain size shows both downstream and upward fining trend through the simulation but to a less extent compared to Cases 1-3 (Figure 3d and h).

4.2 Variability of channel hydraulics

The initial downstream flow changes for Cases 1-4 were small (from normal flow depth of 6.88 m to a depth of 7 m at the river mouth) (Figure 4), representing the imposed initial boundary conditions. However, flow depth at the river mouth increased through the simulations for all the cases, thus the downstream deepening became distinct as the simulations proceeded. By the end of each of the cases, the signature of nonuniform flow hydraulics is quite evident for Cases 1, 2 and 3. Case 1 ended with a 150% increase in flow depth near the river mouth. Case 2 shows an almost two-fold increase in flow depth at the end of the simulation. The flow depth variation in Case 3, although under a constant base-level fall, shows a similar degree of downstream deepening as Case 1. Case 4 shows the least downstream increase (less than 130%) in flow depth among all the simulations.

The initial channel bed slope for all the cases is fixed and constant through the fluvial reach (Figure 5). The channel bed slope at the river mouth decreased rapidly for all simulations, resulting in a downstream decrease in the channel bed slope.



FIGURE 2 Four simulations (Cases 1 through 4) show fluvial-deltaic stratigraphy development through nonuniform flow. Case 1 is a base case where base level was held constant. Cases 2 and 3 model stratigraphic development under base level rise and fall at constant rates of 10 and 3 mm/yr, respectively. Case 4 shows stratigraphic development with constant base level and a deeper basin. Blue lines mark the final flow surface profile and base level

Case 1 ended with a 77% downstream decrease in channel bed slope. Case 2 shows a 70% downstream decrease in channel bed slope from the start to around 2,500 years of simulation. For the second half of the simulation, channel bed slope at river mouth increased to about 50% of the initial value. Case 3 also ended with a 77% downstream decrease in channel bed slope. The downstream decrease of channel bed slope in Case 4 is less than 50%. Temporal changes in channel bed slope show a decreasing trend upstream of the location of the initial shoreline break (0 km) and an increasing downstream trend for Cases 1 and 3. For Case 2, temporal changes in channel bed slope show a dominant decreasing trend. Temporal variation in channel bed slope in Case 4 is very small as evidenced by the nearly vertical contour lines in Figure 5.

Onset of nonuniform flow 4.3

The onset of nonuniform flow is estimated to assess the applicability of backwater length scales for characterizing the extent of nonuniform flow. The transition from uniform to nonuniform flow is estimated to be the location (x_t) , where along-stream changes in flow depth (dH/dx) is less than 5×10^{-6} . This threshold, proposed by Moran et al. (2017), is significantly smaller than the original channel bed slope (2×10^{-4}) at the start of each model run. Upstream of this transition, the along-stream changes in flow depth are considered negligible so that the flow can be treated as uniform. The extent of nonuniform flow (L_{bw}) is calculated as the distance between the transition to the



FIGURE 3 (a–d) Modelled fluvial stratigraphy for the four cases. The stratigraphy is color-coded by median channel bed grain size D_{50} . (e–f) The spatiotemporal trend of grain size variation (normalized by initial median channel bed grain size). Solid black lines are contours of grain size. The white dashed line represents the location of normal to nonuniform flow transition x_t (see Section 4.3 for details). Grain size distribution in deltaic foresets were not modelled and thus are not plotted



FIGURE 4 Spatial and temporal variations in normalized flow depth (H^*). Flow depths (H) in all model cases are normalized by upstream uniform flow depth (H_n) to better portray the magnitude of change. Note that the colormap of each case has a different range. Solid black lines are contours of normalized flow depth. The white dashed line represents the location of normal to nonuniform flow transition x_t (see Section 4.3 for details)

shoreline break (x_{sb}) : $L_{bw} = x_{sb} - x_t$. The backwater length scale is traditionally approximated by L = H/S (Paola and Mohrig, 1996), where *S* is the channel bed slope and *H* is the bankfull flow depth. To account for the varying flow depth and channel bed slope through the backwater zone, the backwater length scale for Case 1–Case 4 is evaluated

using $L_n = H_n/S_n$ and $L_m = H_m/S_m$, where S_n is initial channel bed slope, H_n is the normal flow depth, S_m is the channel bed slope at river mouth and H_m is the bankfull flow depth at the river mouth. The locations of backwater transition are calculated for L_n and L_m as $x_n = x_{sb} - L_n$ and $x_m = x_{sb} - L_m$, respectively. Moreover, the location $(x_{\eta 0})$, where channel

FIGURE 5 Spatial and temporal variations in normalized channel bed (S^*). Channel bed slope (S) in all model cases are normalized by initial channel bed slope (S_i). Solid black lines are contours. The white dashed line represents the location of normal to nonuniform flow transition x_t (see Section 4.3 for details)



FIGURE 6 Temporal variations of the locations of transition from normal flow to nonuniform flow (x_t) , backwater transitions $(x_n \text{ and } x_m)$, and location where channel bed elevation intercepts base level $(x_{\eta 0})$. The black lines mark the shoreline locations

bed elevation intercepts base level, is also calculated for comparison with x_t , x_n and x_m .

The transition from uniform flow to nonuniform flow (x_t) migrates upstream for Case 1, Case 2 and Case 4 (Figure 6).

This transition (x_t) migrates upstream and then downstream for Case 3. The length of nonuniform flow extends as x_t and x_{sb} diverge for all simulations. Since L_n is fixed for all model runs as S_n and H_n are constants, the backwater transition x_n

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covaries with the location of the shoreline break (x_{sb}) . The backwater transition x_m , on the other hand, diverges from the shoreline break as S_m and H_m continuously change through the simulations (Figures 4 and 5). The location $(x_{\eta 0})$ where channel bed elevation intercepts base level co-evolves with the backwater transition x_m except for Case 2, especially during the second half of the model run as L_m decreases. This decrease in L_m is the result of increasing channel bed slope towards the shoreline break (Figure 5).

Research

The region of nonuniform flow extends beyond the backwater transition that is typically estimated with backwater length scales. Previous studies have mostly focused on backwater hydrodynamics over a short time scale (i.e., less than a millennium) when the backwater length scale was found to coincide with important morphodynamic changes in a fluvial-deltaic system (e.g., Fernandas et al., 2016; Ganti, Chadwick, Hassenruck-Gudipati, Fuller, et al., 2016a; Nittrouer et al., 2012). However, the backwater length scale manifests as a result of hydraulic changes (rather than causing the changes). And so, along-stream flow depth change (dH/dx) ultimately drives flow hydraulics, such as flow velocity, as well as sediment flux. Although previous studies show that the backwater length scale is strongly associated with changing along-stream flow depth (dH/dx) over short time scales, the modelling results of this study demonstrate that the two are mostly decoupled over millennia time scales. The onset of along-stream flow depth change (dH/dx) extends much farther upstream than the backwater length scale predicts, suggesting that nonuniform flow could potentially operate over a larger spatial extent than previously considered, especially during base-level rise (e.g., Moran et al., 2017; Parker et al., 2008b). Therefore, the hypothesized temporal changes in paleohydraulics signatures could be preserved in ancient systems, although they may not necessarily correspond to the backwater length scale calculated for a given system. The large extent of the influence of nonuniform flow, as demonstrated by the model results, could be further tested in ancient lowstand fluvial-deltaic systems which are confined within incised valleys across the continental shelf, extending over 100 km (e.g., Anderson et al., 2016), or in modern lowland deltaic systems.

5 | DISCUSSION

5.1 | External influences on the development of nonuniform flow

The results here show that nonuniform flow developed due to a backwater effect is essentially an autogenic process that is modifying the longitudinal river profile and may co-evolve with external controls. For Case 1, where base level was held constant (steady allogenic control), the key for the development of nonuniform flow is the difference in flow depth between upstream normal flow and at the river mouth. Since this difference in flow depth sets the initial boundary condition for the model, the nonuniform flow condition persists throughout the simulation and propagates upstream. Case 2 shows that the downstream deepening of flow depth is enhanced as the flow depth at the shoreline break increases due to base-level rise, resulting in more distinct nonuniform flow. In Case 2, the deepening trend propagates much farther upstream than Case 1 and results in a greater magnitude of downstream deepening.

The model result of Case 3 shows that under base-level fall, channel bed aggradation could persist without incision if downstream deepening of flow depth is sustained through the development of nonuniform flow. As the shoreline break progrades in Case 3, the height of the deltaic foreset decreases from 23 to 1.5 m. Therefore, less sediment is required to build deltaic foreset. As a result, the shoreline break progrades much farther basinward. Although base-level fall tends to reduce the flow depth at river mouth, the decrease in the height of deltaic foreset helps to place the shoreline break into a lower and more distal location, so that the overall effect turns out to be an increase in flow depth at shoreline break and a greater magnitude of downstream deepening in Case 3. This case shows that base-level fall could potentially enhance the development of nonuniform flow (Figures 4 and 6), but the effect might be mitigated under particular patterns of base-level fall if river become graded (Muto et al., 2016).

The most significant difference between Case 4 and Case 1-Case 3 is the deeper basin in Case 4, which produces larger accommodation (Parker et al., 2008a) so that shoreline progradation is limited. Case 4 shows only subtle development of nonuniform flow. The magnitude of downstream deepening of flow barely changes as flow depth at shoreline break is almost constant (Figure 4). However, this does not mean that a deeper basin tends to prevent development of nonuniform flow, but rather it shows that a stagnant shoreline position may limit deepening of the flow depth at the river mouth. Nonuniform flow develops as long as there is a difference in flow depth between the upstream normal flow and river mouth. Once the nonuniform flow condition is established, it will persist even with relatively deep basin and the low rate of shoreline migration. Moreover, Case 4 can be interpreted to reflect a shelf-edge deltaic system, which is usually considered to be greatly influenced by high-frequency sea-level fluctuations near the shelf-slope break (e.g., Bowman & Johnson, 2014; Carvajal & Steel, 2009). Therefore, the development of nonuniform flow in a shelf-edge setting could potentially be enhanced through the cycles of base-level fluctuation (e.g., Parker et al., 2008b).

5.2 | Nonequilibrium river profile

The modelled river profiles never reach an equilibrium state in a traditional sense, whereby system maintains a graded and

<mark>≜§ eage</mark>—WILEY lasin Research FIGURE 7 (a) Root mean square (b) 1.6 (a) (RMS) of changes in channel bed elevation 2 through time (normalized by RMS of Case 1 cross-correlation 1.4 Case 2 channel bed elevation at modelling time step normalized Case 3 1). (b) Cross correlation between channel RMS* Case 4 1.5 bed elevation changes (normalized by cross 1.2 correlation of channel bed elevation at modelling time steps 1 and 2) 1 1 1.000 2.000 3.000 4.000 5.000 1.000 2.000 3.000 4.000 5.000 0 0 time (year) time (year)

static profile with neither aggradation nor degradation and all sediment input bypasses the system (Bijkerk et al., 2016; Muto & Swenson, 2005). This is particularly well demonstrated by flow hydraulics that vary continuously throughout all the model runs (Figures 4 and 5). This is not surprising, as the external conditions and basin physiography for all the model runs do not meet the well-known criteria for developing a graded river profile (Muto et al., 2016). Although the river profiles are not graded, the overall change and geometry of the profiles seem to reach a steady state for Case 2. For example, the root mean square (RMS) of the changes in channel bed elevation across the channel plateau towards the end of Case 2 simulation (Figure 7a), suggesting that the amount of change in channel profile is asymptotically approaching a constant. Moreover, the maximum cross-correlation between river profiles at adjacent time steps also show the same trend, suggesting that the similarity between river profiles become constant (Figure 7b). In other words, the river profiles become self-similar, a condition that characterizes autoretreat, which is a typical fluvial response to base-level rise (Parker et al., 2008a). Cases 1 and 3 do not seem to reach such steady state, thus the river profiles are continuously evolving under the given external conditions. Case 4 shows the tendency to reach the steady state, as both RMS of the changes in channel bed elevation and maximum cross-correlation between river profiles decrease towards the end of the model run (Figure 7). This can be interpreted as the "forced grade" case due to the greater basin depth (Muto et al., 2016).

5.3 Influence of nonuniform flow on the development of stratigraphic patterns

5.3.1 **Channel bed aggradation**

The development of stratigraphic patterns in this study is a result of a changing longitudinal river profile, which evolves directly through channel bed aggradation. Channel bed aggradation has been recognized as an important fluvial response to nonuniform flow hydrodynamics (Ganti et al., 2016b; Nittrouer et al., 2012). Much of the previous work was focused on understanding the role of channel bed aggradation in avulsions that lead to development of delta lobes (Chatanantavet, Lamb, & Nittrouer, 2012; Ganti et al., 2016b). Spatial and temporal variabilities in channel bed aggradation, on the other hand, has been less explored. This study shows that the degree and extent of channel bed aggradation resulting from nonuniform flow is dynamic over millennium time scale but is crucial for the development of stratigraphy.

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The four model cases show a wide range of stratal stacking patterns from progradational (Cases 1, 3 and 4) to aggradational (Cases 1-4) to retrogradational (Case 2). These differences can be evaluated using spatio-temporal variations in the rate of aggradation (Figure 8a-d). Case 1 shows very high rate of channel bed aggradation at the shoreline break at the start of the simulation, with a peak rate decreasing afterwards. The downstream increase in the rate of channel bed aggradation becomes more gradual as the overall rate across the channel bed decreases and becomes uniform near the downstream end, which can be inferred from the spreading of the contour lines. The peak rate of channel bed aggradation at the river mouth indicates that a large amount of sediment was transported through the fluvial system so that the shoreline break progrades. Case 2 also shows a high rate of channel bed aggradation at the shoreline break during the beginning of the simulation. As the peak rate of channel bed aggradation at the shoreline break decreases, a local depositional peak develops around 50 km upstream of the shoreline break (Figure 8b). Case 3 generally shows similar spatio-temporal changes in the rate of channel bed aggradation as Case 1. Case 3 shows that falling shoreline trajectory is not necessarily associated with channel bed degradation and channel bed aggradation continues under base-level fall due to persistent backwater hydrodynamics (Figures 4c and 5c). A local peak of aggradation develops around 100 km downstream of the original shoreline break. In Case 4, the peak rate of aggradation remains at the shoreline break and decreases throughout the



FIGURE 8 (a–d) Spatial and temporal variations in rate of channel bed aggradation (mm/yr). Note that the colormap of each case has a different range. (e–f) Total thickness of the fluvial stratigraphy at the end of model runs

model run. The rate of channel bed aggradation in Case 4 is less compared to other cases due to less developed nonuniform flow (Figures 4d and 5d) and, hence, the thickness of fluvial stratigraphy is limited (Figure 8h). This also suggests that much of the sediment input bypassed the fluvial system and is delivered to the deltaic foreset.

A higher rate of channel bed aggradation is found upstream of the river mouth in Cases 1–3 and results in an increase, then decrease in the thickness of fluvial stratigraphy from the upstream model boundary to the river mouth (Figure 8e–h). Similar trends have also been observed in ancient fluvial– deltaic systems considered to be influenced by backwater effects (Martin et al., 2018). The spatio-temporal variability in channel bed aggradation reflects the diffusive and advective nature of the sediment transport processes, which is evidenced by the upward concave river profiles (Parker et al., 2008b; Swenson, Paola, Pratson, Voller, & Murray, 2005; Wright & Parker, 2005). The upstream local peaks in channel bed aggradation seem to be persistent throughout Cases 1–3 and whether they can contribute to avulsion set-up still needs further investigation.

5.3.2 | Shoreline migration

The location and movement of the shoreline break has significant implications for the architecture and stacking pattern of fluvial–deltaic stratigraphy. From a modelling perspective, the shoreline break sets the downstream boundary of

diffusion-based fluvial-deltaic models (e.g., Parker et al., 2008b; Swenson et al., 2005; Wright & Parker, 2005) so that it determines the extent of the fluvial profile. The extent and shape of fluvial profile is crucial for sediment transport and deposition, and thus influences the resulting stratigraphy. In this study, the rate of shoreline progradation rapidly decreases at the model start and gently decreases for the remaining simulation for Case 1 and Case 2 (Figure 9). The rate of shoreline progradation in Case 2 decreases to zero and then turns negative, indicating shoreline transgression; however, the rate is low enough that the shoreline break is nearly fixed in space around 100 km downstream of original shoreline break (Figure 6b). Case 3 shows an overall increase in the rate of shoreline progradation with the rate accelerating towards the end of the model run as the deltaic foreset diminishes. The rate of shoreline progradation in Case 4 is low compared to other cases and nearly constant throughout the simulation.

The rate of shoreline migration is determined by the amount of sediment supplied to the river mouth and how fast this sediment is consumed to build deltaic stratigraphy. Therefore, a shallow basin (thin deltaic foresets) will favour the migration of the shoreline break (e.g., Case 3) as less sediment is needed to build the deltaic foreset. On the contrary, a deep basin (thick deltaic foresets, e.g., Case 4) and more developed nonuniform flow condition (e.g. Case 2) will limit the sediment supply to the river mouth and thus slow the migration of the shoreline break (Bijkerk et al., 2016; Carlson, Piliouras, Muto, & Kim, 2018). Furthermore, since the rate of shoreline migration is dimensional in this study, it can be compared with some modern and ancient deltaic



FIGURE 9 Temporal trends in the rate of shoreline progradation (m/yr)

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systems. The modelled rate is significantly higher than estimates for some ancient systems (e.g., Goodbred & Kuehl,

2000; Rodriguez, Hamilton, & Anderson, 2000; Wolinsky, Swenson, Litchfield, & McNinch, 2010; Zhang, Ronaldm, & William, 2016). Possible explanations of this discrepancy include: (a) the two-dimensional nature of the model means that sediment is not dispersed laterally (e.g., via lobe switching and alongshore transport). (b) Rate of shoreline migration calculated from ancient deposits usually represent minimum estimates due to the rough time control where the duration of transgression and regression successions in a single base-level cycle cannot be accurately determined (e.g., Zhang et al., 2016). Despite this discrepancy, sustained high rate of shoreline migration has been documented for the late Holocene Mississippi River up to 100-150 m/year (Chamberlain, Törnqvist, Shen, Mauz, & Wallinga, 2018), thus the modelled rate of shoreline migration may not be so unrealistic.

5.3.3 Sediment capture and bypass

The nonuniform flow conditions provide a physical mechanism for sediment partitioning that regulates sediment bypass and deposition. Sediment input is consumed to aggrade the channel bed and build deltaic foresets, thus prograding the shoreline break. These two processes are affected by sediment partitioning between the channel bed and the proportion bypassed to the deltaic foreset, respectively. The width-average increments in channel bed aggradation and foreset growth are estimated as the increases in down-dip sectional areas of fluvial strata and the deltaic foreset, respectively. Cases 1-3 show a rapid initial increase in the percentage of sediment partitioned to channel bed (Figure 10), which corresponds to rapid channel bed aggradation and decreasing of channel bed slope near the downstream end of the fluvial reach. For Case 1, the percentage of sediment captured in channel consistently increases through the simulation, with more than 70% of the total sediment input. Case 2 shows similar pattern as Case 1, except that the percentage of sediment captured by the channel bed is much higher, reaching nearly 100%, which suggests that most of the sediment input was sequestered to the channel bed. The percentage of sediment partitioned on the channel bed is slightly lower for Case 3 during first 1,700 years of the model run compared to Case 1 and increases during the second half of the model run, reaching 90% by the end of Case 3. Case 4 shows a much lower percentage of sediment capture to the channel bed compared to Cases 1-3. Around 20% of the sediment input was partitioned to the channel bed through most simulation time for Case 4, which also corresponds to less thick fluvial stratigraphy (Figure 8h).

Backwater hydrodynamics have been observed to impact sediment partitioning within a fluvial reach, which leads to



FIGURE 10 Percentage of sediment partitioned within the channel fluvial system

channel bed aggradation and an overall downstream increase of flow depth with conditions of rising base level (Moran et al., 2017). The model results herein show that nonuniform flow conditions became more efficient in capturing sediment as the magnitude of downstream flow deepening increases. This is especially well demonstrated with Cases 1, 2 and 3. Although the maximum rate of channel bed aggradation decreases for all the simulations, the overall percentage of sediment partitioning increases as the total length of fluvial reach increases via rapid shoreline progradation for Cases 1-3 (Figure 9). Case 4, on the other hand, seems to show a situation where most sediments bypasses the fluvial system and feeds the deltaic foreset developed in the basin. This low percentage of sediment partitioning is likely the result of a much less developed nonuniform flow condition for Case 4, where the downstream increase in flow depth is only 20% (Figure 4). A recent experimental study demonstrated that fluvial-deltaic systems associated with deeper basins tend to develop less concave and steeper longitudinal profiles than systems associated with shallower basin (Bijkerk et al., 2016), thus facilitating bypass of the sediment and limiting aggradation of the channel bed. This agrees with the model results from this study, which show that less concave and steeper river profile in a deeper basin also corresponds to a less developed nonuniform flow (Case 4) compared to a shallower basin (e.g., Cases 1-3).

Furthermore, Case 3 presents an interesting situation where the channel bed aggrades under rapid base-level fall, rather than triggering channel incision. Traditional sequence stratigraphy models emphasize the base-level control on stratigraphic development, in particular base-level fall, usually results in channel incision, sediment bypass and development of a lowstand system tract. Theoretical and experimental studies also indicate that base-level fall could result in sediment release and channel incision (Kim et al., 2006). Case 3 suggests that the hydrodynamics of nonuniform flow in a fluvial-deltaic system can be sustained during base-level fall. It should be noted that the increased sediment partitioning in Case 3 was accompanied by a rapid decrease in the height of the deltaic foresets, and thus a less pronounced deltaic clinoform geometry. The decrease in the height of the deltaic foreset facilitates shoreline progradation (Figure 9), which results in increasing length of the fluvial reach. Together with the enhanced nonuniform flow condition, sediment partitioning on the channel bed increases. This may also explain the absence of distinct clinoform geometry in lacustrine basins (e.g., Jiang et al., 2013; Lin et al., 2001; Scholz et al., 1998), where basin depth is usually less than their marine counterparts and traditional sequence stratigraphic models are difficult to apply. Moreover, this accelerated rate of shoreline regression associated with diminishing deltaic foreset has been found in foreland settings (Leva López, Kim, & Steel, 2014), which represent an important autogenic behaviour of fluvial-deltaic systems. Alternatively, the development of fluvial stratigraphy in Case 3 may represent a normal regression scenario (Zecchin & Catuneanu, 2013), where the rate of base-level fall is not sufficient to cause erosion of underlying deposits.

6 | IMPLICATIONS FOR STRATIGRAPHIC RECORD

The model results show that nonuniform flow impacts the development of fluvial-deltaic stratigraphy by effecting sediment partitioning and therefore depositional centres (i.e., location of maximum channel bed aggradation). Sediment deposition leads to channel bed aggradation, which potentially preserves the deposited sediment as part of the stratigraphic record. Patterns of stratigraphic development could arise solely through nonuniform flow conditions (e.g. Case 1 simulation), or under simultaneous external forcing (Case 2 and Case 3). Figure 11 shows that the pronounced downstream fining of channel bed grain size serves as a strong indicator of nonuniform flow influence (Cases 1 and 3), even when grain size is averaged across the entire fluvial strata over a spatial extent of tens of kilometres. Since the standard deviation is relatively small, this signal of downstream fining of channel bed grain size could potentially be observed in the rock record. However, for Case 2, although the preserved median grain size (20 km average) in stratigraphy fines downstream, the standard deviation is more than double compared to that of Cases 1, 3 and 4, thus could be less distinct to identify in the rock record, but can also serve as an indicator of base-level rise. Sorting, as indicated by standard deviation



FIGURE 11 (a-d) 20-km average median channel bed grain size (normalized by initial channel bed grain size, $D^* = D_{50}/D_i$) with standard deviation of the entire fluvial strata for four model cases. (e–h): 20-km average of flow depth (normalized by normal flow depth, $H^* = H/H_n$) with standard deviation of the entire fluvial strata for four model cases. Labels for *x*-axis at the top of the figure show the upstream distance from the shoreline break (x_{sb}) normalized by backwater length (L_{bw}) at the end of the model run. Labels for *x*-axis at the bottom of the figure show the actual distances. Dashed lines mark the locations of transition from normal to nonuniform flow (x_t) at the end of each model run

of median grain size, generally decreases and then increases towards downstream (Figure 11a–d). However, if only focusing on the nonuniform flow zone defined by L_{bw} , the sorting will dominantly increase downstream for Cases 1, 3 and 4. This result is consistent with field observations (Petter, 2010) and subsurface data (Martin et al., 2018). This also suggests that (a) it's important to examine sorting beyond backwater length scale to test the modelled trend in grain size distribution, and (b) the presence of a strong peak of standard deviation in median grain size within the backwater zone might indicate base-level rise (Case 2).

The model results also show that downstream deepening of flow depth across the entire fluvial strata could be quite common for ancient fluvial–deltaic systems developed for various basin settings (Figure 11e–h). The standard deviation of averaged flow depth is relatively small for Cases 1, 3 and 4, but not for Case 2. This suggests that vertical variability in flow depth in the rock record is limited and the signal of downstream deepening of flow depth could potentially be resolved (e.g., Petter, 2010). Unlike grain size, which can be directly measured, paleo-flow depth can only be estimated using inverse methods or empirical scaling relationships (e.g., Hajek & Heller, 2012; Bradley & Venditti, 2017). However, to invert the paleo-flow depth and resolve the signal of downstream deepening of flow depth within the backwater reach of an ancient system is still challenging, because most paleohydraulic reconstruction methods are developed for uniform flow (e.g., Leclair & Bridge, 2001; Lynds, Mohrig, Hajek, & Heller, 2014; Bradley & Venditti, 2017). There are typically two assumptions associated with existing inversion methods: (a) flow is uniform and (b) net channel bed aggradation is zero (Leclair, 2006; Leclair & Bridge, 2001; Paola & Borgman, 1991). Both assumptions are violated in nonuniform flow in the backwater reach of a fluvial-deltaic system thus these inversion methods still require testing for nonuniform flow. Further studies are needed to clarify how the distinct signal of backwater-driven downstream increase of flow depth influences and transforms into measurable stratigraphic features, such as cross bed thickness and bar thickness (Wu, Nittrouer, & Swanson, 2018). The spatio-temporal variability in channel bed aggradation (Figure 7a-d) and sediment flux caused by nonuniform flow (Chatanantavet et al., 2012; Nittrouer et al., 2012) are likely to affect the distribution of preserved cross set thickness (Jerolmack & Mohrig, 2005), which could potentially show spatial patterns that indicate backwater influence (Wu et al., 2018).

7 | CONCLUSIONS

River hydrodynamic conditions are modified where a system approaches its terminal basin, characterized by the EAGE

onset of backwater flow. The concept of backwater effect has been used to interpret trends, such as the spatial variabilities in grain size and paleoflow depth, in ancient fluvial-deltaic stratigraphy. The signatures of nonuniform flow preserved in the stratigraphic record are therefore largely inferred, but quantitative tests using morphodynamic models have yet to be produced. This study uses a one-dimensional morphodynamic model to evaluate the impacts of nonuniform flow on the development of fluvial-deltaic stratigraphy. The model results show that (a) the development of nonuniform flow under various external conditions and basin configurations result in the spatial variation in flow depth, channel bed slope and grain size over the delta. (b) Development of stratigraphic stacking pattern is the combined result of spatio-temporally varying deposition on the channel bed and shoreline migration. (c) The region of nonuniform flow extends much farther upstream than predictions from traditional backwater length scales. (d) Together with base-level fluctuation, the development of nonuniform flow can be enhanced to leave distinct stratigraphic features. However, future work is needed to understand how channel hydraulic signatures transform into rock record. This requires improved paleohydraulic reconstruction methods that account for nonuniform flow.

ACKNOWLEDGEMENT

Chenliang Wu acknowledges financial support provided by the ExxonMobil/GSA Student Geoscience Grants from Geological Society of America, David Worthington Named Grant and Meckel Family Named Grant from the American Association of Petroleum Geologists, Post-Graduate Research Grant from the International Association of Sedimentologists and support form Rice University. We thank reviewers Tetsuji Muto, Enrica Viparelli and John Martin for their insightful and constructive reviews. This work is benefited from discussions with Hongbo Ma, Wonsuck Kim, David Mohrig, Gary Parker, Travis Swanson and Andrew Moodie.

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How to cite this article: Wu C, Nittrouer J. Impacts of backwater hydrodynamics on fluvial–deltaic stratigraphy. *Basin Res.* 2019;00:1–18. <u>https://doi.</u> org/10.1111/bre.12385