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Supplemental material

Morphodynamic model

Following the numerical model of Wu and Nittrouer (2019), the morphodynamic model of Parker et al. (2008a, b) is integrated with and the "surface-based" grain size method (Naito et al., 2019) to calculate non-uniform flow hydraulics, channel bed grain size, and simulate the evolution of the alluvial profile. Flow depth (*H*) in the along-stream distance (*x*) is calculated using the backwater equation:

$$\frac{dH}{dx} = \frac{S - C_f Fr^2}{1 - Fr^2}, (1)$$

where C_f is friction coefficient, Fr is Froude number, determined by $Fr = U(gH)^{-0.5}$, with *U* the depth-averaged flow velocity and *g* gravitational acceleration.

Sediment flux is calculated using the bed material load equation of Ma et al. (2017):

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

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

The along-stream variation in sediment flux ($\partial q_s/\partial x$) is calculated and used to assess rate of channel bed aggradation ($\partial\eta/\partial t$) using a modified Exner equation (Paola and Voller 2005):

$$\frac{\partial\eta}{\partial t} = -\mathrm{A}\frac{\partial q_s}{\partial x}, (3)$$

where A is determined by: $\mathrm{A} = (1 + \Lambda)\Omega I_f\left(1 - \lambda_p\right)^{-1} {r_B}^{-1}$, and λ_p is the mean porosity of the channel-floodplain complex, Λ is the mud/sand ratio, Ω is the channel sinuosity, I_f is the flood intermittency, and r_B is the ratio between channel width and width of flood plain; short of measuring these parameters from field data, standard values from modern lowland systems that scale with the ancient basins are utilized (Table 1, Parker et al., 2008b; Moran et al., 2017). The purpose of including the coefficient A in the model is to account for natural processes such as sediment deposition the floodplain. The net effect of adding coefficient A in the model is that the absolute dimensional rate of channel bed aggradation will be reduced to a degree that is more realistic as natural systems (Parker, 2004). The autostratigraphic time and length scales are also adjusted with the coefficient A. During nondimensionalization, the coefficient A is effectively canceled out so that the choice of the set of parameter values that determines the value of A will not affect most of the model results. Changes in channel bed elevation ($\partial\eta$) affect along-stream flow depth and therefore system slope (i.e., Equation 1). Following Moran et al. (2017) and Wu and Nittrouer (2019), the transition from uniform to non-uniform flow is estimated to be the location where along-stream changes in flow depth (dH/dx) is less than 5×10^{-6} and the backwater length is calculated as the distance between this location and the shoreline break.

Table 1. Key model input parameters

Variable	Value	Description
S_f	1.6×10^{-4}	Slope of deltaic foreset
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 deposited in the channel–floodplain complex per unit sand deposited
r_B	60.0	ratio of channel width to flood plain width

Initial condition of the alluvial profile is determined by the initial channel bed slope of the system and the elevation of the downstream end of the alluvial profile, which is set at 7 m below the initial sea level of 0 m. The choice of the downstream elevation of alluvial profile has very limited influence on the long-term development of stratigraphy and hydraulics (Wu and Nittrouer, 2019). Important initial boundary conditions include depth-averaged flow velocity and width-averaged water discharge at upstream boundary, reach-averaged flow depth at both upstream and downstream boundaries. The depth-averaged flow velocity U at the upstream boundary is assumed to be 1.7 ms^{-1}. Width-averaged water discharge q_w at the upstream boundary can be calculated by combining $\tau_b = \rho C_f U^2$, $\tau_b = \rho g H S$, and $H = q_w/U$, where τ_b is the boundary shear stress and ρ is the flow density. Therefore, the reach-averaged flow depth H at the upstream boundary is then calculated as $H = q_w/U$.

The break between the downstream boundary of the fluvial reach and deltaic foreset is effectively the shoreline location before the onset of sediment-starved autoretreat. Therefore, the flow depth at the downstream alluvial boundary (i.e., shoreline break) is calculated as the difference between sea level and bed elevation at shoreline break (Parker et al., 2008b; Wu and Nittrouer, 2019). When sediment-starved autoretreat starts, this geometric break in cross-sectional profile is abandoned and cannot represent the shoreline anymore as shoreline starts rapid retreat. Therefore, to calculate the downstream flow depth during sediment-starved autoretreat, the location of the shoreline needs to be specified first. The location of the shoreline during sediment-starved autoretreat is estimated as the location where the ratio between friction slope and channel bed slope S_{fric}/S is less than 0.3 in previous study (Parker et al., 2008b). This ratio is determined from field data (Parker et al., 2008b). When the friction slope is too low, channel bed shear stress will drop below the critical shear stress needed to transport sediments, thus sediment flux (q_s) drops close to zero, downstream of which can be considered morphodynamic inactive. In this study, we use a threshold S_{fric}/S ratio of 0.1 to ensure that no significant amount of sediments can be transported beyond the calculated shoreline location. Since the sediment flux doesn't become zero beyond the calculated shoreline location, the abandoned part of the alluvial reach is not completely inactive but is nearly static through autoretreat. During sediment-starved autoretreat, the flow depth at the downstream alluvial boundary (i.e., abandoned break between alluvial profile and deltaic foreset) is calculated as the difference between sea level and bed elevation. The advantage of this method is that sediment-starved autoretreat arises without the need of special numerical treatment so that the model is more stable.

Code for autoretreat simulation and production of the figures in the main manuscript (Wu, 2020) is available at https://github.com/ChenliangWuGeo/autoretreat.

Model simulations

We performed one thousand model simulations. Initial channel bed slopes S and basement slopes S_b are generated by Monte Carlo sampling, assuming log-uniform distribution for both parameters bounded by the ranges from 4 $\times 10^{-4}$ to 6 $\times$ 10^{-5} and 10^{-5} to 10^{-3}, respectively. All the simulation runs developed autoretreat. The time scale for sediment-starved autoretreat (T_{ssa}) is calculated as the period from the start of model run to the onset of sediment-starved autoretreat.

Detrending alluvial profile

The alluvial profile is detrended by subtracting a downstream-dipping fixed-slope profile from the channel bed profile. For each time step, the fixed-slope profile is calculated as a straight-line segment that shares the upstream boundary with the alluvial profile and has a slope that equals the initial channel bed slope S. Root mean square of the channel bed profile is then calculated as the root mean square of the detrended channel bed elevation of the active alluvial nodes (from upstream boundary to shoreline break), where abandoned alluvial section during sediment-starved autoretreat is excluded for the calculation. Cross-correlation between profiles at adjacent time steps are calculated using the cross-correlation function in Matlab. The function compares an alluvial profile at one time step to the shifted (lagged) alluvial profile at the adjacent time step.

Backwater conditions of natural rivers

The flow depth for natural rivers illustrated in figure 4B are calculated based on published data. Mouth flow depth are mostly calculated by interpolating and resampling the flow depth of the lower 1/10 reach of the backwater zone for every 1 km based on published data and figure. The normal and mouth flow depths of the Mississippi river are calculated based on low stage (10,000 - 15,000 m^3/s) flow depths from RK 696 to 800 and from RK 31 to 100 (Figure 8A of Nittrouer et al., 2012), respectively. The normal and mouth flow depths of the Trinity river are calculated based on median stage (855 m^3/s) flow depths from 27 to 60 km and from 119 to 125 km (Figure 2 of Smith et al., 2019), respectively. The normal and mouth flow depths of the Tombigbee river are calculated based on low stage (300 m^3/s) flow depths from 89 to 103 km and 224 to 238 km (Figure 2 of Dykstra and Dzwonkowski, 2020), respectively. In addition, the normal and mouth flow depths of the middle Fly river are calculated based on the difference between bank height and mean bed elevation from 400 to 440 km and 0 to 40 km (figure 4 of Day et al., 2008), respectively. The “mouth” of the middle Fly river is in fact the junction with

Strickland river, where local backwater effect can develop, rather than the actual fluvial-marine interface of the Fly river. The normal and mouth flow depths of the Yangtze river are adapted from the flow depths of Shashi-Ouchikou reach and Datong-Wuhu reach (table 3 of Huang et al., 2014), respectively. These flow depth estimates are comparable to another study (Chen et al., 2007). The mouth flow depth of the Brazos river is calculated by averaging the lower 13 km bathymetry of the river (Carlin et al., 2015). The normal flow depth of the Brazos river is calculated using rearranged Manning's relation: $H = (\frac{nQ}{BS^{0.5}})^{0.6}$, where H is the average flow depth, n is the Manning's roughness coefficient taken a value of 0.035, Q is the station discharge at Richmond, Texas (same day as the bathymetry survey), B is channel width measured from Google Earth™, and S is the channel slope.

References

Carlin, J. A., Dellapenna, T. M., Strom, K., & Noll IV, C. J. (2015). The influence of a salt wedge intrusion on fluvial suspended sediment and the implications for sediment transport to the adjacent coastal ocean: a study of the lower Brazos River TX, USA. Marine Geology, 359, 134-147.

Chen, Z., Chen, D., Xu, K., Zhao, Y., Wei, T., Chen, J., ... & Watanabe, M. (2007). Acoustic Doppler current profiler surveys along the Yangtze River. Geomorphology, 85(3-4), 155-165.

Day, G., Dietrich, W. E., Rowland, J. C., & Marshall, A. (2008). The depositional web on the floodplain of the Fly River, Papua New Guinea. Journal of Geophysical Research: Earth Surface, 113(F1).

Dykstra, S. L., & Dzwonkowski, B. (2020). The propagation of fluvial flood waves through a backwater-estuarine environment. Water Resources Research. 56, e2019WR025743, https://doi.org/10.1029/2019WR025743

Huang, H. Q., Deng, C., Nanson, G. C., Fan, B., Liu, X., Liu, T., & Ma, Y. (2014). A test of equilibrium theory and a demonstration of its practical application for predicting the morphodynamics of the Yangtze River. Earth Surface Processes and Landforms, 39(5), 669-675.

Ma, H., Nittrouer, J.A., Naito, K., Fu, X., Zhang, Y., Moodie, A.J., Wang, Y., Wu, B., Parker, G. (2017). The exceptional sediment load of fine-grained dispersal systems: Example of the Yellow River, China. Science advances, 3(5), e1603114.

Moran, K. E., Nittrouer, J. A., Perillo, M. M., Lorenzo-Trueba, J., and Anderson, J. B., 2017, Morphodynamic modeling of fluvial channel fill and avulsion time scales during early Holocene transgression, as substantiated by the incised valley stratigraphy of the Trinity River, Texas. Journal of Geophysical Research: Earth Surface, v. 122, p. 215-234.

Naito, K., Ma, H., Nittrouer, J. A., Zhang, Y., Wu, B., Wang, Y., Fu, X., and Parker, G., 2019, Extended Engelund–Hansen type sediment transport relation for mixtures based on the sand-silt-bed Lower Yellow River, China. Journal of Hydraulic Research, v. 57:6, p. 770-785.

Nittrouer, J. A., Shaw, J., Lamb, M. P., & Mohrig, D. (2012). Spatial and temporal trends for water-flow velocity and bed-material sediment transport in the lower Mississippi River. GSA Bulletin, 124(3-4), 400-414.

Parker, G. (2004), 1D Sediment Transport Morphodynamics With Applications to Rivers and Turbidity Currents [electronic]: Civ. and Environ. Eng., Univ. of Ill. at Urbana-Champaign, Urbana.(Available at http://vtchl. uiuc. edu/people/parkerg/morphodynamics_e-book. htm).

Paola, C., and V. R. Voller (2005), A generalized Exner equation for sedimentmass balance, Journal of Geophysical Research, 110, F04014, doi:10.1029/2004JF000274.

Parker, G., Muto, T., Akamatsu, Y., Dietrich, W. E., & Lauer, J. (2008a). Unravelling the conundrum of river response to rising sea-level from laboratory to field. Part I: Laboratory experiments. *Sedimentology*, 55(6), 1643-1655.

Parker, G., Muto, T., Akamatsu, Y., Dietrich, W. E., & Wesley Lauer, J. (2008b). Unravelling the conundrum of river response to rising sea-level from laboratory to field. Part II. The Fly–Strickland River system, Papua New Guinea. *Sedimentology*, 55(6), 1657-1686.

Smith, V., Mason, J., & Mohrig, D. (2019). Reach-scale changes in channel geometry and dynamics due to the coastal backwater effect: the lower Trinity River, Texas. Earth Surface Processes and Landforms.

Wu, C., & Nitterour, J. A. (2019). Impacts of backwater hydrodynamics on fluvial–deltaic stratigraphy. Basin Research. *in press*.

Wu (2020). ChenliangWuGeo/Autoretreat: first release of autoretreat (Version v1.0). Zenodo. http://doi.org/10.5281/zenodo.3837668.