The importance of erosion in distributary channel network growth, Wax Lake Delta, Louisiana, USA

John B. Shaw* and David Mohrig
Jackson School of Geosciences, University of Texas at Austin, C9000, Austin, Texas 78712, USA

ABSTRACT

A river delta's shape and kinematics are dictated by the three-dimensional evolution of its distributary channels on the delta front, yet measurements of this evolution are scarce. We supply four bathymetric surveys documenting this evolution for Gadwall Pass—part of the Wax Lake Delta, one of the few rapidly prograding regions of the greater Mississippi Delta in coastal Louisiana, United States. This distributary channel extends 2–6 km beyond the sub-aerially emergent delta (dependent on water surface elevation) and bifurcates into four similarly sized distributary channels (average channel width = ~150 m) in this sub-aqueous reach. Distributary channel growth proceeds primarily through erosion of the unchannelized foreset deposit, and growth patterns differ between high and low river flow. During high river flow, high upstream sand supply acts to aggrade the bed both inside and outside of the channel network. Erosion during high flow is focused at the sand shoals that define the sidewalls of the bifurcate channels, causing channel network rearrangement into a single primary channel with the remaining secondary channels branching off of it. During low river flow, bed erosion is focused at channel tips and the beds of all of the sub-aqueous distributary channels, leading to a bayward extension of each channel tip by ≥0.87 channel-widths. Channel-bottom erosion during low river flow is enhanced by tidally modulated currents that support sand suspension and transport in the subaqueous channels during ebb tide while receiving only a small sand supply from upstream.

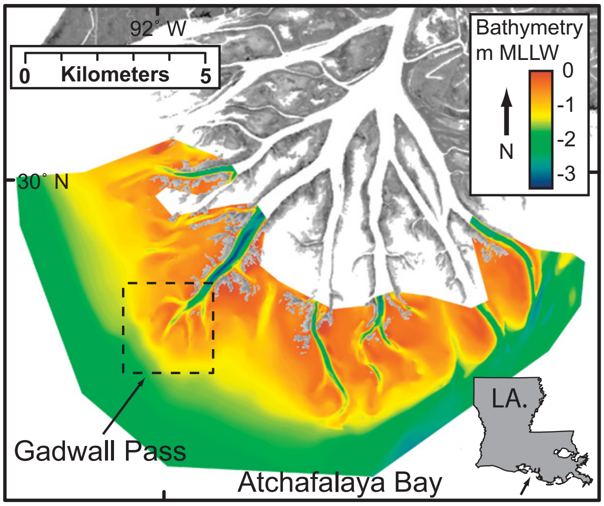

Figure 1. Bathymetric map of Wax Lake Delta (Louisiana, USA) from August 2011, overlain by LANDSAT imagery from 1 October 2011. LANDSAT imagery was acquired when water level at the Amerada Pass tide gauge (National Oceanic and Atmospheric Administration [NOAA] site #11354, 4.2 km east of map) was –0.11 m MLLW (mean lower low water). Dashed box shows location of the time series of bathymetric surfaces shown in Figure 2. Arrow in map of Louisiana shows general location of study.

INTRODUCTION

As an actively prograding sub-delta of the greater Mississippi River Delta (Louisiana, United States), the Wax Lake Delta (WLD, Fig. 1) is viewed by geoscientists, engineers, and ecologists as the primary analogue for land-building processes that can be harnessed elsewhere in the region (Allison and Meselhe, 2010; Paola et al., 2011; Falcini et al., 2012). Sediment accumulation and morphology are generalized from the WLD to other locations (Kim et al., 2009), yet few measurements of channel kinematics in the delta front exist. The transferability of the WLD's morphology to proposed land-building sites cannot be evaluated until the kinematics controlling the morphology-defining channel network are understood.

The WLD is assumed by many studies to be a classic river-dominated delta (Wellner et al., 2005; Falcini and Jerolmack, 2010; Edmonds et al., 2011), i.e., its form is negligibly affected by the small mean tidal range (0.4 m) and wave climate (0.5 m maximum monthly wave height; Syvitski, 2006). This assumption, which has been used to explain the delta's bifurcating distributary channel network (Wellner et al., 2005; Edmonds and Slingerland, 2007), leads to two corollary assumptions regarding bathymetric change on the delta front: (1) bathymetric change occurs only during floods, and (2) depositional processes drive channel extension. This study will show that the assumptions derived from this conjecture of river-domination do not accurately describe channel kinematics on the WLD.

The first assumption is that the majority of bathymetric change on the delta front occurs during periods of high river flow (floods) when most of the sediment arrives at the delta. Consequently, models commonly assume that all sediment motion and change in bed elevation is tied to exceedance of a characteristic water discharge (Kim et al., 2009; Edmonds and Slingerland, 2007; Geleynse et al., 2010). However, evidence from other deltaic systems suggests that considerable sediment reworking can occur during low flow. Mean flows tend to deepen channels on the Mossy Delta in Saskatchewan, Canada (R. Slingerland, 2013, personal commun.), and the full hydrologic cycle is important for sediment transport in a crevasse splay on the Mississippi Delta (Esposito et al., 2013). We measure bathymetric change over both high and low flow conditions, and find that distinct patterns of bathymetric evolution emerge for each condition.

The second assumption is that depositional processes are the primary drivers of channel kinematics on the delta front, with erosion playing a relatively minor role. Existing models of channel network formation predict that networks arise as aggrading channel levees confine a channel laterally (Rowland et al., 2009), and focused deposits downstream of channel mouths ("mouth bars") deflect the flow to either side, causing channel bifurcation (Bates, 1953; Wright, 1977; Wellner et al., 2005; Edmonds and Slingerland, 2007; Falcini and Jerolmack, 2010). However, studies of the WLD channel network initiation (Roberts et al., 1980) and growth to its present-day emergent channel network (Shaw et al., 2013) note channel erosion within the net-depositional system. The present study shows that the channel network growth cannot be understood without the explicit inclusion of erosive processes.

RESULTS

Gadwall Pass distributary channels were measured four times using methods described in the GSA Data Repository[1]. Channels maintain

*Current address: Department of Geology and Geophysics, University of Wyoming, Department 3006, Laramie, Wyoming 82071, USA.

[1]GSA Data Repository item 2014008, survey methods and grain-size analysis, is available online at www.geosociety.org/pubs/ft2014.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

GEOLOGY Data Repository item 2014008 | doi:10.1130/G34751.1

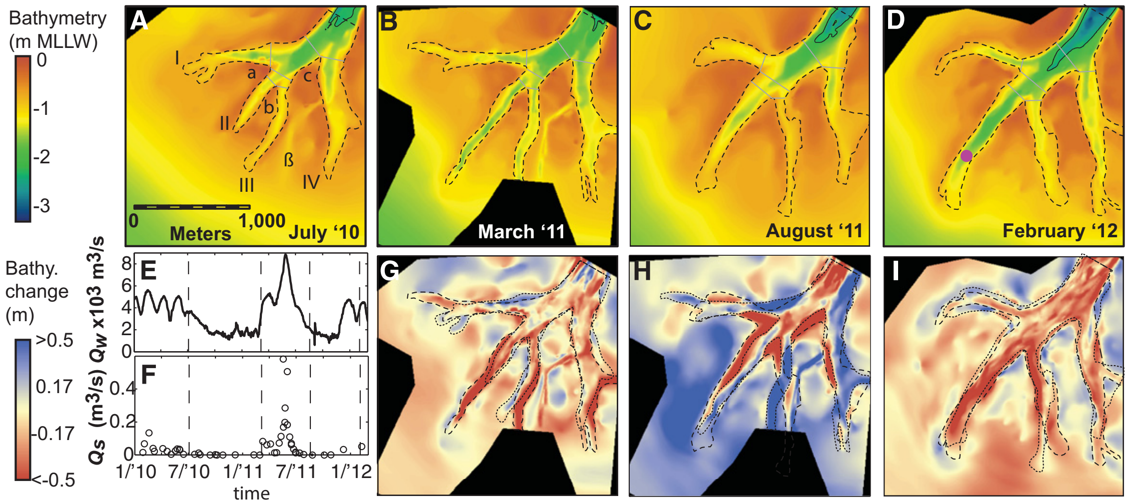

Figure 2. A–D: Bathymetric evolution of Gadwall Pass from July 2010 to February 2012. Black solid line (–2.3 m MLLW [mean lower low water]) separates the bedrock extent (deeper) and alluvial (shallower) regions of the channel. Dashed line marks the channel network boundary (see the Data Repository [see footnote 1] for definition). Gray lines mark transects used to measure bifurcation asymmetry. Distributary channels I–IV and bifurcate shoals a–c are labeled in A. A fifth, downstream deepening channel is labeled β. The pink dot in D shows location of water and sand discharge measurements (Fig. 3). E,F: Water (Q_w) and suspended sand (Q_s) discharge measured at Calumet, LA (U.S. Geological Survey gauge #07381590), upstream of the Wax Lake Delta and survey periods (gray dashed lines). G–I: Difference maps showing bathymetric change between successive surveys. Dotted and dashed lines show the channel network boundary from the earlier and later surveys, respectively.

definition for 2 km (7 channel-widths) beyond the low-tide-emergent delta when water levels are near 0 m MLLW (mean lower low water; Fig. 1). This submerged length increases to 6 km (20 channel-widths) at high tide, when the water level is 0.49 m higher (NOAA, 2012) and much of the WLD is inundated (Allen et al., 2012). At the upstream edge of the survey area (Fig. 2), the channel bed is –2.4 m MLLW, and incised into cohesive muds of pre-delta Atchafalaya Bay that act as bedrock (Shaw et al., 2013) and lie beneath –2.3 m MLLW (black solid lines in Figs. 2A–2D). The trunk channel bifurcates into four distributary channels (I through IV in Fig. 2A) bayward of the bedrock exposure. The relief between channel bed and banks decreases in the downstream direction, reaching zero at the channel tips where bed elevations lie between –1.2 and –0.6 m MLLW. Each channel bifurcation has an associated sand shoal that separates the two adjacent channels (a–c, in Fig. 2A). A fifth channel form is defined between channels III and IV in July 2010 and March 2011(marked "β" in Fig. 2A). This channel grows deeper and wider in the downstream direction in contrast to channels I–IV and is consequently excluded from analysis of the mapped channel network. Beyond this network, the unchannelized delta foreset slopes asymptotically to the bay bottom with a maximum slope of 1×10^{-3}. Grab-samples of bed material downstream of the bedrock reach show that the sub-aqueous channel network is composed of well-sorted sand with median grain diameter up to 210 µm. Grain sizes composing the bed grow finer with distance from the channel tips (Fig. DR2 in the Data Repository). No plants had colonized the field site at the time of the surveys.

Between July 2010 and March 2011 (Figs. 2A and 2B), the WLD received an average water discharge of 1933 $m^3 s^{-1}$, transporting 3.0×10^6 m^3 of suspended sediment (sand + mud, density assumed to be 2650 kg m^{-3}; Figs. 2E and 2F). During this interval, 3.2×10^5 m^3 of bed material was eroded from the study site (Fig. 2G). The average channel bed elevation was eroded by 15%, dropping from –1.26 to –1.45 m MLLW, while the channel-bed area grew by 7%. Erosion focused at all channel tips produced an average downstream extension of 327 m (Table 1). Sand shoals a–c (Fig. 2) aggraded as much as 0.35 m along their margins, and the points of bifurcation at their upstream tips remained roughly stationary on average (Table 1). Mean channel bifurcation asymmetry, defined as the ratio of the larger to the smaller cross-sectional area of two channels at a bifurcation (Fig. 2), had similar values of 1.28 and 1.18 in July 2010 and March 2011 (Table 1), respectively. Downstream of the channel network, a wide, unchannelized swath of bed degradation (<0.3 m) was found. Channel β also incised during this time.

Between March 2011 and August 2011 (Figs. 2B and 2C), the WLD underwent the second largest flood on record. The average water discharge was 4950 $m^3 s^{-1}$, and 4.9×10^6 m^3 of suspended sediment reached the delta (Figs. 2E and 2F), of which 6.0×10^5 m^3 of sediment was deposited in the study area (Fig. 2H). Over this period, the area covered by the channel bed increased by 12%, but the average channel bed elevation aggraded, shallowing by 12% of the previous water depth to –1.27 m MLLW. The regions of largest aggradation outside the channel network were along the western bank of channel II (>0.44 m) and where the previous courses of channels III, IV and β were filled in by deposits >0.37 m in thickness. In contrast to the period of low flow, the largest changes to the channel network shape were at channel bifurcations instead of at the channel tips.

TABLE 1. METRICS DESCRIBING THE EVOLUTION OF GADWALL PASS

Channel tip extension (m)	July 2010 to March 2011	March 2011 to August 2011	August 2011 to February 2012
I	230	–350	130
II	570	160	140
III	X	X	–100
IV	180	–80	260
Average	327	–90	108

Bifurcation migration (m)	July 2010 to March 2011	March 2011 to August 2011	August 2011 to February 2012
A	–40	200	–50
B	20	160	10
C	30	120	30
Average	3	160	–3

Bifurcation asymmetry	July 2010	March 2011	August 2011	February 2012
A	1.35	1.22	2.11	3.02
B	1.10	1.29	2.42	3.30
C	1.40	1.03	2.64	1.58
Average	1.28	1.18	2.39	2.64

Note: The table quantifies the bathymetric evolution of Gadwall Pass, Louisiana, USA. Channel tip extension quantifies the change in location of channel tips I–IV (Fig. 2A). Negative numbers indicate upstream retreat of a channel tip. X's mark where extension could not be measured. Bifurcation migration quantifies the change in location of the sand shoal separating channel bifurcations a, b, and c (Fig. 2A), with negative numbers indicating upstream migration. Bifurcation asymmetry is the ratio of cross-sectional area between the larger and smaller bifurcate channels at bifurcations measured at cross-sections marked by gray lines in Figure 2A.

The erosion of channel sidewalls caused the tips of sand shoals separating the distributary channels to migrate downstream an average of 160 m (Table 1). The asymmetric erosion of channel sidewalls increased the mean bifurcation asymmetry from 1.2 to 2.4. Overall, bathymetric change during flood "selected" channel II and channels I, III, and IV became subordinate channels (Fig. 2C). Change in positioning of the distributary-channel tips was mixed. Channel II extended downstream via growth of subaqueous levees, but channels I and IV retreated upstream as their previous courses were filled with sediment (Table 1; Fig. 2H).

Between August 2011 and February 2012 (Figs. 2C and 2D), the delta returned to low river flow conditions, with average water discharge of 2440 $m^3 s^{-1}$ that transported 2.4×10^6 m^3 of suspended sediment. Distributary channels extended an average of 110 m basinward (Table 1). Channel III did retreat by 100 m, but this was associated with 280 m of eastward migration (Fig. 2I). Sediment deposition of up to 0.33 m was limited to the proximal channel levees, and the position of sand shoals remained roughly stable on average (Table 1). The channelized area increased by 3% and the average channel bed elevation incised by 19% to –1.75 m MLLW. Channel bifurcation asymmetry increased slightly over this period to 2.6 (Table 1).

Flow velocity and sediment transport measurements collected near the tip of channel II in February 2012 (Fig. 3) show how tidal modulation of river currents dominates water and sand transport on the delta front during low river flow. Data were collected over one spring tidal cycle (amplitude = 0.51 m) under stable meteorological conditions (constant 2–4 m s^{-1} winds from the southeast, measured at the NOAA [2012] Amerada Pass Gauge #8764227, located 4.2 km east of the area shown in Figure 1). Water velocity ranged from 0.046 m s^{-1} during rising tide to 0.730 m s^{-1} during falling tide (Fig. 3B). Flow direction did not reverse during rising tide because the incoming river discharge (3800 $m^3 s^{-1}$; Fig. 2E) exceeded the discharge required to fill the tidal prism (1300 $m^3 s^{-1}$), estimated as the subaqueous area of the WLD (72 km^2) times the average rate of change in water-surface elevation during rising tide (0.067 m hr^{-1}). Shear velocity was computed by fitting a logarithmic profile to velocity data. During rising tide, shear velocity fell below the critical shear velocity of sediment motion for the 210 μm sand composing the channel bed (9.2 mm s^{-1}; Wilkerson and Parker, 2011; Fig. 3C). Shear velocity increased during falling tide until conditions were reached where sand was entrained into suspension (21.0 mm s^{-1}; Wilkerson and Parker, 2011; Fig. 3C). Accordingly, width-averaged suspended-sand transport measured with a 76 × 76 mm^2 Helley-Smith sampler placed 0.09 m above the bed with a 105 μm mesh bag recovered only traces of sand during rising tide (width-averaged suspended sand flux through sampler = 4.63×10^{-4} kg m^{-1} hr^{-1}), but increased by three orders of magnitude to 1.33×10^{-1} kg m^{-1} hr^{-1} during falling tide.

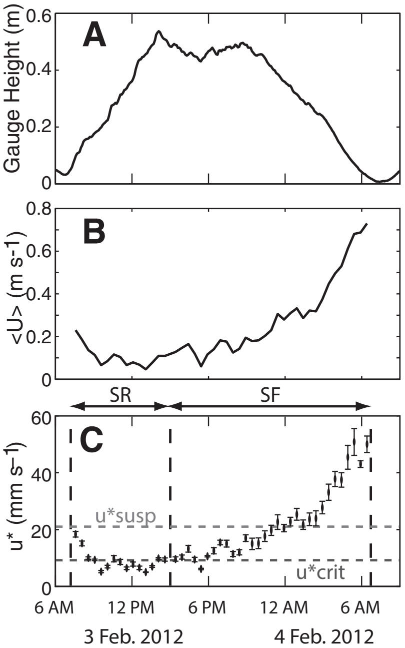

Figure 3. Flow measurements over a tidal cycle, 3–4 February 2012. A: Gauge height measured at Amerada Pass in Atchafalaya Bay (National Oceanic and Atmospheric Administration [NOAA] site #11354). B: Depth-averaged flow velocity (<U>) measured by acoustic Doppler current profiler. SR and SF indicate periods of sampling for rising- and falling-tide suspended sand discharge. C: Shear velocity (u*) calculated from velocity profiles. Thresholds of sediment motion and suspension for the 210 μm sand composing the bed are indicated by u*crit and u*susp, respectively.

DISCUSSION AND CONCLUSIONS

Bed erosion occurred at channel tips during low flow because there was sufficient bed shear stress to move sand at these points on the delta front, and there was little sand input. The strong reduction in sand flux during low river flows is explained by the sedimentary dynamics of backwater flow, where water surface slopes far upstream of the delta are reduced to the point that shear stresses are no longer sufficient to transport significant amounts of sand (Nittrouer et al., 2012). The data presented here (Fig. 3) show that tidal augmentation of low river discharge during ebb tide can produce the sediment-transporting shear stresses necessary to reshape the delta front. While flow and sediment transport measurements were collected at relatively high river flow (3900 $m^3 s^{-1}$), they show that sand transport can be effectively shut down during rising tide and support suspended sand transport during falling tide. In addition to tides, water level setup and drawdown associated with the passage of cold fronts can alter the water surface elevation by up to 1 m and flush up to 40% of the water from Atchafalaya Bay (Feng and Li, 2010). It is therefore likely that cold-front-induced changes in water level could also lead to channel erosion during times of low river flow.

Cold fronts are also a likely candidate for producing the large swaths of bed erosion, up to 0.3 m, observed on the fine-grained, unchannelized, distal delta front (Table DR2) during each of the low-flow bathymetric surveys (Figs. 2G and 2I). Through geochemical analysis of cores from the inner continental shelf near Atchafalaya Bay, Allison et al. (2000) estimated that a typical cold front could resuspend up to 1 cm of mud in water depths <10 m in Atchafalaya Bay and transport it seaward in suspension or as dilute mass flows (Walker and Hammack, 2000; Jaramillo et al., 2009). With an average of 25 cold fronts per year, it is reasonable to expect the observed tens of centimeters of erosion per season.

Mouth bars, the sedimentary deposits that accumulate down-flow from distinct channel mouths as flow expands into a still basin (Bates, 1953; Wright, 1977; Edmonds and Slingerland, 2007), have been suggested to be a key depositional element building the WLD (Wellner et al., 2005). Interestingly, the delta front bathymetry presented here does not easily conform to models of mouth bar deposition. While mouth bars normally exist as topographic elements distinct from lateral channel banks, the bathymetric surveys (Fig. 2) show that channel banks persist continuously to the channel tip. The subaqueous sand shoals (a–c) that separate distributary channels certainly are not active mouth bars because the front of the channel network has extended past them. Although it is possible that these shoals were once mouth bars, it seems unlikely given the lack of mouth bars observed beyond the tips of the current channels during the surveys (Figs. 2A–2D). The concept of a distinct channel mouth is also problematic, because subaqueous channels extend 7–20 channel-widths beyond the emergent channel banks permitting flow spreading over an aerially distributed zone rather than only at a discrete channel mouth. If delta front sedimentation cannot be tied to a channel mouth, then it cannot be called a mouth bar. Taken together, these observations suggest that models of mouth-bar sedimentation do not accurately describe the present-day growth of the WLD.

High and low river discharge conditions have different effects on the channelized and unchannelized portions of the delta front. During high flow (Fig. 2H) when significant sand input occurs from upstream (Fig. 2F), the unchannelized portion of the delta front aggrades everywhere except at channel margins, causing the delta foreset to prograde. The distributary channels of Gadwall Pass do not necessarily track this progradation, with one channel extending and two retreating. Within the channels, the effect of high flow is aggradation of channel beds and erosion of sidewalls of sand shoals. Sidewall erosion leads to an increase in bifurcation asymmetry. What were four similarly sized distributary channels are reorganized into a single, primary pathway through the delta front (channel II) and three secondary channels.

During low flow, minimal bed aggradation occurs due to low sand input from upstream. This sedimentation is limited to the proximal levees of the subaqueous channels (Fig. 2F). Erosion switches to the bottoms and downstream tips of each distributary channel (Figs. 2G and 2I). Channel extension occurring via low-river-flow erosion may be important for developing the radial network of multiple, similarly-sized distributary channels that characterizes WLD. It is worth noting that even though deposition of constructional levees caused channel II to extend during high flow (Fig. 2H), its rate of channel extension via low-flow erosion (Figs. 2G and 2I; 1.6 m day^{-1}) is greater than depositional extension during high flow (Fig. 2H; 1.0 m day^{-1}), showing that erosional extension is larger even for this primary channel.

The data presented here do not conform to the corollaries outlined in the introduction. First, the distinct patterns of erosion and deposition observed at high and low river flows show that only considering a single formative discharge to characterize delta-front dynamics is insufficient; reworking of the delta front during low flow is necessary for evenly distributed channel extension on the WLD. Second, the erosion found under both river conditions is essential for characterizing the growth of the channel network and changes to bifurcation asymmetry. We do not question the river-dominated classification of the WLD as few deltas on oceans are subject to smaller wave and tide forcing. Instead, we urge further study of sediment transport processes on the delta front, and their role in channel network kinematics.

ACKNOWLEDGMENTS

This work was supported by the STC program of the National Science Foundation via the National Center for Earth-Surface Dynamics (grant EAR-0120914), the FESD Delta Dynamics Collaboratory (grant EAR-1135427), and by the Jackson School of Geosciences, University of Texas at Austin. We thank T. Shaw, T. Swanson, E. Eke, M. Liang, J. Lorenzo-Trueba, A. Aronovitz, C. Armstrong, E. Rinehart, M. Hiatt, and A. Pilouras for assistance collecting the data.

REFERENCES CITED

Allen, Y.C., Couvillion, B.R., and Barras, J.A., 2012, Using multitemporal remote sensing imagery and inundation measures to improve land change estimates in coastal wetlands: Estuaries and Coasts, v. 35, p. 190–200, doi:10.1007/s12237-011-9437-z.

Allison, M.A., and Meselhe, E.A., 2010, The use of large water and sediment diversions in the lower Mississippi River (Louisiana) for coastal restoration: Journal of Hydrology (Amsterdam), v. 387, p. 346–360, doi:10.1016/j.jhydrol.2010.04.001.

Allison, M.A., Kineke, G.C., Gordon, E.S., and Goñi, M.A., 2000, Development and reworking of a seasonal flood deposit on the inner continental shelf off the Atchafalaya River: Continental Shelf Research, v. 20, p. 2267–2294, doi:10.1016/S0278-4343(00)00070-4.

Bates, C.C., 1953, Rational theory of delta formation: American Association of Petroleum Geologists (AAPG) Bulletin, v. 37, p. 2119–2162.

Edmonds, D., and Slingerland, R., 2007, Mechanics of river mouth bar formation: Implications for the morphodynamics of delta distributary networks: Journal of Geophysical Research, v. 112, F02034, doi:10.1029/2006JF000574.

Edmonds, D.A., Paola, C., Hoyal, D.C.J.D., and Sheets, B.A., 2011, Quantitative metrics that describe river deltas and their channel networks: Journal of Geophysical Research, v. 116, F04022, doi:10.1029/2010JF001955.

Esposito, C.R., Georgiou, I.Y., and Kolker, A.S., 2013, Hydrodynamic and geomorphic controls on mouth bar evolution: Geophysical Research Letters, v. 40, p. 1540–1545, doi:10.1002/grl.50333.

Falcini, F., and Jerolmack, D.J., 2010, A potential vorticity theory for the formation of elongate channels in river deltas and lakes: Journal of Geophysical Research, v. 115, F04038, doi:10.1029/2010JF001802.

Falcini, F., and 12 others, 2012, Linking the historic 2011 Mississippi River flood to coastal wetland sedimentation: Nature Geoscience, v. 5, p. 803–807, doi:10.1038/ngeo1615.

Feng, Z., and Li, C., 2010, Cold-front-induced flushing of the Louisiana Bays: Journal of Marine Systems, v. 82, p. 252–264, doi:10.1016/j.jmarsys.2010.05.015.

Geleynse, N., Storms, J., Stive, M., Jagers, H., and Walstra, D., 2010, Modeling of a mixed-load fluvio-deltaic system: Geophysical Research Letters, v. 37, L05402, doi:10.1029/2009GL042000.

Jaramillo, S., Sheremet, A., Allison, M.A., Reed, A.H., and Holland, K.T., 2009, Wave-mud interactions over the muddy Atchafalaya subaqueous clinoform, Louisiana, United States: Wave-supported sediment transport: Journal of Geophysical Research, v. 114, C04002, doi:10.1029/2008JC004821.

Kim, W., Mohrig, D., Twilley, R., Paola, C., and Parker, G., 2009, Is it feasible to build new land in the Mississippi River delta?: Eos [Transactions, American Geophysical Union], v. 90, p. 373–374, doi:10.1029/2009EO420001.

Nittrouer, J.A., Shaw, J., Lamb, M.P., and Mohrig, D., 2012, Spatial and temporal trends for water-flow velocity and bed-material transport in the lower Mississippi River: Geological Society of America Bulletin, v. 124, p. 400–414, doi:10.1130/B30497.1.

NOAA (National Oceanic and Atmospheric Administration), 2012, Tides and currents: Lawma, Amerada Pass, Louisiana: http://tidesandcurrents.noaa.gov/data_menu.shtml?stn=8764227%20Lawma,%20AmAmera%20Pass,%20LA&type=Historic+Tide+Data (accessed June 2013).

Paola, C., Twilley, R.R., Edmonds, D.A., Kim, W., Mohrig, D., Parker, G., Viparelli, E., and Voller, V.R., 2011, Natural processes in delta restoration: Application to the Mississippi Delta: Annual Review of Marine Science, v. 3, p. 67–91, doi:10.1146/annurev-marine-120709-142856.

Roberts, H., Adams, R., and Cunningham, R., 1980, Evolution of sand-dominant subaerial phase, Atchafalaya Delta, Louisiana: The American Association of Petroleum Geologists Bulletin, v. 64, p. 264–279.

Rowland, J.C., Dietrich, W.E., Day, G., and Parker, G., 2009, Formation and maintenance of single-thread tie channels entering floodplain lakes: Observations from three diverse river systems: Journal of Geophysical Research, v. 114, F02013, doi:10.1029/2008JF001073.

Shaw, J.B., Mohrig, D., and Whitman, S.K., 2013, The morphology and evolution of channels on the Wax Lake Delta: Journal of Geophysical Research, v. 118, p. 1–22, doi:10.1002/jgrf.20123.

Syvitski, J.P.M., 2006, The morphodynamics of deltas and their distributary channels, *in* Parker, G., and García, M., eds., River, Coastal and Estuarine Morphodynamics: RCEM 2005: London, Taylor & Francis Group, p. 143–150.

Walker, N.D., and Hammack, A.B., 2000, Impacts of winter storms on circulation and sediment transport: Atchafalaya–Vermilion Bay region, Louisiana, USA: Journal of Coastal Research, v. 16, p. 996–1010.

Wellner, R., Beaubouef, R., Van Wagoner, J., Roberts, H.H., and Sun, T., 2005, Jet-plume depositional bodies: The primary building blocks of Wax Lake Delta: Transactions of the Gulf Coast Association of Geological Societies, v. 55, p. 867–909.

Wilkerson, G.V., and Parker, G., 2011, Physical basis for quasi-universal relationships describing bankfull hydraulic geometry of sand-bed rivers: Journal of Hydraulic Engineering, v. 137, p. 739–753, doi:10.1061/(ASCE)HY.1943-7900.0000352.

Wright, L., 1977, Sediment transport and deposition at river mouths: A synthesis: Geological Society of America Bulletin, v. 88, p. 857–868, doi:10.1130/0016-7606(1977)88<857:STADAR>2.0.CO;2.

Manuscript received 20 May 2013
Revised manuscript received 11 September 2013
Manuscript accepted 12 September 2013

Printed in USA