Effect of tides on mouth bar morphology and hydrodynamics

Nicoletta Leonardi,1 Alberto Canestrelli, Tao Sun, and Sergio Fagherazzi

Received 1 November 2012; revised 28 June 2013; accepted 1 July 2013.

[Mouth bars are morphological units important for deltas, estuaries, or rivers debouching into the sea. Several processes affect the formation of these deposits. This paper focuses on the role of tides on shaping mouth bars, presenting both hydrodynamic and morphodynamic results. The effect of tides is analyzed in two end-member configurations: a river with a small tidal discharge compared to the fluvial discharge (fluvial dominated) and a river with a very large tidal discharge (tidal dominated). Mouth bar formation is analyzed using the coupled hydrodynamic and morphodynamic model Delft3D. The presence of tides influences the hydrodynamics of the jet exiting the river mouth and causes an increase in the averaged jet spreading. At low tide the lower water depth in the basin promotes a drawdown water profile in the river and an accelerated flow near the mouth. The resulting velocity field is characterized by residual currents affecting growth and final shape of the mouth bar. Simulations indicate that mouth deposits are characterized by the presence of two channels for negligible tidal discharge, whereas three principal channels are present in the tidal-dominated case, with a central channel typical of tidal inlets. On the basis of our numerical analyses, we present a robust criterion for the occurrence of mouth deposits with three channels. Trifurcations form when the tidal discharge is large with respect to the fluvial one and the tidal amplitude is small compared to the water depth. Finally, predicted mouth bar morphologies are compared with good agreement to river mouths in the Gulf of Mexico, USA.


1. Introduction

[2] Deltas result from the interaction between rivers and marine or lacustrine systems, giving rise to complex depositional patterns. Deltas are vulnerable to both anthropogenic and natural changes, such as sea level rise, variations in sediment supply, increase in storm activity and hurricane frequency, urbanization, and changes in land-use practices [e.g., Day et al., 2007; Edmonds, 2012a; Paola et al., 2011; Svytiksi et al., 2009]. Human activities continuously threaten these valuable environments, and new efforts are required to improve their management and restoration strategies. A vast literature on the overall morphology of deltas is already available; however, little work has been done on the effects of tides on delta morphology, and, in particular, on the effect of tides on mouth bars development.

[3] Researchers have analyzed both large-scale structure of deltaic systems [Edmonds, 2012b; Edmonds et al., 2009, 2011a, 2011b; Edmonds and Slingerland, 2010; Canestrelli et al., 2010, Geleynse et al., 2010, 2011; Jerolmack, 2009; Jerolmack and Paola, 2007; Fagherazzi, 2008; Jerolmack and Swenson, 2007; Fagherazzi and Overeem, 2007] and their morphological units, such as mouth bars [Esposito et al., 2013; Nardin et al., 2013; Nardin and Fagherazzi, 2012; Edmonds and Slingerland, 2007; Wang, 1984; Wright, 1977; Bates, 1953].

[4] As indicated by the conceptual model of Edmonds and Slingerland [2007], when a channelized flow enters a body of water the sediment transport rate decreases and a mouth bar evolves through initial deposition, progradation, and stagnation, finally leading to channels formation and bifurcation. Esposito et al. [2013] successfully tested Edmonds and Slingerland [2007] model for mouth bar evolution by means of field data near the mouth of the Mississippi River.

[5] Deltaic bifurcations occur at the fossilized location of mouth bars, which are crucial to delta progradation and often constitute much of the deltaic system [Edmonds and Slingerland, 2009]. With increasing channel bifurcation order, there is a systematic reduction in channels length, width, and depth, due to a lower jet momentum flux and consequent lower transport distance basinward [Edmonds and Slingerland, 2007, 2009].

[6] Marine processes might play a considerable role in sediment deposition, and their influence is reflected in the diversity of planar configurations and internal stratigraphy of coastal deposits. Among others, wave energy, tidal range, and the degree to which tidal currents determine the flow within the lower reaches of a river have been indicated as important geomorphic agents for delta morphodynamics.
and, in particular, for mouth bar evolution [e.g., Wright and Coleman, 1974; Jerolmack, 2009; Ashton and Giosan, 2011; Geleynse et al., 2010; Nardin and Fagherazzi, 2012; Nardin et al., 2013]. As an example, the mouth bars of the Danube Delta change repeatedly, switching from a condition in which fluvial factors are predominant to a configuration in which marine processes are more relevant [Dolgopolova and Mikhailova, 2008].

[7] Nardin and Fagherazzi [2012] showed how the presence of waves leads to different bar morphologies depending on the relative importance of wave angle and wave-induced bottom shear stress. Large waves do not allow the formation of mouth bars either because of deflection of the river mouth jet, in case of oblique waves, or jet destabilization, in case of frontal waves. Wave angles between 45° and 60° have been identified as the most unfavorable for bar deposition. Nardin et al. [2013] found that waves can reduce the time of mouth bar formation and their distance from the river mouth leading to wave-induced deltas being shorter with respect to deltas experiencing no-waves conditions. According to Edmonds [2012], during the 2011 flood in Louisiana, USA, the Atchafalaya and Mississippi Rivers contributed to the same amount of land building, despite of the Mississippi River carrying a volume of sediments twice as large. This outcome was mainly due to the influence of coastal currents that can keep the sediment near the shoreline or disperse it in the ocean.

[8] Few studies have focused on the morphodynamic role of tides near river mouths; on the contrary, a rich literature is available on tidal inlets and their deposits. Since several tidal processes acting on river mouths are also present in tidal inlets, a comparison between the two systems is deemed necessary.

[9] Tidal inlets are channels maintained by tidal flow that connect the ocean with bays, lagoons, or salt marshes [e.g., FitzGerald et al., 2006]. The classical tidal inlet system consists of an ebb delta at the ocean side built around a deep central channel and two lateral shallower channels [Hayes, 1975; FitzGerald, 1976]. The main channel is ebb dominated while the other two are flood dominated. A flood delta is also present in the bay, resulting from the combination of flood channels and tidal flats [Hayes, 1975; FitzGerald, 1976, 1996, 2006; De Swart and Zimmerman, 2008].

[10] Tidal prism and reversing tidal currents are, to a large extent, the governing forces for inlet dimensions and sediment transport dynamics. River mouths present similar dynamics but only when the tidal discharge is considerably larger than the river discharge [FitzGerald, 1996]. According to Lanzoni and Seminara [2002], even if the tidal range is small, it is still possible to have a tide dominance situation in case of large tidal prism or limited wave action.

[11] Tidal inlets are also strongly affected by external drivers, in the same way that mouth bars are. Waves, alongshore currents, and sea-level rise can have a nonnegligible influence on inlet morphology [Dissanayake et al., 2008, 2009; Lanzoni and Seminara, 2002]. It has been shown that, during periods of increasing tidal prism, the inlets of the barrier islands in the Mississippi delta evolve from a wave-dominated configuration, with broad flood deltas, to transitional forms, with sand shoals choking the mouth throat, and finally to a river mouth, with a deep main channel and a well-developed ebb delta resembling a mouth bar [Levin, 1993]. Dissanayake et al. [2009] used the numerical model Delft3D to illustrate how shore-parallel tidal currents and their time lag with respect to inlet currents have influenced the evolution of the Ameland inlet in the Dutch Wadden Sea.

[12] Numerical models are particularly suitable to understand the behavior of fluvial and deltaic systems at medium to long time scales. Among others, numerical results on delta formation by Geleynse et al. [2010, 2011] show noticeable agreement with natural systems. Their simulations include deltaic mouth bars, distributary formation, and meander-bend evolution. Numerical results from Edmonds and Slingerland [2010] suggest that deltas without vegetation, as those that formed antecedent to the Devonian period, should show more fan-like characteristics, with fewer channels and delta plain lakes. This is mainly due to the fact that vegetation operates as a cohesive agent. In fact, they show that highly cohesive sediments favor formation of bird’s-foot deltas with complex floodplains and rugose shorelines. On the contrary, less cohesive sediments more likely form fan-like deltas.

[13] This work aims at analyzing concurrent effects of tides and fluvial discharge on the evolution of river mouth bars with the numerical model Delft3D. Our investigation is limited to a simple geometry and a homopycnal flow (i.e., with negligible buoyancy); wind waves and Coriolis forces are also neglected. Two extreme cases are taken into account and discussed: (1) a river mouth with a fluvial discharge much larger than the tidal discharge and therefore characterized by unidirectional flow; (2) a river mouth with a tidal discharge much larger than the fluvial discharge, allowing the flow to reverse direction.

[14] As we are going to show, different morphologies correspond to the two aforementioned cases. Therefore, a series of additional numerical experiments with intermediate ratios of river discharge to tidal discharge are also presented in order to identify the transition from one condition to the other. The manuscript is organized as follows: in section 2 we describe the model, the model geometry, and the boundary conditions utilized in the simulations. In section 3 we present key hydrodynamic results derived from the simulations. These results are divided into (i) effect of tides on jet spreading, (ii) effect of tides on the velocity field, and (iii) residual tidal currents. Section 4 deals with the implication of the hydrodynamic results for mouth bar formation and evolution. In section 5 we qualitatively compare our results to the morphology of river mouths in the Gulf of Mexico, USA. A set of conclusions is finally presented in section 6.

2. Model Description and Setup

[15] The formation of river mouth bars is studied using the computational fluid dynamics package Delft3D [Roelvink and Van Banning, 1994; Lesser et al., 2004]. Delft3D is a numerical model composed of a number of integrated modules. Together, they allow the simulation of hydrodynamic flow, sediment transport and related bed evolution, short-wave generation and propagation, and the modeling of ecological processes and water quality parameters [Lesser, 2004]. The Delft3D-FLOW module is able to solve the unsteady shallow-water equations in two (depth-
standing wave (90° between tidal level and tidal velocity fluctuations) and effort. Hence, a suitable landward flow discharge needs to be properly modeled. Factors like channel slope and shape, tidal motion in channels as well the interaction of tides with a riverine discharge can significantly alter tidal dynamics. For example, with an increase in riverine discharge, both the tidal wavelength and celerity decrease, while the tidal amplitude shows a more rapid upstream decay. The relationship between tidal elevation and maximum discharge depends on the nonlinear propagation of the tide in the lower reaches of the river and its floodplains. In river mouths, the tidal propagation falls somewhere between a progressive (0° of phase difference between tidal level and tidal velocity fluctuations) and standing wave (90° of phase difference). The main factors affecting the phase difference are the mouth topography and the river currents [Son and Hsu, 2011]. Savenije et al. [2008] considers the phase lag between high water and slack water as one of the main parameters for the classification of estuaries, and proposes explicit analytical and graphical solutions to determine it. Lanzoni and Seminara [1998] show how the phase lag increases with estuary convergence. Approaching the convergent limit for the estuary, the phase lag reaches 90°. Son and Hsu [2011] indicate the phase difference as a key process for the flux of sediments in the river. Accordingly, the flux of sediment is larger in the progressive wave condition.

Here we assume that the tidal wave mimics a standing wave at the river mouth, so that the phase lag between high water and high water slack is zero. This assumption is valid for strongly convergent estuaries [Lanzoni and Seminara, 1998; Savenije et al., 2008], tidal inlets [Bruun et al., 1978], or for rivers with a length much smaller than the tidal wavelength. A standing wave is also the standard assumption of tidal channel hydrodynamics [Boon, 1975; Fagherazzi et al., 1999; Fagherazzi, 2008]. Thus, we neglect further elaborations connected to the contribution of both landforming discharges and propagation of the tidal signal in the fluvial reaches.

In the tidal-dominated case, a constant river discharge is superimposed to an oscillating tidal discharge having a frequency of 30 deg/h (thus approximating a tidal frequency) and out of phase of 90° with respect to the water level. To impose the phase delay of 90° between discharge and water level at the river mouth we need to account for the propagation of the tide in the domain. In fact, we prescribe the water level at the seaward boundary, while the sinusoidal discharge is prescribed at the landward boundary. Since the two boundaries are distant in space a small phase lag is present.

According to Lanzoni and Seminara [2002], in case of tide dominance, the volume of water exchanged in each tidal cycle is at least 1 order of magnitude larger than river discharge. Therefore, we use a discharge equal to 10% of the maximum tidal discharge in order to simulate a relative small and constant riverine flow in the tide-dominated condition.

In fluvial-dominated conditions the discharge \( Q \) is 1500 m³/s. In the tidal-dominated case, the amplitude of the sinusoidal discharge is 570 m³/s. The choice of the discharge is based on the tidal inlet cross-sectional area \( A \) (m²) and tidal prism \( P \) (m³), further referred as \( A-P \) relationship. In fact, for tidal inlets field observations suggest that equilibrium exists, of the form of

\[
A = CP^q
\]  

[22] where \( C \) and \( q \) are empirical parameters [O’Brien, 1931; Jarrett, 1976; Tambroni et al., 2005; D’Alpaos et al., 2009, 2010].

Specifically, Tran et al. [2012] determined the \( A-P \) relationship for a series of hypothetical inlets using the process-based model Delft3D. Based on their results, we use the values of \( C = 2.14 \times 10^{-4} \) m²·s⁻¹ and \( q = 0.96 \) in order to evaluate the tidal prism in equilibrium with the cross-sectional area of the tidal-dominated case.

In the fluvial-dominated case the sediment concentration at the inflow boundary is constant, while in the tidal-dominated case we impose an equilibrium sand
concentration profile. This condition means that the sediment load entering through the boundary is adapted to the local flow conditions so that neither erosion nor deposition occurs at the river mouth.

[25] For the upper and lower boundaries, we impose a zero-flux boundary condition, consisting in imposing the gradient of the alongshore water level equal to zero (Figure 1). As shown by Roelvink and Walstra [2004], in case of unknown lateral boundary conditions it is preferable to impose the water level gradient and allow the model to determine water levels and current velocities.

[26] Finally, in order to make the solution well posed, we prescribe an oscillating water level at the seaward boundary. The water level varies with a semidiurnal tidal frequency (30 deg/h) and has an amplitude ranging from 0.25 to 2.5 m. Therefore our simulations represent both microtidal and mesotidal environments.

[27] The initial condition for each simulation consists of a constant bathymetric depth (3.5 or 4.5 m) and a water level equal to mean sea level. Both sediments in the basin and sediments delivered by the river and possibly entering from the seaward boundary are noncohesive and have a specific density of 2650 kg/m³, a dry bed density of 1600 kg/m³, and a median sediment diameter D₅₀ of 200 μm.

[28] A sensitivity analysis was performed in order to choose the proper grid size, and the results presented in this work are a good compromise between small computational time and accuracy of the solution. The grid size varies from 60 × 60 m at the margins of the basin to 20 × 60 m in the central part. The time step is set to 0.25 min. Every simulation runs for 12 model hours before starting the morphological evolution. This period is sufficient for a complete adaptation of the hydrodynamic simulation to the dynamic boundary conditions.

[29] In Delft3D, horizontal viscosity coefficients can be determined by means of two different closure models. The first is by specifying a constant or space-varying value of eddy viscosity. The second is by using a large eddy simulation model (HLES) that is added to a background value of eddy viscosity [Van Vossen, 2000; Vittenbogaard and Van Vossen, 2001]. In our simulations, we used the HLES model with a background eddy viscosity equal to 10⁻⁴ m²/s. Before adopting this value, we tested different values of background eddy viscosity, from 10⁻⁵ to 10⁻¹ m²/s, without finding any substantial influence on the simulations results from a qualitative point of view. Moreover, we compared the results from the HLES method with those for a constant value of eddy viscosity. We used 0.1 m²/s [Van der Wegen and Roelvink, 2008], 1 m²/s [Battjes, 1975], 2 m²/s, and 5 m²/s. Also in these cases we did not find any significant qualitative difference.

[30] Bed slope affects sediment transport as indicated by Ikeda [1982]. Two coefficients are used in Delft3D in order to adjust the bed load transport in the direction of the average bed stress (longitudinal slope factor) and in the direction perpendicular to it (transverse slope factor). In the simulations, we use default values of 1 and 1.5 for longitudinal and transverse slope factors, as suggested by Ikeda [1982]. However, these parameters are uncertain and their estimates are based on small-scale experiments. For example, Van der Wegen and Roelvink [2008] suggest a value of 5 for the transverse slope factor, while Dissanayake et al. [2009] use a value of 50. According to our sensitivity analysis, the fluvial-dominated case is slightly affected by variation of either longitudinal or transverse slope factor for values up to 50. The tidal-dominated case is influenced by variations in transverse slope factor, but significant changes only occur for values larger than 20, as the central channel gradually becomes shallower and wider. These findings confirm previous results by Dissanayake et al. [2009], for which a smaller transverse slope factor leads to a narrow and deep central channel, whereas the effect of the longitudinal slope factor is negligible.

[31] In order to account for the difference in time scales between hydrodynamics and morphological evolution, bottom changes per time step are scaled up by a morphological factor [Roelvink, 2006]. The morphological factor is 15 for the fluvial-dominated case and 450 for the tidal-dominated case.

3. Hydrodynamic Results

[32] At river mouths, the width generally exceeds the depth by a factor of at least 4, and width to depth ratios greater than 50 are common [Edmonds and Slingerlind, 2007]. Vertical motions of the water are usually less significant than horizontal ones and therefore the flow can be approximated with the shallow water equations [Özsoy and Unliata, 1982]. Note that the numerical model used in this contribution is in agreement with the shallow water approximation. As rivers empty into oceans or lakes, they expand as a confined jet, experiencing mixing, diffusion with the ambient fluid and decrease in sediment concentration and momentum. The jet is divided into two regions: the zone of flow establishment (ZOFE), having a core of constant velocity, and the zone of established flow (ZOEF), where a similarity function can be assumed for the transverse velocity profile. The transition from ZOFE to ZOEF marks the downstream location at which the turbulence generated by shearing along the margins of the jet affects the entire jet [Bates, 1953]. Jet theory has been largely used to describe rivers and inlets discharging into a standing water body [Bates, 1953; Wright and Coleman, 1974; Özsoy and Unliata, 1982; Wright, 1977; Rowland et al., 2010; Falcini and Jerolmack, 2010; Nardin and Fagherazzi 2012; Nardin et al., 2013].

[33] The presence of tides influences the hydrodynamic of the jet and the velocity field producing residual currents that affect the growth and the final shape of mouth bars. To study these currents, we first focus on the jet width and centerline velocity in the case of fixed bed. The self-similarity of the lateral velocity profile is assumed and the jet spreading is represented by the half width, defined as the lateral distance at which the velocity decreases to 5% of its centerline value [Özsoy, 1986]. According to Tee [1976], residual currents are the velocity field after the removal of all the sinusoidal components of tidal motion. They are time independent and can be obtained by the average over a tidal cycle. From a mathematical point of view, residual currents can be described through the fundamental equations of fluid mechanics simplified with the no-time dependence assumption [De Swart and Zimmerman, 2008; Nihoul and Ronday, 1975]. Tides can generate residual currents by several mechanisms such as nonlinear bottom friction,
nonlinear terms in the continuity equation, and the nonlinear advective term in the momentum equation. Residual currents can be studied using vorticity balance, as shown by Zimmerman [1981]: in order to create residual circulations in a tidal cycle, both the production of residual tidal vorticity and a net flux of vorticity through a closed curve are necessary. Residual currents account for part of the sediment transport [Van der Vegt et al., 2006] and therefore are considered as important indicators for mouth bar morphodynamics in our study.

[34] We compute residual tidal currents by calculating the difference between the velocity field with tides, averaged in a tidal cycle, and the velocity field in the absence of tides. Thus, we deplete the flow field from the influence of the riverine discharge, and we can focus our attention on nonlinear residual currents produced by tidal motion only. Two purely hydrodynamic simulations have been run for the calculation of each residual velocity field: one with tides and the other without tides, using the same bar morphology and a nonmoving bottom.

3.1. Jet Spreading

[35] The hydrodynamics of the problem is investigated for a flat and fixed horizontal bathymetry. According to Ozsoy and Ünlüata [1982], a turbulent jet in the presence of flat bottom with friction is characterized by an exponential growth of the jet width and a decay of the centerline velocity. For large width to depth ratios, as those here considered, the ZOFE is small relatively to the ZOEF [Ozsoy and Ünlüata, 1982] and, thereafter, the analysis will be restricted to the latter.

[36] To compute the velocity and jet half width in the case of flat bottom with friction, the following nondimensional parameters are defined (Figure 2):

\[
\xi = \frac{\bar{x}}{b_0}, \quad \zeta = \frac{y}{b(x)}, \quad \mu = \frac{fb_0}{8h_0}, \quad H(\xi) = \frac{h}{b_0}, \quad B(\xi) = \frac{\bar{b}}{b_0}, \quad U(\xi) = \frac{u}{u_0}, \quad \xi_c = \frac{x_c}{b_0}
\]

[37] where \(b_0\) is the inlet half width, \(b(x)\) is the jet half width, \(h_0\) is the inlet depth, \(h\) is the water depth, \(u\) is the velocity, \(u_0\) is the jet velocity at the inlet, and \(f\) is the Darcy-Weisbach friction coefficient. The parameter \(x_c\) is the end coordinate for the core region and marks the passage between ZOFE and ZOEF. The solution for the unknown centerline velocity and jet half width in case of flat bottom with friction is [Ozsoy and Ünlüata, 1982]

\[
U = \frac{e^{-\alpha \xi}}{\left[ e^{-\alpha \xi} + 2\alpha \left( e^{-\alpha \xi} - e^{-\mu \xi} \right) \right]^{1/2}}
\]

\[
B = \frac{e^{-\alpha \xi}}{\frac{\mu}{I_1} \left[ e^{-\alpha \xi} + 2\alpha \left( e^{-\alpha \xi} - e^{-\mu \xi} \right) \right]^{1/2}}
\]

\[38\] where \(\alpha = 0.05\) in the ZOEF and \(I_1\) and \(I_2\) are two constant values equal to 0.450 and 0.316, respectively. According to these equations, as the water depth decreases, the jet spreading increases, and the jet half width has an exponential dependence on the distance from the river mouth. For simplicity, in this work the jet half width has been approximated using a simple exponential curve of the form:

\[
B(\xi) = a \cdot e^{\lambda \xi}
\]

\[39\] where \(\lambda\) is the nondimensional e-folding length representing the spreading of the jet.

[40] The jet spreading for the river-dominated case is evaluated at different instants of the tidal cycle. The largest spreading occurs at low tide, whereas the smallest spreading corresponds to high tide (Figure 3). The mean

![Figure 2. Sketch of a turbulent jet exiting a river mouth.](image)

![Figure 3. Jet half width for different tidal conditions. The points represent locations having an along \(\xi\) velocity equal to the 5% of the centerline value and have been fitted with equation (4) to obtain the solid lines. Black line: jet spreading at low tide \((h_t = 2.5 \text{ m})\). Green line: jet spreading at high tide \((h_t = 2.5 \text{ m})\). Blue line: averaged jet spreading in time with a tidal amplitude \(h_t = 2.5 \text{ m}\). Red line: jet spreading without tide.](image)
spreading computed by averaging the spreading over several tidal cycles is larger than the spreading in the case without tides (Figure 3). This effect is present in all tested tidal amplitudes (ranging from 0.25 to 2.5 m). Moreover, the higher is the tidal range, the more evident is the increase in average spreading. This result can be explained by a nonlinear relationship between water depth and jet spreading (Figure 4a). To further investigate this aspect, the jet half width has been computed for a series of additional numerical experiments, in which a constant discharge is prescribed at the landward boundary and a constant mean water level is prescribed at the seaward boundary (Figure 5). At the seaward boundary, where we impose the varying water level, the velocity is negligible at high and low tide (slack water), a typical condition of a standing wave. On the contrary, near the river mouth, the maximum flood and ebb velocities occur at high and low tide, with slack water taking place near mean sea level (progressive tidal wave) (Figures 5a and 5b).

[46] The transitional region between the standing wave and the progressive wave conditions can be detected through the cross correlation in time between water level and depth-averaged velocity, which determines the delay (Figure 5c). The discontinuity point indicates the transition between the two conditions.

Figure 4. (a) The λ coefficient as a function of water depth; (b) λ coefficient as a function of the centerline velocity at the inlet boundary condition. Colored dots (Figures 4a and 4b) have been obtained with additional numerical experiments with constant discharge and variable depths (Figure 4a), or with constant depth (4.5 m) and variable discharge (Figure 4b). Continuous lines are representative of the analytical solution of Özsoy. Red dots refer to a Chezy coefficient of 45 m$^{1/2}$/s, while blue dots refer to a Chezy coefficient of 30 m$^{1/2}$/s. Gray arrows indicate the range of depth and velocity of the present work.
This transition is caused by the interaction between riverine discharge and seaward water level. At low tide the water depth in the basin is low, thus promoting a drawdown water profile in the river with accelerated flow near the mouth [Lamb et al., 2012]. On the contrary, as the seaward depth rises, a backwater profile establishes, with milder water gradients and limited flow acceleration near the mouth. Figure 6c shows the relationship between velocity gradient and water level in time. The velocity gradient is maximum when the water level is minimum, demonstrating the presence of an accelerated flow during periods of low water level. The velocity is therefore in phase with tidal elevation, peaking at low tide.

3.3. Tidal Residual Velocities

To determine the effect of tides on mouth hydrodynamics and bar formation we compute the tidal residual velocities by integrating the velocity field in a tidal cycle and subtracting the flow field due to the river discharge in the absence of tides [Zimmerman, 1981].

In the fluvial-dominated case, the residual velocities form two lateral bands that bifurcate within a short distance from the river mouth (Figure 7). These bands are produced by an increase in spreading at low tide, which is only partly compensated by a decrease in spreading at high tide (Figure 3). A residual current is also present in the river, and it is caused by high velocities during low tide. In fact a drawdown water surface gradient establishes at low tide that accelerates the flow (Figure 6). Again this acceleration is only partly compensated by a deceleration due to the backwater effect at high tide, yielding a net residual current exiting the river.

In the tidal-dominated case, the residual velocity is associated with the difference between the confined flow of the turbulent jet during the ebb phase and the sink flow during the flood phase, which is radially distributed across the entire domain [Chao, 1990; Tambroni et al., 2005]. The residual velocity field displays two counterrotating vortices exiting the river mouth (Figure 8), which agrees with previous results in tidal inlets [Chao, 1990; Van der Vegt et al., 2006].

4. Bar Morphology

According to Edmonds and Slingerland [2007], mouth bar formation is due to the reduction of the sediment transport rate when the turbulent jet expands into a standing body of water. More specifically, the decrease in the jet momentum flux is responsible for the formation of mouth bars. After initial deposition of sediments, the mouth bar progrades and aggrades until the depth over the bar is shallow enough to create an upstream fluid pressure that is capable of forcing the fluid around the bar rather than over the bar. The mouth bar then widens, creating a classic triangular deposit in planar view.

Following the notation used by Edmonds and Slingerland [2007], h and h₀ are defined as the water depth above the peak of the river mouth bar and the channel depth at the landward boundary (the water depths being referred to the mean sea level). In order to explore changes in residual currents during bar evolution, additional numerical experiments have been conducted during the growth of a bar, for ratio h/h₀ equal to 0.8, 0.6, and 0.4. Successively, the morphology of the bottom was frozen, i.e., maintained constant in time, preventing the bar from growing or being demolished, and two purely hydrodynamic simulations were run, one with tides and one without them. The residual velocities are computed from the difference between the velocity averaged over a tidal cycle and the velocity
without tide (Figures 7b–7d and 8b–8d). Residual velocities are found to increase when the mouth bar grows. Under dominant fluvial conditions, for example, for a tidal amplitude of 0.75 m, the maximum residual velocity is around 0.05 m/s for \( h/h_0 = 1 \), and 0.15 m/s for \( h/h_0 = 0.4 \).

The presence of tides also promotes variations in bottom morphology. Even though fluvial conditions dominate, tides increase bar widening (Figure 9). At the beginning of bar formation, the greater is the tidal range, the higher are the residual velocities, which in turn increase sediment transport rates. Figure 10 shows longitudinal profiles of the bar for increasing tidal amplitudes. Same colors correspond to same intervals of time. It is possible to notice that the part in gray, representative of the initial stage of mouth bar formation, is higher in the presence of tides. The two lateral bands of residual velocity spread the sediments at the two sides of the river mouth and the bar immediately widens (Figure 10b). When the bar is high enough, the flow is laterally channelized during the ebb phase, when the maximum velocities occur. Hence, the bar confines the flow and favors the bifurcation process. As the magnitude of the two lateral bands of residual velocity increases, the lateral spreading grows, as well as the tendency of the bar to widen. The widening rate also starts prevailing upon the accretion rate, producing a larger but shallower bar (Figure 10). Aggradation rates therefore diminish in the final stages of bar development for large tidal amplitudes (Figure 10).

The tidal-dominated case may be regarded as the sum of the fluvial-dominated case plus the influence of the tidal discharge. In our simulations the constant river discharge is only 10% of the maximum tidal discharge at the river mouth. However, it is enough to accelerate the

![Figure 7](image1.png)  
*Figure 7.* Residual currents for the fluvial-dominated case, at different stages of mouth bar formation. The tidal amplitude is 2.5 m. The isolines in each plot represent the location of the bar.

![Figure 8](image2.png)  
*Figure 8.* Residual currents for the tidal-dominated case at different stages of mouth bar formation. The tidal amplitude is 2 m. The isolines in each plot represent the location of the bar. The two black arrows are indicative of two counterrotating eddies.
process of bar growth, because of a larger sediment discharge due to the presence of the additional riverine flow, and to modify the configuration of the deposit with respect to the fluvial-dominated case. Similar to what happens in tidal inlets, the oscillating flow triggered by the tidal prism leads to the formation of a central channel (Figures 8 and 11) due to strong ebbing currents in the central part of the domain. The presence of the riverine discharge leads instead to the formation of two lateral channels, similar to the fluvial-dominated case. Different from the case of tidal

**Figure 9.** Fluvial-dominated case: cumulative erosion and sedimentation for different tidal amplitudes $h_t$ after 3 days of numerical simulation and morphological factor of 15.

**Figure 10.** Morphology of river mouth bars for the fluvial-dominated case. (a) Longitudinal and (b) transverse sections of river mouth bars. Each plot corresponds to a different tidal range. Same colors correspond to same temporal intervals: gray from 0 to 6 h, blue from 6 to 12 h, gray blue from 12 to 18 h, light blue from 18 to 24 h, light green from 24 to 30 h, green from 30 to 36 h, yellow from 36 to 42 h, pink from 42 to 48 h, red from 48 to 54 h, dark red from 54 to 60 h.
inlets, the two lateral channels are ebb dominated. In fact, for tidal inlets, the basic ebb delta configuration is a central channel, dominated by ebb currents, surrounded by a swash platform and two lateral channels dominated by flood currents [FitzGerald, 1996; Davis, 1997]. In tidal inlets, flood dominance in the lateral channels is due to the different flow fields emerging in ebb and flood: while in flood phase, the flow radially distributes near the inlet and in ebb phase the flow is confined in a central jet [Tambroni et al., 2005].

Figure 12 shows the velocity for three tidal cycles at the seaward mouth and at the right boundary. At the beginning of the simulation the system behaves as a tidal inlet. Therefore, a main central channel forms, dominated by ebb flow, and sediments are deposited at the lateral sides of the channel. Due to the formation of these initial deposits and the presence of river discharge, low slack water is delayed near the river mouth. As a result, a transitional period exists during which slack water is already reached at the right boundary, but there is still a positive flow at the river mouth. The water flows through the main channel and it is laterally deviated because of the presence of the initial deposits. During this period, the system behaves as in the fluvial-dominated case, where a constant discharge is prescribed. Moreover, since this transitional period corresponds to minimum water levels and maximum spreading, the flow is further constrained in the lateral channels with the side deposits acting as a flow barrier.

Figure 12b shows the depth-averaged velocity field around low water slack at the right boundary for a bar elevation corresponding to $h_0/h_0 = 0.6$. The maximum velocities occur in points where the lateral channels are going to form. Once the channels are incised, the confined flow occupies them during ebb, leading to a velocity higher than during flood. To sum up, while a simple bifurcation characterizes fluvial-dominated conditions, the tidal-dominated case displays three principal channels. Finally, larger tides accelerate bar growth and widen the bar deposit similar to what was observed for the fluvial-dominated case (Figures 9 and 11).

To determine the transition between bifurcation (river-dominated case) and the presence of a central channel (trifurcation, tidal-dominated case), we performed a series of additional numerical experiments for two different initial basin depths ($h_0 = 4.5$ m and $h_0 = 3.5$ m) and two different friction coefficients ($C = 45$ m$^{1/2}$/s and $C = 30$ m$^{1/2}$/s). The transition is a function of the nondimensional tidal amplitude and the relative importance of tidal velocity with respect to river velocity (Figure 13). We find a robust criterion for the formation of a central channel ($R^2 = 93\%$):

$$\frac{u_0 - u_t}{u_0 + u_t} < 0.34 - 1.3 \frac{h_t}{h_0}$$

where $u_0$ is the river velocity, $u_t$ the maximum tidal velocity, $h_t$ the tidal amplitude, and $h_0$ the water depth. For a given water depth and tidal velocity, the transition from river-dominated to tidal-dominated conditions requires a small river velocity with small tides. Note that a large ratio $h_t/h_0$ maximizes the drawdown effect on river discharge, thus promoting fast shallow flows in the lateral channels and bifurcation. Moreover, for $u_0 > 2 u_t$ (river discharge double the tidal discharge), the bifurcation forms independently of tidal oscillations in the basin (Figure 13). In summary, oscillations in tidally generated
discharge favor the formation of a central channel, whereas tidal oscillations in the basin favor bifurcation and fluvial mouth bars.

5. Discussion

[59] Based on the results of our numerical analysis, we present the following conceptual model of bar formation under the effect of tides. In the fluvial-dominated case, tides enhance jet spreading (Figure 3), which, in turn, promotes lateral residual currents that facilitate the formation of bifurcating channels (Figure 7). Moreover, at low tide even an incipient mouth bar becomes an obstacle to the flow, given the limited water depths. As a result, more sediments are deposited on the mouth bar, speeding up its initial development (Figure 10). Once the two bifurcating channels are emplaced, strong flow establishes at low tide, magnifying the residual currents (Figure 7). These fast, shallow flows transport large quantities of sediments to the sides, creating a wider mouth bar (Figure 9).

[60] In the tidal-dominated case, the remarkable asymmetry between flood and ebb flow fields creates two residual vortexes that magnify the flow along the central axis (Figure 8). As a result, a central channel forms (Figure 11). The oscillatory flow at the mouth prevents the silting of the central channel and two deposits form at the sides (Figure 11). These deposits trigger two bifurcations and the formation of two lateral channels, also facilitated by the confined flow at low tide (Figure 12).

[61] The final bar configurations display similarities to natural systems. Figure 14 shows the mouth morphology of two rivers, the Apalachicola and Suwanee rivers in Florida, USA, with fluvial discharge and tidal features similar to those considered in this work. The Apalachicola River is fluvial dominated, with a drainage area of approximately 50,000 km² and a mean discharge near Sumatra (FL) of 700 m³/s [U.S. Geological Survey, 2012]. The Suwanee River is tidal dominated, with a drainage area of 26,000 km² and a mean discharge of 170 m³/s near Suwanee city (FL) [U.S. Geological Survey, 2010].

[62] The morphology of the Apalachicola mouth bars is dominated by bifurcations (Figure 14d). These landforms can thus be associated with the relatively small influence of the tidal discharge with respect to the fluvial discharge. On the contrary, in the Suwanee River, the tidal discharge seems to have a considerable role, indicated by the strong discharge favor the formation of a central channel, whereas tidal oscillations in the basin favor bifurcation and fluvial mouth bars.

Figure 12. Tidal-dominated case with tidal amplitude of 2.5 m. (a) Velocity variations in time at point A (green line) and point E (red line). Water level variations at point A (black line) and E (blue line). The location of the points is indicated in Figure 1 (A is the closest to the river mouth, E the farthest). Note that when the red line crosses the zero value (near the minimum in water level), the green line has still a positive value. (b) Velocity field at low water slack for height of the bar corresponding to $h/h_0=0.6$. The isolines indicate the location of the bar. Note that peaks of velocities occur where the ebb-dominated channels form.

Figure 13. Occurrence of bifurcation or trifurcation at a river mouth as a function of nondimensional tidal amplitude and relative ratio between fluvial and tidal velocity. The green empty symbols represent trifurcations while the red-filled symbols are bifurcations. Circles $C = 45 \, m^{1/2}/s$, $h_0 = 4.5 \, m$; diamonds $C = 30 \, m^{1/2}/s$, $h_0 = 4.5 \, m$; triangles $C = 45 \, m^{1/2}/s$, $h_0 = 3.5 \, m$; squares $C = 30 \, m^{1/2}/s$, $h_0 = 3.5 \, m$. The straight green line represents the transition criterion.
bidirectionality of the flow (Figure 14b). Here mouth bars are characterized by trifurcations, we also notice cases in which more than three channels are connected to the same node, and we ascribe this morphology to the complexity of the coastal bathymetry and variability in tidal and river discharge that were not captured in our simulations.

Trifurcations do not form only in the presence of large tides; Storms et al. [2007] show that a river debouching in a deep basin can also produce a central channel. However, in their Delft3D simulations the basin is much deeper than ours, as well as deeper than Apalachicola Bay and the Suwanee river mouth. Similarly, bifurcations can also exist in tidally dominated rivers, locally favored by the complexity of the coastal landscape. Our results are therefore valid in a statistical viewpoint, with tidal-dominated rivers giving rise, on average, to more trifurcations.

By modifying mouth bar morphology, tides can also affect the evolution of the entire delta. Wider and shallower bars favor vertical infilling with respect to lateral progradation. We therefore expect a prevalence of thin and spatially uniform deposits with fewer vertical discontinuities. Lack of steep gradients in mouth deposits inhibits possible avulsions, locking the distributaries in place. Wide mouth bars can also affect nearby distributaries, and fewer distributaries are needed to feed the entire delta front. Spatially extensive deposition would also maintain the mouth bars under water for a longer time, delaying vegetation encroachment and related sediment bioturbation.

Our analysis has implications for future restoration projects of deltaic environments. The basic scheme for deltas restoration consists in diverting sediments from the main channel to subsiding and eroding areas where they can eventually create new wetlands [Paola et al., 2011]. According to Edmonds [2011], the success of such a restoration procedure is based on the rates of land growth estimated under different configurations of river diversion. Thus, hydrodynamic conditions of areas surrounding the river diversion are crucial to determine whether sediments would deposit near the shore or would be moved into the deep ocean. Only in the first case the diversion can effectively build new land.

Figure 14. (c) Locations of Apalachicola River and Suwanee River, Florida, USA. (a) Apalachicola river discharge near Sumatra, FL (USGS station 02359170; from 9 April 2012 to 12 April 2012) and (d) satellite image of Apalachicola mouth bars (29°45′N, 84°35′W). (b) Suwanee river discharge near Suwanee, FL (USGS station 02323592; from 9 May 2012 to 11 May 2012) and (e) satellite image of Suwanee mouth bars (29°17′N, 83°09′W). Data have been obtained from the annual water data report by the U.S. Geological Survey. White arrows indicate channels and white dots indicate mouth bars.
For the fluvial-dominated case, the presence of tides gives rise to residual currents causing the lateral spreading of sediments. These residual currents could eventually transport very fine sediment fractions far away from the river mouth and partially prevent land building. Additionally, vegetation encroachment is obstructed due to the fact that mouth bars tend to be shallower and are maintained under water for a longer time. Vegetation is generally recognized for its role in accelerating sediment deposition by means of flow deceleration and sediment trapping. As vegetation expands, detritus and fine-grained sediments mix into the soil matrix, which gradually becomes less dense, finer, and more cohesive [e.g., Feagin et al., 2009; Mariotti and Fagherazzi, 2010; Nepf, 2012].

According to Edmonds [2012a], finer-graded sediments lead to the creation of new lands characterized by a rough shoreline, with small bays and fewer but larger interior lands, thus favoring trapping of incoming sediments and sheltering land from waves’ erosion. Since vegetation establishment would be obstructed, all these positive feedbacks would be less effective. For the tidal-dominated case, due to the presence of the central channel, we expect the formation of more elongated deposits, with fewer but larger interior sediment islands, providing advantageous conditions for vegetation growth, encroachment, and protection from wave erosion. The aforementioned considerations suggest that tides would make sediment diversions and restoration practices less efficient in a river-dominated case, while they would favor them in the presence of a large tidal prism.

6. Conclusions

Mouth bar formation is a complex phenomenon, owing to the interactions of several processes. Among others, tides play an important role in determining the final configuration of the mouth deposits. Here, the effect of tides on mouth bars has been analyzed using the computational fluid dynamics package Delft3D. The present work focuses on two end-member cases: a river with a constant and significant fluvial discharge debouching in a basin with tides (small tidal prism) and a river in which the fluvial discharge is small compared to tidal fluxes (large tidal prism).

In the fluvial-dominated case, tides increase, on average, jet spreading, producing mouth bars that are wider with respect to the case without tides [e.g., Edmonds and Slingerland, 2007]. Fast flows during low tide triggered by the drawdown water surface profile [Lamb et al., 2012] transport sediments to the sides, increasing the width of the depositional area. Tidal oscillations therefore act as a dispersion process similarly to waves [Jeronmack and Swenson, 2007; Nardin and Fagherazzi, 2012; Nardin et al., 2013]. Moreover, tidal oscillations favor flow bifurcation and the formation of a central mouth deposit. In fact, even shallow deposits can become obstacles for the flow at low tide.

In the tidal-dominated case, a central channel forms giving rise to a bifurcation that is commonly displayed in tidal inlets [Hayes, 1975; Fitzgerald, 1976]. Flow reversal during flood maintains the central channel flushed and produces lateral deposits that trigger two lateral bifurcations and the formation of side channels. The occurrence of a bifurcation or trifurcation depends on the relative strength of the tidal discharge with respect to the fluvial discharge and the ratio between tidal amplitude and water depth. A high fluvial discharge favors a bifurcation, which is always present for a river velocity more than twice the tidal velocity. A high tidal range at the river mouth also favors bifurcations with respect to trifurcations. Two examples from the Gulf of Mexico, USA, agree well with our findings. The Apalachicola Delta, characterized by a large fluvial discharge, is dominated by bifurcations, while the mouth of the Suwanee River, characterized by an oscillating discharge, displays on average more trifurcations.

Acknowledgments. This research was supported by Exxon Mobil Upstream Research Company award EM01830, the ACS-PRF program award 51128-N08, by NSF award OCE-0948213 and through the VCR-LTER program award DEB 0621014.

References


Dissanyake, D. M. P. K., J. A. Roelvink, and M. Van der Wegen (2008), Effect of sea level rise on inlet morphology, VII Conference on Coastal and Port Engineering in Developing Countries, Dubai, UAE.

Dissanyake, D. M. P. K., J. A. Roelvink, and M. Van der Wegen (2009), Modelled channel pattern in schematised tidal inlet, Coastal Eng., 56, 1069–1083.


direct numerical simulation and large eddy simulation (TAICDL), University of Texas at Arlington, Tex.


