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Experimental delta evolution in tidal environments: Morphologic response to relative sea-level rise and net deposition

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Key Points:

- Laboratory experiments of tide-influenced deltas are able to produce composite deltas wherein different processes create varying morphologies across the delta.

- All other parameters equal, experimental tide-influenced deltas show transgressive shorelines as compared to fluvial deltas under relative sea-level rise.
- Net deposition data reveal that the major effect of tides is the removal of fluvial sediment that would otherwise be deposited in the delta topset.

Keywords

Tide-influenced delta; laboratory experiment; fluvial delta; relative sea-level rise; tidal channels

Abstract

Tide-influenced deltas are among the largest depositional features on Earth and are ecologically and economically important as they support large populations. However, the continued rise in relative sea-level threatens the sustainability of these landscapes and calls for new insights on their morphological response. While field studies of ancient deposits allow for insight into delta evolution during times of eustatic adjustment, tide-influenced deltas are notoriously hard to identify in the rock record. We present a suite of physical experiments aimed at investigating the morphological response of tide-influenced deltas subjected to relative sea-level rise. We show that increasing relative tidal energy changes the response of the delta because tides effectively act to remove fluvially deposited sediment from the delta topset. This leads to enhanced transgression, which we quantify via a new methodology for comparing shoreline transgression rates based on the concept of a 'transgression anomaly' relative to a simple reference case. We also show that stronger tidal forcing can create composite deltas where distinct land forming processes dominate different areas of the delta plain, shaping characteristic morphological features. The net effect of tidal action is to enhance seaward transfer of bedload sediment, resulting in greater shoreline transgression as compared to identical, yet purely fluvial, deltaic systems that exhibit static or even regressive shorelines.

Introduction

River deltas are landforms shaped by the mutual interplay of coastal processes and fluvial input of water and sediments delivered by means of interconnected channel pathways (Tejedor, Longjas, Zaliapin, & Foufoula-Georgiou, 2015). Deltas evolving under the influence of tides are of particular importance as most of the largest modern rivers feed either tide-dominated or tide-influenced deltas (Hoitink, Wang, Vermeulen, Huismans, & Kästner, 2017; Tänavsuu-Milkeviciene & Plink-Björklund, 2009), and their associated subaerial and subaqueous landmasses place them among the largest sedimentary environments on Earth (FIGURE 1). Moreover, the extensive fertile plains created by the combined action of fluvial and tidal processes host large human populations (Goodbred & Saito, 2012), providing support for some of the largest economies worldwide (Giosan, Syvitski, Constantinescu, & Day, 2014; Hoitink et al., 2017; Syvitski et al., 2009). Examples of tide-influenced deltaic systems (FIGURE 1) include the tide-dominated Changjiang Delta of the Yangtze River which hosts many of the world's largest cities - Shanghai alone has a population density of 2,145 inhabitants per square kilometer (Tian, Jiang, Yang, & Zhang, 2011) - as well as the tide-dominated Ganges-Brahmaputra Delta which hosts nearly 2% of the world's population although half of its area is within 5 m of mean sea level (Akter, Sarker, Popescu, & Roelvink, 2016). The rise in relative sea-level (RSL) currently being experienced by most deltas worldwide - due to eustatic sea level changes, as well as natural and anthropogenically induced subsidence (Giosan et al., 2014; Syvitski et al., 2009; Syvitski & Saito, 2007; Vörösmarty et al., 2009) - threatens the stability of these landscapes and subsequently endangers the populations and human activities they support (Giosan et al., 2014; Syvitski et al., 2009; Syvitski & Saito, 2007; Vörösmarty et al., 2009). While many recent studies investigate the role played by tides in the morphodynamic evolution of fluvial deltas under steady RSL (Goodbred & Saito, 2012; Hoitink et al., 2017; Rossi et al., 2016), little research exists on the response of deltas subjected to the combined actions of tidal processes and RSL rise (Jerolmack, 2009; Lentsch, Finotello, & Paola, 2018). This may be partially due to the limited number of tidal deltas identified in the rock record (Plink-Björklund, 2012), as several characteristic tidal features that might help one to recognise the signature of tidal processes in fluvio-deltaic deposits (e.g., funnel-shaped tidal channels, tidal mouth bars, etc.) are typically not well preserved (Dalrymple & Choi, 2007). Physical experiments can help bridge this knowledge gap, allowing one to capture delta evolution at a spatial and temporal resolution otherwise impractical in the field (Braat, Lokhorst, & Kleinhans, 2018; Kleinhans, van der Vegt, Terwisscha Van Scheltinga, Baar, & Markies, 2012; Kleinhans, Van Rosmalen, Roosendaal, & Van Der Vegt, 2014; Lentsch et al., 2018; Leuven et al., 2018; Malverti, Lajeunesse, & Méttivier, 2008; Paola, Straub, Mohrig, & Reinhardt, 2009; Stefanon, Carniello, D'Alpaos, & Rinaldo, 2012). Recent studies demonstrated that physical experiments are capable of reproducing several typical tidal features, such as ebb deltas (Kleinhans et al., 2012), tidal channel networks (Stefanon et al., 2010; Vlaswinkel & Cantelli, 2011) and tidal estuaries (Kleinhans, Van Rosmalen, et al., 2014). However, all of these experiments functioned by allowing tides to rework a layer of previously emplaced sediment and, due to the limitations of the physical setup or the goals of the research, no sediment was introduced via a fluvial system. Further research is therefore needed in order to clarify the role played by tidal currents and RSL changes in the morphodynamic evolution of depositional fluvio-deltaic systems. Here we examine, by means of physical laboratory experiments, the morphological response of fluvial deltas, subjected to different rates of steady RSL rise, and tidal forcing of varying amplitudes. We propose two new methodologies, based on analysis of high-resolution topographic scans collected during each experiment, to quantify the evolution of delta plains in terms of net deposition and to measure differences between the predicted and actual shoreline migration. By varying the fluvial-to-tidal energy ratio, sediment discharge and RSL rise rate, we provide novel insight not only into the differences in how tidal and non-tidal deltas respond to RSL rise, but also into how a delta system adapts morphologically as external forcing becomes more dynamic.

Research approach

Experimental setup

We performed our experiments using the “Delta Basin 2” (DB-2) experimental facility at the Saint Anthony Falls Laboratory (SAFL), University of Minnesota. DB-2 is a square 5 x 5 m basin, with an approximate depth of 0.5 m (FIGURE 2a), designed to study delta evolution (e.g., Martin et al., 2009; Lentsch et al., 2018). The sediment used in the present study is rounded walnut-shell sand (D_{50} ~ 320 μm) that, owing to its low density (1350 kg/m^3), can be easily entrained and deposited by both tidal and fluvial currents at this scale (Baumgardner, 2015). Well-mixed sediment and water, whose quantities are independently and remotely controlled, are fed from a corner of the basin to create a quarter-circular delta (FIGURE 2a). The feed point source is located behind a gravel diffuser that helps to evenly distribute sediment and water while minimising localised scour. The base level within the basin is computer controlled through a motorised weir connected to a sonar sensor taking water level measurements every 5 seconds. The main basin is connected to a 2 x 2 m, 0.5 m deep, auxiliary basin by means of two remotely controlled industrial pumps, which deliver sinusoidal tides with varying amplitude and period. During the flood phase of the tidal cycle, water is pumped into the main basin from the auxiliary basin by the flood pump. After the specified upper tidal limit is reached, the flood pump is remotely switched off allowing for a period of slack water in the main basin before the ebb pump is activated. The latter pumps water out of the main basin, thus progressively reducing the water level down to the lower limit of the tidal range, where another momentary period of slack water occurs. Both the tidal amplitude and the tidal period are computer controlled, and any deviation from the specified water elevation is identified by the sonar-measured water depth and communicated back to the control system that progressively refines discharges by adjusting the pump voltage. This feedback correction allows for error of less than 1 mm in water elevation.

In all, 10 experiments were conducted with varying water and sediment discharges, tidal amplitudes and RSL rise rates. Different tidal amplitudes were selected to make the system either tide- of river-dominated, based on the ratio between tidal (Ω_T) and fluvial (Ω_F) power delivered to a unit length of delta shoreline per unit time (power/length, [W/m]) (Table 1). These powers are derived from modified forms of the metrics developed by Baumgardner (2015), which have been used to classify deltas both in the field and in physical experiments (Baumgardner, 2015; Lentsch et al., 2018). Fluvial power (Ω_F) can be represented as the bulk flux of mean kinetic energy delivered to the delta plain, per unit length of shoreline. Being $m=\rho Q$ the mass of flowing water per unit time, where Q represents the river discharge characterized by an average velocity u, the river kinetic energy delivered to a unit length of delta shoreline per unit time (t) is equal to

$$\frac{E_k}{tL_s} = \frac{1}{2}\frac{\rho Q u^2}{L_s} \tag{1}$$

Water velocity can be estimated considering the bottom shear stress (τ_b) due to uniform equilibrium flow (i.e., normal flow) conditions over a bed with mean streamwise slope S:

$$u^2 = \frac{2\tau_b}{\rho C_f} = \frac{2ghS}{C_f} \tag{2}$$

where h is a characteristic flow-depth of the riverine input (~10 mm) and C_f is the drag coefficient (~1.5 x 10^{-3} for laminar flow). Fluvial power finally reads:

$$\Omega_f = \frac{\rho g Q S h}{C_f L_s} \tag{3}$$

Following the approach adopted by Baumgardner (2015), tidal power (Ω_T) is defined as the potential gravitational energy of the tidal prism at high tide relative to the low tide. Ideally representing the tidal prism as triangle of height H_T at a given longitudinal section of the deltaic deposit, the tidal prism per unit length of shoreline reads:

$$\frac{P}{L_s} = \frac{1}{2}\frac{H_T^2}{S_T} \tag{4}$$

where L_s is the length of the delta shoreline and S_T is the slope of the intertidal zone. The potential energy associated to P corresponds to the weight of P multiplied by the tidal height:

$$\frac{E_T}{L_S} = \frac{1}{2}\frac{\rho g H_T^3}{S_T} \tag{5}$$

Since half of the tidal period T is required to lift the whole tidal prism from low tide to high tide, tidal power [W/m] finally reads:

$$\Omega_T = \frac{\rho g H_T^3}{S_T T} \tag{6}$$

No attempts were made to scale the tides or other parameters to any particular modern deltaic system. Rather, we aimed to explore the overall morphologic effect of tides and RSL rise on delta evolution. Previous studies have proved the effectiveness of physical experiments for reproducing stratigraphy and geomorphologies similar to those seen in the field in response to short term base-level cycles meant to represent gradual transgressions or regressions (Ethridge, Germanoski, Schumm, & Wood, 2009). The combined effects of tides and RSL rise are investigated here employing a volume-based approach (see next sections) for which a rigorous geometric scaling is not required nor entirely feasible (see Paola et al., 2009; Kleinhans et al., 2014a). Nonetheless, comparisons to field-scale deltas are still permitted by studying representative length and time scales of specific morphological features and the dynamics of the system, such as channel dimensions and avulsion timescales (e.g., Wickert et al., 2013)

A tidal period equal to 2 minutes was imposed, which is shorter than the characteristic channel-avulsion timescale, but long enough to allow for a quasi-steady tidal flow capable of remobilizing sediments during both the ebb and flood phases (Baumgardner, 2015). To increase the temporal efficiency of the experiments, a quarter-circular platform consisting of medium sand was initially hand-laid to within 5 cm of the delta plain's intended starting elevation. Sea level was then raised and the remaining 5 cm of deltaic deposit thickness deposited via the fluvial system until a delta topset of approximately 2 m radius was formed. As the overall goal of this research is to understand how deltas with varying fluvial and tidal energy respond to sea-level rise, we designed the experiments to decouple the system's response to RSL rise alone from its response to combined RSL rise and tides. Therefore, the amount of sediment discharge (Q_s) was set to theoretically maintain a constant delta topset area (A_T) characterised by a 2 m radius shoreline in equilibrium with the rate of sea-level change ($\dot{H}_{SL}$):

$$Q_s = \frac{\dot{H}_{SL} A_T}{f} \quad (7)$$

where f is the fraction of sediment discharge trapped in the topset. Here a value of 0.9 was used for f, based on test experiments conducted at SAFL (Baumgardner, 2015; Lentsch et al., 2018). It is important to note the relationship (7), which is derived from a simple mass balance, holds theoretically for a purely fluvial system with no tidal influence, for which the shoreline should remain approximately static for the given sea-level rise rate. Hence, any observed deviation in shoreline position in experiments with active tides should be the result of tidal processes.

Data collection

Primary data collection in these experiments consisted of overhead time-lapse photography and periodic digital elevation model (DEM) scans. Overhead photographs were taken by means of a digital single-lens reflex camera affixed above the basin with a wide-angle lens to capture the entirety of the delta surface. The camera's resolution and height above the basin were such that each pixel is approximately 1 mm by 1 mm in Cartesian space. As a wide angle lens was used, the outer portions of the images were subjected to barrel distortion. Every image was therefore orthorectified in a batch process so that all portions of the deposit were correctly scaled. The water delivered into the basin for simulating fluvial discharge was dyed blue with an industrial food colourant to aid in channel identification from overhead photos. This highly concentrated dye is introduced as a drop every few seconds and therefore has no significant effect on net fluvial discharge.

Additionally, DB-2 has a data cart equipped with a high-resolution line scan camera for DEM collection (FIGURE 2b). This camera has a plan view resolution of 1 mm by 1 mm and a vertical resolution of less than 1 mm. In order to collect DEM data, the experiments were periodically paused and the surfaces of the deposits allowed to drain slowly so as to not affect the surface morphology. The data cart took several overlapping scans, which were later stitched together into a single continuous DEM. Water level was then slowly raised to the pre-pause position and the experiment continued until the next scan time.

Data Analysis Methods

DEM data retrieved from our experiments were first employed to investigate the evolution of delta shoreline position through time. Starting from the sediment and water point source, we derived longitudinal profiles of the delta deposit in a radial pattern from θ=1° to θ=89° for a total of 89 profiles per hourly DEM scan (FIGURE 3a). Besides allowing for the study of delta topset evolution through time, the intersection of the delta profile with the varying ocean level provided us with the exact shoreline position $R_t(\theta)$ at each time step. Secondly, in order to investigate differences in transgression rates observed among our experiments, the actual evolution of shoreline position (R_t) was compared to the position that the shoreline would have had at any point in the experiment in the absence of erosion or deposition on the delta surface (R_t*). The latter is easily found as the intersection between the initial delta profile, assumed to be fixed, and the mean sea level at a given time step (FIGURE 3b). Any difference between R_t and R_t* reflects net deposition due to the action of fluvial and/or tidal processes. We therefore introduce a new metric, the transgression anomaly (TA), calculated as:

$$TA = (R_t^*/R_t) - 1 \quad (2)$$

Deltaic systems transgressing faster than expected have positive TA values (i.e., retrograding profile in FIGURE 3b), whereas negative values are observed for systems which transgress more slowly (i.e., prograding profile in FIGURE 3b). Using the same 89 profiles previously employed to derive shoreline position, the initial delta profile was recorded from the first run hour and all subsequent shoreline positions were used to calculate the *TA* at each run hour. The 89 individual *TA* values obtained for each run-hour were then averaged to give one individual *TA* value for each considered time step.

Finally, we propose a new approach to quantitatively compare upstream and downstream net deposition occurring along delta-topset longitudinal profiles, allowing us to distinguish between deltas evolving under varying external forcing. This method, named "Mean Profile Method" (MPM, FIGURE 4), allows for the synthetic characterization of deposition and erosion along the same topographic profiles previously employed to calculate shoreline position and transgression anomaly. For each delta, the difference $\Delta\eta = \eta_{t+1} - \eta_t$ between the topographic profile η at time t and $t+1$ is firstly computed, providing us with the absolute change in delta topset elevation. Then, the linear regression of $\Delta\eta$ provides a mean profile of the delta topset, $\bar{\eta}$, which is separated into an upstream and downstream portion based on the position of the zero-crossing point (i.e., the point where $\bar{\eta}$ changes sign; FIGURE 4). We stress that the characterization of upstream and downstream regions in this method is purely geometrical and does not imply any distinction in physical processes occurring across the delta. The distance from the sediment source to the zero-crossing point represents the upstream delta radius, r_u, while the remaining distance along the profile is designated as the downstream radius, r_d (FIGURE 4).The elevations of the first and last points along the mean profile provide the upstream (*a*) and downstream (*b*) depositional heights, respectively, and represent the maximum and minimum elevation values of $\bar{\eta}$. Finally, net depositional areas are calculated as:

$$\alpha = a \cdot r_u/2 \tag{2}$$

$$\beta = b \cdot r_d/2 \tag{3}$$

where α and β are the areas of upstream and downstream net deposition, and the signs of a and b indicate either positive or negative net deposition. When the zero-crossing point of $\bar{\eta}$ is located outside of the delta topset extent (i.e., when the sign of a and b is concordant), the mean profiles showed either deposition or erosion along their entirety, and net depositional areas read:

$$\alpha = a \cdot R/2 \tag{4}$$

$$\beta = b \cdot R/2 \tag{5}$$

where R is the radius of the delta topset.

As the initial delta geometry was set by deposition of sediment fed from the point source in the corner of the basin, each of the experimental deltas started with slightly different topset areas. Consequently, α and β were normalised by the delta's initial topset area to allow for a direct comparison between different experiments. The normalised net depositional area, α^* and β^*, were finally plotted in Cartesian (α^*,β^*) space, which can be divided into octants representing different net depositional behaviors of the delta topographic profile (FIGURE 5). For ease of identification, the octants followed a similar organization as Cartesian space with octant 1 immediately north of the x-axis in the first quadrant. Following a counterclockwise direction, octant 8 is located immediately south of the x-axis in the fourth quadrant. Mean topographic profiles located along the line of equality (i.e., the $\alpha^*= -\beta^*$ line) have equal, yet opposite, upstream and downstream net deposition, thus yielding a first fundamental distinction between erosional ($\alpha^* > -\beta^*$) and depositional ($\alpha^* < -\beta^*$) profiles (FIGURE 5a). Conversely, mean topographic profiles located along the line of uniformity ($\alpha^* = \beta^*$) are characterised by uniform net deposition across the entire profile, while the α^* and β^* axes capture profiles displaying null downstream and upstream net deposition, respectively (FIGURE 5a). Such a division of the (α^*, β^*) space allowed us to distinguish between 8 different modes of delta topset evolution (FIGURE 5b). Specifically, starting from octant 1 and moving clockwise we found (FIGURE 5b): purely-depositional, upstream-dominated profiles (1st octant); purely-depositional, downstream-dominated profiles (2nd octant); net-depositional, downstream-dominated profiles (3rd octant); net-erosional, upstream-dominated profiles (4th octant); purely-erosional, upstream-dominated profiles (5th octant); purely-erosional, downstream-dominated profiles (6th octant); net-erosional, downstream-dominated profiles (7th octant); and net-depositional, upstream-dominated profiles (8th octant).

Experimental Results

Experimental delta morphologies and evolution

The deltaic deposits shaped in our experiments displayed several distinctive features of tide-influenced deltas. As tidal amplitude increases, the delta-front became progressively more convex, exhibiting compound clinoform morphology (Rossi et al., 2016) (FIGURE 3). Increased tidal-to-fluvial energy ratios caused fluvio-deltaic distributary channels to be more elongated seaward as their lateral mobility was reduced, thus keeping them stable in the same delta-topset position for longer as compared to systems with less tidal influence (see also Lentsch et al., 2018, for a more detailed analysis). Smaller secondary channels were observed to diverge occasionally from the main distributary channels, distributing water and sediments over different portions of the delta topset. The periodic action of tidal currents in delta-topset areas not accommodating active fluvial distributary channels led to the development of tidal channels, growing from the shoreline toward the sediment source due to the progressive erosion of their headward extent driven by flow concentration during the ebb phase (FIGURE 6). When formed, we observed that these tidal channels promote sediment bypass through the delta plain by occasionally capturing water and sediment loads carried by secondary fluvial distributary channels (FIGURE 6h-l). Additionally, tidal channels displayed a strong reduction in their width moving landward, thus resembling the classic “funnel-shaped” planform typically observed in tidal environments (Ahnert, 1960; Lanzoni & D’Alpaos, 2015; Wright, Coleman, & Thom, 1973) (FIGURE 6g-l; see also Figure S1 for further details on the characteristics of active channels). Overall, the tidal-channel networks produced in the experiments exhibited a dendritic structure, with maximum stream orders of three or four according to Strahler’s ordering (Strahler, 1957). Larger tidal channels tended to be straighter, while smaller channels were more sinuous, as commonly observed in unvegetated tidal landscapes (FIGURE 6g-l). Both the persistence and the basinward extent of such tidal channels were dependent on the tidal-to-fluvial energy ratios, as well as on the RSL rise rate, with stronger tidal influence and reduced RSL rates promoting well developed and persistent tidal channel networks**.** Particularly, during the entirety of experiment DB2-1601 (FIGURE 6h), a tide-dominated delta experiencing a steady 0.5 mm/hr RSL rise, the main fluvial distributary channel was mostly pinned to the delta's northernmost boundary by the strong (10 mm amplitude) tides. Periodically a smaller secondary channel branched off from the main channel and attempted to fill in the accommodation being created throughout the delta. The presence of a mature tidal network on the remaining delta plain led to a unique interaction between the secondary fluvial channels and the tidal network. Distributary channels preferentially met the headward extent of the tidal channels and evacuate their sediment load down the channel with very little, yet evenly distributed, overbank deposition. The relationship between the fluvial and tidal systems in DB2-1601 led to the development of a composite delta system, with varying tidal and fluvial morphologies seen across the delta (FIGURE 6h). Indeed, the DB2-1601 delta can be qualitatively discretized into four regions based on the varying influences of tides and fluvial channels observed along the shoreline. The area where the main distributary channel was pinned during the whole experiment was classified as fluvial dominated. The immediate flanks of this channel, which experienced progressively increasing fluvial influence throughout the experiment, were classified as fluvial-influenced. Finally, regions exhibiting little to no fluvial influence, where the tidal network rarely interacted with distributary channels, were classified as tide-influenced and tide-dominated, respectively. Based on this classification the shoreline position, transgression anomaly and mean profile method data can be compared for each delta region and are discussed in the following sections.

Shoreline migration and transgression anomaly

After each run hour, mean shoreline position was calculated from the 89 topographic profiles acquired across the delta (FIGURE 7). For purely fluvial deltas the imposed sediment load sufficed to maintain an approximately constant shoreline radius of about 2 m (FIGURE 7a-c), with both the DB2-1602 (FIGURE 7b) and DB2-1605 (FIGURE 7c) shorelines even exhibiting slight shoreline transgression. Conversely, all the experiments involving tides displayed a regressive mean shoreline position, regardless of the relative dominance of tidal and fluvial forcing. Particularly, experiment DB2-1600 (FIGURE 7g), which had no fluvial input, exhibits the highest rate of transgression, followed by tide-dominated systems (FIGURE 7h-l) and lastly by tide-influenced systems (FIGURE 7d-f). Note that an error with the motorised weir controlling basin level occurred during experiment DB2-1606 (FIGURE 7l) at the 9th run hour, causing a drop in sea level. Subsequent erosion of the active distributary channel formed an embayment and the system transitioned to a tidal estuary. Although the issue was fixed at the 10th run hour, the fluvial system was well entrenched into the incisional valley feeding the embayment. This allowed for rapid marine transgression on the majority of the delta as the distributive channels could no longer deposit sediment to the now submerged portions of the delta plain, as readily identifiable in the shoreline data (FIGURE 7-l).

Transgression Anomaly (*TA*) data (FIGURE 8) show that only the purely tidal experiment (DB2-1600) is characterised by overall positive *TA* values, meaning that the shoreline transgressed more quickly than expected from the initial profile. When sediment is introduced, the systems show negative *TA* values on average, which indicates aggradation along the delta topset. This implies that even though the shorelines in all the experiments do ultimately transgress, they do so at a lower rate than systems with no sediment input. For any given class of deltas (i.e., purely tidal, tide-dominated, tide influenced and purely fluvial), the *TA* variation through time becomes more rapid as the rates of RSL rise increases (FIGURE 8). The systems subjected to high RSL rise rates of 2 mm/hr (DB2-1602, DB2-1603 and DB2-1606) also had higher sediment and water discharge rates, which produce a fluvial system with higher energy (Table 1). As fluvial energy in the system increases, the ability of tides to increase transgression rates decreases. Conversely, higher tidal energies (i.e., amplitude) lead to higher transgression rates over systems with comparable sediment budgets (FIGURE 8). Finally, it is worthwhile noting that even though all experiments were run for 20 hours, *TA* data for the highest RSL rise rates (FIGURE 8c) are truncated because TA could no longer be computed. This is because the initial delta profile would be completely flooded given the cumulative RSL rise at the point when *TA* data are truncated.

Mean Profile Method

The results obtained from the MPM were plotted in Cartesian (α^*, β^*) space together with the multivariate kernel density estimates (KDEs) of the data, which better represent the data distribution (FIGURE 9). The diagrams show a clear transition in the depositional response from systems subjected to low sediment load and slow RSL rise to more dynamic systems. With the exception of DB2-1600, which had no sediment input, the majority of topographic profiles plot in octants corresponding to positive net deposition (FIGURE 9). Using the purely fluvial systems as a baseline, the axes values show that both α^* and β^* values increase as the RSL rise rate increases (FIGURE 9a-c). Conversely, for a given RSL rise rate, the magnitude of upstream depositional area (α^*) does not appear to change significantly if different tidal amplitudes are considered, whereas the downstream depositional area (β^*) decreases as the tidal forcing becomes stronger. Therefore, the depositional character of the profiles is modified by tidal forcing, with deltas strongly influenced by tides exhibiting a higher density of overall erosional profiles (FIGURE 9g-l), which are much less frequent in both purely fluvial (FIGURE 9a-c) and river-dominated systems (FIGURE 9d-f). Particularly, the purely-tidal DB2-1600 system (FIGURE 9g), which was subjected to a steady 0.5 mm/hr RSL rise, mainly shows erosion-dominated profiles, although approximately 30% of the profiles display positive net deposition (octants from 8 to 3) half of which show deposition in the upstream portion of the delta (octants 1 and 8). Introducing fluvial sediment input while keeping the RSL rise rate constant at 0.5 mm/hr causes a marked transition from erosion- to deposition-dominated profiles (FIGURE 9a,d,h), with the relative frequency of downstream-dominated depositional profiles (octants 2 and 3) increasing as the fluvial-to-tidal energy ratio increases. Similarly, the relative importance of downstream deposition is enhanced in systems subjected to 1.0 mm/hr of RSL rise compared to similar systems with a lower RSL rise rate (FIGURE 9b,e,i). In fact, higher RSL rise rate produces an overall decrease in the relative frequency of profiles contained in the 8th octant (i.e., net depositional, upstream dominated) while intensifying the occurrence of downstream-dominated depositional profiles (octants 2 and 3) as well as increasing the magnitude of both α^* and β^*. This tendency is further enhanced by a RSL rise rate equal to 2.0 mm/hr (FIGURE 9c,f,l), for which most of the observed profiles plot in the first and second octants. The only exception is DB2-1606 (FIGURE 9l) for which the problem occurred with the mechanical weir controlling RSL.

Discussion

Experimental delta morphology and evolution

The delta morphologies obtained from the experiments showed clear intertidal zones as identified from channel extents (FIGURE 6; see also Figure S1). Many of the longitudinal profiles captured in DEM data, especially for tide-dominated experiments, show compound clinoforms (FIGURE 3b) (Rossi et al., 2016). Such features have been documented for major tidal delta systems such as the Ganges-Brahmaputra and Yangtze (Chen, Song, Wang, & Cai, 2000; Kuehl, Levy, Moore, & Allison, 1997), and are typically associated with tidal acceleration, which causes strong shear stresses on the inner shelf to form a region of limited deposition separating the subaerial and subaqueous clinoforms (Goodbred & Saito, 2012; John B. Swenson, Paola, Pratson, Voller, & Murray, 2005). To the best of our knowledge, compound clinoforms have not been previously identified in any experimental tidal deltas, here signifying another benchmark for the capability of experimental systems to reproduce tidal morphologies and evolution in a realistic manner.

In addition, tidal forcing exerted a strong control on the mobility of deltaic distributary channels, with tide-influenced and, above all, tide-dominated experimental deltas exhibiting a reduced number of distributary channels, much more planimetrically stable when compared to purely fluvial systems (Baumgardner, 2015; Lentsch et al., 2018; Rossi et al., 2016)(FIGURE 6; see also Figure S1). While the action of tides is known to promote both seaward elongation and stabilization of distributary channels by preventing mouth-bar growth at their inlets (Hoitink et al., 2017; Rossi et al., 2016), thus enhancing sediment bypass through the delta plain (Kästner, Hoitink, Vermeulen, Geertsema, & Ningsih, 2017; Olariu & Bhattacharya, 2006; Rossi et al., 2016; Shaw & Mohrig, 2014; Syvitski & Saito, 2007), it was not clear whether and how these mechanisms would operate where new accommodation is created by a rise in the RSL. We have shown in this study that the bypassing effect of tides on bedload sediment continues to operate during RSL rise, and leads to greater transgression than would be the case for a purely fluvial delta. A major question for field and numerical studies is whether adding fine sediments, which can be transported onshore by tidal action (e.g., Wilson et al., 2015), could counteract this effect. Our experiments also demonstrate the evolution of headward growing tidal channel networks (e.g., Coco et al., 2013; D'Alpaos, Lanzoni, Marani, Fagherazzi, & Rinaldo, 2005; Hughes et al., 2009; Knighton, Mills, & Woodroffe, 1991; Pethick, 1969; van Maanen, Coco, & Bryan, 2015) in deltas that aid in evenly distributing water and sediment load across the delta plain, thus reducing fluvial avulsion frequency and enhancing channel stability thanks to the reduced topographic gradient that is generated (Lentsch et al., 2018).

Finally, it is worth noting that a qualitative similarity exists between composite tide-influenced deltas obtained from our experiments (DB2-1600) and some field cases such as the Niger (Figure 1f) and Ganges-Brahmaputra (Figure 1g) deltas, where different portions of the delta plain can be distinguished based on the varying influences of tides and fluvial channels observed along the shoreline. However, most of tide-influenced deltas worldwide are characterized by dense vegetation cover (Figure 1) and low topographic gradients - ranging from 10^{-5} in large deltaic systems such as the Ganges-Brahmaputra-Meghna, (Wilson and Goodbred Jr., 2015) to 10^{-2} in small fan deltas (e.g., Piper et al., 1990) - arguably different from the unvegetated deltaic plains of our experiments where longitudinal slopes are in the order of 10^{-2} (FIGURE 6). Nonetheless, we note that slopes of delta-topsets observed in our experiments are consistent with values reported in the literature for experiments aiming at being representative of very low gradient tidal deltas consisting of fine sediment with high mobility (e.g., Ganti et al., 2016; Kleinhans et al., 2014; Martin et al., 2009; Wickert et al., 2013). Also, modern analogs of poorly vegetated delta plain examples can be found along the meso-macrotidal coast (tidal range up to 5 m) of the Iranian Hormozgan Province (Sanlaville, 2004), where the Mehran River and the Shur River debouch into the Strait of Hormuz forming two fan-shaped deltas (FIGURE 10) characterized by a mean topset slope of about $0.5 \cdot 10^{-4}$ (SRTM data). Vegetation growth is limited by the hypersaline environment, and the deltaic deposit mainly consist of fine, silty sediments. Continental sediment supply is poorly channelised, and sheet flooding surface runoff occurs during sporadic, yet intense rain events affecting the whole delta (Purser, 1985), all of which likely enhance the morphological similarity with our experimental composite deltas (FIGURE 10).

Shoreline migration and transgression anomaly

Monitoring shoreline migration through time provides suitable indicators of how deltas respond to RSL rise, and it also marks an important boundary between fluvial and submarine transport processes, which exert strong controls not only on the sedimentary structures but also on the preservation potential of stratigraphic sections

(J.B. Swenson, Voller, Paola, Parker, & Marr, 2000). The physical experiments in this study allow for forward modeling and multi-dimensional tracking of shoreline trajectory at a high temporal resolution. For purely fluvial experiments, sediment load rates calculated with equation (1) (Table 1) sufficed to either maintain a pseudo-static shoreline radius (DB2-1604) or prompt a prograding radius (DB2-1602 and DB2-1605, see FIGURE 7). In marked contrast, every experiment with active tides produced a shoreline transgression, regardless of the ratio of fluvial to tidal energy for each system (FIGURE 7). Even though these results might slightly vary depending on the calibration parameter we used to calculate the fluvial sediment transport rate needed to keep pace with RSL rise in equation (7), the only difference between tide-dominated and tide-influenced systems, with all other parameters equal, appears to be the rate at which the shoreline transgresses. In addition, when tidal forcing is present, highly dynamic systems (i.e., systems experiencing higher RSL rise rates and sediment discharge) appear to transgress at more variable rates: faster in some instances while comparable to less dynamic systems in others. When tides are absent, these same highly dynamic systems do not show transgression, implying that tidal processes act to remove significant amounts of bedload sediment from the delta plain. This also suggests, counter-intuitively, that tidal deltas are more sensitive to the rate of RSL rise than non-tidal deltas, despite having a sediment budget large enough to counteract the rise in RSL.

When the transgression anomaly data is plotted (Figure 9), the magnitude of the *TA* value represents how far apart the current shoreline and the static-surface shoreline are in space. Positive *TA* values displayed by the purely tidal delta (DB2-1600, FIGURE

8a) are likely justified by stronger flow concentration and higher velocities within channels occurring during the ebb phase, which causes the system to be ebb-dominated. Hence, net sediment export toward the sea causes the shoreline to transgress more quickly than predicted given the initial delta profile. Indeed, when fluvial sediment load is introduced upstream, deltas exhibit negative *TA* values on average, which indicates aggradation and/or progradation along the delta topset (FIGURE 8a). As the fluvial-to-tidal energy ratio increases, the ability of tides to increase transgression rates decreases (FIGURE 8). Conversely, higher tidal energies lead to higher transgression rates over systems with comparable sediment budgets.

The addition of tidal forcing complicates the morphologic response of a delta as fluvial processes of deposition are clearly modified, especially in the shoreline portion of the delta plain. The ebb-enhanced flow clearly dominates our experimental deltas where, irrespective of the relative strength of fluvial and tidal processes, sediments that would otherwise be deposited in the delta topset are instead transported to the delta foreset, bottomset or the deep ocean. While it is possible that tidal deltas that are vegetated or flood-dominant would respond to RSL rise differently, calculations of sediment budgets for several of the world's tidal deltas show large amounts of sediment transported to the subaqueous delta clinothems (Goodbred & Kuehl, 1999; Harris, Baker, Cole, & Short, 1993; Slingerland, Driscoll, Milliman, Miller, & Johnstone, 2008; Swanson et al., 2008). Early calculations for the sediment budget of the Fly delta could only account for half of the annual sediment load within the tide-dominated regions (Harris et al., 1993). It was later discovered that while some of the missing annual sediment budget is deposited on the Fly river's floodplain (Swanson et al., 2008), a large portion is appropriated to a composite clinothem forming the inner Gulf of Papua shelf (Slingerland et al., 2008). A similar study of the Ganges-Brahmaputra by Goodbred and Kuehl (1999) found that two-thirds of the annual sediment discharge is sequestered by the prograding subaqueous delta and the deep-sea Bengal fan. Studies of the major East Asian tide-dominated and tide-influenced deltas have shown that not all sediment removed from the subaerial delta plain is redeposited in the immediate alongshelf clinothem (Liu et al., 2009). In fact, 25-35% of the sediment load in these delta systems is transported between 500-800 km alongshore before final disposition in a shore-parallel middle-shelf clinothem (Liu et al., 2009). This complex picture of continental margin deposition demonstrates that several components may work in conjunction with tides to enhance sediment flux seawards.

Finally, it is worth noting that in field case studies, tidal channels downdrift of distributary channels with high suspended sediment concentrations have been shown to deliver fines upstream if tidal currents are flood-dominant, like in several of the north-south trending tidal channels west of the Ganges-Brahmaputra-Meghna river mouth in the Bay of Bengal (D K Barua, Kuehl, Miller, & Moore, 1994; Dilip K. Barua, 1990; Wilson & Goodbred Jr., 2015). The ebb-dominated character of our experiments resulted in the lack of a significant landward component of suspended sediment transport, though we propose that fine sediment import in the delta plain combined with the presence of vegetated intertidal areas, which promotes both organic and inorganic sediment deposition (D'Alpaos, Mudd, & Carniello, 2011; Goodbred & Saito, 2012), might enhance the capability of tide-influenced deltas to more efficiently keep pace with rising RSL.

Delta evolution and net deposition

Observing how the shoreline evolves through time gives a general sense of how the delta is responding and implies conditions of nearshore sediment delivery or downstream depositional processes. However, this does not explicitly capture net deposition metrics and gives no indication of the morphological adaption that takes place across the delta plain. Conversely, the mean profile method (MPM) is able to capture delta morphologic response in terms of net deposition. MPM data (FIGURE 9) show that delta evolution in these experiments is most strongly influenced by the rate at which RSL rises. While it is intuitive that higher rates of RSL rise increase the magnitude of both upstream and downstream depositional areas given a fixed tidal amplitude (FIGURE 9), it clearly emerges that a progressively increasing amount of sediment is deposited in the downstream portion of the delta topset (2^{nd} and 3^{rd} octants in FIGURE 9) in systems experiencing higher rates of RSL rise, even though sediment load in each experiment was designed to allow the delta to keep pace with RSL rise. This effect is mitigated by the action of tides, which enhance the relative magnitude of upstream depositional area (α^*) by preventing sediments from being deposited downstream, as clearly shown by the most frequent values of β^*, which decrease progressively as tidal amplitude increases for a given RSL rise rate (FIGURE 9).

These results are consistent with those obtained for shoreline migration and transgression anomaly, where tides enhanced the rate of shoreline transgression (FIGURE 7, FIGURE 8). The deltas that had *TA* values closer to zero show the smallest net deposition while the opposite is true for deltas that diverged the fastest and furthest from a zero *TA* value. While highly dynamic deltas seem to be more suited to counteract the effect of RSL rise, tidal influence in these systems ultimately allowed for a transgression to occur. Therefore, tides appear to reduce sediment trapping in the most downstream portion, by either increasing sediment bypass occurring within the delta plain or removing already deposited sediments, thus inhibiting the fluvial sediment load from outpacing accommodation created by RSL rise. This coincides with delta research of very energetic river mouths, such as the Amazon (Kuehl, DeMaster, & Nittrouer, 1986) and Fly (Harris et al., 1993), which show delta growth primarily in the form of a subaqueous mud clinoform leaving the subaerial delta susceptible to transgression. Both of these delta systems are subjected to tidal forcing and it is believed that tidal focusing inhibits sediment accumulation on the delta topset (Allison, 1998; Nittrouer, Kuehl, Demaster, & Kowsmann, 1986).

Nevertheless, research on the very dynamic Ganges-Brahmaputra delta has shown that, in contrast to most modern deltas, the subaerial delta front has prograded over the last 200 years in the face of Holocene sea-level rise (Akter et al., 2016; Allison, 1998). Even though we forced the experimental deltas to prograde in spite of the tide action by sufficiently raising the sediment supply, this draws attention to several factors present in that system as compared to these experimental deltas. Chief among them is the lack of vegetation on the delta top and absence of multiple grain sizes, particularly of fine cohesive sediments, in these experiments. Vegetation found in mangrove forests, and salt marshes in particular, interacts with fine grained clastics to promote the capture of suspended sediment, and help counteract the rise in RSL. The inclusion of these factors would likely modify the effect of tides observed in our experimental deltas. It is however interesting to note how some profiles of the purely-tidal DB2-1600 experiment are actually characterised by upstream deposition (2^{nd} and 3^{rd} octants in FIGURE 9g), showing that, in spite of the ebb-dominated character of the system, flood-phase tidal currents are capable of redistributing sediment along tidal channel apexes to prevent delta sinking. This observation indicates that, despite the lack of scaling and the absence of some determinants for the morphodynamic evolution, our experiments are able to overall correctly reproduce field-scale delta morphologies and evolution (**FIGURE 10**). Indeed, some of the tide-dominated experiments shown here do replicate the composite nature of the Ganges Delta and other tide-influenced deltaic systems worldwide (FIGURE 1 and FIGURE 10). For example, DB2-1606 (10 mm tides with 2 mm/hr RSL rise) is unique as it is the only experiment that indicates how a delta experiencing regression and valley-style erosion responds to subsequent transgression. The majority of expected profiles for the other delta systems fall in octants 1, 2 and 8 with less than 10% also in octant 3. However, as all of the fluvial discharge was trapped within the embayment, a large portion of the delta was influenced by tides only, which created profiles in octants 4 and 7. Accounting for roughly 16% of the total profiles, these are upstream dominant erosional and downstream depositional (octant 4) and upstream depositional and downstream dominant erosional (octant 7). Profiles from these portions of the delta are directly comparable to the majority of profiles found in DB2-1600 which was a delta system with only tides and no fluvial input. Therefore, different areas of DB2-1606 show parity to multiple delta types as different processes dominated in these sections.

Field studies of asymmetric wave-influenced deltas (e.g., Danube and Brazos deltas) have shown differences in sedimentary facies updrift and downdrift of the main channel outlet. Bhattacharya and Giosan (2003) believe this variation to be due to several modifying processes active in the downdrift area such as lagoonal, lacustrine, fluvial, tidal and vegetation-related sedimentation. More recently, large deltas have been considered a composite system if different portions are controlled by fluvial, tidal or wave processes which give distinctive morphologies across the delta (Goodbred & Saito, 2012). Although DB2-1606 is unique in that it was the only experiment that behaved as an estuary, it was not the only delta system to show varying topset morphologies. Particularly, the low rate of RSL rise allowed for the topset of DB2-1601 (FIGURE 6h) to maintain, throughout the entirety of the experiment, qualitatively distinct regions based on the varying influences of tides and fluvial channels (FIGURE 11a). Specifically, the area where the main distributary channel was pinned during the entirety of the experiment we classified as fluvial dominated, while the immediate flanks of this channel were classified as fluvial influenced. Regions where the tidal forcing appeared to outweigh fluvial influence were identified as tide influenced, whereas tide-dominated areas were characterized by the presence of tidal networks that rarely interacted with distributary channels. Based on this classification, we calculated shoreline position, transgression anomaly and mean profile method data for each delta portion (FIGURE 11b,c,d). While the selection of these areas was qualitatively carried out based on empirical observation, the shoreline position and transgression anomaly values reflect the results obtained for different experiments with varying tidal and fluvial energies. As fluvial power decreases, transgression of the shoreline increases and the *TA* moves closer to values of zero (FIGURE 11c). For the MPM, plotting the mean α^* and β^* shows the preferred profile for each section (FIGURE 11d). This preferential profile transitions from being dominantly upstream depositional (octant 1) to roughly uniform depositional ($\alpha^* = \beta^*$) as the relative strength of fluvial to tidal flows increases. These results clearly show that the influence of varying tidal-to-fluvial energy ratios observed for whole delta experiments is essentially replicated at a sub-delta scale, where the evolution of different delta-topset morphologies is driven by the dominant forcing in that region. Therefore, different morphological responses are to be expected in field-scale composite deltas, such as the Ganges-Brahmaputra-Meghna delta, based on the relative strength of the processes influencing a specific region of the delta.

Conclusions

The experiments presented here have shown that physical models of tidal environments can be extended to depositional systems such as deltas. These results expand an avenue experimental research that previously focused on systems such as estuaries and tidal channel networks. While several complexities found in the field such as multiple grain sizes, vegetation and flood dominant currents are not included, we believe these experiments provide a baseline for comparing the response between fluvial- and tide-dominated delta systems to eustatic variation. Furthermore, we have shown the transgression anomaly to be a useful method for comparing shoreline transgression between deltas experiencing different rates of RSL rise. The newly developed Mean Profile Method not only simplifies upstream and downstream deposition along delta topset longitudinal profiles, but also employs the normalised area change between subsequent profiles as a metric for net deposition. Based on our results generated from these new methods, we find that:

(1) Tidal deltas show transgressive shorelines (i.e., retrogradational shoreline evolution), regardless of tidal strength, under conditions in which fluvial deltas show shoreline stasis or regression (i.e., static or progradational shoreline evolution);

(2) The most dynamic fluvial-dominated delta system showed longitudinal profiles with up to two times the net deposition found in its tide-dominated and tide-influenced counterparts, pointing to sediment removal by tides;

(3) As tidal delta systems become more dynamic (i.e., subjected to higher sediment discharge and higher RSL rise rates), they transition from predominately upstream deposition and downstream erosion to up- and downstream deposition;

(4) Strong tidal forcing can immobilise distributary channels (see Lentsch et al., 2018), leading to composite delta morphologies as tidal and fluvial processes dominate different regions of the delta.

While all of the world's deltas are potentially vulnerable to sea-level fluctuations due to their low topographic gradients, we have shown that tides can possibly compound the local effect of sea-level rise as their ability to remove bedload sediment amplifies the mean eustatic trend. Moving forward, further consideration must be given to the effectiveness of sediment removal in delta systems affected by ebb-dominated tidal currents.

Notation

Ω_T	tidal energy per unit length of shoreline (W/m)
Ω_F	fluvial energy per unit length of shoreline (W/m)
ρ	water density (kg/m^3)
u	mean flow velocity (m/s)
τ_b	bottom shear stress (Pa)
C_f	drag coefficient (-)
h	characteristic flow-depth of the riverine input (m)
Q	water discharge (m^3/s)
Q_s	sediment discharge (m^3/s)
S	delta-topset slope (m/m)
S_T	delta-topset slope of the intertidal zone (m/m)
H_T	tidal amplitude (m)
T	tidal period (s)
L_s	shoreline length (m)
$\dot{H}_{SL}$	rate of sea-level change (mm/hr)
A_T	area of delta topset (mm^2)
f	fraction of sediment discharge trapped in delta topset
η_t	topographic profile at time step t
t	time step (hr)
$\Delta\eta$	difference in successive topographic profiles
$\bar{\eta}$	linear regression of $\Delta\eta$

r_u upstream radius (mm)

r_d downstream radius (mm)

a upstream depositional height (mm)

b downstream depositional height (mm)

α area of upstream net deposition (mm^2)

β area of downstream net deposition (mm^2)

α_* α normalised by initial topset area (-)

β_* β normalised by initial topset area (-)

R radius of the delta topset (mm)

R_t actual shoreline position (mm)

R_t^* projected shoreline position (mm)

TA transgression anomaly (-)

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Time-lapse videos of all the experiments described in text can be found at the following link: https://conservancy.umn.edu/handle/11299/195978

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Figure captions and tables

FIGURE 1 Landsat 8 natural colour composite images (bands 6,5,4) of tide-influenced deltas worldwide. The scale bars each represent a length of 20 km

FIGURE 2 Experimental setup. (a) Overhead image of the Delta Basin-2 (DB-2) facility during experiment DB2-1604. The overhead image is superimposed on the corresponding hillshaded DEM data. The added equipment schematics are not drawn to scale. (b) An image of the data cart that serves DB-2. The cart consists of a high-resolution digital line scan camera capable of generating digital elevation models (DEM) with submillimeter resolution. Several overlapping passes were made over the deposit with the final DEM generated from stitching the passes together

FIGURE 3 (a) View of the DB2-1601 experiment tidal delta showing the locations of every 10th degree longitudinal profile. Profiles were taken every degree from θ=1° to θ=89° and captured delta-topset evolution. (b) An example profile taken from 45° showing the raw topographic data. Features that are identifiable include the shelf break as well as a compound clinoform. Using the current ocean level, the exact shoreline position was found and tracked throughout the experiment. A synthetic sketch defining the transgression anomaly (TA) methodology is also shown. The projected shoreline position, R_t*, is calculated as the intersection of the current mean ocean level and the initial delta plain topographic profile. Conversely, the actual shoreline position, R_t, is derived from the intersection between the current mean ocean level and the actual delta topography. TA is negative when the actual shoreline radius is larger than the projected radius, and positive conversely

FIGURE 4 Schematic representation of the Mean Profile Method (MPM) employed to characterise the net depositional behavior of experimental deltas

FIGURE 5 Geometrical interpretation of the Mean Profile Method in terms of net upstream (α*) and downstream (β*) depositional areas. Mean profiles located along the line of equality ($|\alpha^*| = -|\beta^*|$) have equal, yet opposite, upstream and downstream net deposition, while profiles along the line of uniformity ($\alpha^* = -\beta^*$) display uniform net deposition across the entire profile

FIGURE 6 Overhead images of the analyzed experiments superimposed on the corresponding hillshaded DEM data. All images refer to the midpoint of the experiments, for which values of the mean delta-topset slope ($\bar{S}$) are also reported

FIGURE 7 Shoreline position through time as a function of RSL rise rate and tidal amplitude. The shoreline radius, measured as the distance between the point source and the delta-topset intersection with the current ocean level, was computed for 89 radial topographic profiles taken every run hour. Mean shoreline position is also plotted to show the general shoreline migration trend. Different symbols represent varying tidal-to-fluvial energy ratio, computed following the approach suggested by Baumgardner (2015)

FIGURE 8 Transgression anomaly (*TA*) vs. run time based on different relative sea-level (RSL) rise rate: a) RSL rise rate 0.5 mm/hr; b) RSL rise rate 1.0 mm/hr; c) RSL rise rate 2.0 mm/hr. Different symbols denote varying fluvial-to-tidal energy ratios, while different colours stand for distinct tidal amplitude

FIGURE 9 Scatter plots of normalised upstream (α*) and downstream (β*) net deposition, obtained by the Mean Profile Method, are plotted together with the 2D kernel density estimates (KDEs) of the data (obtained by considering a Gaussian kernel and bandwidths equal to 0.4 x10^{-4} α* and β*). Note that exaggeration in panel "g" is ten times larger than the all the other panels. A diagram depicting what each octant represents in terms of net deposition profiles is also reported

FIGURE 10 Qualitative comparison of experimental and field case tide-influenced deltas. a) experiment DB2-1604; b) Mehran river delta, Hormozgan province, Iran (26.816808° N, 55.519997 E; ©Google Earth); c) Shur river delta, Hormozgan province, Iran (27.037227° N, 55.844624° E; ©Google Earth)

FIGURE 11 Composite delta data for DB2-1601. (a) The delta was separated into four regions based on the observed influence of fluvial vs. tidal processes. (b) Mean shoreline position through time for each region. (c) Transgression anomaly (*TA*) through time for each region. (d) Scatter plot of mean α_* and β_* values for each delta region to indicate the preferred net depositional profile

Table 1: Experimental parameters

Experiment	Sea-level Rise Rate (mm/hr)	Water Discharge (m^3/s)	Sediment Discharge (m^3/s)	Tide Amplitude (mm)	Tide Period (sec)	Tidal Power (W/m)[*]	Fluvial Power (W/m)[*]
DB2-1600	0.5	0	0	10	120	4.09e-2	0
DB2-1601	0.5	5.0e-5	5.0e-7	10.0	120	3.27e-2	8.18e-3
DB2-1701	0.5	5.0e-5	5.0e-7	3.5	120	1.87e-3	6.13e-3
DB2-1604	0.5	5.0e-5	5.0e-7	N/A[†]	N/A[†]	N/A[†]	6.54e-3
DB2-1607	1.0	1.0e-4	1.0e-6	10.0	120	3.85e-2	1.39e-2
DB2-1608	1.0	1.0e-4	1.0e-6	3.5	120	2.34e-3	9.83e-3
DB2-1605	1.0	1.0e-4	1.0e-6	N/A[†]	N/A[†]	N/A[†]	1.14e-2
DB2-1606	2.0	2.0e-4	2.0e-6	10.0	120	4.36e-2	2.45e-2
DB2-1603	2.0	2.0e-4	2.0e-6	3.5	120	2.16e-3	2.13e-2
DB2-1602	2.0	2.0e-4	2.0e-6	N/A[†]	N/A[†]	N/A[†]	2.13e-2

[*]Calculated with the energy-based tidal power metrics and stress-based fluvial power metrics of Lentsch et al. (2018), modified from Baumgardner (2015).

[†]N/A = not applicable; these experiments did not include tides.