CONTROL OF BASIN WATER DEPTH ON CHANNEL MORPHOLOGY AND AUTOGENIC TIMESCALES IN DELTAIC SYSTEMS

BRANDEE CARLSON,1,2 ANASTASIA PILIOURAS,1,3 TETSUJI MUTO,4 AND WONSUCK KIM1
1Department of Geological Sciences, Jackson School of Geosciences, University of Texas at Austin, Austin, Texas, U.S.A.
2Department of Earth, Environmental and Planetary Sciences, Rice University, MS-126, 6100 Main Street, Houston, Texas 77005, U.S.A.
3Earth and Environmental Sciences Division, Los Alamos National Laboratory, Los Alamos, New Mexico, U.S.A.
4Department of Environmental Science, Nagasaki University, Nagasaki, Japan
e-mail: bc15@rice.edu

ABSTRACT: River channel geometry is often controlled by upstream boundary conditions, including fluvial discharge and sediment properties. At the coast, downstream boundary conditions (e.g., tides, waves, water depth) also strongly influence channel formation and evolution. We conducted a set of experiments to determine the effects of basin water depth (i.e., a downstream boundary condition) on the evolution and geometry of fluviodeltaic channels and lobes. Internal dynamics (autogenic processes) in the fluviodeltaic system drive channel avulsion through cycles of sediment storage and release. Experimental results indicate an increase in the timescale of autogenic storage and release with increasing basin water depth. Deeper basin water requires a larger volume to be filled within the delta front, thus more time to complete one autogenic storage and release cycle for a given sediment discharge. While a relationship between delta-front volume and autogenic storage and release timescales is expected, we show that autogenically generated morphological changes in the delta topset and distributary channels also exert control on timescales of storage and release. Deltas building into deeper basins develop steeper topsets, and deeper distributary channels that cause high-magnitude topset slope fluctuations, which contribute to the long autogenic timescales. Deposits in shallow basins exhibit both shallower topset slopes and shallower channels. Channel bed slopes are similar (~0.06) across all experiments, but lateral channel migration rates varied with basin depth. Deltas building into shallow basins had rapid lateral channel migration, such that channels quickly reworked the delta topset. For deep basins, channel migration rates were much slower, so the topset was reworked less often, allowing the topset to build to steeper slopes before being reworked. These experiments indicate an intimate relationship between lateral channel migration and topset aggradation. In addition, the deeper and more stable channels in deeper basins generally developed a wider range of channel widths, some of which produced elongate lobes. We conclude that the downstream boundary has a strong control on fluviodeltaic morphology, which can result in a striking difference in the autogenic timescale.

INTRODUCTION

River deltas are home to a large part of the world’s population, but they are constantly at risk of land loss from sea-level rise and subsidence (Syvitski et al. 2005; Paola et al. 2011). Delta growth is influenced by a suite of internal processes and external controls that affect delta dynamics, morphology, and stratigraphy, as well as the ability to build land (Edmonds and Slingerland 2007, 2009; Hoyal and Sheets 2009; Kim et al. 2009; Geleynse et al. 2011; Powell et al. 2012; Caldwell and Edmonds 2014). The classical ternary diagram for dominant controls on delta morphology describes the importance of fluvial and coastal processes (e.g., waves and tides) in shaping the delta surface and subsurface (Galloway 1975). Recent studies have also highlighted the importance of several other boundary conditions, such as grain size and sediment discharge, that control delta and fan growth (Clarke et al. 2008; Van Dijk et al. 2009, 2012; Kleinhans et al. 2010; Powell et al. 2012; Caldwell and Edmonds 2014). Sea level is generally considered as a moving downstream boundary condition (transgression-regression) (Pirimz et al. 1998; Steel et al. 2003; Burgess et al. 2008; Patruno et al. 2015). Basin water depth—set by sea level—controls clinoform stacking patterns (Uliçny et al. 2002), and we therefore expect that basin water depth influences other characteristics of delta morphology. However, our knowledge of the effects of downstream boundary conditions on fluviodeltaic morphology and dynamics is limited (Uliçny et al. 2002; Edmonds et al. 2011; Sassi et al. 2012; Rossi et al. 2016).

The goal of this study is to determine how basin water depth (i.e., accommodation)—as a downstream boundary condition—influences delta channel and lobe geometry and dynamics using a series of physical delta experiments. Under static sea-level conditions, water depth increases (from ~10 m to ~150 m) with distance from the coast due to the slope of the continental shelf. This study explores the morphological response of deltoid distributary channels and lobes in response to basin depth changes, thus providing insight to the spatio-temporal dynamics of deltas prograding across the shelf.

Even under conditions of no or steady forcing, deltas exhibit temporally varying dynamics due to autogenic or internal processes (Muto and Steel 2004; Kim et al. 2006; Muto et al. 2007; Kim and Jerolmack 2008; Van...
Dijk et al. 2009; Powell et al. 2012). Autogenic processes contribute to changes in both modern landscapes and their stratigraphic products (Kim et al. 2006; Muto and Steel 2004). It is therefore important to understand autogenic processes so that paleo-environmental changes recorded in strata, such as those from tectonics or climate change, can be more accurately distinguished from autogenic signatures.

Previous delta-building experiments show that the autogenic cycle of fan deltas consists of channel incision, lateral channel migration, backfilling (in-channel deposition, which propagates from the delta margin toward the apex due to a decrease of channel slope associated with progradation), followed by sheet-flow flooding over the delta topset, and finally initiation of the next channel incision (Kim and Jerolmack 2008). Each component of this cycle exhibits a control on the topset slope, as sediment is either preferentially stored on the topset or delivered to the shoreline, depending on the phase of the cycle. The slope increases during a sediment storage phase during backfilling and sheet flow and decreases in a sediment release phase—coincident with channel initiation—such that shoreline progradation is not constant through time but instead pulses with each release event (Kim and Jerolmack 2008). The time to complete a single autogenic cycle (up to 10^3 years in natural systems (Kim and Jerolmack 2008)), hereafter called the autogenic timescale, can be related to the upstream boundary conditions, as the inputs of water and sediment influence the rates of channel incision and migration (Van Dijk et al. 2009; Powell et al. 2012). The autogenic storage and release process operates on modern fan-delta systems, such as the fan that builds into Emerald Lake in British Columbia (Goehart and Smith 1998; Kim and Jerolmack 2008), the Kosi River fan in Northern India (Reitz et al. 2010), and in twelve fluvial fans near Sydney, Australia (Scott and Erskine 1994). Stratigraphic evidence of autogenic storage and release cycles is proposed to exist in Gilbert-type delta deposits from the Pliocene–Holocene Crati Basin and the Pliocene Loreto Basin, where deposits interpreted as release events produce packages that extend into their receiving basin 400 m and 2 km respectively (Kim and Jerolmack 2008).

Due to the long timescale over which autogenic storage and release events operate (~ 10^3 years), controlling parameters have primarily been explored through physical experiments and numerical modeling (Kim et al. 2006; Clarke et al. 2008; Kim and Jerolmack 2008; Van Dijk et al. 2009; Powell et al. 2012, and others). The effects of varying water and sediment discharges on fluvio-deltaic channel dynamics have been explored by several previous studies (e.g., Van Dijk et al. 2009; Powell et al. 2012). Van Dijk et al. (2009) found that increasing water discharge in their experiments increased sediment transport rates and therefore sped up erosion and channel backfilling cycles associated with the autogenic timescale. In contrast, Powell et al. (2012) observed that high sediment discharge also caused a morphological feedback that increased the magnitude of the fluvial slope change, which caused the autogenic timescale to increase. Increasing water discharge also increased the autogenic timescale due to higher water discharge, resulting in a more organized fluvial system with fewer but stronger, more persistent channels (Powell et al. 2012). The discrepancy in these two studies might be explainable by the fact that Van Dijk et al. (2009) used a sediment concentration that was an order of magnitude smaller than that of Powell et al. (2012) (ratio of sediment discharge to water discharge of 0.001 in Van Dijk et al. 2009 vs. 0.01 in Powell et al. 2012), which would lead to large differences in the topset slope (Parker et al. 1998; Whipple et al. 1998). However, exact causes for this opposite trend have not yet been addressed.

Downstream boundary effects on dynamics of deltas and alluvial fans have also been explored experimentally (Clarke et al. 2008; Kleinhans et al. 2010; Van Dijk et al. 2012). Experiments conducted by Clarke et al. (2008) investigated the effects of alluvial fan toe-cutting on alluvial-fan evolution. Deposition at a delta or mouths of alluvial fan channels typically induces in-channel deposition and backfilling that drives cycles of channelization and sheet flow. However, the presence of a ditch at the fan toe acted to enhance channelization and incision on the fan surface, as sediment was transported past the toe into the larger accommodation, thus limiting in-channel deposition (Clarke et al. 2008). Building on this idea, Van Dijk et al. (2012) conducted experiments to compare alluvial fans to fan deltas, which differ by the absence or presence, respectively, of standing water at the downstream boundary. They determined that the presence of ponded water downstream for fan deltas led to greater channel incision depth, larger slope fluctuations, and more pronounced channelization-backfilling cycles on fan deltas compared to alluvial fans. This was due to the in-channel deposition promoted by deposition at the channel mouth in standing water that drives channel backfilling (Van Dijk et al. 2012).

Results from experiments by Kim et al. (2013) were similar to that of Clarke et al. (2008). Kim et al. (2013) modeled the behavior of a shelf-edge delta by including a sudden slope break to simulate a transition from a shelf to deep water. They found that once the delta reached the shelf edge, a bypass channel was created that routed all sediment from the upstream source past the delta toe to create a subaqueous deep-water fan downstream. The bypass channel was very stable, experiencing little to no lateral migration and effectively serving as a point source of sediment for the subaqueous deep-water fan (Kim et al. 2013). This result was corroborated by Muto et al. (2016), who found that the channels on a delta prograding into very deep water inevitably became graded and were stabilized within a fluvial valley in the axial delta plain. We therefore hypothesize that delta experiments conducted in a deeper basin will show more persistent channelization with less lateral channel migration as the delta works to fill the deeper water at the delta front.

Here we present the results from a series of delta experiments conducted with a range of basin water depths between 1 cm and 10 cm. The objective was to isolate the effects of basin water depth on channel morphology and the autogenic cycle by keeping upstream boundary conditions constant across all experiments and over time. We compare channel morphology (i.e., depth, width, and slope) between experiments and discuss the size and shape of delta lobes that reflect the channel dynamics and timescales under different basin water depths. We further analyze lateral channel migration and show that signatures of basin water-depth control are recorded during delta lobe building. The basin water depths used here range from the same order of magnitude as the channel depth (1 cm) to one order of magnitude higher than the channel depth (10 cm). We could consider the Wax Lake delta as an example of a system with channel depths that are the same order of magnitude as the water depth in the receiving basin (~ 2.5 m and 3 m, respectively). The Mississippi River delta could be considered a deep-water analogue, where distributary channel depths at the Head of Passes are 15–30 m deep and the receiving basin is 200 m deep.

**METHODS**

**Experimental Setup**

We utilized the Sediment Transport and Earth-surface Processes (STEP) basin and the Experimental Delta Dynamics (EDDy) basin at the University of Texas at Austin. The experiments had basin water depths of 1 cm, 3 cm, 5 cm, 7 cm, and 10 cm, and are hereafter referred to as JKF-1, JKF-3, JKF-5, JKF-7, and JKF-10, respectively. The channel depth in the experiments ranged from 0.1 cm to 1 cm. We therefore chose the basin depth to cover an order of magnitude range in channel depth (i.e., 1–10 cm). We used an inner part of the STEP basin that is 3 m long, 2 m wide, and 0.5 m deep. The EDDy basin is 2.5 m long, 2 m wide, and 0.5 m deep. In STEP, a pump connected to a constant-head tank fed water into the basin, and a dry-screw auger feeder was used to input sediment. A weir was used to maintain base level at a constant elevation. The EDDy basin
recirculates water through a constant-head tank to maintain a constant base level, and a dry screw-auger feeds sediment to the basin.

We kept supply rates of water and sediment supply constant throughout each experiment (Table 1). We controlled feed rates of water and sediment in each tank such that the ratio of water to sediment discharge was constant at 65/1, and the sediment supply rate was 1 cm$^3$/s for all experiments. Sediment grains were a bimodal mix of 50% white 170 μm silica sand and 50% brown 1–2 mm silica sand.

Sediment and water were mixed outside of the basin and fed through a tube into a rock cage emplaced at the corner of two walls arranged to form a 90-degree opening to the down-basin direction. The STEP basin walls were acrylic plastic, and the EDDy basin walls were concrete blocks. Sediment was deposited on a flat, non-erodible, and non-subsiding basement and built a quarter-circle-shaped deposit in plan view. We ran each experiment until the average shoreline position reached ~1 m downstream. The duration of each experiment was therefore variable, depending on the depth of the basin.

### Data Collection

We collected subaerial topography data at one-hour runtime intervals throughout the STEP basin experiments. We utilized a laser altimeter to measure the elevation of the sediment surface with a sub-millimeter vertical resolution and a horizontal resolution of 5 mm in the $x$ (down basin) direction and 5 mm in the $y$ (cross basin) direction. We regularly monitored base level with a sonar transducer. Topographic data are not available for the EDDy basin experiments.

From the topography data, we measured the width and depth of each channel along two transects at 1/3 and 2/3 the distance from the delta apex to the shoreline. For each topography scan, we also measured the average channel slope and the average topset slope. Slopes were measured manually by selecting several points along either the channel or topset in the down-basin direction between the delta apex and the shoreline, calculating the slope between each pair of points, and then averaging the local slopes for each transect. We report the average topset and channel slope for each scan as the mean slope of all measured transects for a given scan.

We collected overhead images every 30 seconds in each experiment, and we sampled images every 5 minutes to digitize the shoreline. We corrected distortion in the images by the camera angle and lens curvature using standard methods (Tal et al. 2012). We approximated the average shoreline position over time as the radius of a quarter-circle area equivalent to that of the delta topset area. For JKF-1 and JKF-10, we also measured plan-view positions of active shoreline segments, i.e., the mouths of active distributary channels. The location of an active shoreline can be quantified through an image analysis of the azimuth angles $\theta$ (0°–90°) at both ends of the active segment that are measured from the initial source point, such that $\theta = 45°$ for the delta centerline and $\theta = 0°$ or $\theta = 90°$ for the shoreline at the basin sidewalls. Pairs of measured values of $\theta$ were plotted against time $t$ to track the locations of active shoreline segments through time, and each segment was expressed as a line spanning the range of $\theta$ for each individual channel. We also identified major channels by the following criteria: (1) the channel must have a clear concentration of flow, identified by the blue dye, that continuously extends from the source point to the shoreline; (2) it must convey water and sediment to the shoreline, resulting in recognizable sediment accretion at the shoreline in successive images; and (3) it may bifurcate, but only the main trunk channel(s) will be included as the major channel, while any secondary channels will be counted as minor. Using time-lapse imagery, we isolated individual delta lobe-building events, where lobes are defined as the basin-ward accumulation of sediment at the mouth of an active channel. Delta lobe-building events were considered to begin with initial channel incision and shoreline progradation and were considered complete when the channel back-filled and migrated away from the initial location. By tracing the shoreline immediately before and after the completion of a lobe-building event, we were able to characterize lobe area resulting from a single sediment release event. We also documented the planform geometry of the lobes in terms of a ratio of cross-basin width to down-basin length.

### EXPERIMENTAL RESULTS

#### Shoreline Progradation

Shoreline migration rate varied between the experiments. All runs had the same upstream conditions of water and sediment inputs, but the basin depth unique to each experiment controlled the overall difference in progradation rate. In each experiment, the shoreline progradation rate decreased with time, as the delta grew and the area of the sloping deltaic surface increased. As one may intuit, experiments with deeper basin water depth had slower shoreline migration compared to experiments with shallower basin water depth (Fig. 1). Figure 1 also includes best-fit power-law curves for the average shoreline positions $s$ over time. Shoreline position increases as a power-law function of time. A delta deposit subject to constant sediment supply and approximated as the radius of a semicircle has an expected exponent of 0.5 (Muto and Swenson 2005; Kim et al. 2006; Kim and Muto 2007). For the present experiments, the exponents of the best-fit curves are within a range (0.35–0.55) that is consistent with previously reported delta experiments (Kim and Jerolmack 2008; Powell et al. 2006).

#### Table 1. — Experimental parameters assigned for each experimental run, where the only parameter that varies between experiments is basin water depth.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>$\Omega_s$ [cm$^3$/s]</th>
<th>$\Omega_w$ [cm$^3$/s]</th>
<th>Basin Water Depth [cm]</th>
<th>Delta Open Angle [degree]</th>
</tr>
</thead>
<tbody>
<tr>
<td>JKF-1</td>
<td>1</td>
<td>65</td>
<td>1</td>
<td>90</td>
</tr>
<tr>
<td>JKF-3</td>
<td>1</td>
<td>65</td>
<td>3</td>
<td>90</td>
</tr>
<tr>
<td>JKF-5</td>
<td>1</td>
<td>65</td>
<td>5</td>
<td>90</td>
</tr>
<tr>
<td>JKF-7</td>
<td>1</td>
<td>65</td>
<td>7</td>
<td>90</td>
</tr>
<tr>
<td>JKF-10</td>
<td>1</td>
<td>65</td>
<td>10</td>
<td>90</td>
</tr>
</tbody>
</table>

#### Table 2. — Autogenic storage and release timescales as determined from time-lapse images and perturbations from shoreline progradation curves.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Time [min]</th>
</tr>
</thead>
<tbody>
<tr>
<td>JKF-1</td>
<td>64</td>
</tr>
<tr>
<td>JKF-3</td>
<td>53</td>
</tr>
<tr>
<td>JKF-5</td>
<td>173</td>
</tr>
<tr>
<td>JKF-7</td>
<td>165</td>
</tr>
<tr>
<td>JKF-10</td>
<td>1200</td>
</tr>
</tbody>
</table>

#### Table 3. — Experimental parameters used to calculate sediment transport during channelized periods of each run.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\rho_s$</td>
<td>2650 [kg/m$^3$]</td>
</tr>
<tr>
<td>$\rho_w$</td>
<td>1000 [kg/m$^3$]</td>
</tr>
<tr>
<td>$g$</td>
<td>9.81 [m/s$^2$]</td>
</tr>
<tr>
<td>$D$</td>
<td>0.001 [m]</td>
</tr>
<tr>
<td>$Q_r$</td>
<td>$65 \times 10^{-5}$</td>
</tr>
<tr>
<td>$S$</td>
<td>0.06</td>
</tr>
</tbody>
</table>

al. 2012). If delta growth occurred at a constant rate, experiments would not exhibit departures from the predicted exponential curves. However, cyclic autogenic processes cause periods of decreased progradation rates and periods of accelerated progradation rates, as a function of autogenic storage and release cycles.

Experiments with shallower basin water experienced more frequent autogenic storage and release cycles. Storage occurs as sheet flow over the fan delta surface induces sediment deposition on the topset, which increases the topset slope until the deposit geometry is prone to a channelization instability (Kim and Jerolmack 2008). Small channels laterally migrate across the deposit until flow collapses into a dominant channel. Periods of release occur as channelized flow incises the topset, degrading the fan surface to a lower slope. Channels deliver sediment to the shoreline, building discreet delta lobe deposits. As the lobe builds into the receiving basin, channel slope concomitantly decreases, thus decreasing transport capacity and sediment transport to the shoreline and inducing in-channel sedimentation. Sediment deposition begins at the shoreline and propagates upstream, and this behavior is termed “backfilling,” and acts both to anneal the channel and to restore the topset to a higher slope. Once the channels are backfilled and the topset surface is regraded, sheet flow resumes and the cycle repeats in a periodic fashion to produce cyclic foreset accretions. Periods of storage and release can be observed in Figure 1, where departures from power-law shoreline growth correlate to autogenic storage and release cycles. For all basin depths, the deviations from the radially averaged long-term growth of the delta indicate either accelerated shoreline migration, which often occurs as

![Figure 1](https://pubs.geoscienceworld.org/sepm/jsedres/article-pdf/88/9/1026/4335219/1527-1404-88-9-1026.pdf)

**FIG. 1.**—Radially averaged shoreline position against time. Autogenic sediment release events as times of increased progradation rates are indicated by arrows. The best-fit curve exponents range from 0.35 to 0.55, where a lower exponent is consistent with rapid shoreline progradation into shallow water while larger exponents are consistent with slow shoreline progradation into deep water.
the result of rapid sediment delivery to the shoreline during channelization, or decelerated shoreline migration, which is associated with reduced delivery of sediment to the shoreline because of sediment storage on the deltaic topset by aggradation during sheet flow and/or channel backfilling (Kim and Jerolmack 2008).

The autogenic time scale ($T_a$) is defined as the duration beginning with a strong channelization until the channel backfills and relocates itself on the topset. The timing of autogenic events was determined in two ways: (1) manual identification of departures of shoreline growth from power-law growth (Fig. 1) and (2) signal-processing methods to determine dominant frequencies in the areal growth time series, described in more detail below. In both cases, the timing of autogenic events was corroborated through examination of overhead time-lapse images for each experiment (Table 2).

To determine the frequency of autogenic events, the time series of delta area was detrended using a linear fit. The residuals are plotted in Figure 2A, where fluctuations between periods of sediment storage and sediment release are indicated by oscillations. Periodograms were calculated to identify the dominant frequencies of storage and release oscillations (Fig. 2B).

Dominant frequencies were converted to cycle duration and plotted as a function of basin depth (Fig. 3). The first dominant frequency is approximately the length of each experimental run (black circles 256–1280 min), while the second frequency (gray asterisk) is similar in duration (within 9–35%) to the manually identified autogenic events in time-lapse images and shoreline growth curves (gray boxes 80–213 min). The discrepancies between the manually identified events and the cycle duration determined through signal processing may arise through averaging event durations, image processing, and/or human error of visually identifying the completion of a storage-and-release cycle.

JKF-10 experienced only one complete autogenic storage and release events during the experiment (>1200 min). Therefore, the autogenic timescale aligns with the first dominant frequency (Fig. 3). Overall, the results indicate that the autogenic storage-and-release timescale increases with basin water depth, from ~100 min for JKF-1 to >1200 min for JKF-10 (Fig. 3). We note that the averaged shoreline in JKF-10 did not show significant variations from the power-law growth trend (i.e., shoreline acceleration and deceleration events) over the course of the experiment; rather, during the first half of the experiment, the measured shoreline position is slightly above the predicted shoreline position, while the second half of the experiment exhibits a shoreline position that is slightly less than predicted. These trends are most clearly observed in the detrended data of delta planform area (Fig. 2A), where one prominent storage period and one prominent release period is clearly identifiable.

**Lobe Development**

Each experiment cycled between periods of smooth and irregular shorelines. Figure 4 shows an example of the shoreline planform pattern change in JKF-5. Prominent channels generally caused local progradation and created an uneven shoreline due to the creation of delta lobes (i.e., sediment deposited at the mouth of a channel (Figs. 4A, D). This period was followed by lateral channel migration and avulsion (Fig. 4B), typically including several small channels on the topset that more broadly redistributed the sediment to smooth the shoreline. Compared to lobe-building events, which were characterized by fewer channels making discrete lobes, periods with many small channels displayed slower shoreline growth that occurred over a more widespread area. Channelization resumed after the period of lateral migration and avulsion ceased. Avulsion ceases when one flow path becomes dominant, thus maintaining a higher discharge and sediment transport capacity relative to the smaller, dispersed channels. The cycle of topset aggradation and channelization is repeated for the duration of each experiment.

We measured individual lobe areas using time-lapse images, as shown in Figure 4, but no strong trend of the lobe area associated with basin water depth was observed (Fig. 5A). We also determined characteristic lobe filling times, $T_f$ (Fig. 5B), by dividing lobe volume by sediment supply rate. This calculation assumes that there is no sediment transport past the delta toe and that all sediment input during a given time contributed to lobe building, an assumption that is reasonable given visual inspection suggesting that sediment loss past the delta toe is minimal and uncertainty is therefore low. Since there were no subaqueous topographic measurements, we multiplied basin water depth by measured lobe area to approximate the lobe deposit volume. The lobe deposit volume was corrected by an estimated porosity for the sediment mixture used, which was the same for all experiments (porosity ~ 40%). Figure 5B shows that increasing basin water depth generally increases the total time to complete building a lobe, a trend that is supported by the duration of observed lobe-building events in the experiments.

**Active Shorelines**

Figure 6 shows the locations of active shorelines through time for JKF-1 and JKF-10, which are expressed along with the progressive spatial expansion of the delta plain. We first mapped the active shoreline locations as radial distances from the delta centerline ($\theta = 45^\circ$ since the delta has a 90$^\circ$ opening angle). We then converted $\theta$ to distance ($L_w$) to plot the arcuate distance of an active shoreline segment measured from the axis of the delta plain ($\theta = 45^\circ$) along the periphery of a circle that has a radius of $s$. Shoreline segments fed by a major channel (black lines, Fig. 6) are distinguished from those fed by a minor channel (gray lines, Fig. 6) using the criteria outlined in the methods section.

There is a set of features common to both runs: (1) nearly the entire shoreline was active in the early stages of delta growth, reflecting a well-developed sheet flow covering a considerable part of the delta plain ($s < 250 \text{ mm for } t < 20 \text{ min in JKF-1}; s < 250 \text{ mm for } t < 50 \text{ min in JKF-10}$); (2) following this sheet-flow stage, multiple discrete channels appeared at the shoreline ($s = 250–450 \text{ mm for } t = 20–50 \text{ min in JKF-1}; s = 250–700 \text{ mm for } t = 50–300 \text{ min in JKF-10}$); (3) the number of coexisting distributary channels tended to progressively decrease over time, such that a smaller fraction of the entire shoreline was active over time ($s = 450 \text{ mm for } t = 50–150 \text{ min in JKF-1}; s = 450–700 \text{ mm for } t = 300–650 \text{ min in JKF-10}$); and (4) thereafter, delta growth was dominated by only a single major active channel or, occasionally, few minor distributary channels ($s > 450 \text{ mm for } t > 150 \text{ min in JKF-1}; s > 700 \text{ mm for } t > 650 \text{ min in JKF-10}$).

The temporal density of active shorelines fed by major channels is much higher in JKF-10 (deep water) than in JKF-1 (shallow water), particularly in the late part of the run. However, channels more frequently returned to the same locations in the shallow-water experiment (JKF-1) compared to the deep-water experiment (JKF-10), indicating faster channel migration in shallow water than in deep water. In shallow water (JKF-1), active shoreline locations appear to be sparse in time, which reflects rapid lateral shifting of active shoreline locations and short residence times of channels. Particularly after $t = 42 \text{ min (} s > 450 \text{ mm}$), major channels were only rarely developed and short-lived, while most of the sediment delivered to the shoreline was conveyed by minor channels. The overall basin-wide lateral migration rate of active shoreline segments in JKF-1 was $0.672/\text{min}$ for $s = 684–893 \text{ cm (} t = 126–260 \text{ min})$. The plots of JKF-10 (deep water) show a pattern different from those of JKF-1. Major channel(s) on the delta plain existed during nearly the entire experiment. The dense temporal distribution of major channels indicates their slow migration rates. The overall basin-wide migration rate of active shoreline segments in JKF-10, which were always fed by major channels, was $0.345/\text{min}$ for $s = 674–811 \text{ cm (} t = 626–887 \text{ min})$, i.e., much lower than the lateral migration rate in JKF-1.
Fig. 2.—A) Detrended delta area through time. Fluctuations represent autogenic storage and release events. B) Periodograms of detrended area. The first frequency peak represents experimental run time. The second frequency peak corresponds to autogenic release events identified both in the time-lapse imagery as well as the departures from power-law growth in the shoreline position curves (Fig. 1).
Channel Morphology

Channel depths were measured from topography data (1 cm, 5 cm, 10 cm) and widths were measured from overhead images (all runs) during periods of strong channelization, which we defined as periods when channels were persistent and actively delivering sediment to the shoreline. Figure 7 shows the channel depths and widths measured at 1/3 and 2/3 the distance from the delta apex to the shoreline, where the smaller circles denote the upstream measurement and the larger circles denote the downstream measurement. The range of channel depths increased with increasing basin depth, where JKF-1 channels were all \(< 10\) mm deep and JKF-10 channel depths ranged from 2 mm to 25 mm. Channels in JKF-10 (deep water) also displayed a wider range of widths (20–330 mm) compared to JKF-1 (shallow water; 25–90 mm). In shallower basins, experiments more frequently cycle through periods of backfilling, and during these times, multiple small channels are present on the topset. However, for JKF-10, only one period of backfilling is demonstrated, so the width measurements are procured from an individual channel that has variable width in time, but is generally wide. Thus, channel-width measurements for shallower basins are representative of both strong, individual channels and multiple smaller channels associated with backfilling, while measurements from the deeper experiment exclusively demonstrate a single channel width.

Topset and Channel Slopes

Figure 8 shows the topset (black) and channel (gray) slopes over time for each STEP basin run. The circle marker represents the mean of all measured channel or topset transects for a given topography scan, and the error bars represent one standard deviation. In all experiments, the average topset slope was greater than the average active channel slope. This
indicates that the topset surface was generally steeper than the active channels, as backfilled abandoned channels and sheet-flow aggradation during the storage phases mostly built the topset surface. Autogenic storage-and-release cycles involve the repeated transition between storage phases that build up the topset slope and release phases with incised channels that remove sediment from the topset and deliver it to the shoreline (Kim and Jerolmack 2008). Note that here we collected topography data every hour, so slope measurements are not necessarily correlated with storage-and-release events, but rather represent a range of existing topset and channel slopes throughout autogenic cycles.

Comparing the slope data across experiments with different basin water depths, we can see that the active channel slopes were generally the same.

**Fig. 5.**—A) Planform area of deposition created during each lobe-building event. B) Lobe-building timescale determined from calculated lobe volume and sediment feed rate.

**Fig. 6.**—Trajectory diagrams for A) JKF-1 and B) JKF-10 showing the locations of the active shoreline segments with growing delta radius (left axis) and time (right axis). Black lines indicate the presence of a major channel, as defined in the methods section, while gray lines indicate the presence of a minor channel. Major channelization is present more often throughout the deepwater experiment (JKF-10) compared to the shallow-water experiment (JKF-1). However, channels more frequently return to the same location in the shallow-water experiment, indicating faster channel migration.
Fig. 7.—A) Channel depths and B) width measured at upper and lower transects. Measurements were collected from topography scans collected during channelized flow (i.e., sediment release event).

Fig. 8.—Topset (black) and channel (gray) slopes over time for A) JKF-1, B) JKF-5, and C) JKF-10 measured from topography scans. Dots represent the average slope measured along 3 to 5 downstream transects in each scan and error bars represent one standard deviation.
The range of topset slopes, however, was considerably larger in the experiments with deeper water, often having much steeper topset slopes compared to the experiments in shallower basins. This results in occasionally very large differences between the topset and channel slopes in JKF-10, compared to JKF-1 and JKF-5 where topset and channel slopes are closer in magnitude. Even in JKF-10, however, there were some periods where topset and channel slopes were nearly identical (e.g., hours 14–15), especially during periods of weak channelization. This may also indicate that the topographic data were collected near the end of an autogenic cycle, such that the channel was already backfilling and rebuilding its slope to that of the steeper topset surface.

**DISCUSSION**

Paleo-sea-level records are often produced from paleoshorelines (Patruno et al. 2015). However, delta shorelines exhibit cyclic temporal changes even under conditions of static sea level (Kim et al. 2006). Here we show that basin depth is a control on distributary-channel geometry, topset slope, and the timescales for sediment delivery to the coast. These results suggest that as deltas migrate across the shelf from shallow to deeper water, delta morphology and morphodynamics should predictably respond to changes in basin water depth—a critical tool for stratigraphers.

**Control of Basin Depth on Channel Morphology, Transport Capacity, and Topset Slope**

Deposits building into deeper water were generally able to develop deeper channel depths and wider channel widths relative to those in shallower water (Fig. 7). This result is generated through a complex relationship between deposit geometry (i.e., channel bed slope) dictated by upstream boundary conditions and morphological response to downstream boundary conditions (e.g., channel mobility and topset geometry).

Experiments across all basin depths displayed a similar channel slope (Fig. 8), indicating that channels incised into the topset surface until they reached the equilibrium slope associated with upstream boundary conditions (i.e., water discharge and sediment supply). In this circumstance, the higher background topset slope and/or increased aggradation of the topset in deeper basins allows the development of relatively deeper channels. Here, a question emerges: How does deeper basin depth cause a steeper topset slope and/or more topset aggradation?

Deeper basins cause channels to migrate more slowly, as it takes longer to fill the space at the delta-front (Fig. 6). Nonchannelized regions on the topset receive sediment intermittently and in small quantities via sheet flow that cascades over the unchannelized topset from the delta apex. In theory, then, these areas should be continuously increasing in slope until the region either receives more flow with higher transport capacity (e.g., sheet flow with higher discharge or channelization) or the topset reaches a slope that is conducive to sediment bypass. The higher topset slopes therefore reflect the gradual aggradation of nonchannelized surfaces. Because the channel migrates slowly in deep basins (Fig. 6), there is more time for the topset to aggrade in deep basins.

Topset aggradation is strongly favored over progradation for non-channelized parts of the topset, and this produced different styles of shoreline progradation for different basin depths. Figure 9 shows images of the shoreline at the same averaged downstream position. Though their radially averaged locations are the same, their planform patterns are significantly different. JKF-10 is featured at runtime = 12 hr, in which the delta prograded asymmetrically and a narrow lobe elongated to the river left-hand side. The figure shows the channel on the lobe just before avulsing to river right-hand side, which made a much shorter path to the shoreline along a steep topset slope, which had a long duration of aggradation. In contrast, JKF-1 had more laterally mobile channels and developed a smoother shoreline that was more radially symmetric compared to JKF-10.
more radially symmetric compared to JKF-10. When the channels in JKF-1 avulsed or migrated to new locations, they made new paths but with lengths similar to the existing channel paths due to the more radially symmetric pattern. These channels maintained small widths as they quickly reached the characteristic channel bed slope and rapidly migrated across the topset surface. In deeper basins, the channels spanned a broader range of widths and depths for the longer duration of channel incision into the steeper delta topset.

For the channelized regions of the deposit, the broader range of channel depths for deeper basins (Fig. 7A) indicates a broader range of boundary shear stresses (τ₉) in channels, where τₑ is calculated as

\[ \tauₑ = \frac{pgHS}{B^2} \]  

(1)

where \( \rho \) is the fluid density (1000 kg/m³), \( g \) is acceleration due to gravity (9.81 m/s²), and \( H \) is channel depth. Assuming normal flow, \( S \) is the average channel bed slope. \( S \) was ~ 0.06 for JKF-1, JKF-5, and JKF-10. For JKF-1, \( H \) ranges from 3 to 7 mm, while \( H \) in JKF-10 ranges from 2 to 25 mm. This yields ranges for boundary shear stress, where \( \tauₑ \) ranges from 1.77 to 4.12 kg/ms² for JKF-1 while \( \tauₑ \) for JKF-10 ranges from 1.18 to 14.715 kg/ms². Boundary shear stress is directly related to the transport capacity of a channel to move sediment, where higher boundary shear stress results in higher sediment transport rates. The Engelund and Hansen (1967) prediction for total sediment load shows this relation:

\[ qₜ = \sqrt{\frac{RgDD}{c_f}} \left( \frac{\tauₑ}{\rho RgD} \right)^{2.5} \]  

(2)

where \( qₜ \) is the sediment flux per unit channel width, \( R \) is the submerged specific gravity (here 1.65 for quartz), \( D \) is the median grain diameter, \( \rho \) is water density, and \( c_f \) is the friction factor. Here we explore the role of channel geometry by casting the Engelund and Hansen (1967) equation in terms of channel depth, \( H \), and channel width \( B \) by substituting Equation 1 for \( \tauₑ \). Then we calculate the friction factor as

\[ c_f = c_z \left( \frac{B}{H} \right)^{2} \]  

(3)

where \( c_z \) is the Chezy coefficient, which is calculated as

\[ c_z = \left( \frac{U}{\sqrt{g}} \right)^2 \]  

(4)

and \( U \), water velocity, is calculated as

\[ U = \frac{Q}{BH} \]  

(5)

where \( Q \), water discharge, is a fixed experimental parameter, \( B \) is channel width, and \( H \) is channel depth.

By assigning experimental parameters (Table 3), we calculate a minimum and maximum transport capacity where median grain size, \( D \), ranges from 170 × 10⁻⁶ to 0.001 m, fluid and sediment densities are 1000 kg/m³ and 2650 kg/m³ respectively, water discharge is 65 × 10⁻⁶ m³/s, acceleration due to gravity is 9.81 m/s², and slope is 0.06, we arrive at

\[ qₜ_{min} = 3.64e^{-10} \frac{B^2H^{3/5}}{F^{1/5}} \quad \text{and} \quad qₜ_{max} = 4.28e^{-8} \frac{B^2H^{3/5}}{F^{1/5}} \]  

(6)

where the maximum transport capacity is calculated with \( D = 1 \) mm and the maximum transport capacity is calculated using \( D = 0.170 \) mm. Both \( B \) and \( H \) were measured throughout the duration of the experiments, and we use the cumulative channel width along each transect, located at 1/3 and 2/3 of the delta length, to account for the possible existence of multiple channels. The resulting sediment flux for JKF-1 (shallow water) ranges from 5.4 × 10⁻⁷ to 3.95 × 10⁻⁴ m³/s and for JKF-10 (deep water) ranges from 2.1 × 10⁻⁸ to 1.3 × 10⁻⁴ m³/s, where wider and deeper channels are predicted to have lower \( qₜ \). The Engelund and Hansen (1967) equation predicts that \( qₜ \) in JKF-10 can be an order of magnitude lower than JKF-1, and this is due to lower flow velocities predicted by Equation 5 for channels that are both wider and deeper compared to shallower and narrower. We compare these predicted sediment fluxes to sediment fluxes observed during strong channelization events using overhead images to track areal shoreline differences as described in Methods and displayed in Figure 4. We calculate the fluxes to range from 3.9 × 10⁻⁶ to 8.3 × 10⁻⁵ m³/s, which aligns with the maximum values of predicted sediment flux. We predict that channels in deeper basins, which can be both deeper and wider than channels in shallower basins, ultimately have lower transport capacity, thus increasing the time required to remove sediment from the topset during autogenic release events, contributing to an increase in \( Tₛ \).

Deeper basins also display a larger range of topset slopes, though channel slope was similar across experiments (Fig. 8). Deeper basins are able to produce steeper topsets because channels laterally migrate very slowly so the topset is not reworked for long periods of time. This produced greater slope fluctuations between the maximum topset slope and the channel slope for deeper basin depth (JKF-10) compared to the shallower basin depth (JKF-1). However, because topsets build to steeper slopes in deeper basins, more sediment must be removed during lateral migration. This introduces an intimate linkage between topset slope and channel migration rates, which is difficult to decompose with the present dataset. Nonetheless, our experimental topography data expressly show greater topset slope fluctuations in the deep basin. Therefore, during a release event, more sediment must be liberated from the topset in the deeper basin relative to the shallower basin.

Thus, deltas in shallower basins have both channels with higher shear stress and smaller topset slope fluctuations, both of which influence the autogenic timescale. Considering the difference between the average topset slopes and the average channel slope, the area of sediment encompassed by the topset slope fluctuations, \( A_T \), is calculated as the area of a triangle, where \( A_T = 0.02 m^2 \) and 0.035 m² for JKF-1 (shallow water) and JKF-10 (deep water) respectively, when the distance from the delta apex to the shoreline is 1 m. The time required for the topset to build up to its maximum and minimum slope, \( t_{slope} \), is calculated as

\[ t_{slope} = \frac{A_T}{qₜ} \]  

(7)

where \( t_{slope} \) for JKF-1 ranges from 0.01 to 10 hours and JKF-10 ranges from 0.5 to 460 hours.

In summary, deposits building in deeper basins built steeper topsets with deeper channels. The steeper topset slopes are linked to the slow lateral migration of channels in deeper basins, where topset slopes continuously aggrade by sheet flow over the topset. The topset angle can continuously build until a channel relocates to a previously unchannelized area, the transport capacity of overland flow increases, or the topset reaches a slope that is conducive to sediment bypass. Channels for experiments in all basin water depths reached a characteristic channel bed slope of ~ 0.06. For deeper basins with steeper topset slopes, the fluctuation between topset and channel bed is greater than that for shallower basins, thus producing relatively deeper channels. Because channels in the deep basin cut to deep depths and display a wider range of channel widths, the channels have larger cross-sectional areas than channels on deltas in shallow water. This therefore leads to decreased velocities in the deep-water experiments (Equation 5), such that the transport capacity of channels in deeper basins was less than that of channels in shallower basins, so the sediment evacuation efficiency from the topset is greater in shallower basins.

Thus far, we have shown that channel dimensions, delta topset geometry, and resulting sediment transport conditions respond to static basin depth as a downstream boundary control. For deposits building into a deep basin,
width lobes, and smaller values correspond to elongate lobes. Lobe aspect direction and width in the radial direction. Higher values correspond to wide lobes and smaller values correspond to elongate lobes. A value of 1 would correspond to lobes of equal length in the down-basin direction and width in the radial direction.

The control of basin depth on lateral migration rates appears to be recorded in lobe-building events as deltas in shallower basins had (1) more numerous deposits for a given amount of time, reflected in the shorter lobe-building timescale and (2) higher $\Omega$ values that have a tighter distribution of geometries.

Scaling Autogenic Storage and Release to Natural Deltas

Alternation between channelization and sheet flow, with concomitant pulses of shoreline progradation, are observed in natural fan-delta systems, such as the Emerald Lake fan delta in British Columbia (Goedhart and Smith 1998). These autogenic storage-and-release cycles may be common to all natural, noncohesive fan deltas. In fact, fan deltas in the rock record show cyclic foreset progradation, which may be the result of autogenic storage-and-release events (Kim and Jerolmack 2008).

The autogenic storage-and-release processes observed in our experiments are comparable to those in natural systems. We recorded fluctuations in topset slope for storage and release events that range from $\sim 20$ to $40\%$ of the mean topset slope, which is within the slope variability of natural rivers presented in Kim et al. (2006). If we consider a depositional fluvial system with a length of 100 km and a mean slope of $1 \times 10^{-4}$ and a release event with a slope fluctuation of $\sim 30\%$, the resulting accumulation would then release about $3 \times 10^{5}$ m$^2$ of sediment per unit of basin width to the shoreline. The extent to which the deposit would extend into the basin would depend on basin water depth. For example, a basin with a mean water depth of 10 m would result in a deposit that would extend 30 km during the release event. However, we emphasize that while the autogenic storage and release occurs in concert with lobe and channel switching and may occur at similar timescales, the autogenic storage-and-release mechanism described here is a laterally averaged process and therefore distinct from a single avulsion event.

CONCLUSIONS

Deltas build strata that record key information about allogenic forcings, such as climate change and tectonics. Embedded in these signals is the complicated record of autogenic cycles, which are generated by internal dynamics. By decomposing stratigraphy into signatures produced as autogenic versus allogenic responses, we can elucidate important events in Earth history, such as shifts in paleoclimate. Autogenic processes respond to boundary conditions, where variations such as grain size, sediment supply, and water discharge may act to increase or decrease autogenic timescales.

Here we show that basin water depth—as a downstream boundary condition—provides control on autogenic timescales. For deeper basins, more volume must be filled at the delta front for delta progradation to occur, and one may intuit that this alone leads to increased duration of the autogenic timescale. Increased autogenic timescales with increasing basin water depth is indeed observed in our experiments. However, we show that basin-filling volume alone does not account for the longer duration of the autogenic timescale. Deeper basin water depth promotes a morphological response of both channels and the delta topset that
culminates in increased autogenic timescales. Channel width and depth tend to increase with increasing basin depth. This results from a complex relationship between lateral channel mobility and topset aggradation. Prolonged aggradation of the topset occurs for deep basins due to low lateral migration rates of channels, and this allows the topset slope to become steeper compared to topset slopes of deposits building into shallower basins. Both the increased channel volume and the greater topset slope fluctuations result in increased autogenic storage and release timescales for deeper basins.

We expect that the autogenic storage-and-release process is manifested in natural systems on $10^3$ year timescales (Kim and Jerolmack 2008). In the rock record, deposits with signatures of deep basin water depth (clinoforms with large foreset relief) should display a broader range channel dimensions, exhibit signatures of slow lateral migration, and consistency in channel location (few avulsions), compared to deposits building into shallow basin water depths.

Separating autogenic from allocogenic signatures requires knowledge of landscape dynamics under different suites of boundary conditions. By thoroughly exploring the role of these controlling parameters for autogenic timescales, we will more accurately interpret the signals of external perturbations recorded in stratigraphy.

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REFERENCES


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