Exploring Sediment Compaction in Experimental Deltas: Towards a Meso-Scale Understanding of Coastal Subsidence Patterns

Samuel Mason Zapp

University of Arkansas, Fayetteville

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Exploring Sediment Compaction in Experimental Deltas: Towards a Meso-Scale Understanding of Coastal Subsidence Patterns

A thesis submitted in partial fulfillment of the requirements for the degree of Master of Science in Geology

by

Samuel Mason Zapp
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John B. Shaw, Ph.D.
Thesis Director

Ralph Davis, Ph.D.
Committee Member

Michelle Bernhardt-Barry, Ph.D.
Committee Member
Abstract

Subsidence in low elevation coastal areas has been extensively researched through direct field measurement, numerical modelling, and stratigraphic reconstruction of ancient sediment deposits. Here I present the first investigation of subsidence due to sediment compaction and consolidation in two laboratory scale river delta experiments. Compactional subsidence rates have never been thoroughly quantified in the experimental setting, though this mechanism is found to be a primary creator of total relative sea level rise which will likely cause coastlines to retreat in the coming years. Spatial and temporal trends in subsidence rates in the experimental setting may elucidate behavior which cannot be directly observed at sufficiently long timescales, except for in a reduced scale model such as the ones studied. I compare subsidence between a control experiment using typical boundary conditions of standard laboratory fan-deltas with an experiment which has been treated with a proxy for highly compressible organic rich marsh or peat deposits. Both experiments have non-negligible compactional subsidence rates across the delta-top which are comparable to our boundary condition relative sea level rise of 250 μm/h. Subsidence in the control experiment, on average 54 μm/h across the low elevation areas of the subaerial delta, is concentrated in very low-elevation (<5mm above sea level) areas near the coast and is likely due to creep induced by a rising water table near the shoreface. The marsh experiment exhibits larger (on average 126 μm/h) and more spatially variable subsidence rates which are controlled mostly by compaction of recent marsh deposits at or very near the sediment surface. These rates compare favorably with field and modelling based subsidence measurements when plotted in dimensionless space. By scaling these results to the field, we find that subsidence “hot spots” may be relatively ephemeral through longer timescales, but average subsidence across the entire low elevation region of a delta can be variable at century and millennial timescales. Subsidence rates in a given decade or century may exceed thresholds for marsh platform drowning, even in the absence of anthropogenic impacts.
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1. Introduction

River deltas are home to several hundred million people, as well as important agricultural, environmental, and energy resources. With global coastlines expected to retreat due to climatically driven relative sea level rise in the coming decades, deltas are becoming increasingly vulnerable to land loss, and simultaneously valuable as a mechanism of land building (Ericson et al., 2006). Wetlands on and around coastal deltas are particularly susceptible to being drowned by relative sea level rise because they grow in very low elevation areas. These threatened wetland environments provide valuable ecosystem services including storm surge protection, carbon sequestration, and water quality regulation (Engle, 2011). In South Louisiana alone, over 5,000 km² of coastal salt marshes have been lost since 1930 (Couvillion et al., 2017), largely due to high, spatiotemporally variable subsidence rates from the compaction of highly compressible marsh deposits in the shallow subsurface (Törnqvist et al., 2008). On the modern Upper Gulf Coast, subsidence within the top several meters of soft sediment deposits can dominate background relative sea level rise (eustatic sea level rise plus tectonic regional subsidence) by as much as a factor of three, at least on short timescales (Jankowski et al., 2017).

Our understanding of the co-evolution between delta morphodynamics, marsh growth, and resulting subsidence remains relatively limited. This incomplete understanding can be attributed to our inability to observe the processes of delta evolution (i.e. aggradation, channel incision, avulsion) and marsh platform growth over sufficiently long timescales (>1000 years), as well as difficulty isolating the various forcing conditions that drive morphodynamic change and mechanisms that contribute to coastal subsidence (Hoyal and Sheets, 2009; Yuill et al., 2009).

Predicted and measured subsidence rates on the Mississippi Delta (as well as on other similar deltas) are both highly spatially variable and operate on multiple timescales. While subsidence rates over annual to decadal timescales can exceed a centimeter per year at a given location (Nienhuis et al., 2017), there is disagreement about the upper bound of annual
subsidence rates and whether or not subsidence maxima correlate with geologic controls such as fine grained Holocene deposit thickness (Jankowski et al., 2017; Byrnes et al., 2019). Millennial scale subsidence rates are even harder to predict but consistently estimated to be significantly lower (Kooi and de Vries, 1998; Meckel et al., 2006; van Asselen, 2011). It remains unclear how long areas of high subsidence can persist on century and longer timescales due to a lack of direct measurement going back further than about 15 years. Additionally, the overprinting of several possible subsidence mechanisms including sediment compaction, faulting, anthropogenic soil drainage and deep fluid withdrawal make it difficult to understand which processes are driving the complexity of observed subsidence rates (Dokka, 2006; Yuill et al., 2009; Chang et al., 2014). Understanding the degree to which natural processes and human activity each impact different subsidence mechanisms is crucial to proposed land loss mitigation plans such as sediment diversions and wetland restoration, which are planned on century timescales.

This study describes internal subsidence behavior of two laboratory scale delta experiments, one treated with a proxy for salt marsh deposits (TDWB-19-2) and one untreated control experiment (TDB-18), in order to better understand how the coupling between deltas and marshes impacts the spatiotemporal variability of subsidence rates throughout delta evolution. Similar reduced-scale experiments have often proven highly effective at creating analogous kinematics and spatial architecture to autogenic behavior observed in field deltas (Paola et al., 2009). They have been particularly useful in understanding “meso-scale” delta evolution that cannot be fully captured by continuous field measurement during active morphodynamic changes, nor reconstructed by stratigraphic interpretation (Paola et al., 2009). For example, the processes of backwater-controlled channel avulsion (Edmonds et al., 2009; Hoyal and Sheets, 2009) and shoreline autoretreat (Muto, 2001) were validated by experimental observations. This study is part of a larger project which aims to assess the impact of salt marshes on a wide range of deltaic processes, from delta-top kinematics to stratigraphic patterns.
No studies have previously described autogenic subsidence due to compaction (hereafter, just subsidence) in a delta experiment. Basin water level is often raised to impose an allogenic relative sea level rise (RSLR\textsubscript{b}), which is the sum of eustatic sea level rise plus spatially uniform subsidence akin to deep-seated lithospheric flexure (Paola, 2000; Paola et al., 2001; Straub et al., 2015). RSLR\textsubscript{b} is assumed to create all accommodation space and allow the delta to aggrade in place, but dynamic subsidence internal to the sediment deposits (\(\sigma\)) has either been ignored or written off as negligible. Here, \(\sigma\) is measured in quiescent, abandoned areas of the delta, and subsidence under active channel belts is not directly considered, though it is likely somewhat larger than \(\sigma\). The sum of \(\sigma\) and RSLR\textsubscript{b} yields total relative sea level rise (RSL). As such, any descriptions of subsidence patterns in the control experiment are novel in and of themselves, and potentially useful to understand autogenic controls on accommodation space in experiments.

Differences in the magnitudes, geometry, and temporal signals in subsidence rates between the control and marsh proxy experiment can be generally attributed to the consolidation of proxy which has similar geotechnical properties (such as porosity and coefficient of consolidation) to highly compressible organic rich marsh deposits, and has a greater potential for compaction than the fluvially transported sediment. These patterns could themselves resemble medium-long term natural subsidence behavior of marsh-rich deltas, such as the Mississippi.

2. Methods

2.1 Experimental Setup

This study analyzes data collected from two laboratory delta experiments which are best described as a partially filled swimming pool being slowly filled with sediment pumped into the system with a hose. The sediment self organizes into a delta under constant forcing conditions such as sea level rise rate (RSLR\textsubscript{b}), basin geometry (2.5 m across), and water (\(Q_w\)) and sediment flux (\(Q_s\)) listed in Table 1. Each experiment was allowed to aggrade for 120 hours
before hour zero for the system to reach a dynamic equilibrium with an average subaerial delta radius of approximately 1.1 m. The experiments have identical boundary conditions except that a proxy for marsh sediments was deposited every two hours of run time in the marsh treatment experiment (TDWB-19-2) while the control (TDB-18) received only fluvially transported sediment. Therefore, significant statistical differences in delta top kinematics, subsidence rates, or stratigraphic patterns can be attributed to the impact of the marsh proxy deposits. From here onwards, TDWB-19-2 will be referred to as the “marsh experiment,” and TDB-18 will be referred to as the “control experiment”.

**Table 1.**

<table>
<thead>
<tr>
<th>Delta Experiment</th>
<th>Experiment Run Time (h)</th>
<th>Qw (gpm)</th>
<th>Qs (kg/h)</th>
<th>RSLRb (mm/h)</th>
<th>Sediment Mixture</th>
</tr>
</thead>
<tbody>
<tr>
<td>Control (TDB-18)</td>
<td>560</td>
<td>2.7</td>
<td>1.40</td>
<td>.25</td>
<td>Strongly cohesive mixture (Straub et al., 2015)</td>
</tr>
<tr>
<td>Marsh (TDWB-19-2)</td>
<td>560</td>
<td>2.7</td>
<td>1.40</td>
<td>.25</td>
<td>Strongly cohesive mixture (Straub et al., 2015), EPK kaolinite marsh proxy</td>
</tr>
</tbody>
</table>

Boundary conditions of both experiments.
Figure 1. (A) An image of the marsh experiment during a marsh distribution cycle. The metal apparatus is a sieve mounted to a low-frequency vibrator which shakes out marsh proxy at queried locations. The darker brown sediment is exposed marsh deposits. (B) Marsh deposits as a fraction of total sediment mass entering the basin for each distribution cycle. Actual marsh deposits generally follow the trend of “ideal” deposit mass based on the number of locations within the marsh window but exhibit scatter around the trend.

The marsh proxy was distributed in all locations between -9 mm below sea level and 5 mm above sea level (hereafter referred to as the “marsh window”) in order to fit a model of salt marsh platform growth (Morris et al., 2002), and to allow the fraction of the delta covered by marsh to roughly approximate marsh coverage in large field deltas. Initial marsh deposit thicknesses from -9 to -5 mm and 0 to 5 mm were set to fill roughly 1.8 times RSLRb, and deposits in the -5 to 0 mm elevation range were set to fill only 0.9 times RSLRb. Total drowning of the marsh platform would occur if $\sigma_s$ were to exceed RSLRb by a factor of 1.8 over a sufficiently long timescale. The proxy was distributed in dry powdered form through a sieve (pictured in Figure 1(A)) and allowed to settle onto the sediment surface in all areas where mean elevation satisfied these rules every two hours. The mass fraction of marsh sedimentation to total sedimentation, or “marsh mass input fraction” for every distribution cycle is plotted in Figure 1(B). The red line represents the ideal mass fraction of marsh based on our elevation model, while the green line shows the actual mass fraction deposited. The actual trend
generally follows the ideal trend, but there is significant scatter (approaching 50% error) due to mechanical imprecision associated with the distribution cart.

2.2 Geotechnical Properties

The cohesive “clastic” sediment mixture delivered by river transport to both experiments ranges in grain size from coarse sand to clay and contains a polymer to increase sediment cohesion. It has been described in Straub et al., (2015). Cores taken from the control experiment have porosities of roughly 50% averaged across the thickness of the entire stratigraphic package. This sediment was not expected to compact significantly, so porosity was assumed to be relatively unchanged as a function of burial depth. The marsh proxy is EPK kaolinite. Its porosity is estimated to be roughly 90% when first deposited. This gives it significantly more compaction potential than the clastic cohesive sediment.

Experimental sediments have relatively similar grain sizes and initial porosities to surficial soil samples from the field, but experience far less overburden loading (less than 1kPa) than field samples which can nearly exhaust their primary consolidation potential within Holocene strata (Keucher, 1994; Keogh, 2020). Additionally, organic rich field samples experience a very high degree of secondary compression due to the collapse of peat particles (Mesri et al., 1997). Our marsh proxy is entirely mineral sediment. These factors represent potential issues for scaling compaction between laboratory and field scale.

A one-dimensional controlled rate of strain (CRS) consolidation test detailed in ASTM D4186 was performed on both the clastic and marsh sediments to assess their vertical strain response (i.e. consolidation) to an imposed load. The test yields a coefficient of consolidation ($C_v$) parameter which represents this constitutive relationship. Samples were run at a set strain rate of 0.5%/h until 15% strain was reached after 30 hours. The strain rate was set near the lowest possible value that the machine could impose in order to mimic the very small sediment overburden loads at experimental scale. This resulted in issues with recorded $C_v$ values, with negative pore pressure differentials and unusable data common at many times during the test.
Based on usable portions of the data, $C_v$ was recorded as $3.78 \times 10^{-10} \text{ m}^2/\text{s}$ for the clastic sediment and $7.11 \times 10^{-10} \text{ m}^2/\text{s}$ for the marsh proxy. These values are lower than those of most clays collected from the field (Carter and Bentley, 1991), casting doubts on their accuracy. A low $C_v$ for the clastic sediment is not entirely surprising because its polymer component likely restricts fluid flow through pores. The marsh proxy, however, is untreated with polymer and expected to have a higher value than the one reported.

2.3 Data Collection and Processing

2.3.1 Digital Elevation & Subsidence Maps

Digital elevation maps (DEMs) of the basin were collected at least once every hour of experimental runtime using a stationary FARO LiDAR scanner similar to Straub et al., (2015) and Li et al., (2017). For the control experiment, “dry” scans were collected at the beginning of each run hour while fluvial input was paused, and “wet” scans were collected 48 minutes into each hour while the fluvial system was active. For the marsh experiment, dry scans were collected every even hour of runtime and wet scans 48 minutes into each run hour. LiDAR generated point clouds were cleaned in MATLAB and gridded into 5x5 mm raster pixels, each representing the median value of all points within the pixel. Each pixel has a vertical uncertainty of approximately 0.71 mm, which is discussed in the following error propagation section.

Subsidence maps for both experiments were generated by differencing digital elevation maps (DEMs) and screening out areas that were flooded by sea level or that received surface water flow at some point during the timestep. This was done to remove all sediment transport processes and isolate subsidence in “quiescent” areas. Areas covered by surface water were removed by a color threshold screen. Differenced DEM values smaller than -1000 μm or greater than 5000 μm were considered erroneous and removed from the dataset. Subsidence rates and magnitudes calculated by using all available pixels not removed by previously described screens are referred to as “delta-wide”, but do not include subsidence due to sediment loading from active channels.
Control experiment DEMs were differenced by appropriately screening each scan, subtracting the previous dry scan from a given scan, and then summing the resulting two-hour DEMs of difference. The marsh experiment DEMs were differenced by subtracting the wet scan collected 72 minutes prior from each dry scan. This was done to exclude marsh accretion in the first 48 minutes of each two-hour period of run time. The resulting subsidence maps were then multiplied by a correction factor of 120/72 to account for “lost” time and make them equivalent to two-hour DEMs of difference. Therefore, subsidence rates can be compared between the experiments at a two-hour temporal resolution. Subsidence rates were also compared at ten-hour timesteps by summing consecutive two-hour subsidence maps. An initial thickness map of each marsh deposit was separately quantified by differencing the wet scan taken right after distribution and the dry scan taken right before, then screening out areas outside of the marsh window. Areas receiving surface water were excluded in the same manner as in the subsidence maps.

Ringing artifacts due to angular measurement error were present in the DEMs of both experiments. Banding was somewhat more pronounced in the marsh experiment subsidence maps, possibly because the magnitudes of all values were multiplied by the previously discussed correction factor. These artifacts were mitigated by running each two-hour DEM of difference through a MATLAB data cleaning script before applying any screens. This script identified the center of ringing as the location of the LiDAR scanner relative to the basin. It then broke each map into a series of radial bands with a width of five pixels. The median value all pixels within each band was subtracted from the initial value of each pixel within a given band to remove the artifact. Real subsidence patterns generally did not align with the orientation of these bands, so they are assumed to be relatively unchanged. Bulk subsidence rates with and without ringing removed were similar, suggesting that this cleaning step did not bias subsidence rates. Additionally, overall subsidence rates throughout the experiments were nearly unchanged, though local subsidence rates at a given time and location were often altered,
presumably now reflecting actual elevation changes rather than instrument artifacts. Due to the slightly elliptical shape of the bands, and the irregularity of their magnitudes and locations, some ringing remains in both datasets.

2.3.2 Error Propagation

The FARO uncertainty in ranging measurement is listed as +/- 1mm. The FARO was positioned above the deposit and collected returns at an approximately 45-degree angle, so vertical uncertainty can be estimated as $1/\sqrt{2}$, or 0.71mm. The standard error of the mean on each pixel of a difference map can be described by the equation,

$$S_z = \sqrt{(S_{z1})^2 + (S_{z2})^2}$$

where $S_{z1}$ and $S_{z2}$ are the vertical uncertainty on one point at times one and two. The standard error on each pixel of the difference map comes out to +/- 1mm. This is larger than almost all observed subsidence magnitudes over timescales of interest in the experimental setting, so values need to be spatially averaged over many pixels to yield workable error bars. Average uncertainty over a number of individual pixel measurements ($S_{z,n}$) is described by the equation,

$$S_{z,n} = \left(\frac{1}{n} \cdot \sqrt{S_z}\right)^2$$

where $n$ is the number of pixels averaged. Error decreases as $n$ increases. For $S_z = +/- 1$mm and $n = 1000$, $S_{z,n}$ is .001mm or 1 micron. Delta-wide or “low-elevation zone” average subsidence values calculated in section 3.1 always average over at least this many pixels, so uncertainty is assumed to be negligible. All spatially distributed subsidence measurements in Figures 6 and 7 are created by averaging over 10x10 pixel areas (50x50 mm) with at least 40 non-NaN values ($n >= 40$). In this case, $S_{z,n} = 0.025$mm or 25 microns. This is significantly smaller than the range of typically measured values and assures the validity of correlations drawn in those analyses.

The error propagation technique outlined in this section assumes a lack of spatial or temporal correlation in errors. I make this assumption only after applying the ringing artifact...
mitigation script described in the previous section. These artifacts are an example of spatially correlated error. For this reason, anomalous subsidence measurements calculated over a small number of pixels should be met with skepticism.

2.3.3 Groundwater Measurements

A vertical array of 1/8" piezometers at 1 cm spacing was placed near the expected mean shoreline of the marsh experiment (1.2m from inlet channel) in order to become buried by a mixture of clastic and marsh proxy sediments. Hydraulic head at each vertical interval was measured every 10 minutes through a MATLAB computer vision script written specifically to identify water levels in each piezometer (see Figure 2).

![Figure 2. An image of water levels in each piezometer taken during experimental runtime. The red marks overlaying the checkerboard at left were generated with a computer vision script in MATLAB. Checkerboard corners were identified to produce a real-world coordinate system for each image. The water level in each well was then identified by a color threshold and assigned an elevation relative to sea level in the experiment.](image)

The dataset was normalized by a control well measuring sea level, so all measurements are relative to sea level. Rapid decreases in hydraulic head can be related to consolidation (i.e. subsidence) of soft sediment through the Terzaghi 1-D consolidation equation (Terzaghi, 1943):

$$C_v \frac{\partial^2 u}{\partial z^2} = \frac{\partial u}{\partial t},$$
where \( \frac{\partial u}{\partial t} \) is the change in pore pressure per unit time, \( dz \) is the change in sediment thickness, and \( C_v \) is the coefficient of consolidation. Ideally, such an analysis would identify sediment consolidation as a driving subsidence mechanism and point to the stratigraphic intervals at which this subsidence occurs. However, difficulty obtaining reliable \( C_v \) values at the very low stresses imposed in laboratory scale delta experiments, combined with variable groundwater discharge from a mobile fluvial system have precluded the procedure to this point.

3. Results

3.1 Overall Subsidence Trends

Both experiments exhibit measurable subsidence in quiescent sub-aerial portions of the delta. The marsh experiment has a higher delta-wide mean subsidence rate of 126 micron/h compared to 54 micron/h for the control experiment. Figure 3 shows that, although delta-wide subsidence rates between the two experiments sometimes overlap, the marsh experiment exhibits significantly more subsidence throughout its evolution. Cumulative average subsidence across the entire subaerial delta top is approximately 7050 \( \mu m \) in the marsh experiment and 2980 \( \mu m \) in the control experiment. Marsh experiment subsidence outpaces the control by a factor of 2.40.

![Figure 3](image)

**Figure 3.** (A) Cumulative sum of delta-wide average quiescent subsidence throughout experimental runtime. (B) Distributions of delta-wide average quiescent subsidence for each 10-hour timestep for both experiments. Delta-wide subsidence rates are generally greater in the marsh experiment relative to the control.
Although the spatial structure of subsidence is very different between the two experiments, subsidence rates are stable throughout the duration of both. Figure 4 shows the overall spatial and temporal trends in subsidence rates in low elevation coastal zones (defined as between sea level and 15 mm above sea level). This elevation window covers a larger area on average in the marsh experiment, but it was chosen to approximate all locations potentially susceptible to being flooded during one cycle of delta lobe building across the basin.

Part C of Figure 4 shows that subsidence rates are constant throughout the entire timescale of each experiment, but regularly fluctuate by more than 100 μm/h across the delta top over ten hour windows. The deposit is constantly aggrading throughout the experiment run time, so the lack of a gradual increase in rates demonstrates that subsidence is uncorrelated with stratigraphic thickness. Low-elevation zone averaged rates rarely exceed imposed sea level rise (RSLRb) for the marsh and never do in the control. Therefore, most of the total relative sea level rise is created by imposed sea level rise, as opposed to compactional subsidence, in both cases.
Figure 4. (A) Overhead image of control experiment at hour 180 overlain with a LiDAR generated low elevation zone (0-15 mm RSL) subsidence map (5 mm x 5 mm resolution) of the previous 10 hours. (B) Overhead image of marsh experiment at hour 520 overlain with a lidar generated low elevation zone subsidence map of the previous 10 hours. (C) Time series of delta-wide average subsidence rates for each ten-hour timestep for both experiments. Subsidence rates are variable through time, but do not exhibit a temporal trend. The constantly imposed sea level rise (RSLR) is also plotted. (D) Time averaged profiles of subsidence as a function of elevation above sea level for both experiments. The window of active marsh deposition for the marsh experiment is shaded yellow.
Parts A and B of Figure 4 show that subsidence is highly concentrated near the coastline in the control, but more dispersed throughout the subaerial delta in the marsh experiment. Subsidence in the control is clearly related to elevation above sea level in Figure 4(D). Subsidence rates follow a bell curve shape which peaks with rates of approximately 250 micron/h at 4-5 mm above sea level and becomes minimal at around 10 mm above sea level. In contrast, marsh experiment subsidence rates are relatively consistent throughout the 15 mm elevation window. They are slightly higher in the window of active marsh deposition, but only decrease by about 10-15% in the centimeter above the marsh window. It should be noted that most locations outside the active marsh window pass through the window at different delta configurations, and therefore have received some amount of marsh deposition.

3.2 Control Experiment Subsidence Pattern

The near-coast subsidence “band” is the only significant subsidence feature in the control experiment. It is relatively consistent in magnitude and orientation relative to the coastline throughout time (Fig. 4D). This feature is well represented in Figure 5(A). The band is consistently parallel to the shoreline, roughly 15 cm across, and subsiding at 250-300 µm/h at its center. As an abandoned delta lobe is flooded by rising sea level, the band moves landward (Fig. 5B). Topography profiles cross the band from T-T’ at the beginning and end of the timestep from hour 170-180. They show bed lowering slightly above sea level, and a slight increase in elevation directly below sea level. The result is a translation of the steepest topography landward at the end of the timestep. Landward movement of the subsiding band with rising sea level is persistent for the entire time that an area of the delta remains free of fluvial reworking.
3.3 Marsh Experiment Subsidence Pattern and Mechanism

The subsidence patterns in the marsh experiment are more complex than in the control, and their geometry cannot be simply described. Subsidence rates are highly transient and do not persist through time at the two-hour scale, our smallest possible timestep of analysis. Figure 6(A) shows the distribution of correlation coefficients between average subsidence at each 5 cm x 5 cm location in a given two-hour subsidence map, and subsidence at each location in the next subsidence map. In greater than 90% of instances, localized subsidence rates (5 x 5 cm resolution) are uncorrelated (r<0.4) with future subsidence rates. This means that subsidence “hot spots” rarely exist through multiple measurements at the same location.
Localized subsidence rates correlate somewhat better with thickness of the most recent
marsh deposit (distributed immediately before the subsidence map is taken) at the same
locations, with $r$ values typically ranging from 0.3 to 0.6 for a two-hour timestep (Figure 6B).
Many of the poorly correlated timesteps occur when only a relatively small portion of the delta
top is map-able for measurement of quiescent subsidence. More complete subsidence and
marsh maps tend to yield moderate to good correlation (see Figure 7). In these cases, the
spatial structure of areas with high subsidence is very similar to the distribution of areas which
have received thick marsh deposits in the most recent distribution cycle. Correlations between
two-hour subsidence and the thickness of the most recent two or three marsh deposits are
worse.

Figure 6. (A) Histogram of correlation coefficients between subsidence at a given two-hour
timestep with subsidence at the next two-hour timestep in the marsh experiment. Each R value
is generated by gridding each subsidence map into 10x10 pixel blocks, averaging value within
every block, plotting the subsidence rate at each block against the subsidence rate at each
block 2 hours later. In the vast majority of timesteps, there is little to no temporal correlation
between subsidence rates. (B) Histogram of correlation coefficients between subsidence at a
given timestep and the thickness of the marsh deposit distributed immediately prior. R values
were gathered in the same manner as in Part A. Subsidence rates generally have a low to
moderate correlation with prior marsh deposit thicknesses at the same location.
3.4 Groundwater Data

Cyclic changes in hydraulic head seem to be primarily attributable to pauses in the experiment and associated losses of fluvial input to the basin, or the proximity of the main channel to the well array (see Figure 8). Figure 8(D) shows that a vertical pressure gradient develops in an area that is being actively loaded by fluvial water and sediment. Piezometers at...
greater burial depth experience some overpressure. Therefore, the deposit is not entirely “well-drained”. However, overpressure is small and not highly persistent through time, making it difficult to apply the Terzaghi consolidation relationship. This dataset could still prove useful in obtaining hydraulic conductivity values for the lithologically diverse facies in the experiment. It could also be used to better understand groundwater-surface water interactions in a system with a highly mobile channel network rapidly traversing the basin. Any signal related to pore water expulsion from consolidating sediments is relatively minimal (<1mm) and obscured by larger overprinting signals.

Figure 8. (A) All hydraulic head measurements relative to sea level from each piezometer throughout experimental run time in “real world” rather than “experimental” time. Measurements that deviate from sea level at the cm scale or register as negative are due to poor hydraulic connectivity in piezometers. (B) Data from hour 450-560 in a deeply buried piezometer (well 4) and a shallowly buried piezometer (well 14). The major signal is daily spikes due to the fluvial system being turned on and off as experiment is paused overnight. (C) Data from hour 454-462 when the well array is located in an abandoned delta lobe (fluvial system is far away). Piezometer measurements track each other closely. (D) Data from hour 508-520 when the well array is in area receiving fluvial input. A small vertical hydraulic gradient is apparent as piezometer measurements diverge.
4. Discussion

4.1 Interpreting Spatial and Temporal Subsidence Trends

The experimental results presented in this paper suggest that the spatial and temporal structure of shallow subsidence rates are strongly influenced by the presence of highly compressible marsh deposits being continuously deposited in low elevation areas of river deltas. Subsidence patterns in the marsh experiment are much more widely dispersed, and therefore more realistic than the entirely elevation-controlled near-shore subsidence seen in the control experiment.

The apparent subsidence band in the control experiment is unlike large scale subsidence patterns observed at the field scale. Soil creep could be responsible for lateral movement of sediment from the shoreface into the shallow nearshore (maybe add an arrow in fig. 5 inset). Although it is not a purely vertical subsidence mechanism like consolidation, creep is commonly experienced in coastal areas such as salt marshes and plays a role in bed lowering at a local scale (Mariotti et al., 2016). Unlike vertical sediment consolidation, which occurs as pore water is expelled, creep can be triggered by a rising water table decreasing friction between grains and causing failure (Mariotti et al., 2019). The water table around the coastline in the experiment is generally rising with sea level, so creep is deemed the most likely mechanism triggering the “subsidence” band. This mechanism and its signature are likely present in all laboratory delta experiments which use a similar cohesive, well graded sediment mixture to the control experiment.

Subsidence in the marsh experiment seems to be mostly controlled by very recent marsh accumulation and must be occurring mostly in the very shallow (<2 mm) subsurface. This claim is supported by the fact that subsidence rates do not increase with increasing stratigraphic thickness as the deposit builds up throughout the experiment.

Remarkably, Figure 4(D) shows significant subsidence occurring outside of the “marsh window,” or area of active marsh deposition. Because the addition of marsh is the only change
between experiments, this suggests that the marsh treatment is influencing subsidence outside the active marsh window. This is unexpected because subsidence is shallow and correlated with recent deposit thickness. The only explanation for these seemingly incompatible results is that a large portion of subsidence in the marsh experiment is controlled by surficial compaction of clay layers, but a degree of marsh deposit compaction continues to occur for some time post-deposition. Overall, it seems that marsh sedimentation decreases the slope of the delta, removing the near-shore creep failure mechanism, and adding a more spatial variable mechanism of marsh deposit compaction/consolidation.

The behaviors described above establish that the experiments, and particularly in the marsh experiment, have strong spatiotemporal variation in subsidence rates. Many field scale deltas such as the Mississippi also exhibit such variation. If subsidence patterns in the marsh experiment are to be considered predictive of field scale behavior, we find the “hot spots” or “local subsidence maxima” often observed on modern deltas (Morton et al., 2003) are highly transient at timescales longer than those measured by continuous field observations. In other words, there is a rapid turnover time from high local rates at short timescales (Karengar et al., 2015; Nienhuis et al., 2017; Byrnes et al., 2019) to much lower millennial-scale or longer basin wide averages (Kooi and de Vries, 1998; van Asselen, 2011). This conflicts with field studies from South Louisiana which insist that millennial-scale compaction rates are controlled by the thickness of the entire Holocene sediment package (Törnqvist et al., 2008; Byrnes et al., 2019), and should therefore continue to be large in areas with a thick sediment package. Some of this discrepancy could potentially be described by issues scaling the sediment consolidation process from laboratory to field scale discussed in the methods section. It could also mean that the overwhelming majority of compactional subsidence is occurring in the very shallow subsurface and is dominantly controlled by recent marsh deposition, as outlined in Section 3.3. Ongoing work to collect porosity values from the experimental deposit will further test this hypothesis.
Recent research comparing bulk densities of Mississippi River Delta sediments over a wide range of lithologies and burial depths shows that most compaction occurs within the top meter in organic rich deposits less than 100 years old (Keogh, 2020). These findings match remarkably well with our experimental results (Figs. 4C, 6, 7), which suggest that most subsidence is occurring in recent marsh deposits much shallower than one channel depth, on the order of one centimeter in experiments and one meter in the field. This near surface bed lowering is continuously replenished by new marsh accretion and therefore, probably does not contribute much to the long-term generation of accommodation space. Even though subsidence is heightened by the addition of marshes to a deltaic environment, their near surface compression likely has relatively little effect on delta-top kinematics. In contrast, nearly half of the total accommodation space in the Holocene Rhine-Meuse delta package has been attributed to peat compaction, which is hypothesized to slow delta progradation and aid in the formation of natural levees (van Asselen, 2011). I do not expect that compaction induced subsidence in quiescent areas of the marsh experiment is deep seated enough to significantly alter these processes. Subsidence under areas receiving active sediment loading, which has not been addressed, could occur at greater depths, though groundwater measurements from Section 3.4 indicates that overpressures during times of loading are relatively small and probably do not have much on an effect on subsidence rates.

4.2 Linking Experimental Subsidence to Meso-scale Dynamics

Results from Figure 4 (C) show that RSLR$_{b}$ nearly always outpaces average subsidence rates across low elevation regions (0-15 mm above sea level) at any given time. However, this is often not the case for modern measurements of field scale deltas (Table 2). This could be due to imperfect scaling between the experimental setting and the field, anthropogenic alteration of modern deltas, a relationship between subsidence rates and the timescale of measurement, or some combination of these factors. A comparison of compactional subsidence to allogenic or imposed relative sea level rise at several experimental and field deltas is illustrated in Table 2.
Table 2.

<table>
<thead>
<tr>
<th>Delta</th>
<th>$\sigma_s$ $\mathrm{(mm/y)}$</th>
<th>RSLR$_b$ $\mathrm{(mm/y)}$</th>
<th>$\sigma_s$/RSLR$_b$</th>
<th>Timescale of Measurement, $t_{\text{meas}}$ (y), *(h)</th>
<th>Compensation Timescale, $t_{\text{comp}}$ (y), *(h)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Control (TDB-18)</td>
<td>54*</td>
<td>250*</td>
<td>0.22</td>
<td>*2-10</td>
<td>*50</td>
</tr>
<tr>
<td>Marsh (TDWB-19-2)</td>
<td>126*</td>
<td>250*</td>
<td>0.50</td>
<td>*2-10</td>
<td>*~50</td>
</tr>
<tr>
<td>Modern Mississippi</td>
<td>7.1</td>
<td>4.3</td>
<td>1.66</td>
<td>~10</td>
<td>~$10^5$</td>
</tr>
<tr>
<td>Modern Mekong</td>
<td>16</td>
<td>4.2</td>
<td>3.81</td>
<td>~10</td>
<td>N/A</td>
</tr>
<tr>
<td>Modern Po</td>
<td>4</td>
<td>4.2</td>
<td>0.95</td>
<td>~10</td>
<td>N/A</td>
</tr>
<tr>
<td>Holocene Mississippi</td>
<td>1-5</td>
<td>~1.5</td>
<td>.67-3.33</td>
<td>~$10^3$</td>
<td>~$10^5$</td>
</tr>
<tr>
<td>Hydrodynamic Model</td>
<td>0.1-1</td>
<td>~1</td>
<td>0.1-1</td>
<td>~5x$10^4$</td>
<td>~$10^5$</td>
</tr>
</tbody>
</table>

Relative importance of compactional subsidence ($\sigma_s$) vs allogenic relative sea level rise (RSLR$_b$) for field scale, experimental, and modelled river deltas at a range of timescales. Rates for Mississippi (Neinhuis et al., 2017), Po (Svyetski et al., 2009), and Mekong (Erban et al., 2014) are based on modern measurements. Millennial scale rates for the Mississippi (Törnqvist et al., 2008) and longer timescale rates from a hydrodynamic compaction model (Kooi and de Vries, 1998) are included to compare subsidence measurements at different timescales. RSLR$_b$ were estimated for field deltas by adding 1mm/yr of deep-seated subsidence to eustatic sea level rise rates. The compensation timescale was calculated for the control experiment and assumed to be roughly the same for the marsh experiment.
The ratio of compactional subsidence to imposed sea level rise ($\sigma_s/RSLR_b$) represents a simple attempt at describing the relative importance of compactional subsidence in the generation of total relative sea level rise (RSL) and associated risks of land-loss due to marine flooding. Its nondimensional nature allows for a crude comparison between laboratory and field scales. $\sigma_s/RSLR_b$ represents a physically meaningful dimensionless parameter because $RSLR_b$ generates accommodation space which allows for marsh deposits to aggrade in place. These deposits then compact, resulting in $\sigma_s$. For field, modelling, and laboratory studies, $\sigma_s/RSLR_b$ varies by no more than one order of magnitude over many orders of magnitude in timescales and physical dimensions of the delta system.

This dimensionless subsidence parameter is on average 0.22 in the control experiment, 0.50 in the marsh experiment, and generally greater than 1.0 in modern measurements of field scale deltas. However, experimental values should be considered conservative because subsidence measurements were only taken in areas that were not being actively loaded by sediment. Dimensionless scaling of subsidence rates in the marsh experiment demonstrates that the overall importance of compactional subsidence in this physical experiment approaches that of field scale deltas, even if it is somewhat lower. This means that shallow compactional subsidence alone can reproduce similar rates and patterns of subsidence to those seen in the field. Figure 9 shows that average low elevation zone compactional subsidence rates from the marsh experiment (2,4,6,8, and 10-hour distributions are each a boxplot) fit a similar trend to field data and numerical models when plotted in dimensionless space.
The relative importance of subsidence due to compaction versus other providers of accommodation space ($\sigma_s/\text{RSLR}_b$), termed the “dimensionless subsidence ratio” here, is plotted against a dimensionless time scale (timescale of measurement normalized by the compensation timescale for each system) for experimental, field, and numerical modelling data. The compensation timescale is roughly equivalent to the time for one channel depth of aggradation to occur across the entire basin, and the timescale necessary to smooth relief across the entire delta (Wang et al., 2011). Because both marsh deposition and subsidence rates are controlled

![Dimensionless subsidence ratio vs. timescale](image_url)

**Figure 9.** Relative magnitudes of compactional subsidence for marsh experiment (boxplots) normalized by RSLR$_b$ along with field and numerical modelling data at a range of timescales (see Table 2). The x-axis is timescale of measurement normalized by compensation timescale. The dashed line is an estimated regression ($\sigma_s/\text{RSLR}_b = -0.36\log(t_{\text{meas}}/t_{\text{comp}}) + 0.87$) through 95th percentile values of subsidence in low-elevation zones of the marsh experiment at 2, 4, 6, 8, and 10-hour timesteps. Large average subsidence values of subsidence across the entirety of low elevation zones declines as the timescale of measurement increases. As the timescale of measurement decreases, it is more likely that measured subsidence rates, even across large areas of the delta top, will exceed a proposed threshold for marsh drowning.

The relative importance of subsidence due to compaction versus other providers of accommodation space ($\sigma_s/\text{RSLR}_b$), termed the “dimensionless subsidence ratio” here, is plotted against a dimensionless time scale (timescale of measurement normalized by the compensation timescale for each system) for experimental, field, and numerical modelling data. The compensation timescale is roughly equivalent to the time for one channel depth of aggradation to occur across the entire basin, and the timescale necessary to smooth relief across the entire delta (Wang et al., 2011). Because both marsh deposition and subsidence rates are controlled
by topography, the compensation timescale effectively resets their spatial distribution, and is
deemed a suitable timescale to use for normalization. While median values of the dimensionless
subsidence ratio remain constant with timescale, the upper bound of dimensionless subsidence
ratios decreases at longer timescales of measurement in all settings, as anomalous highs and
lows are averaged together. In other words, there is a natural variability in the average amount
of subsidence occurring in a delta system at any given time, and that variability decreases as
you measure over longer time windows.

The shortest measurement timescale possible in the experiments, a two-hour timestep
associated with a compensation timescale of roughly 50 hours, scales to several thousands of
years of geologic time when compared to a system such as the Mississippi (with a
compensation time on the order of $10^5$ years). Therefore, all experimental measurements
represent much more time than humans have ever directly observed, at least in terms of the
amount of sediment aggradation and fluvial reworking occurring over two experimental hours.
Even so, there is significant scatter and many outliers in the dimensionless subsidence ratio at
this temporal resolution. A regression through 95th percentile average subsidence values at
2, 4, 6, 8, and 10 hour timesteps (dashed blue line in Figure 9) suggests that subsidence ratios
exceeding the criteria for marsh drowning in the experiment ($\sigma_s/RSLR_b > 1.8$) could commonly
be achieved at much shorted timescales of measurement. The equation of this line is,

$$\frac{\sigma_s}{RSLR_b} = -0.36 \log \left( \frac{t_{meas}}{t_{comp}} \right) + 0.87$$

where $t_{meas}$ is the timescale of measurement and $t_{comp}$ is the compensation timescale of the
system. For $t_{meas}/t_{comp} < 3 \times 10^3$, 95th percentile average subsidence values exceed our
experimental criteria for marsh drowning. In this dimensionless space, the Louisiana Coastwide
Reference Monitoring System (CRMS) data plots as one of these very short timescale scenarios
($t_{meas}/t_{comp} \sim 10^{-4}$) at less than 15 years of continuous measurement. These scaling relationships
suggest that CRMS, the dataset most commonly used to diagnose subsidence problems and
inform future subsidence predictions in South Louisiana, has not been operating for long enough to diagnose the natural temporal variability of delta-wide subsidence rates. Even in the absence of anthropogenic impacts, decadal subsidence rates may naturally exceed thresholds for marsh drowning such as the one set in our experiment after Morris et al., (2002), or the much lower threshold ($\sigma_s + RSLR_b > 3\text{mm/y}$) proposed by Tornqvist et al., (2020).

4.3 Future Work

More precise quantification of the degree of marsh compaction post-burial is forthcoming with stratigraphic and geotechnical analysis of the experimental deposit. This component appears to be less important than surficial compaction in the marsh experiment, but contradictory results from Fig 4(D) require further exploration. An analysis of marsh deposit porosity as a function of burial depth will bear out whether significant marsh compaction is occurring beyond the top several millimeters of the sediment surface. If deeply buried marsh deposits have a much lower porosity, they must be consolidating during active loading from channels, a phenomenon which was not measured in the analyses presented in this study. Overburden pressures can be calculated from groundwater data (Section 3.4) to assess this possibility.

The marsh experiment represents an imperfect, but useful first pass at understanding coupling between river deltas and marshes. Future work will link dynamic subsidence rates to marsh platform stability, delta top kinematics, and stratigraphic stacking patterns of coal seams. Continued field and numerical modelling efforts will extend our ability to reliably predict subsidence rates in low elevation coastal zones past the decadal timescale, where they likely remain variable even under natural conditions.
5. Conclusion

This paper represents the first detailed analysis of subsidence within an experimental delta deposit. Measurements of subsidence rates in abandoned delta lobes in both experiments are non-negligible, at least locally, indicating that they could be an important contributor to total relative sea level rise in all cohesive delta experiments, both past and future. The control experiment experienced rapid bed lowering slightly landward of the coast, likely due to soil creep, but had subsidence rates near zero across the rest of quiescent lobes. This mechanism does not scale well to the field, but probably occurs in many laboratory experiments that do not include a marsh proxy. In contrast, subsidence patterns in the marsh experiment resemble field scale measurements from the Mississippi River Delta and other locations in their spatial and temporal heterogeneity, contribution to total relative sea level rise, mechanism of shallow compaction, and correlation with very recent marsh deposition. The consolidation of buried marsh deposits analogous to peat/coal layers at depth appears to be relatively less important than surficial compaction, although some subsidence outside the active window of marsh deposition notably occurs. A comparison of the marsh experiment to previous field and modelling studies shows that the maximum contribution of compaction to overall relative sea level rise across all low-lying areas of a delta increases as the timescale of measurement becomes shorter. Therefore, active shallow subsidence rates measured throughout a given decade or century could exceed thresholds for marsh drowning simply due to natural variability.
6. References


