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Modelling the effect of marine processes on deltaic wetlands

Leonardi, Nicoletta

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Boston University
MODELLING THE EFFECT OF MARINE PROCESSES ON DELTAIC WETLANDS

by

NICOLETTA LEONARDI

B.Eng., University of Pisa (Italy), 2009
M.Eng., University of Pisa (Italy), 2012

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MODELLING THE EFFECT OF MARINE PROCESSES ON DELTAIC WETLAND

NICOLETTA LEONARDI

Boston University Graduate School of Arts and Sciences, 2016

Major Professor: Sergio Fagherazzi, Associate Professor of Earth and Environment

ABSTRACT

Deltaic wetlands are among the most biodiverse systems on earth, provide important ecosystem services, and are natural buffers against violent storms and hurricanes. Marine processes change the planar configuration and internal stratigraphy of deltaic wetlands, and understanding their contribution to wetlands development and deterioration processes is a key issue for society. In this thesis, field data and numerical models are used to investigate the effect of marine processes on the formation and evolution of deltaic wetlands.

The first part of this work focuses on the effect of micro and meso tides on the hydrodynamics, morphodynamics, and stratigraphy of deltaic distributaries. Data from instruments deployed in Apalachicola Bay (FL) were used to investigate the hydrodynamics of river mouths. Investigating the hydrodynamics is the first step for a deeper understanding of sediment transport processes, and possible implications for the morphological evolution of these depositional environments. The effect of micro and meso tides on the morphology and stratigraphy of mouth bars is then explored by using numerical and analytical tools. Mouth bars are the building units of river deltas and
continuous bifurcations around them allow delta progradation and the formation of new deltaic islands.

The second part of this work focuses on the effect of wind waves on salt marsh deterioration using cellular automata and process-based models. Special attention is given to salt marsh resilience to extreme events, to the effect of variable erosional resistance on the large scale morphodynamic response of salt marshes to wind waves, and to the identification of geomorphic features indicative of wetlands deterioration. Results from cellular automata and process-based models are compared to field and literature data.
PREFACE
River Deltas are among the most productive depositional systems on earth, they house half a billion people, host highly diverse ecosystems and provide a natural defense against storms and hurricanes. They can also constitute important reservoirs for oil and natural gas [Paola et al., 2011; Edmonds 2012]. Therefore, understanding mechanisms responsible for delta formation and evolution is a key issue for society. Deltaic deposits are formed by sediment-laden turbulent jets exiting from river mouths [e.g. Bates, 1953; Wright, 1977; Edmonds and Slingerland, 2007; Fagherazzi et al., 2015]. Despite the fact that several studies have analyzed the overall morphology of deltas, many processes affecting their formation and development remain poorly understood. Marine processes can strongly affect hydrodynamic and sediment transport mechanisms at river mouths and their influence is reflected in the diversity of planar configurations and internal stratigraphy of coastal deposits. Among others, wave energy, tidal range, and the degree to which tidal currents determine the flow within the lower reaches of a river have been indicated as important geomorphic agents for delta morphodynamics and mouth bars evolution [e.g. Wright, 1977; Jerolmack 2007].
Mouth bars are one of the main mechanisms for delta formation, by means of their repetitive deposition and distributary bifurcations around them. Thus, morphological and stratigraphic information on mouth bars are important as they could potentially determine the architecture of the entire delta. When a channelized flow enters a body of water the sediment transport rate decreases and a mouth bar evolves through initial deposition, progradation and stagnation, finally leading to channels formation and bifurcation. With
increasing channel bifurcation order there is a systematic reduction in channels length, width and depth, due to a lower jet momentum flux and consequent lower transport distance basinward [Edmonds and Slingerland, 2007]. Coastal processes influence mouth bars formation and evolution. As an example, tidally induced variations in water level create wider mouth bars which develop faster than in the absence of tides. This is mainly due to the fact that tides increase flow spreading at the channel mouth and that low tidal conditions favor a drawdown water profile and an accelerated flow near the river mouth. Moreover, mouth bars forming in distributaries with a small tidal prism maintain a compact shape while higher tidal discharges dissect the bar into multiple channels. The dissection of mouth bars into multiple channels is dictated by the relative importance of tidal velocity with respect to the ratio between water depth and tidal amplitude [Leonardi et al., 2013].

The final stage of delta plains development is vegetation encroachment and marsh establishment whose maintenance is dictated by their surroundings sediment budgets as well as by external agents, such as wind waves or sea level rise. Particularly, salt marshes have been found to be resilient with respect to vertical dynamics because feedbacks among inundation, organic matter accretion, plant growth, and sediment deposition allow the marsh to keep pace with sea level rise [e.g. Fagherazzi et al., 2013; Kirwan et al., 2010]. On the contrary, recent results indicate that salt marshes are inherently weak with respect to horizontal erosion [Fagherazzi et al. 2013]. Specifically, waves forming in large coastal bays can trigger irreversible salt marsh deterioration even in the absence of sea level rise [Mariotti and Fagherazzi 2013]. As a result, saltmarshes do not display
lateral equilibrium but are always contracting or expanding at rates of meters per year [Fagherazzi 2013].

This work focuses on the effect of wind waves and tides on the earliest stages of deltaic wetland development, as well as on the final stage and maintenance of these depositional environments.
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CHAPTER 1. Interplay between river discharge and tides in a delta distributary

The content of this chapter were published in 2015 in Advances in Water Resources. This paper was co-authored by Sergio Fagherazzi (Department of Earth and Environment, Boston University), and Alexander S. Kolker (Louisiana Universities Marine Consortium).

Abstract

The hydrodynamics of distributary channels has tremendous impact on nutrient and dissolved oxygen circulation, transport of sediments, and delta formation and evolution; yet many processes acting at the river-marine interface of a delta are poorly understood. This paper investigates the combined effect of river hydrograph and micro-tides on the hydrodynamics of a delta distributary. As the ratio between river flow to tidal flow increases, tidal flood duration at the distributary mouth decreases, up to the point when flow reversal is absent. Field measurements in a distributary of the Apalachicola Delta, Florida, USA, reveal that, once the flow becomes unidirectional, high-discharge events magnify tidal velocity amplitudes. On the contrary, while the flow is bidirectional, increasing fluvial discharge decreases tidal velocity amplitudes down to a minimum value, reached at the limit between bidirectional and unidirectional flow. Due to the different response of the system to tides, the transition from a bidirectional to an unidirectional flow triggers a change in phase lag between high water and high water slack. In the presence of high riverine flow, tidal dynamics also promote seaward directed Eulerian residual currents. During discharge peaks, these residual currents almost double
mean velocity values. Our results show that, even in micro-tidal environments, tides strongly impact distributary hydrodynamics during both high and low fluvial discharge regimes.

1.1 Introduction

Flow dynamics in delta distributaries determines the fate and transport of sediments, contaminants, and nutrients with important consequences for deltaic deposits, landform evolution, water quality, and deltaic ecosystems [e.g. Dalrymple and Choi, 2007; Edmonds, 2012; Paola et al., 2011; Ganju et al., 2004; Mariotti et al., 2013]. Herein we investigate the effect of micro-tides on the hydrodynamics of a delta distributary. We particularly focus on the interactions between tides and river hydrograph, and on the system response to tides under low and high riverine discharge conditions. Three main aspects of the problem have been taken into account: i) phase delay between water level and velocity fluctuations; ii) tidal velocity amplitudes; iii) tidally induced Eulerian residual currents. These processes are investigated by means of a two months duration instruments deployment in Apalachicola Bay, Florida, USA, where velocity and water level measurements have been collected at the mouth of a delta distributary, and along its reaches, during two consecutive flooding events.

The effect of tides has been recognized as an important factor controlling both the hydrodynamics of the jet exiting river mouths and the morphology of its sediment deposits [e.g. Geyer et al., 2001; Lanzoni and Seminara, 2002; D’Alpaos et al., 2010; Toffolon and Savenije, 2011; Yang et al., 2014; Canestrelli et al., 2010]. Interactions
between tidal oscillations and riverine runoff have been also closely related to water quality, and nutrient concentrations [e.g. Montani et al., 1998]. The presence of tides has three main consequences: mixing is increased, and the effect of buoyancy is partially suppressed; a bidirectional sediment transport is present; the marine-river interface moves in both the vertical and horizontal direction [e.g. Wright, 1977]. When tides enter the river they behave as waves progressing upstream, they distort, and eventually dissipate due to bottom friction and riverine flow. For a bidirectional flow and a propagating tide, the effect of river discharge on tidal amplitude, wave celerity, and phase lag has been explored [e.g. Cai et al., 2014a,b; Toffolon and Savenije, 2011]. It has been shown that the presence of a river discharge has the same effect than increasing friction by a factor proportional to the riverine to the tidal discharge ratio. Thus, increasing river discharge enhances tidal damping, reduces tidal velocity amplitude, and wave celerity, and increases the phase lag between high water and high water slack [e.g. Cai et al., 2012, Cai et al., 2014b]. However, many aspects of the interaction between tidal hydrodynamic and riverine flow deserve further investigations.

Another important mechanism, connected this time to different riverine discharge conditions, is the backwater effect, which could have important geomorphological implications, when combined with tidal changes in water level. This backwater effect can be also referred to as residual slope [e.g. Savenije, 2005; Cai et al., 2014b]. For a gradually varied flow, velocity varies along the channel; consequently bed slope, water surface slope, and energy line slope all differ from each other, and a backwater profile
estabishes. For instance, in a sub-critical flow, the effect of a control point, such as the sea level at the downstream boundary, propagates upstream creating a water profile that gradually adapts to the downstream conditions. For low flow conditions, the water depth at the shoreline is greater than the normal flow depth, and the water surface profile is concave up (also referred to as M1 curve, Figure 1, green line). On the contrary, a drawdown profile (also called M2 curve, Figure 1.1 pink line) is typically established during high flow conditions, when the normal flow depth is higher than the water depth at the shoreline [e.g. Chow, 1959; Lamb et al., 2012].

Figure 1.1 Sketch of the geometry of an idealized distributary channel. Water levels at mean sea level in case of high (pink line) and low (green line) fluvial discharge. The longitudinal axis, $x$, has origin at the river mouth and is positive in the seaward direction. The longitudinal coordinate, $x^*$, is the point up to which the tide propagate.
1.2 Study site

The lower Apalachicola River is located in the Florida panhandle at the terminus of the Apalachicola-Chattahoochee-Flint (ACF) River system (Figure 1.2 a). Apalachicola River is the largest in Florida in terms of flow rate and it belongs to one of the largest river system in the Gulf of Mexico. The system was formed over the past 6,000 years as the Apalachicola River deposited sediments into a shallow shelf, the distal sands of which were reworked to form sandy barriers, spits, and islands [Osterman et al., 2009]. The bay encompasses about 620 km$^2$ of open water with an average depth of 1.9 m at mean low tide. Approximately 80% of the open water zone is composed of soft, muddy, unvegetated sediments and the remainder is divided between oyster reefs and submerged aquatic vegetation [Chanton and Lewis, 2002; Huang, 2002]. Hydrodynamics forcing observed here are likely to be present in other river mouth settings along the Gulf Coast as they are linked to regional meteorological and oceanographic conditions [Allison et al., 2000; Roberts et al., 1989]. The northern Gulf Coast experiences winter cold fronts that reoccur with a 3-10 day periodicity. During the approaching phase of the front, onshore winds force water and sediment landward. After the front passes, winds shift to the north, leaving behind mudflats and other coastal deposits [Allison et al., 2000; Roberts et al., 1989, Draut et al., 2005]. Discharge in the Apalachicola River has ranged between 110 to 8,235 m$^3$/s from 1929 to present; peak flows generally occur in late winter and early spring, and are highly correlated to rainfall events in Georgia, while low flows occur during late summer and early fall [Elder and Mattraw, 1982; Chanton and Lewis, 2002]. During the period of our study, river discharge ranged from 200 to 3,100 m$^3$/s.
Tides in Apalachicola Bay are mixed diurnal and semi-diurnal. The two main semidiurnal components are the lunar semidiurnal (M2), and solar semidiurnal (S2), with amplitudes (half the tidal range) equal to 0.38 and 0.12 m respectively. The two main diurnal components are the two lunar constituents (K1 and O1), with amplitudes equal 0.43 m and 0.37 m respectively.

In this study, we analyze processes acting at the mouth of a distributary channel in the northern part of the Bay (Figure 1.2 b,c). The area is characterized by a gentle seaward slope and a maximum channel depth of 0.5 m, with respect to adjacent, almost flat areas (Figure 1.2 d).

1.3 Methods

We deployed an Aquadopp Nortek Acoustic Doppler Current Meter (ADCP) at the mouth of a distributary channel (29°45'24.23"N, 84°54'41.87"W, Figure 1.2 b,c), and measured water depth and velocity from January 22 to March 12, 2013. The instrument recorded every 30 minutes at 2Hz, averaging over 60 seconds, and with vertical cell sizes of 10 cm. Water depth was calculated by means of ADCP’s piezometers, and pressure values were corrected from atmospheric fluctuations using data from the NOAA station at Apalachicola (NOAA station ID 8728711). Considering the shallow depth of the channel (maximum depth of 0.5 m), we did not detect significant velocity variations along the vertical, and velocity measurements have been depth-averaged. Horizontal velocities were then rotated to align them with the channel axis (Figure 1.2 c). One RBR TWR2050 pressure transducer for water depth measurements was deployed farther upstream, near
Figure 1.2  A) Study area in the Apalachicola delta, Florida. Location of instruments deployment: ADCP deployed at 29°45'29.44"N, 84°54'43.02"W; RBR deployed at 29°47'58.99"N, 84°59'16.66"W, near the Three Brothers Creek. USGS stations near Sumatra (ID 02359170) and near Chattahoochee (ID 02358000); B), C) Distributary mouth and location of ADCP deployment; D) Cross section of the river mouth, the black point indicates the ADCP location, at the bottom of the distributary.

the Three Brothers Creek (29°47'58.99"N 84°59'16.66"W, Figure 1.2 a), between the channel mouth and the USGS station at Sumatra (USGS Station ID 02359170). The RBR sampled every 15 minutes, 512 points at 2 Hz. Surface elevations and discharge measurements from the USGS stations near Sumatra and near Chattahoochee (USGS Station ID 02358000) were used in our analysis as well (Figure 1.2 a). Atmospheric data
were retrieved from the NOAA station at Apalachicola. Surface elevations were geo-referenced with respect to the National Geodetic Vertical Datum of 1929 (NGVD29). For signal processing, we used wavelet analysis which is a valuable tool to analyze tidal processes deviating from the exact periodicity assumption, typical of classical harmonic analysis [Jay and Flinchem, 1997; Grinsted and Jevrejeva, 2004]. Wavelet transforms expand time series into the time-frequency domain, and allow finding transient periodicities as they are localized in both frequency ($\Delta t$), and time ($\Delta \omega$) [Grinsted and Jevrejeva, 2004]. Similarly to sines and cosines for Fourier basis functions, a mother wavelet ($\psi_0$) is used as the basis in representing other time series. The mother wavelet is then dilated and translated in time: dilatations allow the localization in the frequency domain, while its translation allows localization in time [Jay and Flinchem, 1997]. We used continuous wavelet analysis and the non-orthogonal Morlet wavelet. Heisenberg inequality states that there is a lower limit to the product of frequency and time resolution. There is thus a tradeoff between localization in time and frequency such that if the time resolution is improved the frequency resolution degrades. The Morlet wavelet provides a good balance between frequency and time localization, and provides realistic images of the oscillations in data with non-stationary processes such as river flow [e.g. Labat, 2005]. Continuous wavelet transform is defined as the convolution of the time series with the scaled and normalized wavelet and its complex argument can be interpreted as the local phase. We further use the cross-wavelet transform of the time series velocity and water level, whose complex argument is the relative phase lag between the two signals.
1.4 Results

Figures 3 provides an overview of the data retrieved during the period of study. From January 22 to February 11, the average river discharge at Sumatra (FL) was 300 m$^3$/s. After February 11, the Apalachicola River was interested by two consecutive high discharge events, with discharge peaks of 2350 and 3100 m$^3$/s respectively (Figure 1.3 a). Figure 1.3 b shows surface elevations at our measuring stations along the distributary channel, and the Apalachicola River. With increasing discharge, and consequent increase in normal flow depth, the water surface profile shifted from a concave up profile (M1 curve) to a drawdown profile (M2 curve), in order to gradually adapt the upstream normal flow depth to the downstream water level (Figure 1.3 c). The drawdown profile forced the current within a water depth smaller than the flow depth occurring with low river flow and high tide for a distance of at least 3 km (Figure 1.3 c, point A). Freshwater discharge and incoming tidal waves are both time dependent, but their time scale differ widely as the tidal wave changes hourly, while the freshwater discharge changes in terms of days or week. The outward current $u_0$ eventually becomes dominant as the wave progresses upstream or for an increasing ratio of the riverine flow to the tidal flow. Once the riverine flow becomes important, the instantaneous flow velocity of a moving particle is made of a steady component, created by the riverine discharge, and a time dependent component contributed by tides [e.g. Godin et al., 1999; Lanzoni and Seminara, 1998]. Eventually, if the riverine flow is high enough, the oscillating flow at the river mouth is replaced by a unidirectional flow.
Herein, we define fluvial dominated the state during which there is no flow reversal at the distributary mouth and the tidal discharge is small with respect to the fluvial one. In contrast, we define tidally dominated the condition for which the tidal discharge is large enough to allow flow reversal at the distributary mouth. For the tidally dominated period, minimum velocities throughout the tidal cycle \((u_{min})\) are negative. On the contrary, for the fluvial dominated period, minimum velocities throughout the tidal cycle are positive (Figure 1.4 a, orange line). Figure 1.4 shows velocity and water level measurements for the period of study (Figure 1.4 a), as well as their lower frequency constituents (Figure 1.4 b), diurnal (Figure 1.4 c), and semidiurnal (Figure 1.4 d) constituents. Low-frequency velocities and water levels account for frequencies lower than the diurnal one and are therefore representative for storm surges and sea level variations different with respect to the main tidal constituents, as well as for the riverine flow. On the contrary, diurnal and semidiurnal frequencies represent tidal harmonics.

Tidally averaged velocities at the distributary mouth ranged around 0.1 m/s during low flow and reached 0.25 m/s and 0.5 m/s during the first and the second flood respectively (Figure 1.4 b). During the two high discharge events, the riverine flow was such that there was not flow reversal at the river mouth and minimum velocities throughout each tidal cycle \((u_{min})\) were positive (orange line, Figure 1.4 a). Five main meteorological tides characterized the period of interest. On January 30, February 7 and 24, South-West winds of 5 m/s, 8 m/s and 4.5 m/s respectively caused an increase in water level of around 0.6 m and consequent decrease in flow velocities (meteorological high tide, Figure 1.4 b). On February 15 and March 3, winds of 6 m/s and 10 m/s, coming from
Figure 1.3 A) Apalachicola river discharge measurements at Sumatra (ID 02359170). B) Water levels in the Apalachicola River and in the distributary for the period of study. Blue line: ADCP measurements at the distributary mouth; green line: RBR measurements; pink line: surface levels at the USGS station near Sumatra; red line: surface levels at the USGS station at Chattahoochee. Elevations refer to the vertical datum NAVD29. C) Gradually varied water surface profile for high (pink line, 28 February) and low (green line, 1 February) river discharge. Continuous and dashed lines refer to low tide and high tide respectively. North, decreased water levels in the channel of 0.5 and 0.2 m respectively and increased mean velocity values (meteorological low tides, Figure 1.4 b).
1.4.1 Phase lag between water level and velocity

For a tidal wave, the phase delay ($\varepsilon$) between water level and velocity is defined as the phase lag between high water (HW) and high water slack (HWS, i.e. when the velocity is zero). For a simple harmonic wave, $\varepsilon = \frac{\pi}{2} - (\Phi_z - \Phi_u)$, where $\Phi_z$ is the water level phase and $\Phi_u$ is the velocity phase [e.g. Savenije et al., 2008; Lanzoni and Seminara, 1998]. Thus, when $\varepsilon = 0$ slack water periods occur at high and low tide. When $\varepsilon = \frac{\pi}{2}$, ebb and flood occur at low and high water respectively. The above definition is maintained herein for low frequency fluctuations of the water surface (i.e. meteorologically induced fluctuations).

Figure 1.5 b shows the phase relationship between velocity and water level at the distributary mouth for different frequencies, as obtained from the cross wavelet transform of the two time series. The phase relationship is shown as arrows, with $\varepsilon = 0$ corresponding to an arrow pointing up and $\varepsilon = \frac{\pi}{2}$ corresponding to an arrow pointing to the left. Low-pass downstream water level variations, triggered by meteorological events, always displays a phase lag $\varepsilon \approx \frac{\pi}{2}$, with maximum velocities occurring when sea level is lower than its predicted astronomical value (Figure 1.4 b, 5b). This phenomenon is particularly evident in the first part of the study period, corresponding to low flow conditions and characterized by an approximately constant, and low discharge (23
Figure 1.4. A) Water levels (red line) and Velocity (blue line) measurements for the period of study; Envelopes of maximum (yellow line), and minimum (orange line) velocities throughout a tidal cycle. The fluvial dominated period is indicated with a solid line while the tidally dominated period is indicated with a dotted line. B) Low frequency velocity and water level constituents. C) Diurnal tidal constituents. D) Semidiurnal tidal constituents.

January to 1 February, Figure 1.4 b). Mean velocity values are thus affected by storm surges, and the percent increase in velocity is inversely proportional to the surge increase in mean sea level (Figure 1.6). As a result, velocities in the distributary are higher during meteorological low tides, when wind decreases water level below its astronomical value.
Figure 1.5. A) Sketch of water level, velocity fluctuations, and phase lag between HW and HWS. B) Phase relationship between water level and velocity fluctuations as a function of time, for different frequencies. Phase lag $\varepsilon = 0$ corresponds to a pointing straight up arrow, and $\varepsilon = \frac{\pi}{2}$ corresponds to an arrow pointing to the left.

While surge induced water level and velocity variations are always characterized by a phase lag $\varepsilon \approx \frac{\pi}{2}$, the phase relationship between tidal velocities and water levels is strongly affected by the riverine flow, and appear related to the transition from a bidirectional to an unidirectional flow at the distributary mouth. During the tidally dominated period, the phase-lag between velocity and water level approaches zero and slack waters occur around time of high and low water (Figure 1.5 a,b). Once the unidirectional outflow condition is reached ($u_{\text{min}}>0$), the phase lag approaches $\frac{\pi}{2}$, and high and low water correspond to minimum and maximum velocities respectively (Figure
1.5 a,b grey area). These trends are evident for diurnal and semidiurnal tidal frequencies (pink bands, Figure 1.5 b), and for meteorologically forced constituents. No significant trend emerges for the phase lag between velocities and water levels in the overtidies domain. A phase lag close to zero is what is often found in strongly tapering estuaries, while a phase lag close to $\frac{\pi}{2}$ corresponds to a progressive wave, frequently observed in river channels with constant cross section. This change in phase lag is thus in agreement with the transition from estuarine to riverine conditions. The drawdown backwater curve further creates a situation where the cross-sectional area is more constant. In fact, the width convergence that we observe during low flow is now compensated by an upstream increase in flow depth, and this adds to the riverine behavior.

Figure 1.6 . Mean velocity changes due to changes in water level induced by surges form January 23 to February 1 2013. On the vertical axis the percentage of velocity changes with respect to the mean velocity value. On the horizontal axis, surge induced water level fluctuations.
1.4.2 Tidal Velocities

To qualitatively assess the system behavior under low and high discharge conditions, it is convenient to look at velocity fluctuations for one tidal cycle at low flow and during the highest discharge event (Figure 1.7 a). Around the peak of the second flood, the low tide further enhances spatial accelerations created by the drawdown profile. Due to the change in phase lag between velocity and water level fluctuations, low water levels also correspond to maximum velocities. Therefore, the combined effect of a drawdown profile, and the presence of tides promotes a dramatic increase in peak velocity (points A and A’, Figure 1.7 a). Difference in maximum tidal velocities between high flow and low flow are then higher than differences in minimum tidal velocity (the distance between points A-A’ is 60 % higher than the distance between points B- B’, Figure 1.7 a). It is then worth noticing that, going from low to high discharge conditions, there is a strong increase in tidal velocity amplitude (tidal velocity amplitude, \( \nu_{AB} \), for high discharge conditions almost double the tidal velocity amplitude, \( \nu_{AB} \), corresponding to low river discharge conditions).

Figure 1.7 b,c represent tidal velocity amplitudes for the diurnal and semidiurnal tidal components as a function of the minimum velocity throughout the tidal cycle, \( u_{min} \). \( u_{min} \) is positive in the presence of unidirectional flow, and negative in the case of flow reversal. During the tidally dominated period, tidal velocity amplitudes decrease with \( u_{min} \), for both diurnal and semidiurnal tidal components (left side Figure 1.7 b, c). Vice versa, during the river dominated period, the amplitude of velocity fluctuations increase with flow velocity (right side of Figure 1.7 b,c).
Figure 1.7. A) Tidal velocity fluctuations during one tidal cycle at the distributary mouth (ADCP location) corresponding to low, and high river discharge conditions, and to the water surface profiles indicated in 3C. B) Tidal velocity amplitudes of the semidiurnal tidal constituents as a function of the minimum velocity throughout the tidal cycle. C) Tidal velocity amplitudes of the diurnal tidal constituents as a function of the minimum velocity throughout the tidal cycle.

To explain this behavior we use a simplified tidal prism model. Considering the river discharge, $Q_R$, the instantaneous flow between the estuary and the ocean, $Q$, and the instantaneous volume between the free surface and the low tidal level, $P_i$ (see Figure 1.1), from the continuity equation, it follows that:

$$\frac{dP_i}{dt} = Q_R - Q$$

(Equation 1.1)

For a simple tidal harmonic with period $T$, the Volume $P_i$ can be approximated as:

$$P_i = \frac{P}{2} \left( \sin \left( \frac{2\pi t}{T} \right) + 1 \right)$$

(Equation 1.2)
Where $P$ is the tidal prism. Substituting Equation 1.2 into Equation 1.1, it follows that:

$$Q = Q_R - \frac{\pi P}{T} \cos \left( \frac{2\pi t}{T} \right)$$

(Equation 1.3)

From which it is possible to notice that, as the ratio $Q_R T / \pi P$ increases, the flood duration decreases up to the point when flow reversal is negligible and seawater does not enter the estuary, if gravitational circulation is not present [e.g. Luketina, 1998]. The tidal velocity amplitude, $v$, is then proportional to:

$$v = \frac{Q_{EBB}}{B_0 h_{EBB}} - \frac{Q_{FLOOD}}{B_0 h_{FLOOD}} = \frac{Q_R}{B_0} \left( \frac{H \sin(\varepsilon)}{h_{mst}^2 - \frac{H^2}{4} \sin^2(\varepsilon)} \right) + \pi \frac{P}{T} \left( \frac{2h_{mst}}{h_{mst}^2 - \frac{H^2}{4} \sin^2(\varepsilon)} \right)$$

(Equation 1.4)

Where $h_{mst}$, $h_{EBB}$ and $h_{FLOOD}$ are water depths at mean sea level, ebb, and flood respectively, $H$ is the tidal range, $B_0$ is the width of the distributary mouth, $h_{EBB} = h_{mst} - \frac{H}{2} \sin(\varepsilon)$, $h_{FLOOD} = h_{mst} + \frac{H}{2} \sin(\varepsilon)$. When the river discharge, $Q_R$, is much larger with respect to the tidal discharge, and flow reversal is negligible ($u_{min} > 0$), the first term on the right-end side of Equation 1.4 is dominant and the tidal velocity amplitude increases with increasing discharge. Moreover, $\varepsilon$ approaches $\frac{\pi}{2}$, and low water level to tidal range ratios promote high tidal velocity amplitudes. Thus, when meteorological tides are present, tidal velocity fluctuations can be decreased (in case of meteorological high tide), or further increased (in case of meteorological low tide), in agreement with a more pronounced drawdown profile.
On the other hand, when the tidal discharge is larger than the fluvial one, the second term on the right hand side of Equation 1.4, becomes relevant and tidal velocity amplitudes mainly depends on the volume of the tidal prism. The tidal prism, $P$, is the water volume enclosed between the envelopes of high water slack (HWS) and low water slack (LWS) (Figure 1, and Savenije, 2005):

$$P = \int_{-\infty}^{0} H' B ds$$

(Equation 1.5)

Where $H'(x)$ is the range between HWS and LWS as a function of distance (difference between envelopes of the water levels occurring at HWS and LWS along the estuary). $B(x)$ is the estuary width and $x$ is the longitudinal coordinate, starting from the river mouth and considered positive when moving seaward (Figure 1). Assuming that the wavelength is large compared to the length of the river and that different water levels are reached instantaneously along the distributary channel, the tidal range, $H$, exponentially changes along the estuary axis with damping coefficient $\delta$ ($\delta < 0$ when the tidal wave is damped; $\delta > 0$ when the tidal wave is amplified). The tidal range between water slacks,$H'$, is then related to the tidal range, $H$, by:

$$H' = H \cos(\varepsilon)$$

(Equation 1.6)

Where $\varepsilon$ is the phase lag between HW and HWS, assumed constant along the estuary.
Considering the tidal damping and substituting Equation 1.6 into 1.5 it is possible to obtain the following expression to estimate the tidal prism volume [Savenije, 2005, 2012]:

\[
P \sim \frac{HO}{1 - \delta b} \cos(\epsilon)
\]

(Equation 1.7)

Where \( b \) is the convergence length of the stream width, and \( O \) is the surface of the river interested by tidal oscillations. The tidal prism is thus a function of the tidal range, \( H \), of the surface area where the tide propagates, \( O \), of tidal damping, \( \delta \), and of the phase lag, \( \epsilon \).

While flow reversal at the river mouth is still allowed, increasing river discharge decreases the upstream area where tides propagate (Figure 1, the longitudinal coordinate \( x^* \) moves downstream). Increasing river discharge also decreases tidal amplitudes far from the river mouth (higher absolute values of the damping coefficient), and increases the phase lag (\( \epsilon \)) (Figure 1.3 b, green line; Figure 1.5). All these mechanisms contribute to a reduction in flood volume stored within the river, and a consequential decrease in tidal velocity amplitudes. Moreover, by substituting Equation 1.7 into equation 1.4, it is possible to notice that as \( \epsilon \) increases from 0 to \( \frac{\pi}{2} \), the first term in equation 1.4 gains precedence over the second term. In fact, the first term contains \( H \sin(\epsilon) \), while the second term contains \( H \cos(\epsilon) \). The change in phase lag can be thus considered a clear indicator of the changed hydraulic behavior, and it is one of the reasons behind the non-linear behavior illustrated in Figure 1.7.
1.4.3 Eulerian residual currents

Eulerian residual currents are defined as the averaged velocities at a fixed location, and over a tidal cycle. These second-order currents, driven by the nonlinear tidal dynamics, have been recognized as a significant component of the flow field in shallow areas, and can be relevant to investigate river deltas, and estuaries hydrodynamics [e.g. Hunt and Johns, 1963; Huthnance, 1973; Nihoul and Ronday, 1975; Ianniello, 1979; Zimmerman, 1981; Van der Vegt et al., 2006; Burchard et al., 2010; Wang et al., 2013; Zhou et al., 2014; Olabarrieta et al., 2014]. It is important to notice that the net mass transport of water is not only dependent by the mean velocity, but it is rather controlled by Lagrangean trajectories, obtained as the sum of Eulerian velocity and Stokes drift, with the Stokes drift being a mathematical artifact of the Eulerian framework [e.g. Savenije, 2012].

Several mechanisms contribute to the formation of tidal residual currents, which can be decomposed into three main contributions [e.g. Burchard et al., 2010; Cheng et al., 2011]: (1) The density driven flow which depends on buoyancy gradient; (2) asymmetric tidal mixing, which is connected to the correlation between eddy viscosity and vertical shear, and thus to tidal straining, relevant to tidal asymmetry. (3) vertically averaged tidal mean velocity, connected to the residual riverine flow and to non-linear flow mechanisms, which can be explained by using the various terms in the De Saint-Venant equations including the non-linear frictional term, the non-linear advective term in the momentum equation, and the non-linear term in the continuity equation [e.g. Tee, 1977; Parker, 1991]. Residual currents calculated here are depth averaged, and are thus only
representative of the third component described above. Figure 1.8 shows semi-diurnal (blue arrows) and diurnal (black arrows) residual currents. Residual currents exit from the river mouth for most of the record period, and maintain a direction parallel to the mean flow. Moreover, while low discharge residual currents are of the order of magnitude of the mean flow (compare arrows to the dashed line, Figure 1.8), as the river discharge increases, they almost double. This is because once the unidirectional outflow condition is reached, acceleration of the flow at low tide are higher and not compensated by the corresponding flow deceleration at high tide. In fact, for unidirectional flow $\varepsilon \approx \frac{\pi}{2}$, and the increase in velocity during ebb with respect to the velocity at mean sea level can be written as:

$$v_{EBB} = \frac{Q_R}{B_0 h_{EBB}} - \frac{Q_R}{B_0 h_{msl}} = \frac{Q_R}{B_0} \left( \frac{\frac{H}{2} \sin(\varepsilon)}{h_{msl} \left( h_{msl} - \frac{H}{2} \sin(\varepsilon) \right)} \right)$$

(Equation 1.8)

However, the corresponding decrease in velocity during flood is lower, being equal to:

$$v_{FLOOD} = \frac{Q_R}{B_0 h_{msl}} - \frac{Q_R}{B_0 h_{FLOOD}} = \frac{Q_R}{B_0} \left( \frac{\frac{H}{2} \sin(\varepsilon)}{h_{msl} \left( h_{msl} + \frac{H}{2} \sin(\varepsilon) \right)} \right)$$

(Equation 1.9)

The average over a tidal cycle will thus result in a net seaward directed velocity component. Moreover, the value $v_{EBB} - v_{FLOOD}$ increases with increasing discharge, and with the tidal range to water depth ratio.
Figure 1.8. Eulerian residual currents. Blue arrows correspond to diurnal averaged residual currents. Black arrows correspond to semidiurnal residual currents. The dotted line represents the low pass filtered velocity.

The formation of seaward directed, depth-averaged Eulerian residual currents after the transition from tidal to fluvial dominated conditions is also in agreement with previous numerical model results [Leonardi et al., 2013]. On the contrary, high meteorological tides enhance the possibility of landward-directed residual currents (Figure 1.8).

1.5 Discussion

Despite of Apalachicola Bay being a micro-tidal environment, our results suggest that tides can significantly affect the hydrodynamic of distributary mouths, for both low and high river discharge regimes. When the tidal discharge is large compared to the fluvial one, the presence of a riverine flow increases tidal damping, and decreases the amplitude of tidal velocities. However, once the river discharge becomes sufficiently high, the tidal flow becomes negligible, and a unidirectional outflow establishes, with the system transitioning from tidally dominated to river dominated conditions. Once this fluvial
dominated state is reached, the effect of tides at the distributary mouth is surprisingly amplified. The unidirectional flow promotes a change in the phase relationship between velocity and water level fluctuations, with ebb and flood occurring at low and high tides respectively. Moreover, the amplitude of tidal velocities increases with increasing discharge. The presence of tides in the river dominated case (i.e. during high flow regimes) also promotes the establishment of seaward directed Eulerian residual currents at the distributary mouth. During the two high discharge events, residual currents almost double mean velocity values providing evidence of the possible contribution of tides to the outward mass transport.

All of the above mentioned mechanisms could significantly affect biological and ecological processes at distributary mouths. As an example, tidal velocity amplitudes contribute to nutrient flushing, and have been found to rapidly dilute both phytoplankton biomass and nutrient concentration in the lower reaches of distributary channels [Caffrey et al., 2007; Valiela et al., 1997]. Our results may have implications for sediment transport processes as well. For instance, bathymetric surveys of the Wax Lake Delta have shown that tides may promote the erosion of deltaic channels tips, and play a major role in channel kinematics [Shaw et al., 2013; Esposito et al., 2013]. Moreover, even if rivers are often assumed to decelerate at their mouth, the region connecting the upstream reaches to the offshore plume can be an erosional area when a drawdown profile establishes [Lamb et al., 2012]. Under high discharge regimes, the presence of tides enhances the drawdown profile and increases tidal velocity amplitudes, which in turn may promote channels scour. On the other hand, tides may also affect those areas where
deposition is expected. For instance, when the fluvial discharge is large, higher tidal velocity amplitudes have been connected to wider deposits at distributary mouths, as well as to the most likely occurrence of tidal bedding features. Specifically, lamination extent and difference in mud content between successive layers have been found to increase with tidal velocity amplitudes [e.g. Leonardi et al., 2014; Dalrymple and Choi, 2007]. In the presence of bidirectional flow, low tidal velocity amplitudes are associated to mouth deposits with a compact shape, while higher tidal velocity amplitudes generally correspond to deposits dissected into multiple channels [e.g. Wrigth 1977; Leonardi et al., 2013]. The effect of tides on distributary hydrodynamics could also be exacerbated by the shallow depths typical of distributary mouths, which are of the order of 1-2 m. For these depths, even an oscillation in water level of few tens of centimeters can have dramatic consequences for the velocity field.

1.6 Conclusions

We conclude that even very small tides can strongly impact the velocity field at distributary mouths, during both low and high flow regimes. Specifically, while during low discharge conditions the presence of a river discharge increases tidal damping and decreases tidal velocity amplitudes, during very high flow regimes the effect of tides at the distributary mouths is magnified with a noteworthy increase in tidal velocity amplitudes and seaward directed Eulerian residual currents. High discharge regimes lead to a phase change between water level and velocity fluctuations, with minimum and maximum velocities occurring at low and high tides respectively. This change in phase
lag appears a key determinant for the amplification of tidal velocities. The effect of tides at the mouth of distributaries appears thus intensified during very high discharge conditions when even small water level fluctuations strongly impact the velocity field.

1.7 List of variables

\(Q_R\) River discharge

\(B\) Distributary mouth width

\(b\) Convergence length of the stream width

\(h_{\text{msl}}\) Water depth at mean sea level

\(h_{\text{FLOOD}}\) Water depth during flood

\(h_{\text{FLOOD}}\) Water depth during ebb

\(H\) Tidal range (from HW to LW)

\(H'\) Slack tidal range (from HWS to LWS)

HW High water

HWS High water slack

LW Low water

LWS Low water slack

\(O\) Surface area of the distributary channel where the tide propagates

\(P\) Tidal prism Volume

\(u_0\) Velocity due to the river discharge \(Q_R\)

\(\Delta u\) Difference between tidally averaged velocities for different discharge values

\(\delta\) Damping coefficient
\[ \varepsilon \text{ Phase lag between HW and HWS} \]

\[ \Phi_z \text{ Water level phase} \]

\[ \Phi_u \text{ Velocity phase} \]

\[ \nu \text{ Tidal velocity amplitude} \]

\[ \nu_{\text{EBB}} \text{ Difference between ebb velocities and velocities at mean sea level} \]

\[ \nu_{\text{FLOOD}} \text{ Difference between flood velocities and velocities at mean sea level} \]
1.8 References


CHAPTER 2. Modeling tidal bedding in distributary mouth bars

The content of this chapter were published in 2014 in Journal of Sedimentary Research. This paper was co-authored by Sergio Fagherazzi (Department of Earth and Environment, Boston University), and Tao Sun (ExxonMobil Exploration Company, Houston, Texas, USA).

Abstract

Distributary mouth bars are important morphological units of deltas which develop under a wide range of wave, tidal, and riverine conditions, and are known to form highly productive subsurface oil and gas reservoirs. This paper extends previous work on purely fluvial mouth bars, to mixed systems where tides are also present. Under these conditions mouth bars can display alternate layers of mud and sand that can ultimately determine their vertical permeability. Herein we propose an analytical, process based model to explain tidal bedding characteristics and quantify their extent in mouth bars. Findings from our analytical model are compared with results from the numerical model Delft3D. From landward to seaward and in the absence of tides, our analysis shows that a sand dominated zone is followed by a depositional environment made of sand and mud mixtures, and finally by mud dominated areas. With increasing tidal amplitude, the sand-mud mixture zone is gradually replaced by a lamination zone characterized by alternate tidal bedding. Bedding characteristics in mouth bars are defined using the extension of the lamination area and the difference in mud content between coarse and fine sediment layers. Both quantities tend to increase with increasing tidal amplitude. The lamination
zone grows while the difference in mud content decreases for small ratios of mud to sand settling velocity and mud to sand concentration. Bottom friction strongly affects tidal bedding by reducing the length of the zone where lamination occurs and increasing differences in mud content between successive layers.

2.1 Introduction

Mouth bars are dynamic environments characterized by high potential for sediment preservation [e.g. Esposito et al., 2013]. When a river debouches into a receiving basin, it experiences a decrease in velocity and flow momentum with consequent sediment deposition and mouth bars formation [e.g. Wright and Coleman, 1974]. Mouth bars are one of the main mechanisms for delta formation, by means of their repetitive deposition and distributary bifurcations around them [Edmonds and Slingerland, 2007; Jerolmack and Swenson, 2007]. Thus, morphological and stratigraphic information on mouth bars are important as they could potentially determine the architecture of the entire delta.

When a bar becomes emergent, sediment composition plays a crucial influence on the encroaching vegetation and related fauna [Dyer et al., 2000; Laden and Wang, 2003]. Sediment grain size can also control pollutants, which are more likely to adhere to mud for its cohesiveness and chemical properties. Mud content can therefore be considered an indicator of potential pollution at river mouths [De Groot, 1982]. Mouth bars can constitute important reservoirs for oil and natural gas. As a consequence, the stratigraphy of these depositional environments needs to be fully understood in order
to evaluate fluid flow within the reservoir. In fact, oil production and field development depend on the capability of forecasting deposits heterogeneity at all scales of geological variability [White et al., 2004]. Grain size distribution, bioturbation and bedding style, presence and thickness of mud layers, provide important information on reservoir characteristics. For example, vertical permeability has been found to significantly change across heterolithic planar bedding [Schatzinger and Tomutsa, 1997].

Sediment bed characteristics are strongly influenced by marine processes and can be particularly complex due to the interaction of several external drivers. Stratigraphic evidence confirms the role of waves and tides in reworking mouth bars sediments [Allen and Posamentier 1993, Sydow and Roberts 1994]. Among others, the role of tides on the morphology of coastal deposits and its influence on bed layering has been widely recognized [e.g. Dalrymple and Choi, 2007; FitzGerald 2006; Leonardi et al., 2013].

Tidally induced variations in water level create wider mouth bars which develop faster than in the absence of tides. This is mainly due to the fact that tides increase flow spreading at the channel mouth and that low tidal conditions favor a drawdown water profile and an accelerated flow near the river mouth [Leonardi et al., 2013].

Tidal bedding can be considered the stratigraphic expression of tidal cycles, and its presence has been used to identify energetic tidal conditions [e.g. Reineck and Wunderlich, 1968; Shi, 1991].

During high velocity periods tides allow the deposition of only coarse sediments (typically sand), while during low velocity intervals fine suspended sediments are also able to settle at the bottom. The result in an alternation of sand and mud layers which
forms the so called tidal bedding [Davis and Dalrymple, 2012; James and Dalrymple, 2010]. Tidal bedding is thus associated to layers of different composition, texture and color and is one of several types of rythmites (i.e. sequences of sediments that are produced by cyclic conditions) [Reineck and Singh, 1968; Greb and Alcher, 1995; Davis and Dalrymple, 2012].

Typical features of tidal bedding rythmites are repetitive vertical thickening and thinning of alternating sandstone or siltstone-shale laminae couplets. These variations in thickness of successive layers might record velocities changes due to lunar and solar cycles such as diurnal inequality and neap-spring alternations [Dalrymple et al., 1991; Greb and Archer, 1995]. Detailed analyses of these features allow establishing the relationship between moon and earth over geologic time scales [Dalrymple et al., 1991]. Modern tidal rythmites are present, for example, within the upper estuarine reaches of the Bay of Fundy, Canada, and are common in the Bay of Mont-Saint-Michel in France [Dalrymple et al., 1991; Tessier, 1993; Tessier et al. 1995]. Tidal bedding has been documented in micro-tidal environments as well, for example in the Dyfi River Estuary, UK [Shi, 1991]. Longhitano et al. [2012] and Coughenour et al. [2009] present a review of tidal depositional systems in rock records and show how sedimentary structures generated by tidal hydrodynamics are characterized by great variability. Figure 2.1 shows an example of bed layering which has been interpreted as generated by tidally driven currents [after Longhitano et al., 2012]. Coughenour et al. [2009] summarize the state of the art of quantitative analysis used to identify tidal bedding and tidal periodicities encoded in the
geological record. Initial investigations on tidal facies were based on measurements of the thickness and number of sand laminae [e.g. Visser, 1980; Coughenour et al., 2009].

Figure 2.1. A) Rhythmic, cross-and plane-parallel laminations that have been interpreted as generated from oscillating, tidally-driven currents (Eocene, Itu, Brazil). [after Longhitano et al., 2012]. B) Water level and velocity Apalachicola River, FL (1–4 March, 2013). Water level and velocity measurements have been taken at the mouth of one of the distributaries using an acoustic doppler current profiler (29°45'27.65"N; 84°54'42.53"W). Subsequent approaches have been based on spectral estimation techniques which are mainly divided into parametric and non-parametric methods [e.g. Hayes, 1996]. Among the parametric methods, one of the most common is the maximum entropy method. For the non-parametric methods, the use of periodograms, and in particular of the Schuster
periodogram is usual. Researchers have also used predicted tidal heights and current speeds to model laminae deposition and test the efficiency of tidal cycles preservation in tidal rythmites [Archer 1995; Coughenour et al., 2009].

Despite the fact that several studies have analyzed the stratigraphic architecture of tidal environments and related facies distribution [see Dalrymple and Choi 2007 for a review], the complexity of the problem requires more efforts in order to understand the hydrodynamic mechanisms behind the formation of tidal laminations and their effect on small scale stratigraphic units, such as mouth bars. This paper aims at investigating internal structures of distributary mouth bars forming in micro and meso-tidal environments. Specifically, our goal is to identify key sediment transport processes promoting the formation of tidal laminations. For this purpose, we develop an analytical model and compare its results to numerical simulations carried out with the numerical model Delft3D [Roelvink and Van Banning, 1994; Lesser, 2004], considering a simple geometry and homopycnal effluents. For sake of simplicity, we neglect wind waves and Coriolis forces. The above mentioned external drivers may have a non-negligible influence on mouth bars, from both a morphodynamic and stratigraphic point of view. For example, the presence of wind waves in the exposed location of mouth bars could prevent the preservation of laminae in a form that can be recognized with confidence in ancient deposits.
2.2 Numerical Model Setup

Mouth bar formation and stratigraphy are studied by means of the computational fluid dynamics package Delft3D [Roelvink and Van Banning, 1994; Lesser, 2004]. Delft3D allows the simulation of hydrodynamic flow, sediment transport, and related bed evolution [Lesser, 2004].

The model solves the shallow-water equations in two (depth-averaged) dimensions. These equations are the horizontal momentum equations, the continuity equation, the sediment transport equation, and a turbulence closure model. The vertical momentum equation reduces to the hydrostatic pressure assumption because vertical accelerations are considered small with respect to gravitational acceleration and are not taken into account [Lesser et al., 2004]. The sediment transport and morphology modules account for bed and suspended load transport of cohesive and non-cohesive sediments and for the exchange of sediment between bed and water column. Suspended load is evaluated using the sediment advection-diffusion equation and bed-load transport is computed using empirical transport formulae. The bed load transport formulation used in this work is the one proposed by Van Rijn [1993]. The model also takes into account the vertical diffusion of sediments due to turbulent mixing and sediment settling due to gravity. In case of non-cohesive sediments, the exchange of sediments between the bed and the flow is computed by evaluating sources and sinks of sediments near the bottom. Sources are due to sediments upward diffusion and sinks are caused by sediments dropping out from the flow due to their settling velocities [Van Rijn, 1993]. In case of cohesive sediments, the Partheniades-Krone formulations for erosion and deposition are used.
[Partheniades, 1965]. In their formulation, the critical shear stress for erosion is always greater or equal to the one for deposition, therefore intermediate shear stress conditions may exist for which neither erosion nor deposition occur. This cohesive sediments paradigm is in contrast with common assumptions for non-cohesive sediments for which deposition and erosion always occur simultaneously [Sanford

Figure 2.2 A) Computational domain and boundary conditions B) Sketch of a turbulent jet exiting a river mouth.
and Halka, 1993]. However, the existence of a critical shear stress for deposition is controversial. Winterwerp [2007] recently reviewed the cohesive sediment paradigm by means of literature data and was able to reproduce experiments carried out by Krone [1962] without considering the presence of a critical shear stress for deposition. Thus, he concluded that the so-called critical shear stress for deposition does not exist and it is simply a threshold for resuspension. The latter consideration was also postulated by Krone in its original report [Krone, 1962]. These findings are in agreement with observations of Sanford and Halka [1993] in the upper Chesapeake Bay for which model results show poor agreement with field data observations when the presence of a critical shear stress for deposition is taken into account.

Therefore, we choose to assume gross sedimentation rate of cohesive sediments equal to their settling flux $w_m c_m$, where $w_m$ and $c_m$ are settling velocity and concentration of the cohesive sediment fraction [Winterwerp, 2007]. A possible implication of this hypothesis is an increase in the area where mud deposition is allowed.

Sediment transport and morphology modules in Delft3D allow accounting for multiple sediment fractions. The transport of each sediment class is separately calculated taking into account the availability of each fraction within the bed. The erodible bed (comprising the channel) is divided into multiple layers and for each time step the exposed layer (transport layer) is the only one providing sediments to the flow. At every time step the layer thickness is updated. Within each layer, sediments are assumed to be vertically mixed. In our simulations, sediment erosion during one time step never exceeds the thickness of a layer. We used 75 initial layers of 2 cm thickness.
The model mesh is rectangular with rectangular cells, whose long axis is parallel to the flow (Figure 2.2 a). The simulations are designed to investigate a distributary discharging into a body of water with an initial flat bottom subject to tides of different amplitudes. The tested basin depths are 4 and 4.5 m. The river length is 400 m and has been necessarily limited in order to ensure a reasonable computational time once it was verified that the channel length was not affecting the results.

We further consider a fluvial dominated condition characterized by a river discharge much larger than the tidal one. Our simulations are thus representative of systems having a negligible tidal prism or for peak flow regimes that are high enough to prevent strong flow reversal. Relevant systems could be, for example, the Mississippi and Apalachicola delta, USA, and the Sepik River, Papua New Guinea. Figure 2.1 b, shows an example of a distributary channel in the Apalachicola delta that during flooding behaves as fluvial dominated system with a tidally modulated unidirectional flow, under a progressive wave condition. The hydrodynamic of mouth bars under a fluvial dominated case has been investigated by Leonardi et al., [2013] who show that under these conditions a progressive wave condition at the river mouth is promoted by the establishment of a drawdown profile at low tides and consequent flow acceleration [Leonardi et al., 2013].

The domain has three open boundaries: at the seaward boundary a varying water level is imposed to simulate sea level variations due to tides. For the lateral boundaries, we impose a zero-flux boundary condition, consisting of imposing the gradient of the alongshore water level equal to zero (Figure 2.2a). In the channel, a constant discharge is prescribed with values ranging from 900 m$^3$/s to 2000 m$^3$/s. Water level varies with
semidiurnal frequency (30 deg/h) and simulates tidal amplitudes ranging from 0.25 to 2.5 m. Initial conditions consist of a flat bottom and uniform bed composition with non-cohesive sediments everywhere in the domain. Variable channel width-to-depth ratios have been used (90, 70 and 37) as well as variable friction coefficients (Darcy-Weisbach friction coefficients equal to 0.09, 0.02 and 0.04). Width-to-depth ratios were chosen considering that at the channel mouth the width is generally much larger than depths and that width to depth ratios greater than 50 are common [Edmonds and Slingerland, 2007]. We prescribe a constant sediment input of cohesive and not cohesive sediments for each numerical test. The non-cohesive fraction has specific density of 2650 kg/m³, dry bed density of 1600 kg/m³, and median sediment diameter D₅₀ of 200 μm. Cohesive sediments characteristics were chosen in agreement with values provided by Berlamont [1993]. Specific density is 2650 kg/m³, dry bed density is 500 kg/m³, and settling velocities vary from 0.0001 m/s to 0.001 m/s.

2.3 Theoretical Framework for the Turbulent Jet at a River Mouth

As distributary channels discharge into a body of water, they behave like a turbulent jet, experiencing mixing and diffusion [e.g. Bates, 1953; Canestrelli et al., 2007, 2010; Wright and Coleman, 1974; Özoys and Ünlüata, 1982; Wright, 1977; Rowland et al., 2009; Rowland et al., 2010; Falcini and Jerolmack, 2010; Nardin and Fagherazzi, 2012; Nardin et al 2013; Leonardi et al., 2013].

In coastal areas vertical motions are negligible respect to horizontal ones and the shallow water approximation is widely accepted [e.g. Özoys and Ünlüata, 1982]. Under these
conditions the integral jet theory is generally applicable and the turbulent jet has a symmetrical geometric structure with respect to the longitudinal axis [Abramovich, 1963]. The jet can be divided into two regions: a zone of flow establishment (ZOFE) and a zone of established flow (ZEOF). The first zone is characterized by a core of constant velocity, while the second one is characterized by an exponentially decreasing centerline velocity and a self-similar profile for the transverse velocity. The transition between the two zones is the downstream location at which turbulence generated at the margins of the jet propagates towards the center [Bates 1953, Abramovich, 1963].

Özsoy [1977, 1986] proposed an analytical solution for jet parameters and sediment transport in the nearshore area in the vicinity of tidal inlets. The advection-diffusion equation is used to guarantee the conservation of mass of sediments discharged by the river and experiencing gravitational settling. In this framework an ambient concentration distribution can be taken into account and it is assumed that sediment concentration is small with respect to fluid density with small density variations not contributing to the momentum balance. To compute centerline velocity, jet half-width, and centerline sediment concentrations in case of flat bottom with friction, the following normalized parameters are defined (Figure 2.2 b):

\[
\begin{align*}
\xi &= \frac{x}{b_0}, \quad \zeta = \frac{y}{b(x)}, \\
H(\xi) &= \frac{h}{h_0}, \quad B(\xi) = \frac{b(\xi)}{b_0}, \\
U(\xi) &= \frac{uc}{u_0}, \quad \xi_s = \frac{x_s}{b_0}, \\
R(\xi) &= \frac{r_c}{r_0}, \quad C(\xi) = \frac{c_c}{c_0}, \\
C_A(\xi) &= \frac{c_a}{c_0}, \\
\gamma &= \frac{b_0w}{h_0u_0}, \quad \psi = \frac{u_0}{u_c}, \quad \mu = f \frac{b_0}{h_0}
\end{align*}
\]

(Equation 2.1)
Where $b_0$ is the inlet half-width, $b(\xi)$ is the jet half-width, $h_0$ is the inlet depth, $h$ is the water depth, $r_0$ is the jet core at the inlet, $r_c$ is the jet core, $u_0$ is the centerline jet velocity at the inlet, $u_c$ is the centerline velocity, $c_0$ is the concentration at the inlet, $c_a$ is the ambient concentration, $c_c$ is the centerline concentration, $f$ is the Darcy-Weisbach friction coefficient, $w$ is the sediment settling velocity, $u_{cr}$ is the critical shear velocity, $x_s$ is the end-coordinate for the core region and marks the passage between ZOFE and ZOEF. $x_s$ is found imposing the normalized jet core half-width $R$ equal to zero in the ZOFE equations [Özsoy and Ünlüata, 1982, 1986] (Figure 2.2 b, Equation 2.2). The depth averaged equation of momentum and the advection diffusion equation are then solved, using the quasi-steadiness and self-similarity assumptions. The solution along the centerline in case of flat bottom with friction is, for the ZOFE ($\xi < x_s$):

\[
U(\xi) = 1, \\
R(\xi) = \frac{I_1 e^{-\mu \xi} - I_2 (1 + \alpha \xi)}{I_1 - I_2} \\
B(\xi) = \frac{(1 - I_2) (1 + \alpha \xi) - (1 - I_1) e^{-\mu \xi}}{I_1 - I_2} \\
C(\xi) = \frac{X - (I_1 - I_4)H(B - R)c_A}{[R + I_4(B - R)]H}
\]

(Equation 2.2)

And for the ZOEF ($\xi > x_s$):
\[ U(\xi) = \frac{e^{-\mu \xi}}{[e^{-2\mu \xi} + \frac{2\alpha l_2}{\mu l_1}(e^{-\mu \xi} - e^{-\mu \xi})]^{1/2}} \]

\[ B(\xi) = \frac{e^{\mu \xi}}{l_2}[e^{-2\mu \xi} + \frac{2\alpha l_2}{\mu l_1}(e^{-\mu \xi} - e^{-\mu \xi})] \]

\[ C(\xi) = \frac{X - (I_1 - I_4)HUC_A}{[R + I_4B]HU} \]

(Equation 2.3)

Where \( U(\xi) \) is the non-dimensional centerline velocity, \( B(\xi) \) is the non-dimensional jet half-width, \( C(\xi) \) is the non-dimensional centerline concentration, \( R(\xi) \) is the non-dimensional jet-core half-width, \( \alpha = 0.036 \) in the ZOFE, \( \alpha = 0.05 \) in the ZOEF and \( I_1, I_2 \) and \( I_4 \) are numerical constants equal to 0.450, 0.316 and 0.368 respectively.

The quantities \( X \) and \( M \) are:

\[ X(\xi) = \frac{1}{P}\left\{ \int_0^\xi P M d\xi + 1 \right\} \]

\[ M(\xi) = aHUC_A + \gamma[\psi^2U^2(I_2 - I_5) - (1 - I_3)](B - R)C_A + (I_1 - I_4)HU(B - R)C_AQ \]

(Equation 2.4)

With:

\[ P(\xi) = \exp\int_0^\xi Q d\xi \]

(Equation 2.5)

\[ Q(\xi) = \frac{\gamma[R + I_3(B - R)] - \gamma\psi^2U^2[R + I_5(B - R)]}{HU[R + I_4(B - R)]} \]
Where $I_3$ and $I_5$ are numerical constants equal to, 0.6 and 0.278 respectively.

Note that for $C_A$ (non-dimensional ambient concentration) equal to zero, $M(\xi)$ goes to zero as well. These equations will be used in section 5 as the starting point for the process based model. The variables $X(\xi), M(\xi), P(\xi)$ and $\bar{Q}(\xi)$ can be used to write, in a more convenient form, the normalized equation for sediment concentration distribution integrated across the jet cross section:

$$\frac{dX(\xi)}{d\xi} + Q(\xi)X(\xi) = M(\xi)$$

(Equation 2.7)

### 2.4 Inter-layered Bedding Formation

Herein we introduce conceptual considerations on processes allowing the formation of tidal bedding. Three main factors regulate the formation of tidal laminae: i) availability of at least two sediment fractions is necessary to guarantee the presence of multiple facies. ii) Alternating deposition of these two sediment fractions is also required. In case of one sediment fraction continuously depositing across the entire area, intermittent deposition of a second fraction is sufficient to guarantee the establishment of laminae. iii) Sediment and settling characteristics such as sediment concentration, settling velocity, bottom friction and tidal amplitude may play an important role in defining bedding features.
The alternation of erosion and deposition is mainly dictated by shear velocity variations at the bottom. Tides, as well as varying discharge conditions, are responsible for such bottom shear velocity variability. In the presence of sand and mud, as in our numerical tests, variability throughout the tidal cycle of areas allowing mud deposition, triggered by variations in shear velocity, is expected to be greater than corresponding variability in sand deposits. This is mainly due to high values of settling velocity and critical shear stress for erosion of sand with respect to mud. Under these conditions small shear velocity variations slightly affect the erosion of sand, and sand deposition continues to mainly occur near the river mouth.

From numerical model results, it is possible to evaluate net deposition as the difference between gross deposition (D) and erosion (E) for different instants of the tidal cycle, as a function of different river mouth velocities and water depths.

Two possible cases lead to zero net deposition: for \( E \neq 0 \), net deposition is zero when \( E = D \). For \( E = 0 \), net deposition is zero when \( D = 0 \). These two behaviors can be observed for different sediment fractions: for fine sediment fractions, the shear stress is expected to exceed its critical value near the river mouth (this condition also cover \( E > D \) cases) and the condition \( E = D \) determines locations of points of zero net deposition. For coarse sediment fractions, the shear stress does not exceed its critical value and deposition occurs around the river mouth in a tidal cycle. Erosion is not expected to occur in the rest of the domain due to lower shear stress far from the river mouth. Thus, in the latter case, the reduction in sediment concentration in the water column of sand-size material is the only process leading to zero net deposition.
Mud net deposition is shown in Figure 2.3 b for a tidal range equal to 0.75 m (Figure 2.3 a) and for three instants of the tidal cycle (Figure 2.3 b). For this fine sediment fraction, a region of alternated negative and positive net deposition is present due to shear velocity variations at different instants of the tidal cycle. Here, $x_{\text{min}}$ and $x_{\text{max}}$ are the minimum and maximum longitudinal coordinates where net deposition reaches a zero value within a tidal cycle. The region between $x_{\text{min}} < x < x_{\text{max}}$ is a potential lamination area, assuming the deposition of a second fraction of sediments (e.g. sand) everywhere in this zone.

In Figure 2.3 c, yellow bars indicate simultaneous deposition of both mud and sand, while green bars indicate mud erosion and only sand deposition at that specific tidal instant. The result is a lamination extent going from $x_{\text{min}}$ to $x_{\text{max}}$ and an area characterized by continuous deposition of both sediment fractions beyond $x_{\text{max}}$ that extends up to the limit of sand deposit (sand limit, Figure 2.3 c).

The lamination extent may be substantially reduced if sand deposition does not reach $x_{\text{max}}$. The sand and mud mixture zone and the mud only zones are limited by sediments availability as well.

### 2.5 Spatial and Temporal Patterns of Tidal Laminae

The model Delft3D is able to simulate the formation of mouth bars and the presence of tidal laminations in the deposits. Figure 2.4 a shows a planar view of the morphology of a simulated mouth bar. Figure 2.4 b and 2.4 c show two cross-sections of the mouth bar. Different colors represent different percentages of sediment fractions. Red means mud in
Figure 2.3 A) Water level variations and tidally induced velocity variations in 18 hours. B) Mud net deposition along the mouth bar centerline at three instants within a tidal cycle ($T_{\text{min}}$, $T_0$ and $T_{\text{max}}$). Time instants are indicated in figure 2.3 a. The continuous black line represents net deposition of mud during periods of low velocity (instant $T_{\text{min}}$), while the dashed line represents net deposition for high velocity (instant $T_{\text{max}}$). The dotted line is the net deposition for intermediate velocity (instant $T_0$). $x_{\text{min}}$ is the longitudinal coordinate of zero net deposition at low velocity. $x_{\text{max}}$ is the longitudinal coordinate of zero net deposition at high velocity. C) Green bars represent mud negative net deposition and sand positive net deposition at that specific tidal instant. Yellow bars indicate simultaneous deposition of mud and sand. Red bars indicate areas where there is deposition of only mud. The sand limit is the location beyond which sand deposition ceases. Plus and minus signs indicate positive and negative mud deposition.

the absence of sand. Blue means sand and no mud. It is possible to notice that tidal laminations are only clearly present on the distal side of the mouth bar and appear to
terminate rapidly into the muddy prodelta deposits. The extension of tidal laminae increases during mouth bar shoaling (i.e. the extension grows with elevation). Figure 2.4 also indicates variables useful to describe mouth bar geometry such as mouth bar length ($l_b$) and height ($h_b$).

![Figure 2.4 River mouth bar for $f=0.09$, $c_m=0.6$ kg/m$^3$, $w_m=0.001$, $h_t=2.5$ m. A) Planar view of the river mouth bar morphology. B) Longitudinal section. C) Transverse section. The location of the cross sections is indicated in Figure 2.4a. In Figure 2.4a different colors indicate different mouth bar elevations. In Figures 2.4b and 2.4c, different colors indicate different sediment fraction percentages. Red color means mud. Blue color is sand.](image)
Following a notation similar to Edmonds and Slingerland [2007], \( h_u \) and \( h_l \) are defined as the water depth above the peak of the mouth bar and the river depth at the landward boundary (water depths being referred to mean sea level). According to the authors, the formation of mouth bars goes through different phases. The first phase is the initial deposition due to a decrease in jet momentum and consequent sediment settling. The second phase is connected to flow acceleration at the top of the bar and consequent bar progradation. Finally, the bar stops prograding and starts widening, once it is high enough to force fluid around it. The latter step starts for \( h_u/h_l \) values around 0.6 [Edmonds and Slingerland, 2007].

Mud net deposition and sand net deposition have been evaluated at different instants of the tidal cycle, when velocity is maximum and minimum, and for different stages of the mouth bar evolution. We calculated net deposition for \( h_u/h_l \) equal to 1, 0.6 and 0.4. Figure 2.5a, 2.5b and 2.5c refer to these ratios, showing net deposition for mud and sand, for a tidal amplitude \( h_t \) equal to 2.5 m and for the minimum and maximum velocity during the tidal cycle. Figure 2.5a represents net depositional patterns at the earliest stage of the simulation, when the mouth bar is not formed yet. Net depositional patterns maintain the same trends for small mouth bar elevations. Lamination is going to occur in the area between point A and B due to alternated presence of mud. In the area between points B to C, we are going to have lamination as well, this time due to alternating sand deposition, in the presence of mud.

For \( h_u/h_l \) ratios of 0.6 (Figure 2.5b), the bar is at its prograding stage. The lamination area is extensive and comprises the whole footprint of the bar where fluid flow is accelerated.
For low flow velocities, both mud and sand are able to settle on the top of the bar. However, for high velocities only sand can deposit because of its higher settling velocity, while mud is eroded. Around the centerline, where sand is deposited during the entire tidal cycle, sand and mud layers are produced by intermittent mud deposition. At the two sides of the river mouth, either sand or mud deposits are present due to the absence of the other grain size. The final result is an expansion in time of the lamination area both longitudinally and transversally (Figure 2.4).

For $h_u/h_l$ ratios of 0.4 (Figure 2.5c), channelization around the bar begins. During periods of low velocity, depositional patterns are similar to that observed for previous $h_u/h_l$ ratios. However, during periods of high velocity and low water level, depositional patterns change because the flow is confined at the two sides of the bar. The result is that lamination in front of the bar ceases, while lamination at the two sides increases.
Figure 2.5 A) Mud and sand net deposition for $h_u/h_l = 1$ (beginning of the simulation), B) Mud and sand net deposition for $h_u/h_l = 0.6$ (beginning of bar progradation), C) Mud and sand net deposition for $h_u/h_l = 0.4$ (bar widening). For $h_u/h_l = 0.6$ and 0.4, the isolines in the second and third rows represent the location of the bar. The last column indicates the extent of the lamination area.
2.6 Analytical Model for Facies Distribution

According to considerations presented in section 3, laminae extent due to alternate erosion and deposition of only one sediment fraction is confined between two points, \( x_{min} \) and \( x_{max} \), where net deposition is equal to zero when the tidal flow is minimum and maximum.

Given the analytical formulations for centerline velocity and concentration presented in Section 2, it is possible to evaluate the centerline longitudinal coordinates at which net deposition is equal to zero at every instant of the tidal cycle and for each sediment fraction. Gross deposition, \( D \), and erosion, \( E \) along the centerline are evaluated as:

\[
D = wc(\xi) \tag{Equation 2.8}
\]

\[
E = \delta M \left( \frac{u(\xi)^2}{u_{cr}^2} - 1 \right) \tag{Equation 2.9}
\]

Where \( \delta = 0 \) for \( \frac{u(\xi)^2}{u_{cr}^2} \leq 1 \), and \( \delta = 1 \) for \( \frac{u(\xi)^2}{u_{cr}^2} > 1 \).

For cohesive sediment fractions, \( M \) is the erosion parameter of the Partheniades-Krone formulation. For non-cohesive sediment fractions, \( M \) is obtained from the pick-up function proposed by Van Rijn [1984].

For a sediment fraction such that \( u_0(\xi)^2/u_{cr}^2 \leq 1 \), erosion is prevented at the river mouth and, because the velocity decreases with distance from the mouth, no erosion is expected to occur in the whole domain. Therefore, zero net deposition only occurs when
sediment settling is negligible, i.e. the concentration $C(\xi)$ in the water column is zero (Equation 3). For a sediment fraction such that $u_0(\xi)^2/u_{cr}^2 > 1$, the resulting non-dimensional coordinate, $\xi$, at which net deposition is zero at a certain instant is obtained by imposing $U(\xi)$ (Equation 2.3) equal to $u_{cr}(\xi)$ and by solving the second degree equation in $e^{\mu \xi}$:

$$\xi = \frac{\log(f_1)}{\mu}$$

(Equation 2.10)

Where:

$$f_1 = \frac{b}{2a} + \frac{1}{2a\bar{u}_{cr}} \sqrt{b^2\bar{u}_{cr}^2 + 4a}$$

$$a = e^{-2\mu \xi_s} + \frac{2\alpha l_2}{\mu l_1} e^{-\mu \xi_s} \propto 1 + \frac{2\alpha l_2}{\mu l_1}$$

$$b = \frac{2\alpha l_2}{\mu l_1}$$

$$\bar{u}_{cr}^2 = wc_0 \frac{u_0^2}{M(\frac{u_0^2}{u_{cr}^2} - 1)}$$

(Equation 2.11)

and $\bar{u}_{cr}$ is an approximated value for $u_{cr}$ (see Appendix A). The condition $u_0(\xi)^2/u_{cr}^2 > 1$ most likely occurs for fine sediment fractions for which erosion is present. Figure 2.6 shows how locations of zero net deposition vary with the
dimensionless tidal velocity \( (u_t/u_0) \) for two different sediment fractions (sand and mud). Where, \( u_t \) is the component of the periodic velocity amplitude and \( u_0 \) is the mean velocity throughout the tidal cycle. We consider two instants of the tidal cycle, when the velocity is maximum (thick dashed lines) and minimum (thick continuous lines). Thick black lines correspond to mud net deposition equal to zero. Thick red lines correspond to sand net deposition equal to zero. Points above black lines are characterized by mud positive net deposition. Points under red lines are characterized by sand positive net deposition. Solid pink area (zone 1) is characterized by continuous sand deposition in the absence of mud and, thus, it is sand dominated. The pink area of the plot marked with thin black lines (zone 2) is a mud lamination area and is characterized by coarse and fine layers due to alternating mud deposition with continuous presence of sand during the whole tidal cycle. Solid white area (zone 3) is characterized by continuous sand and mud depositions, resulting in a uniform sand-mud mixture. White area with thin black lines (zone 4), is a sand-mud lamination zone and is characterized by lamination due to alternating of both mud and sand deposition. Grey area, marked with thin red lines (zone 5) is a sand lamination zone characterized by lamination due to alternative sand deposition with continuous presence of mud. In the latter case, sand is present while mud is depositing, but it is not always present throughout the tidal cycle. The solid gray zone (zone 6) is mud dominated and is characterized by continuous mud deposition in the absence of
Figure 2.6 Facies model for river mouth bars as obtained from Equations 2.2 to 2.13. On the longitudinal axis, \( u_t/u_0 \) is the dimensionless tidal velocity. On the vertical axis, \( \xi \) is the dimensionless longitudinal coordinate. Zone 1 is characterized by sand deposits in the absence of mud. Zone 2 is a lamination area due alternating mud deposition. Zone 3 is an area characterized by a homogeneous deposition of mud and sand. Zone 4 is a sand-mud lamination area. Zone 5 is characterized by sand alternating deposition with constant presence of mud. Zone 6 is mud dominated.

According to the plot we can move along the vertical axis to proceed along the jet centerline. In the absence of tides (\( u_t/u_0=0 \)), we expect a sand dominated area near the river mouth, then a sand-mud mixture zone and finally a mud dominated zone far from the river mouth. No lamination is present. With increasing tidal amplitude, a lamination area forms, due to variations in flow velocity. For (\( u_t/u_0=1 \)), the following depositional
environments can be found along the centerline: only sandy deposit near the channel mouth (up to point A), sand and mud interlayered deposit (from points A to D) and finally only mud deposits (above point D).

Figures 7a and 7b illustrate the effect of a possible increase in mud settling velocity and sediment concentration for a generic instant of the tidal cycle. These parameters influence gross deposition and do not affect erosion. Reductions in settling velocity and concentration have the same effect and tend to delay the location of zero net deposition (Figure 2.7a, the black dot longitudinal coordinate is higher than the red one). With decreasing settling velocity (or concentration), the deposition curve intersects the erosion curve in locations where the erosion curve has a low gradient. As a consequence, even small vertical variations in the intersection point (due to velocity variations during the tidal cycle) determine significant deviations along the longitudinal coordinate with a possible increase in lamination length. The red marked area in Figure 2.7a represents positive net deposition for high settling velocity (or concentration) and it is larger than the black marked area, representative of positive net deposition for low values of these quantities. Therefore, for high values of mud settling velocity and concentration, it is reasonable to expect a small lamination area near the river mouth and high mud content in the layers. Figure 2.7b illustrates variations in lamination extent (zero net deposition for mud in Figure 2.6) due to a decreasing mud settling velocity and it is a direct consequence of variations in the intersection points of figure 2.7a.
Figure 2.7 A) Erosion and deposition of mud along a mouth bar centerline. The blue line is the erosion curve. Red and black lines are gross deposition curves for different values of settling velocity, \( w_m \), or sediment concentration, \( c_m \), of the mud at the river mouth. The red line corresponds to high values of \( w \) or \( c_0 \). The red and black dots are points of zero net deposition. Their position varies throughout the tidal cycle. The black point is characterized by a low gradient of the erosion curve. A vertical shift of the erosion curve during a tidal cycle causes larger variations in the longitudinal coordinate of the black dot with respect to the red one. On the contrary, steeper gradients of the upper part of the erosion curve cause high differences between the marked areas at different instant of the
tidal cycle. B) Lines of zero net deposition for the mud as in Figure 2.6 for different settling velocities.

The laminae extension increases basinward for low settling velocities of the mud, while an increase in settling velocity (or sediment concentration) tends to increase landward deposition.

2.7 Comparison between analytical and numerical model

The analytical model proposed in the previous paragraph does not take into account bottom evolution and, as a consequence, the expansion of the lamination area connected to the shoaling of the mouth bar (Figure 2.4, Section 3). Figure 2.8 compares the length scale over which tidal laminae can form predicted by the numerical model compared to that predicted by the analytical model. Note that in our idealized models this length-scale corresponds to the length-scale of the individual laminations themselves (as they are considered continuous). In a natural system, where there are many additional processes at work, the length of individual laminations may be very different from the length of the zone under which they are stable (zone 2, 4, 5 in Fig. 6).

As expected, the length of the lamination zone predicted by the analytical model underestimates the numerical model results. However, there is a significant correlation between the two models with analytical and numerical area of lamination having comparable trends (Figure 2.8).
Figure 2.8 Comparison between lamination lengths measured from numerical model results and calculated from the analytical model.

For the difference in mud content between successive layers, a qualitative comparison between analytical and numerical model can be obtained by looking at its distribution along the centerline. Figure 2.9 shows how $\Delta c$ (blue line) and the maximum mud concentration ($C_{\text{max}}$) (red line) vary along the centerline for a typical run with 2.5 m of tidal amplitude. Given a certain longitudinal coordinate, $\Delta c$ is the average, for multiple tidal cycles, of the difference in mud content between two subsequent layers deposited at each tidal cycle.
Figure 2.9 Percentage difference in mud content between successive layers along a mouth bar centerline (blue line) and maximum mud content (red line). Points A, B, C, D are reported in Figure 2.6 as well.

$C_{\text{max}}$ is the maximum mud content, for all vertical layers. It is possible to see that both curves are characterized by three main zones with different slopes (A-B, B-C and C-D). The three zones correspond to the three different area marked in Figure 2.6 (right y-axis). Proceeding downstream from the channel mouth along the centerline we encounter locations having increasing time of mud deposition (Figure 2.3 c). Mud content per layer (in layers where the percentage of mud is higher than sand) as well as maximum mud content can be reasonably related to the amount of time, throughout the tidal cycle, during which mud is able to deposit.
In the interval from A to B mud deposition increases but it is intermittent during the tidal cycle. All points are characterized by constant deposition of sand. From B to C, mud deposition duration is still increasing. The increased steepness is determined by the fact that, in this area, sand is not always present, favoring a relative increase in mud content. In the interval from C to D mud deposition is constant and the presence of sand is at its minimum, as it is the sediment fraction determining the formation of layers. In this case both mud concentration and mud difference between different layers is at its maximum. Therefore maximum $\Delta c$ occurs where lamination is determined by sand rather than by mud variability. From Figure 2.10 it is possible to note an increase in mud content per layer with increasing longitudinal coordinate. A reduction in the laminae area for small tidal amplitudes is also evident from Figure 2.10b.

### 2.8 Lamination Characteristics

By taking into account numerical and analytical model results, bedding characteristics along the centerline have been defined using the lamination length ($l_l$), defined as the total length where tidal laminations can form, and the maximum difference in mud content ($\Delta c$) between two successive layers (Figure 2.4). The latter difference is the average of different $\Delta c$ values along the mouth bar centerline (Figure 2.4). Parameters $\Delta c$ and $l_l$ have been calculated from our numerical tests for values of $h_u/h_l$ equal to 0.4. To understand how sediment characteristics and hydrodynamic conditions affect $\Delta c$ and $l_l$, we use dimensional analysis and the Buckingham’s ($\Pi$-) Theorem. Assuming constant values of erosion parameters and critical shear velocity for the two sediment fractions, it
follows that both the location of zero net deposition \( \xi \) and the sediment concentration in the water column \( C(\xi) \) (Equation 2.3 and 2.9) at different instant of the tidal cycle.

![River mouth bar longitudinal sections for different Darcy-Weisbach friction coefficients (f), sediment concentrations and mud settling velocities. Red color means one hundred percent of mud, blue color means zero percent of mud, only sand. A) \( f = 0.04, c_m = 1.2 \text{ kg/m}^3, w_m = 0.001, h_t = 2.5 \text{ m} \). B) \( f = 0.04, c_m = 1.2 \text{ kg/m}^3, w_m = 0.001, h_t = 0.5 \text{ m} \). C) \( f = 0.09, c_m = 0.6 \text{ kg/m}^3, w_m = 0.001, h_t = 2.5 \text{ m} \). D) \( f = 0.04, c_m = 1.2 \text{ kg/m}^3, w_m = 0.00025, h_t = 2.5 \text{ m} \). E) \( f = 0.04, c_m = 1.2 \text{ kg/m}^3, w_m = 0.0001, h_t = 2.5 \text{ m} \). F) \( f = 0.04, c_m = 0.8 \text{ kg/m}^3, w_m = 0.0001, h_t = 2.5 \text{ m} \). No Tides.](image)
depend on $f, b_0, h_0, u_0, w, c_0$ and $h_t$, where $w$ and $c_0$ are the settling velocity and concentration at the river mouth of a certain sediment fraction. Considering the above independent variables for both mud and sand, and applying the II-theorem, it is possible to obtain two functional relationships for $\frac{\Delta c}{c_m}$ and $\frac{l_l}{b_0}$ as a function of the following non-dimensional parameters: $f, b_0, h_t, \frac{h_m}{h_0}, \frac{h_s}{h_0}, \frac{w_m}{w_0}, \frac{w_s}{w_0}, \frac{c_m}{c_s}, \frac{w_m}{w_s}$, where $\bar{u}_0$ is the riverine velocity at mean sea level, $c_m$ and $c_s$ are the mud and sand concentrations in the river, $b_0$ is the river mouth half-width, $h_0$ is the bottom depth for mean sea level, $w_m$ and $w_s$ are mud and sand settling velocity respectively. By assuming power-law relationships and by means of a multiple regression analysis we obtain the empirical expressions:

\[
\frac{\Delta c}{c_m} = 1.86 f^{0.3} b_0^{0.2} h_t^{0.4} \frac{h_m}{h_0} \frac{c_m}{c_s}^{0.01} \frac{w_m}{w_s}^{0.24} \frac{w_m}{w_0}^{0.27}
\]

\[
\frac{l_l}{b_0} = 0.79 f^{-0.2} b_0^{-0.2} h_t^{0.2} \frac{h_m}{h_0} \frac{c_m}{c_s}^{-0.04} \frac{w_m}{w_s}^{-0.01} \frac{w_m}{w_0}^{-0.11}
\]

(Equation 2.12)

Figure 2.11 shows comparisons between measured values of $\frac{\Delta c}{c_m}$ and $\frac{l_l}{b_0}$ and values predicted from the two empirical expressions (12).

If mud concentration, $c_m$, is the only varying parameter, its increase would lead to a reduction in lamination length and to an increase in mud deposition near the river mouth. A higher sand concentration, $c_s$, would lead to a larger area where both sediments are available and, therefore, an increase of lamination length. For the extreme case in which sand concentration is much higher than mud concentration and this ratio goes to zero, the
potential extent of lamination is at its maximum, but the magnitude of $\Delta c/c_m$ is minimum due to lower amounts of available mud with respect to sand. The ratio $\frac{w_m}{w_s}$ represents mud settling velocity over sand settling velocity and it is always less than one. If this value approximates zero, mud behaves as a conservative substance and maximum lamination length occurs together with minimum concentration difference between layers, as mud tends to be transported downstream and only a small portion is allowed to deposit. The term $\frac{w_m}{u_0}$ regulates the interaction between mud and the riverine flow. High values of $\frac{w_m}{u_0}$ can be caused by a decrease in riverine velocity or an increase in settling velocity. Both cases lead to an accelerated deposition of the finer sediment fraction near the river mouth and reduce the susceptibility of this fraction to shear stress variations. An increase in $\frac{w_m}{u_0}$ leads to a reduction in lamination length and to an increase in the $\Delta c/c_m$ ratio, due to higher differences between positive net deposition at low and high velocity. These enhanced differences are caused by large variations in longitudinal erosion near the river mouth (erosion curve in Figure 2.7). The latter mechanism also leads to large $\Delta c/c_m$ for large values of $\frac{w_m}{w_s}$ and $\frac{c_m}{c_s}$. The term $\frac{h_t}{h_0}$ is the ratio between tidal amplitude and mean sea level. For high values of $\frac{h_t}{h_0}$ we have an increase in both $\Delta c/c_m$ and $\frac{l_t}{h_0}$ ratios due to large velocity oscillations within the tidal cycle. Large velocity variations expand the area of intermittent deposition, and guarantee alternate deposition of finer and coarser material.
The friction coefficient \( f \) and the width to depth ratio \( \left( \frac{b_0}{h_0} \right) \) of the river mouth determine the dimensionless parameter \( \mu \) (Eq. 1). This friction parameter regulates the decay of both concentration and velocity. Its fast decay results in a decrease in lamination extent and an increase in landward deposition rate for both sediment fractions. On the contrary, \( \Delta c / c_m \) increases with friction due to the sharp drop in erosion along the centerline that promotes, in turn, differences in positive net deposition during the tidal cycle (dotted areas in Figure 2.7).

An analysis of the exponents of each term in Eq. 11 reveals that bottom friction and river mouth geometry have a major role in determining both lamination length and difference in mud content between successive layers. This is mainly due to the influence of bottom friction on hydrodynamic conditions of the flow, which then affect deposition and erosion of both sediment fractions. The exponents are also large, in absolute value, for the non-dimensional tidal range \( \frac{h_t}{h_0} \) and the non-dimensional river velocity \( \frac{w_m}{u_0} \). Exponents of \( \frac{c_m}{c_s} \) and \( \frac{w_m}{w_s} \) are instead smaller, suggesting that the relative concentration and settling velocity have relatively limited effect on the length of the lamination area and difference in concentration between consecutive layers.

2.9 Conclusion

Making confident predictions of geomorphological and stratigraphic features of mouth bars is relevant to understand the rock record of depositional basins. Mouth bars often display alternate layers of coarse and fine material, due to velocity variations throughout
tidal cycles. In our analysis we consider two sediment fractions, mud and sand, and we propose that tidal bedding occurs in areas where alternate deposition and erosion occur, for at least one of the two fractions. We further propose a facies model such that a sand deposit forms near the river mouth followed in order by a lamination zone, a homogeneous sand-mud mixture area and mud deposits. Lamination and sand mud-mixtures form at intermediate distances from the river mouth, while mud deposits are created in the distal parts of the domain. The lamination area increases with increasing tidal amplitude. Tidal bedding properties are defined by means of bedding extension.
along the centerline and differences in mud content between successive layers. Both quantities have been found to increase with tidal amplitude. Lamination extension grows with decreasing ratios of mud over sand concentration and settling velocity, while the difference in mud content in successive layers has been found to increase with increasing settling velocity and concentration. Differences in mud content between different layers tend to increase far from the river mouth. The lamination area also tends to shift basinward during mouth bar evolution. Finally, bottom friction has been found to be one of the main drivers for lamination. According to our dimensional analysis, an increase in friction reduces lamination length and increases the difference in mud content between successive layers, by means of flow deceleration and early sediment deposition. While this work focuses on tidal laminae, the results could be extended to a broader suite of processes. The same theory could be used to explain seasonal scale processes, such as fluvial discharge variations, with consequent application to longer cycles common in natural systems. The broader conclusion could apply to a variety of processes in that an unsteady flow field with bimodal sediment distribution can result in discrete laminations.

2.10 Appendix

According to the analytical solution [Özsoy and Ünlüata, 1982], vertical flux of material is expressed as:

$$\bar{Q}(\xi) = -w m c \left(1 - \frac{u^2}{u_{cr}^2}\right)$$
The source term of Equation 2.12 is the sum of potential deposition \((w_m c \frac{u^2}{u_{cr}})\) and erosion \((w_m c \frac{u^2}{u_{cr}})\). To substitute the above expression for erosion with the widely accepted Partheniaes-Krone formulation (Equation 2.8), erosion values around the river mouth are made equivalent by substituting \(u_{cr}^2\) with \(\hat{u}_{cr}^2 = w_c 0 \frac{u_0^2}{M(u_{cr}^2 - 1)}\) for \(u(\xi)^2 > 1\).

Considering the extension of the lamination area \((l_l)\) and the maximum difference in mud content between successive layers \((\Delta c)\) as dependent on friction, concentration and settling velocities of sand and mud, channel geometry, tidal amplitude and channel velocity, two functional relationships \((F1\ and\ F2)\) exist such that:

\[
F1(\Delta c, f, b_0, h_0, u_0, h_t, w_m, c_m, c_s, w_s) = 0
\]

\[
F2(l_l, f, b_0, h_0, u_0, h_t, w_m, c_m, c_s, w_s) = 0
\]

(Equation 2.14)

Following the \(II\)-theorem, these expressions can be rewritten utilizing seven non-dimensional variables. Assuming that the relationship among them is a power law, we obtain:

\[
\frac{\Delta c}{c_m} = \theta_c \frac{\alpha_c}{b_0} \frac{\beta_c}{h_0} \frac{\gamma_c}{h_t} \frac{\delta_c}{c_m} w_m \frac{\epsilon_c}{w_s} \frac{k_c}{\hat{u}_0}
\]
\[ \frac{l_t}{b_0} = \vartheta_l f^\alpha_l \frac{b_0}{h_0} h_t^{\gamma_l} \frac{c_m}{c_s} \frac{\delta_l w_m}{w_s} \frac{\epsilon_l w_m}{u_0} \]  

(Equation 2.15)

Where: \( \vartheta_c = 1.86; \ \alpha_c = 0.3; \ \beta_c = 0.2; \ \gamma_c = 0.4; \ \delta_c = 0.01; \ \epsilon_c = 0.24; \ \kappa_c = 0.27 \) and

\[ \vartheta_l = 0.79; \ \alpha_l = -0.2; \ \beta_l = -0.2; \ \gamma_l = 0.2; \ \delta_l = -0.04; \ \epsilon_l = -0.01; \ \kappa_l = -0.11 \]

The table below reports the incremental change in the data variance (\( \sigma^2 \)) with the removal of each term of Equation 2.14. The first column corresponds to Equations 2.15 variance; after that, each column corresponds to the variance of Equations 2.15 after removal of the term in the corresponding first row.

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<td>0.17</td>
<td>0.18</td>
<td>0.33</td>
</tr>
</tbody>
</table>

(Table 2.1)
2.11 References


White, C. D., B. J. Willis, S. P. Dutton, J. P. Bhattacharya, and K. Narayanan

Sedimentology, statistics, and flow behavior for a tide-influenced deltaic sandstone, Frontier Formation, Wyoming, United States In: G.M. Grammer and P.M. Harris, and G.P. Eberli (eds.) Integration of outcrop and modern analogs in reservoir modeling: AAPG Memoir 80, p. 129 – 152


CHAPTER 3. How waves shape salt marshes

The content of this chapter were published in 2014 in Geology. This paper was co-authored by Sergio Fagherazzi (Department of Earth and Environment, Boston University).

Abstract

We present high resolution field measurements of five sites along the United States Atlantic Coast and cellular automata simulations to investigate the erosion of marsh boundaries by wave action. According to our analysis, when salt marshes are exposed to high wave energy conditions their boundary erodes uniformly. The resulting erosion events follow a Gaussian distribution, yielding a relatively smooth shoreline. On the contrary, when wind waves are weak and the local marsh resistance gains importance, jagged marsh boundaries form. In this case, erosion episodes have a long-tailed frequency magnitude distribution with numerous low magnitude events but also high magnitude episodes. The logarithmic frequency magnitude distribution suggests the emergence of a critical state for marsh boundaries, which would make the prediction of failure events impossible. Internal physical processes allowing salt marshes to reach this critical state are geotechnical and related to the non-homogeneity of salt marshes whose material discontinuities act as stress raisers.
3.1 Introduction

Large salt marsh losses have been documented all around the United States, Asia and Europe [e.g. Fagherazzi, 2013; Kirwan et al., 2010; Castillo et al., 2000].

If salt marshes continue declining, we risk losing their valuable ecosystem services. Among others, salt marshes mitigate the impact of hurricane and tsunami, provide habitats for a variety of aquatic animal species and mediate the exchange of sediments and contaminant between the marine and terrestrial environment [e.g. Fagherazzi et al., 2013].

Salt marshes are very resilient with respect to vertical dynamics because feedbacks among inundation, organic matter accretion, plant growth, and sediment deposition allow the marsh to keep pace with sea level rise [e.g. Fagherazzi et al., 2012; Kirwan et al., 2010]. On the contrary, recent results indicate that salt marshes are inherently weak with respect to horizontal erosion [Fagherazzi et al. 2013]. Specifically, waves forming in large coastal bays can trigger irreversible salt marsh deterioration even in the absence of sea level rise [Mariotti and Fagherazzi 2013]. As a result, saltmarshes do not display lateral equilibrium but are always contracting or expanding at rates of meters per year [Fagherazzi 2013].

Understanding mechanisms controlling salt marsh erosion is, thus, of crucial importance for the correct management and preservation of coastal environments. Although salt marsh dynamics has been widely investigated, many processes are still poorly understood. Existing studies adopt process based or empirical models to estimate the location and size of erosion events. An alternative conceptual model for marsh erosion
can be based on simple stochastic models that could help extracting universal features of the processes at play. Simple models, having few rules governing the interaction among individual components, can lead to the emergence of complex systems, displaying “structures with variability” [Goldenfeld and Kedanoff, 1999; Murray 2007, Fagherazzi 2008].

Herein we present a cellular automata model and high resolution field measurements of marsh boundaries to explain erosional trends under different wind-wave exposures. Wind waves erosion is arguably the main mechanism controlling marsh edge retreat, as confirmed by both numerical and empirical investigations [e.g. Mariotti and Fagherazzi 2013; Fagherazzi et al., 2013; Marani et al., 2011]. According to our results, salt marshes that are highly exposed to wind wave power are retreating uniformly. On the contrary, low wave power conditions correspond to long tailed distribution of erosion events, which create rougher marsh fronts.

3.2 Study sites

We focus on five sites located in Plum Island Sound, Massachusetts, USA, and Virginia Coast Reserve, Virginia, USA (Figure 3.1). The Plum Island Sound Estuary is a coastal-plain estuary characterized by extensive marshes. Tides are semidiurnal and the average tidal range is 2.9 m. Tidal prism has been estimated as 32 Mm$^3$. Prevailing winds come from the westerly quadrant; winds having greatest frequency are from the North, while winds with the greatest duration are from Southwest [Pinot, 1965]. The site Refuge North (RN) is the most sheltered from wave action and the slowest eroding one, followed in
order by Stackyard Road (SR), and Refuge South (RS). The Virginia Coast Reserve encompasses more than 100 Km of dynamic system of barrier island, shallow lagoons and salt marshes separated by deep tidal inlets. Tides are semidiurnal with a mean tidal range of 1.2 m. Mean, high, and low water levels respect to mean sea level are 0.68 and -0.70 m. Prevailing winds come from the South-West and North-East quadrants. North-East and South West directions, parallel to the barrier island, also correspond to the highest fetch values [Fagherazzi and Wiberg, 2009]. The two locations taken into account are Chimney Pole (CP) and Hog Island (HI). Field data [Mc. Loughlin et al., 2014] and model results [Mariotti et al., 2010] indicate CP as the most susceptible to wind waves erosion. Plum Island Sound and Virginia Coast Reserve salt marshes are both characterized by prominent scarps at their seaward edges, typically 1.5 m or more above adjacent tidal flats. Overhanging marsh profiles are typical of wind wave eroding marsh boundaries, while gently sloping edges indicate stable configurations [e.g. Fagherazzi et al., 2013].
Figure 3.1 Study areas. A) Plum Island Sound, Massachusetts, USA and Virginia Coast Reserve, Virginia, USA. B) Study sites in Plum Island Sound. From the North: Refuge North (RN, 42°45'00 N 70° 48’ 00 W), Stackyard Road (SR, 42°44’00 N 70° 49’ 00W) and Refuge South (RS, 42°44’00 N 70° 48’ 00W). C) Study sites in Virginia, Chimney Pole (CP, 37° 27’ 00 N 75° 42’ 00 W) and Hog Island (HI, 37°23’ 00 N 75°42’ 00 W)

3.3 Methods

The stochastic model consists of a 2D square lattice (Figure 3.2) whose elements $i$ have randomly distributed resistance $r_i$, comprised between 0 and 1. Each cell has erosion rate $E_i$:

$$E_i = \alpha P^\beta \exp \left( -\frac{r_i}{P} \right)$$

(Equation 3.1)

Where $P$ is the wave power and $\alpha$ and $\beta$ are constants equal to 0.35 and 1.1 respectively [Schwimmer et al., 2001]. The first part of Equation 3.1 is in agreement with classical theoretical and empirical investigations on salt marsh boundary erosion. According to
these, the retreat rate is proportional to wave power and follows a power law relationship, having exponent close to one [e.g. Schwimmer et al., 2001; Marani et al., 2011]. The second part of Eq. 1 is meant to take into account the variety of biological and geomorphologic processes affecting each portion of the marsh. Among others, seepage erosion, crab burrowing, vegetation and sediment cohesion make difficult to predict which portion of the marsh will collapse first. Equation 1 is such that the local variability of marsh resistance is particularly relevant when the wave power is low (in this case, for example, the presence of vegetation could actively prevent the failure of a certain marsh portion but only for small waves). On the other hand, when wind waves power is very high (for example during storms), local marsh characteristics play a secondary role and different marsh elements are eroded at a similar rate, as their resistance is small compared to the main external driver. In this case the exponential in Equation 3.1 goes to one and every cell has the same erosion rate. At each time step, only neighbors of previously eroded cells are susceptible to erosion (herein, being neighbors means having one side in common, Figure 3.2). Each neighbor is eroded with probability \( p_i = \frac{E_i}{\sum E_i} \), where the sum refers to all cells that can potentially be eroded for a given time step. A cell is also automatically eroded if it remains isolated from the rest of the domain (Figure 3.2 c, crossed cell). In fact isolated cells would represent isolated marsh stacks that are disintegrated fast as they are attacked by waves from several directions. A similar model has already been adopted, in a different context, for the chemical etching of disordered solids [Kolwankar et al., 2002]. However, our model uses a different formulation for the erosion rate \( E_i \), and the sum at the denominator of the erosion probability \( p_i \) only runs
over the boundary elements, rather than over the entire domain. Moreover, we automatically remove isolated cells.

Marsh contours have been tracked using a Real-Time Kinematic-Global Positioning System (RTK-GPS) and an electronic Total Station. Data were collected with an average resolution of 1 m. When marsh contours were characterized by significant variations in boundary geometry, measurements were taken up to 20 cm apart. Marsh boundaries have been monitored every September from 2008 to 2013 for the three sites in Plum Island Sound. For the two sites in the Virginia Coast Reserve, measurements have been taken on March 2008 and August 2010. We define as magnitude of an erosion event, for a given year and for a certain point along the marsh boundary, the shortest distance of that point from the marsh boundary of the subsequent year. We define as marsh boundary sinuosity the ratio between the boundary length and the straight line distance between the two boundary extremities.

### 3.4 Model Results

Figure 3.3 shows results for two simulations run for small (0.1 W) and high (20 W) wave power P. Simulations have been run on a grid 250 x 250 cells and have been stopped after removal of half of the domain cells. As it is possible to notice, in case of high wave power, marsh erosion proceeds uniformly along the marsh shoreline and generates a profile which is rough, at the scale of the single cell, but smooth at a larger scale. This is because each cell has similar resistance if compared to the main external driver. On the contrary, low wave power conditions correspond to the development of a jagged
Figure 3.2 Sketch of the cellular automata model. A) Possible domain configuration after removal of two domain elements. Each cell has erosion rate $E_i$; grey cells represent erodible elements and have erosion probability $p_i$. B) Possible domain configuration after removal of two other cells. C) The crossed cell remains isolated from the rest of the boundary and it is thus automatically removed.

Boundary. Indentations are produced by different erosion rates of individual cells which affect the global system behavior.

From a statistical view point, the system behaves differently for the two extreme conditions of very low and very high wave power. The frequency distribution of erosion events in a time
Figure 3.3 Results of the stochastic model: marsh boundary under low and high wave power conditions. Simulations have been run on a 250x250 cells grid and have been stopped after removal of half of the domain cells.

interval $\Delta t$ approaches a Gaussian distribution in the case of high wave power (Figure 3.4a). On the contrary, in the case of low wave power, the frequency distribution is characterized by a long tailed power law distribution (Figure 3.4b). For the low wave power case, a long time is required to erode very resistant cells. However, once the most resistant cells are eliminated, several weak sites remain exposed and can be rapidly removed with consequent generation of large scale failures.
Figure 3.4 Frequency magnitude distributions of erosion events for low (A) and high (B) wave power conditions. \( n \) is the number of eroded cells. \( N(n) \) is the number of times \( n \) cells are eroded within a time interval \( \Delta t \). Red points are model results. Blue lines are the interpolation of model results using a logarithmic (A) and a Gaussian (B) frequency magnitude distribution.

Similar results have been shown by Kolwankar et al. [2002] who demonstrate that, when the etching power of the solution approaches zero, their model is identical to classical invasion percolation, with reaction rate limited by the invasion percolation threshold [e.g. Wilkinson and Willemsen, 1983; Roux and Guyon, 1989; Desolneux et al., 2004].

Despite of the occurrence of large scale events, in the case of low wave power exposure, the remaining domain cells are very resistant due to the selected removal of the weakest sites. This differential removal is allowed by resistance variability among different cells. For a given low wave power condition, if cells resistance \( (r_i) \) values maintain the same mean but their range of variability \( (\Delta r_i) \) is reduced (e.g. \( r_i \in [0.3; 0.7] \) instead of \( r_i \in [0; 1] \)), the domain starts eroding uniformly as for the high wave power condition. This uniform erosion rate leads to an accelerated erosion (Figure 3.5) and is thus unfavorable to the maintenance of the domain elements.
Figure 3.5. Time ($\tau_i$) required to erode half of the domain particles as a function of $\Delta r_i$ and for a wave power $P$ equal to 0.1 W. $\Delta r_i$ is the amplitude of the range within which the cells resistance $r_i$ can vary. Resistance ranges have different amplitudes but the same mean, which is equal to 0.5. For example, $\Delta r_i$ equal to 0.3 means that $r_i$ values span from 0.35 to 0.65. The time ($\tau_{ii}$) required to erode half of the domain cells for a certain $\Delta r_i$ range has been normalized by the time $\tau_1$ correspond to a $\Delta r_i$ equal to 1 (i.e. $r_i \in [0; 1]$).

3.5 Field data analysis

In Plum Island Sound average erosion rates for the period of record (September 2008-September 2013) are: 0.2 m/year at RN; 0.35 m/year at SR; 0.45 m/year at RS. For the two sites in the Virginia Coast reserve, average erosion rates (March 2008-August 2010) are 0.75m/year at HI; 1 m/year at CP. The frequency magnitude distribution of biennial erosion events for each point along marsh shorelines is presented in Figure 5. The most sheltered sites in Plum Island Sound (Refuge North, RN and Stackyard Road, SR) have a logarithmic frequency magnitude distribution. Moreover, the lower is the exposure to
waves the longer is the tail of the power law (Figure 3.6, slope coefficient for RN is lower than the one for SR). For site RS, the power law distribution starts getting closer to the Gaussian distribution and an intermediate condition arises, characterized by a shorter tail and maximum number of erosion events not corresponding to minimum magnitudes anymore.

Figure 3.6. Frequency-magnitude distributions of erosion events for three Sites in Plum Island Sound and two sites in the Virginia Coast Reserve. From left to right: Refuge North, Stackyard Road, Refuge South, Hog Island, Chimney Pole. $n$ is the erosion event (m) occurring $N(n)$ times.

For the two sites along the Virginia Coast reserve, Chimney Pole is the most exposed and its erosion events follow a Gaussian distribution. The frequency magnitude distribution of erosion events in Hog Island is intermediate between the Gaussian and the logarithmic distribution. Thus, in both Plum Island and Virginia Coast Reserve, the lower is the site exposure to wave action, the longer is the power-law tail of the erosion events distribution. The frequency magnitude distribution of erosion events is also clearly recognizable from marsh boundary profiles (Figure 3.7).
Figure 3.7. shows marsh boundary profile for two of the five sites taken into account (SR and CP). CP seems maintaining the same profile while gradually retreating. On the contrary, the SR boundary is more jagged and erosion events of different size happen along its contour. The sinuosity of the marsh boundary at Chimney Pole is equal to 1.17 in 2008 and 1.19 in 2010. At Stackyard Road, the sinuosity of the marsh boundary is equal to 1.65 in 2008 and 1.75 in 2010.

3.6 Discussion and conclusions

Our simple model appears to capture important marsh boundary features and to give some new insight into salt marsh erosional processes. High resolution field measurements at five locations along the United State Atlantic Coast confirm numerical results. Our investigations have been related to high or low wave power exposure as well as to weaker or more resistant marsh platforms (given a fixed wind wave exposure). High exposed sites are characterized by uniform rate of marsh retreat along the shoreline, with erosion events following a Gaussian frequency magnitude distribution. On the contrary, less exposed sites show a long tailed frequency magnitude distribution with numerous small events and few (but not negligible) bigger events, which are unpredictable and can happen despite of a reduced wave action. The fact that sites in Plum Island have longer
tailed distributions than the study sites in Virginia can be related to the reduced likelihood of marsh slumping in microtidal environments with respect to macrotidal marshes. For the latter case, higher marsh scarps promote undercutting and tensional break development [e.g. Schwimmer, 2001].

We retain that, in the case of low wave exposure, the system could reach self-organized criticality (SOC). Self-organization refers to the ability of certain non-equilibrium systems to develop structures and patterns in the absence of any fine tuning from external agents. Criticality refers to the fact that all the members of the system influence each other and that local instabilities generate broader-scale order disturbances [e.g. Bak, 1987; Anderson, 1996; Fonstad and Marcus, 2003].

The power law relationship is a necessary condition for self-organized criticality and it is frequently used to test whether SOC is present. According to Bak [1987, 1989], if frequency data fit a power law distribution over a range of event magnitudes, the system is likely self-organized and could be at a critical state. However, the power law relationship is a necessary but not sufficient condition for self-organized criticality as it could also happen in a range of non-SOC systems. Other necessary conditions (but not sufficient, as sufficient conditions are unknown) for the development of SOC are: i) existence of a quasi-steady critical state at which the system self-organizes; ii) an internal mechanism by which the system can reach this critical state; iii) the response of the system to perturbations varies in magnitude independently of perturbations size; iv) presence of mechanisms for the system energy dissipation; v) presence of many degrees of freedom within which internal processes can operate.
Conditions iv) and v) are easily verified as saltmarshes are inherently dissipative systems characterized by a continuous loss in both potential and kinetic wave energy, consequent sediment removal and further energy dissipation. Condition iii) is confirmed by the logarithmic distribution of erosion events which spans over multiple length scales. For conditions i) and ii) we propose that the critical state for marsh boundaries is the one promoting the removal of weak sites and consequent exposure of more resistant and uniform marsh portions. The critical state would, thus, be the one maximizing salt marsh resistance to wave action. In our simplified model this condition corresponds to the contour approaching the percolation cluster made of the slowest eroding sites and surrounded by easily erodible ones. Field data confirm this assumption, considering that the slope of the logarithmic distribution of the less exposed sites (RN) is close to 1.53 and thus in agreement with classical invasion percolation problems [e.g. Kolwankar et al., 2003; Desolneux et al., 2003]. Finally, internal processes allowing the system to reach its critical state are geotechnical mechanisms connected to system discontinuities. Discontinuities enhance wave stresses and lead to cracks development. In the presence of cracks, the system approaches the minimum energy state independently from external agents [e.g. Roylance 2001]. Failures of marsh portions became thus possible and independent from any fine external tuning. In our simplified model these discontinuities correspond to contact area between clusters having relatively high or low erosional rates. In a natural system discontinuities could correspond to contact surfaces between marsh blocks weakened by groundwater seepage. Stress concentration along these discontinuities favors the failure of weak “marsh clusters” of different sizes. For example
marsh slumping, triggered by cracks and seepage, can lead to large scale events which in turn promote long tailed distribution. Once the weakest sites are removed, more resistant marsh portions are uncovered, which are difficult to erode. Thus, variability in marsh properties allows marsh boundaries to be “armed” against wind wave action by means of selected removal of weak elements.
3.7 References


Wilkinson, D., and Willemsen, J. F., 1983, Invasion Percolation - a New Form of
Percolation Theory: Journal of Physics a-Mathematical and General, v. 16, no. 14,
p. 3365-3376.
CHAPTER 4. Effect of local variability in erosional resistance on large-scale morphodynamic response of salt marshes to wind waves and extreme events

The content of this chapter were published in 2015 in Journal of Geophysical Research Letters. This paper was co-authored by Sergio Fagherazzi (Department of Earth and Environment, Boston University).

Abstract

The presence of natural heterogeneities is an integral characteristic of salt marshes and needs to be account for, as local feedbacks could influence the large scale morphodynamic evolution of these wetlands. Herein, we use field data and a cellular automata model to investigate salt marsh response to wave action under different wave energy conditions and frequency of extreme events. Our results suggest that salt marsh response to wind waves is tied to their local properties. In case of low wave-energy conditions, a local variability in marsh resistance might lead to the unpredictable failure of large marsh portions with respect to average erosion rates. High wave-energy conditions, while overall lead to faster marsh deterioration, produce constant and predictable erosion rates. A high occurrence of extreme events causes salt marshes to reach the highest likelihood of large failures for a lower wave exposure, while also leading to smoother, and more uniformly deteriorating marsh boundary profiles. Salt marshes subject to weak wave energy conditions are the most susceptible to variations in the frequency of extreme events. On the contrary, salt marshes exposed to energetic
waves remain relatively unchanged in front of such frequency variations. This suggests that variations in time in the morphology of salt marsh boundaries could be used to infer changes in frequency and magnitude of external agents.

**4.1 Introduction**

The successful outcome of coastal restoration strategies relies upon understanding of the response of salt marshes to both natural and anthropogenic forcing. Under a climate change scenario, increasing attention has been drawn to these wetlands, as their role as natural buffers against violent storms and hurricanes could be crucial for protecting coastal communities. One of the main concerns about the resilience of salt marshes to climate change is their response to a possible increase in the frequency of extreme events, such as extra-tropical storms and hurricanes. As a matter of fact, evidence from observations show that climatic extremes have changed [e.g. Easterling et al., 2000; Goldenberg et al., 2001; Webster et al., 2005], and losses caused by catastrophes in the United States (which for the insurance industry are storms causing damaging for more than $5 million), have already been steadily growing in the past few decades [e.g. Easterling et al., 2000].

Located at the delicate interface between marine and terrestrial environments, salt marshes dissipate wave energy [e.g. Chen and Zhao, 2011; Moller et al., 2014], while also providing a unique habitat for many floral and faunal species, filtering nutrients and sediments from the water column, and storing large amounts of carbon over decennial and millennial timescales [e.g. Zedler et al., 2005]. Despite their valuable services, salt
marshes have been lost worldwide at increasing rates [e.g. Fagherazzi, 2013; Marani et al., 2011; D’Alpaos, 2011]. Salt marshes have proven to be very resilient to relative sea level rise [e.g. Fagherazzi et al., 2013], but they appear unable to maintain their horizontal extent at the geological timescale, and are extremely weak with respect to wave action [e.g. Schwimmer and Pizzuto, 2000; Marani et al., 2011; Mariotti and Fagherazzi, 2013; Ganju et al., 2013; Bendoni et al., 2014; Leonardi et al., 2014].

The complicated nature of marsh erosion is exacerbated by the variety of biological and ecological processes acting along marsh boundaries, as well as by their spatial heterogeneity which needs to be accounted for. Numerous studies have successfully investigated some of these processes. Among others, spatial and temporal nonlinearity of coastal wetlands erosion have been linked to variability of the vegetation cover [e.g., Mullarney and Henderson, 2010; Temmerman et al., 2003; Van der Wal et al., 2008], crab burrowing [e.g. Xin et al., 2009; Bertness et al., 2014], and soil properties, with less dense, fine, and cohesive soils conferring higher resistance to the marsh platform [e.g. Feagin et al., 2009; Howes et al., 2010].

Considering the myriad processes at play, it is difficult to incorporate all of them in numerical models, while also maintaining reasonable computational times and model generality. One of the challenges is thus to develop ways to resolve causality, while overcoming the limitations and specificity of scenario-based predictions [e.g. Coco et al., 2007; Larsen et al., 2014]. In this context, very simple models, with few stochastic rules have been found to be effective, and able to reproduce the large scale behavior of natural
systems [e.g. Murray and Paola, 2003; Fonstad et al., 2003; Bak et al., 1987; Goldenfeld and Kadanoff, 1999].

Herein we adopt an existing stochastic model with a minimum level of detail to reproduce marsh erosion by waves [Leonardi and Fagherazzi, 2014]. Despite its simplicity, the model accurately predicts important marsh boundary dynamics, and is in agreement with long-term field data measurements. The model encompasses two marsh boundary features: the existence of an external agent acting as erosive process (wave power), and the presence of small scales heterogeneities within the system.

In the first part of this work we verify the model accuracy by means of long-term detailed measurements of marsh boundary erosion, wave power, and soil shear strength at three sites in Plum Island Sound, Massachusetts, USA. In the second part of this work, we use the model to explore salt marsh response to increasing wave exposure and to the occurrence of increasingly frequent, and randomly distributed extreme events. We also focus on differences between erosion rates of homogeneous marsh boundary portions, with respect to the erosion rate of the system as a whole.

4.2 Methods

4.2.1 Field Data

We collected field data at three marsh sites in Plum Island Sound, Massachusetts, USA (Figure 4.1). Plum Island Sound is a shallow meso-tidal estuary, with mean tidal range of 2.6 m. Due to the sheltering effect of the estuary, wind waves are locally generated, and of short period (2 to 3 seconds). Dominant wind directions are from west and northwest,
and the highest wind waves are generally created by northeast winds (Nor’easters) [e.g. Fagherazzi et al., 2014]. The three sites taken into account are Refuge North, Stackyard Road, and Refuge South (Figure 4.1, white circles).

![Map of Plum Island sound showing field sites](image)

**Figure 4.1.** A) Field sites in Plum Island sound, Massachusetts USA: Stackyard Road (42°44’00 N 70° 49’ 00 W), Refuge North (42°45’00 N 70° 48’ 00 W), and Refuge South (42°44’00 N 70° 48’ 00 W).

We tracked the marsh contour at the three sites, for stretches of shoreline extending around 100 meters alongshore, every year, from 2008 to 2013, using a real time kinematic global positioning system and an electronic total station. Marsh contours were then used to calculate the magnitude of the erosion events at every point along the marsh shoreline.
We further used a vane shear test to measure soil strength at each location, and over regular spatial intervals along the marsh shoreline (Figure 4.2, Table 4.1). In-situ measurements of soil strength have the advantage of partially taking into account vegetation presence, and density. In fact, it is generally recognized that plants increase resistance to shearing, act as sticking agents, increase soil roughness, and contribute to the overall cohesion of the soil by mobilizing their tensile strength [e.g. Baets et al., 2008; Temmerman et al., 2005].
Figure 4.2. Marsh boundary profile and soil strength measurements for the three study sites in Plum Island Sound. A) Marsh boundary at Stackyard Road (42°44’00 N 70° 49’00W). B) Marsh boundary at Refuge South (42°44’00 N 70° 48’00W). C) Marsh boundary at Refuge North (42°45’00 N 70° 48’00 W). Points along the marsh boundaries indicate locations were soil strength measurements were taken. D, E, F) Field measurements of soil shear strength collected along the marsh boundary at Stackyard Road, Refuge South, and Refuge North respectively. Pin locations are indicated in Figure A, B and C. On the vertical axis of each subplot there is the depth at which soil strength measurements were taken. Zero values correspond to the marsh surface. On the horizontal axis there is the value of soil shear strength (daPa). G) Normalized frequency-magnitude distribution of soil shear strength for the three sites in Plum Island Sound. On the horizontal axis there is the soil shear strength, on the vertical axis the frequency of occurrence of a certain shear strength value.
Table 4.1. numerical values of soil strength measurements. Pins location is indicated indicated in Figure 4.2.

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<td>0.63</td>
<td>200.00</td>
<td>0.57</td>
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</tr>
<tr>
<td>0.63</td>
<td>0.21</td>
<td>180.00</td>
<td>0.51</td>
<td>0.25</td>
<td>150.00</td>
<td>0.43</td>
<td>0.51</td>
<td>150.00</td>
<td>0.43</td>
</tr>
</tbody>
</table>

The table above shows the numerical values of soil strength measurements for different locations and elevation levels. The pins location is indicated as in Figure 4.2.
4.2.2 Wave power calculation

To calculate wave power for the period of interest, and for the three sites taken into account, we collected wind direction and wind speed data from a nearby station (NOAA station ID. IOSN3). Wave climate was then computed following the equations of Young and Verhagen [1996], and by taking into account fetch lengths and water depth at the three different sites (Figure 4.3).

Figure 4.3. Wave power calculated at the three field sites in Plum Island Sound, Massachusetts, USA: A) Refuge North (42°45’00 N 70° 48’ 00 W) B) Stackyard Road (42°44’00 N 70° 49’ 00W). C) Refuge South (42°44’00 N 70° 48’ 00W).

4.2.3 Cellular automata model

The model consists of a two-dimensional square lattice whose elements, \( i \), have randomly distributed resistance, \( r_i \). The critical soil height \( H_{cl} \) for boundary stability is calculated from soil shear strength values and is assumed as representative of soil resistance, as it is a convenient way to take into account general soil and ambient conditions. The erosion
rate of each cell, $E_i$, which represents the erosion of an homogeneous marsh portion, is defined as:

$$E_i = \alpha P \beta \exp \left(-\frac{H_e i}{H}\right)$$

(4.1)

Where $\alpha$ and $\beta$ are non-dimensional constants set equal to 0.35 and 1.1 respectively [Schwimmer, 2001; Leonardi and Fagherazzi, 2014], $P$ is the wave power, and $H$ is the mean wave height. The first part of Equation 4.1 follows classical theoretical and empirical investigations of salt marshes erosion. According to these, the retreat rate is proportional to wave power [e.g., Schwimmer, 2001; Marani et al., 2011]. The second part of Equation 4.1 is meant to take into account variability in soil resistance due to biological and ecological processes affecting each portion of the marsh. The erosion rate has two extreme limits, such that for low wave power values the system is highly disordered (each element $i$ has a different erosion rate), while for very high wave power, only a weak disorder is present (all elements have similar erosion rates).

A cell’s neighbors are elements having one side in common with the cell itself (Figure 4.4). The model follows three rules: i) only neighbors of previously eroded cells can be eroded. Therefore, only cells having at least one side in common with previously eroded elements are susceptible to erosion; ii) at every time step one element is eroded at random with probability $p_i = \frac{E_i}{\sum E_i}$; iii) A cell is removed from the domain if it remains isolated from the rest of the boundary (no neighbors).
Since erosion rates of individual sites are independent of each other, the time interval $\Delta t$ before a cell is eroded with probability one is

$$\Delta t = \frac{1}{\sum E_i'},$$

where the sum refers to cells that can be removed for a given time step (see supplementary material).

Model formulation, as well as its parametrization, is then based on the following considerations: i) under the action of an external agent, weaker areas generally erode more easily than resistant ones; ii) when wave forces are well above the local threshold for erosion everywhere, resistance heterogeneity is low, while when forces are in some places above and in some places below the local threshold for erosion, resistance heterogeneity is high. iii) Discontinuities in soil resistance can act as a stress raiser and lead to the development of cracks whose failure can happen independently from the action of waves (i.e. even when wave power is very low). As an example, in the absence of wind waves, the failure of a large marsh block takes place when the crack has reached a critical length beyond which the system can lower its energy if the crack is allowed to grow longer [e.g. Roylance, 2001; Terzaghi, 1996]. When the system is eroding slowly, a failure of this kind can be relatively large with respect to average erosion rates. On the contrary, it is more likely for a large failure to fall around average erosion rates, when the erosion proceeds rapidly.

An exponential function has been found to be the easiest and most analytically tractable way to represent a threshold-like dependence on soil shear strength and to reproduce differential erosion mechanisms typically found in salt marshes, and presented in the above paragraph.
The cellular automata model allows avoiding complex and time-consuming models based on partial differential equations by using abstracted rules to simulate realistic morphologies. The limitation of the model is that important processes and feedbacks are neglected. For example, waves shoaling, breaking or refraction are not taken into account. In the first part of the manuscript, the model is run using the wave power calculated at the three sites. In the model, cells resistance values are based on soil strength measurements conducted in the field. Critical heights values of the marsh boundary, $H_{cl}$, are randomly and normally distributed with the same mean and standard deviation than the ones measured in the field (see supplementary material). In the second part of the manuscript the model is run with a uniform wave power, interrupted by randomly distributed high intensity events, having magnitude much higher than the background value. The frequency of these extreme events has been varied between 0 and 20 percent.
Figure 4.4. Sketch of the cellular automata model. Each cell has an erosion rate $E_i$ equal to: $E_i = \alpha P^\beta \exp\left(-\frac{H_{ci}}{H}\right)$. Where $\alpha$ and $\beta$ are non-dimensional constants, $P$ is the wave power, and $H$ is the mean wave height. At every time step one element is eroded at random with probability $p_i = \frac{E_i}{\sum E_i}$ [Leonardi and Fagherazzi, 2014].

A) To initialize the simulation, a fictitious row of eroded cell is assumed (grey cells with central dot). Any cell belonging to the first row ($E_{11}, E_{12}, E_{13}, E_{14}, E_{15}$, see also yellow line) is susceptible to erosion because it has one side in common with previously eroded cells (at time step zero, these are the gray cells). B) At time step 1, one of the domain cells is eroded at random, with probability depending on its erosion rate. For example, in this sketch, the cell $E_{14}$ is eroded. It is more likely for the weakest element to be eroded. However, due to the fact that this is a random process another element could be eroded as well. All cells along the yellow line are susceptible to erosion ($E_{11}, E_{12}, E_{13}, E_{14}, E_{15}, E_{10}$). Cells $E_9$ is a newly exposed cell that can be eroded because it is a neighbor of $E_{14}$. C) Among all the cells along the yellow contour cell $E_8$ is eroded; now cells $E_8$ and $E_{10}$ can be eroded as well; D) Cell $E_{10}$ is eroded. Cells along the yellow contour can be eroded as well. E) Cell $E_{10}$ is eroded. Cell $E_{15}$ is now an isolated cell because it has no neighbors and is automatically removed from the domain. Isolated cells are isolated marsh stacks that are more susceptible to erosion because they are attacked by wind waves from several directions, and for this reason are more easily disintegrated.

4.3 Results and Discussions

In agreement with existing numerical models and field data [e.g. Fagherazzi et al., 2014], Refuge South is, on average, the most exposed to wind waves (average wave power $P_{avg} = 25 \, W$), followed in order by Stackyard Road ($P_{avg} = 13 \, W$), and Refuge North ($P_{avg} = 10 \, W$) (Figure 4.3).

Figure 4.5 show the normalized frequency-magnitude distribution of erosion events for the three sites in Plum Island Sound (red points), and corresponding model results (blue lines). In the model the magnitude of erosion events corresponds to the number of eroded cells, while for the field data the magnitude of erosion events has been evaluated using marsh boundary contours. Model results well agree with field data, and for increasing
wave exposure a frequency-magnitude distribution with a gradually shorter tail is predicted. Refuge North and Stackyard Road display a logarithmic frequency magnitude distribution of erosion events, while Refuge South, the most exposed site, approaches a frequency-magnitude distribution that is intermediate between a logarithmic and a Gaussian. In fact, the frequency-magnitude distribution of erosion events has two extreme limits: i) when the ratio between salt marsh resistance and wave power is very high (e.g. very low wave power or very high soil resistance), the frequency-magnitude distribution approaches a logarithmic distribution; ii) when the ratio between salt marsh resistance and wave power is low, the frequency-magnitude distribution of erosion events approaches a Gaussian distribution. In the first case, even if a long time is required to erode very resistant cells, when a group of weak sites remain uncovered large failure

Figure 4.5. On the horizontal axis, the magnitude of erosion events \( (n^*) \), normalized by the magnitude of the largest erosion event. On the vertical axis, how many times an erosion event of a given magnitude did happen \( (N(n))^* \), normalized by the maximum frequency of occurrence. Red points denote field data while blue lines are model results.
events may occur as weak sites can be eroded much more rapidly than resistant ones [Leonardi and Fagherazzi, 2014].

Figure 4.6 shows results for two sets of simulations run for high (Figure 4.6 A, B), and low (Figure 4.6 C, D) wave power conditions (low and high cell resistance to wave height ratios respectively), with and without the occurrence of extreme events.

In the absence of extreme events, and for high wave energy conditions (Figure 4.6 A), the model predicts uniform erosion rates, leading to a smooth marsh boundary profile. On the contrary, when the wave power is low (Figure 4.6 C), a jagged marsh boundary forms (see movies in supplementary material). The jagged marsh boundary profile is due to the presence of heterogeneities in marsh resistance (small scale disorder) that affect the large scale system. In the latter case, the simulated marsh contour approaches the exterior boundary of an invasion percolation problem [e.g. Willemsen et al., 1983; Gouyet et al., 1988].

Figure 4.6. Geometry of the marsh boundary after erosion of half of the model domain, in case of high (A, B), and low (C, D) wave energy conditions, with and without extreme events. Blue indicates water; green indicates marsh platform.
From a qualitative point of view, when the system is subject to the occurrence of randomly distributed extreme events, the salt marsh is characterized by more uniform erosion rates (Figure 4.6). According to this trend, when salt marshes are already subject to high wave energy conditions, their boundary profile remains relatively unchanged when extreme events are added in the simulation. On the contrary, when the system is subject to low wave-energy conditions, the presence of frequent extreme events creates noticeably more uniform marsh boundary profiles (figure 4.6 C, D). In fact, during an extreme event, the presence of high waves prevents the selected removal of weak elements, and reduces the possibility of large scale failures of weak marsh clusters.

To better understand this process, changes in the following variables have been analyzed:
erosion time (Figure 4.7 A), which is the time necessary to erode half of the domain particles; Incision length, which is the cross shore amplitude of the position of the distal eroded cells (Figure 4.6 C, 4.7 B); boundary length (Figure 4.7 C), representative of marsh roughness; size of marsh failures (Figure 4.7 D), which corresponds to the maximum number of eroded cells within a short time interval. These variables have a similar trend, and decrease with the ratio of cell-resistance to wave power, in agreement with fast deteriorating and uniform marsh boundary profiles. Moreover, the presence of randomly distributed erosion events accelerates salt marshes erosion (i.e. the time necessary to erode half the domain cells decreases, Figure 4A). However, given the same number of eroded cells, incision length, size of marsh failures, and boundary length decrease with increasing frequency of the extremes (Figure 4.7 B, C, D).
The reduction in incision length with wave energy and with the number of extreme events suggests the existence of a possible non-linear mechanism, connected to the occurrence of unexpected marsh failures of different size, and to the interplay between the erosion rates of homogeneous marsh units and their effect on the large scale system deterioration. Specifically, while increasing wave power causes salt marsh to deteriorate faster, it also leads to low occurrence of unpredictable large scale events (failure size and incision length decrease), which prevents salt marshes portions far from the seaward side of the domain to be suddenly exposed to waves.

Figure 4.7. Erosion time (A), Incision length (B), boundary length (C), and size of marsh failure (D) as a function of wave power, for different frequency of extreme events. Values on the vertical axis have been normalized to their maximum value.
To investigate this problem, we define the global to local erosion ratio as the ratio between the erosion velocity of the entire system (calculated as the incision length divided the erosion time), and the average erosion rate of the single cell. For a given wave power and thus, for an average erosion rate of isolate cells, $E_i$, this ratio represents the large scale response of the system with respect to the response of its small scale, homogeneous units. It is important to notice that this metric of salt marsh vulnerability is relative to a specific wave energy, as salt marshes generally erode faster for increasing wave exposure. Large values of the global to local erosion ratio do not necessarily correspond to high rates of marsh deterioration; they rather indicate the presence of small erosion rates punctuated by large failure events. Figure 4.8 A displays the global to local erosion ratio as a function of wave power. Four different zones can be identified. Zone 0 is characterized by a nearly null erosion velocity, as well as very low single-cell erosion rates. It represents favorable conditions for salt marshes which would, in this case, deteriorate very slowly. Zone 1 is characterized by large scale erosion rates growing faster with respect to the erosion rate of isolated cells. In this regime, the effect of waves is maximized due to the fact that wave power is sufficiently high to incise salt marshes, but low enough to allow the occurrence of large scale failures. In this zone, marsh non-homogeneity contributes to the creation of the most vulnerable conditions for marsh deterioration. In Zone 2, the erosion rate of the entire system grows slower with respect to the erosion rate of the single units. Finally, in Zone 3 the local and global erosion rates coincide.
Variations of the global to local erosion ratio as a function of extreme events frequency is shown in Fig. 4.8b. As the probability of extremes increases, the peak of the ratio increases as well, moving toward lower wave energy conditions, and is characterized by steeper gradients (therefore, any change in wave power around the peak area would have greater effects). Thus, the presence of extreme events increases large scale erosion rates with respect to erosion rates of smaller units. The maximum value of the ratio is shifted, with respect to the case without extremes because, for an average wave power, the presence of exceptionally high wave power intervals contributes to more uniform, and faster erosion rates (Figure 4.7). Our results suggest that the response of salt marshes to wave action derives from the complex interaction of its smaller units, whose resistance affects the organization of the system at a larger scale. This causes the relative salt marsh vulnerability with respect to an external agent to be maximized for low wave energy conditions. Moreover, natural small scale heterogeneities along marsh boundaries affect salt marsh morphodynamic response to extreme events. Salt marshes that are generally exposed to low wave energy conditions are the most susceptible to changes in the frequency of extreme events as well as to changes in mean wave energy. On the contrary, highly exposed salt marshes remain relatively unaffected by increasingly frequent extreme events. This suggests that variations in time in the morphology of salt marsh boundaries could be used to infer changes in frequency and magnitude of external agents. The global to local erosion ratio also affects the long-term predictability of marsh deterioration. Fast rates of deterioration are relatively constant and predictable. Low
erosion rates, with high global to local ratios, are very unpredictable with large marsh failures punctuating long periods of small erosion rates.

Figure 4.8. A) Ratio between global and local erosion rates as a function of wave power in the absence of extreme events. B) Ratio between global and local erosion rates as a function of wave power for different frequencies of extreme events.
4.4 References


Moller, I. et al. (2014), Wave attenuation over coastal salt marshes under storm surge conditions, Nature Geoscience, 7(10), 727–731.


Temmerman, S., T. J. Bouma, G. Govers, Z. B. Wang, M. B. De Vries, and P. M. J. Herman (2005), Impact of vegetation on flow routing and sedimentation patterns:


CHAPTER 5. The relationship between wave power and salt marsh erosion is linear

The content of this chapter is currently under Review in Proceeding of the National Academy of Science (submission date: May 22, 2015). This paper was co-authored by Neil Kamal Ganju (United States Geological Survey, Woods Hole Science Center, Woods Hole, Massachusetts, USA), and Sergio Fagherazzi (Department of Earth and Environment, Boston University).

Abstract

Salt marsh losses have been documented worldwide due to land use change, wave erosion, and sea level rise\textsuperscript{1-10}. It is still unclear how resistant salt marshes are to extreme storms and whether they can survive multiple events without collapsing. Based on a large dataset of salt marsh erosion rates collected around the world\textsuperscript{11-16}, here we determine the general response of salt marshes to wave action under normal and extreme weather conditions.

As wave energy increases, salt marshes response to wind waves remain linear and there is not a critical threshold in wave energy above which salt marsh erosion drastically increases.

We apply our general formulation for salt marsh erosion to historical wave climates at eight salt marsh locations affected by hurricanes in the United States. Based on the analysis of two decades of data, we find that violent storms and hurricanes contribute to less than one percent to long-term salt marsh erosion rates. In contrast, moderate storms
with a return period of 2.5 months are those causing most salt marsh deterioration. Therefore, salt marshes appear more susceptible to variations in mean wave energy, rather than to changes in the extremes. The intrinsic resistance of salt marshes against violent storms and their predictable erosion rates during moderate events should be taken into account by coastal managers in restoration projects and risk management plans.

### 5.1 Significance statement

In recent years there has been a flurry of restoration projects aimed at mitigating the impact of coastal storms using salt marshes and vegetated surfaces (the so called ‘living shorelines’). Based on a large dataset of salt marsh erosion and wave power measurements collected all around the world, we found that this data can collapse into a unique linear relationship. Our results clearly show that long-term salt marsh deterioration is dictated by average wave conditions, and it is therefore predictable. We find that violent storms and hurricanes contribute to less than one percent to long-term salt marsh erosion rates. This result can be of high value for coastal restoration projects and for detailed coast-benefit analyses.

### 5.2 Main findings

The potential of salt marshes to serve as natural buffers against violent storms appears even more important in view of significant threats imposed by climate change, such as increased storminess and higher hurricane activity registered in the past decades\(^\text{17-20}\). Recent research results show that salt marshes reduce wave energy during storms, and possibly mitigate storm surges\(^\text{21-23}\). These results triggered a flurry of planned coastal restorations centered on the concept of ‘living shorelines’\(^\text{22}\), which use vegetated surfaces
to reduce the impact of hurricanes\textsuperscript{21-24}. Yet little is known about the endurance of salt marshes against wave action, and whether such ecosystems can survive extreme events. Most of marsh erosion occurs at its seaward boundary, where the effect of waves is concentrated\textsuperscript{3, 11}. Wave erosion constitutes one of the main contributions to salt marshes deterioration, and even very small waves can cause failure of large salt marsh blocks\textsuperscript{2, 9, 26, 27}.

Despite the complexity of the problem, some studies have identified a correlation between wave energy and lateral rates of marsh erosion\textsuperscript{11, 12}. Erosion of marsh edges by wave action is caused by many different mechanisms, such as the indentation of V-shaped notches into linear stretches of shoreline, or cliff undercutting when lower sediment layers are eroded more rapidly than the overhanging root-mats\textsuperscript{2, 12, 27}. Varying resistance to wave erosion can be caused by biotic and abiotic factors, such as geotechnical characteristics of the sediments\textsuperscript{25}, vegetation characteristics\textsuperscript{26}, height of the marsh scarp, and presence of mussels or crab burrowing\textsuperscript{28}.

However, existing studies have mainly focused on individual marsh locations, and do not provide a universal relationship applicable to multiple, and ecologically diverse systems. Herein we combine wave energy and marsh erosion data, from different locations in the United States, Australia, and Italy. We show that this data can collapse into a unique relationship (Figure 5.1):

\[ E^* = a^*P^*, \ a^* = 0.67 \]

Equation 5.1
The linear relationship between dimensionless wave power \( P^\ast \) and dimensionless erosion rate \( E^\ast \) is obtained by dividing field measurements by the averaged conditions at that site. Non-dimensionalization allows filtering out the diverse resistance of marsh boundary at individual locations. Field measurements display a linear behavior \((R^2=0.62, p<.05)\), as shown by their average over sub-intervals (grey dots). The non-dimensionalization has been done by assuming that, if a linear relationship is valid for individual data points, then a linear relationship is valid for the averages as well such that

\[ E = aP, \quad E_{avg} = a_{avg}P_{avg}. \]

A general relationship, valid for all sites, is then obtained, and reads: \( E^\ast = a^\ast P^\ast \), where \( E^\ast = E/E_{avg}, \quad P^\ast = P/P_{avg}, \) and \( a^\ast = 0.67 \). Some of the data also account for the occurrence of major storms. As an example, data for Barnegat Bay and Plum Island Sound account for the passage of Hurricane Sandy, ranked as 1/900 year event\(^2\) (see also Figure 5.12, for detailed salt marsh erosion measurements immediately before and immediately after Hurricane Sandy).

Two important observations lie behind the linear nature of the relationship: the first is that salt marsh erosion continuously occurs, even under low wave energy conditions, suggesting the absence of a critical threshold in wave energy below which no erosion is expected. This result underlines the importance of relatively low wave energy conditions for salt marsh lateral retreat. The second observation is that, as wave energy increases, salt marshes do not respond with a catastrophic collapse (e.g. absence of exponential growth in erosion rates), highlighting the intrinsic resilience of salt marshes to extreme events. Scatter can arise from several sources of uncertainties, such as different methods used for the calculation of wave power, and to estimate erosion rates.
Figure 5.1. Relationship between dimensionless wave power ($P^*$), and dimensionless erosion rate ($E^*$) in salt marshes ($R^2=0.62$, $p<0.05$). Solid gray circles indicate values obtained by averaging data points over regular intervals, to emphasize the overall linear trend. The gray area is the uncertainty of the prediction of $E^*$ over a range of coefficients with 95% bounds, which are equal to 0.64, and 0.7. The non-dimensionalization has been done by assuming that, if a linear relationship is valid for individual data points, then a linear relationship is valid for the averages as well such that $E = aP$, and $E_{avg} = a_{avg}P_{avg}$. A general relationship, valid for all sites, is then obtained, and reads: $E^* = a^*P^*$, where $E^* = E/E_{avg}$, $P^* = P/P_{avg}$, and $a^* = 0.67$. 
Figure 5.2. Dimensionless wave power $P^*$, (blue line) and dimensionless erosion rate $E^*$, (pink blocks) for each study site. Wave Power values, $P^*$, are daily average. Yearly erosion rate values and bounds (pink blocks) were obtained using the regression coefficients calculated for the linear relationship between wave power and erosion rate. Major storms affecting the areas of interest are indicated.

We use this general relationship to investigate long-term salt marsh behavior, under realistic wave energy conditions. For this purpose, we collected meteorological data for 23 year period (from 1991 to 2014), at eight different salt marsh locations in the United States, and computed the corresponding wave energy time series (Figure 5.2, Figure 5.5, Figure 5.6, see also Methods).

The areas taken into account were chosen to maximize the occurrence of major hurricanes (Figure 5.7). We use wave energy and Equation 5.1 to estimate yearly salt marsh erosion rates (Figure 5.2). The erosion rate maintains a similar value in different years and at different locations. Moreover, the years characterized by the occurrence of extreme events, such as hurricanes or tropical depressions, do not necessarily correspond to peaks in erosion rate.

We further categorize wind data according to the Beaufort wind scale, and assess the contribution of each wind category to total erosion rate of the entire period of record (Figure 5.3, Figure 5.8). The highest contribution to marsh edge erosion comes from moderate but frequent weather conditions (wind speed ranging from 10 to 40 km/h), while violent storms and hurricanes (wind speed above 65 km/h) contribute to less than one percent to the total marsh edge erosion. This result is due to the linear nature of the relationship between wave power and erosion rate, and to the short duration of extreme
events. In fact, while the action of moderate weather conditions spans most of the study period, the erosion potential of extreme events is concentrated within a few days per year.

Figure 5.3 Average contribution of different wind categories to salt marsh erosion rates: Calm, 0.1% ± 0.05%; light air, 4.0% ± 1.9%; Light breeze, 5.0% ± 2.7%; Gentle breeze, 36% ± 8.3%; Moderate breeze, 18.0% ± 3%; Fresh breeze, 24.0% ± 5.7%; Strong breeze, 7.0% ± 2.5%; Near gale, 5.0% ± 3%; Gale 0.2% ± 0.1%; Strong gale 0.2% ± 0.1%; Storm 0.2% ± 0.07%; Violent storm 0.2% ± 0.05%; Hurricanes 0.1% ± 0.05%. Plots refer to the entire period of record (see also Figure S4 for the contribution of each wind category to a specific field site).

This behavior can be well explained in terms of geomorphic work. Following a magnitude-frequency analysis\textsuperscript{30}, we can multiply the magnitude of marsh retreat for a given wind event by the event's frequency to find the wind event that does the most geomorphic work.

This product attains a maximum indicating the frequency at which the largest portion of the work is accomplished\textsuperscript{30}. The same concept, applied to our test cases, shows that the maximum erosion is attained for frequent and low wave energy conditions, occurring
with a return period of 2.5 ± 0.5 months (Figure 5.4, Figure 5.9). Our results suggest that events occurring with a monthly frequency such as, for instance, winter storms associated to cold-front passages in the Gulf of Mexico, USA, lead to more marsh erosion than hurricanes occurring at a decadal timescale. Therefore extreme storms are not the dominant threat to salt marsh stability as they are to other coastal environments. As an example, beach dunes generally dissipate wave energy during mild storms, whereas they collapse during hurricanes. Moreover, while the response of sandy beaches to external drivers presents multiple stable states, and the effect of storms is amplified or mitigated depending on environmental conditions, the response of salt marshes is constant across different geographic regions and for different climatic conditions. Our analysis might be only valid for salt marshes and is not applicable to brackish or freshwater tidal marshes. The linear relationship between wave energy and erosion, and the fact that salt marsh erosion rates vary little from year to year enable the prediction of the long-term fate of these environments and the estimation of their life-cycle. Even if salt marshes are constantly deteriorating at a slow rate, their predictable response to a wide range of storms and the possibility of forecasting both their life span and mitigation effects make these landforms well suitable for ecosystem based-coastal defense.
Figure 5.4 For Virginia Coast Reserve: frequency-magnitude distribution of the dimensionless wave power, $P^*$ (dashed black line). Total erosion (continuous black line), and dimensionless erosion rate, $E^*$ (dashed blue line) as a function of $P^*$, and of its return time in months. For the Virginia Coast Reserve the return period of wind waves causing maximum erosion is 3 months. The average return period for all sites is $2.5 \pm 0.5$ months. (See also Figure S5 for the plot of geomorphic work for individual bays).

5.3 Methods

5.3.1 Erosion rate and wave power data

We conducted an extensive search, finding the majority of available literature data of marsh erosion as a function of wave power $^{11-16}$.

We further used salt marsh erosion measurements from Plum Island Sound, MA (USA), and in the Barnegat Bay-Little Egg Harbor system. Erosion measurements in Plum Island Sound were collected at three different sites (Figure 5.12), from 2008 to 2013. For one of the sites, field measurements were also collected immediately before and immediately after the occurrence of hurricane Sandy (see also Figure 5.12 to notice that little erosion
was present during such event). Erosion measurements in Barnegat Bay were obtained by
digitalizing more than 100 km of marsh shoreline using aerial images (1930, 2007 and
2013) from the digital orthophotography of New Jersey. These data-sets consist of 0.3 m-
GSD pixel resolution natural color (2007, 2013), and black and white (1930),
orthoimages covering the state on New Jersey (Figure 5.13).

5.3.2 Meteorological data

Data used to compute wave power are available at the National Data Buoy Center
(http://www.ndbc.noaa.gov/). Specifically, we use the following stations: VENF1 27.072
N 82.453 W, for Tampa Bay; SRST2 29.683 N 94.033 W, for Vermillion Bay; LLNR
293 29.212 N 88.207 W, for Lake Borgne; LLNR 1205 27.907 N 95.353 W, for
Galveston Bay; CLKN7 34.622 N 76.525 W, for Pamlico Sound; LLNR 830 40.251 N
73.164 W, for Barnegat Bay; LLNR 168 38.461 N 74.703 W, for Delaware Bay; CHLV2
36.905 N 75.713 W, for Virginia Coast Reserve (Figure S1, S2, S3).

5.3.3 Wind waves

Variables necessary for wave power calculation are obtained from meteorological data
presented in the above paragraph. Average water depth and fetch measurements for each
individual site are presented in Figure 5.5, 5.6.

To compute wave climate, we follow the equations of Young and Verhagen. Wave
height, H, is computed from the wave energy, W, through the expression $W = \rho g H^2 / 8$,
while the wave power is $P = W c_g$, where $c_g$ is the group velocity.
The dimensionless wave energy, \( \varepsilon = \frac{g^2 W}{U^4} \), and peak frequency, \( \nu = \frac{f U}{g} \), are related to the non-dimensional fetch \( \chi = \frac{g x}{U^2} \), and dimensionless water depth \( \delta = \frac{g d}{U^2} \) through the expression:

\[
\varepsilon = 3.64 \cdot 10^{-3} \left\{ \tanh A_1 \tanh \left[ \frac{B_1}{\tanh A_1} \right] \right\}^{1.74}
\]

Equation 5.2

Where \( g \) is the gravitational acceleration, \( U \) is the reference wind velocity at an elevation of 10m, \( f \) is wave frequency, \( x \) is the fetch, \( d \) is water depth and \( A_1 = 0.493 \delta^{0.75} \), \( B_1 = 3.13 \cdot 10^{-3} \chi^{0.57} \). The dimensionless peak frequency is:

\[
\nu = 0.133 \left\{ \tanh A_2 \tanh \left[ \frac{B_2}{\tanh A_2} \right] \right\}^{-0.37}
\]

Equation 5.3

Where \( A_2 = 0.331 \delta^{1.01} \), and \( B_2 = 5.215 \cdot 10^{-4} \chi^{0.73} \).

For data points presented in Figure 5.1, and relative to the Barnegat Bay-Little Egg Harbor system, we use the Coupled-Ocean-Atmosphere-Wave-Sediment-Transport (COAWST) Modeling System\textsuperscript{34} to reconstruct the long term wave climate in the area. Water level variations typical of a tidal cycle, and wind speed and direction measurements collected from the National Data Buoy Center (see above paragraph) are used as model input. See also 35 for more details about the model. The numerical model is used to more conveniently obtain wave power values at the scale of the entire Bay (Figure 5.13, 5.14).
5.3.4 Return period

For the calculation of the return period we use the maxima method, which consists in breaking up the initial sequence of data into monthly blocks, extracting the maximum observation for each block, and fitting an extreme value distribution. Assuming independence between different months, a well-established model for extreme wave heights is based on the Gumbel distribution\(^{36}\) (Figure 5.10), which reads:

\[
F(x; a, b) = \exp \left( - \exp \left( - \left( \frac{x - b}{a} \right) \right) \right), \quad -\infty < x < \infty
\]

Equation 5.4

With \(a\), and \(b\) being the distribution parameters. If \(M_H^k\) is the maximum value during the \(k\)-month, and \(F(x)\) the variable Gumbel distribution function, the \(N\)-year return period, \(s_N\) is:

\[
F(s_N) = 1 - 1/N
\]

Equation 5.5

For the Gumbel distribution, it follows that the return value for the monthly maximum is:

\[
s_T = b - a \log(-\log(1 - 1/T))
\]

Equation 5.6
5.4 Supplementary information

Figure 5.5 Location of bays and wind stations used in the study for the calculation of wave power, erosion rate, and total erosion.
Figure 5.6 Detailed view of bays and wind stations used in the study for the calculation of wind waves and total erosion: A, Galveston Bay, Texas; B, Vermillion Bay, Louisiana; C, Virginia Coast Reserve, Virginia; D, Delaware Bay, Delaware; E, Lake Borgne; F, Pamlico Sound, North Carolina; G, Tampa Bay, Florida; H, Barnegat Bay, New Jersey. Maximum fetch (\(x\), white line), and average water depths (\(d\)) are indicated as well.
Figure 5.7 Path of major hurricanes that affected the areas of interest from 1991 to 2013 (See Table S1 for storms category, and active dates). Locations of studied bays are indicated as well.
Figure 5.8 Contribution of different wind categories to the erosion rate of each site. Plots refer to the entire period of record.

Table 5.1 Major hurricanes and tropical storms that have affected the areas of interest; Storm name, active date, and category according to the Saffir-Simpson hurricane scale are reported. Data are available at (http://www.weather.gov/).
Figure 5.9 Frequency-magnitude distribution of the dimensionless wave power, $P^*$ (dashed black line), total erosion (continuous black line), dimensionless erosion rate, $E^*$, (dashed blue line), as a function of $P^*$. The return period (months), corresponding to a given wave power, $P^*$, is indicated as well.
Figure 5.10 Frequency of wave events computed using a Gumbel distribution. On the horizontal axis wave height observations are indicated, while on the vertical axis the reduced variable $\log(-\log(1 - F(x)))$ is shown. The blue lines fit data according to the Gumbel distribution (See Methods).
Figure 5.11 Non dimensional wave power and non-dimensional erosion rates for each individual study site. Colors used are the same than in Figure 1. A) Lake Borgne, USA; B) Western Port Bay, AU; C) Virginia Coast Reserve, USA (a); D) Virginia Coast Reserve, USA (b); E) Venice Lagoon, IT; F) Delaware Bay, USA; G) Plum Island Sound; H) Barnegat Bay, USA.
Figure 5.12 A) Location of field measurements collected in Plum Island Sound, USA. Boundary measurements have been taken for stretches of shoreline of around 100 m. The three locations are Refuge North (42°45′00 N 70° 48′ 00 W), Stackyard Road (42°44′00 N 70° 49′00 W) and Refuge South (42°44′00 N 70° 48′00 W). Marsh contours have been tracked using a Real-Time Kinematic-GLOBAL Positioning System (RTK-GPS) and an electronic Total Station every year from 2008 to 2013. Data were collected with an
average resolution of 1 m. When marsh contours were characterized by significant variations in boundary geometry, measurements were taken up to 20 cm apart.

B) Marsh boundary at Stackyard Road. Different colors are measurements at different years. The boundary presented in this plot is eroding at an average rate of 0.35 m/y. Is it possible to notice that the erosion occurred during Hurricane Sandy (1/900 year event) is negligible for great part of the shoreline, and smaller than the average yearly erosion.

Figure 5.13 Erosion rates (m/y) in Barnegat Bay. Erosion rates were obtained by digitalizing more than 100 km of marsh shoreline using aerial images (1930, 2007 and 2013) from the digital orthophotography of New Jersey (see also Methods). Erosion rate has been obtained by comparing images from 1930, and 2007 and from 2007 to 2013, dividing by the number of years, and averaging between the two data-sets.
Figure 5.14 A) Wind rose for the Barnegat Bay-Little Egg Harbor system (National Data Buoy station LLNR 830 40.251 N 73.164 W) for the period from 1991-2013. B) Bathymetry of the Barnegat-Bay Little Egg Harbor system. Yellow lines indicate the boundaries of the computational domain. The model used is the Coupled-Ocean-Atmosphere-Wave-Sediment-Transport (COAWST) Modeling System [Warner et al., 2008; Warner et al., 2010]. In COAWST the ocean model ROMS, the atmospheric model WRF, the wave model SWAN, and the modules of the Community Sediment Transport Model are fully coupled by means of the Model Coupling Toolkit. The ocean model ROMS is a three-dimensional, free-surface, terrain-following model solving finite difference approximations of the Reynolds Averaged Navier-Stokes equations, using hydrostatic and Boussinesq assumptions [Chassignet et al., 2000; Shchepetkin and McWilliams, 2005; Kumar et al., 2011].

The 3D primitive equations for the wave-averaged currents in horizontal orthogonal curvilinear and terrain following vertical coordinate systems, are given by the following equations:

\[ \frac{\partial}{\partial t} \left( \frac{H_z^c}{mn} u \right) + \frac{\partial}{\partial \xi} \left( \frac{H_z^c}{n} u \right) + \frac{\partial}{\partial \eta} \left( \frac{H_z^c}{m} v \right) + u \frac{\partial}{\partial \xi} \left( \frac{H_z^c}{n} u \right) + u \frac{\partial}{\partial \eta} \left( \frac{H_z^c}{m} u \right) + \frac{1}{mn} \frac{\partial}{\partial s} \left( \frac{w_s^c}{mn} u \right) + u \frac{\partial}{\partial s} \left( \frac{w_s^c}{mn} u \right) = - \frac{H_z^c}{n} \frac{\partial \phi^c}{\partial \xi} \big|_z + \frac{H_z^c}{mn} \frac{f v}{mn} + H_z^c f v^st + H_z^c v^st \frac{1}{n} \frac{\partial v}{\partial \xi} - \frac{1}{m} \frac{\partial u}{\partial \eta} \big|_z - \frac{1}{mn} \frac{w_s^c}{mn} \frac{\partial}{\partial s} \left( \frac{H_z^c}{mn} \right) + \frac{H_z^c}{mn} P^c + \frac{H_z^c}{mn} F^{w^c} + \frac{H_z^c}{mn} D^c \]
\[ -\frac{\partial}{\partial s} \left( u'w' - v \frac{\partial u}{H_z^2 \partial s} \right) + \ddot{p} \]

Equation 5.7

**y-momentum balance:**

\[
\frac{\partial}{\partial t} \left( \frac{H_z^2 v}{mn} \right) + \frac{\partial}{\partial \xi} \left( \frac{H_z^2 u}{n} \right) + \frac{\partial}{\partial \eta} \left( \frac{H_z^2 v}{m} \right) + v \frac{\partial}{\partial \xi} \left( \frac{H_z u_{st}}{n} \right) + v \frac{\partial}{\partial \eta} \left( \frac{H_z v_{st}}{m} \right) + \frac{1}{mn} \frac{\partial}{\partial s} (w_s u) + v \frac{\partial}{\partial s} \left( \frac{w_{st}}{mn} \right) = -\frac{H_z^2}{m} \frac{\partial \phi^c}{\partial \eta} \left|_{z} - \frac{H_z^2}{mn} f u - \frac{H_z^2}{mn} f u_{st} - H_z^2 u_{st} \left( \frac{1}{n} \frac{\partial v}{\partial \xi} - \frac{1}{m} \frac{\partial u}{\partial \eta} \right) - \frac{1}{mn} w_{st} \frac{\partial}{\partial s} (v) + \frac{H_z^2}{mn} F^\eta + \frac{H_z^2}{mn} F^{\omega \eta} + \frac{H_z^2}{mn} D^\eta \right. \\
- \frac{\partial}{\partial s} \left( \dddot{v} - v \frac{\partial u}{H_z^2 \partial s} \right) + \ddot{p} \eta
\]

Equation 5.8

**continuity equation:**

\[
\frac{\partial}{\partial t} \left( \frac{H_z^2 u}{mn} \right) + \frac{\partial}{\partial \xi} \left( \frac{H_z^2 (u + u_{st})}{n} \right) + \frac{\partial}{\partial \eta} \left( \frac{H_z^2 (v + v_{st})}{m} \right) + \frac{1}{mn} \frac{\partial}{\partial s} (w_s + w_{st}) = 0
\]

Equation 5.9

In the equations \( m^{-1} \) and \( n^{-1} \) are the Lamé coefficients, \( u, v, w_s \) are the mean Eulerian velocity in the horizontal (\( \xi, \eta \)), and vertical (\( s \)) directions. \( H_z^2 \) is the grid cell thickness, \( f \) is the Coriolis coefficient, \( \phi^c \) is the geopotential function, \( (F^\eta, F^{\omega \xi}) \) are non-wave body forces, including wind shear stress and bottom shear stress; \( (F^{\omega \eta}, F^{\omega \xi}) \) are non-conservative wave-induced accelerations; \( (D^\xi, D^\eta) \) are parameterized horizontal momentum mixing terms. Overbars represent averages, and a prime represents turbulent fluctuations. To include the effect of surface waves, the momentum equations require, as input, information for wave height, wave energy dissipation, propagation direction, and wavelength. These information are obtained from SWAN (Simulating Waves Nearshore), which solves the transport equations for waves action density, and accounts for shoaling, refraction, wind waves generation, wave breaking, bottom dissipation, and non-linear wave interactions [Booij et al., 1999]. In the COAWST modeling system, ROMS provides SWAN with free surface elevation and currents. Currents are computed by taking into account the vertical distribution of the current profile, as well as the relative depth of surface waves [Warner et al., 2008]. Model domain (comprised in between yellow lines) includes Barnegat Bay, Little Egg Harbor, part of Great Bay to the South, and Manasquan inlet to the north. The computational grid consists of 160 East-West and 800 North-South grid points with seven vertical layers having equal depth. Cell sizes vary from 40 to 200 m, with the grid being refined at the inlets and around complex
morphological features. The bathymetry is based on the National Ocean Service Hydrographic Survey data (NOAA NOS 2012, Defne and Ganju, 2014). At the seaward boundaries we prescribed water level variations typical of a tidal cycle and used a combination of Chapman, Flather, and gradient boundary conditions. At the westward boundary we prescribed a radiation boundary condition allowing tidal energy to propagate landward.

The model was calibrated by changing the bottom roughness coefficient to attain the best agreement with measurements collected at the Barnegat Bay Little Egg Harbor estuary. The Brier-Skill-Score [e.g. Murphy and Epstein 1989] was used to assess model calibration. Skill assessments of the model ranged from very good to excellent [Defne and Ganju, 2014]. More specific information about model calibration and setup can be found in Defne and Ganju [2014]. The model was run for wind speed ranging from 5 to 30 m/s, and for different wind directions. Wind data (speed and direction) from 1991 to 2013 were retrieved from a nearby NOAA station (station ID LLNR 830). Wave information for each point along the marsh shoreline, corresponding to different wind directions and wind speed conditions, were used to reconstruct the long-term average wave climate in the area.
5.5 References


15. Tomkins, K., McLachlan, G., Coleman, R., Hurst, T. & Glover, M., Coastal bank erosion data, Western Port Bay. v3. CSIRO. Data Collection. 102.100.100/12889 (2013).


Chapter 6. Macroscopic indicators of salt marsh erosion rates in Barnegat Bay, New Jersey, USA

The content of this chapter is the final draft of a research paper that will be submitted to Journal of Geophysical Research: Earth Surface. This paper was co-authored by Neil Kamal Ganju (United States Geological Survey, Woods Hole Science Center, Woods Hole, Massachusetts, USA), and Sergio Fagherazzi (Department of Earth and Environment, Boston University).

Abstract

We investigate the relationship between wind waves, salt marsh erosion rates and the shape of the marsh boundary by using aerial images, and the numerical model COAWST. We found that, on average, salt marsh erosion remains constant in time, and appears to be independent of the occurrence of extreme events. We also found a significant relationship between salt marsh erosion rates and the shape of marsh boundaries: slowly eroding salt marshes are rougher with fractal dimension higher than rapidly deteriorating salt marshes. While average erosion rates remain constant, individual erosion events are unpredictable. Specifically, for low wave energy conditions, there are higher probabilities of isolated and far from the mean, high magnitude erosion events. Finally, we confirm the existence of a significant relationship between salt marsh erosion rate and wind waves exposure. These results suggest that variations in time in the morphology of salt marsh boundaries could be used to infer changes in frequency and magnitude of external agents.
6.1 Introduction

Located at the interface between marine and terrestrial environments, salt marshes are ecosystem-based flood defenses and help reducing the impact of storms and hurricanes on coastal communities [e.g. Temmerman et al., 2013; Moller et al., 2014; Chen et al., 2011; Zhao and Chen, 2013]. In recent years, these vegetated surfaces have been at the center of many restoration projects based on the concept of “living shorelines” [e.g. Temmerman et al., 2013; Fagherazzi, 2014], and the value of their storm-protection services has been estimated up to $5 million per km$^2$ in the United States [Costanza et al., 2008]. Salt marshes also provide other important ecosystem services such as nutrients removal, a unique habitat for many floral and faunal species, and large amount of carbon stored over decennial and millennial time scales [Zedler et al., 2005]. In spite of their valuable services, salt marsh losses have been observed worldwide and their survival is threatened by human activities, sea level rise, and wave action. Salt marshes lateral retreat is frequently of the order of meters per year [e.g. Fagherazzi, 2013]. Feedbacks between sediment deposition, organic matter accretion, and plant growth generally allow salt marshes to keep peace with sea level rise. On the contrary, salt marshes are inherently weak with respect to wave action because wind waves impact bare sediments below the vegetation surface where material is easily removable [e.g. Fagherazzi et al., 2013]. Therefore, understanding the response of salt marshes to wind waves is a key issue for society because is at the base of restoration projects aimed at the maintenance and improvements of wetlands ecosystem services.
A direct relationship between wave energy and salt marsh lateral retreat has been frequently recognized [e.g. Schwimmer 2001; Marani et al., 2011; McLoughlin et al., 2015]. However, many existing studies only focus on small spatial scales (shoreline lengths of the order of hundreds of meters), and less attention has been paid to the frequency-magnitude distribution of erosion events, as well as to the possible influence of different wind wave exposures and erosion rates on the shape of marsh shorelines. The frequency-magnitude distribution of erosion events is relevant to risk assessment in coastal wetlands, as it allows gaining more information about the occurrence of erosion events far from average conditions. In fact, the occurrence of unexpected erosion episodes constitutes a catalyst for questioning the effectiveness of risk assessment plans and coastal protection schemes. There is thus a need for further information about the possible unpredictability of isolated salt marsh failures as well as for ways to incorporate infrequent episodes into vulnerability estimates, and shoreline protection strategies. Moreover, many studies use wave power as representative for wind wave’s exposure. While this approach has been frequently proved as successful, when enough information is available wave thrust could be a better indicator for the action of wind waves on wetlands boundaries since it constitutes the actual force contributing to the dynamic equilibrium of marsh portions [e.g. Tonelli et al., 2011; Bendoni et al., 2014; Francalanci et al., 2013].

In this manuscript we focus on erosion processes, and wave action at the boundary of salt marshes in the Barnegat-Bay Little-Egg Harbor system. After presenting the study site and the adopted methodology, we focus on the frequency-magnitude distribution of
erosion events, and on the relationship between the shape of marsh boundaries and erosion rates. The erosion analysis has been carried out for extensive stretches of shoreline (order of hundreds of kilometers). We further explore the long-term relationship between erosion rates and wave thrust at the marsh boundary. Finally, a set of discussions and conclusions are presented.

6.2 Study site

The Barnegat-Bay-Little Egg Harbor system is located along the central New Jersey coastline in Ocean County (between 39°31’ and 40°06’ N latitude and 74°02’W and 74°20’W longitude), and consists of three shallow coastal bays: Barnegat Bay, Manahawkin Bay, and Little Egg Harbor. The system takes stretches of shoreline of around 70 km parallel to the mainland, and comprises a shallow, lagoon type estuary, separated from the Atlantic Ocean by a narrow barrier island complex. Lagoon width ranges from 2 to 6 km, and water depth ranges from 1 to 5 m, with an average value of 1.5 m. Most of the western side of Barnegat Bay is deeper (~1-3 m depth) than the Eastern side (~1 m depth). The deepest area runs along the Intracoastal Waterway and spans along the bay length [Kennish, 2001; Defne et al., 2014; Aretxabaleta et al., 2014]. The barrier is breached at Barnegat Inlet and Little Egg Inlet. Barnegat Inlet has a mean width of 400 m, and a maximum water depth of 15 m. Little Egg Inlet has a width of approximately 2 km, and average water depth of 10 m. Figure 6.1 shows the bathymetry of the area and long-term statistics of wind speed and direction (from 1991 to 2013). The
area is characterized by strong wind seasonality. During the fall and winter seasons, winds predominantly come from Northwest.

Figure 6.1 A) Bathymetry of the Barnegat-Bay Little Egg Harbor system. Yellow lines indicate the boundaries of the computational domain. B) Wind rose for the area for the period from 1991-2013.
Figure 6.2 A) example of digitalized shoreline. The location is the white circle in Figure 6.2C. B) Fractions of total digitalized shoreline as a function of different erosion rates. C) Erosion rates map measured from 1930 to 2007. D) Erosion rates map measured from 2007 to 2013.

to West directions. During spring and summer, the jet stream retreats northward, and winds are from the South to the Southwest. In summer, subtropical high pressures generally produce warm and humid southerly breezes. Northeast winds, associated with Atlantic storms known as “northeasters,” do not strongly impact the region.
Tides are semidiurnal, the tidal range outside the estuary is over 1 m, but it rapidly attenuates within the estuary. Tidal oscillations within the Bay are around 20% of the offshore water level because in the 6 h interval between low and high water there is not enough time for the water to flow through the inlets so that the water level can match the one offshore [Aretxabaleta et al., 2014].

In the past, the lagoon was almost completely fringed by salt marshes. In 1888, salt marshes area was approximately 14850 ha, while in 1995 salt marshes extent was estimated to be 9940 ha; 65% of this decline has been directly attributed to land use and losses due to development, while the remaining lost has been attributed to natural or human disturbances [Lathrop and Bognar, 2001].

### 6.3 Methods

#### 6.3.1 Aerial photograph

To determine the geometry of the salt marsh boundary and related erosion rates we used three aerial images (1930, 2007 and 2013) from the digital orthophotography of New Jersey. These data-sets consist of 0.3 m-GSD pixel resolution natural color (2007, 2013), and black and white (1930), orthoimages covering the state on New Jersey. For each aerial photograph we digitalized more than 100 km of shoreline, corresponding to great part of the interior marsh boundary. Using the digitalized aerial images, we then computed salt marsh erosion from 1930 to 2013 and from 2007 to 2013 (Figure 6.2).
For a 0.3 m GSD orthoimage, the horizontal position accuracy for a given year ($E_i$) was designed not to exceed 1.2 m, and according to the National Standard for Spatial Data Accuracy, the RMSE for the 95% confidence is 0.88 m. Root mean square errors can be assumed as representative for rectification errors. We further accounted for a digitalization error ($E_i$) of 0.5 m for each aerial image. This value was considerate appropriate, and is a precautionary value, with respect to the actual digitalization error evaluated by digitalizing multiple times same stretches of shoreline and considering maximum distances between the same digitalized marsh boundary profile. Total position uncertainty ($U_t$) was estimated as [Cowart et al., 2011; McLoughlin et al., 2015]:

$$U_t = \pm \sqrt{E_r^2 + \Sigma E_i^2} \quad \text{(Equation 6.1)}$$

where the sum refers to different years. Total position uncertainty, $U_t$, is $\pm 1.5$ m and corresponds to a maximum uncertainty of 0.2 m/y when comparing the 2007 and 2013 aerial images, and 0.02 m/y when comparing the 1930 and the 2007 images.

To compute shoreline fractal dimension, we followed the classic Minkowski–Bouligand method [e.g. Dubuc et al., 1989]. Fractal dimension is a useful metric and allows evaluating the degree of marsh boundary complexity by computing how fast length measurements increase or decrease as the measuring scale becomes larger or smaller.

Generally speaking, given a topological set ($S$), of dimension $n$, for any $\varepsilon > 0$, let $N_\varepsilon(S)$ be the minimum number of $n$-dimensional cubes of side-length $\varepsilon$ needed to cover $S$. The fractal dimension, $D$, is such that:
\[ N_\varepsilon(S) \sim 1/\varepsilon^D \text{ as } \varepsilon \to 0 \]  
(Equation 6.2)

And:

\[ D = -\lim_{\varepsilon \to 0} \frac{\log N_\varepsilon(S)}{\log(\varepsilon)} \]  
(Equation 6.3)

If a limit curve, or surface is smooth and differentiable, its fractal dimension, \( D \), equals the topological dimension, \( d \). For a rough and non-differentiable curve the fractal dimension may exceed the topological dimension and take values in between \( d \) and \( d + 1 \).

Apart, from average erosion values, we are also interested in evaluating the frequency-magnitude distribution of erosion events. In particular, it is especially useful to have information about the possible occurrence of unexpected and above average erosion rates, belonging to the tail of the erosion events distribution. To gain more information about these tail events, we approximated the frequency-magnitude distribution of erosion events by means of a logarithmic frequency-magnitude distribution. The main advantages of using a distribution that is normal in the logarithm of a parameter are: i) the distribution is positive-definite, and it is thus suitable to represent quantities that cannot be negative, such as erosion rate values; ii) For many natural processes, where the possibility of events falling far from the average exists, a logarithmic distribution often provides a better fit with respect to Gaussian-like functions.

For a log-normal distribution the probability density function, \( f_\varepsilon \), is:
\[ f_x(x, \mu, \sigma) = \frac{1}{x\sigma\sqrt{2\pi}} e^{-\frac{(lnx-\mu)^2}{2\sigma^2}} \]

(Equation 6.4)

Where \( \mu \) is the variable mean, and \( \sigma \) is the variance. With increasing variance, the tail of the frequency magnitude distribution of erosion events gets longer, distribution skewness increases, and the possibility of erosion events far from the mean increases as well.

### 6.3.2 Wave condition and Model Description

In this study we used the Coupled-Ocean-Atmosphere-Wave-Sediment-Transport (COAWST) Modeling System [Warner et al., 2008; Warner et al., 2010]. In COAWST the ocean model ROMS, the atmospheric model WRF, the wave model SWAN, and the modules of the Community Sediment Transport Model are fully coupled by means of the Model Coupling Toolkit. The ocean model ROMS is a three-dimensional, free-surface, terrain-following model solving finite difference approximations of the Reynolds Averaged Navier-Stokes equations, using hydrostatic and Boussinesq assumptions [Chassignet et al., 2000; Shchepetkin and McWilliams, 2005; Haidvogel et al., 2008; Kumar et al., 2011].

The 3D primitive equations for the wave-averaged currents in horizontal orthogonal curvilinear and terrain following vertical coordinate systems, are given by the following equations:

\( x \)-momentum balance:

\[ \frac{\partial}{\partial t} \left( \frac{H_z^c}{mn} u \right) + \frac{\partial}{\partial \xi} \left( \frac{H_z^c u}{n} \right) + \frac{\partial}{\partial \eta} \left( \frac{H_z^c}{m} u \right) + u \frac{\partial}{\partial \xi} \left( \frac{H_z u^{st}}{n} \right) + u \frac{\partial}{\partial \eta} \left( \frac{H_z v^{st}}{m} \right) + \]
\[
\frac{1}{mn} \frac{\partial}{\partial s} (w_s u) + u \frac{\partial}{\partial s} \left( \frac{w_s^{St}}{mn} \right) = - \frac{H_z^c}{n} \frac{\partial \phi^c}{\partial \xi} |_z + \frac{H_z^c}{mn} f v + \frac{H_z^c}{mn} f^v^{St} + \\
H_z^c v^{St} \left( \frac{1}{n} \frac{\partial v}{\partial \xi} - \frac{1}{m} \frac{\partial u}{\partial \eta} \right) - \frac{1}{mn} w_s^{St} \frac{\partial}{\partial s} (u) + \frac{H_z^c}{mn} f^\xi + \frac{H_z^c}{mn} F^w^\xi + \frac{H_z^c}{mn} D^\xi \\
- \frac{\partial}{\partial s} \left( \frac{w^{St} w}{H_z^c} \right) + \hat{F}^\xi
\]

(Equation 6.5)

y-momentum balance:
\[
\frac{\partial}{\partial t} \left( \frac{H_z^c}{mn} v \right) + \frac{\partial}{\partial \xi} \left( \frac{H_z^c u}{n} v \right) + \frac{\partial}{\partial \eta} \left( \frac{H_z^c v}{m} v \right) + v \frac{\partial}{\partial \xi} \left( \frac{H_z^c u^{St}}{n} \right) + v \frac{\partial}{\partial \eta} \left( \frac{H_z^c u^{St}}{m} \right) + \\
\frac{1}{mn} \frac{\partial}{\partial s} (w_s u) + v \frac{\partial}{\partial s} \left( \frac{w_s^{St}}{mn} \right) = - \frac{H_z^c}{m} \frac{\partial \phi^c}{\partial \eta} |_z - \frac{H_z^c}{mn} f u - \frac{H_z^c}{mn} f^u^{St} - \\
H_z^c u^{St} \left( \frac{1}{n} \frac{\partial v}{\partial \xi} - \frac{1}{m} \frac{\partial u}{\partial \eta} \right) - \frac{1}{mn} w_s^{St} \frac{\partial}{\partial s} (v) + \frac{H_z^c}{mn} F^\eta + \frac{H_z^c}{mn} F^w^\eta + \frac{H_z^c}{mn} D^\eta \\
- \frac{\partial}{\partial s} \left( \frac{v^{St} v}{H_z^c} \right) + \hat{F}^\eta
\]

(Equation 6.6)

continuity equation:
\[
\frac{\partial}{\partial t} \left( \frac{H_z^c}{mn} \right) + \frac{\partial}{\partial \xi} \left( \frac{H_z^c (u + u^{St})}{n} \right) + \frac{\partial}{\partial \eta} \left( \frac{H_z^c (v + v^{St})}{m} \right) + \frac{1}{mn} \frac{\partial}{\partial s} (w_s + w_s^{St}) = 0
\]

(Equation 6.7)

In the equations \( m^{-1} \) and \( n^{-1} \) are the Lamé coefficients, \( u, v, w_s \) are the mean Eulerian velocity in the horizontal (\( \xi, \eta \)), and vertical (\( s \)) directions. \( H_z^c \) is the grid cell thickness, \( f \) is the Coriolis coefficient, \( \phi^c \) is the geopotential function, \( F^\eta, F^\xi \) are non-wave body forces, including wind shear stress and bottom shear stress; \( F^w^\eta, F^w^\xi \) are non-
conservative wave-induce accelerations; \((D^\xi, D^\eta)\) are parameterized horizontal momentum mixing terms. Overbars represent averages, and a prime represents turbulent fluctuations. To include the effect of surface waves, the momentum equations require, as input, information for wave height, wave energy dissipation, propagation direction, and wavelength. These information are obtained from SWAN (Simulating Waves Nearshore), which solves the transport equations for waves action density, and accounts for shoaling, refraction, wind waves generation, wave breaking, bottom dissipation, and non-linear wave interactions [Booij et al., 1999; Ris et al., 1999]. In the COAWST modeling system, ROMS provides SWAN with free surface elevation and currents. Currents are computed by taking into account the vertical distribution of the current profile, as well as the relative depth of surface waves [Warner et al., 2008; Olabarrieta et al., 2011].

Model domain is presented in Figure 6.1A (yellow line) and it includes Barnegat Bay, Little Egg Harbor, part of Great Bay to the South, and Manasquan inlet to the north. The computational grid consists of 160 East-West and 800 North-South grid points with seven vertical layers having equal depth. Cell sizes vary from 40 to 200 m, with the grid being refined at the inlets and around complex morphological features. The bathymetry is based on the National Ocean Service Hydrographic Survey data (NOAA NOS 2012, Defne and Ganju, 2014). At the seaward boundaries we prescribed water level variations typical of a tidal cycle and used a combination of Chapman, Flather, and gradient boundary conditions. At the westward boundary we prescribed a radiation boundary condition allowing tidal energy to propagate landward.
The model was calibrated by changing the bottom roughness coefficient to attain the best agreement with measurements collected at the Barnegat Bay Little Egg Harbor estuary. The Brier-Skill-Score [e.g. Murphy and Epstein 1989] was used to assess model calibration. Skill assessments of the model ranged from very good to excellent [Defne and Ganju, 2014]. More specific information about model calibration and setup can be found in Defne and Ganju [2014]. The model was run for wind speed ranging from 5 to 30 m/s, and for different wind directions. Wind data (speed and direction) from 1991 to 2013 were retrieved from a nearby NOAA station (station ID LLNR 830). Wave information for each point along the marsh shoreline, corresponding to different wind directions and wind speed conditions, were used to reconstruct the long-term average wave climate in the area.

### 6.3.3 Wave thrust

In this study we are interested in the contribution of wind waves to salt marshes lateral retreat. The action of wind waves on salt marshes has been found to be modulated by tidal elevations [e.g. Tonelli et al., 2010]. Laboratory and numerical experiments [e.g. Tonelli et al., 2010; Bendoni et al., 2014; Fracalanci et al., 2013] confirm that the action of wind waves at the edge of salt marshes is more severe when water levels are close and slightly below the marsh platform. Specifically, wave forcing has been found to increase with tidal elevation, up to the point when the marsh is submerged, and then it rapidly decreases. When wind waves are below the marsh platform, they are partially reflected and their high increases; once the bank is submerged, wind waves are affected by water
depth on the top of the surface, they rapidly break, and a large amount of energy is
dissipated due to the presence of vegetation [e.g. Tonelli et al., 2010; Moller et al., 2014].
We used wave thrust as a proxy for the action of wind waves on marsh boundaries. Wave
thrust is the integral along the vertical of the dynamic pressure of waves. Wave thrust
allows taking into account both wave and salt marsh boundary features, such as platform
elevation with respect to mean sea level, and orientation of the boundary with respect to
wave direction. Moreover, wave thrust calculation is at the base for physically based
models aimed at describing specific failure mechanisms such as toppling failures. For the
latter, marsh blocks are schematized as simple dynamical systems which respond to an
external forcing by oscillating around an equilibrium configuration [e.g. Bendoni et al.,
2014]. The dynamic pressure of wind waves can be calculated as:

\[ p_d(z, t) = \rho g K_p(z) \eta(t) \]  
\hspace{10cm} \text{(Equation 6.8)}

Where \( K_p(z) \) is the pressure-response factor accounting for the dynamic component due
to water particle acceleration, and can be calculated as:

\[ K_p(z) = \frac{\cosh(kh + z)}{\cos(kh)} \]  
\hspace{10cm} \text{(Equation 6.9)}

Therefore, wave thrust can be calculated by integrating the dynamic pressure from the
bottom to the water surface. To take into account the dependence of wave thrust on tidal
elevation, we followed results from Tonelli et al. [2010], and prescribed an up to 60%
reduction in wave thrust proportional to the water depth on the marsh platform. As part of
this project we also developed a subroutine for wave thrust calculation for the COAWST modeling suite (see supplementary material).

6.4 Results

6.4.1 Boundary erosion analysis

Figure 6.2a provides a detail of one of the digitalized maps. The location is indicated in Figure 6.2c (white circle). Figure 6.2b reports the fractions of the entire digitalized shoreline subject to different erosion rates (percentages based on a comparison between the 1930 and 2013 orthoimages). More than half of the digitalized shoreline is eroding at rates ranging from 0.25 to 2 m/y. Around 40% of the shoreline is eroding slower than at 0.25 m/y, or is not eroding at all. Erosion peaks are around 4 m/y and affect points in proximity of Little Egg Harbor.

Figure 6.2 also shows erosion rates in the area (m/y) from 1930 to 2007 (2c) and from 2007 to 2013 (2d). Rates of change of shoreline position varied spatially along marsh edges. However, average erosion rates occurring during the two different time frames show similarities in terms of both magnitude and spatial distribution. This indicates that average erosion rates remained constant for the two different temporal scales. The highest erosion rates occur in proximity of Little Egg Inlet, around points of greatest projection in Little Egg Harbor, and at the eastern portion of Barnegat Bay. Generally speaking, the eastern part of the bay appears more vulnerable with respect to the west side.

Fractal dimension was used to estimate shoreline roughness, and the existence of a potential relationship between erosion rates and morphology of salt marsh boundaries.
Figure 6.3a shows the relationship between the fractal dimension of the marsh boundary (calculated from the 2013 orthoimage), and erosion rates for the period from 1930 to 2013. Fractal dimension has been found to be significantly, and negatively correlated with erosion rates in the area ($R^2=0.60$, $p<0.005$). The significant correlation suggests that rapidly eroding areas have smoother marsh boundary profiles with respect to the slowly eroding ones.

Figure 6.3 A) relationship between erosion rate (horizontal axis), and fractal dimension (vertical axis) Solid black circles indicate values obtained by averaging data over regular bins to emphasize the overall trend; B) Relationship between erosion rate (horizontal axis), and the standard deviation of the erosion events (vertical axis). C, D, E example of frequency magnitude distribution of erosion events for different stretches of shoreline, and corresponding shoreline images. For shoreline locations, see corresponding colored circles in Figure 6.2C.

To gain further information about salt marsh erosion processes, apart from average erosion values, it is also useful to investigate the frequency magnitude distribution of erosion events.
The likelihood of an erosion event to fall far away from average erosion rate values was estimated by approximating the frequency magnitude distribution of erosion events using a log-normal distribution, and by computing corresponding standard deviation values.

Figure 3b show standard deviation, $\sigma$, values as a function of erosion rate. The standard deviation is negatively correlated to average erosion rates in the area. This indicates that unexpected erosion episodes far from the mean more likely occur in slowly eroding areas. On the contrary, low standard deviation values, corresponding to high average erosion rates, indicate that in fast degrading marshes the majority of erosion episodes fall around the mean. Figure C, D, and E are example of frequency magnitude distributions of erosion events and marsh boundary profiles at three different sites. Sites locations are indicated in Figure 6.2 (see colored circles Figure 6.3C, D, E and Figure 6.2). Again, is it possible to notice that a longer tail of the frequency-magnitude distribution of erosion events corresponds to a slowly eroding marsh with a rougher boundary profile.

### 6.4.2 Wind wave exposure analysis

A series of simulation were conducted to explore the general system susceptibility to winds having different intensity and directions. Figure 6.4A represents average and maximum values of wave power at marsh boundaries, as a function of wind direction, and for two different wind speed values equal to 5m/s (left panel), and 10 m/s (right panel). Lighter colors correspond to average values of wave power for the entire bay. Darker colors represent the maximum values of wave power within the entire bay. In a
similar fashion, Figure 6.4b represents average (lighter colors), and maximum (darker colors) wave power in the entire Bay, for different wind directions, and two different wind speed values. B) Average wave thrust (light color), and maximum wave thrust (dark color) in the entire Bay, for different wind directions, and two different wind speed values. C) Average wave thrust (light color), and maximum wave thrust (dark color) in the entire Bay, for different wind directions, and two different wind speed values. Wave thrust has been reduced proportionally to the water depth on the top of the marsh scarp.

Figure 6.4 A) Average wave power (light color), and maximum wave power (dark color) in the entire Bay, for different wind directions, and two different wind speed values. B) Average wave thrust (light color), and maximum wave thrust (dark color) in the entire Bay, for different wind directions, and two different wind speed values. C) Average wave thrust (light color), and maximum wave thrust (dark color) in the entire Bay, for different wind directions, and two different wind speed values. Wave thrust has been reduced proportionally to the water depth on the top of the marsh scarp.
colors) wave thrust values for the entire bay. Figure 6.4c represents wave thrust values that accounts for the decline in wave forcing once the water level overtop the marsh scarp. Wave power and wave thrust show a similar dependence from wind direction, and both increase with wind speed. Wind directions parallel to the bay and coming from the West, correspond to high fetch values and to the highest wave action at marsh boundaries. Figure 6.5 represents average wave thrust values calculated from 1991 to 2013 for the entire bay, and for areas fringed by salt marshes. Wave thrust values in Figure 6.5 correspond to values averaged throughout every tidal cycle, and then further averaged from 1991 to 2013. Values were obtained by associating at every point of the boundary and at every wind speed, direction, and water level, wave thrust values obtained from the numerical tests carried out to assess general system susceptibility to different winds. If a reduction in wave thrust when the marsh is submerged is taken into account (Figure 6.5b), average wave thrust values for the period of interest strongly decline but maintain a similar spatial distribution. The more noticeable differences in the spatial distribution of the two different wave thrust calculations are at eastern corner of Great Bay, and at west side of Barnegat Bay. Generally speaking, the eastern part of Barnegat Bay, the South-East side of Great bay, and points of greatest projection in Little Egg Harbor appear to be the
Figure 6.5 Average wave thrust values for the entire bay from 1991-2013. With and without the reduction in wave thrust during salt marsh submergence.

most exposed to the action of wind waves. Figure 6.6 shows the relationship between wave thrust and salt marsh erosion rates. Each cell of the computational grid contains multiple digitalization points and erosion rates in these plots have been averaged within each cell. Colored bands are areas contain 95% of the data points. Despite of the noticeable scatter, wave thrust and salt marsh erosion rates appear significantly related. Moreover, the correlation between the two variables slightly increases when taking into account a decrease in wave forcing with marsh submergence. Scatter may arise from a variety of different sources. Among others, differences in soil resistance, vegetation density and bioturbation significantly affect salt marsh erosion rates.
6.5 Discussion

From a long-term analysis of salt marsh lateral retreat in the Barnegat Bay and Little-Egg Harbor system, we observed that shoreline erosion rates remain constant in time.

Specifically, erosion rates (m/y) evaluated for the period from 1930 to 2007, are similar and of the same order of magnitude than erosion rates occurring from 2007 to 2013. This is especially relevant considering that the second period of interest was characterized by the occurrence of hurricane Sandy, which is one of the most destructive hurricanes in the history of New Jersey. This storm was ranked as 1/900 year event, and the return period for this storm

Figure 6.6 Relationship between erosion rate and wave thrust.
tide has been estimated to be 1570 years, based on generalized extreme value return curves [Brandon et al., 2014].

The fact that erosion rates remain constant in time suggests that extreme events do not necessarily correspond to the occurrence of exceptionally high salt marsh erosion rates.

This could be due to a bypass mechanism such that, once salt marsh start being submerged, wave action significantly decreases due to wave energy dissipation on the top of marsh surfaces [e.g. Tonelli et al., 2010; Chen et al., 2011; Moller et al., 2014]. Therefore, even if the storm surge associated to hurricane Sandy created severe damages to the shore (for example to sandy beaches), it may have had a reduced impact to salt marshes which are more vulnerable when the water level is around mean sea level.

Constant erosion rates could be also connected to the absence of a critical threshold in wave energy above which salt marsh erosion rates drastically increases. In fact, the response of salt marsh to increasing wave energy has been frequently recognized as linear [e.g. Marani et al., 2011; Schwimmer et al., 2001]. These considerations highlight the potential of salt marshes as natural buffers against the action of violent storms.

The emergent relationship between salt marsh erosion and shoreline roughness suggests that the shape of marsh boundaries may be considered as the geomorphic signature of the magnitude of erosion events, and wetland-vulnerability to wind waves. Specifically, while rougher shorelines (high fractal dimension) correspond to slowly eroding salt marshes, smoother shorelines (low fractal dimension) correspond to rapidly degrading coastlines. The relationship between marsh boundary profiles and erosion rates is also recognizable from the frequency-magnitude distribution of erosion events. While
rapidly eroding shorelines are characterized by a Gaussian frequency-magnitude distribution of erosion events, slowly eroding shorelines display a logarithmic frequency magnitude distribution. Similar results have been found by Leonardi and Fagherazzi [2014, 2015] for salt marshes at five locations along the United States Atlantic Coast. It has been shown that, under different wave energy conditions, the different shape of the frequency-magnitude distribution of erosion events, as well as different marsh boundary rugosities are attributable to variability in soil resistance along marsh boundaries. Under a climate change scenario, a relationship between the erosion rate and the shape of marsh boundaries could be relevant as shoreline morphological feature could be potentially used to infer changes in external agents, such as changes in mean wave climate or in the frequency of extreme events.

### 6.6 Conclusions

We digitalized more than 100 km of marsh shoreline, and calculated salt marsh erosion rates for the Barnegat Bay and Little Egg Harbor system. We further computed shoreline fractal dimension, and the standard deviation of erosion events. Finally we correlated erosion rates to wave thrust values. Our analysis led to three main conclusions: i) salt marsh erosion rates remain constant in time and do not appear strongly affected by the occurrence of extreme events. The spatial distribution of salt marsh erosion rates remains constant as well. This suggests that, on average, salt marsh erosion rates are predictable. ii) The second results is that the magnitude of salt marsh erosion rates is recognizable from the morphology of the marsh boundary. Rapidly deteriorating salt marshes have
relatively smooth marsh boundary profiles. On the contrary, slowly eroding salt marshes have been found to have rougher marsh boundary profiles. Moreover, while average erosion rates appear predictable, individual erosion events may significantly differ from the average. Specifically, when average erosion rates are low, the frequency magnitude distribution of erosion events follows a long tailed distribution. This means that, occasionally, very big failure events far from the mean may happen. On the contrary, when erosion rates are high, erosion events are more predictable. iii) Finally, we confirm the existence of a relationship between salt marsh erosion rates and wave energy exposure. Specifically, we found a significant correlation between wave thrust and shoreline change.
6.7 References


Moller, I. et al. (2014), Wave attenuation over coastal salt marshes under storm surge conditions, Nature Geoscience, 7(10), 727–731.


FINAL REMARKS

The effect of wind waves and tides on deltaic wetlands is important and influence the hydrodynamic and morphological evolution of these coastal environments.

As an example, even very small tides can strongly impact the velocity field at distributary mouths, during both low and high flow regimes. Specifically, while during low discharge conditions the presence of a river discharge increases tidal damping and decreases tidal velocity amplitudes, during very high flow regimes the effect of tides at the distributary mouths is magnified with a noteworthy increase in tidal velocity amplitudes. In the presence of tides, mouth bars can also display alternate layers of coarse and fine material, due to velocity variations throughout tidal cycles. Lamination extent along mouth bar centerlines as well as differences in mud content between successive layers increases with tidal velocity amplitudes. Lamination extension also grows with decreasing ratios of mud over sand concentration and settling velocity, while the difference in mud content in successive layers has been found to increase with increasing settling velocity and concentration.

The effect of wind waves on salt marsh deterioration has been explored as well. Model results and field measurements suggest that sites exposed to high wave energy conditions are characterized by uniform rate of marsh boundary retreat, with erosion events following a Gaussian frequency magnitude distribution. In contrast, less exposed sites show a long tailed frequency magnitude distribution with numerous small events and few (but not negligible) bigger events, which are unpredictable and can happen despite of
relatively low wave energy conditions. Salt marshes exposed to high wave energy conditions are characterized by smooth marsh boundary profiles. In contrast, salt marshes subject to low wave energy conditions have rougher shoreline profiles.

Salt marshes have been also found to be very resilient to violent storms and hurricanes. As wave energy increases, salt marsh response to wind waves remain linear and there is not a critical threshold in wave energy above which salt marsh erosion drastically increases. Violent storms and hurricanes contribute to less than one percent to long-term salt marsh erosion rates. In contrast, moderate storms with a return period of around 2.5 months are those causing most salt marsh deterioration. Results suggest that even if salt marshes are constantly deteriorating at a slow rate, their predictable response to a wide range of storms and the possibility of forecasting both their life span and mitigation effects make these landforms well suitable for ecosystem based-coastal defense
BIBLIOGRAPHY


Kumar, N., G. Voulgaris, and J. C. Warner (2011), Implementation and modification of a three-dimensional radiation stress formulation for surf zone and rip-current applications, Coastal Engineering, 58(12), 1097–1117,

doi:http://dx.doi.org/10.1016/j.coastaleng.2011.06.009.


doi:http://dx.doi.org/10.1016/j.jhydrol.2005.04.003


Moller, I. et al. (2014), Wave attenuation over coastal salt marshes under storm surge conditions, Nature Geoscience, 7(10), 727–731.


Tomkins, K., McLachlan, G., Coleman, R., Hurst, T. & Glover, M., Coastal bank erosion data, Western Port Bay. v3. CSIRO. Data Collection. 102.100.100/12889 (2013).


White, C. D., B. J. Willis, S. P. Dutton, J. P. Bhattacharya, and K. Narayanan
Sedimentology, statistics, and flow behavior for a tide-influenced deltaic sandstone, Frontier Formation, Wyoming, United States In: G.M. Grammer and P.M. Harris, and G.P. Eberli (eds.) Integration of outcrop and modern analogs in reservoir modeling: AAPG Memoir 80, p. 129 – 152


NICOLETTA LEONARDI
Boston University
Department of Earth & Environment
675, Commonwealth Ave, Boston, MA, 02215
Tel: +1 (617) 353-4009
nicleona@bu.edu
nicoletta_leonardi@hotmail.it

ACADEMIC APPOINTMENT
Assistant Professor (Lecturer), University of Liverpool, UK - Liverpool Institute for Sustainable Coasts and Oceans, UK. (expected January 2016-).

EDUCATION
Ph.D. Earth and Environment, Boston University, (September 2012-current)
M.Eng. Civil and Hydraulic Engineering, University of Pisa (July 2012).
110 summa cum laude /110,
dissertation title: “Predicting river mouth bars growth in deltaic environment: the role of tides”
B.Eng. Civil Engineering, University of Pisa (December 2009).
110 summa cum laude /110,
dissertation title: “Sediment erosion of stilling basins downstream of rock river training structures: an experimental study”

TEACHING AND PROFESSIONAL EXPERIENCE
Teaching Fellow, ES105 Environmental Earth Science (undergraduate) Boston University, (Fall 2015).
Teaching Fellow, ES317 Introduction to Hydrology (undergraduate) Boston University, (Spring 2015).
Teaching Fellow/ Instructor, ES543 Estuaries and Nearshore Processes (undergraduate and graduate, part of the class requires students field work) Boston University, (Fall 2014).
Teaching Fellow, ES543 Estuaries and Nearshore Processes (undergraduate and graduate, part of the class requires students field work) Boston University, (Fall 2013).
Visiting Scholar, Earth Science Department, Boston University (January 2012-May 2012).
Laboratory assistant, Hydraulic engineering laboratory, University of Pisa (2010)
Security responsible for temporary and mobile construction works (Italian decree Law D.Legis 81/08 Ex. Law 494/96 University of Pisa (2009).
FELLOWSHIP AND AWARD

Director’s Award for Outstanding Teaching (2015).
This award recognizes “outstanding teaching achievements and teaching fellows who went above and beyond the call of duty in the teaching mission of Boston University”

National Center for Earth-surface Dynamics fellowship (2015): this fellowship covers the participation to the Summer institute on Earth-surface Dynamics (SIESD) at Tulane University to study: “Self-organization in landscapes and its residue in the stratigraphic record”

Long Term Ecological Research Network award, Luquillo, Puerto Rico, LTER site, (2015): this fellowship cover the participation to the LTER ASM meeting “Resistance, resilience, and vulnerability to high-energy storms: A gradient perspective” in Aspen, Estes Park, CO

Cum Laude distinction, University of Pisa (2012)

Denton Scholarship, Departmental Scholarship Boston University (2012),
This is conferred to individuals “whose application rose to the top of those submitted by a very strong group of candidates”

Cum Laude distinction, University of Pisa (2009).

PUBLICATIONS

Peer-reviewed journal publications:


Yao, H., Leonardi, N., Fagherazzi S. Sediment transport in a surface–advected estuarine plume (accepted: Continental Shelf Research).


(Figure 11 selected to appear as featured image on JGR: Oceans).

*Under Review*

Leonardi, N., Ganju N., K., and Fagherazzi S., The relationship between wind waves and salt marsh erosion is linear (under review; target: *Proceeding of the National Academy of Science*).

*In preparation (final draft; to be submitted soon)*

Leonardi, N., Ganju N., K., and Fagherazzi S., Development of a routine for marsh shoreline erosion and wave attack for COAWST

*Invited Talks*

Leonardi N., Using geometry to infer landscape stability, Utrecht University, NL (July, 2015).

Leonardi N., Building and Eroding Land, the role of marine processes, University of Exeter, UK (June 2015).


*Conference presentations and posters*

(* denotes oral presentation of the first author)*


Leonardi N.*, and Fagherazzi S., Ecougeomorphic Heterogeneity Sculpts Salt Marshes; AGU Fall meeting, *Coastal Geomorphology and Morphodynamics*, SFO, USA, 15 – 19 December 2014.


Leonardi N., Fagherazzi S., Ecogeomorphological variability arms salt marshes, VCR Long Term Ecological Research, LTR, All Scientist Meeting, VA, USA, March 2014.


Yu P., J. Littrell, A. Morey, A. Li, Leonardi, N., and S. Fagherazzi. Dune Morphology and Vegetation on Crane Beach, MA: Reaction and Evolution following Hurricane Sandy.


JOURNAL REVIEWS

SELECTED FIELD WORK


Apalachicola Bay (FL), (2013) Hydrodynamic survey, topographic survey, instruments deployment (ADCP, CTD, RBR), Sediment analysis.

Virginia Coast Reserve (VA), (2014) Topographic survey, vegetation monitoring, instruments deployment (Soil salinity sensors, light sensors).

TECHNICAL SKILLS AND COMPETENCES
Python, Matlab, Unix Shell scripting,
Topographic survey: technical instruments and data analysis,
ADCP, ADV, RBR: technical instruments and data analysis.

LANGUAGES
Italian (Mother tongue), English (Full professional proficiency).

DRIVING LICENSES
European driving license B, Massachusetts driving license D