1. INTRODUCTION

Salt marshes are delicate landforms at the boundary between the sea and land. These ecosystems support a diverse biota that modifies the erosive characteristics of the substrate and mediates sediment transport processes. Here we present a broad overview of recent numerical models that quantify the formation and evolution of salt marshes under different physical and ecological drivers. In particular, we focus on the coupling between geomorphological and ecological processes and on how these feedbacks are included in predictive models of landform evolution. We describe in detail models that simulate fluxes of water, organic matter, and sediments in salt marshes. The interplay between biological and morphological processes often produces a distinct scarp between salt marshes and tidal flats. Numerical models can capture the dynamics of this boundary and the progradation or regression of the marsh in time. Tidal channels are also key features of the marsh landscape, flooding and draining the marsh platform and providing a source of sediments and nutrients to the marsh ecosystem. In recent years, several numerical models have been developed to describe the morphogenesis and long-term dynamics of salt marsh channels. Finally, salt marshes are highly sensitive to the effects of long-term climatic change. We therefore discuss in detail how numerical models have been used to determine salt marsh survival under different scenarios of sea level rise.

Numerical modeling is one powerful tool that can be used to quantify the nonlinear feedbacks between salt marsh ecosystems, morphology, and sediment transport processes [Costanza and Voinov, 2004; Mcleod et al., 2010]. Numerical models can be used to test hypotheses regarding salt marsh processes, quantify the exchanges of energy and material across the intertidal landscape, and shed light on the long-term evolution and resilience of these systems.

Salt marsh models differ in the spatial scales that are considered, the processes that are simulated within the models, and ultimately, the output that is generated by the simulations [Rybczyk and Callaway, 2009]. At the smallest scale are the zero-dimensional models that simulate processes (i.e., net primary production and elevation change) at a single point within a marsh. Next are the models that simulate morphodynamics (i.e., sedimentation, channel development, and erosion) across a marsh platform (a two-dimensional model) or a marsh transect (a one-dimensional model). These models are said to be “ecogeomorphic” if they additionally consider the feedbacks between marsh vegetation and physical processes such as sedimentation and erosion. Finally, at the largest scale are the landscape models that simulate processes over entire coastlines or estuaries. In general, landscape models excel at simulating general trends at large spatial scales while the smaller-scale models often provide more mechanistic algorithms to simulate wetland processes.

Salt marsh models usually simulate long time scales, from years to centuries, and therefore particular attention must be devoted to integration errors, which accumulate in the results and predictions. To this end, empirical frameworks based on data collected at the temporal scale of interest are often more suitable, whereas mechanistic models based on a detailed description of the short-term physics might lead to long-term errors.

Here we present a synthesis of several approaches to salt marsh modeling. All models follow a general conceptual framework for salt marsh evolution (Figure 2) [Fagherazzi et al., 2004a; Ogden et al., 2005, Sklar et al., 1990]. First, we review physical processes; modeling sediment fluxes across the marsh platform and modeling marsh boundary and channel evolution. Next, because the presence or absence of halophytic vegetation on the marsh surface is fundamental to salt marsh evolution [Fagherazzi et al., 2004b], we review the simulation of aboveground and belowground production. We then shift to larger-scale simulations and review models that simulate coastal marsh evolution at the landscape scale. Finally, because much of the recent salt marsh modeling work has been in response to concerns regarding the effects of rising sea levels on salt marsh evolution and resilience [Kirwan et al., 2010], we review how numerical models have been used to determine the fate of salt marshes under different scenarios of sea level rise.

Figure 2. Simplified scheme of the interactions between ecology and geomorphology in salt marshes [after Fagherazzi et al., 2004a].
2.1. Empirical Models of Marsh Sedimentation

Empirical models start from field measurements of spatial sedimentation patterns (Figure 3) followed by statistical analyses in order to relate the observed sedimentation patterns to environmental variables [e.g., French et al., 1995; Leonard, 1997; Temmerman et al., 2003a; Van Proosdij, 2005b]. First, many studies have identified that sedimentation rates decrease with increasing platform elevation [e.g., Cahoon and Reed, 1995; Stoddart et al., 1989]. This is simply explained by the fact that lower portions of the marsh platform are flooded more frequently, higher and longer, so that more sediment is supplied and deposited. Second, platform sedimentation rates are found to decrease with increasing distance from tidal channels and from the seaward marsh edge [e.g., French et al., 1995; Leonard, 1997; Reed et al., 1999; Temmerman et al., 2003a], which may be explained by progressive sediment deposition along flow paths starting from the channels or marsh edge and directed to the inner portions of the marsh platform [Christiansen et al., 2000]. The underlying mechanism is that suspended sediments start depositing as soon as the flow reaches the marsh platform, where velocities are much smaller than in the tidal flow paths starting from the channels or marsh edge and directed to the inner portions of the marsh platform.
channel [D’Alpaos et al., 2007a]. Sedimentation is also
favored by the dense vegetation cover that exerts significant
friction and therefore rapidly decreases tidal current veloc-
ties and turbulence once the water flows from the channels
into the platform vegetation canopy [Christiansen et al.,
2000; Leonard and Luther, 1995; Yang, 1998]. Some field
studies have further highlighted the role of vegetation in
trapping sediments directly on aboveground plant structures
[Stumpf, 1983].

On the basis of the above described mechanisms, a
spatially explicit empirical sedimentation model has been
proposed by Temmerman et al. [2003b]. The model describes
the spatial variations in platform sedimentation rates using an
empirical data set of

\[ \text{SR} = k \cdot \text{e}^H \cdot \text{e}^{m \cdot D}, \text{e}^{n \cdot D_k}, \]  \hspace{1cm} (1)

where SR is the sedimentation rate (g m$^{-2}$ per time unit); $H$ is
the platform surface elevation (m relative to tidal datum); $D_o$ is the distance to the nearest tidal channel or marsh edge (m); $D_k$ is the distance to the nearest tidal channel or marsh edge (m); $H$, $D_o$, and $D_k$ are calculated from a remote sensing image from which the tidal
channel network and marsh edge are extracted. Most GIS
software programs offer algorithms to do this [see, e.g.,
Temmerman et al., 2003b]. Figure 3 shows an example of the
spatial implementation of this model for a specific tidal
marsh, illustrating that observed sedimentation patterns are
reasonably well reproduced [Temmerman et al., 2005a].

2.2. Physical Models of Marsh Sedimentation

[14] Rinaldo et al. [1999a] were one of the first to propose
a set of simplified hydrodynamic equations that describe the
two-dimensional depth-averaged flow field over a tidal
marsh platform. Their model basically assumes that the tide
propagates instantaneously (i.e., by immediate vertical
adjustment of a flat water surface) through the tidal channel
network dissecting the marsh platform, and that the flow on
the marsh platform is dominated by a balance between water
surface slope and friction. It is further assumed that the marsh
platform is flat, that the friction is constant in space and time,
that spatial variations in water surface above the platform are
much smaller than the average water depth, and that the
length of the marsh platform is much smaller than the tidal
wavelength. Under these assumptions, Rinaldo et al. [1999a]
reduced the shallow water equations to a Poisson approxi-
mation of the form [see Rinaldo et al., 1999a; Fagherazzi
et al., 2003]:

\[ \nabla^2 \eta_l = \frac{\lambda}{D_o} \frac{\eta_0}{\tau} \]  \hspace{1cm} (2)

where $\eta_l$ is the local deviation of the water surface from its
instantaneous average value, $\eta_0$, $\lambda$ is a constant bottom fric-
tion coefficient; and $D_0$ is the average water depth above the
platform. This model is used to calculate water surface slopes
above the marsh platform at any time during a tidal cycle and
to derive flow directions at any location above the platform
following the direction of steepest water surface slope.

Depth-averaged flow velocities are calculated from:

\[ \nabla \eta_l = -\frac{\lambda}{D} U. \]  \hspace{1cm} (3)

The model of Rinaldo et al. [1999a] has been used in
later publications to address several aspects of tidal marsh
morphodynamics, including the transport and deposition of
suspended sediments on the marsh platform [D’Alpaos et al.,
2007a]. Suspended sediment transport may be generally
modeled by an advection–diffusion equation of the form:

\[ \frac{\partial C}{\partial t} + \nabla \cdot (UCD - k_D \nabla C) = E - S \]  \hspace{1cm} (4)

where $C$ is the depth-averaged suspended sediment concentra-
tion; $D$ is the local water depth; $U$ is the local depth-
averaged flow velocity field; $k_D$ is a diffusion coefficient; $E$ is
the local erosion rate; and $S$ is the local sedimentation rate.

For cohesive sediments, $E$ and $S$ are generally modeled as
[Partheniades, 1965]:

\[ E = E_0 \left( \frac{\tau}{\tau_s} - 1 \right) \]  \hspace{1cm} (5)

\[ S = w_s C \left( 1 - \frac{\tau}{\tau_s} \right) \]  \hspace{1cm} (6)

where $\tau$ is the local bed shear stress; $\tau_s$ is the critical bed shear
stress for erosion ($E = 0$ when $\tau < \tau_s$); $E_0$ is an empirical
erosion coefficient; $\tau_s$ is a critical bed shear stress for sedi-
mentation ($S = 0$ when $\tau > \tau_s$); and $w_s$ is the settling velocity of
the suspended sediment. On a vegetated marsh platform,
$\tau < \tau_s$ in most cases, so that erosion may be neglected.

[15] D’Alpaos et al. [2007a] simulated the transport and
deposition of suspended sediment on a marsh platform
assuming that the tidal channels are the sources of the
suspended sediment. Their model results basically show that
simulated sedimentation patterns are governed by a decrease
in sedimentation rate with increasing distance from the
channels, as a consequence of progressive sediment settling
along simulated flow paths that are more or less perpendicu-
lar to the channels.

[16] While the simplified hydrodynamic scheme of Rinaldo
et al. [1999a] and D’Alpaos et al. [2007a] assumes that vir-
tually all water and sediment is supplied to the marsh plat-
form through the channels, field data have shown that
considerable water volumes (up to 60% of total volume) are
directly transported as sheet flow from the marsh edge
[French et al., 1995; Temmerman et al., 2005a]. Field data
further show that the partitioning of flow through the channels versus over flow over the marsh edge is controlled by the depth of flooding of the marsh platform. The deeper the marsh platform and its vegetation canopy are submerged, the greater the percentage of water that flows over the marsh as sheet flow rather than through the channels [Temmerman et al., 2005a]. Therefore the hydrodynamic scheme of Rinaldo et al. [1999a] and D’Alpaos et al. [2007a] is better suited to capture the initial stages of the marsh flooding and drainage, when the water depth on the platform is small. Since the peaks in channel velocity are reached around bankfull [French and Stoddart, 1992], this scheme provides excellent estimates of the formative discharge in the channel network. [17] Platform flow directions may also considerably change during single tides, and these changes seem to occur around the moment of submergence of the microtopographic relief or submergence of the vegetation canopy [Christiansen et al., 2000; Davidson-Arnott et al., 2002; Torres and Styles, 2007].

2.3. Coupling Vegetation and Marsh Sedimentation

[18] Temmerman et al. [2005b] presented a physically based model approach accounting for the interacting effects of inundation depth, vegetation canopy structure, and platform microtopography. They combined the three-dimensional hydrodynamic model Delft3D with a sediment transport model in presence of vegetation [Nepf, 1999]. In contrast to the work of Rinaldo et al. [1999a], this hydrodynamic model does not assume instantaneous vertical adjustments of a flat water surface in the tidal channels; instead the model explicitly simulates the interactions of tidal propagation through the channels and over the platform. The model of Rinaldo et al. [1999a] is therefore ideal in situations where a small tidal excursion and limited water depths on the marsh platform result in a simplified system behavior. On the other hand, high rates of change of tidal level give rise to a complex hydrodynamics characterized by sheet flow over the marsh platform and relevant fluxes from the marsh boundaries. [19] An important aspect of the model is that it accounts for the influence of the vegetation canopy on the vertical profile of the velocity and related drag, adding an extra source term of friction force, \( F(z) \), caused by rigid vertical plant structures to the momentum equations:

\[
F(z) = 0.5 \rho_0 \Phi(z)n(z)\left|U(z)\right|^2 U(z) \tag{7}
\]

where \( \rho_0 \) is the fluid density; \( \Phi(z) \) is the diameter of plant structures at height \( z \); \( n(z) \) is the number of plant structures per unit area at height \( z \); and \( U(z) \) is the horizontal flow velocity at height \( z \). The model further includes turbulence effects of the vegetation canopy. [20] The turbulence closure used in this model is a classical \( k-\varepsilon \) model, which is then modified to include extra source terms for \( k \) (turbulent energy generation) and for \( \varepsilon \) (turbulent energy dissipation) as a consequence of the vegetation. The extra source terms for \( k \) and \( \varepsilon \) are dependent on vegetation parameters such as the diameter of the stems and spacing in between the stems. The closure scheme was calibrated to measured turbulence data from a laboratory flume experiment with vegetation [see Bouma et al., 2007]. [21] It is important to stress that nonlinear friction due to vegetation is a very complex process, and equation (7) is just a simplified approximation trying to account for the presence of plant structures on the marsh platform.

[22] The model of Temmerman et al. [2005b] showed that the vegetation canopy has a crucial impact on the spatial flow and sedimentation patterns in a tidal marsh, while the influence of the platform microrelief (around 0.3 m) is minimal. When vegetation is considered, the simulated tidal flow is concentrated toward the channels, because friction is much lower in the bare channels than on the densely vegetated platform. Consequently, flood flow velocities are much higher in the channels (~0.6 m s\(^{-1}\)) than on the vegetated platform (~0.1 m s\(^{-1}\)). Given the size of the channels and their spacing on the marsh platform, most of the tidal flow rate is carried through the channels. Hence the platform is flooded from the channels with flow directions more or less perpendicular to the channel edges. In accordance with this flow pattern, simulated sedimentation rates decrease with distance from the channel edges (Figure 4b). Therefore, for vegetated marsh surfaces, the simplified model of Rinaldo et al. [1999a] is an excellent approximation of tidal hydrodynamics. In contrast, when vegetation is assumed to be absent, the flow is less concentrated in the channels, so that the speed of flood propagation through the channels and over the platform is comparable, and the platform is flooded in part from the seaward marsh edge. Consequently, simulated sedimentation patterns on the platform also depend on the distance from the marsh edge, leading to channel infilling (Figure 4c). Temmerman et al. [2005b] conclude that the presence of marsh vegetation has a profound impact on (1) the development of natural levees along channels and (2) the maintenance and even formation of dense channel networks [see also Temmerman et al., 2007]. [23] Temmerman et al. [2005b] further explain the occurrence of changes in platform flow directions during a single tidal cycle, as a consequence of gradual submergence of the vegetation canopy. At the onset of platform flooding, flow directions are always perpendicular to the channels. However, as the vegetation canopy is submerged, the relative difference in friction between the channels and platform decreases, so that larger-scale sheet flow from the marsh edge onto the platform becomes increasingly important. The model simulations show that this partitioning of flow through channels versus over the marsh edge is strongly controlled by inundation depth and by the height and density of the vegetation canopy. With deeper inundation depths, the percentage of water that is supplied through the channels decreases, and that is why distance from channels plays a minor role in explaining sedimentation patterns on low-lying marshes [Temmerman et al., 2005a]. Furthermore, the taller and denser the vegetation canopy, the more water that is supplied through the channels, which results in stronger...
developed turbulent flow given the quadratic dependence on velocity might be unrealistic for the entire tidal cycle, since the velocities over the marsh are such that the flow during slack water is likely in transition between a viscous and a turbulent regime.

3. MODELING MARSH BOUNDARY EVOLUTION

[26] Salt marshes develop in intertidal zones when conditions are sufficiently benign to allow for plant growth. When sediment supply is sufficient, salt marsh vegetation can accumulate extensive amounts of fine-grained sediment, which can result in the formation of a salt marsh plateau [Allen, 1989]. Consolidation of clays and silts on this plateau is strengthened by the rooting activities of the vegetation. Higher elevation and sediment stability improve plant growth, resulting in a positive feedback between increased sedimentation and increased plant growth.

[27] Because the remaining tidal flat is not accumulating sediment in equal amounts, the edge between the salt marsh and the tidal flat becomes increasingly steep and vulnerable to wave attack. This can cause the formation of strongly erosive marsh cliffs, which capture large amounts of wave energy as they are often near vertical. The combined effects of increased consolidation and direct protection by vegetation maintain a steep cliff that moves inland along a prolonged front (Figure 5). The front can be maintained for decades and destroy extensive areas of salt marsh as it translates landward [Van der Wal et al., 2008]. This process is a consequence of the inevitable increase in elevation of the marsh relative to the surrounding tidal flat and is crucial for understanding marsh dynamics [Van de Koppel et al., 2005]. Moreover, external forcing like large storms, sea level rise, and variations in sediment supply can strongly determine the evolution of the marsh scarp and influence the coupling between vegetation dynamics and morphology [Mariotti and Fagherazzi, 2010].

[28] Current research provides information that can be used to distinguish between endogenous and exogenous causes of marsh erosion. In locations where cliff erosion is an endogenous process, the tidal flat sometimes reemerges in front of the cliff, becoming again suitable for plant growth. This has been observed in a number of marshes along the Westerschelde Estuary, Netherlands [Van de Koppel et al., 2005; Van der Wal et al., 2008]. Thus the dynamics of vegetation patches in front of actively eroding salt marsh cliffs can be used as indicators of endogenous marsh erosion. Moreover, as young salt marsh vegetation can be (but not always) more diverse than the older marsh plateau, cliff erosion can be interpreted as complex natural dynamics that leads to salt marsh rejuvenation, thus maintaining structural and species biodiversity in salt marsh ecosystems.

[29] Mariotti and Fagherazzi [2010] presented a one-dimensional numerical model for the coupled long-term evolution of salt marshes and tidal flats. The model focuses on the migration of the boundary between the two landforms as a function of wind waves, sediment erosion, and deposition, as well as the effect of vegetation on sediment...
where vegetation might dictate the style of boundary erosion, the erodibility of the marsh margin to wave impact. How-

ever, vegetation does not have a major effect on inland reduction of the height and erosive power of waves propagating but included in the erodibility of the marsh boundary are not described in detail effects of vegetation roots and sediment characteristics on the waves are unable to affect the scarp. In this model the dynamics. Numerical simulations demonstrate that a vertical marsh slope forms during marsh retreat and that vegetation determines the rate of marsh progradation and regression (Figure 5). Mariotti and Fagherazzi [2010] relate the erosion of the marsh boundary to wave characteristics by:

\[ R = \begin{cases} 
0 & P < P_{cr} \\
\beta(P - P_{cr}) & P > P_{cr}
\end{cases} \]  

(8)

where \( R \) is the rate of boundary erosion; \( P \) is the wave power per surface unit dissipated by breaking at the marsh boundary; and \( P_{cr} \) is a threshold value for erosion, below which the waves are unable to affect the scarp. In this model the effects of vegetation roots and sediment characteristics on the erodibility of the marsh boundary are not described in detail but included in the \( \beta \) parameter. While vegetation clearly reduces the height and erosive power of waves propagating inland [Le Hir et al., 2007; Gedan et al., 2011], Feagin et al. [2009] argue that vegetation does not have a major effect on the erodibility of the marsh margin to wave impact. However, vegetation might dictate the style of boundary erosion, favoring the formation of a steep scarp, undercutting, and cantilever or toppling failure.

[30] Tonelli et al. [2010] used a numerical model solving the coupled Boussinesq-nonlinear shallow water equations to evaluate the effect of wave action on marsh boundaries as a function of tidal elevation and wave height. Results show that the wave thrust on the marsh scarp strongly depends on tidal level. The thrust increases with tidal elevation until the marsh is submerged and then rapidly decreases. Therefore, when the marsh is flooded, waves affect the marsh boundary less, and the maximum lateral erosion occurs when the water elevation is just below the marsh platform.

[31] Marsh cliff movements have also been shown to be independent of wave action, with debris that collects and stabilizes the cliff during periods without waves. Often sediments fill the spaces between blocks toppled or slumped from the scarp, forming a gentle slope [Allen, 1989].

[32] The effect of increased sea level on the dynamics of salt marsh edge erosion is still unknown. The amount of wave energy that can reach the edge of the cliff is determined to a large extent by the depth and slope of the tidal flat that often exists in front of a (sedimentary) marsh. Sea level rise will increase the relative depth of this tidal flat, and hence more wave energy will be imposed upon the salt marsh edge. Moreover, as the water level increases, it will become increasingly difficult for new vegetation to establish on the tidal flat that is exposed by the retreating edge [Fagherazzi and Wiberg, 2009; Mariotti et al., 2010].

[33] Simulations carried out by Mariotti and Fagherazzi [2010] indicate that a low rate of sea level rise increases wave dissipation and sediment deposition while a high rate of sea level rise leads to wave erosion and regression of the marsh boundary (Figure 6). Hence, there is the possibility that edge erosion becomes more severe, and recovery is hampered, by sea level rise, further squeezing salt marshes between increased human occupation at the landward side, and increased sea level at the estuarine side. More research is clearly needed to address this important point.

4. DYNAMICS OF MARSH CHANNELS

[34] A large body of literature exists describing salt marsh channel initiation and development [e.g., Yapp et al., 1916, 1917; Pestreng, 1965; Redfield, 1965, 1972; Beeftink, 1966; Gardner and Bohn, 1980; French and Stoddart, 1992; Steel and Pye, 1997], the hydrodynamics of salt marsh channels [e.g., Boon, 1975; Pethick, 1980; French and Stoddart, 1992; Rinaldo et al., 1999a, 1999b; Temmerman et al., 2005b; Fagherazzi et al., 2008], and their morphometric features [e.g., Fagherazzi et al., 1999; Rinaldo et al., 1999a, 1999b; Marani et al., 2003; Novakowski et al., 2004; Feola et al., 2005; Marani et al., 2006]. In spite of their fundamental role in the ecomorphodynamic evolution of salt marsh systems, only in the last few years have numerical models been developed to describe the morphogenesis and long-term dynamics of salt marsh channels [e.g., Fagherazzi and Furbish, 2001; Fagherazzi and Sun, 2004; D’Alpaos et al., 2005, 2006; Marciano et al., 2005; Perillo et al., 2005; Minkoff et al., 2006; Kirwan and Murray, 2007; Temmerman et al., 2010; Boon et al., 2010].
In one isolated case, the initiation and evolution of tidal channel networks have been described in the framework of a scaled laboratory model [Stefanon et al., 2010]. Although recent studies have helped refine our understanding of salt marsh channel dynamics, the dominant mechanisms and chief processes governing the initiation and development of these fundamental geomorphic features of the tidal landscape are not completely understood and still under debate. Moreover, more research is needed to quantify the sensitivity of the models to inevitable errors in the description and sediment transport processes, and how these errors might affect morphological predictions.

### 4.1. Tidal Channel Initiation and Development

It is generally agreed that the incision and subsequent elaboration of a channel network on a tidal platform is one of the chief morphological processes involved in the evolution of the tidal landscape. Tidal channel initiation can be ascribed to the concentration of tidal fluxes over a surface, which could be a sand or mudflat, or a terrestrial region that has been encroached by salt water due to sea level rise or breach opening on a littoral barrier. The concentration of tidal fluxes over the surface, possibly induced by the presence of small perturbations of bottom elevations, produces local scour as a consequence of the excess bed shear stress, thus favoring the initiation of drainage patterns characterized by shelving banks (see, e.g., Figure 7). Tidal fluxes further concentrate within the forming channel due to its increasing cross-sectional area and decreasing flow resistance as a result of the increase in the depth of flow within the channel [Fagherazzi and Furbish, 2001]. The increased flow velocity associated with reduction of the relative bottom roughness in the channel with respect to the adjacent marsh platform leads to higher bottom shear stresses and channel erosion. Consequently, erosion and deepening of the channel creates a positive feedback mechanism between erosion and channel formation which leads to the development of the incised tidal patterns.

The above described mechanism of channel initiation and development is outlined by observational evidence and by a number of conceptual and numerical models [e.g., Beeftink, 1966; French and Stoddart, 1992; Fagherazzi and Furbish, 2001; D’Alpaos et al., 2006; Temmerman et al., 2007]. During the earlier stages of the evolution of tidal

![Figure 6. Progradation and erosion rates of the marsh boundary computed with the model of Mariotti and Fagherazzi [2010] as function of RSLR and sediment concentration. Positive values indicate progradation, and negative values indicate erosion. (a) With vegetation. (b) Without vegetation [after Mariotti and Fagherazzi, 2010].](image)

![Figure 7. Sketch of the process of channel formation starting from a nearly flat bottom configuration. Small perturbations of bottom elevations enhance flux concentration, leading to bottom erosion and the initiation of a channel in which tidal fluxes further concentrate, thus increasing channel dimensions in a self-sustained process.](image)
channels, the tidal network develops via headward growth and tributary initiation through the carving of incised cross sections, where the local shear stress exceeds a critical shear stress for erosion. As the channels evolve and progressively drain larger portions of the marsh landscape and therefore capture a larger tidal prism, their cross-sectional areas expand to accommodate the increasing discharge.

[37] Field observations and related conceptual models describing the coupled evolution of salt marshes and channel networks support the above scenario [e.g., D’Alpaos et al., 2005; Kirwan and Murray, 2007; Temmerman et al., 2007]. As an example, Figure 8 shows some snapshots of the progressive development of salt marsh channels obtained by using the morphodynamic model of D’Alpaos et al. [2005] that is based on the simplified hydrodynamic model proposed by Rinaldo et al. [1999a, 1999b] (see section 2.2) which reduces the two-dimensional shallow water equations to a Poisson boundary value problem. The channels cut through an idealized rectangular domain, limited by impermeable boundaries on the top and on the lateral sides and flanked by an existing channel on the bottom side. Small incisions are initiated at sites along the bottom channel and then progressively grow because of the increase in flowing discharges enhanced by network development at sites in which the local bottom shear stress, controlled by water surface gradients, exceeds a threshold value for erosion. The model reproduces several observed characteristics of real tidal networks; however, the simulated channels are only statistically similar to natural ones [D’Alpaos et al., 2005]. In fact, only statistical parameters like drainage density, unchanneled length, and area probability distributions are correctly reproduced, whereas the exact location of each tidal channel can vary from simulation to simulation.

[38] The dynamics of the system is characterized by a competition among developing networks to capture the available watershed area, by the scouring of the channel cross sections due to the action of the flowing discharges, and by the feedbacks between network expansion and discharge concentration at the tips of the network. Similar results have also been obtained by Fagherazzi and Sun [2004] and Kirwan and Murray [2007] utilizing process-oriented models that account for the role of local gradients of a Poisson–parametrized water surface [Rinaldo et al., 1999a], an approach which is particularly suitable for shallow tidal areas of limited extent. Marciano et al. [2005] used the Delft3D hydrodynamic model, coupled with sediment transport, to simulate the formation of large-scale tidal patterns in a short tidal basin.

[39] The process of network incision is agreed to be rather rapid: Steers [1960] reported a channel headcut growth of up to 5–7 m yr\(^{-1}\), Collins et al. [1987] observed a headward erosion of more than 200 m in 130 years, Wallace et al. [2005] measured a mean extension rate of 6.2 m yr\(^{-1}\), and D’Alpaos et al. [2007b] documented a mean annual rate of headward growth of about 11 m yr\(^{-1}\). After an initial stage of rapid evolution, which gives the network a basic imprinting, the network structure undergoes a slower elaboration and is characterized by the adjustment of channel geometry to variations in the local tidal prism through network contractions and expansions. At equilibrium, the maximum bottom shear produced by tidal currents is just below the critical value for erosion at the channels tips, and therefore the channels do not extend any further. At this point the network displays a distribution of unchanneled lengths with an exponential trend [D’Alpaos et al., 2005, 2007b].

[40] The presence of vegetation may promote channel incision [D’Alpaos et al., 2006; Temmerman et al., 2007] and influence the planimetric evolution of tidal channels because of its stabilizing effects on surface sediments and channel banks [e.g., Redfield, 1972; Garofalo, 1980; Gabet, 1998; Marani et al., 2002; Fagherazzi et al., 2004b; Kirwan et al., 2006; Alpaos et al., 2005, 2007b].
Redfield [1965] reports on changes in tidal channel cross-sectional geometry, through deepening and narrowing, as a consequence of the vertical accretion and horizontal progradation of the adjacent vegetated marsh surface. Such an observation supports the concept of inheritance of the major features of channelized patterns dissecting the salt marshes from previously existing sand or mudflat underlying marsh deposits [e.g., Redfield, 1965; Allen, 2000; Marani et al., 2003]. Following Redfield [1965], Hood [2006] proposed a conceptual model of channel growth suggesting that tidal channels might be the result of depositional rather than erosional features, in a rapidly prograding delta. In particular environments (Bahía Blanca Estuary, Argentina) the interaction between crabs (Neohelice granulatus) and halophytic plants (Sarcocornia perennis) favors the formation of salt marsh creeks [Perillo et al., 2005; Minkoff et al., 2006]. The bioturbation effects linked to crab-plant interactions exerts, in this case, a relevant role in driving the development of salt marsh creeks, thus overcoming water surface gradients.

4.2. Scaling Properties of Tidal Networks

Tidal networks display basic geometric properties common to natural terrestrial patterns [e.g., Pestrang, 1965; Fagherazzi et al., 1999; Steel and Pye, 1997; Novakowski et al., 2004] but lack the scale invariance features that are peculiar to fluvial channel networks [Rinaldo et al., 1999a, 1999b; Marani et al., 2003; Feola et al., 2005]. A marked absence of scale-free distributions implies that there is no similarity of the part and the whole within the tidal landscape, in sharp contrast to what happens in fluvial basins where ubiquitous power laws occur [Rodríguez Iturbe and Rinaldo, 1997]. This reflects the many conflicting processes acting at overlapping spatial scales that affect the relevant morphodynamics, thus hindering simple geometric relationships of the types observed in fluvial basins to hold throughout the actual range of tidal scales. Analyses claiming the contrary may indeed have been misled by the subtleties of network comparisons [e.g., Novakowski et al., 2004]. As pointed out by Hack [1957], fluvial basins tend to become longer and comparatively narrower as their size increases:

\[ L \sim A^\theta \]  

where \( A \) is the total contributing area at any site of a fluvial basin; \( h \) is an exponent typically equal to 0.57; and \( L \) is the embedded mainstream length, defined as the longest streamwise distance, measured along the network from the outlet to the watershed divide. Hack’s law does not seem to be applicable to tidal networks [Feola et al., 2005]. Another relevant morphological measure is introduced by computing the downstream unchannelled length, \( \ell \), evaluated along the flow streamlines determined through the steepest descent direction of the water surface from every salt marsh site to the nearest channel [Marani et al., 2003]. The properties of the probability density function of this length, \( \ell \), are physical indices of network capability to drain the basin and thus provide an appropriate definition of drainage density. A clear tendency to develop watersheds described by exponential decays of the probability distributions of unchannelled lengths has been observed for about 140 watersheds within 20 salt marshes in the lagoon of Venice, thereby confirming the existence of scale-free features.

[44] Salt marsh channels are also highly sinuous, with meanders that are geometrically similar to their fluvial counterparts [Marani et al., 2002; Solari et al., 2002; Fagherazzi et al., 2004b], although displaying characteristics, such as meander sinuosity, which appear to vary not only from one salt marsh to another, but also within distances of a few hundred meters [Marani et al., 2002]. Strong spatial gradients in discharge favor the development of relevant spatial gradients of characteristic geometric features (e.g., chiefly wavelength and width) whereas the bidirectionality of the discharge shaping the tidal channels often implies a meander evolution that departs from terrestrial meanders [Fagherazzi et al., 2004b]. Solari et al. [2002] indicate that tidal oscillations give rise to symmetric oscillations of the point bar-pool pattern around the locations of maximum curvature, without triggering a net migration of the meander if the tide is periodic with zero mean. Moreover, salt marsh channels tend to be characterized by stable planform configurations, with channel migration rates consistently slower than those experienced by fluvial rivers. This is particularly the case of channels cutting through densely vegetated platforms. Dense vegetation, in fact, tends to freeze lateral channel migration while bank undercutting and slumping favors the formation and growth of meanders [Redfield, 1972; Garofalo, 1980; Gabet, 1998; Marani et al., 2002; Fagherazzi et al., 2004b]. Block collapse through a combination of cantilever and toppling failures produces bank migration [Allen, 1989], whereas the persistence of failed bank material, which temporarily protects the bank from erosion, decreases the rates of lateral migration [Gabet, 1998].

5. COUPLING VEGETATION AND SEDIMENTARY PROCESSES IN SALT MARSHES

Salt marsh macrophytes maintain the elevation of marshes by trapping organic sediments [e.g., Gleason et al., 1979; Leonard and Luther, 1995; Li and Yang, 2009] and through direct deposition of organic sediments [e.g., Turner et al., 2001; Nyman et al., 2006; Langley et al., 2009; Neuhaus, 2008]. Both trapping and organic deposition is positively correlated with plant biomass [e.g., Gleason et al., 1979; Li and Yang, 2009; Morris et al., 2002], which is controlled, in part, by the elevation of the marsh platform [e.g., Morris et al., 2002]. Thus there is a strong feedback between the elevation of salt marshes and marsh vegetation (Figure 9).

[46] In salt marsh models all these feedbacks are expressed through process-based equations, which can be implemented in a code using numerical schemes having different levels of approximation and complexity. Here we provide a detailed description of the key equations used for the coupling between vegetation and sedimentary processes.
Figure 9. Schematic showing the relationship between mean high tide, biomass, and the roots:shoots ratio. The cartoons are based on measurements at North Inlet, South Carolina [Morris et al., 2002; Mudd et al., 2009].

5.1. Feedbacks Between Marsh Vegetation and Platform Elevation

The biomass of salt marsh macrophytes is a function of a number of factors, including tidal amplitude [e.g., Kirwan and Guntenspergen, 2010], latitude, temperature [McKee and Patrick, 1988; Kirwan et al., 2009], sediment supply [Fragoso and Spencer, 2008], and CO₂ concentration [Langley et al., 2009]. In a single estuary, however, there is a distinct elevation range that is occupied by marsh vegetation [e.g., Redfield, 1972; Orson et al., 1985; Morris et al., 2005], with marsh vegetation occupying elevations approximately between mean sea level and mean high tide [McKee and Patrick, 1988; Kirwan and Guntenspergen, 2010] (Figure 9). The biomass and productivity of macrophytes varies strongly within this window [e.g., Morris et al., 2002; Spalding and Hester, 2007].

In a given estuary with relatively constant tidal amplitude, temperature, and sediment supply, marsh elevation is the dominant factor in determining plant biomass [e.g., Morris et al., 2002]. For example, Morris et al. [2002] found that at North Inlet, South Carolina, Spartina alterniflora was most productive at sites 55 cm below mean high tide (in a location where the tidal amplitude was ~60 cm).

Because plants are most productive at some optimum elevation in relation to mean high tide, a negative feedback between plant growth, sea level rise, and sedimentation can occur [e.g., Morris et al., 2002]. If the marsh elevation is lower than the optimum elevation, an increase in the depth of flooding during tides leads to a decrease in plant productivity and therefore a decrease in sedimentation.

On the basis of these observations, Morris et al. [2002] put forward the following equation relating standing biomass of halophyte vegetation, B, to the difference between mean high tide and marsh elevation, D:

\[ B = aD + bD^2 + c \]  \hspace{1cm} (10)

where the parameters a, b, and c depend on vegetation type and marsh location.

This equation provides a simple quantitative feedback between salt marsh ecology (vegetation biomass) and morphology (marsh elevation), and forms the basis for several models of salt marsh evolution [Mudd et al., 2004; D’Alpaos et al., 2005; Morris, 2006; Kirwan and Murray, 2007; Mariotti and Fagherazzi, 2010].

Following this model, an increase in the rate of sea level rise would lead to the drowning of the marsh; eventually the marsh would become too deep for vegetation to survive. Fagherazzi et al. [2006] and Marani et al. [2007] concluded that this negative feedback is responsible for the bimodal distribution of elevations in typical estuaries with vegetated marshes situated near mean sea level and unvegetated mudflats below.

The model of Morris et al. [2002] was derived in tidal marshes dominated by Spartina spp. Elsewhere, for example, in Mediterranean and northern European marshes, interspecific competition among numerous halophytic species determines biomass to be an increasing function of elevation [Marani et al., 2004; Silvestri et al., 2005].

For example, Marani et al. [2010] use a logistic model for the vegetation biomass:

\[ \frac{dB}{dt} = \frac{r(z)B}{d} (d - B) - m(z)B \]  \hspace{1cm} (11)

where the biomass B = pd is expressed as the product of vegetation fractional cover, p, and the carrying capacity of the system, d (maximum biomass per unit area); and r and m are the reproduction and mortality rates of the halophytic plants, which depend on the marsh elevation z. By assigning specific functions to r and m the two observed trends of vegetation biomass reaching a maximum for a determined elevation [Morris et al., 2002] or always increasing with elevation [Silvestri et al., 2005] can be simulated within the same framework.

5.2. Inorganic Sedimentation Enhanced by Marsh Vegetation

The presence of plants on the marsh surface can enhance inorganic sedimentation. Gleason et al. [1979] measured sedimentation rates as a function of stem density in an
Fluvial deposits and a vegetation belt are collected at the interface of the tidal marsh and the river. The vegetation reduces the velocity of the incoming flow and modifies the hydrodynamic conditions of the flow. The reduction of velocity leads to a decrease in turbulence and an increase in sedimentation. The drag coefficient, which is a function of the vegetation density, is also reduced. The equation for the drag coefficient is given by:

\[ C_D = 2 \left( \frac{\alpha v}{Re} + \alpha_1 \right), \]

where \( \alpha_v \) is a coefficient reported to be 0.9 by Nepf [1999] and \( u \) is the flow velocity on the marsh. In equation (12) the bulk drag coefficient \( C_D \) decreases as the element spacing decreases or \( ad \) increases, therefore reducing the TKE when a thick vegetation canopy is present. [58] In fact, the drag coefficient is also related to the characteristics of marsh vegetation. Tanino and Nepf [2008] found that the drag coefficient in an array of emergent cylinders can be described as:

\[ C_D = 2 \left( \frac{\alpha_v}{Re} + \alpha_1 \right), \]

where \( \alpha_1 \) is a function of the solid fraction of stems within the flow. The solid fraction is the area of a single cylinder in cross section times the number of cylinders per unit volume, or \( \pi d_a^4 / 4 \). The Reynolds number based on the plant stems, \( Re_c \), is a function of flow velocity, the diameter of the stems, and the kinematic viscosity of water: \( Re_c = u d_a / \nu \). The stem diameter and projected area per unit volume of marsh vegetation have been found to follow power law functions of biomass [Mudd et al., 2004, 2010]:

\[ a_c = \alpha B^{3/4} \]

\[ d_c = \mu B^{1/4} \]

where \( B (\text{M L}^{-2}) \) is the biomass per unit area of marsh macrophytes and \( \alpha, \beta, \mu, \) and \( \varphi \) are empirical coefficients. [57] Turbulence helps to maintain particles in suspension so reductions in turbulent kinetic energy may enhance particle settling on salt marshes [e.g., Leonard and Luther, 1995; Nepf, 1999; Christiansen et al., 2000; Leonard and Croft, 2006]. The upward velocity of sediment particles, \( w_{up} \), can be determined by the Rouse equation [Mudd et al., 2000; Orton and Kineke, 2001]:

\[ w_{up} = \kappa w \left( \frac{\tau}{\rho_w} \right)^{1/3} \]

where \( \kappa \) is a constant and \( \tau \) is the shear stress. The project area of the plants per unit 872 volume, \( \alpha \):

\[ K = \alpha^2 u^2 (C_D a_d)^{2/3}, \]
where \(\kappa\) is von Karman’s constant, assumed to be 0.38, \(\tau\) is the shear stress, and \(\rho_w\) is the density of water. Several authors correlated shear stress to total kinetic energy:

\[
\tau = \omega k
\]

where \(\omega\) is a constant of proportionality found to vary between 0.19 and 0.21 [Kim et al., 2000; Soulsby and Dyer, 1981; Stapleton and Huntley, 1995].

6. MODELING BELOWGROUND ORGANIC PRODUCTION

6.1 Particle settling can then be calculated by subtracting the upward particle velocity caused by turbulence from the settling velocity of particles in still water. Turbulence is produced in stem wakes [Nepf, 1999; Neumeier, 2007], but this is counteracted by a reduction in velocity due to increased drag. The net effect of denser vegetation is to reduce turbulent kinetic energy [e.g., Mudd et al., 2010].

6.2 Particles suspended in tidal waters can also be captured directly by plant stems and leaves [e.g., Palmer et al., 2004; Li and Yang, 2009]. Li and Yang [2009] measured particle capture on a Spartina alterniflora marsh near Shanghai, China, and found it to increase with increasing stem density. Palmer et al. [2004] conducted laboratory experiments and determined that particle capture was a function of stem diameter, stem density particle diameter, flow velocity, the concentration of suspended sediment (C), and flow depth. Because stem diameters and stem densities can be related to aboveground biomass (B) [e.g., Morris and Haskin, 1990; Mudd et al., 2004, 2010], the rate of mass directly captured by plant stems per unit area of the marsh (Qc) can also be related to biomass and flow characteristics.

\[
Q_c = \frac{NK}{V} \mu^\gamma \sigma C \mu^\psi B^{\phi/(\gamma - \phi)}d_p \frac{1}{\alpha},
\]

where \(h\) is the flow depth; \(\kappa\), \(\gamma\), and \(\sigma\) are empirical coefficients; and \(d_p\) is the diameter of suspended sediment particles. Using a numerical model that accounted for both plant capture and enhanced settling due to marsh vegetation, Mudd et al. [2010] concluded that in typical marshes (with flow velocities \(\leq 0.2\) m s\(^{-1}\)), settling will dominate inorganic sedimentation.

Currents on the marsh platform are typically too weak to cause erosion [e.g., Christiansen et al., 2000; Wang et al., 1993], but vegetation can focus flow around patches [e.g., Bouma et al., 2009] and in channels [e.g., Temmerman et al., 2005a, 2007], leading to enhanced erosion.

6.3 Plant biomass affects the accumulation of sediments and subsequent salt marsh evolution by trapping mineral and organic particles previously suspended in the water column, by contributing aboveground plant litter to the surface, and by the direct inputs of belowground organic matter as the result of belowground root production, turnover, and decomposition [Morris and Bowden, 1986]. Most salt marsh models, however, ignore belowground production.

[64] It is clear that, under certain conditions, belowground production, and subsequent organic matter accumulation, can account for a relatively large fraction of marsh accretion. Callaway et al. [1997] measured organic content and \(^{137}\)Cs-derived accretion rates in three Spartina alterniflora dominated salt marshes in the northern Gulf of Mexico and found that organic accretion ranged between 7.7% and 21.8% of total accretion (by mass). Nyman et al. [2006] found that organic accretion could reach up to 40% of the total accretion rate in Spartina alterniflora and Spartina patens dominated saltwater sites in Louisiana. Chmura et al. [2003] reviewed 24 studies of organic matter accumulation on 85 salt marsh sites across a broad geographic area (the Gulf of Mexico, the northeastern Atlantic, the Mediterranean, the northeastern Pacific, and the northwestern Atlantic) and determined an organic matter accumulation rate as high as 1713 g m\(^{-2}\) yr\(^{-1}\) at a site in Louisiana.

[65] Despite the relative lack of data, an expanding interest in issues in which belowground processes are critically important (i.e., salt marsh evolution and marsh sustainability in the face of sea level rise) has led to the development of some evolution models that do simulate belowground production. To do so, the modeler is confronted with three challenges: (1) the development of algorithms that describe the rate of belowground production, (2) simulating the distribution of roots within the sediment column, and (3) the development of algorithms that define the rate and proportion of root-derived organic matter that is incorporated into the sediment matrix and contributes to marsh elevation change.

[66] The rate at which dead roots are incorporated into marsh sediments is related to aboveground biomass, and can be stated as:

\[
M = G - \frac{\partial B_{ag}}{\partial t}
\]

where \(M\) is a mortality rate in (M L\(^{-2}\) T\(^{-1}\)); \(G\) is a growth rate (M L\(^{-2}\) T\(^{-1}\)); and \(B_{ag}\) is aboveground biomass. Belowground mortality can then be calculated by determining the roots:shoots ratio of the dominant marsh macrophyte under a variety of environmental conditions. Mudd et al. [2009] reported that belowground biomass of Spartina alterniflora increased with increasing aboveground biomass in North Inlet, South Carolina. In contrast, the roots:shoots ratio at that site was inversely related to aboveground biomass [Mudd et al., 2009] (Figure 9); this relationship could be approximated by a linear function:

\[
\frac{B_{bg}}{B_{ag}} = \theta_{bg}D + D_{mbn}
\]

where \(B_{bg}\) is belowground biomass and \(\theta_{bg}\) and \(D_{mbn}\) are the slope (L\(^{-1}\)) and the intercept (dimensionless) of the relationship between the roots:shoots ratio and the depth below MHHW. This relationship is likely site-specific: Darby and Turner [2008a, 2008b] found that fertilization of Spartina alterniflora in Louisiana resulted in increased aboveground biomass but had no effect on belowground biomass. Because aboveground biomass is related to the elevation of the marsh...
platform, this elevation also controls belowground biomass. 

Nyman et al. [2006], for example, found that root growth 
increased if flooding depth increased, but did not report 
aboveground biomass at their sampling sites. 

Of the salt marsh models that mechanistically simulate 
belowground production within a sediment column, most 
[i.e., Callaway et al., 1997; Rybczynk et al., 1998; Cahoon 
et al., 2003; Mudd et al., 2009; Kairis and Rybczynk, 2010] 
are derivative of Morris and Bowen’s [1986] single-year 
sediment cohort model, originally developed for a freshwater 
tidal marsh on the North River, Massachusetts. One advantage 
of the cohort approach is that this framework can simulate the percent sediment organic matter and bulk density 
with depth, and this output can be compared to actual sediment cores for model calibration and validation. In Morris and Bowen’s [1986] original model, simulated live root biomass within a vertical sediment column was assumed to be greatest at the surface and to decrease exponentially with depth and defined as:

$$ R = R_0 e^{-kD} $$  \hspace{1cm} (21)

where $R$ is the root biomass (g cm$^{-2}$) at depth $D$ (cm); $R_0$ equals the weight of roots at the sediment surface (g cm$^{-2}$); and $k$ is root distribution parameter (cm$^{-1}$). The parameter $k$ essentially describes the exponential root distribution with depth (greatest at the surface). The integrated root biomass over all depths is defined as the integral of equation (21):

$$ R_t = \int R_0 e^{-kD} dD $$  \hspace{1cm} (22)

where $R_t$ is defined as the total live root biomass in the soil column (g cm$^{-2}$). By specifying $R_t$ (usually as a function of aboveground biomass) and assuming that all live belowground biomass is contained with a known rooting depth, one can then use equation (22) to solve for $R$, and $k$ [Morris and Bowen, 1986].

Labile organic matter, because it decays, does not contribute to marsh accretion, and deposition of low-density, uncompacted organic sediments is offset by compaction. The change in organic carbon can be modeled as:

$$ \frac{dC_l}{dt} = k_l C_l + m \chi_l $$  \hspace{1cm} (23)

where the subscript $l$ refers to the labile pool; $C$ (M L$^{-3}$) is the organic carbon per unit volume; $m$ (M L$^{-3}$ T$^{-1}$) is the mortality rate per unit volume ($M$ is equal to $m$ integrated over the depth of the rooting zone); $k_l$ is the decay coefficient; and $\chi_l$ is the fraction of dead root matter that is labile. Decay rates depend on a number of factors, including (possibly) sulfate concentration and oxygen supply [e.g., Silver and Miya, 2001]; many of the factors proposed to control the rate of organic matter decay vary with the depth below the 45 sediment surface. Some authors have suggested depth-dependent decay coefficients [Conn and Day, 1997; Rybczynk et al., 1998]. Others, however, have conducted measurements of decay in marsh sediment that show no depth dependence [e.g., Blum, 1993].

Compaction can be modeled as [e.g., Gutierrez and Wangen, 2005]:

$$ E = E_0 - C_l \log \left( \frac{\sigma_{off}}{\sigma_0} \right) $$  \hspace{1cm} (24)

where $E$ (dimensionless) is the void ratio; $C_l$ (dimensionless) is the compression index; $E_0$ (dimensionless) is the void ratio at the reference stress, $\sigma_0$ (M T$^{-2}$ L$^{-1}$); and $\sigma_{off}$ (M T$^{-2}$ L$^{-1}$) is the effective stress. The long-term rate of vertical accretion, in units of length per time, is determined by dividing the rate of accumulation of refractory organic material by the density of compacted organic material [e.g., Mudd et al., 2009]. Assuming that all root material is made of refractory carbon and that compressed organic material has a density of 0.1 g cm$^{-3}$, the maximum vertical accretion rate from organic sediments is approximately 1.7 cm yr$^{-1}$, based on Chumura et al.’s [2003] highest reported organic accumulation rate. Mudd et al. [2009], using a model that incorporated compaction, organic decay, and measured productivity and mortality of Spartina alterniflora at North Inlet, South Carolina, United States, calculated a theoretical maximum organic accumulation rate of ~2200 g m$^{-2}$ yr$^{-1}$.

Kirwan et al. [2009] found that among North American marshes, Gulf Coast and southwest Atlantic marshes were the most productive, corroborating the results of Chumura et al. [2003]. Kirwan et al. [2009] estimated that an increase of 4°C could boost productivity by up to 40%, but even if this increase in productivity was mirrored in the production of belowground biomass, this would mean a maximum rate of vertical accretion from organic sediment of 2.4 cm yr$^{-1}$. This productivity gain could potentially be enhanced because of greater atmospheric CO$_2$; Langley et al. [2009] compared the production of fine roots in plots with ambient CO$_2$ and plots of CO$_2$ of ambient +340 ppm and found that fine root production increased by 75% and 35% over two field seasons in the plots with elevated carbon dioxide. Again assuming all of this additional root matter was refractory carbon, and assuming a compacted density of organic matter of 0.1 g cm$^{-3}$, one could estimate the maximum possible vertical accretion rate due to organic sedimentation as 4.2 cm yr$^{-1}$. This rate, however, should be considered extreme as organic sediments are highly compressible [e.g., Mesri et al., 1997] and organic material typically contains 20–90% labile carbon, as indicated by the steep decline in organic material content with depth found in virtually all marsh cores [e.g., Sharma et al., 1987].

**7. SALT MARSH LANDSCAPE-SCALE ECOSYSTEM MODELING**

The objective of an ecosystem-based landscape model is to minimize the computation of physical processes in order to expand the resolution and forecast of the resultant effects on biological systems. Given their relative simplicity, ecosystem-based models can be applied at high resolution to large areas and study the spatial interactions among different ecosystem units. The use of spatially explicit models thus
1129 where change in plant biomass is the result of plant produc-
1130 tions during long-term sea level rise (SLR) and has been 
1131 [78] Simulation results can be combined with site-
1132 dependent ecosystem services to coastal landscapes and 
1133 communities [Craft et al., 2009]. 
1134 [79] SLAMM was used to simulate the effects of SLR 
1135 on carbon (C) sequestration and water quality improvement 
1136 (denitrification) in the Altamaha River Estuary, Georgia, 
1137 United States. The simulation used the IPCC A1B mean sea 
1138 level rise scenario predicts that forest and tidal freshwater 
1139 marsh declined by 24% and 38%, respectively, by 2100 
1140 (Figure 11). Salt marsh is predicted to decline by 8% whereas 
1141 brackish marsh area increases by 4%. The model predicts 
1142 large increases in transitional salt (high) marsh, tidal flat, and 
1143 estuarine open water as sea level rises. Delivery of ecosystem 
1144 services related to water quality improvement (denitrifica-
1145 tion) and carbon (C) sequestration declines as salt marsh 
1146 is submerged and forest and freshwater marsh convert to 
1147 brackish marsh habitat. 
1148 [80] Overall, in the estuary model, results show that tidal 
1149 wetland area is reduced by 12%, whereas denitrification 
1150 and C sequestration are reduced by 10% and 19%, respec-
1151 tively. The disproportionately large decrease in denitrifica-
1152 tion is attributed to loss of tidal forests which have high 
1153 rates of denitrification relative to the more saline wetlands. 
1154 The results of this example illustrate one way of coupling 
1155 ecosystem-based studies of wetland structure and function 
1156 with landscape models such as SLAMM to predict how 
1157 the delivery of wetland ecosystem services will change in 
1158 response to sea level rise in the coming century. 
1159 
1160 8. SALT MARSH EVOLUTION AND SEA LEVEL RISE 
1161 [81] At the most fundamental level, a marsh must gain 
1162 elevation at a rate faster than or equal to the rate of sea level 
1163 rise to maintain its vertical position in the intertidal zone 
1164 [Reed, 1995]. Historically, rates of sediment deposition and 
1165 organic accretion have been similar to rates of sea level rise in 
1166 most marshes worldwide. Long-term accretion rates derived 
1167 from measurements of Pb-210 indicate that sea level rates 
1168 and accretion rates have been similar over approximately the 
1169 last century and that accretion rates are fastest in regions with 
1170 
1171 15 of 28
1212 rapid sea level rise [Friedrichs and Perry, 2001; French, 2006]. Direct measurements of elevation change on annual to decadal time scales also suggest a connection between marsh accretion and sea level; most salt marshes and mangroves around the world are accreting faster than or equal to the rate of historical sea level rise plus any local subsidence [Cahoon et al., 2006]. Indeed, sediment cores from many marshes reveal stratigraphic properties that change little over a few thousand years during periods of relatively slow sea level rise [e.g., Redfield, 1965].

8.1. Feedbacks Between Accretion, Submergence, and Sea Level Rise

1224 [82] Observations of long-term stability and platform maintenance inspire conclusions that marsh ecosystems must be capable of rapidly adjusting to changes in rates of sea level rise [Friedrichs and Perry, 2001]. Several processes, both biological and physical, are likely responsible for the tight coupling between sea level and marsh accretion. From a purely physical perspective, sediment deposition rates on the marsh platform are largely controlled by the duration and frequency of tidal flooding [Marion et al., 2009]. Rates of mineral deposition are highest in low-elevation salt marshes that are inundated for long periods of time and lowest in high-elevation marshes that are periodically flooded [Pethick, 1981; Bricker-Urso et al., 1989]. Similarly, rates of deposition at a single location within a salt marsh are highest when tides and inundation depths are highest [Temmerman et al., 2003b]. Therefore, if an increase in the rate of sea level rise is accompanied by more extensive platform flooding, mineral deposition rates will increase. More recent work suggests that this feedback is enhanced by growth characteristics of marsh macrophytes. Long-term measurements of Spartina alterniflora, for example, demonstrate that its productivity is strongly correlated to interannual variations in sea level [Morris et al., 2002] and that it grows fastest at relatively low elevations within the intertidal zone [Mudd et al., 2009]. Feedbacks between flooding and accretion are less understood for organic rich marshes, though organic matter decay rates are likely slower in frequently flooded, anaerobic soils. Finally, sediment eroded from one portion of a marsh can be a source of sediment aiding vertical accretion in surviving marshland. For example, eroding marsh edges often have levees just behind the point of wave impact. Similarly, expansion of channel networks in response to accelerated sea level rise may deliver more sediment to portions of the platform that were previously sediment deficient [D’Alpaos et al., 2007a; Kirwan et al., 2008]. These types of ecogeomorphic feedbacks likely explain the persistence of wetlands within the intertidal zone over thousands of years in the stratigraphic record [Redfield, 1965] and observations of accretion rates that are highest in regions with historically high rates of sea level rise [Friedrichs and Perry, 2001].

[84] Nevertheless, widespread observations of marsh submergence today indicate that there are limits to the ability of ecogeomorphic feedbacks to preserve the position of a marsh
1267 within the intertidal zone. Stratigraphic evidence and tidal
1268 gauge records indicate that sea level rise rates were less than
1269 1 mm yr\(^{-1}\) for most of the last 2000 years and began accel-
1270 erating toward modern rates of about 2–3 mm yr\(^{-1}\) in the
1271 18th or 19th centuries [Donnelly et al., 2004; Church and
1272 White, 2006; Jevrejeva et al., 2008; Gehrels et al., 2008].
1273 Perhaps in response, marshes around the world appear to be
1274 degrading. The replacement of high marsh vegetation by low
1275 marsh vegetation in some salt marshes in New England
1276 began at roughly the same time sea level began to accelerate
1277 [Donnelly and Berta, 2001]. Marsh elevations appear to
1278 be deepening relative to sea level in South Carolina [Morris
1279 et al., 2005], channel networks appear to be expanding in
1280 South Carolina, New York, and Maryland [Kearney et al.,
1281 1988; Hartig et al., 2002; Hughes et al., 2009], and large
1282 amounts of marshland are being converted to open water in
1283 Louisiana, Maryland, Italy, and southeastern England [e.g.,
1284 Reed, 1995; Day et al., 1999; Kearney et al., 2002; van der
1285 Wal and Pye, 2004]. In fact, historical rates of marsh loss
1286 correlate with historical changes in relative sea level rise rates
1287 in coastal Louisiana and the Chesapeake Bay Estuary
1288 [Stevenson et al., 1986; Swenson and Swarzenski, 1995].
1289 It is difficult to determine the influence of sea level acceleration
1290 remains difficult, however, since the effects of sea level rise
1291 alone cannot be isolated in natural wetlands. Sediment supply
1292 also exerts a strong control on marsh expansion and decline.
1293 A pulse of sediment can lead to a wide range of feedbacks
1294 that encourage marsh expansion [e.g., Mudd, 2011]. Sedi-
1295 ment supply reduction and increased subsidence rates are at
1296 least partially responsible for marsh loss in Chesapeake Bay,
1297 coastal Louisiana, and Venice Lagoon marshes [Kearney et al.,
1298 2002; Reed, 1995; Marani et al., 2007]. The main
1299 cause of loss of coastal wetlands in the Mississippi Delta is
1300 the isolation of the river from the delta through the construc-
1301 tion of levees, which dramatically reduced sediment inputs to
1302 the salt marshes [Day et al., 2005]. Moreover, identifying sea
1303 level rise as the primary driver of marsh loss is complicated
1304 by observations that some marshes are submerging despite
1305 vertical accretion rates that exceed relative rates of sea level
1306 rise [Kirwan et al., 2008] and that most marshes today have
1307 elevations increasing faster than historic rates of sea level rise
1308 [Cahoon et al., 2006]. In fact, Kirwan and Temmerman
1309 [2009] concluded that factors other than historical sea level
1310 acceleration were most likely responsible for widespread
1311 patterns of marsh submergence.

### 8.2. Simulating Marsh Evolution Under Sea Level Rise

Numerical models may help determine the direct
1314 influence of sea level rise on marsh survival since they offer
1315 the distinct advantage of being able to isolate sea level as a
1316 forcing variable. Models of platform elevation, for example, show a tendency for marshes to become deeper within the
1318 tidal frame in response to an increase in the rate of sea level
1319 rise alone [e.g., French, 1993; Allen, 1995] and can become
1320 too deep to support vegetation growth at high rates of sea
1321 level rise (i.e., >10 mm yr\(^{-1}\)) [Morris et al., 2002]. While
1322 these models show vegetated intertidal surfaces to be rela-
1323 tively resilient to changes in sea level, D’Alpaos et al.
1324 [2007a] suggest that relatively small changes in platform
elevation can lead to channel network expansion, and dis-
turbance to vegetation can trigger rapid marsh degradation
[Marani et al., 2007; Kirwan et al., 2008].

Although a wide variety of numerical models exists
[see also Kirwan and Temmerman, 2009, and references
therein], most are based on the assumption that marsh
acceleration rates should increase with inundation due to sea
level rise:

\[ \frac{dz}{dt} = k_z \]

(26)

where \( \frac{dz}{dt} \) represents the change in marsh elevation through
time (i.e., its accretion rate) and \( k_z \) is a proxy for the duration
and frequency of inundation. At its most basic level,
equation (26) predicts that a decrease in elevation relative to
sea level, \( z \), will be accompanied by an increase in the
accretion rate. The processes actually responsible for this
relationship (implicitly incorporated into \( k_z \)) vary between
models and include many of the biological and/or physical
feedbacks discussed in earlier sections (e.g., sections 5
and 8.1). In particular, vertical elevation adjustment in most
recent models is accomplished primarily through the sedi-
mentation-inundation feedback and the enhanced growth of
plants with moderate increases in inundation and its effect
on mineral sediment trapping and organic accretion.

### 8.3. Multiple Stable Equilibria

Several basic patterns evolve out of any model that
relies on an inundation-dependent accretion scheme. First,
when sea level rise rates are constant a marsh will asym-
ptotically evolve toward an elevation where accretion rates are
in equilibrium with sea level. If initial marsh elevations are
relatively low (i.e., frequently inundated), accretion is rapid
and the surface builds higher in the tidal frame. If marsh
elevations are relatively high (i.e., infrequently flooded),
accretion is slow, and sea level rise causes the marsh surface
to become lower in the tidal frame. Morris et al. [2002]
explain that the particular equilibrium elevation is con-
trolled by the rate of sea level rise, the amount of sediment
available for mineral accretion, and the rate of vegetation
growth. Where mineral sediment concentrations decrease
with distance from the channel or marsh edge, equilibrium
elevations would be expected to decrease toward the marsh
interior, resulting in a levee, gently sloping marsh surface
[Mudd et al., 2004]. Conversely, for a marsh dominated by
organic accretion, the marsh platform would be expected to
evolve toward a flat intertidal surface.

Since marsh elevations evolve toward a condition
where vertical accretion balances sea level rise, the equilib-
rium solution of equation (26) can be rewritten as:

\[ m = \frac{k_z}{z} \]

(27)

where \( m \) represents the rate of sea level rise. Following
Morris et al. [2002], a change in the rate of sea level rise must
be accompanied by a change in marsh elevation. For an
acceleration in the rate of sea level rise, the elevation of the
1374
8.4. Lags Between Sea Level and Morphologic Change

[90] The idea that marshes must become more inundated before they can accrete at faster rates leads to a common observation in numerical experiments that there is a lag between changes in sea level and morphology [French, 2006; Kirwan and Murray, 2008a; Kirwan and Temmerman, 2009; D’Alpaos, 2011]. As a brief example, models of salt marsh evolution suggest that in response to a step change in the rate of sea level rise from 1 to 3 mm yr \(^{-1}\), a marsh will lose about 10 cm of elevation relative to sea level before accretion rates equilibrate to the new rate of sea level rise. If sea level is rising at 3 mm yr \(^{-1}\), then in the absence of deposition, it would take 33 years to deepen 10 cm and equilibrate. Of course, the marsh is also building elevation during this time, so the adjustment period (i.e., the lag) is actually much longer (~100 years) [Kirwan and Temmerman, 2009] (Figure 13). As a consequence, marshes are always moving toward, but never reaching equilibrium with a continuously varying sea level rise typical of real eustatic oscillations [French, 2006; Kirwan and Murray, 2008a] (Figure 13b). Moreover, simulated marshes never reach equilibrium with a continuously accelerating rate of sea level rise. In scenarios of future acceleration, accretion rates mimic sea level rise rates that occurred about 30–40 years previously [Kirwan and Temmerman, 2009] (Figure 13a). [91] The idea that marshes must become more inundated before they can accrete at faster rates leads to a common observation in numerical experiments that there is a lag between changes in sea level and morphology [French, 2006; Kirwan and Murray, 2008a; Kirwan and Temmerman, 2009; D’Alpaos, 2011]. As a brief example, models of salt marsh evolution suggest that in response to a step change in the rate of sea level rise from 1 to 3 mm yr \(^{-1}\), a marsh will lose about 10 cm of elevation relative to sea level before accretion rates equilibrate to the new rate of sea level rise. If sea level is rising at 3 mm yr \(^{-1}\), then in the absence of deposition, it would take 33 years to deepen 10 cm and equilibrate. Of course, the marsh is also building elevation during this time, so the adjustment period (i.e., the lag) is actually much longer (~100 years) [Kirwan and Temmerman, 2009] (Figure 13). As a consequence, marshes are always moving toward, but never reaching equilibrium with a continuously varying sea level rise typical of real eustatic oscillations [French, 2006; Kirwan and Murray, 2008a] (Figure 13b). Moreover, simulated marshes never reach equilibrium with a continuously accelerating rate of sea level rise. In scenarios of future acceleration, accretion rates mimic sea level rise rates that occurred about 30–40 years previously [Kirwan and Temmerman, 2009] (Figure 13a).
patterns of inundation for most of the 20th century and may just now be approaching a new equilibrium state. Alternatively, if recent increases in sea level rise rates are interpreted as a gradual acceleration, then properties of the marsh are always a few decades behind the physical conditions driving the change. In that case, observations of marsh adjustment such as transgression of plant zonation or expanding channel networks are underestimated, and more change should be expected even if rates of sea level rise were to stabilize [Kirwan and Murray, 2008a].

8.5. Threshold Sea Level Rise Rates for Marsh Survival

While models of salt marsh evolution generally point to ecosystem resiliency, widespread observations of marsh submergence indicate that under some conditions, marshes simply cannot survive. In particular, marshes in estuaries with low tidal ranges and little mineral sediment appear to be vulnerable [Reed, 1995]. One limit to the survival of intertidal wetlands is the growth of vegetation itself. In the absence of vegetation, a number of ecogeomorphic processes (e.g., peat collapse, wave and channel erosion, and lack of accretion) lead to the rapid loss of elevation, precluding the return of vegetation [DeLaune et al., 1994; Fagherazzi et al., 2006; Marani et al., 2007; Kirwan et al., 2008]. Thus for a marsh to maintain its position in the intertidal zone, its elevation must never become so low that vegetation dies. [95] As noted earlier, an acceleration in the rate of sea level rise leads to a loss of marsh elevation relative to sea level rise and an increase in tidal flooding. While this tends to increase mineral deposition, plant growth, and organic accretion, at some point, platform elevations become so low that vegetation cannot survive. Kirwan et al. [2010] summarized the conditions that lead to platform submergence in five numerical models and found that threshold rates of sea level rise for marsh survival vary by more than 2 orders of magnitude depending on an estuary's tidal range and sediment
availability Figure 14). At low tidal ranges and suspended
sediment concentrations, marshes submerged at rates of sea
level rise of only a few millimeters per year. However, under
more favorable conditions, modeled marshes could survive
several centimeters of sea level rise per year. Variation
between models is surprisingly low given that they were
designed to represent marshes from around the world (e.g.,
Venice Lagoon (Italy), North Inlet (South Carolina), and
Scheldt Estuary (Belgium-Netherlands)) and consequently
incorporate different dominant processes and approaches
to modeling ecogeomorphic feedbacks (two treat organic
accretion in detail while others treat mineral accretion in
detail, and one does not model vegetation growth at all).

Apparent the coupling between inundation and accretion
that is generally common to all models (k in (equation (26))
leads to similar results regardless of the actual processes that
are responsible for such a link or how they are modeled
[Kirwan and Temmerman, 2009].

Although wave erosion is not incorporated into these
particular simulations, its effect is likely to enhance the
threshold condition by introducing a positive feedback
between inundation and erosion. In relatively shallow water,
an increase in inundation leads to a decrease in bottom fric-
tion and therefore an increase in erosion rate and further
platform lowering [Fagherazzi et al., 2006]. Although veg-
etation growth can mitigate the feedback, when sea level rise
lowers platform elevations so that vegetation cannot grow,
waves quickly erode low intertidal surfaces into ones that are
permanently inundated and incapable of regrowing vegeta-
tion [Marani et al., 2007]. In fact, this process is likely
responsible for the bimodal distribution of elevations in
coastal lagoons where very few surfaces are at low intertidal
elevations [Fagherazzi et al., 2006].

8.6. Predictions for the Future

[97] The threshold conditions identified in the previous
section (Figure 14) offer insight into how coastal wetlands
will respond to future sea level rise. As a first approximation,
the fate of any particular marsh can be evaluated simply
by knowing the suspended sediment concentration and
tidal range of the estuary. Using the Plum Island Estuary,
Massachusetts, United States, as an example (SSC = 1 mg
L\(^{-1}\) and TR = 3 m), the model ensemble predicts a threshold
rate of sea level rise of 4 mm yr\(^{-1}\). While stable at historical
rates of relative sea level rise (∼3 mm yr\(^{-1}\) locally), even a
slight acceleration would push marshes in this estuary into
the unstable portion of the graph. On the other hand, with
high sediment availability (e.g., Yantze River Delta, China;
SSC = 1000 mg L\(^{-1}\) and TR ∼ 5 m), the modeling frame-
work predicts marshes to be stable at virtually any realistic
sea level rise rate. Kirwan et al. [2010] conclude that while
a moderate IPCC style acceleration in the rate of sea level
rise would threaten marshes in a few estuaries, most would
remain stable. However, faster accelerations in the rate of sea
level rise (>1 m by 2100 [Rahmstorf, 2007]) would submerge
all but the most sediment rich marshes [Kirwan et al., 2010].

[98] Although point-based models of vertical elevation
change converge to fairly similar results (Figures 14 and 15a),
there is considerably less confidence in how elevation tra-
jectories will vary spatially, particularly where influenced by
feedback between inundation and wave erosion [Fagherazzi et al., 2006; Marani et al., 2007]. Incorporating these types of complexities, common in real intertidal systems, remains an important goal for the next generation of salt marsh models.

Existing spatially explicit models help solve some of these problems. Large-scale, landscape-style models consider separately the evolution of different vegetation types, and some treat organic accumulation in detail [e.g., Reyes et al., 2000; Craft et al., 2009; Kairis and Rybczyk, 2010]. These models document the transgression of vegetation zones and demonstrate that loss of a particular type of marsh occurs well before the complete drowning of all marshland [Craft et al., 2009; Reyes et al., 2000; Kirwan and Murray, 2008b; Kairis and Rybczyk, 2010] (Figure 11). Although salt marsh edge erosion is not comprehensively treated [e.g., Mariotti and Fagherazzi, 2010], landscape models offer insight into the potential expansion of marshes in the upslope direction. Simulations on the Georgia Coast, for example, suggest that brackish marshes and shrub-marsh transition areas will expand [Craft et al., 2009]. Similarly, simulations on the Fraser River Delta, Canada, predict that, in the absence of dykes, marsh expansion would more than account for loss of salt marsh due to sea level rise over the next century [Kirwan and Murray, 2008b]. Nevertheless, treatments of processes in these landscape-scale models are necessarily simplistic and in some cases may overestimate predicted change [Kirwan and Guntenspergen, 2009]. More detailed, process-based models consider important interactions between the channel network and marsh platform, between wave erosion and vertical accretion, and the competition between marsh erosion at the seaward edge and marsh expansion at the landward edge. However, at present, these tend to be most suitable for exploring the general evolution of a schematic wetland or simulating a very specific location (i.e., a portion of the Venice Lagoon, Italy).

### 9. CONCLUSIONS

The first generation of models of salt marsh evolution simulated deposition and accretion processes only along the vertical dimension (point models [see Allen, 1994; Woolnough et al., 1995]). These models are simple and of great conceptual value, but fail to represent the richness of the marsh landscape. In recent years, several researchers introduced the spatial distribution of sediment fluxes and vegetation characteristics in their modeling frameworks. Existing spatial models range, with increasing complexity, from simple empirical models that predict sedimentation patterns as a function of topographic variables [Temmerman et al., 2003a] to physically based models that simulate water and sediment flow paths on the basis of simplified hydrodynamic schemes [D’Alpaos et al., 2007a; Rinaldo et al., 1999b] or on the basis of a full hydrodynamic description of the feedbacks between tidal flow and vegetation [Temmerman et al., 2005b]. All of these approaches have their potentials and limitations. For example, the numerically simpler models allow the computation of long-term morphological changes as the result of

1558 waves and vegetation zonation. High marshes are subject to different processes than low marshes (e.g., the role of organic accumulation increases with elevation), and species with different tolerances to flooding interact in ways that can enable or inhibit survival of neighboring species [Morris, 2006]. Moreover, point-based models miss potentially important interactions between different portions of the marsh system. For example, channel erosion that accompanies platform deepening may bring extra sediment to interior portions of the marsh platform [D’Alpaos et al., 2007a, 2007b; Kirwan et al., 2008]. Alternatively, gradual platform submergence may be accelerated by the positive

![Figure 15. Response of (a) marsh elevation and (b) accretion rate to a conservative sea level acceleration (IPCC A1B scenario [Bindoff et al., 2007]) as predicted by five point models of salt marsh evolution. The heavy blue line denotes sea level at spring high water (Figure 15a) or the sea level rise rate (Figure 15b). Since sea level rise rates tend to exceed accretion rates, marsh elevations adjust to sea level acceleration by becoming lower relative to sea level (i.e., more inundated) (Figure 15a), which enhanced vertical accretion (Figure 15b). (Experimental conditions: spring tidal range = 1 m, suspended sediment concentration = 30 mg L⁻¹) [from Kirwan et al., 2010].](image-url)
platform sediment fluxes and the interaction with other eco-
geomorphological units of tidal marshes, such as the channel
dynamics [D’Alpaos et al., 2007a]. Complex hydrodynamic
modeling provides an opportunity to gain fundamental
insights in sediment flux mechanisms, such as the role of
vegetation-flow interactions [Temmerman et al., 2005b].

[102] A key component of numerical models of salt marsh
evolution is the coupling between geomorphology and ecol-
ogy. This coupling must be quantitative, i.e., described by
process-based equations that can be included in numerical
codes. Moreover, the parameters of these equations should be
derived by field or laboratory experiments. In sections 5 and
6 we presented a brief overview of the parametric equations
currently used in salt marsh models, but these studies are still
depicted in their infancy, and more research is clearly needed to
address the influence of biota on morphology and sediment
transport. In fact, several of the expressions used in current
models were derived for specific geographical locations, and
their inclusion in global models is of doubtful validity. The
first interdisciplinary studies involving engineers, biologists,
and geologists have started to address these important feed-
backs, providing the basic blocks for the next generation of
marsh models [see, e.g., Mudd et al., 2009; Kirwan et al.,
2009; Fagherazzi et al., 2011].

[103] Sediment transport dynamics in current models of
salt marsh evolution are extremely simplified and are only a
starting point for the representation of these systems. More
refined models will need to account for complex erosive and
depositional processes in cohesive sediments, the effect of
biota on sediment transport processes, the sedimentology of
organic matter, and the effect of grain size distribution on
erosion and deposition.

[104] Most of the models presented in this review describe
the morphological evolution of salt marshes as a continuous
process regulated by slowly varying inputs of sediment and
sea level rise. In reality both erosion and deposition are sto-
chastic in nature, with infrequent events like storms, hurri-
canes, and heavy rainfall producing most of the geomorphic
work. Storms trigger wave attack of marsh boundaries and
removal of the vegetation mat [Priestas and Fagherazzi,
2011], as well as enhance sediment remobilization in the
subtidal area leading to high transport and deposition on the
marsh surface [Martinot and Fagherazzi, 2010]. The inclu-
dion of the variability of external drivers in salt marsh models
is still in a primordial phase [see, e.g., Rybczyn and Cahoon,
2001], and more research is clearly needed on this important

[105] A few conclusions arise from the applications of
marsh models to climate change and the dynamics of sea
level rise. First, the point-based vertical evolution models
clearly indicate that large swaths of marshland will persist
under conservative projections of sea level rise during the
next century, but will submerge under faster scenarios. In
either case, sea level change will be accompanied by a low-
erdraining of platform elevations that will lead to a migration of
vegetation zones, and landscape models predict the loss of
some vegetation types. More detailed geomorphic models
predict that platform lowering will be accompanied by an
expansion of the channel network, wave scour, and lateral
erosion of the marsh edge. However, these models are for
now unable to discern between a few basic outcomes. For
example, the models presented herein cannot determine
whether upland expansion of marshes can compensate for
erosion of the salt marsh edge and vertical submergence of
the platform. They cannot quantify the relative importance
of the sediment delivery to the marsh interior from an
expanding channel network or the positive feedback between
wave erosion and inundation.

[106] Further studies are also needed for uncertainty
quantification and error estimation of the model results. In
fact, several of the numerical frameworks presented herein
are based on simplified assumptions, on sparse data sets very
often site-specific, and on parameters that might display high
variability in nature.

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22 of 28


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