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On the morphodynamic stability of intertidal environments and the role of vegetation

Nabil Kakeh[a,*], Giovanni Coco[b], Marco Marani[c,d]

[a] Department of Applied Physics, Universitat Politecnica de Catalunya, Barcelona, Spain
[b] School of Environment, University of Auckland, Auckland, New Zealand
[c] Earth and Ocean Sciences Division and Civil and Environmental Engineering Department – Duke University, USA
[d] Department of Civil, Environmental and Architectural Engineering – University of Padua, Padova, Italy

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ABSTRACT

We describe the coupled biotic and abiotic dynamics in intertidal environments using a point model that includes suspended sediment deposition, wave- and current-driven erosion, biofilm sediment stabilization, and sediment production and stabilization by vegetation. We explore the effects of two widely different types of vegetation: salt-marsh vegetation and mangroves. These two types of vegetation, which colonize distinct geographical areas, are characterized by different biomass productivities and stabilization mechanisms. We show that changing vegetation and biofilm properties result in differing stable states, both in their type and number. The presence of the biofilm exerts a dominant control on the tidal flat (lower intertidal) equilibrium elevation and stability. Vegetation controls the elevation of the marsh platform (i.e., the upper intertidal equilibrium). The two types of vegetation considered lead to similar effects on the stability of the system despite their distinct biophysical interactions.

1. Introduction

Ecogeomorphology, or biomorphodynamics, is a novel discipline that studies how the feedbacks between biotic and abiotic processes shape Earth's surface. The importance of biological effects on sedimentary processes was first discussed by Darwin, who, in 1881, studied the role of bioturbation on sedimentary dynamics [1]. Darwin predicted that the study of sediment transport in natural environments would have to incorporate the effects of biological processes. Earth science totally disregarded Darwin's suggestion and only after more than 100 years are we now beginning to understand that biology plays a key role in morphodynamics through feedbacks that affect water and sediment transport processes. The presence of such feedbacks implies that not only biological processes affect geomorphological evolution, but also that organisms are in turn affected by the evolving geomorphology [2]. The importance of ecogeomorphology is aptly embodied by the case of tidal landscapes, such as estuaries and coastal embayments, where it has been extensively studied (e.g., [3–9]).

Tidal landscapes are dissected by a network of tidal channels [6,10] separating relatively flat areas (tidal platforms) positioned at different, but characteristic, elevations: subtidal areas (permanently submerged), tidal flats (usually non-vegetated zones approximately located between low tidal level and mean sea level), and vegetated upper intertidal areas (marshes and mangroves, located between mean sea level and high tidal level). Here we focus on the accretionary dynamics and transformations of these characteristic landforms, regulated as they are by two-way interactions among physical and biological processes [11]. Vegetation plays a key role in the upper intertidal zone, where it produces organic soil, traps suspended sediment, and stabilizes the soil surface, thus exerting a strong control on the elevation of salt-marsh areas (e.g., [9,12–15]). Marsh vegetation also fundamentally affects flow resistance [16], and thus interacts with channel formation processes [17,18] and the topography-hydrodynamics feedback widely regarded as the chief network formation mechanism [6].

Concerns about marshes ability to maintain their elevation relative to an increasing sea level rise have spurred a considerable amount of research focusing on the mechanisms regulating marsh accretion [3–5]. These works explored the existence, and nature, of equilibrium configurations under different scenarios of rate of relative sea-level rise, inorganic sediment availability, and tidal amplitude.

Here we build on the approach described in [4] and [11], who couple equations describing vegetation and "vertical" elevation dynamics in a 0D framework, finding that alternative stable states exist and that changes in the rate of sea level rise and in other forcings (e.g. sediment availability) can induce transitions from one state to another.

* Corresponding author. Tel.: +34 934 01 60 88.
E-mail address: nabil.kakeh@upc.edu (N. Kakeh).

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These previous studies have highlighted the importance of vegetation in affecting the feedback processes driving morphological evolution, yet only a limited range of vegetation species have been examined. The focus has been usually directed towards species present at higher latitudes (e.g. along the Atlantic and Mediterranean coasts). Yet, a large part of the world's tidal environments, in tropical and subtropical areas, are dominated by mangroves [19]. Mangrove trees are characterized by bio-stabilization effects similar to those occurring in marshes, but their geomorphic implications have only just began to be explored. At present most studies deal with the interactions between mangroves, hydrodynamics and sediment transport (e.g. [20]), by means of small-scale field experiments and only recently a large-scale nonlinear morphodynamic model has been developed to study the evolution of the whole biophysical system [21,22].

Here we develop a general model that couples physical and biological processes typical of tidal environments and we explore the role of different vegetation types in determining possible stable states and the transitions among them. We also explore the importance of benthic biofilms generated by microphytobenthos (a collection of microalgae and microbial assemblages colonizing intertidal soils, e.g. [23]), a known but poorly understood bio-geomorphic factor, which regulates erosion and sediment transport in tidal flats and in the upper intertidal zone. Following [11], we use a stability analysis approach to look at the evolution of the coupled sediment-vegetation-system. While using their same description for the physical system, we adopted different modeling frameworks to describe different vegetation types, from grassy alophytic vegetation, e.g. Spartina, to woody mangroves. We subsequently analyzed the dynamical implications of the different biological processes associated with such vegetations species.

2. Methodology

We use the 0-D modeling approach proposed by Marani [11] to describe the average mean tidal platform elevation $z(t)$ (with respect to mean sea level - MSL) and the vegetation density of biomass $B(t)$. Additionally, we consider the time evolution of the biofilm biomass $\chi(t)$. As our interest lies in studying the long-term stability of the system we have assumed that intra-annual variations of vegetation and platform elevation are negligible.

2.1. Physical model

Fig. 1 shows a sketch of the physical system considered. The platform is subject to a sinusoidal tide, of period T (we used here T= 12 hrs, the fundamental periodicity of tidal oscillations) and amplitude H. The suspended sediment concentration flux from the water column is deposited on the platform as a result of settling and trapping processes. Moreover, root growth produces organic soil (after oxidation of the labile organic material) and cause soil expansion which contributes to platform accretion. Finally, waves may cause platform erosion and produce sediment transport.

According to [11], we can express the long-term evolution of the platform as:

$$\frac{dz}{dt} = Q_S(z,B) + Q_T(z,B) + Q_O(B) - E[z,B] - R \tag{1}$$

Q_S is the settling flux, which is proportional to the settling velocity w_s:

$$\frac{dz_s}{dt} = \frac{w_s}{\rho_b} C(z,B,t) \tag{2}$$

where dz_s/dt is the elevation change due to the settling flux, ρ_b is the sediment bulk density, $\rho_b = \rho_s(1-\lambda)$($\rho_s = 2650\,\mathrm{kg/m^3}$for an inorganic material), and $\lambda = 0.5$is the porosity (i.e. the final porosity once the sediment is deposited and compacted). The instantaneous sediment concentration on the tidal platform $C(z, B, t)$ (mass of sediment per unit volume of the water column) depends on the vegetation density (through trapping) and the instantaneous water depth. The tidal elevation with respect to the local mean sea level (MSL) is $h(t)$ and the instantaneous water depth is equal to $D(t) = h(t) - z$.

As noted in [11], the time scale of change of $h(t)$ (order of tens of minutes), is much more rapid than the time scale of change of $z(t)$(order of years to tens of years). Hence the settling deposition over a tidal cycle can be averaged by assuming $z(t)$ to be constant. Consequently, the mean settling flux to the platform surface can be expressed as:

$$Q_S(z,B) = \frac{w_s}{\rho_b}\frac{1}{T}\int_T C(z,B,t)dt \tag{3}$$

Trapping is the process by which particles suspended in the water column are intercepted by vegetation and are eventually deposited on the soil surface. The sediment flux to the soil surface associated with trapping can be expressed as:

$$Q_T(z,B) = \frac{\alpha_T B^{\beta}}{\rho_b}\frac{1}{T}\int_T C(z,B,t)dt \tag{4}$$

where $\beta = 0.382$and $\alpha_T = 1.02 \cdot 10^6 d_{50}^2 U^{1.7}$ m/s $(\mathrm{m^2/g})^{\beta}$(this formulation and the associated parameterizations are taken from [11]). The average tidal flow velocity and grain size are here assumed to be $U = 0.02$ m/s(a typical order of magnitude) and $d_{50} = 50\,\mu$m, respectively.

In order to solve Eqs. (3) and (4) we must determine the sediment concentration, $C(z, B, t)$, over the tidal platform. If we assume that the concentration is constant over the water column, the total mass of suspended sediment per unit area is equal to $D(t) \cdot C(t)$. We can now express the rate of change in the mass of sediment in the water column as the sediment exchanged between the tidal platform and the surrounding environment minus the sediment deposited on the marsh surface due to settling and trapping [4,24]:

$$\frac{dC_s}{dt} = \frac{d(D\cdot C)}{dt} = \hat{C}\frac{dh}{dt} - (w_s + \alpha_T B^{\beta})C \tag{5}$$

The first term represents the sediment flux exchanged between the platform and the lagoon. When the tide is rising, the flux is positive, from the lagoon to the platform. When the tide is falling, the the flux is negative, from the platform to the lagoon. If C_0 is the average suspended sediment concentration in the lagoon, we can express the term $\hat{C}$as [11,25,26]:

$$\hat{C}(z,t) = \begin{cases} C_0 & \text{when } \frac{dh}{dt} > 0 \\ C(z,t) & \text{when } \frac{dh}{dt} < 0 \end{cases} \tag{6}$$

The term $Q_O(B)$ in Eq. (1) represents the elevation change due to soil expansion driven by organic matter production, accumulation and decomposition. According to various authors ([12,26,27]) this term can be considered to be proportional to the biomass, B:

$$Q_O(B) = \frac{dz_O}{dt} = \gamma B \tag{7}$$

where γ is a parameter that depends on soil and vegetation characteristics and on the decomposition rate (largely controlled by soil aeration [13]), and may take a value in the range $1-3$ (mm/yr)$(\mathrm{m^2/g})$(a value of 2 is used unless otherwise indicated).

Erosion is produced by the sediment being resuspended from the platform surface. The erosion rate is usually expressed as a function of the normalized shear stress exceedance $\overline{\tau}$[28–30]:

$$\overline{\tau}[D(t), U_w, B, \chi] = \frac{\tau[D(t), U_w] - \tau_c[B,\chi]}{\tau_c[B,\chi]} \tag{8}$$

where τ and τ_c are the shear stress and the critical shear stress. We consider here that τ is locally generated by wind waves (contributions by tidal flows being significant only in macrotidal environments). It is calculated from wind wave height and wave period,