Application of a barrier island translation model to the millennial-scale evolution of Sand Key, Florida

Riccardo Masetti [a], Sergio Fagherazzi [b,*], Alberto Montanari [a]

[a] Department DIRAC, University of Bologna, V.le Risorgimento 2, 40126 Bologna, Italy
[b] Department of Earth Sciences and Center for Computational Sciences, Boston University, Boston, MA 02215, USA

ARTICLE INFO

ABSTRACT

The evolution of a barrier island and its stratigraphic architecture is investigated with the numerical model BIT (barrier island translation). The model simulates, with simplified equations, the effects of various processes (wind waves, storm surges, sea-level oscillations) on sediment location and characteristics. The presented formulation is able to reproduce both the cross-shore profile and the distribution of sediment facies in time.

The model is based on two appropriate simplifying assumptions: (i) the conservation of the total mass of sediments and (ii) the conservation of the equilibrium cross-shore beach profile. Particular attention is devoted to the treatment of sediment reworking by wind waves and barrier overwash, which is induced by extreme events like storms and hurricanes, and causes a relative transport of sediments from the body of the sand barrier to the top of the island and to the back-barrier area. The model is applied to Sand Key, Florida. Model results show a satisfactory reproduction of the geometry and stratigraphy of the inner shelf, with the correct position of every stratigraphic facies within the barrier island. Simulations under different scenarios of sea-level rise during the last 8000 yr indicate that the rate of overwash and lagoonal deposition are critical for the survival of the barrier island under past sea-level oscillations.

1. Introduction

Sand barriers are elongated islands or spits parallel to the shore, whose back-barrier area is characterized by the presence of a bay, a lagoon or a salt marsh (Hoyt, 1967). Originated by the deposition of sand by waves, wind and currents, barrier islands are geological formations with an extremely variable morphology. They are constantly shaped by wind, tides and waves and, on a longer time scale, they can shift landward or seaward according to the oscillations of sea level and variations in sediment supply. Barrier islands can be found along the 13% of the worldwide shoreline (Dean and Dalrymple, 2002), mostly in areas with a relatively small tidal range and abundance of siliceous–clastic sediments (Dean and Dalrymple, 2002; Davis and Fitzgerald, 2002).

Several theories have been proposed to explain the formation of a sand barrier (deBeaumont, 1845; Shepard, 1960; Hoyt, 1967). For deBeaumont (1845), barrier islands form from offshore bars. Hoyt (1967) put forward the hypothesis that a barrier island starts

with the formation, along the shoreline, of either a dune or a beach ridge. This phase is then followed by an increase in sea level, which determines the flooding of the area landward of the ridge. Therefore, barrier islands are not limited to areas where lagoonal deposition and long-shore transport are abundant, and the Hoyt (1967) theory can explain the absence of open-marine sediments landward of several barrier systems (van Straaten, 1965; Shepard, 1960; Zenkovitch, 1962); overcoming in this way the main weakness of the deBeaumont (1845) theory, which does not explain how an offshore bar can emerge above mean sea level. In specific coastal locations where along-shore currents carry large volumes of sediments, barrier islands can form from the elongation of spits (Gilbert, 1890).

Sand barrier evolution is thus driven by the oscillations of mean sea level, which, through sediment transport by waves and currents, cause the landward or the seaward translation of the barrier in response, respectively, to an increase or a drop of mean sea level. The processes responsible for barrier island evolution span several time scales with high frequency components forcing on long-term effects. In particular, the wave action and storm surge currents can produce two distinctive results. In normal conditions, the wave action reworks the sediment substrate producing an equilibrium shoreface profile; during extreme

* Corresponding author. Tel.: +1 617 353 2092.
E-mail address: Sergio@bu.edu (S. Fagherazzi).

events like severe storms and hurricanes, storm surges cause a large transport of sediments (overwash) from the shoreface to the back-barrier area.

Barrier islands are important for marine ecosystems and for mainland protection: for these reasons the scientific community has proposed several approaches based on conceptual and numerical models to study their morphological evolution (see Fagherazzi and Overeem, 2007, for a review). With their conceptual model, Evans et al. (1985) investigated the influence of the antecedent topography on the morphology of a barrier island. This approach allows investigating the effects of the reworking by waves, aeolian action and lagoonal transport. Evans et al. (1985) identified five different sediment facies in barrier island sediment cores, which are the direct consequence of the action of several processes.

The conceptual model of Evans et al. (1985) was applied to the study area of Pinellas County, in Florida (USA), where Sand Key is located, which is a region characterized by a significant long-shore sediment transport, reduced wave activity, an unerodible bedrock, and a scarcity of Quaternary sediments. Evans et al. (1985) understood that, under these conditions, the role of the antecedent topography is critical; their conceptual model hypothesizes that the sand barriers of Pinellas County were shifting landward, as confirmed by the presence, seaward of the islands, of lagoonal sediments, which are usually accumulated in the back-barrier area. This translation was caused by an increase of mean sea level occurred during the late Holocene. Evans et al. (1985) indicate that the island formed on a bedrock terrace and that landward migration of the barrier was stopped by grounding on this morphological feature.

A different approach, called STM (Shoreface Translation Model), was proposed in the pioneering work of Cowell et al. (1995), who simulated the translation of a sand barrier with a determined shape fixed "a priori". Since the barrier geometry is predetermined, the model is unable to replicate the formation and, eventually, the drowning of the island, which entails deep modifications in shape. Moreover, different stratigraphic facies are not considered in this formulation. The simulated processes are: the sediment gains and losses through the long-shore and inlet currents, and the sediment transport caused by hurricanes and severe storms. In this model, the beach profile is set equal to the equilibrium profile indicated by Dean (1991).

An extension of STM could be considered the stratigraphic model GEOMBEST recently elaborated by Stolper et al. (2005). GEOMBEST has the merit of considering the presence of two sediment classes, sand and mud, and a lithified bedrock under the beach profile. The bedrock influences the evolution of the barrier island reducing sediment supply and, consequently, the rate of landward translation. A limit of this approach is the analysis of the dynamics of only one grain class of sand. For this reason, the model is unable to describe the complex stratigraphy of a barrier island, which includes overwash, aeolian, transitional and shoreface sand facies (Evans et al., 1985).

A more complex stratigraphic model has been proposed by Storms et al. (2002). In this model, the processes of erosion and deposition of sediments are independently treated as a function of grain size. Therefore, it is possible to observe, during the evolution of the beach profile, the formation of several stratigraphic units characterized by different erodibility and grain size distribution. The model is very successful in reproducing the sorting of sediments along the inner shelf and the related stratigraphy, as well as the formation of barrier islands at the shoreline location. However, the model does not differentiate among different processes responsible for sediment redistribution in the inner shelf.

The aim of the work proposed here is to develop a new numerical model, named BIT (barrier island translation), which is able to simulate the evolution of a sand barrier, overcoming the limits of the approaches previously described. Contrary to the STM and GEOMBEST models here we introduce a series of equations that mimic all the processes at play. The morphological evolution of the barrier island stems from the application of these equations to the shelf sediments. A similar approach has been adopted in Storms et al. (2002), but their formulation did not represent each process separately.

Our formulation falls in the category of exploratory models, as defined by Murray (2003). Following Murray (2003), our goal is not the accurate description of the processes at play but the understanding of the interactions and feedbacks responsible for the long-term barrier island evolution. We therefore describe four different processes acting on barrier island sediments with simplified equations having a limited number of parameters. The four processes are: (i) sediment reworking by waves in the shoreface area; (ii) sediment diffusion by waves in the inner shelf; (iii) sediment overwash during storms and hurricanes; and (iv) lagoonal deposition in the back-barrier area. In this way, we can (1) understand the interplay of different processes in the evolution of the barrier island system; (2) track the deposition of different facies (each of them produced by a specific process) and their preservation in the geological record. The model is implemented in a spatial domain extending on a vertical plane parallel to the direction of the average wave action and it is based on two essential hypotheses: (1) the conservation of the total mass of sediments in the domain and (2) the conservation of the beach equilibrium profile.

Sediment losses along the equilibrium profile as, for instance, sediment abrasion, offshore sediment export, and the divergence of long-shore sediment transport can be simply included in the model as external inputs or outputs, by just adding/subtracting sediment at selected locations at each time step.

The proposed model is similar to the conceptual model of Evans et al. (1985) because it takes into account the same processes and produces a similar stratigraphic architecture; however, in our case the spatial domain extends beyond the beach toe and thus the model is able to reach more general results.

2. Model description

The numerical model, proposed herein, aims to simulate the dynamics of the principal processes that govern the formation and evolution of a barrier island. The model includes sea-level oscillations and sediment distribution operated by waves and currents. Each process determines the deposition of a distinct sediment facies, separately schematized in the spatial domain. Therefore, at any temporal step, it is possible to recognize six different stratigraphic units: bedrock, transitional, which is the oldest sediment present on the shelf before the translation of the barrier, usually represented by Pleistocene sand; overwash, which is the sediment eroded at the shoreline and then moved on top of the island and to the back-barrier area by hurricanes and severe storms; shoreface, which is the result of the reworking of the beach profile by wind waves; aeolian, composed of material transported by the wind and then deposited on top of the barrier; lagoonal, which consists of fine sediment accumulated in the lagoon behind the sand barrier.

The processes of erosion and distribution of sediments are schematized with differential equations and then solved with finite differences. The integration domain extends on the vertical plane perpendicular to the coastline (Fig. 1), between the vertical
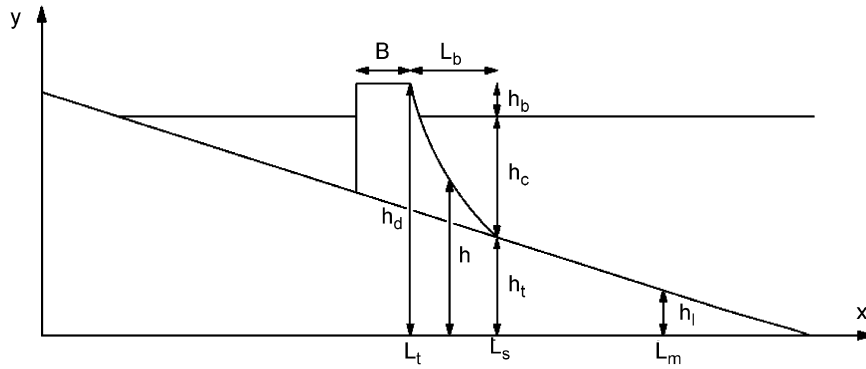
Fig. 1. Integration domain of the model and symbols used in the text.

line passing through the point L_t, which is the top of the foredunes, and the vertical line passing through the point L_m, which is the closure depth for sediment reworking. The spatial partition is obtained subdividing the domain in rectangular cells 0.01 m high and 50 m long. In the study case, the spatial domain extends for almost 30 km.

The model requires the following inputs: (1) the sea-level curve; (2) the rate of lagoonal deposition; (3) the rate of overwash; and (4) the initial shelf profile. The stratigraphic data are organized in a matrix of integers; therefore erosion and deposition of sediments can only occur for a multiple of the cell height, which is set here to 0.01 m. Every matrix entry corresponds to a stratigraphic unit (bedrock, overwash, transitional, shoreface, aeolian, and lagoonal).

The model equations are derived imposing the following conditions:

- The wave action causes the reworking of the beach profile towards the equilibrium configuration defined by Dean (1991). The effect of the wave action extends from the shoreline to the beach toe, indicated with the symbol L_s in Fig. 1.
- The sediment resuspension and deposition in the inner shelf is modeled with a diffusion equation, acting from the beach toe L_s to the maximum depth L_m.
- The exposure of bedrock stops the erosion of the beach profile, which is thus unable to reach the equilibrium configuration.
- The amount of sand moved to the back-barrier area by overwash is a function of the height and width of the subaerial part of the barrier.
- In the lagoon located between the sand barrier and the mainland, there can be deposition of sediments due to the sediment inputs from rivers.
- The upper part of the barrier island is reworked by aeolian processes.

Each condition listed above is described in the following paragraphs with simplified equations.

2.1. Reworking of the beach profile

The reworking of the beach profile occurs between the points L_t and L_s, which represent the spatial limits of the shoreface. Dean (1991) proposed a monomial law to define the equilibrium configuration of a beach reworked by wave action:

$$h_{eq} = h_d - ax^m, \tag{1}$$

where a (L^{1-m}) is a constant depending on grain size (Dean and Maurmeyer, 1983; Kriebel et al., 1991), x (L) represents the horizontal distance between the study point and the beginning of the beach, and m is a parameter, usually considered equal to 2/3.

The model assumes that the wave action reworks the beach profile towards an equilibrium configuration described by Eq. (1). The dynamics of erosion and deposition are schematized with the following relationship (Fagherazzi et al., 2003):

$$\frac{\partial h}{\partial t} = K' \frac{\partial^2}{\partial x^2}(h - h_{eq}), \quad h_t < h < h_d, \tag{2}$$

where h (L) is the depth of the study point and K' (L^2T^{-1}) is a diffusion coefficient. h_t (L) and h_d (L) are the elevations of the points L_s and L_t, previously described. The diffusion scheme changes the bottom elevation at a rate linearly proportional to the difference between the current and the equilibrium profile, as already presented by Kriebel and Dean (1993), and then redistributes the removed or deposited material in equal parts between the contiguous inshore and offshore locations.

This formulation is similar to the one presented in Plant et al. (1999) in the context of sandbar migration. However, Plant et al. (1999) model focuses on a relatively short timescale (seasonal to interannual) during which the sandbars form at the shoreline, migrate seaward, and then decay offshore. At our timescale (centuries to millennia) the shoreline mean profile averages out, as supported by the data reported in Plant et al. (1999), so that a constant equilibrium beach profile can be adopted in Eq. (2).

Since the reworking of sediments by waves, schematized with Eq. (2), is less effective in deep areas, it is assumed that the diffusion coefficient linearly decreases with depth according to the following relationship:

$$K' = K\left(\frac{h - h_t}{h_d - h_t}\right), \tag{3}$$

where the limestone bedrock is exposed, the reworking of the beach profile does not take place (Pilkey et al., 1993; Wright, 1995; Thieler et al., 1995, 2000). We also assume that the rate of bedrock erosion by wave action is negligible at the timescales adopted in our simulations

2.2. Inner-shelf sediment redistribution

In the inner shelf, we model the redistribution of sediments by wind waves, which acts beyond the beach toe L_s to the maximum depth of sediment reworking L_m, which represents the lower limit of the model spatial domain. In this region, bottom shear stresses produced by wind waves rework the sediment, smoothing bottom asperities. Sediment redistribution in the inner shelf can be schematized with a diffusion equation applied to bottom elevations, as already adopted by Flemings and Jordan (1989) and Jervey (1988) to model depositional systems. Herein we assume that the diffusion linearly decreases until reaching zero at the

depth h_t:

$$\frac{\partial h}{\partial t} = \left(\frac{h - h_i}{h_t - h_i}\right)\frac{\partial^2 h}{\partial x^2}, \quad h_i < h < h_t, \tag{4}$$

where α ($L^2 T^{-1}$) is the diffusion coefficient.

2.3. Overwash

The model assumes that overwash is the process responsible for the landward translation of the barrier in response to an increase in sea level. Overwash is related to storm surges produced by extreme atmospheric events. We divide the overwash into two phases: the erosion of sediment along the beach profile and its corresponding deposition on top of the island or in the back-barrier area. The phenomenon of overwash is schematized assuming that, at any temporal step, the first shoreface element of the barrier island is eroded of a quantity θ (LT^{-1}), which is equal to

$$\theta = \frac{A}{h_s}, \tag{5}$$

where h_b (L) is, at any temporal step, the difference between the maximum elevation of the barrier and mean sea level, and A ($L^2 T^{-1}$) is a constant of proportionality that depends on climatic conditions, as the frequency of hurricanes and severe storms.

Detail studies of barrier island overwash in New England (Leatherman and Zaremba, 1987; Leatherman et al., 1977) show that storm surges move large volumes of sediments (up to $400\,m^3/m$ for Nauset Spit, MA) in the back-barrier area for a distance of the order of 200–300 m while maintaining a constant maximum elevation of the barrier subaerial structure between 2 and 3 m.

Based on these observations, the process of overwash is stopped in our model when the barrier reaches an elevation $h_{b,max}$ of 2 m above mean sea level, uniformly extended for a distance B (Fig. 1). B represents the maximum width of the barrier and it is set herein equal to 300 m.

The material eroded by overwash is supposed to be deposited on the subaerial part of the barrier, precisely in the first element having an elevation below the maximum elevation $h_{s,max}$. When the maximum height is achieved, the sediment is deposited in the successive element, thus producing a net translation of the system towards inland.

Washover deposits form lobes in specific areas where gaps between sand dunes facilitate the concentration of flow and sediments. Therefore, the transport of sediment to the back-barrier is highly localized rather than uniform (Leatherman and Zaremba, 1987; Leatherman et al., 1977). Only by averaging among several storms at the millennial timescale we obtain a uniform migration of the barrier inland landward. This notwithstanding, since we are focussing on the long-term migration of the barrier island, we assume a uniform sediment transport to the back-barrier area during overwash events.

2.4. Lagoonal deposition and aeolian sediment reworking

In our model, the lagoonal deposition is the deposition of fine sediments in the accommodation space between the barrier island and the mainland. In some cases, it can be considered the first cause of the formation of barriers in front of a river delta. The lagoonal deposition has a remarkable influence on the evolution of a sand barrier, because it determines the configuration of the shelf profile, over which the island will then translate (Roy et al., 1994; Stolper et al., 2005). A large lagoonal deposition can lead to a reduction of accommodation space in the back-barrier area, thus

increasing the probability of survival during sea-level rise (Cowell et al., 2003). As showed by Evans et al. (1985), lagoonal facies can be found in the geological record to a distance of almost 2500 m offshore the barrier island. Our model assumes that the lagoonal deposition always fills the lower point in the back-barrier area at the rate prescribed in the simulation.

Aeolian sediment transport is very important in the formation of the subaereal part of the barrier island. Our model treats the wind action with a simplified approach, which considers that the subaerial part of the barrier, for a thickness equal to 1 m, is reworked by the wind at each time step. Therefore, in our model, wind action is not described as an active process by a specific equation, but it is simply implemented as a change in stratigraphic facies in the subaerial part of the island.

2.5. Model implementation

At each time step, the model performs the following operations: (a) the sea level is updated following a prescribed sea-level curve; (b) the equilibrium beach profile is computed with Eq. (1) starting from the most offshore emerged point; (c) the difference between the actual beach profile and the equilibrium profile is determined at each point between the shoreline and the beach toe, and utilized in Eq. (2) to compute the beach accretion or erosion. The reworked sediment at this time step is defined as shoreface; (d) the sediments are redistributed in the inner shelf (from the beach toe to the maximum depth) using Eq. (4); (e) a volume of sediment determined by Eq. (5) is removed from the beach profile and deposited in the subaerial part of the island. The sediment is deposited at the closest location from the shoreline having an elevation below the maximum dune elevation. If the dune elevation is reached with only part of the sediment volume removed from the beach, the remaining is deposited in the next point available toward the mainland, and so on until either all the overwash sediment is deposited or the maximum width of the barrier is reached. In the second case, the sediment volume in excess is returned to the beach. The sediment emplaced at this time step is defined as overwash facies; (f) a prescribed volume of lagoonal sediments is deposited in the back-barrier area, starting from the lowest point; (g) the top of the island is changed in aeolian facies.

3. Characteristic scales and barrier island drowning

The simplified formulation introduced allows a direct comparison of the timescales of the processes at play. Several studies indicate that the beach profile recovery occurs at the timescale of months (e.g., Kriebel and Dean, 1993). If we consider that the profile is reconstituted 2 months after a storm, then it is possible to obtain a value of $K = 3 \times 10^6\,m^2/yr$. Furthermore, from Eq. (2), the characteristic rate of beach adjustment toward the equilibrium configuration can be derived as $R_b = K(h_b + h_c)/L_b^2$. With the beach and shelf geometrical parameters reported in Table 1, we then obtain a rate of beach profile adjustment toward the equilibrium configuration $R_b = 74\,m/yr$. This value is of the same order of magnitude of the sandbar migration rate, 0.1 m/day,

Table 1
Simulation parameters

h_b	2 m	B	300 m
h_c	−10 m	S	0.001
R	0.002 m/yr	a	$0.15\,m^{1/3}$
K	$3 \times 10^6\,m^2/yr$	m	2/3
L_b	700 m		

reported in Plant et al. (1999), thus confirming the fast evolution of the beach morphology and profile.

The comparison of R_b with the characteristic rate of sea-level rise during the late Holocene, say $R = 0.002$ m/yr, clearly shows that the adjustment of the shoreface profile happens at a rate much faster than sea-level oscillations, so that it is reasonable to suppose that the beach profile is preserved constant at a geological timescale.

Contrary to shoreface processes, overwash displaces sediments from the shoreface to the back-barrier at rates of the same order magnitude of sea-level changes, given the rare occurrence of intense storms.

If Q is the volume of sediments transported by overwash and aeolian processes from the shoreface to the back-barrier area per unit time, and H the average horizontal distance covered by the sediments, then the vertical rate of barrier island translation is

$$R_b = \frac{Q}{V}HS, \qquad (6)$$

where V is the volume of barrier island per unit length from the beach toe to the back-barrier bay and S is the shelf slope that transforms horizontal in vertical translations. From our model geometry (Fig. 1), we can set, to a first approximation, $V = (B+L_b/2)(h_b+h_c)$ and $H = B+L_b/2$.

Thus, with the parameters in Table 1, an average of 28 m^3/yr/m needs to be transported from the shoreface to the back-barrier in order to maintain the barrier in equilibrium. If this amount is considerably reduced, the barrier island will drown.

To further investigate the drowning of the barrier, in the first set of simulations we consider a simplified geometry with a linear shelf with a slope of 0.001, a barrier island with a width of 300 m, and a beach toe 10 m below mean sea level (Fig. 2 and Table 1). We suppose that intense storms are frequent enough to transfer the necessary volume of sediments to the back-barrier area, so that the barrier island translates at a constant rate with an increase in sea level (Fig. 2a). If the sediment volume transported during overwash events is not sufficient, then the barrier drowns, reforming at the shoreline location (Fig. 2b).

If we measure the translation of the barrier island, we can clearly notice that a barrier destined to drown slowly retreats on the shelf, a direct consequence of an increase in accommodation space in the back-barrier area. Plotting the barrier position on the shelf against time, it is possible to show that a decrease in

sediment volume transferred by overwash implies a reduction of barrier translation rate (Fig. 3).

4. Case study: Sand Key, Florida

The model is then applied to the shelf profile in front of Sand Key, Florida (Fig. 4). In this area, the bedrock is composed of Miocene to Pliocene limestone and dolomitic formations and it is covered by unlithified Pleistocene and Holocene sand and shell (Cherry et al., 1970). The thickness of this unit of sediments is limited to few meters and sometimes, for large tracts, it is altogether absent. The fluvial input into Tampa Bay is extremely low (Brooks and Doyle, 1998).

The stratigraphy of Sand Key shows a Pleistocene muddy sand layer overlying the Miocene limestone. Above the Pleistocene facies there is a thick layer of muddy and shelly sand typically found in low energy environments (lagoonal areas) the next unit is composed of shelly sand characteristic of shoreface environments. The presence of back-barrier deposits offshore the island has been interpreted as a proof of the landward translation of the island in the recent geological past (Evans et al., 1985; Brooks et al., 2003).

The simulations carried out on the Sand Key profile span a period of almost 8000 yr. Since sea-level oscillations are critical for barrier island evolution, three different sea-level curves were
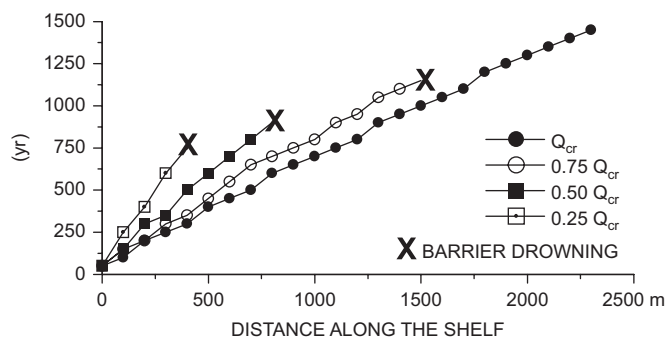
Fig. 3. Barrier island migration as a function of overwash sediment discharge. If the migration rate does not contrast sea-level rise, the barrier drowns (location with a X). Q_{cr} is the minimum overwash sediment discharge that preserves the barrier island.
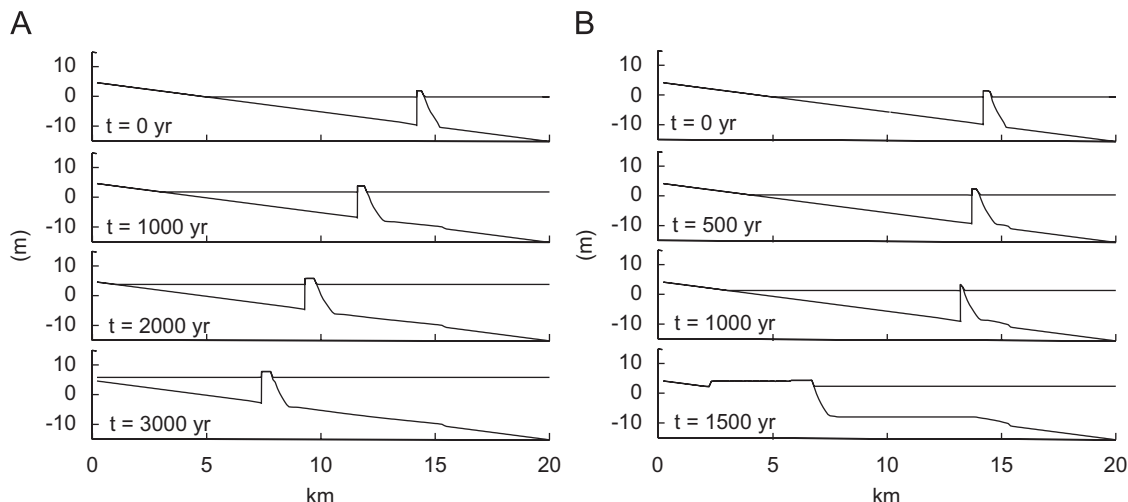
Fig. 2. Sea-level rise and barrier island translation: (a) a sufficient sediment volume is transported by overwash; the barrier island is stable and conserves its shape. (b) The sediment volume transported in the back-barrier is not enough and the barrier island drowns.
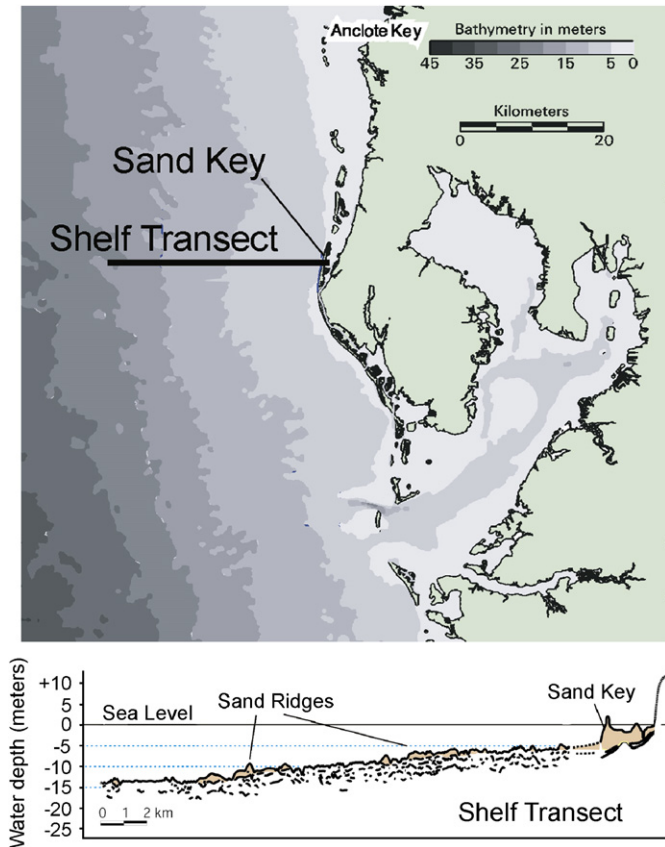
Fig. 4. Location of Sand Key, Florida, and corresponding shelf profile (modified after Locker et al., 2003).
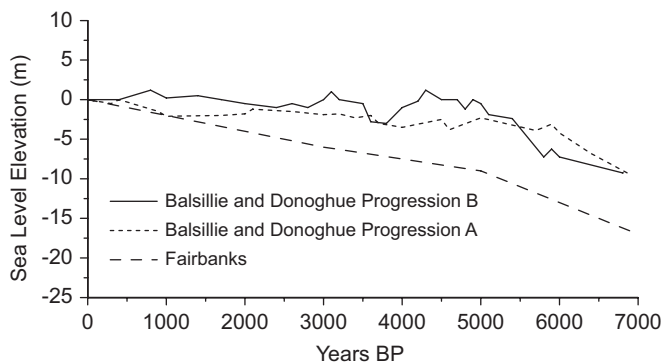

Fig. 5. Sea-level curves utilized in the simulations (after Fairbanks, 1989; Balsillie and Donoghue, 2004).

considered. The first one, derived from Balsillie and Donoghue (2004) and herein called Progression A (Fig. 5), is based on a series of analysis, made by several researchers on offshore stratigraphic samples in the Gulf of Mexico (Kuehn, 1980; Robbin, 1984; Fairbanks, 1989, 1990; Faught and Donoghue, 1997; McBride 1997). The second one, called Progression B (Fig. 5) (Balsillie and Donoghue, 2004), was derived from onshore stratigraphic samples (Tanner et al., 1989; Morton et al., 2000; Blum et al., 2002; Stapor and Stone, 2004; Balsillie and Donoghue, 2004). Progression B has not been thoroughly verified, but introduces for the first time high stand periods above the current sea-level elevation that might have occurred in the recent past. Progression B allows investigating the evolution of a sand barrier under a significant decrease of mean sea level, which can stop the landward translation of the

island and start its seaward progradation. Another set of simulations was carried out considering the sea-level curve of Fairbanks (1989), reconstructed from coral reef dating (Fig. 5). This curve has few oscillations and shows a monotonous increase of mean sea level in our study period.

The shelf profile and the barrier island of Sand Key are reported in Fig. 4 (Locker et al., 2003). Large areas of the shelf are blanketed by a thin layer of sediments or have a bedrock substrate, with several depressions and changes in slope that might have controlled the evolution of the barrier in the geological past. The total barrier island relief from the top of the dunes to the beach toe is almost 7 m, while its subaerial part is more or less 2 m high. The stratigraphic profiles reported in Locker et al. (2003) allowed the estimation of the total volume of sediments present along the transect, without considering lagoonal deposits.

5. Simulation of the evolution of the barrier island of Sand Key

The simulation of the evolution of Sand Key in the last 8000 yr is based on the following initial conditions: (a) the total volume of sediments available along the shelf is considered equal to the amount currently present in the study area; (b) the bedrock profile is assumed identical to the current one; (c) the barrier island, with the current morphology, is located in correspondence of a flat shelf area situated at the beginning of our shelf domain.

Given the lack of information regarding long-shore sediment fluxes, we decided to neglect, to a first approximation, the divergence of the long-shore sediment transport, sediment abrasion, and the offshore export of sediments.

The following values were used for the model parameters (see also Fagherazzi et al., 2003): $K = 2 \times 10^{-6}\,\mathrm{m}^2/\mathrm{month}$, $\alpha = 10^{-6}\,\mathrm{m}^2/\mathrm{month}$, $A = 0.3\,\mathrm{m}^2/\mathrm{month}$, $a = 0.0015\,\mathrm{m}$, the time step in the model is 1 yr.

The initial position of the barrier corresponds to the farthest offshore location of lagoonal facies detected by Brooks et al. (2003). In the first simulation, we assume a rate of overwash $A = 4\,\mathrm{m}^2/\mathrm{month}$, the Progression of Fairbanks (1989), and a lagoonal deposition of 0.0015 m/yr limited to the first 2.5 km of shelf from the mainland shoreline. This rate of lagoonal deposition, calibrated through simulations, permits the sand barrier to reach exactly its current position showed in Fig. 4. The barrier island evolution is showed in Fig. 6. The model reproduces with good agreement the current stratigraphic profile (Fig. 7; it is important to note that the Aeolian sediment will be always reproduced with good agreement, since they are simply emplaced on top of the island). In several locations during the translation, the barrier slows down and almost stops despite the continuous increase in sea level; when it resumes the translation, it leaves along the shelf large volumes of sediment.

The sand barrier decreases its migration rate when the shelf becomes steeper, because, in this case, it is easier for the island to reach the maximum height $h_{b,\max}$ in the subaerial part since the accommodation space in the back-barrier area diminishes. In this configuration, the sediment transport by overwash stops because the barrier island is fully developed with a width equal to the assigned overwash maximum distance and an elevation corresponding to the maximum dune height. Successively, an increase in mean sea level causes a decrease of the relative elevation of the subaerial part of the barrier, triggering overwash and landward translation. Consequently, the translation speed increases and the barrier leaves behind part of its volume. In particular, when the barrier island restarts its landward translation after a pause, all the sediments at the base of the barrier are below the shoreface closure depth, since the barrier has accreted vertically.
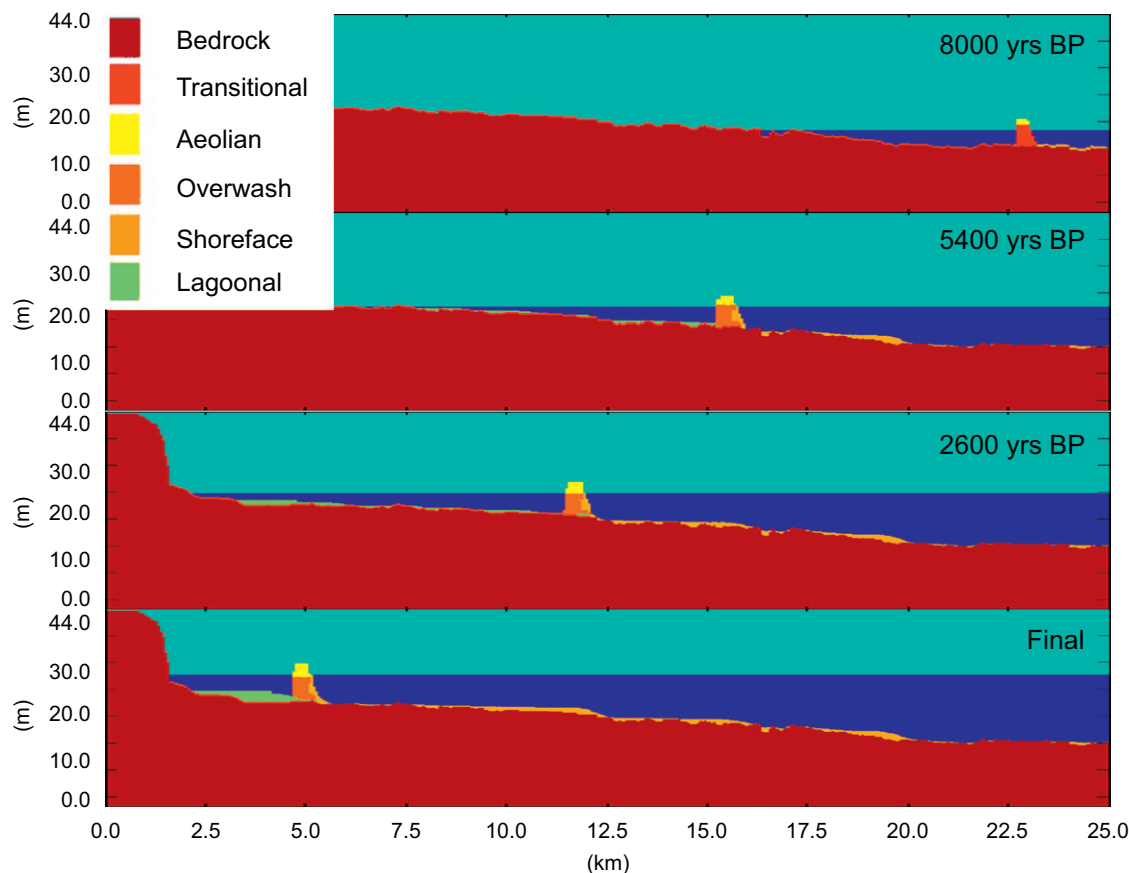
Fig. 6. Evolution of the barrier island of Sand Key with the sea level Progression of Fairbanks (1989) and a rate of overwash $A = 4\,\mathrm{m^2/month}$.
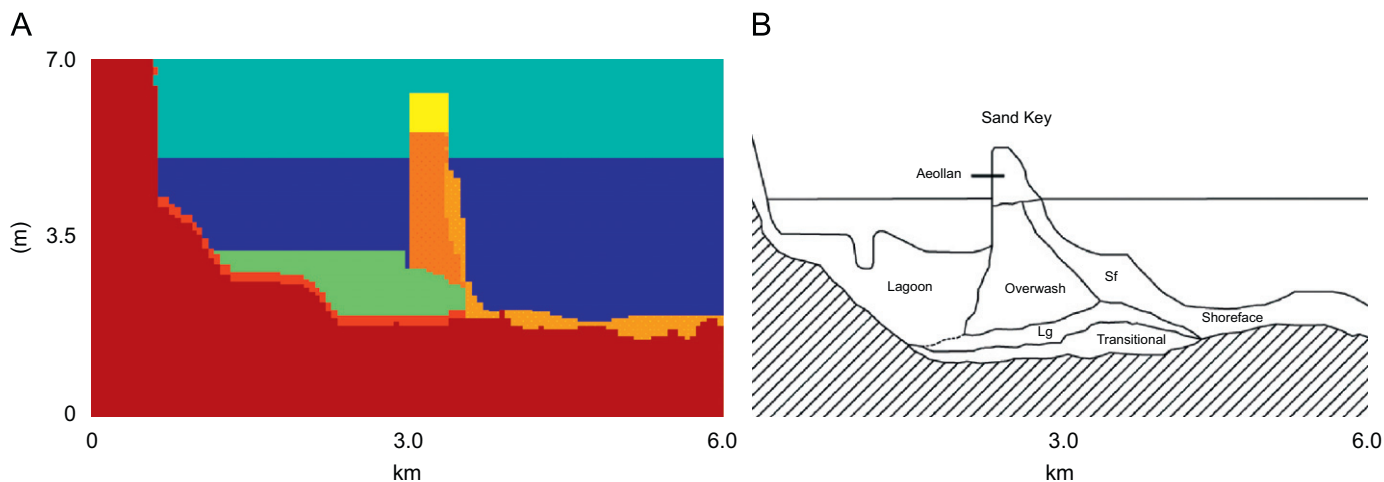

Fig. 7. Comparison between (a) the stratigraphic profile of Sand Key simulated by the model and the (b) profile reconstructed from sediment cores (modified after Evans et al., 1985).

These sediments cannot be redistributed by our wave action formulation (Eq. (2)), and they thus form a sand deposit (Fig. 6). The sand deposits along the shelf are then the result of a landward translation of the barrier island over an uneven substrate rather than being the remains of drowned barrier islands (cf. Figs. 4 and 6). If we compare the barrier island migration rate to the morphology of the shelf, we notice that the barrier stops at three locations for several centuries, and at all locations the bedrock substrate is steep, thus confirming our hypothesis (Fig. 8).

This mechanism would not explain the presence of sediment deposits with positive relief (Fig. 4), but these convexities are

more likely large bedforms or sediment bars produced by sediment reworking by waves and currents. A different value for the closure depth does not modify the evolution of the barrier, but increases the thickness of reworked sediment at the shelf bottom.

When the rate of overwash is low or sea level rises too fast, the upward growth of the barrier cannot be fast enough to counteract the increase in sea level. In this scenario, the island drowns and large volumes of sediments are transported to the mainland shoreline. A simulation carried out with a lower rate of overwash ($A = 0.2\,\mathrm{m^2/month}$) shows the drowning of the barrier (Fig. 9). All the barrier sediments are transported to the mainland shoreline in
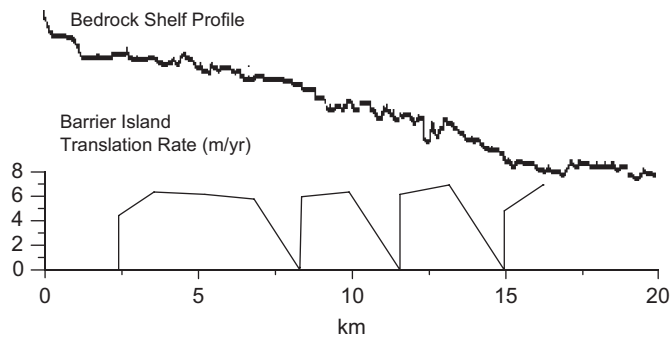
Fig. 8. Comparison between bedrock shelf profile and barrier island translation rate at each shelf location.

few decades. As soon as the barrier elevation is below mean sea level, the sediment is remobilized by the action of waves through Eq. (2). If the barrier drowns, most of the sediment is transported by the waves directly to the modern shoreline and not deposited at intermediate positions on the shelf. The shelf is in fact too flat to allow the reforming of the barrier island at intermediate locations. Moreover, if the barrier island had reformed just recently at the shoreline, its stratigraphy would be different from the one derived from the cores. In Fig. 10, we compare the stratigraphic architecture at the shoreline for a simulation with and without barrier island drowning. It is easy to see that in Fig. 10a all the different sediment facies (transitional, aeolian, shore-face, overwash, and lagoonal) reported by Evans et al. (1985) are present. On the contrary, during the drowning of the barrier, all the sediments are reworked by waves and transformed in shoreface, determining in this way the presence of only two sediment facies: shoreface and overwash, which can be found at the shoreline on the top of the barrier remains (Fig. 10b).

A set of simulations is carried out considering Progressions A and B of Balsillie and Donoghue (2004). With Progression A, the barrier island is able to reach its current position, but only considering a rate of overwash higher than that of the first simulation. This was necessary because the mean sea level does not rise constantly, but oscillates. The simulated stratigraphic architecture is similar to the one reported in Evans et al. (1985). The evolution of Sand Key was simulated also considering the Progression B of mean sea level (Fig. 11). In this case, the barrier rapidly drowns due to the high oscillations of sea level. When sea level decreases, as in the case of Progression B, the island starts prograding seaward. This happens because the sediments eroded along the shelf profile is transported landward until the barrier island; these sediments cannot be removed by overwash and therefore accumulate on the barrier shoreface.

6. Discussion and conclusions

Our numerical model, developed to simulate the dynamics of the evolution of a barrier island, is able to correctly reproduce the current configuration of the shelf profile in front of Sand Key, Florida. Our results stress the importance of a separate account for the effects of high and low frequency components of sediment reworking. Wind waves have a fundamental role in the reworking of the beach profile, while storm surges during extreme events induce the translation of the sand barrier. Model simulations show how the translation of the barrier is dictated by the abundance or scarcity of sediments. In particular, the rate of overwash influences the evolution of the sand barrier. The barrier survives when overwash is frequent enough to allow island

translation, otherwise the accommodation space in the back-barrier grows with sea-level rise, and the island will eventually drown during the next major storm (Figs. 2 and 3).

Our model formulation does not include the influence of along-shore sediment transport on barrier formation and evolution. Moreover, the bi-dimensional formulation presented herein precludes the reformation of the barrier island by spit extension (Gilbert, 1890), which is a typical three-dimensional process. External fluxes or losses of sediments, like sediment abrasion or offshore sediment export, could deeply modify the evolution of the barrier island, and clearly deserves adequate investigation in future research. A sensitivity analysis on the model parameters reveals the importance of the overwash rate in the barrier island evolution (Figs. 2 and 3). High overwash rates allow the barrier to translate inshore thus counteracting an increase in sea level. On the contrary, small washover volumes decrease the barrier translation rate, increasing the chance of barrier drowning during sea-level rise.

A recent conceptual model based on the distribution of marine sediments and on the facies ages in the west-central Florida shelf indicates that around 7000–8000 yr BP the low rate of sea-level rise and the abundance of Pleistocene material probably led to the development of barrier islands in excess of 20 km offshore the present coast. The islands then translated during sea-level rise until reaching the current position (Brooks et al., 2003). Our model simulations respect the sediment architecture beneath the barrier island reported in Brooks et al. (2003), with a pre-Holocene unit unconformably overlain by lagoonal facies, which in turn is overlain by a ravinement surface and open-marine facies. Brooks et al. (2003) report the presence of lagoonal facies up to 20 km offshore. In our model instead the sediment reworking by wave action transforms all the lagoonal sediment in shoreface deposits. However, some lagoonal facies can be preserved if the depth of shoreface reworking is reduced in Eq. (2). This notwithstanding our model results show that the presence of lagoonal facies in the shelf is a condition sufficient but not necessary for the past existence of barrier islands at that location, and facies preservation in the geological record depends only on the thickness of the sediments and the depth of sediment reworking by waves.

Model simulations indicate that barrier island translation was not uniform, but occurred in definite steps as a function of bottom topography (shelf slope) and sediment availability (Fig. 8). A different hypothesis concerning the history of the barrier islands in this shelf location indicates a recent formation near the present location by sediment aggradation and shoal emergence (Davis et al., 2003). Under this scenario pre-historic barriers might have been present during periods of lower sea level, but they just formed and disappeared without significant translation. This hypothesis is partly justified by the age of the current barriers, which is less than 3000 yr BP (Stapor et al., 1991). Our model can also provide critical insight under these conditions. The barrier could indeed disappear if the average volume of sediments moved in the back-barrier area by overwash is not sufficient to maintain the translation speed (Fig. 9), but then the facies architecture of the newly formed barrier would not be compatible with coring data (Fig. 10). More importantly, given the gentle slope of the west-central Florida shelf, the drowning process could have happened only once during the last 8000 yr. In fact, after the disappearance of the barrier, the shelf would have been completely submerged and a new barrier could have formed only near the present location (e.g., Fig. 11). The model can also explain the recent age of the barrier deposits reported in Stapor et al. (1991) as a by-product of sediment reworking. Model simulations show that the age and the stratigraphic architecture constantly evolve during barrier island translation, with old sediment
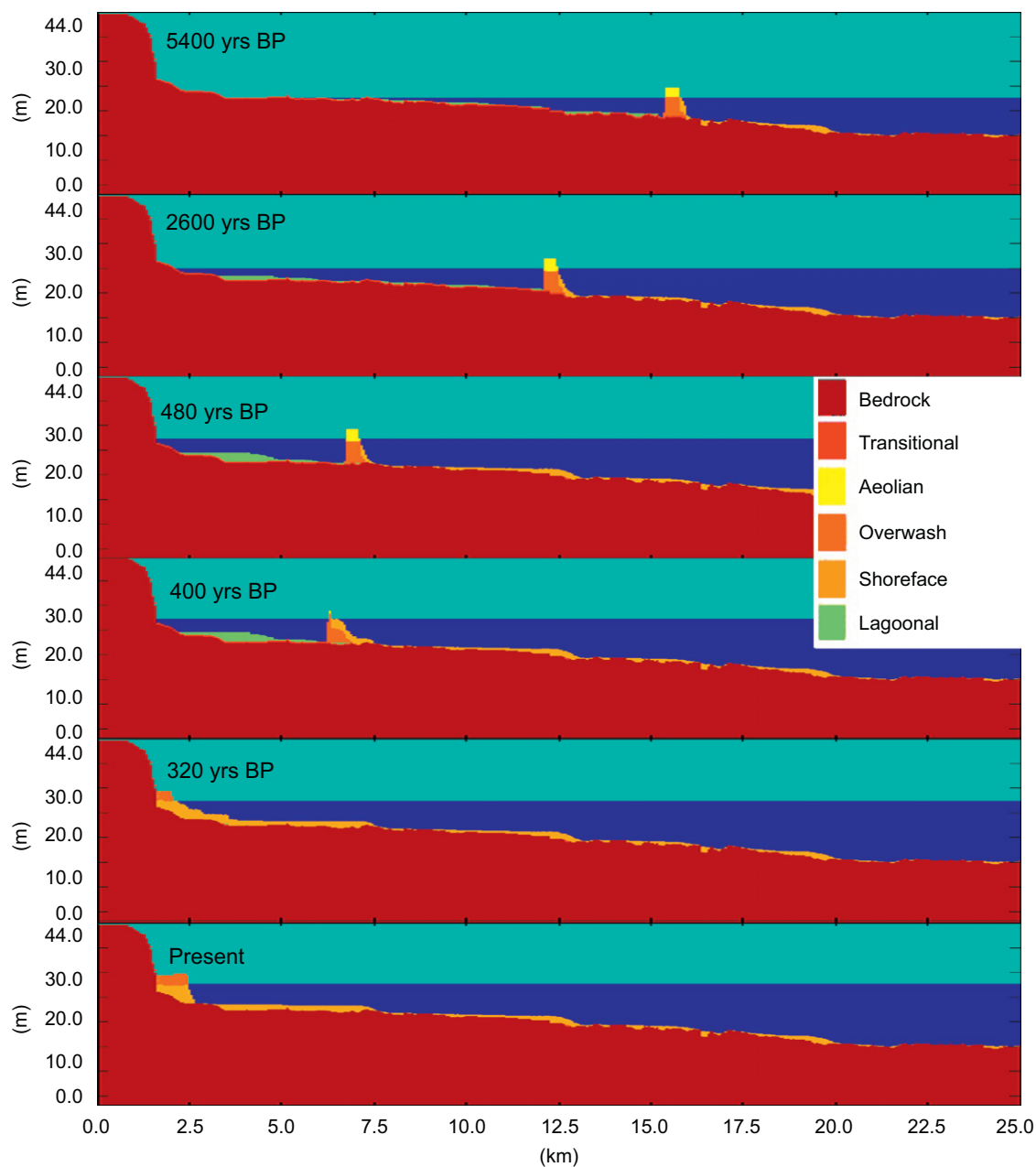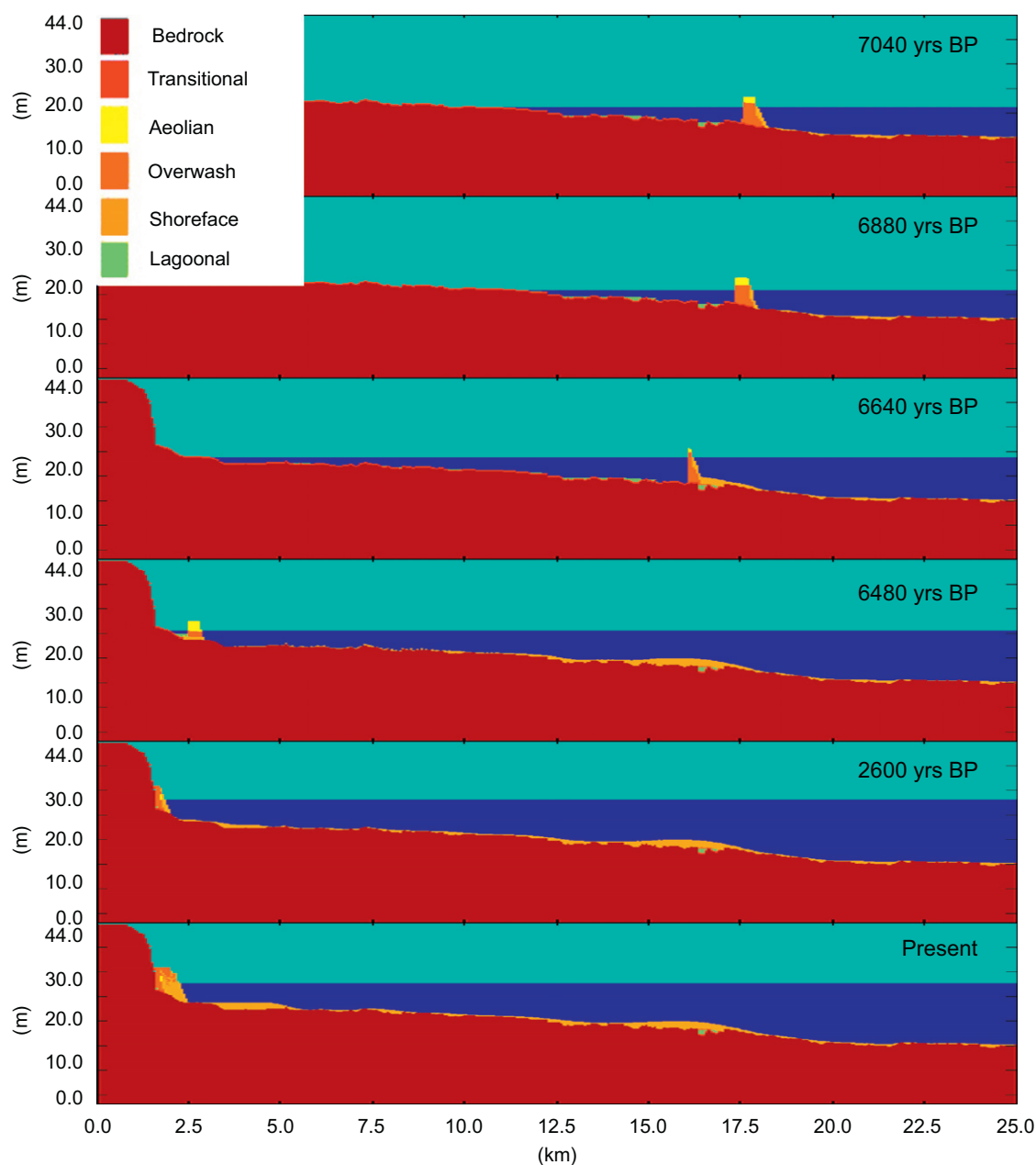
Fig. 9. Evolution of the barrier island of Sand Key with the sea level Progression of Fairbanks (1989) and a rate of overwash $A = 2\,m^2/month$.
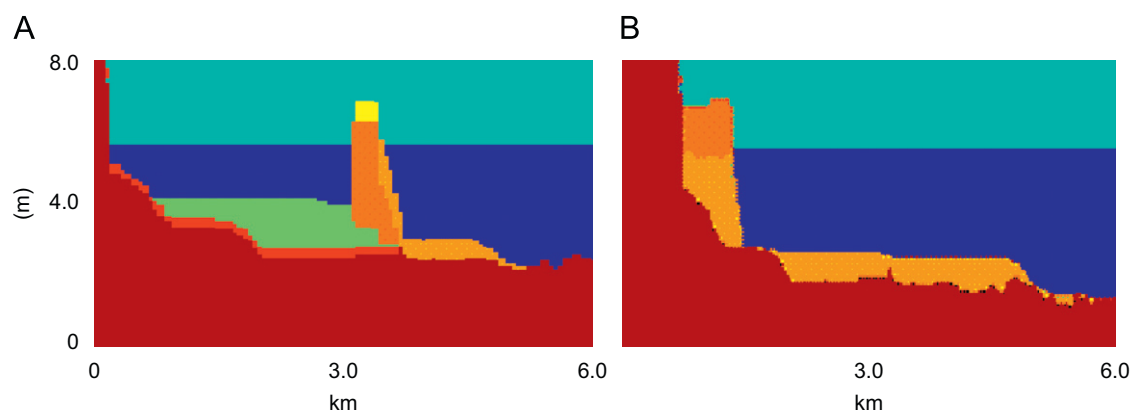
Fig. 10. Comparison between the stratigraphic profile of Sand Key at the end of the simulation considering the Progression A of mean sea level and a overwash rate of (a) $3\,m^2/month$ and (b) $2\,m^2/month$ (after the drowning of the barrier).

Fig. 11. Evolution of the barrier island of Sand Key with the sea level Progression B of Balsillie and Donoghue (2004).

deposited in the shelf and new sediment entrained from the bottom. As suggested by model simulations, the most relevant evidence of barrier island translation versus cycles of drowning and reforming can be derived only by a careful analysis of the overwash process involving an estimate of the volume of sediments moved in the back-barrier area and a characterization of hurricane and storm frequency in the last 8000 yr. Finally, our simulations are in contrast with the hypothesis of Evans et al. (1985) of a reduction of barrier translation in correspondence of a flat substrate morphology. On the contrary, results show that the barrier translation reduces in locations where the shelf is steep, since a reduction in accommodation space in the back-barrier allows the barrier to develop faster in the vertical direction, keeping pace with sea-level rise (Fig. 8).

On the other hand, a flat shelf terrace favors the translation of the island, followed by sediment deposition at the bottom (Figs. 6 and 8). Therefore, large sediment bodies present along the shelf are not the remains of barriers previously disappeared, but are more likely the direct consequence of the landward translation of the island over an uneven shelf bottom.

References

Balsillie, J.H., Donoghue, J.F., 2004. High resolution sea level history for the Gulf of Mexico since the last glacial maximum. Report of Investigations No. 103, Florida Geological Survey.

Blum, M.D., Carter, A.E., Zayac, T., Goble, R., 2002. Middle Holocene sea-level and evolution of the Gulf of Mexico coast (USA). Journal of Coastal Research, Special Issue 36, 65–80.

Brooks, G.R., Doyle, L.J., 1998. Recent sedimentary development of Tampa Bay, Florida: a microtidal estuary incised into tertiary platform carbonates. Estuaries 21 (3), 391–406.

Brooks, G.R., Doyle, L.J., Suthard, B.C., Locker, S.D., Hine, A.C., 2003. Facies architecture of the mixed carbonate/siliciclastic inner continental shelf of west-central Florida: implications for Holocene barrier development. Marine Geology 200 (1–4), 325–349.

Cherry, R.N., Stewart, J.W., Mann, J.A., 1970. General hydrology of the middle Gulf area, Florida. US Geological Survey, Investment No. 56, 96pp.

Cowell, P.J., Roy, P.S., Jones, R.A., 1995. Simulation of large-scale coastal change using a morphological behaviour mode. Marine Geology 126, 45–61.

Cowell, P.J., Stive, M.J.F., Niedoroda, A.W., Swift, D.J.P., Vriend, H.J., de Buijsman, M.C., Nicholls, R.J., Roy, P.S., Kaminsky, G., Cleveringa, J., Reed, C.J., de Boer, P.L., 2003. The coastal tract (Part 2): applications of aggregated modeling of low-order coastal change. Journal of Coastal Research 19 (4), 828–848.

Davis, R.A., Fitzgerald, D., 2002. Beaches and Coasts. Blackwell Science, 448pp.

Davis, R.A., Yale, K.E., Pekala, J.M., Hamilton, M.V., 2003. Barrier island stratigraphy and Holocene history of west-central Florida. Marine Geology 200 (1–4), 103–123.

Dean, R.G., 1991. Equilibrium beach profiles—characteristics and applications. Journal of Coastal Research 7 (1), 53–84.

Dean, R.G., Dalrymple, R.A., 2002. Coastal Processes with Engineering Applications. Cambridge University Press, New York.

Dean, R.G., Maurmeyer, E.M., 1983. Models of beach profile response. In: Komar, P., Moore, J. (Eds.), CRC Handbook of Coastal Processes and Erosion. CRC Press, Boca Raton, FL, pp. 151–165.

deBeaumont, L.E., 1845. Lecons de Geology Pratique, 7me Lecon-Levees de Sables et Galets, Paris.

Evans, M.W., Hine, A.C., Belknap, D.F., Davis Jr., R.A., 1985. Bedrock controls on barrier island development: west-central Florida coast. Marine Geology 63, 263–283.

Fagherazzi, S., Overeem, I., 2007. Models of deltaic and inner continental shelf landform evolution. Annual Review of Earth and Planetary Sciences 35, 685–715.

Fagherazzi, S., Wiberg, P.L., Howard, A.D., 2003. Modeling barrier island formation and evolution. In: Paper presented at Coastal Sediments 2003, Clearwater Beach, Florida, 18–23 May, 9pp.

Fairbanks, R.G., 1989. A 17,000-year glacio-eustatic sea level record; influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation. Nature 342, 637–642.

Fairbanks, R.G., 1990. The age and origin of the "Younger Dryas climate event" in Greenland ice cores. Paleoceanography 5 (6), 937–948.

Faught, M.K., Donoghue, J.F., 1997. Marine inundated archaeological sites and paleofluvial systems: examples from a karst-controlled continental shelf setting in Apalachee Bay, northeastern Gulf of Mexico. Geoarchaeology 12, 416–458.

Flemings, P.B., Jordan, T.E., 1989. A synthetic stratigraphic model of foreland basin development. Journal of Geophysical Research: Solid Earth and Planets 94 (B4), 3851–3866.

Gilbert, G.K., 1890. Lake Bonneville: US Geological Survey Monograph 1, 438pp.

Hoyt, J.H., 1967. Barrier island formation. Geological Society of America Bulletin 78, 1125–1136.

Jervey, M.T., 1988. Quantitative geological modelling of siliclastic rock sequences and their seismic expression. In: Wilgus, C.K., Posamentier, H., Roos, C.A., Kendall, C. (Eds.), Sea Level Changes—An Integrated Approach, vol. 42. SEPM Special Publication, pp. 47–69.

Kriebel, D.L., Dean, R.G., 1993. Numerical simulation of time-dependent beach and dune erosion. Coastal Engineering 9, 221–245.

Kriebel, D.L., Kraus, N.C., Larson, M., 1991. Engineering methods for predicting beach profile response. In: Paper presented at Coastal Sediments '91 ASCE, New York, pp. 557–571.

Kuehn, D.W., 1980. Offshore transgressive peat deposits of southwest Florida. Evidence for a late Holocene rise of sealevel. M.S. Thesis, Department of Geology, Pennsylvania University, 104pp.

Leatherman, S.P., Zaremba, R.E., 1987. Overwash and aeolian processes on a United States Northeast coast. Sedimentary Geology 52 (3–4), 183–206.

Leatherman, S.P., Williams, A.T., Fisher, J.S., 1977. Overwash sedimentation associated with a large-scale northeaster. Marine Geology 24, 109–121.

Locker, S.D., Hine, A.C., Brooks, G.R., 2003. Regional stratigraphic framework linking continental shelf and coastal sedimentary deposits of west-central Florida. Marine Geology 200 (1–4), 351–378.

McBride, R.A., 1997. Seafloor morphology, geologic framework, and sedimentary processes of a sand-rich shelf offshore Alabama and northwest Florida, northeastern Gulf of Mexico. Ph.D. Dissertation, Department of Oceanography and Coastal Sciences, Louisiana State University, 509pp.

Morton, R.A., Paine, J.G., Blum, M.D., 2000. Responses of stable bay-margin and barrier-island systems to Holocene sealevel highstands, western Gulf of Mexico. Journal of Sedimentary Research 70, 478–490.

Murray, A.B., 2003. Contrasting the goals, strategies, and predictions associated with simplified numerical models and detailed simulations. In: Iverson, R.M., Wilcock, P.R. (Eds.), Prediction in Geomorphology, vol. 135. AGU Geophysical Monograph, pp. 151–165.

Plant, N.G., Holman, R.A., Freilich, M.H., Birkemeier, W.A., 1999. A simple model for interannual sandbar behavior. Journal of Geophysical Research-Oceans 104 (C7), 15755–15776.

Pilkey, O.H., Young, R.S., Riggs, S.R., Smith, A.W.S., Wu, H.Y., Pilkey, W.D., 1993. The concept of shoreface profile of equilibrium—a critical review. Journal of Coastal Research 9 (1), 255–278.

Robbin, D.M., 1984. A new Holocene sea-level curve for the upper Florida Keys and Florida reef tract. Miami Geological Society Memoir 2, 437–458.

Roy, P.S., Cowell, P.J., Ferland, M.A., Thom, B.G., 1994. Wave dominated coasts. In: Carter, R.W.G., Woodroffe, C.D. (Eds.), Coastal Evolution: Late Quaternary Shoreline Morphodynamics. Cambridge University Press, Cambridge, pp. 121–186.

Shepard, F.P., 1960. Rise of sea-level along northwest Gulf of Mexico. In: Shepard, F.P., et al. (Eds.), Recent Sediments, Northwest Gulf of Mexico. American Association of Petroleum Geologists, pp. 338–344.

Stapor, F.W., Stone, G.W., 2004. A new depositional model for the buried 4000 BP New Orleans Barrier: implications for sea-level fluctuations and onshore transport from a nearshore shelf source. Marine Geology 204, 215–234.

Stapor, F.W., Mathews, T.D., Lindforskearns, F.E., 1991. Barrier-island progradation and Holocene sea-level history in Florida. Journal of Coastal Research 7 (3), 815–838.

Stolper, D., List, J.H., Thieler, E.R., 2005. Simulating the evolution of coastal morphology and stratigraphy with a new morphological-behaviour model (GEOMBEST). Marine Geology 218, 17–36.

Storms, J.E.A., Weltje, G.J., Van Dijke, J.J., Geel, C.R., Kroonenberg, S.B., 2002. Process-response modeling of wavedominated coastal systems: simulating evolution and stratigraphy on geological timescales. Journal of Sedimentary Research 72 (2), 226–239.

Tanner, W.F., Demirpolat, S., Stoper, F.W., Alvarez, L., 1989. The "Gulf of Mexico" late Holocene sea level curve. Transactions of the Gulf Coast Association of Geological Societies 39, 553–562.

Thieler, E.R., Brill, A.L., Cleary, W.J., Hobbs III, C.H., Gammisch, R.A., 1995. Geology of the Wrightsville Beach, North Carolina shoreface: implications for the concept of shoreface profile of evolution. Marine Geology 126, 271–287.

Thieler, E.R., Pilkey Jr., O.H., Young, R.S., Bush, D.M., Chai, F., 2000. The use of mathematical models to predict beach behaviour for US coastal engineering: a critical review. Journal of Coastal Research 16 (1), 48–70.

van Straaten, L.M.J.U., 1965. De bodem der Waddenzee. In: Abrahamse, J., Buwalda, J.D., van Straaten, L.M.J.U. (Eds.), Het Waddenboek. Thieme, pp. 75–151.

Wright, L.D., 1995. Morphodynamics of Inner Continental Shelves. CRC Press, Boca Raton, FL, 256pp.

Zenkovitch, V.P., 1962. Fundamentals of the Coastal Science. Moscow, AS USSR Publishers (in Russian).