Model for a dune field with exposed water table

Marco C. M. de M. Luna\textsuperscript{a}; Eric J. R. Parteli\textsuperscript{a,b}; Hans J. Herrmann\textsuperscript{a,c}

\textsuperscript{a}Departamento de Física, Universidade Federal do Ceará - 60455-760, Fortaleza, CE, Brasil.
\textsuperscript{b}Programa de Pós-Graduação em Engenharia Química, Universidade Federal do Ceará, 60455-900, Fortaleza, CE, Brazil.
\textsuperscript{c}Institut für Baustoffe IJB, ETH Hönggerberg, HIF E 12, CH-8093, Zürich, Switzerland.

Abstract

Aeolian transport in coastal areas can be significantly affected by the presence of an exposed water table. In some dune fields, such as in Lençóis Maranhenses, northeastern Brazil, the water table oscillates in response to seasonal changes of rainfall and rates of evapotranspiration, rising above the ground during the wet season and sinking below in the dry period. Understanding dune mobility in an environment with varying groundwater level is essential for the study of long-term evolution of many dune fields. Here we apply a model for aeolian dunes to study the genesis of coastal dune fields in presence of an oscillating water table. We find that the morphology of the field depends on the time cycle, $T_w$, of the water table and the maximum height, $H_w$, of its oscillation. Our calculations show that long chains of barchanoids alternating with interdune ponds such as found at Lençóis Maranhenses arise when $T_w$ is of the order of the dune turnover time, whereas $H_w$ dictates the growth rate of dune height with distance downwind.

Key words: Coastal dunes, Water table, Wind erosion, Sand transport, Dune model

1. Introduction

It is widely accepted that dune morphology depends fundamentally on the wind directionality and on the amount of sand available for transport (Wasson and Hyde, 1983). While longitudinal and star dunes form under bi- and multidirectional wind regimes, respectively, the best understood types of dune are formed by unidirectional wind. In this case, crescent-shaped barchans occur if the sand availability is low, while transverse dunes, which display nearly invariant profile in the direction orthogonal to the wind, form if the ground is covered with sand. The shape of dunes can be significantly modified due to natural agents such as vegetation growth (Hesp, 2002; Tsoar and Blumberg, 2002; Barbosa and Domínguez, 2004) or cementation of sand by mineral salts (Schatz et al., 2006). Modeling has brought many insights on dune formation in environments with stabilizing vegetation (Nishimori and Tanaka, 2001; Baas, 2002; Durán and Herrmann, 2006a; Yizhaq et al., 2007; Nield and Baas, 2008a,b; Luna et al., 2011). The influence of sand induration on the shape of dunes has been also investigated (Herrmann et al., 2008; Rubin and Hesp, 2009). The water table can also play a fundamental role in sediment transport on coasts and in deserts (Kocurek et al., 1992; Ruz and Meur-Perec, 2004; Kocurek et al., 2007; Mountney and Russell, 2009). The dynamics of the water table in some aeolian sand systems is connected with seasonal variations in climate and rainfall (Levin et al., 2009), and has been modeled by many different authors (de Castro Ochoa and Muñoz Reinoso, 1997; Kocurek et al., 2001). However, quantitatively little is known about dune field morphodynamics in presence of a varying groundwater level.

One example of dunes evolving in presence of a dynamic water table is the coastal dune field known as “Lençóis Maranhenses”, located in the State of Maranhão, northeastern Brazil (Fig. 1). Long chains of laterally linked barchans (the so-called “barchanoids”) extending over several kilometers constitute the dominant dune morphology at Lençóis (Fig. 1a). Dunes at Lençóis alternate with freshwater lagoons that form in the rainy seasons and, at some places, nearly disappear in the summer. In spite of the field research undertaken in the past (Gonçalves et al., 2003; Parteli et al., 2006; Levin et al., 2007; Kadau et al., 2009), it has remained unclear how the oscillating water table is contributing to the long-term dynamics and to the shape of dunes at Lençóis.

Understanding the effect of the water table on dune field evolution may help predict coastal dune mobility and elucidate the history of many ancient deserts (Davidson-Arnott and Pyskir, 1988; Enzel et al., 1999; Mountney and Thompson, 2002; Chen et al., 2004; Grotzinger et al., 2005; Bourke and Wray, 2011). A physically based model that accounts for a mathematical description of sand transport and for the evolution of the terrain in response to aeolian drag and variations in the groundwater level can provide a helpful tool in the investigation of dune fields with exposed water table.

In the present work, we adapt a recently developed con-
tinuum model for aeolian dunes (Sauermann et al., 2001; Kroy et al., 2002) to study the formation of coastal dune fields in an area of exposed groundwater. This model has been applied to study different types of dune encountered in nature, and has proven to reproduce the shape of real dunes yielding quantitative agreement with measurements (Sauermann et al., 2003; Parteli et al., 2006; Durán and Herrmann, 2006a,b; Parteli and Herrmann, 2007; Durán et al., 2008; Parteli et al., 2009, 2011). The dune model encodes a mathematical description of the turbulent wind field over the terrain with a continuum model for saltation — which consists of grains hopping in ballistic trajectories close to the ground and ejecting new particles upon collision with the bed (Bagnold, 1941). Here we add to the dune model a water table that can rise and sink seasonally, thus affecting local rates of sand transport during the evolution of the dune field. In particular, we aim to understand the genesis and evolution of dunes in presence of a dynamic water level in order to shed light on the conditions leading to the dune morphologies observed at Lençóis Maranhenses.

2. Lençóis Maranhenses

The National Park of Lençóis Maranhenses is located on the coastal area of the Maranhão State, northeastern Brazil. The area comprises 155 thousand hectares delimited by the coordinates S 02°19’ and 02°45’, and W 42°44’ and 43°29’ (Figs. 1a, b). Climate at Lençóis is semi-humid tropical with sparse vegetation, air humidity of 68% and annual average temperature about 28.5°C (IBAMA, 2003; Floriani et al., 2004).

The landscape of Lençóis is characterized by the presence of long chains of barchanoids (Fig. 2) giving the impression of a crumpled sheet — hence the origin of the name: “Lençóis” means “sheets” in portuguese. These dunes extend along 75 km and penetrate inland to distances larger than 20 km. They are detached from the coast by a deflation plane of width between 600 m and 2000 m, and are migrating on top of dunes of older generations (Tsoar et al., 2009). Insights on the formative process and early development stages of dunes at Lençóis Maranhenses have been gained from recent field research (Gonçalves et al., 2003; Parteli et al., 2006). Sand deposited by tides on the beach is eroded by the wind thus forming small barchans with height between 50 cm and 1 m close to the beach (Parteli et al., 2006). As the barchans advance downwind, they become larger, link laterally and give place to barchanoidal chains that can reach heights of 30 m. The sand of the dunes is composed by quartz grains of mean diameter varying between 120µm and 350µm (IBAMA, 2003), which are values around the average diameter $d = 250µm$ previously reported for grains of sand dunes in other fields (Bagnold, 1941; Pye and Tsoar, 1990).

Wind regime at Lençóis is strongly unidirectional. The corresponding sand rose shows that the strongest winds, which can reach velocities over 8 m/s (Jiménez et al., 1999), blow mainly from the East (c.f. Fig. 3a). The vector length of each direction of the sand rose gives the potential rate of sand transport from that direction, or the drift potential (Tsoar, 2001),

$$DP = \sum k u^2 |u - u_t| f,$$  \hspace{1cm} (1)

where $k \approx 7.3$, $u$ is the wind velocity (in m/s) at a height of 10 m and $f$ is the fraction of time the wind was above $u_t$, which is the threshold wind velocity for saltation (Tsoar,
The value of $u_t$ can be roughly estimated (assuming that the constituent sand is not wet and neglecting vegetation cover, cohesion between the grains or sand induration) through the scaling with the threshold shear velocity for saltation, $u_{st}$, and as such, $u_t$ scales with the square root of the grain diameter (Bagnold, 1941; Tsoar, 2001). The resultant drift potential ($RDP$) is obtained by calculating the vector sum of the drift potential for each one of the directions of the sand rose. Wind directionality can be then quantified in terms of the ratio $\beta = RDP/DP$, where $DP$ is the sum of the magnitude of the drift potential for all directions. A value of $\beta$ close to unity or close to zero means unidirectional or multidirectional wind regime, respectively. At Lençóis, $\beta$ is approximately 0.97, which is consistent with the unidirectional wind regime of the area (Tsoar et al., 2009).

Wind power at Lençóis is negatively correlated with rainfall, as depicted in Fig. 3b. During the dry season from August to December month-averaged values of wind velocity are much higher than during the wet season. From January to July, when almost 93% of the rainfall takes place (Jiménez et al., 1999), the interdune areas are inundated due to the high water table thus forming ponds (Fig. 2). The lagoons, placed amidst very clean sand, have no inlet or outlet and are exclusively filled by rain water. Their bottom is covered by a soft brown or green sheet of algae and cyanobacteria (Kadau et al., 2009). The interdune ponds cover 41% of the area of Lençóis (Levin et al., 2007). Levin et al. (2007) estimate that the average interdune pond area at the end of the rainy season is 7 ha, while most of the ponds are longer in the cross-wind direction than in the along-wind direction. Using the definition of Robinson and Friedman (2002), the ponds at Lençóis have a mean circularity — where the circularity of a sphere is one — of about 0.47 (Levin et al., 2007). A large fraction of the interdune lakes formed in the rainy season disappear in the dry season. The landscape of Lençóis appears to change continuously, indeed dune mobility implies that the lakes often reappear in different places with different contours. The effect of the fluctuating water level on dune morphology at Lençóis is still poorly understood.

3. The model

The model used in the calculations of the present work consists of a set of mathematical equations that describe the average surface shear stress ($\tau(x, y)$) over the topography, the mass flux ($q(x, y)$) of saltating particles and the time evolution of the surface resulting from particle transport (Sauermann et al., 2001; Kroy et al., 2002; Durán and Herrmann, 2006b; Durán et al., 2010). Here, the model is extended in order to account for the water table. In the model the following steps are solved in an iterative manner.
stress: graphically induced perturbations in the average shear surface is calculated solving a set of analytical equations developed by Weng et al. (1991). The wind model computes, first, the Fourier-transformed components of the topographically induced perturbations in the average shear stress:

\[
\tilde{\tau}_x = \frac{\tilde{h}_x k_x^2}{|k|} \frac{2}{U^2(l)} \left\{-1 + \left(2 \ln \frac{l}{z_0} + \frac{1}{k_x^2} \right) K_1(2\sigma) K_0(2\sigma) \right\},
\]

\[
\tilde{\tau}_y = \frac{\tilde{h}_y k_y}{|k|} \frac{2}{U^2(l)} 2\sqrt{2\sigma} K_1(2\sqrt{2\sigma}),
\]

where \( x \) and \( y \) are the components parallel, respectively, perpendicular to the wind direction, \( \vec{k} \) is the wave vector, and \( k_x \) and \( k_y \) its coordinates in Fourier space; \( \tilde{h}_x \) is the Fourier transform of the height profile, which is defined as the envelope comprising the sand landscape and the level of exposed water; \( U \) is the normalized vertical velocity profile, \( l \) is the inner layer depth of the flow and \( L \) is 1/4 the mean wavelength of the Fourier representation of the height profile; \( \sigma = \sqrt{\alpha K_2 z_0}/l \); \( K_0 \) and \( K_1 \) are modified Bessel functions, and \( z_0 \) is the aerodynamic roughness of the surface. The average shear stress is obtained, then, from the equation,

\[
\bar{\tau} = \tau_0(\tilde{\tau}_0/\tau_0 + \tilde{\tau}),
\]

where \( \tau_0 \) is the undisturbed shear stress over the flat ground and \( \tau_0 \equiv |\tilde{\tau}_0| \).

(i) Given an initial topography, e.g. a terrain with dunes or smooth sand hills, the average shear stress field over the surface is calculated solving a set of analytical equations developed by Weng et al. (1991). The wind model computes, first, the Fourier-transformed components of the topographically induced perturbations in the average shear stress:

\[
\bar{\tau} = \tau_0(\tilde{\tau}_0/\tau_0 + \tilde{\tau}),
\]

where \( \tau_0 \) is the undisturbed shear stress over the flat ground and \( \tau_0 \equiv |\tilde{\tau}_0| \).

(ii) Next, the local sand flux over the landscape above the water level is calculated. The cloud of saltating grains is considered as a thin fluid-like layer moving over the surface. Once the wind speed exceeds the saltation threshold, the sand flux increases, first, exponentially due to multiplicative processes inherent to the splash events at grain-bed collisions. However, the wind strength decreases as more grains enter saltation, since these have to be accelerated at cost of aeolian momentum (Nishimura and Hunt, 2000; Almeida et al., 2006). After a saturation distance, the wind is just strong enough to sustain transport and the sand flux is maximal. By using mass and momentum conservation, and by explicitly accounting for the flux saturation transients, the following equation is derived for the height-integrated mass flux of sand per unit time and length (Sauermann et al., 2001),

\[
\tilde{\nu} \cdot \tilde{q} = (1 - |\tilde{q}|/q_s) |\tilde{q}|/\ell_s,
\]

where \( q_s = (2v_s^2 \alpha/g) \rho_s u_{4t}^2 ([u_{4t}/u_{4t}]^2 - 1) \) is the saturated flux and \( \ell_s = (2v_s^2 \alpha/g\gamma)/([u_{4t}/u_{4t}]^2 - 1) \) the characteristic length of flux saturation, whereas \( u_{4t} = \sqrt{\bar{\tau}/\rho_a} \) is the wind shear velocity and \( \rho_a = 1.225 \text{ kg/m}^3 \) is the air density; \( u_{4t} \), the impact threshold (Bagnold, 1941), is about 80% the fluid threshold velocity \( u_{4t \text{ ft}} \sim 0.26 \text{ m/s} \) required to initiate saltation; \( g = 9.81 \text{ m/s}^2 \) is gravity, and the average grain velocity, \( v_s \), is computed by taking the steady-state wind velocity within the saltation layer (Sauermann et al., 2001; Durán and Herrmann, 2006b), whereas \( \alpha = 0.43 \) and \( \gamma = 0.2 \) are empirically determined parameters (Sauermann et al., 2001; Durán and Herrmann, 2006b).

(iii) The local height, \( h(x, y) \), of the sand landscape evolves according to the equation

\[
\partial h/\partial t = -\tilde{\nu} \cdot \tilde{q}/\rho_b,
\]

where \( \rho_b = 1650 \text{ kg/m}^3 \) is the bulk density of the sand. If the local slope exceeds 34°, the unstable surface relaxes

Figure 4: Longitudinal profile of a transverse dune, \( h(x) \), and the separation bubble, \( s(x) \), both rescaled by the dune height, \( H \). The shear stress, \( \tau(x) \), and the sand flux, \( q(x) \), are normalized by the upwind shear stress, \( \tau_0 \), and the saturated flux, \( q_s \), respectively. The water level \( h_w(x) \) is rescaled by its maximum value, \( H_w \). Wind direction is from left to right.
through avalanches in the direction of the steepest descent. Avalanches are considered to be instantaneous since their time-scale is much smaller than the one of dune motion. The downslope flux of avalanches is calculated with the equation,

\[ q_{\text{aval}} = k \left[ \tanh(\nabla h) - \tanh(\theta_{\text{dyn}}) \right] \frac{\nabla h}{|\nabla h|}, \]

where \( k = 0.9 \) and \( \theta_{\text{dyn}} = 33^\circ \) is the so-called “dynamic” angle of repose (Durán et al., 2010). The calculation of the flux due to avalanches followed by the update of the angle of repose (Durán et al., 2010). The calculation of the avalanches applies to the whole sand topography, i.e. including the surface below the water level. The only difference between the transport under and above the water table concerns erosion due to the action of the wind. The surface below the water level is obviously protected from erosion and can only evolve in time due to avalanches following the deposition of sand incoming from dunes above the water level. At the interface with the exposed water, the sand flux vanishes and the incoming volume of sand is, then, instantaneously accreted to the local surface, at the first grid element where the sand surface is below the water level.

(iv) Thereafter, the level of exposed water, \( h_w(t) \), is updated. It is assumed that the water table is roughly proportional to rainfall (Tsoar et al., 2009), as observed in real situations (Kocurek et al., 1992). The water table oscillates in time \( t \) with a period \( T_w \) and maximum level \( H_w \) according to the equation:

\[ h_w(t) = H_w \sin[2\pi t/T_w], \]

which is chosen here to represent the nearly sinusoidal behaviour of average rainfall, as monitored over several seasons, in areas of coastal dunes (Jiménez et al., 1999; Tsoar et al., 2009) (see Fig. 3b).

The calculations are performed using open and periodic boundaries in the directions longitudinal \( (x) \) and perpendicular \( (y) \) to the wind, respectively.

4. Simulations of coastal dune fields

The initial surface is a sand beach, which is modeled as a flat transverse sand hill of height 1.5 m, width 80 m and Gaussian longitudinal profile, as depicted in Fig. 5. The transverse profile of the hill has small, random fluctuations of amplitude of the order of the grain diameter, \( d \approx 250 \mu m \) (Luna et al., 2011). The sand hill is subjected to a wind of constant average shear velocity \( u_s \). We take \( u_s \) values within the range 0.3–0.4 m/s, which corresponds to measured average velocity values of sand-moving winds in areas close to Lençóis Maranhenses (Jiménez et al., 1999; Parteli et al., 2006; Tsoar et al., 2009). However, it must be emphasized that the full range of variability of \( u_s \) in the dune field of Lençóis Maranhenses has not been measured, and so the aforementioned values are rather approximate values based on knowledge of average \( u_s \) values of sand-moving winds in other dune areas of northeastern Brazil (Jiménez et al., 1999; Tsoar et al., 2009).

The calculation of a coastal dune field in the absence of water table has been studied in previous modeling (Durán et al., 2010; Luna et al., 2011). As shown from these works, the condition for the genesis of a coastal dune field is a saturated influx coming from the sea. In that case net deposition occurs at the windward foot of the sand surface. The hill does not evolve into a migrating dune (Katsuki et al., 2005), instead it remains fixed, increases in size and flattens (Durán et al., 2010). The flat sand surface resulting from the initial hill is unstable and develops undulations — the so-called “sand-wave instabilities” (Elbelrhiti et al., 2005) — which evolve into small transverse dunes of height between 50 cm and 1 m. Once the transverse dunes reach the bedrock, they become unstable and give place to barchans (Reffet et al., 2010; Parteli et al., 2011). Average dune size increases with distance as small barchans merge and collide with larger, more slowly migrating dunes downwind. In this manner, a sand hill subjected to a saturated flux becomes a source of sand for a field of barchans. Figure 6 shows a snapshot of the calculation of dune field genesis in the absence of water table (Durán et al., 2010; Luna et al., 2011).

The calculations including the water table are also performed using a saturated flux condition at the inlet, i.e. \( q_{\text{in}} = q_s \). Furthermore, the initial surface is adapted by including an upwind area of width \( \Delta x \) protected from erosion, c.f. Fig. 5. This area represents a simple model for the backshore of an accreting beach (Dingler, 2005). At the backshore there is less potential to remove sand thus favouring accumulation (Tsoar, 2000). In fact, if the water table rises above the sand upwind of the crest of the hill, as illustrated in Fig. 5, then, at a later time the sand surface is fully subdued by the water level as the hill flat-
tens. In this case, input of sand ceases and no dune field forms. In order to avoid this problem and to make the model more realistic, we consider that the sand hill has a constant profile in the upwind backshore area, which has width $\Delta x \approx 25$ m. Whereas in reality the sand surface close to the beach and water table may have profiles much more complex than in the model depicted in Fig. 5 (Kocurek et al., 2001), modeling of the detailed process leading to the accumulation of sand at the backshore and the formation of the sand beach is out of the scope of the present work. Indeed, using different profiles for the sand beach or different values of $\Delta x$ does not change the results presented in the next Section, provided there is a sufficient amount of sand above the water level that serves as an upwind source of sand for the dune field.

5. Results and discussion

Next, the genesis of dunes is studied by accounting for the presence of a water table that oscillates as described in Section 3. The seasonal rise and lowering of the water table constrains dune dynamics. The evolution of the field depends on the quantities controlling the dynamics of the water table, namely the time period ($T_w$) and the amplitude ($H_w$) of the oscillation (c.f. Eq. (9)).

One relevant quantity for dune field morphology is the time cycle, $T_w$, of the water table relative to the migration or turnover time, $T_m$, of the barchan, i.e. the time needed for the barchan to cover a distance approximately equal to its own width (Allen, 1974). The turnover time of a barchan of height $H$ is given by the equation (Hersen et al., 2004; Durán et al., 2010),

$$T_m \approx a \frac{H^2}{Q}, \quad (10)$$

where $Q(u_s) = q_s/\rho_s$ is the bulk sand flux associated with the wind shear velocity $u_s$, and the proportionality constant, $a$, is approximately equal to 3 (Parteli et al., 2011). During the time where the water table is above the ground, the dunes are locally supply-limited thus leading to deflation of the crests. In the limit where $T_w$ is much longer than $T_m$, dunes are nearly leveled off during inundation. In fact, if $T_w \gg T_m$, then the dune migration velocity, which scales as $1/T_m$, is much larger than the velocity of the oscillation of the water table (which is proportional to $1/T_w$). In this case, the crests of the dunes move large distances downwind while the lowest portions of the dunes remain under water. Consequently, sand above water level is spread throughout the field forming extensive flat regions in the wet season whereas the wind has ample time to reshape the dunes again the dry season. Conversely, if $T_w \ll T_m$, the amount of sand taken from the dunes during the wet season is small, thus leading to negligible accumulation in interdune areas. Furthermore, only a small amount of sand is transported between dunes in the dry season, such that isolated dunes separated by large flat areas dominate the morphology of the field.

An interesting scenario occurs when $T_w$ becomes of the order of $T_m$. In the wet season, the lowest portions of the dunes remain under water while the dune crest can migrate some distance downwind, thus leading to a decrease in dune height. Indeed, the amount of sand deposited at the interdune is large enough to construct small dunes in the dry season. Due to their fast relative migration velocity, the small bedforms emerging between the dunes collide with the larger dunes in their front, thus linking the barchans laterally and leading to the formation of barchanoidal chains. Therefore, the typical morphology observed at Lençóis, with chains of barchanoids alternating with interdune water ponds, arises when the size of dunes is such that their reconstitution time ($T_m$) is of the order of the time cycle of the water table ($T_w$). Indeed, taking $Q \approx 210$ m$^3$/year, which corresponds to $u_s = 0.36$ m/s, and $T_w = 1$ year, the condition $T_m \sim T_w$ is fulfilled by dunes with heights of the order of 10 m, which is consistent with the scale of dunes at Lençóis.

In Fig. 7 we present snapshots of calculations obtained with $u_s = 0.36$ m/s and different values of $T_w$, namely 1 month (a), 1 year (b) and 10 years (c). Each one of Figs. 7a, 7b and 7c shows one snapshot taken at a time where the water level is maximum, i.e. $h_w = H_w$ (bottom), and another for the case $h_w = -H_w$ (top), where $H_w$ has...
Figure 7: Snapshots of the calculation of dune field genesis after \( \sim 1000 \) years, obtained with wind shear velocity \( u_* = 0.36 \) m/s and amplitude of the water table \( H_w = 1.0 \) m. The period of the water table oscillation, \( T_w \), is (a) 0.1 year, (b) 1.0 year and (c) 10 years. In each figure, the images on top and on bottom show calculation snapshots when the water level is minimum and maximum, respectively. The blue color represents the water. Wind direction is from left to right.

The morphology of the field also depends on the maximum height, \( H_w \), reached by the water. Figure 8 shows snapshots of simulations performed using the nominal value \( T_w = 1 \) year and different values of \( H_w \): 20 cm (a), 1 m (b) and 2 m (c). As can be seen from Fig. 8, the average size of dunes increases with distance at a slower rate as \( H_w \) becomes larger. The average dune height at a distance of 5 km from the beach is about 28 m when \( H_w = 20 \) cm, and 12 m for when \( H_w = 1 \) m. For \( H_w = 2 \) m, the dune height is roughly constant throughout the field and is about 4 m. Figure 9 shows the dune height at \( x \approx 5 \) km as a function of \( H_w \) for different values of the wind shear velocity.
velocity, \( u_* \). We see that \( u_* \) and \( H_w \) play opposite roles for the height of dunes, i.e. while dunes become larger for increasing values of shear velocity, the water table reduces the spatial gradient of dune height in the direction of sand transport.

![Figure 8: Snapshots of the calculation of dune field genesis after \( \sim 1000 \) years, obtained with wind shear velocity \( u_* = 0.36 \) m/s and time cycle of the water table \( T_w = 1.0 \) year. The amplitude of the water table oscillation, \( H_w \), is (a) 20 cm, (b) 1.0 m and (c) 2.0 m. The calculation snapshot shown in each case corresponds to an instant where the water level is maximum. The blue color represents the water. Wind direction is from left to right.](image)

Figure 9: Dune height at a distance of \( \sim 5 \) km from the beach as a function of the maximum water level, \( H_w \), for different values of the wind shear velocity, \( u_* \).

Dune height increases with distance more slowly in the presence of an oscillating water table due to cyclic phases of destruction and construction associated with the rise and sinking of the water level. When the water table rises above the ground, sand blown from the dunes accumulates into wet interdune areas, leading to flattening of the dunes. The sinking of the water table in the dry season exposes the sand deposited in the lowest areas, which serve, then, as internal source for dunes throughout the field. The maximum level, \( H_w \), reached by the water table within the wet season determines the net amount dune’s sand deposited into flatlands — whereas deposition occurs either due to sand blown from exposed areas of the limbs or through avalanches along the slip-face, since the shear stress is assumed to be zero within the separation bubble (cf Section

![Figure 10: The plot shows the relative interdune area when the water level is minimum (squares) and the relative area of interdune water ponds when the water table reaches its maximum height (stars), for different values of the oscillation amplitude, \( H_w \). The calculations were performed with \( u_* = 0.36 \) m/s and \( T_w = 1 \) year, and the results shown in the figure refer to the area between 2.8 km and 5.6 km downwind from the beach. The data were averaged over 10 – 20 snapshots at time instants between 500 and 4000 years, when the water level was larger (stars) and smaller (square) than 95\% of \( H_w \) and \(-H_w\), respectively.](image)
3) — and thus the evolution of dune height in the field. Construction and destruction of dunes due to a fluctuating water table have been documented from field observations at the back-island dune fields on Padre Island, Texas (Kocurek et al., 1992). These fields constitute areas where the genesis and formative stages of dunes in the presence of a water table can be observed. Early portions of the fields on Padre Island are reduced to a nearly planar surface during the winter, when rainfall exceeds evaporation and the water table rises. Dunes are then reconstructed during the summer when groundwater is below the surface (Kocurek et al., 1992). As on Padre Island, the early landscape of the dune fields produced in the simulations consists of 50 cm high bedforms, which are nearly leveled in the wet season and then reform in the dry season. As also noted by Kocurek et al. (1992), dunes in the “mature” stage of the field far downwind from the beach have more chance of survival during wet seasons due to their larger size. Dune behaviour depends thus on the maximum water level relative to the equivalent sand thickness, \( \delta \), i.e. the thickness of sand if the barchan was leveled (Fryberger and Dean, 1979). Since the volume of a barchan is given by the equation \( V \approx cW^3 \), with \( c \approx 0.017 \) (Durán et al., 2010), and since \( V \approx \delta \cdot W^2 \), we obtain,

\[
\delta \approx cW, 
\]

which can be also written, using the linear relation between \( W \) and the dune height, \( H \) (Durán et al., 2010), as \( \delta \approx c \cdot [12H + 5] \). Therefore, the smallest dunes of width \( \sim 10 \) m and height \( \sim 1 \) m are nearly leveled in the wet season when \( H_w \) is around 20 – 30 cm. In fact, this prediction is supported by historical imagery of the dune field near Paulino Neves, located close to Lençóis Maranhenses, where it can be seen many examples of smaller dunes nearly deflating during the interval between the images (we thank an anonymous referee for this encouraging remark).

The effect of the oscillation amplitude of the water on the morphology of the field can be further elucidated through Fig. 10. This figure shows that the total interdune area during the dry season decreases with the amplitude of the water table oscillation. The larger \( H_w \) the larger the relative amount of sand contained in the interdune area, compared to the amount of sand in the dunes, and the smaller the relative area that is free of sand in the dry season. In contrast, the relative area of exposed water during maximum water level increases with \( H_w \), provided \( H_w \) is not too large. Figure 10 suggests that, for intermediate values of \( H_w \) smaller than 1.5 m, the relative interdune area can serve as a proxy for the maximum water level relative to dune height at a given location of a dune field. When the value of \( H_w \) increases beyond 1.5 m, the fraction of dune surface above the water level becomes so small that erosion of dune crest leads to flat sand sheets of thickness comparable to the dune height. As shown in Fig. 8c, these interdune sand sheets can fill large areas of the lagoons between dunes, thus leading to a decrease in the value of \( f_w \) as the water level becomes larger. Indeed, the average value of \( f_w \) for \( H_w = 2 \) m is slightly smaller than the one for \( H_w = 1.5 \) m, as shown in Fig. 10. For such large values of \( H_w \), the morphology of the field deviates from the typical landscape of barchanoidal chains observed at Lençóis Maranhenses (c.f. Fig. 8). Therefore, we take smaller values of \( H_w \), namely around 1 m.

Figure 11a shows an image of barchanoidal dunes at Lençóis Maranhenses, which are located within area I in Fig. 1a, downwind of a recently investigated field of transverse dunes of heights between 7 m and 10 m (Parteli et al., 2006). In Fig. 11b, we show a snapshot of a calculation obtained with \( H_w = 60 \) cm, while \( u_r = 0.36 \) m/s and \( T_w = 1 \) year are taken in consistence with the conditions at Lençóis Maranhenses. In both Figs. 11a and 11b we count 14 dune crests along the longitudinal cut indicated in the corresponding images. Furthermore, we calculate the total relative area of the terrain \( (f_w) \) which is covered by water, and find \( f_w = 28\% \) and 31\% for the dunes in the image and in the simulation, respectively.

![Figure 11](image)

**Figure 11:** (a) Image of barchanoids at Lençóis Maranhenses, within an area of 986 m × 1964 m; (b) barchanoids produced in the simulation using \( u_r = 0.36 \) m/s, \( H_w = 100 \) cm and \( T_w = 1 \) year.

Further downwind in the field of Lençóis Maranhenses, in area II of Fig. 1a, dunes are larger and can reach heights of about 15 – 20 m. In Fig. 12a, an image of barchanoids within area II is shown, whereas a snapshot of a simulation performed with \( H_w = 1 \) m is displayed in Fig. 12b. The value of \( f_w \) calculated for the real dunes is about 34\%, while for the dunes in the simulation we obtain \( f_w = 28\% \).

Therefore, the relative surface area \( (f_w) \) that is covered by water in the wet season as obtained from our calcula-
value of $f$ to the formation of well-separated barchans and a larger interdune flats of different sizes, which further contributes the sand is distributed in an irregular manner leading to permanent lagoons, indeed in some portions of the field and 12). In reality, the area of Lençóis also hosts large dune ponds between chains of barchanoids (c.f. Figs. 11 and 12). It has to be emphasized that the value of $f$ estimated from our calculations corresponds only to interdune ponds between chains of barchanoids (c.f. Figs. 11 and 12). In reality, the area of Lençóis also hosts large permanent lagoons, indeed in some portions of the field the sand is distributed in an irregular manner leading to interdune flats of different sizes, which further contributes to the formation of well-separated barchans and a larger value of $f_w$.

6. Conclusions

We extended the dune model introduced by Sauermann et al. (2001) to study dune formation in presence of a seasonally varying water table. The model was applied to investigate the genesis of dunes from a sand source in an environment where groundwater level oscillates in response to variations of rainfall and evapotranspiration rates. We found that the morphology of the field is dictated by the quantities controlling the dynamics of the water table, namely the time period ($T_w$) and the amplitude ($H_w$) of the oscillation. Chains of barchanoids, such as those at Lençóis Maranhenses dune field in northeastern Brazil, are obtained when $T_w$ is of the order of the dune turnover time, while $H_w$ determines the rate at which dune height increases in wind direction. We could reproduce barchanoidal dunes with the same size as those at Lençóis by using realistic values for $T_w$ (1 year) and $H_w$ (between 60 cm and 1 m) and an average wind shear velocity $u_\ast = 0.36 \text{ m/s}$ that is within the ranges observed at the real field. Furthermore, we found that the maximum water level affects the total relative interdune area in the dune field, $f_w$.

Finally, it would be interesting to combine the present model for dune formation in presence of a water table with the model for sand transport with vegetation growth (Durán and Herrmann, 2006a; Durán et al., 2008; Luna et al., 2011) in order to investigate the combined effect of those natural agents on the genesis and evolution of coastal dune fields. In particular, the present model does not account for the formation of residual dune ridges (RDRs), which are formed by vegetation growing along a line upwind of the dune and indicate the position of the dune during the rainy period in each year (Levin et al., 2009). Seasonal variations in vegetation growth due to the oscillating water table affect the dynamics of sand transport within the dune field and can be thus important for the evolution of the morphology of dunes in the field. In fact, residual transverse ridges stabilized by vegetation could be obtained previously using a simple extension of the dune model for isolated transverse dunes with seasonally varying vegetation growth rates (Luna et al., 2009). Therefore, a complete dune model that accounts for the evolution of RDRs is essential for future modeling studies of dunes at Lençóis Maranhenses and on a large scale in dune fields of northeastern Brazil (Levin et al., 2009).

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