Nonlinear modelling of shoreface-connected ridges; Impact of grain sorting and interventions

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Abstract

The evolution of shoreface-connected sand ridges on micro-tidal inner shelves and the variations in the mean grain size over these ridges are investigated with process-based models. A review of previous studies is presented, as well as new results that concern the influence of grain sorting on the finite-amplitude behaviour of the ridges, the application of the model to La Barrosa beach and the role of wave-topography feedbacks. The ridges initially form due to morphodynamic self-organisation, in which the presence of waves and a storm-driven current are crucial. Predicted growth time scales, migration speeds, topography and spatial pattern of the mean grain size agree with field data collected on micro-tidal shelves in the case that both bedload and suspended load sediment transport are accounted for, together with spatially non-uniform wave orbital motion. The model can not successfully explain the presence of large-scale ridges observed on La Barrosa inner shelf, because strong and complex behaving tidal currents occur in that area.

Nonlinear model simulations show that on the long term the height of the ridges evolves towards a finite, constant value, whilst their migration speed hardly changes during the evolution. In the saturated stage the ridges have asymmetrical profiles, with steep slopes on the downstream sides. The maximum variation in mean grain size also tends to a constant value and during the evolution the spatial lag between the patterns in the mean grain size and topography decreases. The processes that cause these changes are identified and explained. Model results can be obtained for transverse bottom slopes up to 50% of their observed values on micro-tidal shelves. Extrapolation of results to realistic values of the inner shelf slope yields, in case of Long Island shelf, a final height that agrees with observed ridge heights, but the modelled variation in mean grain size is small compared to field data. Finally, the response of ridges to large-scale interventions is considered. Experiments reveal that extraction of sand on the inner shelf causes a decrease of the sand volume stored in the surf zone.

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1. Introduction

Shoreface-connected sand ridges (hereafter abbreviated as SFR) are elongated bed-forms that are observed on the inner shelf of many coastal seas. Examples include the east US coast (Swift et al., 1985, see Fig. 1), Argentina, Germany and Holland (van de Meene and van Rijn, 2000). These ridges evolve during storms due to the joint action of large waves (stirring the sediment) and a longshore storm-driven flow (causing net sand transport). Geological analysis has revealed that their formation took place during the Holocene period (Swift et al., 1985; van de Meene et al., 1996). The ridges extend from the offshore end of the shoreface to the outer shelf, thereby forming an angle of 20°–35° with respect to the coastline. The orientation of the crests is upward: their seaward ends are shifted upstream with respect to their attachments to the shoreface. The alongshore spacing between successive crests ranges between 2 and 8 km and lengths of individual crests are between 10 and 25 km. Their height is between 1 and 6 m in water depths between 4 and 20 m. The North American ridges and some of the Argentine ridges are asymmetric, with the seaward side steeper than their landward side. The ridges slowly migrate in the direction of the storm-driven current with a characteristic celerity of 1–10 m yr⁻¹.
Furthermore, a persistent pattern of grain sorting is observed over the ridges: as shown in the right panel of Fig. 1, the finest sediments are observed on the seaward side of the ridges.

Gaining more knowledge about the conditions under which sfcr form and evolve is important for various reasons. One of them is that they may affect the stability of pipelines and oilrigs, due to lateral movements. Knowing the location of the ridges is also crucial for navigation. Moreover, sfcr are possible sources for sand mining and they are potential oil reservoirs because of their high preservation potential and good textural characteristics. Finally, the presence or absence of sand ridges seems to be linked to the overall stability of the coastal system (van de Meene, Boersma, and Terwindt, 1996).

As reviewed by Dyer and Huntley (1999) theories that explain the origin of sfcr fall into two categories. The first is that they are relict features produced during a period of sea level rise. The second is that they are the result of dynamic interactions between water motion and the mobile bed. It was first demonstrated by Trowbridge (1995) that sfcr can also form due to morphodynamic self-organisation. He studied the stability properties of a storm-driven longshore current, with a cross-shore gradient, on a sandy shelf bounded by a straight coast and with a transverse bottom slope. His model consists of the depth-averaged shallow water equations (ignoring Coriolis and bottom friction terms), supplemented with a bottom evolution equation and a formulation for the sediment transport (linear dependence on the net current). He also applied the quasi-steady assumption, i.e., the flow responds instantly to the changing bottom, since the characteristic hydrodynamic timescale (order of days) is much smaller than the morphodynamic timescale (order of centuries for the ridges).

This model is able to predict the growth and down-current migration of topographic features that have a similar shape as that of observed ridges in the field. Falqué et al. (1998) extended the model of Trowbridge (1995) by including Coriolis and bottom friction terms in the momentum equations, thereby allowing for production of vorticity. Moreover, they accounted for the effect of bed slopes on the magnitude and direction of sediment transport. They concluded that vorticity dynamics does not play a dominant role in the formation of the ridges and that the inclusion of bed slopes in the sediment transport results in the emergence of a preferred mode of which the longshore wavelength is in the range of values that are commonly observed in the field.

A limitation of the two models discussed above is that they only deal with storm-driven currents on micro-tidal shelves. Although this assumption is justifiable for e.g. the American shelf, it fails near the Dutch and German coasts where strong tidal currents (0.7–1.1 m s⁻¹, at the surface) occur. The influence of tides has an impact on the morphodynamic response: even during fair weather conditions bottom stresses are sufficiently strong to erode and transport sediment. Calvet et al. (2001b) demonstrated that the formation of either sfcr, tidal sand ridges or a combination of these features depends on the actual tidal conditions, the intensity of the storm-driven current and the fraction of time during which storms prevail. This model still suffered from two serious drawbacks. First, the ratio of the timescales related to the growth and migration of the ridges was not in accordance with field data. Second, the instability mechanism appeared to be very sensitive to the profile of the storm-driven current.

In a study by Calvet et al. (2001a) a new model for ridge formation on a micro-tidal shelf was analysed. This model accounts for both bedload and suspended load transport of sediment and also explicitly describes spatially non-uniform wave orbital motion. The mechanism is that sediment is stirred by the waves and subsequently transported by the storm-driven current. Suspended load transport is computed by solving the concentration equation in which a validated formulation for the sediment pick-up function is adopted. They concluded that the growth of the ridges in their model is mainly due to suspended load transport, whereas the migration is mainly controlled by bedload transport. Typical growth timescales of the most preferred mode are of order 1000 yr and migration velocities are several metres per year. The results of this model are hardly sensitive to the specific cross-shore profile of the storm-driven current.
current. The authors also explained their results in terms of physical mechanisms, extending arguments that were already introduced by Trowbridge (1995).

In Walgreen et al. (2002) the models of Calvete et al. (2001a,b) were combined in order to predict under what conditions sfc, tidal sand ridges or a combination of these features would form. Also, they studied the effect of adding higher tidal harmonics on the formation of bedforms. These so-called overides exist because tidal dynamics is nonlinear. In particular, self-interaction of a principal tide (e.g., the semi-diurnal lunar tide with a period $T_{M2}=12\,h\,25\,m$) results in the presence of a higher harmonic (in this case, the $M_4$ tide, with a period $T_{M4}=T_{M2}/2$). Walgreen et al. (2002) showed that, in case of the Belgian and Dutch shelf, the conditions are such that both types of bedforms can emerge. Overides can have a large effect on migration speeds of the ridges, but not on their growth rates.

Walgreen et al. (2003) studied the influence of grain sorting on the initial formation of sfc. The results were that, for increasing influence of sorting, the growth rate of the most preferred mode decreases and its spacing and migration speed slightly increases. The flow-sediment interaction also results in a mean grain size pattern which is characterised by finer (coarser) sediment on the downstream (upstream) side of the crests, as is also observed in the field.

The objectives of the present paper are twofold. The first is to study the effect of grain size sorting on the finite-amplitude evolution of sfc. The second objective is to show how sfc respond to large-scale interventions. Therefore, a new model will be derived, which combines ingredients of the models of Calvete et al. (2001a) and Walgreen et al. (2003). More extended versions of these models were presented in two recent studies (Lane and Restrepo, 2007; Vis-Star et al., 2007a). In particular both of them deal explicit feedbacks between waves and the ridges. Here, the latter study will be discussed in some detail, as it reveals that these feedbacks imply a new physical mechanism by which sfc form.

The model equations are presented in Section 2, followed in Section 3 by the method of analysis. Results will be shown in Section 4 and the response of ridges to interventions is studied in Section 5. We end with a discussion and the conclusions.

2. Model formulation

2.1. Geometry

Motivated by the results of previous studies (Trowbridge, 1995; Calvete et al., 2001a) we consider the possible formation of sfc due to internal feedbacks between a storm-driven flow and the mobile, sandy bed of the inner shelf. A highly idealised model is used, in which the shelf is schematised as a semi-infinite domain, bounded on the landward side by the transition from inner shelf to shoreface (Fig. 2). A Cartesian coordinate system is used with $x$, $y$, $z$-axes pointing in the cross-shore, alongshore and vertical direction, respectively. In the absence of any ridges the bathymetry is modelled as $z=-H(x)$, with $H=H_0+\beta x$ for $0\leq x<L_s$ and $H=H_s$ for $L_s\leq x<\infty$.

Thus, the undisturbed water depth $H$ increases linearly on the inner shelf (width $L_s$) and is constant on the outer shelf. The slope of the inner shelf is defined as $\beta=(H_s-H_0)/L_s$, where $H_0$ is the depth at the transition shoreface-inner shelf. Typical values of the parameters are $H_0=15\,m$, $H_s=20\,m$, $L_s=5\,km$ and $\beta=1\times10^{-3}$.

2.2. Hydrodynamics

The water motion is governed by the depth-averaged shallow water equations:

$$\frac{\partial \vec{v}}{\partial t} + (\vec{v} \cdot \nabla) \vec{v} + f\vec{e}_z \times \vec{v} = -g \nabla z + \frac{\vec{e}_b - \vec{e}_h}{\rho D}, \quad (1)$$

$$\frac{\partial D}{\partial t} + \nabla \cdot (D \vec{v}) = 0. \quad (2)$$

Here, $\vec{v}$ is the depth- and wave-averaged velocity, $f$ the Coriolis parameter, $\vec{e}_z$ is a unit vector in the vertical, $\vec{e}_b$ the wind shear stress vector, $\vec{e}_h$ the bottom stress vector, $g$ the acceleration due to gravity and $\rho$ the density of water. Furthermore, $t$ is time and $\nabla$ is the two-dimensional (horizontal) nabla vector. The local water depth is given by $D=z_s-z_b$, with $z_s$ the level of the free surface and $z_b$ the bottom level, both measured with respect to the undisturbed water level $z=0$. This model only represents conditions during storms, when the water motion is characterised by strong waves (amplitude of the local near-bottom wave orbital motion is $u_{\text{max}}$) and a storm-driven net current. The latter is forced by the wind stress and by a prescribed longshore pressure gradient. It is assumed that during fair weather conditions no sediment is transported, as the flow velocities are supposed to be below the critical velocity of erosion. Forcing of currents by wave-induced radiation stresses, Stokes drift, the cross-shore wind stress, density gradients and tides are neglected. Furthermore, waves are assumed to be monochromatic and nearly symmetrical. Some additional processes are considered in other studies (tides in Calvete et al. (2001b), radiation stresses and Stokes drift in Lane and Restrepo (2007), random waves in Vis-Star et al. (2007a)), but they are not crucial to model the dynamics of sfc.

During storms the amplitude of the wave orbital motion is much larger than the magnitude of the storm-driven current. If this is applied to the quadratic friction law, which relates the instantaneous bed shear stress to the instantaneous velocity

![Fig. 2. Sketch of the geometry of the model, representing the inner shelf (width $L_s$) and part of the outer shelf of a coastal sea. Further information is given in the text.](image-url)
field, it follows (in case that the angle between waves and currents is small)

\[ \tau_b = \rho r \bar{v}, \quad r = \frac{2}{\pi} \zeta_f u_w. \]  

(3)

Here, \( r \) is the bottom friction coefficient and \( \zeta_f \) the drag coefficient, which is assumed to be constant (typical value 0.002). Furthermore, \( u_w \) is the amplitude of the wave orbital motion at the edge of the wave bottom boundary layer. Here, the simple parameterisation of Calvete et al. (2001a) is adopted, i.e.

\[ u_w = u_{w0} \left( \frac{H_0}{H} \right)^{m/2}, \]  

(4)

with \( u_{w0} \) the amplitude of the near-bed wave-orbital motion at the transition from inner shelf to shoreface (\( x=0 \)) and \( m \) an exponent. This expression provides a good fit for results of a simple wave transformation model. Typical values of the parameters are \( u_{w0} \sim 1 \text{ m s}^{-1} \) and \( m \sim 1.6 \). Note that this formulation excludes feedbacks between ridges and waves, because \( u_w \) only depends on the reference depth \( H \) and not on the actual bed level \( z_p \). More sophisticated formulations for \( u_w \) derived from linear wave theory (see Mei et al., 2005, Chapter 3), are employed by Vis-Star et al. (2007a) and Lane and Restrepo (2007).

The boundary conditions imposed are that the cross-shore velocity component \( u \) vanishes at \( x=0 \), hence no net exchange of water occurs between inner shelf and shoreface. Furthermore, \( u \) vanishes at large distances from the coast.

2.3. Sediment characteristics and mass balance

The sediment considered in the model is a mixture of grains with two different sizes \( d_1, d_2 \). Typical values on the inner shelf are \( d_1 \sim 0.25 \text{ mm} \) and \( d_2 \sim 0.5 \text{ mm} \). The mean grain size is denoted by \( d_m \).

Evolution equations for the bed level and grain size distribution are derived from a one-layer model for bottom (Fig. 3). Following Seminara (1995) and references herein, the bed consists of a well-mixed active layer that overlies a substrate with a constant distribution in time.

Mass conservation for each size class yields

\[ (1 - p) \left( F_i \frac{\partial z_b}{\partial t} + L_a \frac{\partial F_i}{\partial t} \right) = -\nabla \cdot \left( \bar{q}_i \right), \quad (i = 1, 2). \]  

(5)

Here, \( p \) is the bed porosity, \( F_i \) the probability density function for grains of size \( d_i \), \( z_b = -H + h' \) the bed level, with \( H \) the reference level and \( h' \) the perturbation, and \( L_a \approx 0.02 \text{ m} \) is the thickness of the active layer. The instantaneous volumetric sediment transport per unit width during storms of grains with size \( d_i \) is indicated by \( \bar{q}_i \); in (5) it is averaged over a wave period. Using the constraint that \( F_1 + F_2 = 1 \), Eq. (5) can be rewritten as

\[ (1 - p) \frac{\partial z_b}{\partial t} = -\left[ \nabla \cdot \left( \bar{q}_1 \right) + \nabla \cdot \left( \bar{q}_2 \right) \right], \]  

(6a)

\[ (1 - p) L_a \frac{\partial F_i}{\partial t} = - \left[ (1 - F_1) \nabla \cdot \left( \bar{q}_1 \right) - F_1 \nabla \cdot \left( \bar{q}_2 \right) \right]. \]  

(6b)

2.4. Sediment transport

The volumetric transport of sediment per unit width, \( \bar{q}_i \), contains contributions due to bedload and suspended load processes, denoted as \( \bar{q}_{bi} \) and \( \bar{q}_{si} \), respectively:

\[ \bar{q}_i = \bar{q}_{bi} + \bar{q}_{si}. \]  

(7)

Field data analysed by Green et al. (1995) revealed that during storms large amounts of sand are transported as suspended load. Expressions for \( \bar{q}_{bi} \), \( \bar{q}_{si} \) are based on the formulations for a single grain size and corrected for availability of sediment in that size class and dynamic hiding effects. The latter refers to the fact that fine grains hide between the coarser grains, with the consequence that they do not experience the full stress of the water motion acting on the bed.

The formulations for bedload sediment transport are

\[ \bar{q}_{bi} = g_{bi} \bar{q}_b, \quad \bar{q}_b = \frac{3}{2} v_s u_w \left( \bar{v} - \bar{v}_h u_w \nabla h' \right), \quad g_{bi} = \left( d_i \right)^{c_b} \]  

(8)

Here, \( g_{bi} \) is the transport capacity function for sediment with size \( d_i \) that is transported as bedload, coefficient \( c_b \) indicates the strength of the hiding. This formulation is based on the concept of dynamic hiding, as discussed in Ribberink (1987), Seminara (1995) and references herein. It means that fine grains are less exposed to the bed shear stress than the coarse grains. Consequently, the value of the coefficient \( c_b \) in the expression for \( g_{bi} \) is positive (\( c_b \sim 0.75 \)). The parameterisations for a single grain size are based on arguments discussed by Ballard and Inman (1981) for the transport during storms. The first contribution to \( \bar{q}_b \) represents a stirring of the sediment by waves and the subsequent transport by the wave-averaged current. The second contribution to \( \bar{q}_b \) (proportional to coefficient \( \lambda_h \)) accounts for the preferred down-slope movement of sediment.
These equations are linearised about the basic state and the linear stability analysis is performed. The latter describes a steady storm-driven flow \( V(x) \) over a shelf, characterised by a topography \( H(x) \) and a distribution of fine and coarse sand \( F_1(x) \) and \( F_2(x) \), respectively, that only vary in the cross-shore (x) direction. Since the aim of this study is to analyse the dynamics of sfcr, it is convenient to split the model variables (velocity, free surface elevation, wave orbital velocity amplitude, fraction of fine grains and bottom depth) in a basic state contribution and a perturbation on this basic state. Thus, \( \Psi = \Psi_0 + \psi \), where \( \Psi = (u, v, z_s, u_w, F_1, z_h) \) and
\[
\Psi_0 = (0, V(x), U_w, s_0 + \zeta, F_1, -H(x)),
\]
\[
\psi = (u', v', \eta, u_w', f_1', H').
\]

The vectors \( \Psi_0 \) and \( \psi \) contain the information about the basic state and the perturbations, respectively. We assume that \( F_1 \) is independent of the cross-cross coordinate x. Note that in the present model the parameterisation (4) implies that \( u_w = 0 \).

First, the basic state is investigated. Substitution of \( \Psi = \Psi_0 \) in the longshore momentum balance (1) yields for the longshore flow
\[
V(x) = \frac{\tau_{xy}/\rho - g s_0 h}{r},
\]
with friction coefficient \( r \) defined in Eq. (3). The basic state velocity consists of a steady component, which is driven by a prescribed alongshore free surface pressure gradient \( s_0 \) and an alongshore wind stress \( \tau_{xy} \).

Next, a linear stability analysis is carried out, following the concepts that are discussed in the review paper of Dodd et al. (2003). Substitution of the solutions \( \Psi \) in the equations of motion and next assuming the amplitude of the perturbations to be small, the result is a linear system of differential equations. A detailed derivation of these terms is given in Walgreen et al. (2003). Here, we only review the basic principles.

The linear equations for the small perturbations allow solutions of which the amplitude can grow (or decay) exponentially in time. For example, the perturbation in the bed level reads
\[
h' = \Re \{ A_0 h(x) e^{(i\omega)t} \}
\]
and similar expressions for the other perturbations. Here, \( \Re \) denotes the real part of the solution, \( A_0 \) is an initial amplitude, \( k \) is the longshore wavenumber (a free parameter), \( h(x) \) describes the cross-shore structure of the perturbation and \( \Omega \) is the complex wave frequency. The latter can be split in a real and imaginary part:
\[
\Omega = \Omega_r + i \Omega_i,
\]
with \( \Omega_r = \omega \) the growth rate and \( \Omega_i \) the frequency. Clearly, if \( k = 0 \) the perturbation is alongshore uniform. In the case that \( k \neq 0 \) the solution (11) describes a wave that travels along the coast with phase speed \( c = -\Omega/k \).

It appears that \( \Omega \) are the eigenvalues of the linearised system and \( h(x) \) are the corresponding eigenfunctions. Hence, for any choice of wavenumber \( k \), a set of modes with different cross-shore structures, is found. If the growth rates \( \omega \) of all perturbations are negative, then they all decay and the basic state is stable.
Conversely, if there are perturbations with positive growth rates the basic state is unstable. In the latter case, the perturbation with the largest growth rate is called the initially most preferred mode or simply the linearly dominant mode.

3.2. Nonlinear analysis

The model variables that are split in a basic state contribution and a perturbation on the basic state, i.e. \( \Psi = \Psi_0 + \psi \), are substituted into the system of equations (momentum: (1), mass balance: (2), bottom evolution: (6a) and fraction fine grains: (6b)). The properties of the basic state solutions are used to derive the equations for the perturbed variables, which can be written symbolically as

\[
S \frac{\partial \psi}{\partial t} = \mathcal{L} \psi + \mathcal{N} (\psi).
\]

(13)

Here, \( \Psi = (u', v', \eta', u'''_w, f', h') \) and \( S \) is a matrix which contains the temporal information of the perturbations. Furthermore, \( \mathcal{L} \) is a matrix operator containing the linear contributions and \( \mathcal{N} (\psi) \) is a vector operator containing the nonlinear contributions.\(^1\)

Matrix \( S \) only has one nonzero element, due to the term \( \partial Z_j / \partial t \) in Eq. (6a). This is consistent with the rigid lid and quasi-steady assumptions used in deriving the model.

In this study a spectral method is applied to solve the system (13) of partial differential equations. The perturbations are expanded in known eigenmodes of the linear system, together with their \( k = 0 \) contributions, of which the amplitudes are to be determined. This approach is chosen because the linear solutions already show a high resemblance with the observed ridges. Here, it is assumed that \( u'''_w = 0 \). For the expansion of the nonlinear solutions in the linear eigenmodes, as described in Calvete et al. (2002), a truncated series of eigenmodes is selected. The solutions are computed in a domain of alongshore length \( L = 2\pi / K_0 \), with periodic boundary conditions in the longshore (y) direction.

The structure of the different modes will be indicated by \( (j, n_j) \), where \( j \) is the alongshore mode number and \( n_j \) the cross-shore mode number. Furthermore, \( (J, N_J) \) denotes the largest alongshore and cross-shore mode numbers considered in the expansion. The numerical values of the truncated numbers \( J \) and \( N_J \) should be chosen in such a way that the properties of the final solution do not change if more modes are included. If this is the case then the main physics are represented correctly by the selected modes. The nonlinear solutions of the equations of motion are thus written as

\[
u' (x, y, t) = u_0 (x, t) + \sum_{j=1}^{J} \sum_{n_j=1}^{N_J} \hat{u}_j (t) \hat{u} (x) e^{iK_j y} + c.c.
\]

(14)

and similar expressions for \( v', \eta', f' \) and \( h' \). Here, \( u_0 \) represents the perturbation with an alongshore-uniform structure, \( \hat{u}_j (t) \) is the amplitude of the cross-shore velocity component with mode number \( (j, n_j) \) and \( \hat{u}_j (x) \) is its known cross-shore structure. This expansion is inserted in the nonlinear system (13) and next projected on the adjoint linear eigenfunctions, resulting in equations for the amplitudes of the different modes. Details about the projection and numerical methods are discussed in Walgreen (2003) and Calvete and de Swart (2003).

4. Results

4.1. Parameter values: Long Island shelf (default case)

The default case is representative for the Long Island shelf (\(~40^\circ N\)) on the Atlantic coast of North America, thus the Coriolis parameter \( f \approx 1 \times 10^{-4} \text{ s}^{-1} \). The bottom profile for the inner shelf (width \( L_x = 5.5 \text{ km} \)) and outer shelf is shown in Fig. 2. Experiments are performed with a mild bottom slope in the basic state; \( \beta = 1.3 \times 10^{-4} \), which corresponds to an increase in depth over the width of the inner shelf from \( H_0 = 14.0 \text{ m} \) to \( H_x = 14.7 \text{ m} \), as indicated in Fig. 4. This slope is approximately 10% of its observed value in this area. It is chosen because solutions for larger values of \( \beta \) are more difficult to interpret, whereas their overall properties do not differ from those obtained for a milder slope. Observations (Niederodora and Swift, 1981; Lentz et al., 1999) show that typical values of the parameters that force the storm-driven current \( V (x) \) along the coast (see Eq. (10)) are

\[ \tau_0 = 0.25 \text{ N m}^{-2}, \quad s_0 \sim 2 \times 10^{-7}, \quad u_{w0} \sim 1 \text{ m s}^{-1}. \]

The resulting current amplitude at the coast is \( U \sim 0.4 \text{ m s}^{-1} \) and the cross-shore profile of \( V (x) \) is shown in Fig. 4 (right panel). In the basic state equal amounts of fine and coarse sediment are present: \( F_1 = F_2 = 0.5 \). The mean grain size is \( d_m = 0.35 \text{ mm} \) and for a standard deviation of \( \sigma_0 = 0.50 \) (default)
these values correspond to a fine grain size class with a diameter of $d_1=0.25$ mm and a coarse fraction with $d_2=0.50$ mm. A typical value for the advective part of the bedload transport in the model is $Q_a = \frac{1}{2} v_w u_{w0} H_0 U \sim 1 \times 10^{-3} \text{m}^2\text{s}^{-1}$.

The advective suspended load transport is $Q_s = v_w u_{w0} H_0 U \sim 6 \times 10^{-4} \text{m}^2\text{s}^{-1}$. Other parameter values are the typical magnitudes of the bed slope transports: $\frac{1}{2} v_b \lambda H_{w0} u_0 L_s \sim 3 \times 10^{-3} \text{m}^2\text{s}^{-1}$ for bedload and $\frac{1}{2} v_b \lambda H_0 u_{w0} u_0 H_0^2 / L_s \sim 2 \times 10^{-7} \text{m}^2\text{s}^{-1}$ for suspended load. The default case is defined by hiding in suspended and bedload sediment transport, with values of $c_s = -1.1$ and $c_b = 0.75$, respectively. This means a reduced suspended load transport of coarse grains (with respect to that in case of uniform sediment) and a reduced bedload transport of fine grains.

4.2. Linear analysis: default case

The growth rate curve of the linear perturbations evolving on the basic state of the model is shown in Fig. 5. The default values for the model parameters are used. The largest possible growth rate $\Omega_{LM}$ is found for a mode that has an alongshore wavenumber $k = K_M \sim 0.8 \text{ km}^{-1}$. This is called the (initially) preferred mode. The corresponding wavelength is $\lambda = 2\pi k^{-1} \sim 7.6 \text{ km}$, and its e-folding timescale for the growth is $\Omega_{LM}^{-1} \sim 2.3 \times 10^3 \text{ yr}$. The spatial structure of bottom and fraction of fine sand of the first and second cross-shore mode for longshore wavenumber $k = K_M$ are shown in Fig. 6. The greyscale is used to indicate the height of the bottom perturbations ($h'$), contour lines show the perturbation in the fraction of fine grains ($f'_1$). The first mode clearly shows a phase shift between the pattern of $h'$ and $f'_1$, whereas the shift is hardly present in the second mode. This is because the amplification of the preferred mode is mainly due to suspended load sand transport. For the other modes, which grow slower, the bedload transport is more important with respect to the suspended load transport. In Walgreen et al. (2003) it is discussed that the hiding in bedload favours a $90^\circ$ phase shift between the fraction of fine sand and the topography, while suspended load favours a $90^\circ$ phase shift.

4.3. Application to La Barrosa inner shelf

Runs with the linear model have also been performed with values of the parameters that are representative for the inner shelf of La Barrosa. This shelf is located on the south coast of Spain, west of the city of Cadiz. In the EU-project HUMOR this area was selected as a test area for all models that were developed within that project.

Bathymetric surveys have revealed that on this inner shelf a patch of six large-scale ridges are observed (Fig. 7). The average distance between successive crests, measured in the alongshore direction, is about 5 km and their typical height is 4 m. Although the crests do not show a clear up-current orientation, it is interesting to investigate whether the present model is able to simulate the formation of ridges that to some extent resemble the ridges observed on this shelf.

Values of the parameters were determined by analysing available field data about offshore wave characteristics (from buoys), the wind climatology and sediment samples. From this it was found that the depth at the transition inner shelf-shoreface is 10 m, the width of the inner shelf is 4.2 km and it has a slope of $1.35 \times 10^{-3}$. Furthermore storms (offshore wave heights larger than 2.5 m) occur 5% of the time and the amplitude of near-
bottom wave orbital motion is 1 ms$^{-1}$. Furthermore the average wind stress during storms is 0.25 Nm$^{-2}$ and sediment characteristics are comparable to those at Long Island inner shelf. The model predicts the formation of ridges with a wavelength of 3 km, which is in the same order as that of the observed ridges. The corresponding timescale of growth is of order 100 yr. However, the spatial pattern of the most preferred mode differs substantially from that of the observed ridges. One reason for this discrepancy might be that La Barrosa inner shelf is characterised by meso-tidal conditions; tidal current amplitudes are up to 1 ms$^{-1}$ (Lobo et al., 2000). Therefore, additional runs were performed with a version of the model that also accounts for tidal currents, in the way described by Walgreen et al. (2002). However, the results were not in closer agreement with the data. It should be remarked here that in the model it was not possible to describe the complex behaviour of tidal currents as observed in this area. The tidal dynamics is so complicated because of the proximity of the narrow Gibraltar Strait to the east, which in this area widens towards the Atlantic Ocean on the west.

The conclusion is that the model can not satisfactorily explain the presence of the ridges on La Barrosa inner shelf by morphodynamic self-organisation. As discussed in Dyer and Huntley (1999) and references herein, they are probably generated by other mechanisms, e.g. by migration of the ebb-tidal deltas of rivers in combination with shoreline retreat related to sea level rise.

4.4. Effect of feedbacks between waves and ridges

So far, results were presented for parameterisation (4) of the near-bed wave orbital velocity amplitude $u_{w0}$, which obeys $u_{w0}^*$=0. In Vis-Star et al. (2007a) and Lane and Restrepo (2007) wave properties are calculated by employing physical equations, rather than using a parameterisation, and consequently, $u_{w0}^*$ ≠ 0. Here, results are shown which were obtained with the model of Vis-Star et al. (2007a), modified such that it accounts for grain sorting and it assumes monochromatic waves. Values of parameters were chosen identical to those in Section 4.1, except that the bed slope parameter $\lambda_{\delta}$ was adjusted such that a maximum growth rate was obtained for a realistic longshore wavenumber $k$. In the left panel of Fig. 8 the growth rate curve is shown for the new model. Clearly, growth rates are significantly larger than those in the

Fig. 7. Left: bathymetric map of the inner shelf of La Barrosa (south coast of Spain), indicating the presence of large-scale ridges. The x-axis points to the south. Right: cross-shore bottom profile of La Barrosa inner shelf in the area indicated by the square box in the left subplot.

Fig. 8. Left: growth rate versus longshore wavenumber calculated with a model that explicitly accounts for feedbacks between sfe and waves. Right: contour plot of the corresponding most preferred mode. The vectors denote the perturbations in the wavevector of the waves.
that the level minim variations through this initial amplification. The dashed line indicates the linear solution for the total height. Default case. The long time scale is discussed in the text.

Fig. 5. The time step used in the time integration is 10 yr. This turned out to be sufficient: adding more modes or decreasing the time step did not affect the behaviour of the solutions.

Fig. 9 (left panel) shows for the default case, starting from a state in which all bottom modes have very small amplitudes, the temporal evolution of the maximum and minimum heights (with respect to the level of the undisturbed bottom), as well as the evolution of the total height of the ridges. The panel on the right shows the same quantities, but for the fraction of fine sediment. In addition, Fig. 10 shows the evolution of amplitudes, of both bed level $h'$ and fraction $f_j$ of fine sediment, of the five modes that have the largest values at the final time of the simulation ($t \sim 25 \times 10^3$ yr).

The results shown in Fig. 9 indicate that, in the initial stage of evolution, the height of the bedforms and the variation in the fraction of fine grains increase exponentially in time. After some time (in the present case after about $15 \times 10^3$ yr) amplitudes are large enough for nonlinear processes to become important. From then on, the evolution of the system starts to deviate from that predicted by the linear stability analysis (see the dashed lines in the two panels of Fig. 9). In Fig. 10 it can be seen that nonlinear interactions between different modes are such that they cause excitation of modes which are linearly damped, such as the (2,2), (3,3) and (3,2) mode. This causes a reduction in growth of the height of both the bedforms and the size fractions and after about $20 \times 10^3$ yr they even become constant. The total height of the bottom perturbations in this saturated stage is approximately 0.72 m, which is slightly larger than the total change in depth.

4.5. Nonlinear analysis: Long Island shelf

We now discuss the results of runs performed with the nonlinear model which uses parameterisation (4) for the near-bed wave orbital velocity amplitude. In this case the smallest longshore wavenumber considered in the spectral expansions (14) is $K_0=K_M$, i.e., that of the initially most preferred mode. The number of modes used in the nonlinear evolution was truncated at $J=25$ (longshore) and $N_J=16$ (cross-shore) for this value of $\beta$. These modes are indicated by the filled dots in Fig. 10. Time evolution of the amplitude of the five modes which have the largest values at the final time $t=25 \times 10^3$ yr. The left and right panels show amplitudes of bottom modes and fraction of fine grains, respectively. Modes are denoted as $(j, n)$, where $j, n$ are the longshore and cross-shore mode number. Default case.
over the inner shelf in the basic state. The total variation in the fraction of fine and coarse grains is rather small: less than 1% of their characteristic values $F_1=F_2=0.5$ in the basic state. Fig. 10 demonstrates that the saturated stage is still dominated by the (1,1) mode, which is also the initially preferred mode.

The spatial patterns of the perturbations in both the bottom and fraction of fine sediment at different times during the evolution are shown in Fig. 11. The different grey values correspond to different bed levels (light: crests, dark: troughs); contour lines represent values of $F_1$. The bedforms which develop from the initial state (all bottom modes have small amplitudes) soon resemble sfr. They migrate in the direction of the mean storm-driven flow (negative $y$-direction). Initially, a spatial phase shift of approximately $90^\circ$ between bottom perturbation and fraction of fine sediment is present. This is consistent with the results of a linear stability analysis, as presented in Fig. 6. During the nonlinear stage of the evolution the spatial patterns of the bottom and fraction of fine sediment become asymmetrical. Most noticeable are the mild gradients on the landward (and upstream) sides and the steep gradients on the seaward (and downstream) sides. The ratio of the seaward slope and landward slope, once a constant amplitude of the ridges is reached, is 6:1. In the saturated stage the migration speed of the ridges hardly deviates from that predicted by the linear stability analysis.

A new result is that the distance between location of ridge crest and the location where the finest sediment occurs decreases during the nonlinear evolution stage. In the final state this distance is so small that the finest sediment is almost located at the crests; likewise the coarsest sediment is found almost in the troughs. This tendency is rather difficult to relate to field data, as in general the finest sediments are observed at measurable distances downstream of the ridges. We will return to this aspect in the discussion. Also, note that the time to reach the final saturated stage is quite long. However, it should be realised that in the simulations a rather small value of the transverse bottom slope $\beta$ has been used. As will appear later on (Section 4.7), the saturation timescale becomes considerably smaller if $\beta$ is increased.

4.6. Process analysis

To understand the physics responsible for the observed nonlinear evolution of the ridges and the corresponding sorting of sediment a detailed process analysis was carried out by Walgreen (2003). Here, we summarise the main results.\(^2\) The analysis was done by studying the convergence of sediment transports, because Eqs. (6a) and (6b) shows that they induce changes in bed level and size fractions. A distinction was made between transports of fine and coarse grains, transports related to suspended load and bedload processes and transports related to advective processes and bed slope processes. The latter are proportional to the coefficients $\lambda_5$ and $\lambda_6$ in Eqs. (8) and (9).

First, we focus on the temporal behaviour of the bedforms. Eq. (6a) shows that it is not necessary to distinguish between fine and coarse sediment fractions because the different contributions are added. It turns out that, as long as the system has not reached its final saturated state, the convergence of the suspended load sediment transport near the crests is much larger than that of the bedload transport. Conversely, in areas between crests and troughs the convergence of the bedload transport is dominant. Thus, suspended load transport determines the growth, while bedload transport mainly determines the migration of the ridges in the initial phase of the development of the ridges.

\(^2\) A copy of the thesis of Walgreen can be obtained from the first author.
Information about the growth and saturation mechanism can be extracted from Fig. 12, which shows the convergence of the advective and that of the bed slope suspended load transport above the crests as a function of the total height of the ridges (left panel) and the ratio of these two terms (right panel). It appears that the advective transport causes growth, whilst the bed slope transport causes decay.

During the initial stage the advective part dominates over the bed slope part, but the magnitude of the latter rapidly increases with respect to that of the former, until ultimately a balance is established with a negligible total convergence of sediment transport on the crest. This result is consistent with the earlier conclusion of Calvete and de Swart (2003) that, within the present model context, the bed slope term is important to reach the saturation.

To unravel the observed changes in the spatial lag between the ridges and the locations of extrema in the mean grain size it is necessary to distinguish between erosion and deposition of fine and coarse grains, as can be traced back from Eq. (6b). The investigation of the nonlinear grain sorting process, using the full spectral model, is a difficult task because of the many nonlinear terms involved.

However, test experiments revealed that not all nonlinear terms are crucial for the evolution of the ridges and the grain sorting. It was found that the saturation still occurs if the nonlinear contributions in the equations for the bottom and the fraction of fine grains are excluded in the computations.

Upon ignoring the nonlinear terms in the equations describing the dynamics of the fraction of fine sediment, Eq. (6b) can be approximated by

\[
-T_{b3} \nabla \cdot \overline{q^b} - T_{s3} \nabla \cdot \overline{q^s} \approx |T_{a} + T_{b} T_{s3} | q_{o} \frac{\partial f_1}{\partial y}.
\]

Here, \(T_{b} \) and \(T_{s} \) are known coefficients (Walgreen et al., 2003) and it is used that in the basic state the suspended load sediment transport \(q_{o,0} = (0, q_{o,0}) \) dominates over the bedload transport. For the default parameter setting \( T_{b3} F_1 F_2 (G_{o1} - G_{o2}) < 0 \), which is a consequence of the dynamic hiding effects in bedload transport. The coefficients \( G_{o1}, G_{o2} \) denote the values of the sediment transport capacity functions \( G_{o1} \) and \( G_{o2} \) which are defined in Eq. (8). Furthermore, \( T_{b3} > 0 \) and the contribution between the brackets is positive. The variables that change in time are \( \overline{q^b} \) and \( \overline{q^s} \), which are the perturbed sediment transport for bedload and suspended load in case of uniform sediment, and \( f_1 \). Thus, the spatial pattern of \( f_1 \) is determined by the patterns of convergence of bedload and suspended load transport. Analysis of the model revealed that in the initial, linear stage of ridge growth, the bedload contribution on the left-hand side of Eq. (15) is smaller than the suspended load contribution. However, during the nonlinear evolution stage the bedload term becomes dominant over the suspended load term. This causes the maximum in fine sediment, which is initially located downstream of the ridges, to shift in the course of time towards the crests.

4.7 Nonlinear analysis: sensitivity to parameter values and wave formulation

Additional experiments with this nonlinear model for \( f_1 \) were carried out for different values of the transverse bottom slope; all other parameters had their default values. The results shown in Fig. 13 indicate that the final height of the ridges and the variation in the mean grain size increase for a higher transverse bottom slope of the inner shelf. The former trend is the same as was already found in the model for uniform sediment. Saturation occurs faster for higher values of \( \beta \). The bottom patterns clearly show an increase in asymmetry in the final state of the evolution of the ridges for larger slopes of the inner shelf (results not shown). The solutions become unbounded at some time during the saturation process for values of \( \beta \) larger than \( 0.55 \times 10^{-3} \), which corresponds to approximately 50% of the value of the observed transverse bottom slope of the Long Island inner shelf.

Sensitivity experiments were also carried out by varying the standard deviation \( \sigma_0 \) of the sediment mixture. Compared to the case of uniform sediment the final height of the ridges hardly changes, while increasing \( \sigma_0 \) causes an increase in the amplitude of \( f_1 \). The saturation time also increases for a larger difference between the two grain sizes. Another series of experiments shows that the properties of the solution in the saturated state depend on the value of coefficient \( c_b \) in the hiding function for bedload: increasing \( c_b \) enhances the variation in the mean grain size.
Finally, we present results of a nonlinear model for sfcr that has recently been developed by Vis-Star et al. (2007b). It uses physical equations to calculate the amplitude of the near-bed wave orbital motion, but it ignores wave-topography feedbacks ($u''_b=0$) and it is designed for uniform grains. In the left panel of Fig. 14 the bed level is shown at different times at an alongshore transect which is located halfway the inner shelf. Parameter values in this experiments have been chosen such that external conditions similar to those of the default case discussed in the previous section. On the outer shelf waves have a height of 1.4 m, a period of 11 s and their wavevector makes an angle of 20° with the x-axis (normal to the coast). Also this model shows the initial formation of sfcr which, in the course of time, increase their height and become more asymmetrical. After about 1500 yr of continuous storms their height is almost constant. With this model the dependence of the saturation height and saturation timescale on the wave characteristics can be investigated. In the right panel of Fig. 14 the final height of the ridges is plotted as a function of wave height at the outer shelf for different values of the transverse bottom slope. The ridge height decreases with increasing wave height and increases if a larger bottom slope is considered. The dependence of ridge height on wave height is due to the fact that bedslope sediment transport in the model, which is responsible for the saturation behaviour of the ridges, increases more rapidly with wave height than the advective sediment transport (which causes the growth of the ridges).

4.8. Response of ridges to large-scale interventions

Human interventions, like the construction of coastal defense structures, mining of sand and dredging of navigation channels, can have large consequences for the stability of coasts (sand may be transported away from the beach) and therefore require profound studies. de Swart and Calvete (2003) investigated the response of finite-amplitude sfcr to interventions using a nonlinear model for uniform sediment. They proceeded by first constructing a field of sfcr with finite heights. Next, an intervention was imposed and the subsequent response of the model was analysed. Three types of interventions were studied: extraction of sand from a ridge, nourishment of sand on the inner shelf and the dredging of a navigation channel.

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Fig. 13. Total height of bedforms in the saturated state, total variation in the fraction of fine grains in the saturated stage (in the figure called: final $f_i$) and time needed for saturation as a function of the slope of the inner shelf in the basic state. The arrow indicates the default value of $\beta$, for which case the evolution in time towards the saturated stage is shown in Fig. 9.

Fig. 14. Results obtained with the nonlinear model for sfcr of Vis-Star et al. (2007b). Left panel: Bed profile at different times at an alongshore transect which is located at $x=L/2$, where $L=5.5$ km is the width of the inner shelf. Right panel: Final (saturation) height of the ridges versus wave height at the outer shelf, for different values of the transverse bottom slope $\beta$. 
Here, we discuss the case of a sand extraction of $2 \times 10^6$ m$^3$ from the inner shelf, due to dredging of a navigation channel (2 km wide, constant depth). The parameter values used are representative for Long Island shelf. Uniform sediment was assumed and the value of the transverse bottom slope ($\beta = 1 \times 10^{-4}$) was about 10% of the observed slope. This was necessary to model saturation behaviour of the ridges without encountering numerical problems. Consequently, timescales are quite long, but the results of Fig. 13 suggest that in case of a realistic $\beta$ the same qualitative results may be expected, but timescales are a factor 10 smaller. Furthermore, the smallest longshore wavenumber included in the spectral expansions (14) was now $K_0 = K_M/8$. This means that, unlike in the previous section, seven so-called subharmonics of the initially most preferred mode were taken into account. This was done in order to be able to impose an intervention on an isolated ridge in the domain, as the length of the domain is now large with respect to the longshore wavelength of the ridges.

The situation immediately after dredging is shown in the upper left panel in Fig. 15. Note that only five ridges are present in the domain, whereas seven subharmonics are included. This means that in this case the initially most preferred mode is not the dominant mode in the saturated state. The mechanism responsible for this lengthening of the ridges is discussed in Calvete and de Swart (2003). The situation at 1000 yr after the intervention is shown in the top right panel of Fig. 15. As can be seen the ridge recovers on a timescale of centuries, so apparently the system adjusts to its original state. Since sand was withdrawn from the inner shelf there must be a net import of sand from the adjacent areas. The lower left panel of Fig. 15 indeed shows that the sand volume of the inner shelf restores to its original equilibrium value, but the evolution takes place on a much longer timescale than that on which the ridge field recovers. The lower right panel of Fig. 15 shows that sand is transported from both the nearshore area and outer shelf to the inner shelf. Thus, in these experiments the presence of the so-called $k=0$ perturbation, having an alongshore-uniform structure, turns out to be crucial. The reason is that only this perturbation can provide for net changes of the sand volume stored in the inner shelf.

This experiment suggests that extraction of sand from the inner shelf has negative consequence for the beach, because the volume of sand in the nearshore zone will decrease. The same conclusion was found in case of extracting sand from sfcr. Remarkably enough, dumping of sand on the inner shelf does result in only a weak transfer of sand to the nearshore zone: most of the surplus will be transported to the outer shelf. The reason for this behaviour appears to be due to the structure of the $k=0$ perturbation, which causes preferred seaward transport of sand.

5. Discussion and conclusions

The results discussed in this paper demonstrate that the long-term evolution of sfcr and the variations in the mean grain size over these ridges can be simulated and analysed with a nonlinear process-based model. The latter assumes two grain size classes, includes dynamic hiding effects in both suspended and bedload sediment transport and is based on the one-layer concept for the bottom evolution.

It was shown that, starting from an initial state without bedforms, a pattern of sfcr with finite amplitudes develops due to morphodynamic self-organisation. Initially, the height of these ridges increases exponentially, but nonlinear effects cause a reduction of the growth, such that height tends to a constant value. At the same time, the maximum variation in the mean grain size evolves towards a constant value. Saturation is reached as a result of a balance between the deposition related to the advective part of the suspended sediment transport and erosion related to the bed slope part over the ridges. Identical conclusions were drawn with regard to the saturation behaviour of other types of bedforms in coastal seas, such as tidal sand ridges (Roos et al., 2004), estuarine bars (Schramkowski et al.,...
and bars in the surf zone (Garnier et al., 2006). The use of multiple grain sizes does not significantly change the final height and shape of the bottom topography that were obtained by Calvete et al. (2002) with a model based on a single grain size. The shape of the ridges changes from symmetrical in the linear stage towards asymmetrical in the nonlinear stage. The asymmetry ratio of the seaward over the landward slope for the default case is $\sim 6:1$ after reaching the equilibrium amplitude, whereas the maximum side slope of the ridges is around $0.05^\circ$.

The model was also applied as a tool to study the impact of large-scale interventions on the inner shelf (sand mining, dredging operations) on sfc and the stability of the coast. It was found that the response of the system to interventions is an adjustment back to its original equilibrium state. In the case that the intervention concerns extraction of sand from the inner shelf a net import of sand from the adjacent shoreface and outer shelf takes place. This result suggests that mining sand may have negative consequences for the stability of the nearshore zone, because its sand volume will decrease.

For the default case studied in this paper a transverse bottom slope $\beta$ of approximately 10% of its observed value in the field (Long Island inner shelf) was chosen. This was done because solutions for larger values of $\beta$ were more complicated to analyse, whilst their overall characteristics were identical to those obtained for the default value of the bottom slope. Numerically stable solutions of the model could be obtained until a slope of approximately 50% of its observed value. For larger values of $\beta$ solutions become unbounded at some time during the evolution.

To allow a comparison between model outcome and field data, the results for the final height of the ridges, maximum variation in mean grain size and saturation timescale as functions of $\beta$ (see Fig. 13) were extrapolated to realistic values of the slope of Long Island inner shelf. This yields a final height of $\sim 6$ m and a saturation timescale of $\sim 1 \times 10^3$ yr. These values agree rather well with those obtained from field data.

A shortcoming of the model is that the modelled maximum variation in the fraction of fine and coarse sediment over the ridges are much smaller than those measured in the field. Extrapolating towards realistic slopes still indicates variations in the fraction of fines of only 0.004 ($\sim 1\%$ change on the initial value of 50% fine grains in the basic state). For a diameter of the fine and coarse grains of 250 $\mu$m and 500 $\mu$m, respectively, this corresponds to a total variation in the mean grain size of approximately 2 $\mu$m. Observations suggest that the variations in the mean grain size are in the range of 200 $\mu$m.

Another point of discussion concerns the result that the distance between the location of ridge crest and the location where the finest sediment occurs decreases during the nonlinear evolution stage. The linear analysis showed a spatial shift between the variations in the mean grain size and the ridge topography, with the maximum fraction of fine grains located just seaward (downcurrent) of the crest. This phase shift is also observed in the field (cf. right panel of Fig. 1). The spatial shift changes if the ridges obtain an asymmetrical profile: in the final state this distance is so small that the finest sediment is located almost at the crests. This tendency is rather difficult to relate to field data, as most fine sediments are observed at measurable distances downstream of the ridges.

The results mentioned above may indicate that not all physical processes are yet incorporated correctly in the model. Recent studies indicate that feedbacks between bedforms and waves have a profound effect on the initial formation of sfc. So far, their influence on the long-term evolution of the ridges remains to be investigated. It should also be realised that the assumptions leading to the equation describing the evolution of the grain size fraction might not be valid for ridges with a significant height. The validity of the one-layer concept, which assumes a thin transport layer that is uniformly mixed over its depth and negligible interaction with the underlying substrate, might not be valid for the long timescales that are related to the formation of large-scale bedforms. This could be in principle overcome by using, for example, a two-layer model (cf. Ribberink, 1987) that accounts for vertical sorting. However, a major drawback of this approach is the large number of additional parameters which are poorly known.

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