

1 UNDERSTANDING COASTAL  
2 MORPHODYNAMIC PATTERNS FROM  
3 DEPTH-AVERAGED SEDIMENT  
4 CONCENTRATION

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6 This review highlights the important role of the depth-averaged sediment concen-  
7 tration (DASC) to understand the formation of a number of coastal morphodynamic  
8 features that have an alongshore rhythmic pattern: beach cusps, surf zone transverse  
9 and crescentic bars, and shoreface-connected sand ridges. We present a formulation  
10 and methodology, based on the knowledge of the DASC (which equals the sediment  
11 load divided by the water depth), that has been successfully used to understand the  
12 characteristics of these features. These sand bodies, relevant for coastal engineering  
13 and other disciplines, are located in different parts of the coastal zone and are char-  
14 acterized by different spatial and temporal scales, but the same technique can be  
15 used to understand them. Since the sand bodies occur in the presence of depth-averaged  
16 currents, the sediment transport approximately equals a sediment load times the cur-  
17 rent. Moreover, it is assumed that waves essentially mobilize the sediment and the  
18 current increases this mobilization and advects the sediment. In such conditions, know-  
19 ing the spatial distribution of the DASC and the depth-averaged currents induced  
20 by the forcing (waves, wind, and pressure gradients) over the patterns allows infer-  
21 ring the convergence/divergence of sediment transport. Deposition (erosion) occurs  
22 where the current flows from areas of high to low (low to high) values of DASC. The  
23 formulation and methodology are especially useful to understand the positive feed-  
24 back mechanisms between flow and morphology leading to the formation of those  
25 morphological features, but the physical mechanisms for their migration, their finite-  
26 amplitude behavior and their decay can also be explored.

## 1. INTRODUCTION

27 Coastal zones are highly valued worldwide for their natural beauty, the recreational  
28 opportunities they offer and the economic benefits that result from tourism, shipping, and  
29 fishing industries. As a result, more than half the world's population (and the percentage  
30 is growing) has settled along this narrow strip of the world's surface [Komar, 1998] and its  
31 preservation has turned out to be important for social, economic and ecological reasons.  
32 Sandy coasts, which are about 25% of the coasts on a global scale [Short, 1999], are highly  
33 dynamic, and increasing our knowledge of such complex systems is necessary to build more  
34 reliable engineering tools. Field data collected in the *swash and surf zones* and on the  
35 continental shelf of sandy coasts often reveal the presence of undulations in the sandy bed  
36 and the shoreline (hereafter referred to as *morphodynamic patterns*), indicating that they  
37 are an integral part of the coastal system. (Italicized terms are defined in the glossary,  
38 after the main text.) Many of these morphodynamic patterns show a remarkable spatial  
39 periodicity along the shore (Figure 1). Understanding the dynamics of these alongshore  
40 rhythmic patterns is important to increase our general knowledge about coastal processes  
41 and, thereby, our capacity to predict the short/long-term evolution (erosion/accretion) of  
42 the coastal system.

43 Crescentic bars (also called *rip-channel systems*, Figure 1a) are well known examples  
44 of alongshore rhythmic morphologic patterns that commonly occur in the surf zone [van  
45 Enckevort *et al.*, 2004, and references therein]. A crescentic bar consists of an alongshore  
46 sequence of shallower and deeper sections alternating shoreward and seaward (respec-  
47 tively) of a line parallel to the shore in such a way that the bar shape is undulating in  
48 plan-view. In some cases the undulation is quite subtle, the bar being almost straight,

but occasionally it features pronounced crescent moons with the horns pointing shoreward and the bays (deeps) located seaward. The deeper sections are called *rip channels* because strong seaward directed currents called *rip currents* [Dalrymple et al., 2011] are concentrated there. Patches of transverse bars are other distinct morphologic features observed in the surf zone [Gelfenbaum and Brooks, 2003; Wright and Short, 1984; Ribas and Kroon, 2007; Pellón et al., 2014, and references therein] (Figures 1b,c). They consist of several sand bars that extend perpendicularly to the coast or with an oblique orientation and the alongshore distance between bars can be remarkably constant. They are typically attached to the shoreline but they have been occasionally observed attached to a *shore-parallel bar*. Patches of shoreface-connected sand ridges are examples of larger scale features that occur on the *inner shelf*. They consist of several elongated sandy bodies of a few kilometers, oriented at an angle with respect to the shoreline, and separated an approximately constant alongshore distance [Dyer and Huntley, 1999, and references therein]. Beach cusps are well known morphologic features with an alongshore rhythmicity that occur at the swash zone (Figure 1d). Beach cusps can be described as lunate embayments (lowered areas of beach level) separated by relatively narrow shoals or horns (raised areas of beach level) [Coco et al., 1999, and references therein]. These four features are located in different parts of the coastal zone (i.e., at different water depths), and are characterized by different spatial and temporal scales, as shown in Table 1 and Figure 2.

The relevance of these alongshore rhythmic patterns for coastal engineering is being increasingly recognized for several reasons. Firstly, studying their dynamics allows identification of important physical mechanisms that control coastal evolution. In particular, it increases our understanding of the effective sediment transport in areas of the coastal zone

72 where there is still a significant lack of knowledge on such important process (e.g., swash  
73 zone and inner surf zone) [*Soulsby*, 1997]. Secondly, these alongshore rhythmic morpho-  
74 dynamic patterns have a direct impact on the shoreline by creating areas of erosion and  
75 deposition [*Komar*, 1998; *MacMahan et al.*, 2006]. The presence of beach cusps and trans-  
76 verse bars implies an erosion of the shoreline in their embayments, and crescentic bars  
77 and shoreface-connected ridges affect wave refraction and breaking, creating patterns in  
78 the nearshore flow circulation that can cause erosional hot spots [*Sonu*, 1973; *Wright and*  
79 *Short*, 1984; *Benedet et al.*, 2007]. Furthermore, beach cusps are notable morphodynamic  
80 features because they occur in the swash zone, a region whose dynamics are not yet well  
81 understood but which forms the physical interface between the land and the sea, where  
82 the effects of erosion/deposition are most clearly seen. In the surf zone, sandy bars are a  
83 natural protection of the beach: waves dissipate part of their energy on the bars and the  
84 bars can also provide sand to the beach as they can migrate onshore. Furthermore, the  
85 alongshore migration of surf zone bars can cause (additional) erosion/deposition patterns  
86 near coastal structures that are generally not considered in engineering projects. It is also  
87 important to understand the horizontal circulation induced by surf zone bars since the as-  
88 sociated currents enhance transport and exchange of pollutant or floating matter [*Castelle*  
89 *and Coco*, 2013]. Further, although surfers take advantage of rip currents occurring in  
90 between sand bars to move offshore, such currents are dangerous for swimmers, being  
91 one of the most lethal natural hazards worldwide [*Dalrymple et al.*, 2011]. On the con-  
92 tinental shelf, shoreface-connected ridges are of interest to coastal engineering as sources  
93 for extraction of sand (e.g., for beach nourishment or for the construction industry) and  
94 because they are located in areas where wind turbine fields are present or planned [*van*

95 *de Meene and van Rijn*, 2000]. Also, due to their alongshore migration, they can produce  
96 the infilling of navigation channels and affect pipeline burial. Shoreface-connected ridges  
97 also have an interest for biologists since they provide favorable conditions for benthic life  
98 and fish [*Slacum et al.*, 2010], in particular on their sheltered landward side (where grain  
99 size is smaller). From a geologic point of view, all these morphodynamic features are of  
100 interest because they lead to depositional rhythmic patterns that can be detected in the  
101 stratigraphy, and thus provide insight into the long-term evolution of the coast. In par-  
102 ticular, shoreface-connected ridges, having evolved over thousands of years, can be traced  
103 in and dated from cores [*McBride and Moslow*, 1991].

104 Rhythmic morphodynamic patterns are the result of waves and currents that erode  
105 and transport sediment by exerting *shear stresses* at the sandy sea bed. The conver-  
106 gence/divergence of sediment transport produces bed level changes, which feedback into  
107 the wave and current fields. Rhythmic morphologic patterns grow very often due to *feed-*  
108 *back mechanisms* (the so-called self-organization theory), without a corresponding spatial  
109 pattern in the hydrodynamic forcing (the latter being essential in the so-called template  
110 theories) [*Coco and Murray*, 2007]. The key message of the present contribution is that,  
111 despite the fact that beach cusps, surf zone bars and shoreface-connected ridges have dif-  
112 ferent scales and occur in different areas of the coastal zone, they nevertheless have one  
113 important aspect in common: their formation, migration and long-term evolution can be  
114 explained by the *advection* of the *depth-averaged sediment concentration (DASC)* by the  
115 *depth-averaged current*. As each feature is associated with different types of water motion,  
116 each has its own typical spatial distribution of sediment concentration. The aim of this  
117 contribution is to highlight the important role of the spatial distribution of the DASC in

118 the development of these alongshore rhythmic coastal morphodynamic patterns. Previous  
119 studies focusing on these distinct features will be reviewed and linked, thereby showing  
120 that, by using a specific formulation of the equations, the convergence/divergence of sed-  
121 iment transport can be understood in a remarkably simple way, from the joint action  
122 of the gradients in the DASC and the current perturbations produced by the evolving  
123 morphologic pattern. This formulation is a powerful tool to get insight into the underly-  
124 ing feedback mechanisms that explain why features with a specific spatial pattern (e.g.,  
125 *up-current orientation* of shoreface-connected ridges and transverse bars, see Figure 2)  
126 grow and migrate [e.g., *Falqués et al.*, 2000; *Calvete et al.*, 2001; *Caballeria et al.*, 2002;  
127 *Ribas et al.*, 2003; *Calvete et al.*, 2005; *Dodd et al.*, 2008; *Ribas et al.*, 2012]. The physical  
128 mechanisms for the saturation of the growth of the features or for their decay can also  
129 be explored with this technique [e.g., *Garnier et al.*, 2006, 2008; *Vis-Star et al.*, 2008;  
130 *Garnier et al.*, 2013].

131 The first step is to present and discuss the formulation and methodology, based on the  
132 DASC, which have been successfully used to understand and model the characteristics  
133 of coastal patterns. In existing publications, different versions of this formulation were  
134 presented, corresponding to the specific morphodynamic features being studied. Here we  
135 will present the overall theory, the underlying hypotheses and the physical interpretation  
136 of the equations. The model framework and most important physical laws and processes  
137 governing the dynamics of the currents, the waves and the sediment at the coast are  
138 presented in section 2. Since this contribution focuses on the morphologic evolution,  
139 some technical details of the hydrodynamic processes will be given in appendices. The  
140 formulation of the equations, with the DASC being the main focus, is derived in section 3,

141 and the methodology that allows understanding rhythmic pattern formation is described.  
142 The second step is to review the key studies that apply this formulation to the development  
143 of the four specific morphologic patterns mentioned above (Figure 2): crescentic bars  
144 (section 4), transverse bars (section 5), shoreface-connected sand ridges (section 6) and  
145 beach cusps (section 7). A physical and transparent explanation, based on the DASC, will  
146 be provided of why alongshore rhythmic patterns of a certain shape grow and sometimes  
147 migrate. Each of these four sections can be read independently of the others. Finally,  
148 the most important conclusions are summarized in section 8 and a list of important open  
149 issues for future research is included in section 9.

150 The four selected patterns have in common the presence of a coastline and an underlying  
151 topography with a cross-shore slope (sloping beach, sloping shelf), which clearly distin-  
152 guishes an alongshore and a cross-shore coordinate. Also, they occur on wave-dominated  
153 sandy coasts (without vegetation) that are uninterrupted in the alongshore direction at  
154 the length scale of the studied feature. We do not cover other coastal morphodynamic  
155 patterns such as ripples, megaripples, tidal sand waves, tidal sand banks, sorted bed-  
156 forms, cusped shorelines and km-scale shoreline sand waves. A review on ripples, tidal  
157 sand banks and tidal sand waves can be found in *Blondeaux* [2001]. *Gallagher* [2011], and  
158 references therein, studied the formation of megaripples. A review on sorted bedforms  
159 (or rippled scour depressions, related to a physical mechanism based on sediment sorting)  
160 and large-scale cusped shorelines was presented by *Coco and Murray* [2007]. Km-scale  
161 shoreline sand waves have been studied by *van den Berg et al.* [2012] and references  
162 therein.



## 2. COASTAL MORPHODYNAMICS, THE MODEL FRAMEWORK

163 Coastal morphodynamics is the research field that studies the mutual interactions be-  
 164 tween the sea bed morphology and coastal hydrodynamics through sediment transport  
 165 [*Wright and Thom, 1977*]. These interactions are included in the process-based coastal  
 166 area models [*Amoudry and Souza, 2011*] (Figure 3). The sea bed level and the shoreline  
 167 of sandy coasts change due to the divergence/convergence of sediment transport, which  
 168 itself is driven by the bed shear stresses exerted by the flow velocities related to the cur-  
 169 rents, the incoming waves and the *turbulence*. Changes in bed level in turn affect these  
 170 hydrodynamic processes, so feedback mechanisms occur.

171 It is important to keep in mind that those processes can occur at several time scales so  
 172 that the corresponding variables and equations are commonly time-averaged to just keep  
 173 the dynamics at the scale of interest. In particular, each morphological feature has its  
 174 own morphodynamic time scale,  $T_m$ , defined as that at which significant morphological  
 175 changes occur. This scale is roughly  $T_m = O(10^4 s)$  for beach cusps,  $T_m = O(10^5 s)$  for  
 176 surf zone bars and rip channels and  $T_m = O(10^{10} s)$  for shoreface-connected sand ridges.

177 Figure 4 shows the frame of reference commonly used in coastal morphodynamic models.  
 178 The domain represents a sea that is bounded by an alongshore uniform coast. The  $y$ -axis  
 179 is oriented in the alongshore direction, the  $x$ -axis is perpendicular to it, with  $x$  the distance  
 180 to the coastline and the  $z$ -axis is vertical.

### 2.1. Coastal sediment transport and bed evolution

181 Conservation of sediment mass is the key equation of coastal morphodynamics and,  
 182 after some assumptions that are described in Appendix A, can be cast into

$$183 \quad (1 - p) \frac{\partial h}{\partial t} + \vec{\nabla} \cdot \vec{q} = 0 . \quad (1)$$

184 Here  $\vec{\nabla} = (\partial/\partial x, \partial/\partial y)$  is the horizontal nabla operator,  $\vec{q}(x, y, t)$  is the *net* transport  
 185 of sediment per unit width (total volume of sediment crossing the horizontal unit length  
 186 per unit time,  $m^2s^{-1}$ ),  $h(x, y, t) = z_b(x, y, t) - z_{b0}(x)$  is the bed elevation with respect  
 187 to an alongshore uniform background bathymetry,  $z = z_{b0}(x)$ , and  $p$  is *sediment porosity*  
 188 (typically  $p \sim 0.4$ ). The adjective *net* means that  $\vec{q}(x, y, t)$  results from a time-average  
 189 on a time interval that is short enough with respect to the morphological time scale we  
 190 are interested in. Equation (1) states that the bed level rises ( $\partial h/\partial t > 0$ ) at the locations  
 191 where sediment transport converges ( $\vec{\nabla} \cdot \vec{q} < 0$ ) and vice versa (Figure 7).

192 To evaluate bed level changes sediment transport must therefore be computed. Sediment  
 193 transport in the coastal environment is a complex process that depends on the mechanics  
 194 of sediment grains subject to forces exerted by waves and currents. It takes place both  
 195 in suspension (*suspended load*) and in contact with the bed (*bed load*, which may include  
 196 *sheet flow*) [Soulsby, 1997]. Sediment transport is still poorly understood and hard to  
 197 predict accurately [Amoudry and Souza, 2011], due to the complexity of the processes  
 198 involved. On the other hand, field observations suggest that the dynamics of beach cusps,  
 199 rhythmic surf zone bars and shoreface-connected ridges is associated to the action of  
 200 intense currents involving net water mass flux. These observations motivate the working  
 201 hypothesis that the net sediment transport,  $\vec{q}$ , depends on the *depth-averaged current*,  $\vec{v}$ ,  
 202 (net water volume flux per unit width divided by net water depth, see section 2.2) through  
 203 the formula

$$204 \quad \vec{q} = \alpha \vec{v}, \quad (2)$$

205 where  $\alpha$  is the total sediment load (including bedload and suspended load). This formula  
 206 is inspired by the case of suspended load transport with a vertically uniform concentration,

207 in which case the expression is exact and  $\alpha$  is the depth-integrated volumetric sediment  
 208 concentration ( $m^3/m^2$ ). As is further explained in Appendix A, additional sediment  
 209 transport occurs in the cross-shore direction due to a number of sources including gravity  
 210 combined with bottom slope, *wave nonlinearities* and *undertow*. Here it is assumed that  
 211 the joint action of the various cross-shore sediment transport sources, not described by  
 212 equation (2), determines an equilibrium cross-shore beach and inner shelf profile. This  
 213 profile is chosen as the background bathymetry,  $z_{b0}(x)$ , and in the absence of current,  $\vec{v}$ ,  
 214 it is assumed to be stable. The possible unbalance in cross-shore sediment transport due  
 215 to any deviation,  $h(x, y, t)$ , just tends to drive the bathymetry back to equilibrium. This  
 216 is represented by a slope term that is added to equation (2), which becomes

$$\vec{q} = \alpha \vec{v} - \gamma \vec{\nabla} h . \quad (3)$$

218 The rationale behind equation (3) is that oscillatory motions mobilize (or *stir*) the sed-  
 219 iment, due to either the orbital velocities at the bed or the turbulent vortices created  
 220 by breaking waves, without producing a transport. The current increases the stirring  
 221 and transports the sediment (as illustrated in Figure 6). The stirring is represented by  
 222 the *sediment load*  $\alpha$  and the slope coefficient  $\gamma$ , which can depend (nonlinearly) on local  
 223 quantities such as the current magnitude  $|\vec{v}|$ , the amplitudes of the wave orbital velocity  
 224 and the turbulence-induced velocity, the sediment properties and the water depth  $D$ . If  
 225 the velocities at the bed are smaller than a critical value,  $\alpha$  and  $\gamma$  are zero. This formu-  
 226 lation works reasonably well, in the sense that it captures the overall characteristics of  
 227 the processes [*Fredsoe and Deigaard, 1992; Soulsby, 1997; Camenen and Larroudé, 2003*].  
 228 However, it has important limitations that are discussed in detail in Appendix A. The  
 229 most important are that cross-shore sediment transport plays a passive role, driving the

230 bathymetry to the alongshore uniform equilibrium and that the sediment transport is in  
231 equilibrium with the local hydrodynamics so that possible lags are neglected.

232 The sediment load  $\alpha$  in the first term of equation (3) is the total volume of sediment  
233 in motion per horizontal area unit ( $\text{m}^3/\text{m}^2$ ), and can also be interpreted as a ‘stirring  
234 function’ [e.g., *Falqués et al.*, 2000] if this term is understood as describing sediment  
235 being stirred (by waves and currents) up to load  $\alpha$  and then being transported by the  
236 current. Characteristic values of  $\alpha$  range from  $10^{-5} \text{ m}^3/\text{m}^2$ , for bedload conditions, to  
237  $10^{-3} \text{ m}^3/\text{m}^2$ , for total load conditions [*Soulsby*, 1997]. Table 2 shows examples of the  
238  $\alpha$  function for six standard sediment transport formulas that can be cast in the form  
239 of equation (3), all of them described in detail in *Soulsby* [1997]. An illustration of the  
240 applicability of many of these different sediment transport parameterizations (and others)  
241 is given in *Camenen and Larroudé* [2003]. As described in *Soulsby* [1997], the existing  
242 formulas have been extensively calibrated, although mostly in *wave flumes* or outside  
243 the surf zone. Under breaking waves, the strong turbulent vortices can have a significant  
244 amplitude at the bed and add to the sediment stirring by the current and the wave orbital  
245 velocity [*Voulgaris and Collins*, 2000; *Butt et al.*, 2004]. This process is not included in  
246 any of the standard sediment transport formulas (Table 2) but can be included with an  
247 adequate expression of the  $\alpha$  function. For example, *Reniers et al.* [2004] added this  
248 process in the  $\alpha$  of the Soulsby and van Rijn formula and *Ribas et al.* [2011] modified it  
249 and showed its importance for the dynamics of rhythmic surf zone bars. In the surf zone  
250 applications the stirring by turbulent vortices as implemented by *Ribas et al.* [2011] will  
251 be included.

## 2.2. Coastal hydrodynamic processes

252 As explained in the previous section, to calculate the sediment transport some hydro-  
253 dynamic variables, such as the current, the water depth and the wave orbital velocity,  
254 must be computed. Water motion in the coastal zone occurs at different time scales. On  
255 the coasts studied here, incoming waves are the most obvious motion to the eye. The  
256 characteristic time scale of waves is provided by their period,  $T_w$ , which typically ranges  
257 between 1 and 20 s. At shorter time scales turbulent motions take place. Since the rela-  
258 tive amount of sediment carried by the water motion is small (volumetric concentration  
259 of sediment hardly reaches  $O(10^{-2})$ ) the bed level typically changes at a characteristic  
260 time scale (morphodynamic time scale),  $T_m \gg T_w$ . Therefore, it is sufficient to consider  
261 time-averaged hydrodynamic variables and thus filter out the fast dynamics at the time  
262 scales of waves and turbulence. This means that the hydrodynamics is decomposed into  
263 two components: a) mean motions and b) "fast" fluctuating motions. Of course, waves  
264 and turbulence affect the dynamics of the system but their effects are considered only  
265 through averages that are described by the corresponding hydrodynamic forces on the  
266 mean motions. Therefore, all hydrodynamic variables are time-averaged on a time scale  
267  $T_w$ . An exception will be made when describing the dynamics of the *swash zone*, where  
268 the time average needs to be made on a shorter time scale, filtering only the turbulent  
269 motions but not the waves.

270 Another important assumption is that we focus on morphodynamic features located in  
271 shallow waters and the horizontal scales involved in these features are at least one order  
272 of magnitude larger than the vertical scales. It is therefore reasonable to expect that their  
273 dynamics can be understood within the framework of the depth-integrated shallow water  
274 approximation [*Phillips*, 1977; *Mei*, 1989; *Svendsen*, 2006]. Thus, the hydrodynamic

275 variables describing the mean hydrodynamic motions (i.e., the dynamics of the water  
 276 columns) are the depth-averaged current, i.e., the time-averaged water volume flux per  
 277 unit width divided by the time-averaged water depth,  $\vec{v}(x, y, t)$  (hereinafter simply referred  
 278 to as current), and the time-averaged free surface level,  $z_s(x, y, t)$ .

279 Conservation of water mass is one of the fundamental laws for the mean hydrodynamic  
 280 motions. Its depth-integrated formulation reads

$$281 \quad \frac{\partial D}{\partial t} + \vec{\nabla} \cdot (D\vec{v}) = 0, \quad (4)$$

282 where  $D = z_s - z_b$  is the time-averaged water depth. The quantity  $D\vec{v}$  is the volumetric  
 283 flux of water per unit width entering a water column [Svendensen, 2006]. Equation (4) states  
 284 that if there is convergence of water flux (i.e.,  $\vec{\nabla} \cdot (D\vec{v}) < 0$ , meaning that a net quantity of  
 285 water flows into the water column) an increase in water depth will occur (i.e.,  $\partial D/\partial t > 0$ ,  
 286 for instance by increasing the free surface level  $z_s$ , see Figure 5a). Note that, in the swash  
 287 zone, an extra term may appear on the right hand side (RHS) of equation (4), related to  
 288 the infiltration of water into the bed [Dodd *et al.*, 2008].

289 The momentum balance for time and depth-averaged currents

$$290 \quad \frac{\partial \vec{v}}{\partial t} + \vec{v} \cdot \vec{\nabla} \vec{v} = -g\vec{\nabla} z_s + \frac{\vec{\tau}_b}{\rho D} + \frac{\vec{\tau}_w}{\rho D} + \frac{1}{\rho D} \vec{\nabla} \cdot (\mathbf{R} - \mathbf{S}), \quad (5)$$

291 is the other fundamental law governing the mean motions. The LHS (left hand side) is  
 292 the horizontal acceleration of the water columns and the RHS consists of the forces per  
 293 mass unit acting on them. The first term on the RHS represents the pressure gradient  
 294 force per unit mass due to gradients of the free surface level. The second term involves  
 295 the net bed shear stress,  $\vec{\tau}_b$ , which produces frictional forces on the flow and also the wind  
 296 can produce forces described through the free surface shear stresses,  $\vec{\tau}_w$ . The turbulent

297 Reynolds stress tensor,  $\mathbf{R}$ , and the wave radiation stress tensor,  $\mathbf{S}$ , are 2D second order  
 298 symmetric tensors that describe the net depth-integrated transfer of momentum that are  
 299 due to turbulence and waves, respectively. Their divergence, whose x- and y-components  
 300 are (e.g.,  $\vec{\nabla} \cdot \mathbf{S}$ )

$$301 \quad \frac{\partial S_{xx}}{\partial x} + \frac{\partial S_{xy}}{\partial y} \quad , \quad \frac{\partial S_{yx}}{\partial x} + \frac{\partial S_{yy}}{\partial y} \quad , \quad (6)$$

302 results in a force acting on the water columns. For beach cusps  $\mathbf{S}$  is absent since the time  
 303 average is made on a time scale shorter than  $T_w$ , filtering the turbulent motions but not  
 304 the waves. Moreover, for large scale features  $O(1-10 \text{ km})$ , appearing on the continental  
 305 shelf, the Coriolis volumetric force is added on the RHS of equation (5).

306 Knowing the bed level  $z_b(x, y, t)$ , the system of the hydrodynamic equations (4) and (5)  
 307 is not closed mainly because the stress tensors depend on the fast fluctuating hydrody-  
 308 namic components, i.e., turbulence and waves. Turbulent stresses play a secondary role  
 309 and are modeled with the standard eddy viscosity approach [Svendsen, 2006] so that they  
 310 are proportional to  $\nabla \vec{v}$  components through a mixing coefficient that depends on wave  
 311 energy dissipation. However, wave radiation stresses are crucial in the surf zone as they  
 312 provide the main driving force for the currents. They depend on *wave energy density*, on  
 313 the propagation direction and on the ratio  $c_g/c$  ( $c_g, c$  being the group and phase celerities)  
 314 [Longuet-Higgins and Stewart, 1964; Svendsen, 2006]. Essentially, when waves approach  
 315 the coast and feel the sea bottom they start *refracting, shoaling* and breaking, varying  
 316 their energy density and direction. These changes cause in turn gradients in the radiation  
 317 stresses producing net forces on the water column. The net bed shear stresses in equa-  
 318 tion (5),  $\tau_{bi}$ , are parameterized in terms of  $\vec{v}$ , the wave orbital velocity at the edge of the  
 319 boundary layer and a friction coefficient that depends on  $D$ , sediment size and unresolved

320 small bedforms. This is not straightforward and different options can be used [e.g., *Fed-*  
 321 *dersen et al.*, 2000]. The specific equations and parameterizations used to describe the  
 322 different features can be found in *Caballeria et al.* [2002] (surf zone bars), *Calvete et al.*  
 323 [2001] (shoreface-connected ridges), and *Dodd et al.* [2008] (beach cusps).

324 Although the individual wave motions are not resolved in the present formulation, the  
 325 knowledge of the time-averaged properties of waves is nevertheless crucial. These include  
 326 wave energy density, energy dissipation, orbital velocity amplitude, angle and wavenum-  
 327 ber. In fact, all these quantities can be computed in terms of the root-mean-square height,  
 328  $H(x, y, t)$ , the wavenumber,  $k(x, y, t)$ , and the wave angle,  $\theta(x, y, t)$  (Figure 4) and these  
 329 three variables can be evaluated using the dispersion relation, the wavenumber irrotation-  
 330 ality and the wave energy balance. The details and the corresponding set of equations,  
 331 which are subsequently coupled to equations (4) and (5), are described in Appendix B.

332 A common assumption regarding coastal morphodynamics is the so-called *quasi-steady*  
 333 *approximation*. It consist of dropping out all the time derivatives from the hydrodynamic  
 334 equations (4) and (5) but not from the bed evolution equation (1). It is not an essential  
 335 step for the methodology explained in this contribution but it facilitates the physical  
 336 interpretation of the equations. For instance, the mass conservation equation (4) becomes  
 337  $\vec{\nabla} \cdot (D\vec{v}) = 0$ , which means that there is no net water transport into or out from the water  
 338 column. A gradient in  $D$  then implies a change in  $\vec{v}$  (Figure 5b shows a 1D example of a  
 339 current increase due to a decreasing water depth).

340 The quasi-steady assumption means that the hydrodynamics is in equilibrium with the  
 341 morphology all the time, i.e., the hydrodynamic variables are assumed to adapt instan-  
 342 taneously to the bed level so that the former vary only when the latter changes. This



343 assumption suppresses any oscillatory solution of equations (4) and (5) like *infragravity*  
 344 *waves, shear waves, low frequency eddies* and tidal waves. The first three types of motion  
 345 occur with periods ranging from about 20 s to  $O(10^3 \text{ s})$  [Reniers et al., 2004] while tides  
 346 occur at periods of  $O(10^4 \text{ s})$ . The quasi-steady approximation can be applied if the os-  
 347 cillatory water motions do not affect significantly the morphologic evolution. This is not  
 348 the case for beach cusps, which are in fact closely linked to the unsteady wave motion as  
 349 it is expressed in the uprush and backwash of the waves. On the other hand, despite low  
 350 frequency eddies may affect crescentic bar dynamics [Reniers et al., 2004] and infragravity  
 351 waves (edge waves) had earlier been thought to be the primary cause of rhythmic surf zone  
 352 features (Holman and Bowen [1982] and others, see sections 4.2 and 5.2), it is nowadays  
 353 accepted that these low-frequency oscillatory motions are not essential for the formation  
 354 of rhythmic bars in the surf zone [Blondeaux, 2001; Coco and Murray, 2007]. Similarly,  
 355 although tidal oscillations mildly affect the evolution of shoreface-connected sand ridges,  
 356 they are not essential for explaining their formation [Walgreen et al., 2002]. The quasi-  
 357 steady assumption is therefore applied to understand the dynamics of surf zone rhythmic  
 358 bars and shoreface-connected sand ridges.

### 3. FORMULATION AND METHODOLOGY BASED ON THE DEPTH-AVERAGED SEDIMENT CONCENTRATION

#### 3.1. Bed evolution equation

359 A formulation of the bed evolution equation based on the *depth-averaged sediment*  
 360 *concentration* is now derived. For this, we substitute  $\vec{q}$  from equation (3) into equation (1)  
 361 to obtain the so-called bed evolution equation (BEE),

$$362 \quad (1 - p) \frac{\partial h}{\partial t} = -D\vec{v} \cdot \vec{\nabla}C - C \vec{\nabla} \cdot (D\vec{v}) + \vec{\nabla} \cdot (\gamma \vec{\nabla}h), \quad (7)$$

363 where  $\mathcal{C} = \alpha/D$  is the total sediment load divided by the water depth. In the present  
 364 contribution,  $\mathcal{C}$  is interpreted as a depth-averaged sediment concentration (DASC) and  
 365 it includes both bedload and suspended load. Some authors [*Falqués et al.*, 2000] have  
 366 called it ‘potential stirring’ but here we use the name DASC because it is related to a  
 367 variable that can be measured. In the outer surf zone and the continental shelf, bottom  
 368 changes occur at depths larger than  $O(1 \text{ m})$ . In the inner surf zone and the swash zone,  
 369 water depths range between  $0.1 - 1 \text{ m}$ . Given that  $\alpha \leq 10^{-3} \text{ m}^3/\text{m}^2$  (see section 2.1),  
 370 characteristic values of  $\mathcal{C}$  range from  $10^{-7}$  to  $10^{-2} \text{ m}^3/\text{m}^3$  ( $\mathcal{C}$  could be higher only in  
 371 the very shallow swash zone). The left hand side (LHS) of equation (7) quantifies the  
 372 bottom changes. The first term on the RHS describes the erosion/deposition produced  
 373 due to the advection of  $\mathcal{C}$  by the depth-averaged current  $\vec{v}$  when there are gradients of  $\mathcal{C}$   
 374 (section 3.2 is devoted to explain in depth the physical interpretation of this term). The  
 375 second term on the RHS describes the deposition (erosion) that occurs when water flux  
 376 converges (diverges). The third term on the RHS is a slope-induced diffusive term and  
 377 tends to damp the gradients in bed level.

378 If the quasi-steady hypothesis can be assumed (i.e., for surf zone and inner shelf fea-  
 379 tures), the mass conservation equation (4) becomes  $\vec{\nabla} \cdot (D\vec{v}) = 0$  (see section 2.2) and the  
 380 BEE becomes

$$381 \quad (1 - p) \frac{\partial h}{\partial t} = -D\vec{v} \cdot \vec{\nabla} \mathcal{C} + \vec{\nabla} \cdot (\gamma \vec{\nabla} h) . \quad (8)$$

382 In the application at the swash zone (beach cusp development), where the quasi-steady  
 383 hypothesis does not hold, equation (7) is used, with an additional term related to water  
 384 infiltration into the bed (see section 7). An equation similar to equation (8) was first  
 385 derived and used for the nearshore by *Caballeria et al.* [2002].

### 3.2. Erosion/deposition processes

Equation (8) gives the time evolution of the bed level deviations at any location as a function of the water depth,  $D$ , the depth-averaged current,  $\vec{v}$ , and the gradient of the DASC,  $\mathcal{C}$ . It is not a closed equation since it needs the knowledge of  $\vec{v}$  and the distribution of  $\mathcal{C}$ . The powerful advantage of equation (8), with respect to the original equation (1), is that it allows for an interpretation of the erosion/deposition processes, in terms of  $\vec{v}$  and  $\vec{\nabla}\mathcal{C}$ , which might be known from field observations, from numerical simulations or just qualitatively from physical reasoning.

According to equation (8),  $\vec{v} \cdot \vec{\nabla}\mathcal{C} > 0$  will tend to induce bed erosion ( $\partial h/\partial t < 0$ ) and  $\vec{v} \cdot \vec{\nabla}\mathcal{C} < 0$  will tend to induce bed accretion ( $\partial h/\partial t > 0$ ). In words, *any current with a component in the direction of the gradient in  $\mathcal{C}$  will produce erosion and any current with a component that opposes this gradient will cause accretion* (Figure 8). This behavior can be physically understood from the fact that  $\mathcal{C}$  is in local equilibrium with the flow, i.e., it is the depth-averaged sediment concentration of the water column corresponding to the stirring by the local hydrodynamics (section 2.1 and Appendix A). If  $\mathcal{C}$  increases along the flow ( $\vec{v} \cdot \vec{\nabla}\mathcal{C} > 0$ ), water with little amount of  $\mathcal{C}$  will move to places where the stirring by the hydrodynamics allows for larger  $\mathcal{C}$ . Therefore, more sediment will be picked up from the bed underneath the water column, which will hence be eroded (Figure 8a). The contrary will happen if  $\mathcal{C}$  decreases along the flow (Figure 8b).

### 3.3. Linearized bed evolution equation

In order to understand the dominant mechanisms involved in the initial formation of the features of interest, it is convenient to assume that the state of the system is a superposition of an initially alongshore uniform steady state (the equilibrium state already

defined in section 2.1) and a perturbed state, with small amplitude perturbations that evolve from the equilibrium state [Dodd *et al.*, 2003]. The equilibrium state represents the mean dynamic balance in the absence of rhythmic features. It consists of an alongshore uniform equilibrium profile  $z_{b0}(x)$  (already mentioned in section 2.1), a depth-averaged sediment concentration  $\mathcal{C}_0(x)$ , a water depth  $D_0(x)$  that includes the wind- or wave-induced set-up/set-down, and often an *alongshore current*  $V_0(x)$ . The set-up (set-down) is an over-elevation (under-elevation) of the free surface level in the coastal zone forced by the cross-shore transfer of momentum after waves break or by wind-induced cross-shore forces. The alongshore current is forced by the alongshore momentum transfer produced after oblique waves break, by wind-induced alongshore forces or by free surface gradients. A schematic representation of the alongshore current and the wave-induced set-up can be seen in Figure 4.

Small perturbations in bed level,  $h(x, y, t)$  (the bed level deviations defined in section 2.1, but now assumed to be small), concentration,  $c(x, y, t)$ , depth,  $d(x, y, t)$ , and current,  $(u(x, y, t), v(x, y, t))$ , are added to the equilibrium. The total variables then read

$$z_b = z_{b0} + h, \quad \mathcal{C} = \mathcal{C}_0 + c, \quad D = D_0 + d \quad \text{and} \quad \vec{v} = (0, V_0) + (u, v). \quad (9)$$

Substituting these expressions into equation (8) and only retaining the terms that are linear in the small quantities ( $u, v, c, d$  and  $h$ ), yields the linearized BEE,

$$(1 - p) \frac{\partial h}{\partial t} = -D_0 u \frac{d\mathcal{C}_0}{dx} - D_0 V_0 \frac{\partial c}{\partial y} + \frac{\partial}{\partial x} \left( \gamma_0 \frac{\partial h}{\partial x} \right) + \frac{\partial}{\partial y} \left( \gamma_0 \frac{\partial h}{\partial y} \right). \quad (10)$$

Here  $\gamma_0$  is the equilibrium value of the  $\gamma$  coefficient in equation (8).

Equation (10) shows that the small bed level changes of a known equilibrium state can be analyzed just from the perturbations that the bottom produces in the cross-shore com-

429 ponent of the current,  $u$ , and the alongshore gradients of the perturbation of the DASC,  
 430  $\partial c/\partial y$ . The first RHS term of equation (10) leads to deposition (erosion) if  $u d\mathcal{C}_0/dx$  is  
 431 negative (positive). The second RHS term of equation (10) leads to deposition (erosion) if  
 432  $V_0 \partial c/\partial y$  is negative (positive). Note that if  $V_0 = 0$ , the second RHS term disappears and  
 433 erosion/deposition processes only depend on  $u$  and the cross-shore gradient of the equilib-  
 434 rium DASC,  $d\mathcal{C}_0/dx$ . The last two RHS terms of equation (10) have a diffusive effect on  
 435 the bed perturbations. The first derivation of a linearized BEE similar to equation (10)  
 436 for the nearshore was made by *Falqués et al.* [1996].

### 3.4. Erosion/deposition patterns: global analysis

437 The equations and analysis of the previous sections, based purely on equations (8) or  
 438 (10), are local in the sense that these equations describe the bed level evolution in one  
 439 location due to local convergence/divergence of the sediment transport. Since this con-  
 440 tribution aims at understanding the development of morphologic patterns that grow and  
 441 migrate on the whole domain, it is essential to understand the erosion/deposition patterns  
 442 occurring on the whole domain. As an example, given a morphological feature consist-  
 443 ing of alongshore alternating bars and troughs, Figure 9 shows what erosion/deposition  
 444 patterns would produce (a) pure growth, (b) growth and down-drift migration, (c) pure  
 445 down-drift migration, (d) decay and down-drift migration, and (e) pure decay of the fea-  
 446 ture. Thereby, it is essential to analyze the effect integrated on the whole domain of the  
 447 different terms in equations (8) or (10) in order to evaluate their influence on growth,  
 448 decay or migration of the features. As a first step, this can be done in a qualitative way,  
 449 i.e., by visual observation of the erosion/deposition patterns created by each of the terms  
 450 (and comparing with the patterns in Figure 9). For instance, in the case of the first

451 term on the RHS of equation (8), and consistently with the local analysis presented in  
 452 section 3.2, if the regions with  $h > 0$  and the current opposing the gradients in DASC (or  
 453  $h < 0$  and the current running with the gradients in DASC) dominate over the regions  
 454 where the contrary occurs, this term will contribute to the growth of the feature. Al-  
 455 ternatively, a quantitative global analysis of the equations can be performed by taking a  
 456 specific average over the horizontal domain of the different terms in equations (8) or (10).  
 457 The global effect of each term on the development of the morphological patterns can then  
 458 be studied quantitatively. The technical details of how such a quantitative global analysis  
 459 is performed are given in Appendix C.

460 In the next sections it will be demonstrated that, for a wide range of alongshore rhythmic  
 461 morphological patterns, the global effect of the first RHS term of equation (10) essentially  
 462 contributes to the initial growth of the features, the second RHS term essentially con-  
 463 tributes to their alongshore migration (in the presence of an alongshore current), and  
 464 the last two terms produce decay of the features. This is a powerful result: the 'sponta-  
 465 neous' breaking of alongshore uniformity of the nearshore bathymetry and the emergence  
 466 of alongshore rhythmic morphological patterns can be understood by knowing only the  
 467 cross-shore gradient of the equilibrium DASC,  $d\mathcal{C}_0/dx$  and the cross-shore perturbation  
 468 of the induced horizontal currents,  $u$  (first RHS term of equation (10)).

### 3.5. Methodology to use the DASC to explain pattern development

469 The development of four different alongshore rhythmic morphodynamic patterns are  
 470 explained in the next sections: two surf zone patterns (crescentic bars and transverse  
 471 bars), one at the continental shelf (shoreface-connected ridges) and one at the swash zone

472 (beach cusps). The following three steps are taken to use the DASC to understand the  
473 formation of these morphological patterns.

474 First, it is essential to have information on the gradients in DASC. This quantity is  
475 difficult to measure in the surf zone due to the highly complex dynamics of sediment  
476 transport under breaking waves [*Soulsby, 1997*] and it is also highly unknown in the  
477 swash zone. Thereby, very often the formulations are inferred from laboratory data and  
478 theoretical reasoning. Different parameterizations found in the literature (like those in  
479 Table 2) lead to different results for  $\mathcal{C}$ , which can strongly affect the morphological changes.  
480 Here, physical reasons will be presented as to what  $\mathcal{C}$  profiles are expected in the areas  
481 where features develop. Also, one of the formulas in Table 2 (which has been extensively  
482 calibrated against data) will be applied to substantiate the reasoning. Since the bed  
483 evolution depends on the gradients of the concentration (section 3.2), it is crucial that the  
484 parameterizations of sediment transport adequately represent not only the magnitude of  
485 DASC, but especially the gradients of DASC.

486 Second, some information on the hydrodynamics induced by the growing feature is  
487 needed. This information can be obtained by measurements and/or with the hydrody-  
488 namic module of the morphodynamic models. The latter is usually quite robust, i.e., there  
489 is little difference between the different models, even though different parameterizations  
490 are used for the bed shear stresses, the turbulence-induced effects, wave energy dissipa-  
491 tion through breaking, etc. Here, the focus will be on describing the horizontal currents  
492 (and especially the cross-shore component  $u$ ) associated with each feature, discussing in  
493 a qualitative way the essential physical processes that create these currents.

494 Third, a global analysis (either qualitative or quantitative) of the linearized BEE (10)  
495 must be performed in order to understand the erosion/deposition patterns, created by the  
496 joint action of the horizontal currents and the gradients in DASC, that causes the initial  
497 formation of the features. This allows understanding the initial shape of the pattern,  
498 its initial growth and migration rates, and under which climate conditions the feature  
499 develops.

500 Finally, a global analysis (either qualitative or quantitative) of the nonlinear BEE (8) is  
501 performed to understand the finite amplitude behavior of the features: *saturation of the*  
502 *growth* and changes in shape and migration rate. In some cases, this analysis also allows  
503 explaining the destruction of the features by certain climate conditions.

## 4. CRESCENTIC BARS

### 4.1. Characteristics of observed crescentic bars (and rip channels)

504 Crescentic bars are located in the surf zone of micro to meso-tidal sandy beaches [*Lipp-*  
505 *mann and Holman, 1990; van Enckevoort et al., 2004; Lafon et al., 2004*] (Figures 1a  
506 and 2g). The alongshore spacing between crescentic bar horns is relatively constant for a  
507 specific system. They have been reported at different scales with a mean spacing ranging  
508 from tens of m up to 2-3 km. Crescentic bars are sometimes also called rip channel sys-  
509 tems because the rip channels are a striking and well known characteristic of them [*van*  
510 *Enckevoort and Ruessink, 2003*]. Note, however, that rip channels, i.e., bed depressions or  
511 cross-shore oriented channels in the surf zone where rip currents concentrate, can also be  
512 observed without the presence of crescentic bars (see, e.g., *MacMahan et al. [2005]* and  
513 also section 5).



514 Crescentic bars are linked to shore-parallel bars, which are alongshore uniform sand  
515 bars parallel to the coast. The latter form in medium sand beaches during high-energy  
516 wave events. Crescentic bars develop out of the shore-parallel bar during decreasing wave  
517 energy (Figure 10), i.e., during post-storm conditions. In the widely accepted beach  
518 state classification [*Wright and Short*, 1984; *Lippmann and Holman*, 1990], such process  
519 is classified as the down state transition from the Longshore Bar and Trough state to  
520 the *Rhythmic Bar and Beach (RBB) state*. Crescentic bars can become shore-parallel  
521 again in the reverse (up state) transition if wave energy increases again (Figure 10).  
522 The latter process is called bar straightening or morphologic reset. Recent studies have  
523 stressed the effect of wave obliquity in the transitions between shore-parallel and crescentic  
524 bars, revisiting the traditional classification of *Wright and Short* [1984]. They found that  
525 crescentic bars seem to develop preferably for normal wave incidence and bar straightening  
526 occurs for highly oblique waves [*Holman et al.*, 2006; *Thornton et al.*, 2007; *Splinter et al.*,  
527 2011; *Price and Ruessink*, 2011].

528 Along beaches with crescentic bars the shoreline often features undulations with a sim-  
529 ilar alongshore spacing. Since this spacing is typically significantly larger than the one  
530 of ordinary beach cusps, these undulations are called megacusps [*Thornton et al.*, 2007].  
531 The horns of the crescentic bars can connect to the shoreline and to the megacusp system  
532 during long-lasting conditions of low wave energy (down state transition from the RBB  
533 state to the *Transverse Bar and Rip (TBR) state* [*Wright and Short*, 1984; *Ranasinghe*  
534 *et al.*, 2004]). The resulting transverse bar system is a particular case of the four different  
535 types of transverse bar systems that will be discussed in section 5.

## 4.2. Existing theories for their formation

536 The origin of crescentic bars was first explained with the so-called hydrodynamic tem-  
537 plate theory, in which the morphologic pattern is the result of a pre-existing similar pattern  
538 in the hydrodynamics [see the review by *Coco and Murray, 2007*]. More specifically, their  
539 formation was attributed to the pattern of near-bed velocities associated with *edge waves*  
540 [*Bowen and Inman, 1971; Holman and Bowen, 1982*], which are alongshore propagating  
541 trapped waves. Edge waves with alongshore spacings at the crescentic bar scale can be  
542 generated by infragravity oscillations associated with the incident wind or swell waves.  
543 However, more recent studies have shown that the edge wave hypothesis is only partially  
544 consistent with available field data [*Coco and Murray, 2007*].

545 The second theory, which was first suggested by *Hino [1974]*, is that crescentic bars  
546 emerge as a *morphodynamic instability* of the system with a shore-parallel bar. That  
547 is, they emerge from a positive feedback between wave-driven currents and morphology,  
548 starting from any perturbation of the featureless state. The first study modeling the  
549 formation of a crescentic bar from a shore-parallel bar by *self-organization* was that of  
550 *Deigaard et al. [1999]*. *Falqués et al. [2000]* described in more detail the physical mecha-  
551 nisms involved, emphasizing the role of the depth-averaged sediment concentration (called  
552 potential stirring in that paper). The instability mechanism was called ‘bed-surf instabil-  
553 ity’ (term introduced by *Falqués et al. [1996]*) because it is essentially due to the positive  
554 feedback between the sea bed perturbations and the distribution of wave breaking. Later  
555 on, *Calvete et al. [2005]* used a more realistic model that reproduced many of the observed  
556 characteristics of crescentic bars and confirmed the important role of the DASC in cres-  
557 centic bar formation. The self-organized origin of crescentic bars has been supported by  
558 numerous other modeling studies [*Caballeria et al., 2002; Damgaard et al., 2002; Ranas-*

559 *inghe et al.*, 2004; *Reniers et al.*, 2004; *Klein and Schuttelaars*, 2006; *Garnier et al.*, 2008;  
 560 *Smit et al.*, 2008] making this theory currently more widely accepted than the hydrody-  
 561 namic template theory. In the next section, the role of the DASC in the transformation  
 562 of shore-parallel bars into crescentic bars will be discussed based on the studies of *Falqués*  
 563 *et al.* [2000], *Calvete et al.* [2005] and *Garnier et al.* [2008].

### 4.3. Role of DASC in the formation mechanism

564 For the sake of simplicity, and since crescentic bars mainly develop for relatively small  
 565 wave incidence angles, the focus of this section is on the case of normally incident waves  
 566 (i.e., no alongshore current,  $V_0 = 0$ ). The case of oblique waves will be discussed later in  
 567 section 4.4. As explained in section 3, the joint effect of the gradients in DASC and the  
 568 horizontal circulation induced by the growing feature creates the erosion/deposition pat-  
 569 terns that explain why the feature grows. The three steps of the methodology (section 3.5)  
 570 to explain crescentic bar formation when  $V_0 = 0$  are: 1) describing the cross-shore dis-  
 571 tribution of the DASC (i.e.,  $d\mathcal{C}_0/dx$ ), 2) understanding the horizontal circulation induced  
 572 by the growing feature (i.e.,  $u$ ), and 3) analyzing the erosion/deposition patterns with  
 573 the linearized BEE (10) and the knowledge of  $d\mathcal{C}_0/dx$  and  $u$ . These steps are done in the  
 574 three following subsections.

#### 4.3.1. Depth-averaged sediment concentration profile

576 As stated before, measuring the sediment concentration in the surf zone is difficult,  
 577 hence available data are scarce. However, the theory presented here is based on a simple  
 578 and robust property of the sediment concentration in the surf zone of barred beaches.  
 579 For moderate wave conditions, waves break predominantly over the shore-parallel bar  
 580 inducing a strong sediment concentration over it. This intuitive property is confirmed by

581 most of the sediment transport formula applied for the surf zone. More precisely, in surf  
 582 zones that are characterized by a shore-parallel bar waves can break on the bar (somewhat  
 583 seaward of it) if they have a sufficient height (Figure 11). This causes an intense sediment  
 584 resuspension in that area (by wave orbital velocities and turbulent vortices, if included),  
 585 i.e., the sediment load  $\alpha$  (equation 3) is maximum at a certain point on the seaward flank  
 586 of the bar. Furthermore, the water depth has a local minimum at the top of the bar and  
 587 increases onshore and offshore of the crest. Therefore, the depth-averaged concentration  
 588 (DASC,  $\mathcal{C} = \alpha/D$ ) is maximum at a location  $x = x_m$  slightly seaward from the crest.  
 589 Thus, there is an offshore-directed gradient in  $\mathcal{C}$  for  $x < x_m$  and an onshore-directed  
 590 gradient for  $x > x_m$ . This qualitative behavior is reproduced by all the formulations for  
 591 nearshore sediment transport included in Table 2. As an example, the middle panel of  
 592 Figure 11 shows the DASC profile obtained with the Soulsby-van Rijn formula [Soulsby,  
 593 1997], extended to include an extra sediment stirring produced by turbulent vortices  
 594 [Reniers et al., 2004; Ribas et al., 2011].

#### 595 4.3.2. Rip current circulation

596 The horizontal circulation that is produced over an incipient crescentic bar (i.e., a shore-  
 597 parallel bar with small-amplitude channels) is the well known rip current circulation.  
 598 For normally or nearly normal wave incidence, breaking waves over the small-amplitude  
 599 crescentic bar induce a circulation cell with offshore flow at the channels and onshore flow  
 600 at the *shoals* (Figure 12a). This is a robust characteristic that has been observed in the  
 601 field [MacMahan et al., 2006; Moulton et al., 2013] and in wave-basin experiments [Haller  
 602 et al., 2002; Castelle et al., 2010a], and that is also commonly reproduced by models  
 603 [Garnier et al., 2008; Dalrymple et al., 2011]. Thereby, the cross-shore flow perturbation

604  $u$  (equation 10) can be assumed to have its maximum seaward-directed value ( $u > 0$ ) in  
 605 the channels (i.e., where  $h < 0$ ) and its maximum shoreward-directed value ( $u < 0$ ) over  
 606 the shoals (i.e., where  $h > 0$ ). The basic physics underlying this circulation is explained  
 607 in Appendix D.

### 608 4.3.3. Formation mechanism

609 Now, the joint morphodynamic effect of the gradients in DASC ( $\mathcal{C}$ ) and the horizontal  
 610 circulation can be inferred from the linearized BEE (10). Since we focus on the case of  
 611 normally incident waves, there is no alongshore current in the equilibrium state ( $V_0 = 0$ )  
 612 and the second RHS term of equation (10) drops out. The gradient in  $\mathcal{C}$  is offshore-  
 613 directed over the bar crest, where the current flows onshore (negative  $u$ , see Figure 12a).  
 614 Thereby,  $u d\mathcal{C}_0/dx < 0$  in equation (10), which means that the current carries sediment  
 615 from offshore, where  $\mathcal{C}$  is largest, to the shoal (see section 3.2 and Figure 8). In the  
 616 channels, it is the other way around: the current flows offshore and it carries sediment from  
 617 the channel to offshore. In this way, the circulation will further erode the channels and  
 618 deposit the sand on the shoals. Thus a positive feedback will occur that will enhance both  
 619 the circulation and the bed undulation and the initially shore-parallel bar will develop rip  
 620 channels flanked by shoals (Figure 12b). In addition to that, and given that the position  
 621 of the maximum in DASC,  $x = x_m$ , is close to the crest, the rip currents extend offshore  
 622 of this location and cause deposition of sediment seaward of it because the gradient in  $\mathcal{C}$   
 623 has an opposite sign there. Similarly, the onshore flowing part of the circulation will cause  
 624 erosion seaward of the shoals. Thus, the combination of the DASC and the circulation  
 625 creates not only channels and shoals on the bar but a mirrored pattern offshore of the  
 626 bar (Figure 12b). The addition of the initial shore-parallel bar and the double rows

627 of alternating shoals and channels produces an undulation of the bar in plan-view, with  
628 onshore protruding sections coinciding with the shoals over the bar and offshore protruding  
629 sections at the rip channels. This is the typical crescentic bar morphology (Figure 10).

#### 4.4. Finite amplitude behavior

630 The self-organization models cited in section 4.2 are able to reproduce the initial for-  
631 mation of the crescentic bars with the appropriate shape, but they are unable to explain  
632 the saturation of growth of crescentic bars. The latter process was first simulated by  
633 *Garnier et al.* [2008] (Figure 13). The specific mechanisms for the growth saturation  
634 were explained by *Garnier et al.* [2010] with the use of the global analysis (methodology  
635 described in Appendix C). The saturation of bar height, preventing the accreting shoals  
636 to reach the sea surface, occurs mainly due to a weakening of the positive feedback (term  
637  $-D\vec{v} \cdot \nabla C$  in equation (8)) rather than to an increase of the damping caused by the dif-  
638 fusive transport. The positive feedback weakens but does not vanish: it balances with  
639 the diffusive term (which remains constant), and therefore the latter is also essential to  
640 the saturation process. The weakening of the positive feedback is related to changes in  
641 bar shape rather than to the growth in amplitude. It turns out that the most important  
642 change in shape is that the shoals widen and the channels narrow. More details of the  
643 global analysis applied to the full saturation of crescentic bars are given in Appendix E.  
644 Furthermore, the shoals shift shoreward and the channels seaward with the result that the  
645 bars move overall onshore. This last result shows that the current circulation associated  
646 with well developed crescentic bars system contributes to the attachment of the crescentic  
647 bars to the shore observed in the field [*Ranasinghe et al.*, 2004; *Garnier et al.*, 2008].

648 Another important finite amplitude behavior of the crescentic bar systems is the bar  
 649 straightening. Using the global analysis, *Garnier et al.* [2013] explained why a developed  
 650 crescentic bar straightens due to wave obliquity. Oblique waves inhibit the formation of  
 651 rip channels and straighten crescentic bars because they weaken the rip current intensity  
 652 and cause a down-wave shift of the rips with respect to the channels (i.e., a phase lag  
 653 between the rips and the channels). This weakens the positive feedback between flow and  
 654 morphology given by the term  $-D\vec{v} \cdot \nabla \mathcal{C}$  in equation (8). A more detailed explanation is  
 655 included in Appendix E.

#### 4.5. Discussion

656 All the analysis presented so far in this section concerns a single crescentic bar system. In  
 657 nature, two crescentic bars can coexist in the same beach at different cross-shore positions  
 658 [*van Enkevort et al.*, 2004; *Castelle et al.*, 2007; *Price and Ruessink*, 2011]. In general,  
 659 such double bars do not behave independently. The outer crescentic bar may emerge from  
 660 self-organization (independently of the inner bar), it then induces alongshore variability  
 661 in the onshore hydrodynamics, which in turn forces the morphologic response of the inner  
 662 bar. This behavior is called morphologic coupling [*Castelle et al.*, 2010b]. It is important  
 663 to notice that even in the case of morphologic coupling, the self-organization feedbacks  
 664 between flow and morphology described in section 4.3 still affect the evolution of both  
 665 bars [*Coco and Calvete*, 2009; *Thiebot et al.*, 2012]. Particularly, for a double crescentic  
 666 bar system the DASC profile exhibits a local maximum over each bar system.

667 Although the present contribution is dedicated to rhythmic patterns observed in open  
 668 beaches, it should be stated that crescentic bars are also observed in embayed beaches  
 669 [*Short*, 1999; *Holman et al.*, 2006] that are beaches laterally bounded by headlands or

670 coastal structures. The presence and characteristics of those bars are then conditioned by  
671 the length of the beach (i.e., distance between headlands) but they still seem to emerge  
672 from the basic positive feedback described in section 4.3 [*Castelle and Coco*, 2012].

673 The results presented in this section are taken from previous studies that consider  
674 idealized simplified conditions. Particularly these studies consider an initial bathymetry  
675 that is alongshore uniform [*Falqués et al.*, 2000; *Garnier et al.*, 2008, 2010] or a bathymetry  
676 with a very specific variability [*Garnier et al.*, 2013]. Furthermore, the incoming wave  
677 field is assumed to be time invariant and alongshore uniform. Other modeling studies have  
678 discussed the variability in the wave forcing [*Reniers et al.*, 2004; *Castelle and Ruessink*,  
679 2011] or in the initial bathymetry [*Tiessen et al.*, 2011; *Smit et al.*, 2012]. They show that  
680 this affects the characteristics (e.g., spacing between rip channels) and the dynamics (e.g.,  
681 growth times) of the crescentic bars. However, the feedbacks between flow and morphology  
682 associated with the advection of DASC by the rip currents described in section 4.3 still  
683 play a key role.

## 5. TRANSVERSE BARS

### 5.1. Characteristics of observed transverse bars

684 Apart from the crescentic bars discussed in section 4, the surf zone can also display an-  
685 other kind of morphodynamic feature consisting of several transverse bars separated by an  
686 approximately constant alongshore distance (Figures 1b,c and 2f). The alongshore spac-  
687 ing is defined as the distance between successive bar crests. They are typically attached to  
688 the shoreline and extend into the seaward direction, either approximately perpendicular  
689 to the coastline or with a certain oblique orientation if an alongshore current is present. If  
690 the crests are shifted in (against) the direction of the alongshore current we use the term



691 *down-current (up-current) oriented bars* (Figure 14). However, we use the general term  
 692 transverse bars to refer to all of them, a term introduced by *Shepard* [1952] to distinguish  
 693 them from the shore-parallel bars. In the presence of an alongshore current, they migrate  
 694 downdrift with migration rates up to 40 m/day [*Hunter et al.*, 1979; *Konicki and Holman*,  
 695 2000; *Ribas and Kroon*, 2007; *Pellón et al.*, 2014]. Amplitudes (from a point in the bar  
 696 crest to a point in the trough) can range from 0.3 to 2 m [*Konicki and Holman*, 2000; *De*  
 697 *Melo Apoluceno et al.*, 2002; *Pellón et al.*, 2014; *Gelfenbaum and Brooks*, 2003]. In some  
 698 cases, the bars have been observed to show an asymmetry of the alongshore shape (the  
 699 down-current flank being steeper than the up-current flank [*Pellón et al.*, 2014]). Various  
 700 types of transverse bars (in their characteristics and origin) have been reported in the  
 701 literature (Table 3). In order to distinguish between them, we first follow the classifica-  
 702 tion made by *Pellón et al.* [2014], based on the differences in bar length scales and in the  
 703 environment where they are observed.

704 **Type (1): TBR bars.** The most common type is that conforming the transverse  
 705 bar and rip (TBR) state in the standard beach state classifications [*Wright and Short*,  
 706 1984; *Lippmann and Holman*, 1990; *Castelle et al.*, 2007]. The TBR bars are commonly  
 707 observed in open beaches under medium-energy conditions. They are typically wide and  
 708 short-crested (Figure 1b) and their origin is the merging of a crescentic bar into the beach  
 709 [*Sonu*, 1973; *Wright and Short*, 1984] (they have been mentioned in section 4.1), so that  
 710 their spacing is strongly related to that of the pre-existing crescentic bar. They can be  
 711 approximately perpendicular to the shore [*Hunter et al.*, 1979; *Wright and Short*, 1984]  
 712 or down-current oriented (Figure 14b) when incoming waves arrive with a predominant  
 713 obliquity [*Lafon et al.*, 2004; *Castelle et al.*, 2006]. As in the case of crescentic bars, TBR

714 bars also show strong and narrow rip currents flowing seaward in the troughs and wider  
715 and weaker onshore flows over the crests [*Short*, 1999].

716 **Type (2): Medium energy finger bars.** These transverse bars (Figure 14d) have  
717 been observed in open microtidal beaches under medium-energy conditions [*Konicki and*  
718 *Holman*, 2000; *Ribas and Kroon*, 2007; *Ribas et al.*, 2014] and they always coexist with  
719 shore-parallel (or crescentic) bars. The term finger bars refers to their thin and elongated  
720 nature, and distinguishes them from the wider and shorter TBR bars. These bars are  
721 ephemeral (residence time from one day to one month), attached to the low-tide shoreline  
722 or, occasionally, to the shore-parallel bar [*Konicki and Holman*, 2000; *Price and Ruessink*,  
723 2011]. *Ribas and Kroon* [2007] and *Ribas et al.* [2014] have shown that they are linked to  
724 the presence of obliquely incident waves that create a significant alongshore current and  
725 that they are up-current oriented.

726 **Type (3): Low energy finger bars.** These transverse bars (Figure 14c) are persistent  
727 features in fetch-limited beaches without a shore-parallel bar [*Falqués*, 1989; *Bruner and*  
728 *Smosna*, 1989; *Eliot et al.*, 2006; *Pellón et al.*, 2014]. Only *Bruner and Smosna* [1989]  
729 and *Pellón et al.* [2014] gave information concerning both their orientation and the forcing  
730 direction. At the two sites, the bars were down-current oriented with respect to the  
731 alongshore current generated by the wind-waves.

732 **Type (4): Large scale finger bars.** These transverse bars (Figure 14a) are char-  
733 acterized by long cross-shore spans of O(1 km) and develop across both the surf and  
734 the *shoaling zone*. They are generally observed to be persistent features in low-energy  
735 microtidal environments [*Niederoda and Tanner*, 1970; *Gelfenbaum and Brooks*, 2003],  
736 typically oriented almost perpendicular to the shore. Although their dynamics is less

737 understood, the wave focusing caused by refraction of normal incident waves by the bars  
738 seems to be essential [*Niederoda and Tanner, 1970*]. The recent study of *Levoy et al.*  
739 [2013] describes bars with similar cross-shore spans, but in a macrotidal medium-energy  
740 environment. Consequently, such bars can be governed by different drivers and will not  
741 be dealt specifically in the present study.

742 Table 4 shows an alternative classification of transverse bars based on their orientation,  
743 an important property that will turn out to depend critically on the DASC profile. The  
744 orientation down-current or up-current is sometimes difficult to differentiate in the field  
745 as this require the identification of the main forcing, so only the sites where the latter has  
746 been identified are included in Table 4. For this, a forcing analysis must be performed if the  
747 incoming waves have two dominant directions or in the presence of tidal currents [*Pellón*  
748 *et al., 2014*]. The slope of the part of the beach where the bars appear is also indicated  
749 in Table 4. The shore-normal large-scale finger bars appear on flat terraces (e.g., slope of  
750 0.003). The beach profiles below the shore-normal TBR bars and the down-current bars  
751 are similar: gentle-sloping upper (or low-tide) terraces. Up-current bars appear for larger  
752 beach slopes (0.02-0.04) in the subtidal zone [*Ribas et al., 2014*].

## 5.2. Existing theories for their formation

753 As occurred for the case of crescentic bars (section 4.2), during the 80's and the 90's the  
754 formation of rhythmic patches of transverse bars was commonly conceived to be caused  
755 by hydrodynamic template models, in which rhythmic morphologic patterns are forced  
756 solely by edge waves [e.g., *Holman and Bowen, 1982*]. However, as discussed by *Coco and*  
757 *Murray [2007]*, such theory is hardly consistent with observations by a number of reasons,  
758 the most outstanding being that the template theory neglect the (strong) interactions

759 between the hydrodynamics and the evolving bed level. In addition, in case of oblique  
760 wave incidence, the edge waves are progressive and they would cause a nonstationary flow  
761 pattern that moves much faster than the transverse bars migrating downdrift.

762 During the last two decades other hypotheses have been preferentially adopted. A  
763 first distinction has to be made between the TBR bars, which form from the welding  
764 to the shore of a previous crescentic bar [*Ranasinghe et al.*, 2004; *Garnier et al.*, 2008],  
765 and the finger bars, which grow from alongshore uniform conditions. The hypothesis  
766 that will be here adopted for the formation of transverse finger bars is that the feedback  
767 between components of the fluid/topography system can lead to their development (self-  
768 organization hypothesis, first proposed by *Sonu* [1968]). Some of the initial studies in  
769 this line of thought [*Barcilon and Lau*, 1973; *Hino*, 1974; *Christensen et al.*, 1994; *Falqués*  
770 *et al.*, 1996] had important shortcomings but were certainly pioneering and distinguished  
771 between the bed-flow instability (term introduced by *Falqués et al.* [1996] to refer to  
772 the positive feedback between the sea bed and an alongshore current) and the bed-surf  
773 instability (positive feedback between the bed and the breaking waves, already described in  
774 section 4.3). The subsequent studies [*Caballeria et al.*, 2002; *Ribas et al.*, 2003; *Klein and*  
775 *Schuttelaars*, 2005; *van Leeuwen et al.*, 2006; *Garnier et al.*, 2006; *Ribas et al.*, 2012] have  
776 been more satisfactory: shore-normal, up-current and down-current oriented bars with  
777 realistic spacings have been obtained and the self-organization mechanisms underlying  
778 transverse bar formation and the role of DASC have been explained in more detail. The  
779 knowledge gained in these studies is discussed in the next section.

### 5.3. Role of DASC in the formation mechanism

780 As explained in section 3, the joint effect of the gradients in DASC and the horizontal  
 781 circulation induced by the growing feature creates the erosion/deposition patterns that  
 782 explain why the feature grows. The case of transverse bars is more complicated than  
 783 that of crescentic bars (section 4.3) for two reasons. Firstly, there are different types of  
 784 transverse bars with distinct orientations and growing under different beach conditions  
 785 (Table 4). Secondly, some of them develop with the presence of an alongshore current,  
 786  $V_0$ . The three steps of the methodology (section 3.5) to explain pattern formation from  
 787 the first RHS term of equation (10) are taken in the three following subsections. At the  
 788 end, the role of the second RHS term of that equation (important if  $V_0 \neq 0$ ) is discussed.

### 789 **5.3.1. Depth-averaged sediment concentration profile**

790 The sediment concentration profiles, corresponding to the beach conditions in the dif-  
 791 ferent types of transverse bars, are here described based on simple physical arguments,  
 792 similar as in section 4.3.1. Shore-normal and down-current oriented bars typically emerge  
 793 in terraced profiles with gentle slopes under normal and oblique waves (section 5.1). Waves  
 794 dissipate their energy slowly across a wide saturated surf zone (Figure 15a), with the wave  
 795 orbital velocity amplitude decreasing onshore across the surf zone. In the case of oblique  
 796 wave incidence, an alongshore current is also generated, which typically has a maximum  
 797 somewhere in the middle of the surf zone. Under such conditions, the combined action  
 798 of the wave orbital velocities, the depth-averaged current (and the turbulent vortices, if  
 799 included) will produce a DASC profile,  $\mathcal{C}(x)$ , that has a maximum somewhere in the outer  
 800 part of the surf zone. Thereby, across the terrace there is an offshore-directed gradient  
 801 of  $\mathcal{C}$ . This behavior is reproduced by all the formulas given in Table 2. As an exam-  
 802 ple, the third panel of Figure 15a shows the DASC profile obtained with the Soulsby-van

803 Rijn formula [*Soulsby*, 1997], extended to include an extra sediment stirring produced by  
804 turbulent vortices [*Reniers et al.*, 2004; *Ribas et al.*, 2011].

805 On the other hand, up-current oriented bars occur in the steepest parts of profiles with  
806 shore-parallel bars (section 5.1), either in the inner surf zone or in the seaward side of  
807 the bar. In such situation, incident waves shoal before the crest of the shore-parallel bar  
808 (thereby increasing the orbital velocity amplitude), break over the bar, then reform over  
809 the trough and finally break again in the inner surf zone (Figure 15b). The  $\mathcal{C}$  profile  
810 across the shore-parallel bar, with a local maximum slightly seaward of the crest, has  
811 already been discussed in section 4.3.1. Somewhere in the inner surf zone, a second local  
812 maximum in  $\mathcal{C}(x)$  is also obtained, related with the second breaker zone. The type of  
813 breaking occurring there and the fact that waves dissipate their remaining energy in a  
814 relatively narrow area, with strong breaking-induced turbulent vortices, can make that the  
815 latter contribute significantly to the sediment resuspension (third panel of Figure 15b).  
816 Such process can increase significantly the DASC across the inner surf zone. Also, the  
817 second local maximum in the alongshore current profile in such relatively steep inner surf  
818 zones can be quite close to the shoreline. For all these reasons, the second local maximum  
819 in  $\mathcal{C}(x)$  is found very close to the shoreline and there is an onshore-directed gradient of  $\mathcal{C}$   
820 across the inner surf zone (Figure 15b). No experimental validation of the DASC profile  
821 in such complex natural surf zones is presently available.

### 822 **5.3.2. Horizontal flow pattern over transverse bars**

823 The horizontal circulation that occurs over incipient transverse bars depends critically  
824 on the orientation of the bars (blue streamlines in Figure 16). Such circulation is well  
825 established for the TBR bars and it is the same type of rip current circulation occurring

826 over crescentic bars (discussed in section 4.3.2). Rip current flow seaward in the troughs  
827 between bars (either shore-normal or down-current oriented) and onshore currents are  
828 observed over the bars [*Wright and Short*, 1984; *MacMahan et al.*, 2006; *Dalrymple et al.*,  
829 2011] (Figure 16a). For the case of large-scale finger bars and low-energy finger bars, ob-  
830 servations of the induced currents are scarce but they indicate the same type of circulation  
831 as for the TBR bars. An interesting experiment in a laboratory wave basin was made by  
832 *Niederoda and Tanner* [1970]. On a shore-normal (short-crested) finger bar, an onshore  
833 current was measured over the bar crest, which diverged close to the beach to flow in the  
834 seaward direction through the troughs. An onshore-directed current over the crest of a  
835 low-energy finger bar (with a shore-oblique orientation) was also observed in the field by  
836 *Falqués* [1989].

837 The physical processes driving the hydrodynamic circulation over approximately shore-  
838 normal transverse bars can be qualitatively explained from wave-induced forces. Focusing  
839 of wave energy due to refraction and wave breaking is enhanced over transverse bars and  
840 this creates onshore directed currents (model studies that support this explanation are  
841 *Caballeria et al.* [2002] and *van Leeuwen et al.* [2006]). Such currents are forced to diverge  
842 near the shoreline into two alongshore parallel feeder currents that converge in the trough  
843 and flow seaward as a rip current (similar to the case of crescentic bars, see Appendix D).

844 In the case of shore-oblique finger bars, which always coexist with a significant along-  
845 shore current (Table 4), other hydrodynamic processes that induce a meandering of the  
846 alongshore current can be more important. Due to frictional forces and mass conservation,  
847 the current experiences a seaward deflection over up-current oriented bars and a shore-  
848 ward deflection over the up-current troughs (Figure 16c), as explained in more detail in

Appendix F. No observations of such current circulation induced by up-current oriented finger bars in open beaches are available but model confirmation was given by *Ribas et al.* [2003], *Garnier et al.* [2006], and *Ribas et al.* [2012]. Such circulation (current deflection over up-current bars) opposes the one due to wave-induced forces and is only dominant for obliquely incident waves. In the case of down-current oriented bars, the alongshore current experiences the opposite deflection, veering towards the shore over the crests and towards the sea over the troughs (Figure 16b), so that the corresponding current perturbations are reinforced by those created by wave-induced forces.

### 5.3.3. Formation mechanism and transverse bar orientation

The cross-shore profile of the DASC plays a crucial role in explaining the orientation of the growing transverse bars. Indeed, according to the first RHS term of the linearized BEE (10), for seaward increasing  $\mathcal{C}$  ( $d\mathcal{C}_0/dx > 0$ ) a shoreward current perturbation ( $u < 0$ ) causes sediment deposition and a seaward current perturbation ( $u > 0$ ) causes erosion. Since this is the type of flow occurring on the crests and troughs, respectively, of shore-normal or down-current oriented bars, (Figure 16a,b), a positive feedback between flow and morphology occurs making the bar system grow. Note that shore-normal or down-current oriented bars are observed on terraced planar beaches, where  $d\mathcal{C}_0/dx > 0$  across the terrace (Figure 15a). In other words, shore-normal/down-current bars are formed because the onshore-directed flows over their crests carry sediment from offshore, where  $\mathcal{C}$  is largest, to the crests (see also Figure 8). This formation mechanism is similar to that of crescentic bars (discussed in section 4.3.3). Such a growth mechanism can be dominant for shore-normal waves and oblique waves because both the meandering of the alongshore current over down-current bars and the wave-induced forces create an onshore



872 current perturbation over the crests. Notice that the origin of TBR bars is the merging  
 873 of a pre-existing crescentic bar into the beach, i.e., they do not grow from an alongshore  
 874 uniform planar beach. However, the mechanism described in this paragraph, based on  
 875 the DASC profile, explains why such TBR bars can maintain their shape without being  
 876 destroyed, being the most frequently occurring beach state in some beaches (e.g., 55% in  
 877 Palm beach, Australia, with a residence time of some 20 days [*Ranasinghe et al.*, 2004]).

878 In contrast, for shoreward increasing  $\mathcal{C}$  ( $d\mathcal{C}_0/dx < 0$ ) a seaward ( $u > 0$ ) current pertur-  
 879 bation causes sediment deposition and a shoreward current perturbation ( $u < 0$ ) causes  
 880 erosion. This is the type of flow occurring on the crests and troughs, respectively, of  
 881 up-current oriented bars (Figure 16c). A positive feedback therefore takes place and the  
 882 bars grow. Note that up-current oriented bars are observed in steep inner surf zones and  
 883 seaward slopes of shore-parallel bars, where  $d\mathcal{C}_0/dx < 0$  (Figure 15b). This mechanism  
 884 only works if the angle of wave incidence is large. If waves are less oblique, the meander-  
 885 ing of the alongshore current that creates a positive  $u$  over the up-current oriented crests  
 886 become less effective whilst the wave-induced forces (onshore-directed over the crests)  
 887 become more effective, inhibiting bar growth.

888 Whilst the role of the first RHS term of equation (10) is mainly related to the growth  
 889 or decay of the bars, the second RHS term of that equation turns out to be mainly re-  
 890 lated to the migration of the bars [*Garnier et al.*, 2006; *Ribas et al.*, 2012]. Thereby,  
 891 analyzing transverse bar migration is more complicated because it depends on the along-  
 892 shore gradients of the perturbations of the DASC. The migration direction depends on  
 893 the alongshore phase shift between the bathymetry and the perturbation of the depth-  
 894 averaged concentration,  $c$ . If the maximum of  $c$  is located around the crests of the bars

895 this term will produce pure downdrift migration (like in Figure 9c). The picture is even  
896 more complicated because, often, the first RHS term of equation (10) not only explains  
897 the growth but also adds to the migration (in case of an alongshore phase shift between  
898 the maximum in  $u$  and the maximum in  $h$ ).

#### 5.4. Finite amplitude behavior

899 *Garnier et al.* [2006] reproduced, for the first time, the saturation of growth of trans-  
900 verse bars (with all possible orientations, see Figure 17). The overall characteristics of  
901 finite-amplitude bars were similar to those of the initially growing bars, only differences  
902 up to a factor of 2 occurred in spacings and migration rates. The shape of finite amplitude  
903 bars included typical nonlinear characteristics like the asymmetry of the alongshore shape,  
904 as observed in the field [*Pellón et al.*, 2014], and the asymmetry between offshore flow (rip  
905 current) and onshore flow, in accordance to observed rip current systems [*Short*, 1999].  
906 Other nonlinear phenomena like merging of individual bars, and oscillatory behavior (dy-  
907 namic equilibrium) was also reproduced. *Garnier et al.* [2006] also made for the first time  
908 a quantitative global analysis (see Appendix C) to understand the physical reasons for  
909 the saturation of transverse bar growth. Essentially, two possible different scenarios were  
910 found for the saturation: (i) the damping term, related with the downslope gravitational  
911 transport (the second RHS term in equation (8)), strengthens so that it eventually bal-  
912 ances the instability source or (ii) the production term, related with the instability due to  
913 the gradients in DASC (the first RHS term in equation (8)), weakens so that it becomes  
914 balanced by the damping term. This means that saturation can occur, depending on the  
915 type of transverse bars, either because the finite-amplitude shape of the bars enhances  
916 downslope transport (i) or because it weakens the instability mechanism (ii).

## 5.5. Discussion

917 An important aspect that deserves discussion is that in the modeling studies on trans-  
 918 verse bar formation that included the sediment stirring by turbulent vortices [e.g., *Ribas*  
 919 *et al.*, 2012], the perturbations of the sediment load  $\alpha$  (equation 3) were neglected. Numer-  
 920 ical experiments in which these perturbations were maintained resulted in an output that  
 921 was highly sensitive to numerical parameters (number of grid points and their distribution  
 922 over the computational domain). This model behavior was in line with that of models for  
 923 the initial formation of shoreface-connected sand ridges (section 6.5) when perturbations  
 924 in sediment stirring by waves were included [e.g., *Vis-Star et al.*, 2007]. Given that these  
 925 numerical instabilities were not standard (e.g., due to a too large timestep), this suggests  
 926 that the presently available sediment transport formulations are not yet sufficiently accu-  
 927 rate to correctly describe spatial variations in  $\alpha$  due to the bars. As a consequence, the  
 928 role of the perturbations of  $\alpha$  into the second RHS term of equation (10) is unknown to a  
 929 great extent. Some insight has been provided by *Thiebot et al.* [2012], who modeled the  
 930 nonlinear development of rhythmic surf zone bars, including the perturbations of  $\alpha$ , in  
 931 a beach with two shore-parallel bars. In the case of oblique waves with a large angle of  
 932 incidence, they reproduced the formation of down-current oriented bars at the inner surf  
 933 zone (where  $d\mathcal{C}_0/dx > 0$ , so in agreement with the theory presented here). However, they  
 934 showed that the second RHS term of equation (10) also contributed significantly to bar  
 935 growth.

## 6. SHOREFACE-CONNECTED SAND RIDGES

### 6.1. Characteristics of observed shoreface-connected sand ridges

936 These large-scale bed forms (horizontal extents of several km) are observed on inner  
937 continental shelves (depths of 5-80 m) with a sandy bottom and where storms frequently  
938 occur (Figure 18). Inner shelves are characterized by a transverse bottom slope of about  
939 1 m/km, i.e., substantially smaller than that in the surf zone, but much larger than that  
940 on the outer shelf. Shoreface-connected sand ridges, hereafter called 'ridges', occur on  
941 both meso-tidal and micro-tidal shelves and manifest themselves in patches. They are  
942 for example present on shelves along the east coast of the United States [*Duane et al.*,  
943 1972; *Swift et al.*, 1985], Argentina [*Parker et al.*, 1982], Germany [*Antia*, 1996] and the  
944 Netherlands [*van de Meene and van Rijn*, 2000].

945 The ridges have an along-shelf spacing between successive crests that varies from 1 km  
946 to 8 km. They have asymmetrical profiles, with their steepest slope on the landward  
947 sides, where their sediment is relatively coarse. Heights of the ridges are in the range  
948 1-12 m and they migrate in the direction of the storm-driven currents with velocities of  
949 1-10 m yr<sup>-1</sup>. Interestingly, their crests are persistently up-current oriented with respect to  
950 the wind-driven alongshore current that occurs during storms. *Swift et al.* [1978] already  
951 pointed out that these facts suggests that the ridges evolve during storms, when high  
952 waves and intense storm-driven currents cause abundant erosion and transport of sand.  
953 In contrast, during fair weather conditions the ridges would be inactive, because bottom  
954 shear-stresses do not exceed the critical stress for erosion of sand. Some ridges turn out  
955 to be moribund features, i.e., they are no longer active under present-day hydrodynamic  
956 conditions [*Goff et al.*, 1999].

957 Other large-scale bed forms that occur on continental shelves are tidal sand ridges.  
958 Although their dimensions are similar to those of shoreface-connected sand ridges, they

959 appear on outer shelves and when tidal currents are stronger than  $0.5 \text{ m s}^{-1}$ . Furthermore,  
960 the orientation of tidal sand ridges is related to tidal currents, in the sense that they are  
961 rotated cyclonically with respect to the direction of the dominant tidal current [see *Dyer*  
962 *and Huntley*, 1999, and references herein]. The focus of this section is on shoreface-  
963 connected sand ridges, as their formation is related to gradients in the DASC.

## 6.2. Existing theories for their formation

964 A number of theories have been suggested to explain the origin of shoreface-connected  
965 sand ridges. In early studies it was argued that they could be relict features from before  
966 the Holocene transgression [e.g., *Swift et al.*, 1972], or that they evolved from relict fea-  
967 tures [e.g., *McBride and Moslow*, 1991], such as former dunes and ebb-tidal deltas, which  
968 were flooded due to the rising sea level and subsequently being reworked by waves and  
969 currents. Indeed, observations have shown that sediment transport during storm condi-  
970 tions is significant in ridge areas [*Swift et al.*, 1978]. For example, the ridges along the  
971 east coast of the US and on Dutch inner shelf started to form several thousands years ago  
972 and they are active under the present hydrodynamic conditions [*Swift et al.*, 1978; *Duane*  
973 *et al.*, 1972; *van de Meene and van Rijn*, 2000].

974 Although there is nowadays ample evidence that many ridges are not relict features,  
975 there is less consensus about the dominant physical processes that control their evolu-  
976 tion. *Swift et al.* [1978] suggested that helical circulation cells in the vertical plane might  
977 transport sand from troughs to crests, thereby resulting in a positive feedback. In this  
978 study it was also stated that tides would not be a primary forcing agent of the ridges, as  
979 the latter occur on both micro-tidal and meso-tidal inners shelves. *Niedoroda et al.* [1985]  
980 argued that during storms the ridges receive sand from the nearshore zone, because the

981 wind creates downwelling conditions leading to an offshore-directed (Ekman) flow near  
982 the bottom. This sand would be subsequently reworked by waves and currents.

983 A limitation of the studies cited above is that they did not demonstrate how ridges  
984 would form and what mechanism explains their characteristics. In this regard, a major  
985 step forward was made by *Trowbridge* [1995], who demonstrated, by means of a linear  
986 stability analysis, that formation of bed forms resembling ridges can be simulated with  
987 a model that governs interactions between storm-driven currents and the sandy bed on  
988 an idealized inner shelf with a sloping bottom (Figure 4). His model assumed bed load  
989 sediment transport to be a constant times the current velocity. The constant reflects the  
990 spatially uniform stirring of sediment by waves. Subsequent investigations, based on linear  
991 stability analyses, have shown that the growth of the ridges is mainly caused by suspended  
992 load transport [*Calvete et al.*, 2001] and that the effect of wave shoaling and refraction,  
993 resulting in spatially non-uniform stirring of sediment, is to enhance their growth [*Vis-*  
994 *Star et al.*, 2007]. The effect of tides on the initial formation of the ridges, subject to  
995 both bed load and suspended load transport, was examined by *Walgreen et al.* [2002].  
996 They showed that tidal currents only mildly affected the shoreface-connected sand ridges  
997 and merely resulted in tidal sand ridges on the outer shelf. These findings supported  
998 the earlier hypothesis of *Swift et al.* [1978]. When accounting for different grain sizes  
999 [*Walgreen et al.*, 2003], the model was capable of explaining the observed distribution  
1000 of mean grain size over the sand ridges (coarser sediment on the landward side of the  
1001 crests). Here, the concepts of *Trowbridge* [1995] and follow-up studies will be discussed  
1002 in more detail, because they highlight again the importance of gradients in the DASC

1003 for the dynamics of these bed forms, whilst at the same time explaining many aspects of  
1004 observed ridges.

### 6.3. Role of DASC in the formation mechanism

1005 The role of the DASC in the physical mechanism underlying the initial growth and  
1006 orientation of shoreface-connected sand ridges can be understood following the three steps  
1007 of the methodology (section 3.5). This is done in the next subsections, emphasizing the  
1008 details required to understand ridge formation. Migration is also discussed at the end.

#### 6.3.1. Depth-averaged sediment concentration profile

1010 Consider the situation during storms, as shown in Figure 19, when the sediment is  
1011 stirred by the waves and a storm-driven current (typically of order  $0.5 \text{ m s}^{-1}$ ) flows along  
1012 the coast. Under these conditions, the sediment transport is proportional to the sediment  
1013 load  $\alpha$  and the current as described by equation (3). The sediment load will decrease in  
1014 the offshore direction, because waves will be less efficient in stirring sand from the bottom,  
1015 and the DASC ( $\mathcal{C} = \alpha/D$ ) will decrease in the offshore direction as well.

#### 6.3.2. Horizontal flow pattern over ridges

1017 When the storm-driven current encounters an up-current oriented ridge, conservation of  
1018 water mass will force the flow component perpendicular to the ridge to increase, causing  
1019 an offshore deflection of the current over the ridges. Likewise, the current will have an  
1020 onshore component in the troughs. The result is a meandering storm-driven flow. This  
1021 mechanism also acts in the case of up-current transverse finger bars and it was described  
1022 in section 5.3.2 and, in more detail, in Appendix F.

#### 6.3.3. Formation mechanism

1024 From the considerations above, it immediately follows that in the ridge areas the gradi-  
 1025 ent of the DASC has a negative projection on the current  $\vec{v}$  (i.e.,  $\vec{v} \cdot \vec{\nabla} \mathcal{C} < 0$ , Figure 19).  
 1026 Likewise,  $\vec{v} \cdot \vec{\nabla} \mathcal{C} > 0$  in the trough areas. Thus, according to the first RHS term of the  
 1027 BEE (8), a positive feedback mechanism occurs, leading to growth of the ridges.

1028 Note that the formation mechanism of the ridges resembles that of up-current oriented  
 1029 bars in the surf zone (section 5.3). The differences between the two features, besides their  
 1030 geomorphological characteristics (size, location and time-scales), concern the sediment  
 1031 stirring process and the mechanisms causing the offshore deflection of the current over  
 1032 the crests (see also Appendix F).

1033 In the models that describe the dynamics of sand ridges, bed load transport and sus-  
 1034 pended load transport play a different role in the formation of bottom patterns [*Calvete*  
 1035 *et al.*, 2001]. Suspended load transport is linked to a DASC that is proportional to  $u_b^3$ ,  
 1036 where  $u_b$  is the amplitude of near-bed wave orbital motion. This is because the models  
 1037 assume settling lag effects to be small, so the mass balance of suspended sediment reduces  
 1038 to an approximate balance between erosion of sediment (modeled as being proportional to  
 1039  $u_b^3$ ) and deposition of sediment (assumed proportional to DASC). This term is primarily  
 1040 responsible for the growth of the ridges and its divergence produces maximum deposition  
 1041 approximately at the crest of the ridges. Thereby it essentially leads to growth of the  
 1042 ridges without migration (like in Figure 9a). On the other hand, bed load transport is  
 1043 modeled as in equation (3), with depth-integrated sediment concentration  $\alpha$  (i.e., DASC  
 1044 times depth) being proportional to  $u_b^2$ . Its divergence produces maximum deposition  
 1045 between the crest and the subsequent trough, thus produces essentially pure downdrift  
 1046 migration of the feature (like in Figure 9c). Although bed load transport is considerably



1047 weaker than suspended load transport, it is this component that controls the downstream  
1048 migration of the ridges. The largest deposition of the total sediment transport takes place  
1049 slightly downstream of the crests, which produces their growth and downstream migration  
1050 (like in Figure 9b).

#### 6.4. Finite amplitude behavior

1051 Following the general theory of section 3, saturation of bed forms towards finite heights  
1052 must be due to either changes in the distribution of DASC, or from changes in the flow  
1053 over the bottom pattern, or from an increase of the diffusive term. Regarding the finite  
1054 amplitude behavior of ridges, this problem was studied by several authors [*Calvete et al.*,  
1055 2002; *Calvete and de Swart*, 2003; *Vis-Star et al.*, 2008; *de Swart et al.*, 2008]. They  
1056 derived a nonlinear model from projection of the equations of motion onto the adjoint  
1057 eigenmodes of the linearized system. The result are differential equations that govern  
1058 the time evolution amplitudes of the different bottom modes. Amplitudes of flow modes  
1059 follow from algebraic equations, as it is assumed that the flow adjusts instantaneously  
1060 to a new bed level. The results showed that, after an initial phase in which ridges grow  
1061 exponentially, they saturate and reach a finite height on timescales of several thousands of  
1062 years (Figure 20). The resulting profiles of the ridges are highly asymmetrical, with steep  
1063 stoss sides and mild lee sides, consisting with what is observed in the field. Moreover,  
1064 smaller-scale bed forms, with length scales of a few hundred meters, are superimposed on  
1065 the ridges. Note that these small scale bed forms have the size of sand waves. A detailed  
1066 analysis revealed that both the small-scale bed forms and the diffusive sediment transport  
1067 induced by bed slopes are responsible for the saturation of the ridges to a constant height.

1068 Using a more sophisticated wave model, *Vis-Star et al.* [2008] was able to demonstrate  
1069 patch behavior of the ridges.

1070 One of the limitations of nonlinear spectral models is that they do not allow for varia-  
1071 tions of mean sea level, because that would affect the spatial structure of the eigenfunc-  
1072 tions. On the other hand, field data clearly suggest that ridges are affected by sea-level  
1073 changes [*Swift et al.*, 1978]. These considerations have motivated the development of an  
1074 alternative nonlinear model for shoreface-connected sand ridges, which is based on finite-  
1075 difference techniques. A recent study by *Nnafie et al.* [2014a] shows that changes in mean  
1076 sea level can have a profound impact on the long-term evolution of the ridges. In partic-  
1077 ular, when sea level rise is too fast compared to typical deposition rates, the ridges drown  
1078 and become moribund features.

1079 In a separate study, *Nnafie et al.* [2014b] investigated the impact of extraction of sand  
1080 from fully developed ridges. Their main findings are that the intervened ridge partly  
1081 restores, on a time scale of centuries, albeit that its final volume of sand is smaller than its  
1082 volume prior to the intervention. The sand needed for filling the extraction pit originates  
1083 from different sources, such as the downstream trough, the part of the ridge upstream of  
1084 the pit and the outer shelf and nearshore zone.

## 6.5. Discussion

1085 There are a number of open issues with regard to further understanding of the dynamics  
1086 of shoreface-connected sand ridges. The first is that nonlinear spectral models had prob-  
1087 lems to simulate the ridges for a realistic value of the shelf slope (typically  $10^{-3}$  m/m).  
1088 Nevertheless, output of these models could be used to make educated estimates of expected  
1089 heights and saturation time scales (i.e., the time at which finite heights are reached) for

1090 realistic shelf slopes. This is because the models showed that, for the range of shelf slopes  
1091 that finite amplitude ridges were simulated, the height and saturation time scale depended  
1092 linearly on the shelf slope and the inverse of the shelf slope, respectively. Assuming these  
1093 relationships to hold for larger shelf slopes as well, modeled heights and saturation time  
1094 scales were extrapolated to realistic shelf slopes. It turned out [*Calvete and de Swart,*  
1095 *2003; Vis-Star et al., 2008*] that the results thus obtained agreed fairly well with field  
1096 data. An important breakthrough in this respect was achieved with a recently devel-  
1097 oped nonlinear finite difference code [*Nnafie et al., 2014b*], which is capable of simulating  
1098 ridges for realistic shelf slopes. Moreover, this model confirmed that the earlier applied  
1099 extrapolation method, as discussed above, was indeed correct.

1100 A second discussion point concerns the feedback between wind waves and ridges (as  
1101 occurs for transverse bars, see section 5.5). Their effect on the initial formation of ridges  
1102 was examined by *Lane and Restrepo* [2007] and *Vis-Star et al.* [2007]. Outcomes were  
1103 different: the former study revealed no growth of bed forms, whereas the latter study  
1104 showed that the growth of the ridges was significantly enhanced by perturbations in  
1105 wave stirring. The latter study already provided a physical reason for this enhanced  
1106 growth, and this finding was confirmed in a later study [*Nnafie et al., 2011*], in which  
1107 an independent, numerical morphodynamic model was used. These studies reveal that  
1108 allowing for directional spreading of waves seems a necessary condition to properly account  
1109 for these feedbacks in nonlinear models.

1110 A third interesting extension would be to study the potential interactions between  
1111 storm-driven sand ridges and other bed forms, such as tidal sand waves and megaripples.  
1112 As a first step, this could be done by improving formulations for bottom roughness that are

1113 related to smaller scale bed forms. Ultimately, such interactions should be studied with  
1114 (at least quasi) three-dimensional models, as tidal sand waves and megaripples are the  
1115 result of flow circulations that act in the vertical plane. This approach is quite challenging,  
1116 as since tidal sand wave/megaripples evolve on much shorter time scales than the ridges.  
1117 Thus simulations would require small grid sizes and small time steps.

## 7. BEACH CUSPS

### 7.1. Characteristics of observed beach cusps

1118 Beach cusps are alongshore rhythmic features of the swash zone (Figures 1d and 2e),  
1119 the region that is quasi-periodically covered and uncovered by successive waves. Beach  
1120 cusps consist of lunate embayments separated by relatively narrow shoals or horns, the  
1121 apices of which point seaward, see Figure 21. These horns and embayments, which are  
1122 ostensibly areas of deposition and erosion respectively, have sometimes been observed to  
1123 be accompanied by corresponding areas of, respectively, erosion and deposition further  
1124 seaward. Beach cusps typically have horn-to-horn distances, or spacings of 1 – 50 m,  
1125 and the spacing is proportional to incoming wave period and beach slope. The reader is  
1126 referred to *Coco et al.* [1999] for a comprehensive review of the main features of cusps,  
1127 their development and occurrence, in laboratory and field conditions. An excellent set of  
1128 images of cusp development is provided by *Almar et al.* [2008], in which their morphology  
1129 can clearly be seen.

1130 As described by *Coco et al.* [1999, 2000], cusps can occur on different beach slopes,  
1131 with different sediment sizes, and under different wave conditions. However, they are  
1132 predominantly features of steeper beaches, and most observations are on beach slopes of  
1133 between 0.08 to 0.16 [*Coco et al.*, 1999]. Also, typically sediments are relatively coarse

1134 and well-sorted, almost all observed cusped beach grain sizes are  $> 0.2$  mm, with modal  
1135 values being about 0.5 mm (medium to coarse sand [see, e.g., *Soulsby*, 1997]), but with  
1136 sometimes considerably larger sediment sizes [e.g., gravel, *Coco et al.*, 1999]. Frequently,  
1137 there is also evidence of sorting of grain sizes, with coarser sediments typically accumu-  
1138 lating on the horns, and finer ones in bays [*Coco et al.*, 1999]. Cusps are observed under  
1139 partially reflective wave conditions (i.e., where a significant proportion of the wave energy  
1140 is reflected back out to sea), for normal or near-normal wave incidence. Consistent with  
1141 the partially reflective nature of waves in cusp systems, it is most commonly observed  
1142 that for cusp formation waves break either by plunging or collapsing, in other words by  
1143 expending a lot of their energy in a narrow region near to the shore (the shorebreak), and  
1144 then running up (and back down) the beachface. They are rarely observed where spilling  
1145 breakers occur (which is consistent with the steeper slopes on which they are observed),  
1146 and also less commonly where there is no breaking.

1147 Field data is equivocal regarding whether beach cusps are erosive or accretionary fea-  
1148 tures, although more recent literature seems to point toward their being a combination of  
1149 the two [*Coco et al.*, 2004a; *van Gaalen et al.*, 2011]. What does seem clear now is that  
1150 individual cusps can merge [*Almar et al.*, 2008], forming larger local spacings. Further,  
1151 once formed, cusp systems can be removed both by erosion (i.e., storms) or by contin-  
1152 ued accretionary (i.e., low energy) conditions. This implies that cusps are by their nature  
1153 ephemeral features, which are likely to persist longest on a falling tide [*Coco et al.*, 2004a].

## 7.2. Existing theories for their formation

1154 More controversy surrounds the mechanism of beach cusp formation. The theory that  
1155 pertained predominantly, prior to about 1993, was that of edge waves scouring out the

1156 observed beach patterns [*Guza and Inman, 1975; Guza and Bowen, 1975*]. The amplitude  
1157 of the zero-mode edge waves achieves a maximum at the shore, decays exponentially  
1158 offshore and varies sinusoidally alongshore. In the edge wave theory for beach cusps, the  
1159 pattern is “carved” into the beach by the hydrodynamics. The hydrodynamic motion  
1160 is therefore often viewed as a *template*, which then imposes that pattern on the beach,  
1161 which is in contrast to all the earlier explanations of morphodynamic pattern formation  
1162 mentioned herein. *Coco et al. [1999]* have compared these edge wave theories (in the form  
1163 of observed cusp spacings and wave periods) to numerous historical data-sets and found  
1164 that there is reasonable agreement. On the other hand, *Coco et al. [1999]* also conclude  
1165 that observed wave breaking is predominantly plunging in character, and too dissipative  
1166 to account for the edge wave model being operative in at least 50% of data-sets. Indeed,  
1167 in some field studies edge waves were not present during the initiation of beach cusps  
1168 [*Masselink et al., 2004*].

1169 The main other explanation for beach cusps has its origins in morphodynamics, and  
1170 relies on a positive feedback between bed and water motions. *Werner and Fink [1993]* and  
1171 *Coco et al. [2000]* presented compelling arguments, based on simple models that utilize  
1172 Newtonian dynamics to simplify swash flow as a series of balls moving up and down  
1173 a beach that erode or accrete [see also *Coco et al., 2001*]. From an initial alongshore  
1174 uniform beach, cusps emerge as a larger-scale organized pattern of water motions and  
1175 bathymetry. The resulting cusp spacing showed a correlation with the swash excursion  
1176 (the distance measured along the beach from the base of the swash to the position at which  
1177 maximum run-up is achieved), in line with data [*Werner and Fink, 1993; Coco et al.,*  
1178 *2000*]. Significantly, this type of model was shown generally to agree with observations

1179 made within an experiment designed specifically to monitor beach cusp development from  
1180 an initially plane beach [*Coco et al.*, 2003].

1181 To date, the only other attempt to understand and describe cusp development has been  
1182 by *Dodd et al.* [2008], who formulated a fully coupled morphodynamic model in which  
1183 equations describing the fluid flow are coupled to bed change, thus (in theory) allowing  
1184 feedback processes, but also allowing true hydrodynamic motions. In other words, this  
1185 description allowed both edge wave and self-organization processes to be operational.  
1186 The model reproduced the formation of beach cusps with spacings consistent with field  
1187 observations. Notably, the introduction of infiltration, consistent with steeper beaches  
1188 with coarser sediments, promoted cusp development. It should be noted, however, that  
1189 the model of *Dodd et al.* [2008] did not include sediment settling lags (as explained in  
1190 Appendix A), an effect shown to be important in promoting deposition of suspended  
1191 sediment in the upper swash [*Pritchard and Hogg*, 2005]. This deposition occurs because  
1192 sediment that was entrained in the inner surf zone or lower swash soon begins to settle  
1193 out because the flow is decelerating for most of the uprush, but only finally comes out  
1194 of suspension in the upper swash. This sediment is not all re-entrained in the backwash  
1195 because of the smaller velocities in this region and the corresponding lag in entrainment.  
1196 The absence of this process probably overemphasizes the importance of infiltration in the  
1197 model.

1198 The edge wave theory was examined by conducting purely hydrodynamic experiments.  
1199 Significantly, cusp-like circulations did develop, but were more ephemeral, often evol-  
1200 ving to larger scales, but with some evidence that edge waves were indeed being excited.  
1201 Overall, it was concluded that edge waves might play a part in initiating cusp devel-

1202 opment, but that (a) these were part of an instability mechanism, and (b) an erodible  
1203 bed significantly enhanced this mechanism. In the next section, the role of the depth-  
1204 averaged concentration (DASC) in beach cusp formation through self-organization will be  
1205 described.

### 7.3. Role of DASC in the formation mechanism

1206 As mentioned in sections 2 and 3, beach cusps have unique characteristics (compared to  
1207 the other three morphodynamic features studied here), related to the fact that they occur  
1208 in the swash, where it is essential to retain the back and forth water movement driven  
1209 by successive waves (i.e., non-steady hydrodynamic conditions). This is because there  
1210 is no obvious wave average to be calculated in this region [see *Brocchini and Peregrine,*  
1211 *1996*]. Further, waves in the swash are highly nonlinear, even when non-breaking, so  
1212 parameterizing their effect in a satisfactory way in a time-averaged description is difficult.  
1213 Furthermore, and more fundamentally, because of the high Froude number flows that can  
1214 occur in the swash, the hydro- and morphodynamic time scales are no longer necessarily  
1215 distinct from each other, so the quasi-steady approximation cannot formally be applied.  
1216 So, the equations and variables are not wave-averaged and the quasi-steady hypothesis  
1217 does not hold. Furthermore, the BEE must be solved during the wave cycle, because  
1218 of the aforementioned non-separation in time scales, and because the differences in the  
1219 sediment transport during *uprush* and *backwash* processes are what lead to gradients, over  
1220 a wave cycle, in the depth-averaged sediment concentration. Note also that this does not  
1221 mean that a wave-averaged picture of the morphodynamics is not useful, just that the  
1222 wave-averaging is best done after the modeling in order to reveal the dynamics [see *Dodd*  
1223 *et al., 2008*].



### 1224 **7.3.1. Depth-averaged sediment concentration during a wave cycle**

1225 In the non-steady swash conditions, it is instructive to first see how the DASC,  $\mathcal{C}$ , varies  
 1226 during a swash cycle on an alongshore uniform beach (Figure 22). At the shore-break  
 1227 the wave collapses (or plunges) onto the beach and the water rushes up the beachface  
 1228 (the *uprush*). The uprush is initially of very high velocity ( $O(2 \text{ m/s})$ ) and sometimes  
 1229 initially supercritical. Thus, at the tip of the uprush, where depths are very small,  $\mathcal{C}$   
 1230 achieves a maximum (recall that  $\mathcal{C}$  responds immediately to the flow) and (at the same  
 1231 instant in time) decays, but is still significant, seaward (Figure 22a). Thus sediment is  
 1232 set in motion at the start of the uprush, thus eroding the beach in the lower swash.  
 1233 As the uprush diminishes so does  $\mathcal{C}$ . Therefore, deposition pertains over the rest of the  
 1234 uprush. The flow reverses first in the lower swash, and the *backwash* develops, which  
 1235 soon encompasses the whole swash region as gravity accelerates the flow seaward, thus  
 1236 increasing  $\mathcal{C}$ , which can then achieve values comparable with those in the uprush late  
 1237 in the backwash (Figure 22b). Therefore, sediment is once more mobilized and this time  
 1238 transported offshore. The backwash therefore erodes the upper and mid-swash, eventually  
 1239 depositing sediment in the lower swash. The net sediment transport depends on the  
 1240 balance of these two processes (see *Masselink and Kroon* [2006], and also *Dodd et al.*  
 1241 [2008] for simulation of this process).

### 1242 **7.3.2. Non-steady flow on a cusp system**

1243 In an incipient cusped system, the horns (slightly elevated areas of beach level) are  
 1244 separated by embayments (slightly lowered areas of beach level), along with the corre-  
 1245 sponding regions of lowered and elevated level further offshore. If we now consider such a  
 1246 morphology on a non-erodible beach we can see the effect on the circulation. The purely

one-dimensional motion described in section 7.3.1 becomes two dimensional. At horns, uprush is diverted to either side of the horn (because of the shape of the horn), into the adjacent embayments. The backwash then occurs, with significantly more water at the embayments. Additionally, there is a phase lag that accompanies this circulation. As a normally incident and initially plane wave front approaches the beach it is first affected by the seaward regions of lowered (just seaward of horns) and elevated (just seaward of embayments) level, with the result that just offshore of the horns waves propagate faster (because the bed level is lower and the wave speed is proportional to  $\sqrt{gD}$  [Svendsen, 2006]), and so encounter the shoreline (and break) a little ahead of the wave front at embayment locations. At the horn location the uprush therefore occurs slightly ahead of that at embayments, exacerbating the effect of the flow divergence at the horn because of the absence of water at the embayments in the early uprush. Also, the relatively weak backwash at horns finishes ahead of that at embayments, because of the earlier uprush and the larger beach slope at horn locations, which also means that the next wave is relatively unaffected by the preceding backwash there. However, as the embayment backwash ends, it frequently interacts with (and therefore further delays) the next incident wave at that location. The result is that the horn locations are uprush dominated while the embayments are backwash dominated. What emerges is an overall (wave-averaged) circulation pattern as depicted in Figure 21.

### 7.3.3. Formation mechanism

In order to understand the role of DASC in creating the erosion/deposition patterns that explain cusp formation, the bed evolution equation (BEE) during a wave cycle must be analyzed. As pointed out in section 3.1, in swash zone morphodynamics the BEE

1270 takes the form of equation (7) because of its second RHS term can no longer be assumed  
 1271 negligible as the quasi-steady hypothesis does not hold [Dodd *et al.*, 2008]. Here, we  
 1272 rewrite equation (7) [following Dodd *et al.*, 2008] to include a (positive) vertical infiltration  
 1273 velocity  $w$  (which also requires an alteration to equation (4), as mentioned in section 2.2),

$$1274 \quad (1 - p) \frac{\partial h}{\partial t} = -D\vec{v} \cdot \vec{\nabla} \mathcal{C} - \mathcal{C} \vec{\nabla} \cdot (D\vec{v}) + \mathcal{C}w + \vec{\nabla} \cdot (\gamma \vec{\nabla} h) . \quad (11)$$

1275 Note also that, although  $\mathcal{C}$  is still a depth-averaged sediment concentration,  $\alpha$  is now more  
 1276 appropriately seen as a bed mobility parameter rather than as a wave stirring function.  
 1277 The depositional effect of infiltration can clearly be seen in equation (11) (i.e., the third  
 1278 RHS term is always positive).

1279 The contribution to erosion or deposition of the first and second RHS terms during  
 1280 the uprush and backwash in the case of alongshore uniform conditions are depicted in  
 1281 Figure 23a. In the upper plot we see  $-\overline{\mathcal{C} \vec{\nabla} \cdot (D\vec{v})}$  and  $-\overline{D\vec{v} \cdot \vec{\nabla} \mathcal{C}}$  for uprush and backwash  
 1282 (note that the overbars here denote time averages over only those phases of the swash),  
 1283 and their net contribution during the whole wave cycle. The pattern of erosion/deposition  
 1284 produced by the sum of the net contributions of these two terms is shown in the middle  
 1285 plot, and the corresponding change in bed level is shown in the lower plot. During  
 1286 the uprush, the flow divergence term is positive,  $-\overline{\mathcal{C} \vec{\nabla} \cdot (D\vec{v})} > 0$ , because the flow is  
 1287 converging ( $\vec{\nabla} \cdot (D\vec{v}) < 0$ ), and therefore this term leads to deposition. Conversely, during  
 1288 the uprush the concentration gradient term is negative,  $-\overline{D\vec{v} \cdot \vec{\nabla} \mathcal{C}} < 0$ , because the flow is  
 1289 moving toward a region of high DASC, the tip of the swash. However, this erosive effect  
 1290 is mainly limited to the lower swash, where the stronger gradients in  $\mathcal{C}$  induce a removal  
 1291 of sediment just seaward of the horns. In the backwash these effects reverse, with the  
 1292 flow divergence term removing sediment in the upper swash (as  $\vec{\nabla} \cdot (D\vec{v}) > 0$ ), and the

1293 concentration gradient term depositing in the lower swash. The latter occurs because the  
 1294 fast (often supercritical) offshore flow meets relatively static water thus creating a very  
 1295 large DASC gradient effect such that offshore flow moves down this gradient, and thus  
 1296 deposits rapidly. Note that on an alongshore uniform beach these effects will often, to  
 1297 a first order of approximation, be in balance, i.e., the two effects cancel each other out  
 1298 (solid lines in Figure 23a, top panel). In such a situation, infiltration (or lack thereof) can  
 1299 tip the balance in favor of deposition (erosion).

1300 In the presence of incipient cusps, the balance described in the paragraph above is  
 1301 broken. The flow during the uprush remains relatively unchanged but, during the back-  
 1302 wash, it is diminished in the horns and enlarged at the embayments (see section 7.3.2).  
 1303 Thereby, in the horns, the net erosion/deposition is dominated by the uprush processes  
 1304 whilst in the embayments the net change is dominated by the backwash processes (see  
 1305 Figure 23b,c). On the horns, the balance is shifted to deposition in the mid- and upper  
 1306 swash via flow convergence ( $-\overline{\mathcal{C} \vec{\nabla} \cdot (D\vec{v})} > 0$ ), and erosion in the lower swash via DASC  
 1307 gradient ( $-\overline{D\vec{v} \cdot \vec{\nabla} \mathcal{C}} < 0$ ). Therefore, an incipient horn grows further, accompanied by  
 1308 eroded areas just seaward, and thus there is positive feedback. At incipient embayments,  
 1309 conversely, the erosive effect of the increased divergent flow ( $-\overline{\mathcal{C} \vec{\nabla} \cdot (D\vec{v})}$ ) in the upper and  
 1310 mid swash, and the accompanying depositional effect of the DASC gradient ( $-\overline{D\vec{v} \cdot \vec{\nabla} \mathcal{C}}$ )  
 1311 in the lower swash predominate. Thereby, the embayments are further eroded, accompa-  
 1312 nied by regions of deposition just seaward, again leading to positive feedback. Moreover,  
 1313 any infiltration effects will mean that some uprush does not return as surface flow, thus  
 1314 further enhancing this effect.

#### 7.4. Finite amplitude behavior

1315 The formation mechanism described in the previous section is operational when the  
1316 cusps are of small amplitude. Later, limiting effects come into play. Eventually, the horns  
1317 may no longer experience run-up, and therefore become static, which in turn will tend  
1318 to equalize uprush and backwash at embayments, thus reducing differences in erosion  
1319 and deposition there. Similarly, exaggerated alongshore beach gradients (on the sides of  
1320 horns) may lead to local accelerations and erosion, and therefore the erosion of the flanks  
1321 of the horns. A model result showing the development of a large-amplitude cusp system  
1322 can be seen in Figure 24, in which we can see the bathymetry and the current in different  
1323 simulation times [Dodd *et al.*, 2008]. Sriariyawat [2009] undertook long-term simulations,  
1324 and applied the global analysis to the cusp system, and in doing so obtained a variety  
1325 of finite amplitude states, some apparently physical and some not. This points to the  
1326 difficulty of describing dynamic equilibria in a highly energetic moving boundary problem  
1327 (because small effects can tip the balance one way or another). There is therefore a need  
1328 for further work in this area.

#### 7.5. Discussion

1329 From a combination of field and laboratory work, and numerical studies, a reasonable  
1330 understanding of beach cusp formation has emerged. Nonetheless, some issues remain  
1331 unresolved as yet. The most obvious is what dictates the alongshore spacing of the cusps.  
1332 It is often difficult to pinpoint a clear physical reason for a length scale selection, even if  
1333 it can be shown mathematically or numerically that a certain length scale has the fastest  
1334 growth rate. As discussed by *Coco et al.* [1999], the field and laboratory data correlate  
1335 reasonably well with both edge wave length scales (both subharmonic and synchronous)

1336 and swash excursion, and that the latter is to be expected even if edge waves play a  
1337 part in the cusp development. This leads to the possibility that the self-organization  
1338 mechanism and the edge wave hypothesis are not mutually exclusive, even if they perhaps  
1339 they do not both pertain simultaneously. Interestingly, the work of *Dodd et al.* [2008] gives  
1340 some support for this, because, as was mentioned above, in their numerical experiments  
1341 they observed weak, cusp-like circulations develop on non-erodible beaches. It was clear,  
1342 however, that beach erodibility significantly enhances this mechanism.

1343 Note also that, as explained in section 7.3, the case of beach cusps is different from  
1344 those presented earlier. A wave-resolving approach is best taken here to represent the  
1345 dynamics of the feature. Thus, the current represented by  $\vec{v}$  simultaneously mobilizes  
1346 and transports the sediment. This is in contrast to the wave-averaged studies presented  
1347 heretofore, in which entrainment is produced by waves and current and the mobilized  
1348 sediment is transported by the current. Therefore, splitting sediment transport diver-  
1349 gence into depth-averaged concentration divergence and current divergence (as done in  
1350 section 3.1), makes less sense, because there is no separate mechanism (wave stirring)  
1351 for creating gradients in  $\mathcal{C}$ . Nonetheless, the interpretation embodied in equation (11)  
1352 can successfully be used to understand the feedback mechanisms. Finally, using equa-  
1353 tion (11) implies that the sediment transport responds (effectively) immediately to flow  
1354 changes, which is reasonable for transport by bedload or by suspension of coarse grains  
1355 (see Appendix A). However, if smaller grain sizes are suspended then a modified form of  
1356 equation (11) is necessary because sediment is not immediately deposited as flow deceler-  
1357 ates (settling lag). This can be an important effect in the swash, leading to deposition in  
1358 the upper swash [*Pritchard and Hogg, 2005*].

## 8. CONCLUDING REMARKS

1359 In this contribution a general formulation and methodology are presented to infer the  
1360 erosion and deposition patterns of sediment transport only from the gradients in the  
1361 depth-averaged sediment concentration (DASC) and the spatial structure of the current.  
1362 This can be applied whenever the sediment transport is equal to a sediment load times the  
1363 current, which means assuming that waves essentially stir the sediment and the current  
1364 augments this stirring and advects the sediment. Note that, in the application to the  
1365 swash zone, the stirring is due only to the current. The DASC is then defined as the  
1366 total sediment load divided by water depth. In this formulation the current is the depth-  
1367 averaged current and its dynamics must be described within the context of a time- and  
1368 depth-averaged shallow water model. The key point of the formulation is a bed evolution  
1369 equation (BEE) that describes the bed changes solely in terms of the advection of the  
1370 DASC by the current (section 3). This applies when the time scale on which the bed  
1371 evolves is much larger than the hydrodynamic time scales (i.e., when the quasi-steady ap-  
1372 proximation is applicable). When these time scales are comparable, the bed level changes  
1373 depend also on the DASC itself and the convergence/divergence of the water volume flux,  
1374 but the methodology can still be generalized. In the former, most common case, deposi-  
1375 tion (erosion) occurs where the current flows from areas of high to low (low to high) values  
1376 of DASC (section 3.2). Thus, analyzing the resulting erosion/deposition patterns leads to  
1377 an understanding of why different features with a specific pattern grow in different coastal  
1378 zones. In addition, the DASC, in combination with knowledge of the currents associated  
1379 with emerging patterns, and the BEE, provides insight into some important aspects of  
1380 the finite-amplitude behavior, such as the saturation of the growth of the features. In

1381 particular, the BEE can be integrated over a region of the coastal zone, and then provides  
1382 quantitative information about the mechanisms behind growth, saturation and migration  
1383 of morphologic features (global analysis). Note that this methodology is not a modeling  
1384 technique since it is not closed, the currents must be externally provided from models  
1385 (either hydrodynamic or morphodynamic), from observations or from physical reasoning.  
1386 Rather, it is a way of gaining physical understanding of why alongshore rhythmic patterns  
1387 of a certain shape grow and sometimes migrate or a way of making predictions of what  
1388 type of pattern will emerge.

1389 This methodology has proven to be a powerful tool with which to gain insight into the  
1390 feedback mechanisms between the morphology and the hydrodynamics, and so to explain  
1391 the formation of four morphologic features in the coastal zone that display alongshore  
1392 rhythmic patterns: crescentic and transverse surf zone bars, shoreface-connected ridges  
1393 and beach cusps. The key mechanism for the growth of crescentic bars can be under-  
1394 stood from a seaward increase of the DASC at the bar zone: above shoals (channels) the  
1395 DASC decreases (increases) along the onshore (offshore) directed current, causing deposi-  
1396 tion (erosion) (section 4.3). Similarly, down-current or shore-normal transverse bars, with  
1397 their onshore current perturbations on the crests, create a positive feedback in the case  
1398 of seaward increasing DASC (section 5.3). On the other hand, in the case of shoreward  
1399 increasing DASC the positive feedback occurs if transverse bars are up-current oriented  
1400 because this enhances the convergence of sediment transport in the seaward current per-  
1401 turbations that occur over the up-current crests. Similarly, on the inner continental shelf,  
1402 the combination of the DASC increasing onshore and the offshore (onshore) directed  
1403 currents over the up-current oriented ridges (troughs) causes deposition (erosion) (sec-



tion 6.3). The application to cusps in the swash zone is more complicated because the bed evolves at the same time scale as the free surface and the convergence/divergence of the flow caused by free surface changes cannot be ignored in the BEE (section 7.3). However, we can still apply these ideas to analyze these features. At an area of raised bed level (incipient horn) there is reduced backwash, because return flow is channeled into adjacent regions. There is therefore net deposition in the mid and upper swash (flow convergence), and net erosion in the lower swash (onshore directed DASC gradient during the uprush), both because the reduced backwash fails now to counteract these effects in the uprush. Related to this, in regions of relatively reduced bed level (incipient embayments) there is an excess of backwash, which leads to net erosion in the upper and mid swash (flow divergence), and deposition in the lower swash (onshore directed DASC gradient in the backwash).

For nearly normal wave incidence there is no significant alongshore current and the alongshore gradients in DASC do not affect the development of the features (as discussed in section 3.3). As a result, the formation of crescentic bars and shore-normal transverse bars (which grow for normal wave conditions) can be fully understood because their dynamics are controlled just by the cross-shore DASC profile and the cross-shore current perturbations (first RHS term in the linearized BEE (10)). For the features that develop when a significant alongshore current is present (e.g., shore-oblique transverse bars and shoreface-connected ridges), the second RHS term in the linearized BEE (10), related to the alongshore current and the alongshore gradients in DASC, also affects bed changes. Often, it causes only alongshore migration of the features and the growth/decay is still fully described by the first RHS term. However, for highly oblique waves in the surf

1427 zone, the second RHS term can contribute to the changes in amplitude and the first RHS  
1428 term can contribute to the migration [*Ribas et al.*, 2012; *Thiebot et al.*, 2012]. All this  
1429 makes the analysis more complicated as the second RHS term, which is related to the  
1430 perturbations produced in DASC by the growing features, is difficult to model given our  
1431 limited knowledge of the sediment transport processes.

1432 The feedback mechanisms involved in the formation and subsequent dynamics of the  
1433 morphodynamic patterns addressed here have been confirmed with a number of morpho-  
1434 dynamic models [*Garnier et al.*, 2008; *Dodd et al.*, 2008; *Ribas et al.*, 2012; *Nnafie et al.*,  
1435 2014a, and references therein]. Some of these models have been calibrated against field  
1436 data and shown to give reliable predictions for other situations in beach morphodynamics  
1437 [e.g., *Reniers et al.*, 2004; *Smit et al.*, 2008; *Castelle et al.*, 2010b]. This gives ample sup-  
1438 port for the results from the DASC formulation presented here. However, the models are  
1439 often quite sensitive to the parameterization used for sediment transport in case of oblique  
1440 wave incidence in the surf zone [e.g., *Klein and Schuttelaars*, 2005]. The results of our  
1441 contribution reveal the following reasons for such modeling problems. Firstly, differences  
1442 in DASC profiles can produce completely different patterns. This is especially clear in the  
1443 case of transverse bars: if the DASC decreases (increases) offshore inside the surf zone,  
1444 the bars will grow with an up-current (down-current) orientation. Secondly, the migration  
1445 of the surf zone features can be very sensitive to the sediment transport parameteriza-  
1446 tion because it depends on the alongshore phase lags between the bathymetry and the  
1447 perturbations in the depth-averaged concentration and the latter has unknown functional  
1448 dependences on the perturbations in the water depth, wave orbital velocity, current and

1449 turbulent eddies. This might explain why migration of the surf zone features is generally  
1450 not well modeled at present [*Garnier et al.*, 2006, 2008; *Ribas et al.*, 2012].

1451 The formulation presented here can also be used in the opposite sense: knowing the  
1452 characteristics of the observed bars, the distribution of the DASC can be inferred in a  
1453 qualitative manner. This may be particularly useful in determining the validity of dif-  
1454 ferent sediment transport parameterizations depending on weather and wave conditions,  
1455 morphology and beach conditions. The results presented here are also of interest for  
1456 coastal engineering because they can be used to improve numerical models and existing  
1457 integrated transport formulas (e.g., CERC formula for total alongshore transport rate or  
1458 cross-shore transport formulas), which, at present, neglect the effect of alongshore rhyth-  
1459 mic morphologies. For instance, *Splinter et al.* [2011] modeled the cross-shore migration  
1460 of shore-parallel bars using a parameterization to describe the effect of possible along-  
1461 shore rhythmicities. These results can aid in designing beach nourishments or coastal  
1462 structures, or to understand the complex morphodynamic pattern evolution under time-  
1463 varying forcing conditions. They can also be a guide for the design of field experiments:  
1464 by measuring the currents and the cross-shore distribution of the DASC, the nature of the  
1465 underlying morphodynamic rhythmic pattern can be assessed. Finally, the formulation  
1466 and methodology that we have presented here could also be applied to other natural sand  
1467 features whose dynamics can be described by depth-averaged shallow water models, such  
1468 as sand bars in rivers [*Zolezzi et al.*, 2012], sand bars or shoals in tidal embayments (inlets  
1469 or estuaries) [*de Swart and Zimmerman*, 2009] and tidal sand banks [*Blondeaux*, 2001].

## 9. FUTURE RESEARCH

### 9.1. Field observations

1470 During the last two decades much theoretical research has been done on the develop-  
1471 ment of alongshore rhythmic patterns based on the self-organization hypothesis. While  
1472 the model studies have gone from very idealized to quite realistic, in some circumstances,  
1473 including many of the possibly relevant processes, there is a lack of the necessary quantita-  
1474 tive validation of those studies with field observations. Comparisons of the characteristics  
1475 of the modeled features with field observations that have been made so far are mostly  
1476 qualitative and more quantitative model-data comparison should be a guide to improve  
1477 the models so that they can better reproduce the observations. One reason is that field  
1478 observations that are adequate to test the proposed feedback mechanisms are limited.  
1479 Also, a detailed analysis of such existing observations during the events of formation and  
1480 evolution of the morphologic features is often missing. Therefore an important issue in  
1481 future research is provision of high quality measurements of the variables involved in the  
1482 dynamics of the morphologic features during the events of pattern development.

1483 Measurement of the characteristics of the patterns (alongshore spacing, orientation,  
1484 migration rate, amplitude, shape) is important, together with the ambient bathymetry  
1485 (i.e., that without bed-forms). In the swash and surf zones, some of these bathymetric  
1486 variables are nowadays measured with remote sensing techniques and thereby they are  
1487 available with a good time and space resolution (e.g., *Konicki and Holman* [2000]; *van*  
1488 *Enckevort et al.* [2004]; *Holman et al.* [2006]; *Ribas and Kroon* [2007]). Others (such as  
1489 the amplitude of the features and the ambient bathymetry) must be measured in situ,  
1490 and such data are expensive and thereby scarce. In particular, the lack of large scale  
1491 bathymetric surveys near in time to an event of pattern development is very often the most

1492 important limitation for successfully applying the models and comparing their results with  
1493 the observations. Important climate (wave and atmospheric) conditions that are needed  
1494 to verify the mechanisms vary from feature to feature. For surf zone bars, offshore wave  
1495 height, period and direction (and also tide and wind properties, if significant at the site)  
1496 are typically available. In the case of shoreface-connected ridges, obtaining estimations of  
1497 wind and wave conditions and coastal currents during storms is a challenge. For beach  
1498 cusps, wave conditions at the beachface (i.e., type of breaking, as well as height, period  
1499 and direction) are important quantities and they are hardly measured. Measurements of  
1500 in/exfiltration would also aid in determining cusps dynamics.

1501 Measurements of DASC and the horizontal circulation would provide a direct exper-  
1502 imental validation of the formulation and the methodology presented here. Note that,  
1503 although the sediment transport formulas used in models are widely used formulas, they  
1504 are largely based on laboratory calibrations and it is often unclear how appropriate they  
1505 are for field conditions [*Soulsby, 1997; Amoudry and Souza, 2011*]. The cross-shore distri-  
1506 bution of DASC is important to know, ideally both when features are present and absent.  
1507 Measuring the alongshore profile of the perturbation of DASC over the growing features  
1508 would be particularly difficult, but it would help in understanding its potential role in the  
1509 growth and migration of the features. Measurements of the horizontal circulation induced  
1510 by the growing features would then reveal a detailed picture of the mechanisms discussed  
1511 in this contribution. In particular, measurements of the DASC profile in the inner surf  
1512 zone of natural beaches (i.e., beneath broken waves, with the resuspension produced  
1513 by breaking-induced turbulent velocities), and of current deflections over shore-oblique  
1514 transverse bars and shoreface-connected ridges would be unique. Sediment concentration

1515 profiles can be measured with acoustic backscatter systems (ABS) and acoustic Doppler  
1516 current profilers (ADCP), which provide reliable data when waves are not breaking. How-  
1517 ever, measuring sediment concentration under breaking waves (i.e., inside the surf zone)  
1518 remains a challenge due to the interference of air bubbles and strong turbulent vortices,  
1519 in addition to the difficulty to maintain the tripods well anchored in such high energy en-  
1520 vironment. Also, errors in estimating the water depth (i.e., when no bathymetric surveys  
1521 are available) lead to critical inaccuracies in the DASC profile ( $C = \alpha/D$ ).

## 9.2. Laboratory experiments

1522 *Wave basin* experiments are also likely to reveal important dynamics and confirm (or  
1523 otherwise) hypotheses. Experiments like these would have to be at large scale, because of  
1524 scaling difficulties with sediment grain sizes, and careful control would have to be exerted  
1525 over extraneous effects (e.g., the re-reflection of waves, and the generation of seiching  
1526 modes). If these issues are carefully addressed such experiments are likely to be useful  
1527 in examining DASC and circulation patterns in great detail, as well as providing very  
1528 high resolution bathymetric data-sets. Note that, compared to the natural variability  
1529 in the field, in a wave basin the forcing conditions are controlled, so that features and  
1530 mechanisms to be studied could be isolated. Beach cusp generation would seem the  
1531 most auspicious case to examine, because of the relatively small spatial scales (such an  
1532 experiment could be at prototype scale). Edge wave activity could be carefully monitored,  
1533 along with monitoring of sediment sorting (although water levels within the beach would  
1534 have to be carefully considered to ensure that they were consistent with field values).

### 9.3. Modeling

1535 Necessary future modeling research again depends on the features considered, but,  
1536 generically, should focus on testing the effect of heretofore neglected processes, improving  
1537 representation of some of those included (not that this is not necessarily just incremental,  
1538 but may require wholly different approaches), and developing better numerical techniques  
1539 so that modeling can be carried out in circumstances that were previously prohibitively  
1540 difficult. Note that models have been very useful to isolate and study processes and  
1541 mechanisms but, at a certain point, the effect of including the neglected processes must  
1542 be checked.

1543 The cross-shore sediment transport processes have so far been assumed to play a passive  
1544 role (see section 2.1). The essential mechanisms behind the development of the features  
1545 are unraveled with this assumption but including a more accurate cross-shore transport  
1546 description would be an important step forward because the latter explains beach profile  
1547 dynamics (e.g., the formation and migration of the shore-parallel bars). To study this  
1548 issue, (quasi) three-dimensional models should be developed because including a descrip-  
1549 tion of the vertical structure of the flow and the intra-wave oscillatory motion [*Putrevu*  
1550 *and Svendsen, 1999*] is mandatory to successfully describe cross-shore sediment transport  
1551 processes. This should be a priority to gain more understanding on the development of  
1552 crescentic bars since the transformation of a shore-parallel bar into a crescentic bar of-  
1553 ten occurs whilst the bar migrates onshore [*Short, 1999*]. The intra-wave approach is,  
1554 as has been mentioned earlier (section 7), intrinsic to the modeling of beach cusps, but  
1555 a better description of the boundary layer in the swash zone would also be highly de-  
1556 sirable because it is unsteady and reverses during the swash event, and its impact on

1557 sediment movement is still not fully understood [see *Barnes and Baldock*, 2010; *Briganti*  
1558 *et al.*, 2011]. Including cross-shore sediment transport processes would be also desirable  
1559 to model shoreface-connected ridges because they could be affected by the net exchange  
1560 of sediment between the inner shelf and the nearshore zone. The present-day models  
1561 consider only a weak exchange, but that assumption is controversial [*Kana et al.*, 2011;  
1562 *Schwab et al.*, 2013]. *Niedoroda et al.* [1985] already pointed out that during storms the  
1563 offshore Ekman flow near the bottom might bring large amounts of sediment to the inner  
1564 shelf, thereby feeding the ridges.

1565 There is one aspect of the current-driven sediment transport considered in the present  
1566 contribution (i.e., the sediment transport occurring in the presence of depth-averaged  
1567 currents) that deserve attention in future research. As has been mentioned, in some ap-  
1568 plications it has thus far been assumed that the sediment load in equation (3) is unaffected  
1569 by perturbations [*Ribas et al.*, 2012] (for numerical reasons). However, inclusion of this  
1570 effect can be important in the case of oblique wave incidence and understanding their role  
1571 on transverse bar formation should be a priority in future research. Moreover, a more  
1572 realistic description of suspended load transport would include considering the time and  
1573 space dynamics of the suspended sediment concentration, rather than assuming that the  
1574 concentration is always in equilibrium with local hydrodynamics as done in most of the  
1575 present models. This would allow for time and space lags in the sediment exchange with  
1576 the bed. This was already included by *Reniers et al.* [2004] for crescentic bar dynamics  
1577 and the effect was minor. However, *Murray* [2004] described the formation of transverse  
1578 bars and rip-channels in the surf zone, in which the lags in the sediment exchange with the  
1579 bed were crucial for the growth. Also, we know that these lags are important for beach



1580 cusp formation (section 7.5). So future surf zone models for transverse bar formation  
1581 should take these lags into account.

1582 Studying the evolution of alongshore rhythmic patterns with the sand being composed of  
1583 multiple grain sizes is also a challenge for the future (notice that describing the dynamics  
1584 of the smaller grain sizes in suspension would require taking settling lags into account).  
1585 Field data [e.g., *Baptist et al.*, 2006] reveal a positive correlation between density and  
1586 diversity of benthos communities on one hand and fining of sediment on the other hand.  
1587 Modeling and understanding the distribution of mean sediment grain size and sorting  
1588 over finite-amplitude shoreface-connected sand ridges, that are important in the context  
1589 of modeling ecology of coastal, remain a challenge. So far studies have considered such  
1590 problems only during the initial formation of the bed forms [e.g., *Walgreen et al.*, 2003],  
1591 when vertical sorting can be ignored.

1592 Some aspects of wave climate also require study. Spreading of wave incidence angle and  
1593 period has been considered in modeling crescentic bars [*Reniers et al.*, 2004; *Smit et al.*,  
1594 2008; *Castelle et al.*, 2010b] but not for transverse bars. These effects could affect the sed-  
1595 iment stirring process (and hence the DASC) and might alter the formation mechanisms  
1596 of the features. The effect of including a wave spectrum on cusps has been considered by  
1597 *Coco et al.* [2001], but has not yet been taken into account in fluid dynamical modeling  
1598 [*Dodd et al.*, 2008]. Dealing with storms that have different durations and intensities, and  
1599 with different ocean swell conditions is another challenge. This would require the appli-  
1600 cation and statistical analysis of Monte Carlo simulations. Also, modeling the interaction  
1601 between transverse bars and low frequency hydrodynamic oscillations is another issue that  
1602 deserves further exploration.

1603 Including tides is important for surf zone bars because tidal variability moves the shore-  
1604 line and the surf zone back and forth at a time scale comparable to the characteristic  
1605 time scale of bar morphodynamics. This can have an important effect on transverse bar  
1606 formation, especially in the intertidal features. Also, the tides may cause a significant  
1607 variability on the water depth over the crescentic bars. In spite of this, the effect of tides  
1608 has been generally ignored in the modeling because of the mathematical complexity of  
1609 the problem, wherein we have a moving boundary (the shoreline), which renders both an-  
1610 alytical and numerical description difficult. This is in fact the same problem encountered  
1611 in modeling beach cusps, where the swash variability must be taken into account. There-  
1612 fore, the development of numerical techniques suited to obtaining high accuracy solutions  
1613 in this type of non-steady boundary problem would be desirable. These improvements  
1614 should be a priority in describing the long-term evolution of beach cusps to a dynamic  
1615 equilibrium (see section 7.5).

1616 Modeling the potential interaction between different features is another challenge.  
1617 *Castelle et al.* [2010b] and *Tiessen et al.* [2011] examined the effect of pre-existing bed-  
1618 forms (of the same type) on subsequent development, but studying the interaction of  
1619 different types of features would be also interesting. For instance, up-current finger bars  
1620 in open beaches tend to occur when an outer crescentic bar is present [*Ribas et al.*, 2014].  
1621 In this sense, future modeling studies of open-beach finger bars should use a non-linear  
1622 model and start with a realistic initial bathymetry, incorporating pre-existing larger-scale  
1623 variability. In the same way, a system of shoreface-connected ridges in the inner shelf can  
1624 affect the dynamics of the smaller scale surf zone bars.

1625 Finally, a big challenge for the future is to incorporate biologic variables to the mor-  
 1626 phodynamic models in order to study the interaction between the morphologic features,  
 1627 hydrodynamics and vegetation or benthic life and fish. Study of how the transport of  
 1628 pollutants is modified in the presence of morphodynamic patterns (due, e.g, to rip current  
 1629 circulation) is also relevant.

## APPENDIX A: DERIVATION OF THE BED EVOLUTION EQUATION (2) AND THE SEDIMENT TRANSPORT EQUATION (3)

1630 The dynamics of the depth-integrated volume concentration of sediment in suspension,  
 1631  $\alpha_s$  (different from  $\alpha$  in equation (3) that includes both the suspended and the bedload  
 1632 contributions) can be described with the following simple advection equation

$$1633 \quad \frac{\partial \alpha_s}{\partial t} + \vec{\nabla} \cdot (\alpha_s \vec{v}) = (E - D) , \quad (\text{A1})$$

1634 stating that the total suspended load in the water column changes due to both the advec-  
 1635 tion by the current and the exchange with the bed [Murray, 2004]. The latter is described  
 1636 by the entrainment function  $E$  (the upwards flux due to stirring by waves, currents and  
 1637 turbulence) and the deposition function  $D$  (the downwards flux due to the settling of  
 1638 the grains towards the bed due to gravity). Other terms can be added to equation (A1),  
 1639 such as a slope term and a horizontal diffusive term [Amoudry and Souza, 2011], but  
 1640 the objective here is keeping it as simple as possible. The bed evolution equation then  
 1641 becomes

$$1642 \quad (1 - p) \frac{\partial z_b}{\partial t} + \vec{\nabla} \cdot \vec{q}_b = -(E - D) , \quad (\text{A2})$$

1643 stating that the total exchange of sediment with the bed results in changes in bed level  
 1644 [Amoudry and Souza, 2011], and that the total exchange is due to both divergence of

1645 bedload transport, where  $\vec{q}_b \propto \alpha_b \vec{v}$  (and  $\alpha_b$  is the sediment load as bedload), and exchange  
 1646 with suspended load. Substituting  $E - D$  from equation (A1) into equation (A2), and  
 1647 defining  $\vec{q}$  as in equation (3) with  $\alpha = \alpha_s + \alpha_b$ , yields

$$1648 \quad (1 - p) \frac{\partial z_b}{\partial t} + \frac{\partial \alpha_s}{\partial t} + \vec{\nabla} \cdot \vec{q} = 0, \quad (\text{A3})$$

1649 Since the dynamics of suspended sediment grains (described by equation (A1)) occurs at  
 1650 hydrodynamic time scales, when the quasi-steady approximation is applied,  $\partial \alpha_s / \partial t = 0$ ,  
 1651 and this term drops out from equation (A3). This means that temporal lags between flow  
 1652 and suspended load are disregarded. Another assumption is that  $\alpha_s$  depends only on the  
 1653 local hydrodynamics, i.e., the spatial lags are not considered because the length scale of  
 1654 sediment settling or picking-up processes is much smaller than that of the morphodynamic  
 1655 features. As a result,  $\alpha_s$  is considered as a known function of the flow, which shortcuts  
 1656 solving equation (A1). After applying all these assumptions, equation (A3) leads to  
 1657 equation (1), with the sediment transport described by equation (3) and  $\alpha$  being a local  
 1658 quantity (meaning that sediment load is in equilibrium with the local hydrodynamics).

1659 Three additional assumptions are implicit in our sediment transport formulation (3).  
 1660 First of all, sediment sorting is not accounted for and a single grain diameter is considered.  
 1661 Although systematic gradients in grain size can be observed on the studied features (e.g.,  
 1662 across shoreface connected ridges), the hypothesis here is that this is not essential for  
 1663 explaining their formation. Second, as a result of the joint action of waves and currents  
 1664 acting in different directions,  $\vec{q}$  may not be parallel to  $\vec{v}$  in some cases. In other words, the  
 1665 proportionality factor  $\alpha$  may be a second order tensor rather than a scalar. Again, this is  
 1666 not considered because we assume that these effects are not essential for the formation of  
 1667 our morphological patterns. Finally, in addition to the sediment transport driven by the

1668 depth-averaged currents (first term on the RHS of equation (3)), in the coastal zone there  
 1669 are also cross-shore sediment transport processes driven by, e.g., the waves alone (due  
 1670 to nonlinearities and streaming), the vertical structure of the currents (e.g., undertow  
 1671 [*Svendsen, 2006*]) and the gravity-driven transport. The joint action of these three latter  
 1672 components controls the long-term dynamics of the cross-shore profile (i.e., time scales of  
 1673 weeks-months, see *Ruessink et al. [2007]*) and is typically at least one order of magnitude  
 1674 smaller than the transport driven by the alongshore current or the rip current circulation.  
 1675 Since the working hypothesis here is that those latter currents, in combination with the  
 1676 DASC distribution, control the dynamics of the morphological features of interest, we  
 1677 assume that the cross-shore transport processes build an alongshore uniform equilibrium  
 1678 profile which is stable, so that the deviations from it just cause a net diffusive transport  
 1679 (second term of the RHS of equation (3)).

## APPENDIX B: WAVE EQUATIONS

1680 As it has been stated in section 2.2, the knowledge of wave radiation stresses, wave  
 1681 orbital velocity and wave energy dissipation is necessary to solve the hydrodynamic equa-  
 1682 tions for the currents. Although the description of the incoming surface gravity waves  
 1683 is generally complicated, it is sufficient for our purpose to assume waves have random  
 1684 heights with a Rayleigh distribution characterized by the root-mean-square height,  $H$ ,  
 1685 but a narrow spectrum in frequency and direction. The simplest set of equations describ-  
 1686 ing their transformation from deep water to shore includes the dispersion relation, the  
 1687 wavenumber irrotationality relation and the wave energy balance. The dispersion relation  
 1688 (with Doppler shift) reads

$$1689 \quad \omega = \sqrt{gk \tanh(kD)} + \vec{v} \cdot \vec{k} . \quad (\text{B1})$$

1690 where  $g$  is gravity,  $\vec{k} = (k_x, k_y)$  is the wavenumber vector and  $k$  its modulus. The absolute  
 1691 frequency (frequency with respect an observer at rest on Earth) is  $\omega$  and is assumed  
 1692 constant. The wavenumber irrotationality (conservation of wave crests) reads

$$1693 \quad \frac{\partial k_x}{\partial y} = \frac{\partial k_y}{\partial x} . \quad (\text{B2})$$

1694 The depth-integrated wave energy balance reads

$$1695 \quad \frac{\partial E}{\partial t} + \vec{\nabla} \cdot ((\vec{v} + \vec{c}_g)E) + \mathbf{S} : \vec{\nabla} \vec{v} = -\mathcal{D} , \quad (\text{B3})$$

1696 where  $E = \rho g H^2 / 8$  is the wave energy density (energy for horizontal area unit),  $\rho$  is  
 1697 water density,  $\vec{c}_g$  is the group velocity vector, and  $\mathcal{D}$  is the wave energy dissipation rate,  
 1698 which must be parameterized. In the surf zone, the main source of energy dissipation is  
 1699 wave breaking [parameterized following, e.g., *Thornton and Guza, 1983*], which is much  
 1700 larger than dissipation by bed friction so that the latter is neglected. In the shoaling zone,  
 1701 however, wave energy dissipation by bed friction must be accounted for. According to the  
 1702 notation in tensor algebra,

$$1703 \quad \mathbf{S} : \vec{\nabla} \vec{v} = S_{xx} \frac{\partial v_x}{\partial x} + 2S_{xy} \frac{\partial v_x}{\partial y} + S_{yy} \frac{\partial v_y}{\partial y} . \quad (\text{B4})$$

1704 The wave transformation equations described above can be solved to find the wavenum-  
 1705 ber  $k$ , the wave angle (angle between the direction of propagation of the wave and the  
 1706 shore normal direction, see Figure 4)  $\theta$  ( $k_x = -k \cos \theta$ ,  $k_y = k \sin \theta$ ), and the wave energy  
 1707  $E$  as a function of  $x, y$  and  $t$ . Expressions for the wave radiation stresses, group velocity  
 1708 and orbital velocity amplitude at the bed, as function of  $E, k$  and  $\theta$ , are obtained from  
 1709 linear wave theory [*Mei et al., 2005*]. In some applications, it is also necessary to describe  
 1710 the dynamics of the *roller*, i.e., the aerated mass of water located on the shoreward face  
 1711 of breaking waves. This is achieved with an extra equation for the balance of the roller

1712 energy density. The set of equations above describe wave refraction (by topography and  
 1713 currents), shoaling and breaking, the wave processes that are essential for the creation  
 1714 of the morphodynamic features of interest. More complex wave characteristics, like a  
 1715 spectral dispersion of wave frequency and direction, and other processes in wave propa-  
 1716 gation, such as wave diffraction and reflection, are not accounted for. The potential role  
 1717 of these neglected wave processes and properties on morphodynamic pattern formation is  
 1718 discussed in section 9. The specific wave equations used to describe the different features  
 1719 can be found in *Ribas et al.* [2012] (surf zone bars) and *Vis-Star et al.* [2007] (shoreface  
 1720 connected ridges). A detailed description of the depth-integrated momentum balance and  
 1721 the wave equations is given in *Phillips* [1977] and *Svendsen* [2006].

### APPENDIX C: QUANTITATIVE GLOBAL ANALYSIS

1722 In order to perform a quantitative global analysis of the BEE, as described in sec-  
 1723 tion 3.4, the starting point is to multiply the nonlinear BEE, equation (8), by the bed  
 1724 level perturbation associated to the feature,  $h(x, y, t)$ , and to integrate this equation over  
 1725 the domain. By defining the average of  $f(x, y)$  over the computational domain (with the  
 1726 alongshore distance of the domain,  $L_y$ , being a multiple of the alongshore spacing of the  
 1727 feature)

$$\bar{f} = \frac{1}{L_x L_y} \int_0^{L_y} \int_0^{L_x} f(x, y) dx dy \quad (\text{C1})$$

1729 the integrated nonlinear BEE multiplied by  $h$  reads

$$(1 - p)h \overline{\frac{\partial h}{\partial t}} = -\overline{h D\vec{v} \cdot \nabla \mathcal{C}} - \overline{\gamma |\nabla h|^2}. \quad (\text{C2})$$

1731 The second RHS term has been obtained by integrating by parts and using the alongshore  
 1732 periodicity of  $h(x, y, t)$ . Importantly, the LHS term can be written as

$$1733 \quad \overline{h \frac{\partial h}{\partial t}} = \frac{1}{2} \frac{d}{dt} (\overline{h^2}) \quad (\text{C3})$$

1734 because the computational domain is constant in time. Therefore, the LHS term is pro-  
 1735 portional to the time derivative of the potential energy of the pattern, that is, the gravita-  
 1736 tional potential energy of bed sediment grains (like the potential energy of water surface  
 1737 gravity waves). If such derivative is positive (negative), the morphologic pattern will  
 1738 grow (decay). Accordingly, we define the global growth rate of an alongshore rhythmic  
 1739 morphologic pattern as

$$1740 \quad \Omega = \frac{1}{\overline{h^2}} \overline{h \frac{\partial h}{\partial t}} . \quad (\text{C4})$$

1741 The meaning of the global growth rate becomes clearer if we consider an idealized  
 1742 morphological pattern consisting of a sinusoidal bed wave with growth/decay given by  $\Omega_s$   
 1743 and alongshore propagation celerity  $c_s$ ,

$$1744 \quad h(x, y, t) = \exp(\Omega_s t) \hat{h}(x) \cos(\kappa(y - c_s t) + \psi(x)) \} . \quad (\text{C5})$$

1745 Here,  $\hat{h}(x)$  is a function that stands for the cross-shore structure of the bed wave,  $\kappa$   
 1746 is its alongshore wavenumber and  $\psi(x)$  accounts for the possible differences in spatial  
 1747 lags at each cross-shore position (i.e., yielding an obliquely oriented feature). For such  
 1748 morphologic pattern, it can be proved that

$$1749 \quad \overline{h \frac{\partial h}{\partial t}} = \Omega_s \overline{h^2} , \quad (\text{C6})$$

1750 so that its growth rate coincides with the global growth rate as defined in equation (C4),  
 1751  $\Omega_s = \Omega$ .



1752 Returning to the general case, by inserting the global growth rate, equation (C4), into  
 1753 the integrated BEE, equation (C2), the master equation governing the growth/decay of  
 1754 the pattern follows as

$$1755 \quad \Omega = \frac{1}{h^2}(\mathcal{P} - \Delta) . \quad (\text{C7})$$

1756 where

$$1757 \quad \mathcal{P} = -\frac{1}{1-p} \overline{h D\vec{v} \cdot \nabla \mathcal{C}} , \quad \text{and} \quad \Delta = \frac{1}{1-p} \overline{\gamma |\nabla h|^2} . \quad (\text{C8})$$

1758 Notice that  $\Delta > 0$  so that it always causes decay of the pattern and it is called the  
 1759 damping term. Thus, any growth of the pattern must be described by  $\mathcal{P}$ , which is called  
 1760 the production term. Notice that the production term measures the cross-correlation  
 1761 between  $h(x, y, t)$  and the quantity  $D\vec{v} \cdot \nabla \mathcal{C}$  that has been discussed in section 3.2 as  
 1762 being responsible of the erosion/deposition processes driven by the joint action of the  
 1763 gradients in DASC and the currents. Consistently with the local analysis presented in  
 1764 that section, if the regions where  $h > 0$  and the current opposes the gradients in DASC  
 1765 (or  $h < 0$  and the current runs with the gradients in DASC) dominate over the regions  
 1766 where the contrary occurs,  $\mathcal{P} > 0$ . Then, if the production term is positive and larger  
 1767 than the damping term, the pattern will grow. If the opposite is true, the pattern will  
 1768 decay. If  $\mathcal{P} = \Delta$  the pattern can change its shape or migrate but its global amplitude will  
 1769 remain constant.

1770 Regarding the alongshore migration, for the case of a sinusoidal wave, it is seen that its  
 1771 propagation celerity fulfills

$$1772 \quad \frac{\overline{\partial h}}{\partial y} \frac{\partial \overline{h}}{\partial t} = -c_s \overline{\left(\frac{\partial h}{\partial y}\right)^2} . \quad (\text{C9})$$

1773 Therefore, this brings us to define the global migration celerity of any alongshore rhythmic  
 1774 morphologic pattern as

$$1775 \quad c = -\frac{1}{\left(\frac{\partial h}{\partial y}\right)^2} \frac{\partial h}{\partial y} \frac{\partial h}{\partial t}. \quad (\text{C10})$$

1776 As far as we know, this quantitative global analysis of pattern growth/decay and mi-  
 1777 gration in the nearshore was first presented and applied in *Garnier et al.* [2006]. The  
 1778 methodology was extended and further interpreted by *Vis-Star et al.* [2008].

## APPENDIX D: PHYSICS OF BATHYMETRICALLY INDUCED RIP CURRENT CIRCULATION

1779 To understand the physics behind rip current circulation (blue streamlines in Figure 12),  
 1780 we first consider the hydrodynamics originated by a shoal on an otherwise alongshore  
 1781 uniform topography in an area of breaking waves. The momentum carried by the waves is  
 1782 described by the *radiation stress tensor*,  $S_{xx}, S_{yy}, S_{xy} = S_{yx}$  (see section 2.2 and Appendix  
 1783 B). This momentum is released at breaking originating a hydrodynamic force in the  
 1784 wave-averaged momentum equation through the divergence of the wave radiation stresses  
 1785 (equation 5). It is important to notice that in case of normal wave incidence  $S_{xx}$  is larger  
 1786 than  $S_{yy}$  because of the anisotropy caused by wave propagation direction [*Mei*, 1989;  
 1787 *Svendsen*, 2006]. In general, the wave height reduction at breaking produces an onshore  
 1788 directed hydrodynamic force on the water motions and hence a set-up of the mean sea  
 1789 level  $z_s = z_s(x)$ . Since breaking is induced by a reduction in water depth, there is more  
 1790 energy dissipation over the shoal than at its deeper sides. Therefore, there is more set-  
 1791 up, i.e., a higher water level, shoreward of the shoal than shoreward of its sides. This  
 1792 difference in water level can not be balanced by the alongshore gradients in  $S_{yy}$  because

1793 they are smaller. Therefore, the water flows alongshore from behind the shoal down to  
 1794 the deeper area and produces a higher level there. Again, this is not balanced by the  
 1795 cross-shore gradients in  $S_{xx}$  and the water flows offshore. In this way, a circulation cell is  
 1796 created with onshore current over the shoal and offshore current at the sides (Figure 12).  
 1797 Apart from this simple physical explanation, the conclusion that there is no possible  
 1798 steady solution balancing gradients in radiation stresses with pressure gradients without  
 1799 a circulation is readily seen by working out the momentum balance equations. When there  
 1800 are a number of shoals separated by channels, seaward flowing currents or rip currents are  
 1801 originated at the channels in this manner.

## APPENDIX E: GLOBAL ANALYSIS FOR FINITE AMPLITUDE BEHAVIOR OF CRESCENTIC BARS

1802 Following the analysis by *Garnier et al.* [2010], the first step is a careful analysis of the  
 1803 production ( $\mathcal{P}$ ) and the damping ( $\Delta$ ) terms introduced in Appendix C (equation C8) when  
 1804 the bed level deviations  $h$  (difference in bed level with respect to straight bar situation)  
 1805 of the shoals and channels increases. Since  $\mathcal{P}$  and  $\Delta$  describe the tendency to grow  
 1806 or to decay (respectively) of the shoals and channels, their competition determines the  
 1807 instantaneous growth rate  $\Omega$  (equation C7). Initially,  $\Omega > 0$  so that the crescentic shape  
 1808 of the bar grows but, as the amplitude increases,  $\Omega$  decreases and it eventually becomes 0.  
 1809 At this stage the pattern does not grow anymore, i.e., saturation occurs. Although there  
 1810 is a slight increase of  $\Delta$  that contributes to the decrease of  $\Omega$ , the latter is mainly due to a  
 1811 weakening of the production term, that is, a weakening of the positive feedback between  
 1812 morphology and circulation (Figure 25a). Then, it can be shown that the cross-shore flow  
 1813 component together with the cross-shore gradients in DASC dominate the production

1814 term,

$$1815 \quad \mathcal{P} = -\frac{1}{1-p} \overline{hD\vec{v} \cdot \nabla \mathcal{C}} \simeq -\frac{1}{1-p} \overline{uhD \frac{\partial \mathcal{C}}{\partial x}}. \quad (\text{E1})$$

1816 Finally, it is found that the decrease of the production term when the amplitude increases  
 1817 is controlled by the cross-correlation between cross-shore flow and bed level perturba-  
 1818 tion,  $\mathcal{S} = \overline{uh}/\|u\|\|h\|$ . To elucidate which of the characteristics of the finite-amplitude  
 1819 crescentic bars causes the decrease in the latter quantity, numerical experiments of the  
 1820 hydrodynamics over a fixed bathymetry with a crescentic bar are done. Increasing the  
 1821 amplitude of the shoals and channels but keeping the shape,  $\mathcal{S}$  hardly decreases. In con-  
 1822 trast, widening the shoals and narrowing the channels,  $\mathcal{S}$  significantly decreases because  
 1823 the onshore current  $u$  over the shoals strongly weakens with the result that the whole cir-  
 1824 culation cell weakens. A similar effect (but less significant) is obtained with the shoreward  
 1825 (seaward) shift of the shoals (channels) and the overall seaward shift of the bar.

1826 A similar analysis of the production and damping terms is carried out to investigate  
 1827 the influence of the wave incidence angle on the transitions between shore-parallel bars  
 1828 and crescentic bars [*Garnier et al.*, 2013]. Again, the damping term does not play an  
 1829 important role as it hardly depends on the wave angle. Both the inhibition of crescentic  
 1830 bar formation for oblique wave incidence and the bar straightening for increasing wave  
 1831 angle are caused by a weakening of the production term (see Figure 25b). And, again,  
 1832 this is in turn controlled by a decrease of the cross-correlation between cross-shore flow  
 1833 and bed level perturbation,  $\mathcal{S}$ . It is shown that, by increasing the wave angle, this term  
 1834 decreases due to both a weakening of the rip current intensity and a down-wave shift of  
 1835 the rips (i.e., a phase lag between the rips and the channels). This decrease of  $\mathcal{S}$  explains  
 1836 the weakening of the positive feedback between flow and morphology.

## APPENDIX F: PHYSICS OF CURRENT MEANDERING OVER SHORE-OBLIQUE BARS OR RIDGES

1837 There are two potential hydrodynamic mechanisms that can create a meandering of the  
1838 alongshore current over the crests of shore-attached oblique bars or ridges (blue stream-  
1839 lines in Figures 16b,c and 19). The first one is water mass conservation in the cross-bar  
1840 direction. Consider first the case where the bar is up-current oriented. When the along-  
1841 shore current flows over an up-current oriented bar, the cross-bar component of the current  
1842 becomes larger due to water mass conservation (because depth decreases, the mechanism  
1843 is sketched in Figure 5b). Since bar length is much larger than bar width, the along-bar  
1844 component hardly changes. This gives an offshore current deflection (positive  $u$ ) over  
1845 up-current bar crests. *Trowbridge* [1995] showed quantitatively that, in the potential flow  
1846 approximation of his idealized model, this was the mechanism responsible for the offshore  
1847 deflection over the shoreface connected sand ridges. The second mechanism is related to  
1848 the frictional torques created by depth changes as the alongshore current flows over the  
1849 bar. When the current runs from the trough to the crest of an up-current oriented bar,  
1850 it experiences a clockwise rotation because friction is larger over the crest than it is at  
1851 the trough. This again gives an offshore current deflection (positive  $u$ ) over up-current  
1852 bar crests. This effect was described by *Zimmerman* [1981] and was recognized as crucial  
1853 in the context of tidal currents over tidal sand banks. *Ribas et al.* [2012] showed that  
1854 frictional torques were the essential mechanism to produce the offshore current deflection  
1855 over the up-current oriented transverse bars in their model. If the bars are down-current  
1856 oriented, both mechanisms create an onshore deflection of the current (negative  $u$ ) over  
1857 bar crests.

**GLOSSARY**

1858 **Advection:** Forward carrying by a current.

1859 **Alongshore current:** Current along the coast. In case of the surf zone it refers to  
1860 the current driven by the breaking waves with oblique incidence.

1861 **Backwash:** Flow of water down the swash zone during the backward motion of a  
1862 wave incoming at the beach face.

1863 **Bedload transport:** Sediment transport corresponding to particles that are in  
1864 frequent contact with the sea bed (sliding, rolling or bouncing).

1865 **Bed shear stresses:** Tangential forces per unit area exerted on the sea bed by a  
1866 flow (and vice versa).

1867 **Cross-shore sediment transport:** Sediment transport in the cross-shore direc-  
1868 tion driven by the combination of the waves (asymmetry, skewness and streaming), the  
1869 undertow and the gravity.

1870 **Depth-averaged current (or current):** Time-averaged water volume flux per  
1871 unit width, after filtering out the fast oscillatory motions, divided by the time-averaged  
1872 water depth (also called mass-transport current in the literature).

1873 **Depth-averaged sediment concentration (DASC):** The total volume of mo-  
1874 bilized sediment in a water column (including bedload and suspended load) per water  
1875 volume unit in case of sediment transport by a current.

1876 **Down-current orientation:** Orientation of a transverse bar or ridge so that its  
1877 offshore end is shifted downstream.

1878 **Edge wave:** Surface gravity wave that propagates along the coast and is trapped  
1879 against it in such a way that its amplitude decays roughly exponentially in the seaward  
1880 direction with an e-folding distance of the order of the alongshore wavelength.

1881 **Feedback mechanism:** Loop wherein the hydrodynamics affects the morphology  
1882 via the sediment transport and the morphology affects in turn the hydrodynamics setting  
1883 the solid boundaries of the water body. Starting in a perturbed equilibrium situation,  
1884 the changes in hydrodynamics cause changes in morphology and these may in turn rein-  
1885 force (damp) the changes in hydrodynamics. In such case the feedback is called positive  
1886 (negative).

1887 **Infragravity wave:** Surface gravity waves of lower frequency than the incident  
1888 wind or swell waves, with wave periods ranging from about 20 s to a few minutes.

1889 **Inner shelf:** Region in the nearshore spanning from water depths of a few meters  
1890 to tens of meters, between the surf zone and the middle continental shelf (where the  
1891 along-shelf circulation is usually in geostrophic balance).

1892 **Linear wave theory:** A simplified (linearized) description of surface gravity waves,  
1893 applicable to waves of small height to depth ratio and steepness.

1894 **Low frequency eddies:** Horizontal eddies in the surf zone generated by incident  
1895 wave groups that evolve at time scales of  $O(30 \text{ min})$  and have length scales of  $O(100 \text{ m})$ .

1896 **Morphodynamic instability:** A perturbation growing out of a morphodynamic  
1897 equilibrium due to a positive feedback between flow and morphology so that a new mor-  
1898 phologic pattern showing higher complexity level than the equilibrium emerges.

1899 **Morphodynamic pattern:** Spatial pattern in the morphology and the water mo-  
1900 tions due to their mutual coupling.

1901 **Net:** Adjective applied to variables that result from a time averaging over a time  
1902 scale shorter than that of interest.

1903 **Quasi-steady approximation:** Approximation in coastal morphodynamics where  
1904 the flow is assumed to be steady at each time over the morphology at that time, even  
1905 though the morphology is changing slowly with time. Mathematically this means dropping  
1906 out all the partial time-derivatives from the hydrodynamic equations.

1907 **Reynolds stress tensor:** Stress tensor in the depth and time-averaged momen-  
1908 tum balance equations that accounts for the momentum flux from the turbulent flow  
1909 fluctuations into the mean motions.

1910 **Rhythmic Bar and Beach (RBB) state:** Beach state characterized by a rhythmic  
1911 shoreline and one or more crescentic bars in the beach state classification of *Wright and*  
1912 *Short* [1984].

1913 **Rip channel:** Elongated bed depression or channel trending shore-normal (or  
1914 nearly) in the surf zone where commonly a rip current occurs.

1915 **Rip-channel system:** Patch of several rip channels along the coast.

1916 **Rip current:** Jet-like seaward flowing current that can easily reach  $O(1 \text{ m s}^{-1})$  and  
1917 can be very dangerous for beach users. Rip currents may be due to many causes, one of  
1918 them are breaking waves in a surf zone with one or more rip channels.

1919 **Root-mean-square wave height:** Square root of the mean squared wave height,  
1920 taking all the waves in a wave record.

1921 **Saturation of the growth:** The process whereby an instability mode growing  
1922 out of an unstable equilibrium stops its growth and a new equilibrium displaying certain  
1923 pattern is reached.



1924 **Sediment load:** The total volume of mobilized sediment per horizontal area unit in  
1925 case of sediment transport by a current (also called transport capacity, stirring function  
1926 or depth-integrated sediment concentration).

1927 **Sediment porosity:** Measure of the void (i.e., "empty") spaces in bed sediment  
1928 and is the fraction of the volume of voids over the total volume, between 0 and 1.

1929 **Sediment transport:** Movement of sediment particles driven by the forces exerted  
1930 by water motion.

1931 **Self-organization (process):** Process where some form of global order (pattern)  
1932 emerges out of the local interactions between the components of an initially disordered  
1933 system. This process is spontaneous: the spatial characteristics of the emergent patterns  
1934 do not require spatial variations in the forcing. It is often triggered by random fluctuations  
1935 that are amplified by positive feedback.

1936 **Shear wave:** Oscillatory water motion in the surf zone in case of oblique wave  
1937 incidence originated by a shear instability of the alongshore current. It consists of a  
1938 meandering of the current that propagates downstream with a celerity of the order of the  
1939 current magnitude and with a period similar to infragravity waves but with significantly  
1940 smaller wavelengths.

1941 **Shoal:** Sand deposit with higher bed levels than the surrounding area.

1942 **Shoaling zone:** Nearshore zone offshore the surf zone where the waves feel the sea  
1943 bed and thereby change propagation direction (refraction), wave amplitude and shape.

1944 **Shore-parallel bar:** Elongated shoal parallel to the shore (also called linear bar,  
1945 alongshore-uniform bar or straight bar).

1946 **Surf zone:** Nearshore zone spanning from the beach face to the breaker line, where  
1947 waves break and propagate onshore as bores.

1948 **Suspended load transport:** Sediment transport corresponding to particles that  
1949 are advected by the current in suspension within the water flow.

1950 **Swash zone:** Zone of the beach face that is covered and uncovered by water as the  
1951 water front moves up and down following the incoming waves.

1952 **Transverse Bar and Rip (TBR) state:** Beach state characterized by transverse  
1953 bars associated to the horns of a crescentic bar that have been attached to the shoreline  
1954 in the beach state classification of *Wright and Short* [1984].

1955 **Turbulence:** Random and fast motion of water as small eddies, characterized by  
1956 small length and time scales.

1957 **Undertow:** A nearshore seaward-directed near-bed current that is fed by the  
1958 return flow from broken waves and is caused by the unbalance between the vertical dis-  
1959 tribution of wave radiation stresses and pressure gradients (also called bed return current  
1960 in the literature).

1961 **Up-current orientation:** Orientation of a transverse bar or ridge so that its offshore  
1962 end is shifted upstream.

1963 **Uprush:** Flow of water up the swash zone during the forward motion of a wave  
1964 incoming at the beach face.

1965 **Wave basin:** Laboratory basin with width and length of comparable magnitude  
1966 and a wave maker on one side and a beach or wave-absorbing surface on the opposite side  
1967 to observe the 3D behavior of waves and related processes.

1968 **Wave energy density:** Total mechanical energy of the wave water motion per  
 1969 horizontal area unit.

1970 **Wave flume:** Long and narrow wave-basin of a width much smaller than length to  
 1971 observe the 2D behavior of waves and related processes.

1972 **Wave orbital velocity:** Velocity of the water parcels associated with a wave.

1973 **Wave radiation stresses:** Time-averaged and depth-integrated flux of momentum  
 1974 caused by the wave oscillatory motion only, i.e., excluding the contribution of hydrostatic  
 1975 pressure related to the mean-surface elevation.

1976 **Wave refraction:** Change in wave direction due to a change in phase celerity. It  
 1977 can occur because of bathymetric changes or due to the action of a current and it causes  
 1978 a reduction of the angle between wave crests and the coastline when waves approach the  
 1979 shore.

1980 **Wave shoaling:** Change in wave height due to the reduction in water depth when  
 1981 waves approach the shore. For small angles of wave incidence wave heights first slightly  
 1982 decrease and then increase significantly before breaking.

## NOTATION

$x$  spatial coordinate in the cross-  
 shore direction  
 $y$  spatial coordinate in the along-  
 shore direction  
 $z$  spatial coordinate in the vertical  
 direction  
 $t$  time  
 $\vec{v}$  Depth-averaged current

- $z_b$  sea bed level
- $z_s$  time-averaged sea level
- $D$  time-averaged water depth
- $H$  root-mean-square wave height
- $E$  wave energy density
- $\rho$  sea water density
- $g$  gravitational acceleration
- $\vec{\tau}_b$  bed shear stress
- $\vec{\tau}_s$  wind surface shear stress
- R** Reynolds turbulent stress tensor
- S** wave radiation stress tensor
- $u_b$  root-mean-square amplitude of  
the wave orbital velocity near the  
bed
- $k$  wavenumber of the incident waves
- $\theta$  wave propagation angle with re-  
spect to the shore normal ( $-x$   
axis)
- $\vec{q}$  sediment transport (volume of  
sediment crossing a vertical sur-  
face per width unit and time unit)
- $\alpha$  sediment load (total sediment  
volume in a water column per  
horizontal area unit)
- $\gamma$  sediment diffusivity coefficient

$h$	bed level deviation with respect to alongshore uniform long-term equilibrium
$z_{b0}$	bed level corresponding to the alongshore uniform long-term equilibrium
$p$	bed sediment porosity
$\mathcal{C} = \alpha/D$	depth-averaged sediment concen- tration (DASC)
$\mathcal{C}_0(x)$	depth-averaged sediment concen- tration for the alongshore uni- form long-term equilibrium
$D_0(x)$	water depth for the alongshore uniform long-term equilibrium
$V_0(x)$	alongshore current for the along- shore uniform long-term equilibrium
$c$	small perturbation in DASC
$d$	small perturbation in water depth
$u$	small perturbation in the cross- shore component of $\vec{v}$
$v$	small perturbation in the along- shore component of $\vec{v}$
$x_m$	cross-shore position of a local maximum in DASC
$w$	vertical infiltration velocity in the swash zone

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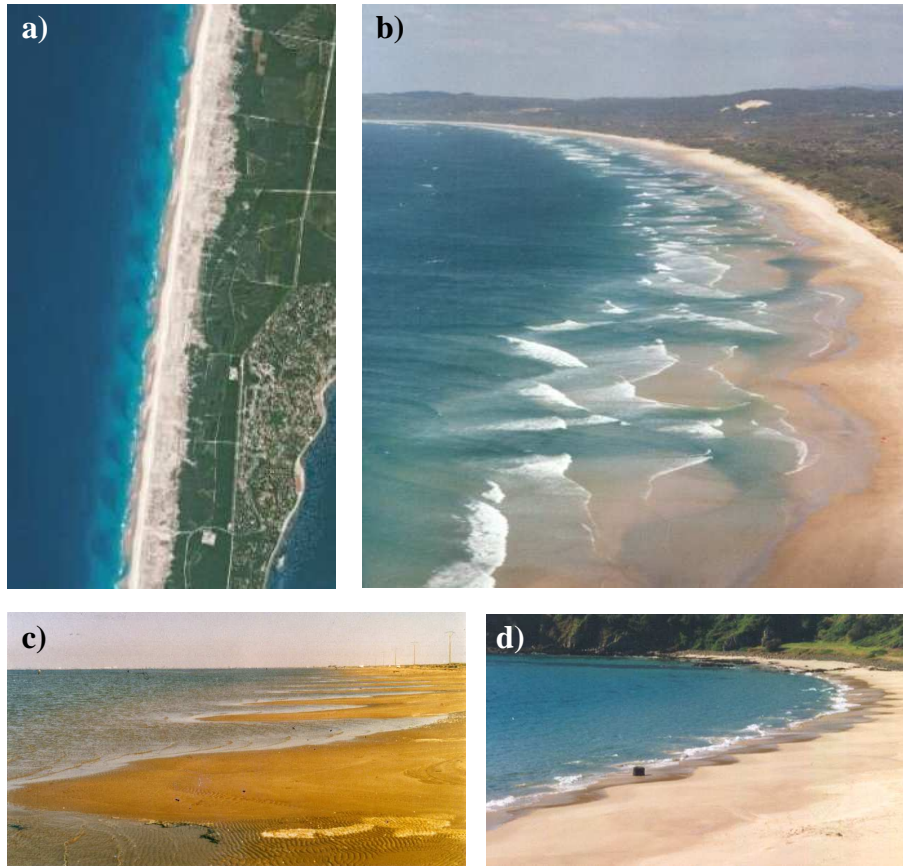
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**Figure 1.** Pictures of (a) a crescentic bar at the Truc Vert beach, France (spacing order of hundreds of m; source: Google Earth, image from NASA), (b) transverse bars at Byron Bay beach, Australia (spacing order of a few hundreds of m), (c) transverse bars at the Ebro delta, Spain (spacing order of a few tens of m), and (d) beach cusps at an Australian beach (spacing order of a few tens of m). The three latter photographs were taken by the authors.

**TABLE 1.** Examples of the coastal sandy features with alongshore rhythmic patterns described in the different sections of the manuscript

Coastal feature	Coastal part	Spatial scale	Temporal scale	Section <sup>a</sup>
Crescentic bars/ rip-channel systems	Surf zone	0.1-3 km	hours-days	4
Transverse bars	Surf zone	10-750 m	hours-days	5
Shoreface-connected ridges	Inner shelf	1-8 km	centuries-millenia	6
Beach cusps	Swash zone	1-50 m	minutes-hours	7

<sup>a</sup> In each section, a list of references with feature observations is included that substantiate the length and time scales.

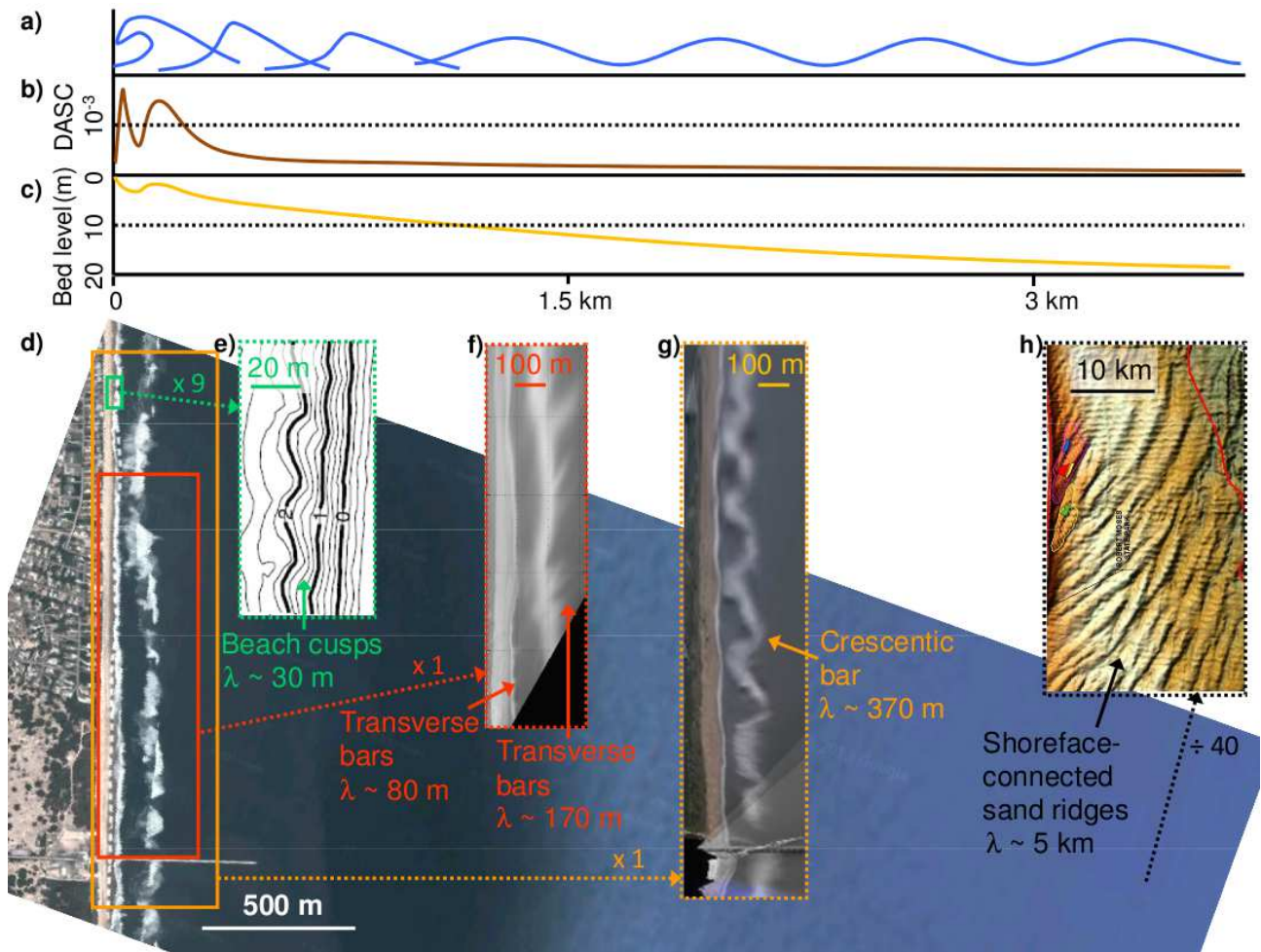
**TABLE 2. The Sediment Load  $\alpha$  in Standard Sediment Transport Formulas.**

Formula name <sup>a</sup>	Stirring processes	Reference	Sediment load $\alpha$ <sup>b</sup>
Bijker bedload	Waves/currents	<i>Bijker</i> [1968]	$\frac{A_B 0.40 d_{50}}{\ln(12D/\Delta_r)} \exp\left(\frac{-0.27g(s-1)d_{50}}{\mu_B(u_*^2 + 0.016 u_*^2)}\right)$
Engelund-Hansen	Currents	<i>Engelund and Hansen</i> [1972]	$\frac{0.04 c_D^{3/2}  \bar{v} ^4}{g^2 (s-1)^2 d_{50}}$
Ackers-White	Currents	<i>Ackers and White</i> [1973]	$C_{AW} d_{35} \left(\frac{ \bar{v} }{u_*}\right)^n \left(\frac{F_{AW} - A_{AW}}{A_{AW}}\right)^m$
Bailard bedload	Waves <sup>c</sup>	<i>Bailard and Inman</i> [1981]	$\frac{\epsilon_B c_f u_b^2}{g(s-1) \tan \phi_i}$
Bailard suspended load	Waves <sup>c</sup>	<i>Bailard and Inman</i> [1981]	$\frac{\epsilon s c_f u_b^3}{g(s-1) w_s}$
Grass	Waves/currents	<i>Grass</i> [1981]	$A_G \left( \bar{v} ^2 + \frac{0.08}{c_D} u_b^2\right)^{(n_G-1)/2}$
Soulsby-van Rijn	Waves/currents	<i>Soulsby</i> [1997]	$A_s \left(\left( \bar{v} ^2 + \frac{0.018}{c_D} u_b^2\right)^{1/2} - u_c\right)^{2.4}$

<sup>a</sup> If not mentioned in the name of the formula, it describes total load transport (bedload plus suspended load).

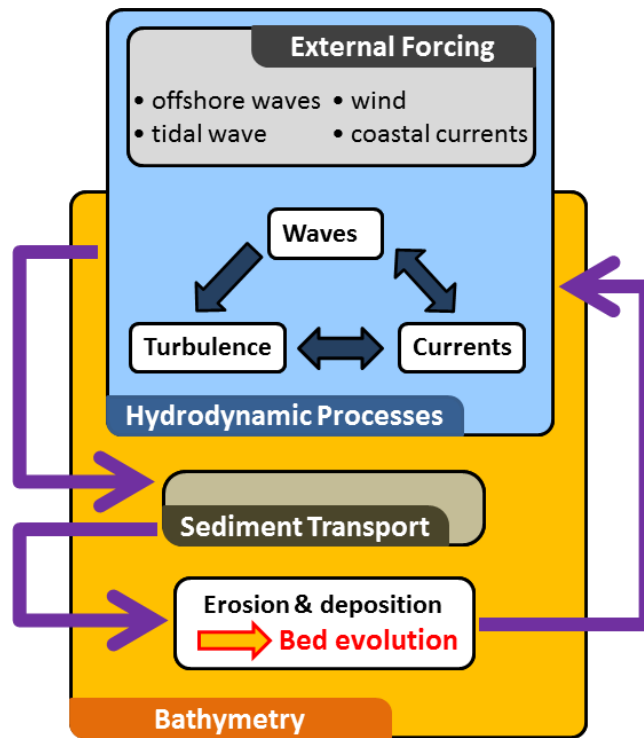
<sup>b</sup> In the formulas,  $d_n$  is the grain diameter for which n% of the grains are finer,  $s$  is the density ratio of grain and water,  $u_*$  is the total friction velocity due to current alone,  $c_D$  is the drag coefficient applicable to depth-averaged current and  $c_f$  is the drag coefficient applicable to wave orbital velocities at the bed. More details and the meaning of the other variables and parameters can be found in *Soulsby* [1997].

<sup>c</sup> In the two Bailard formulas, wave orbital velocity amplitude is assumed to be much larger than depth-averaged currents, which is only valid for weak currents.

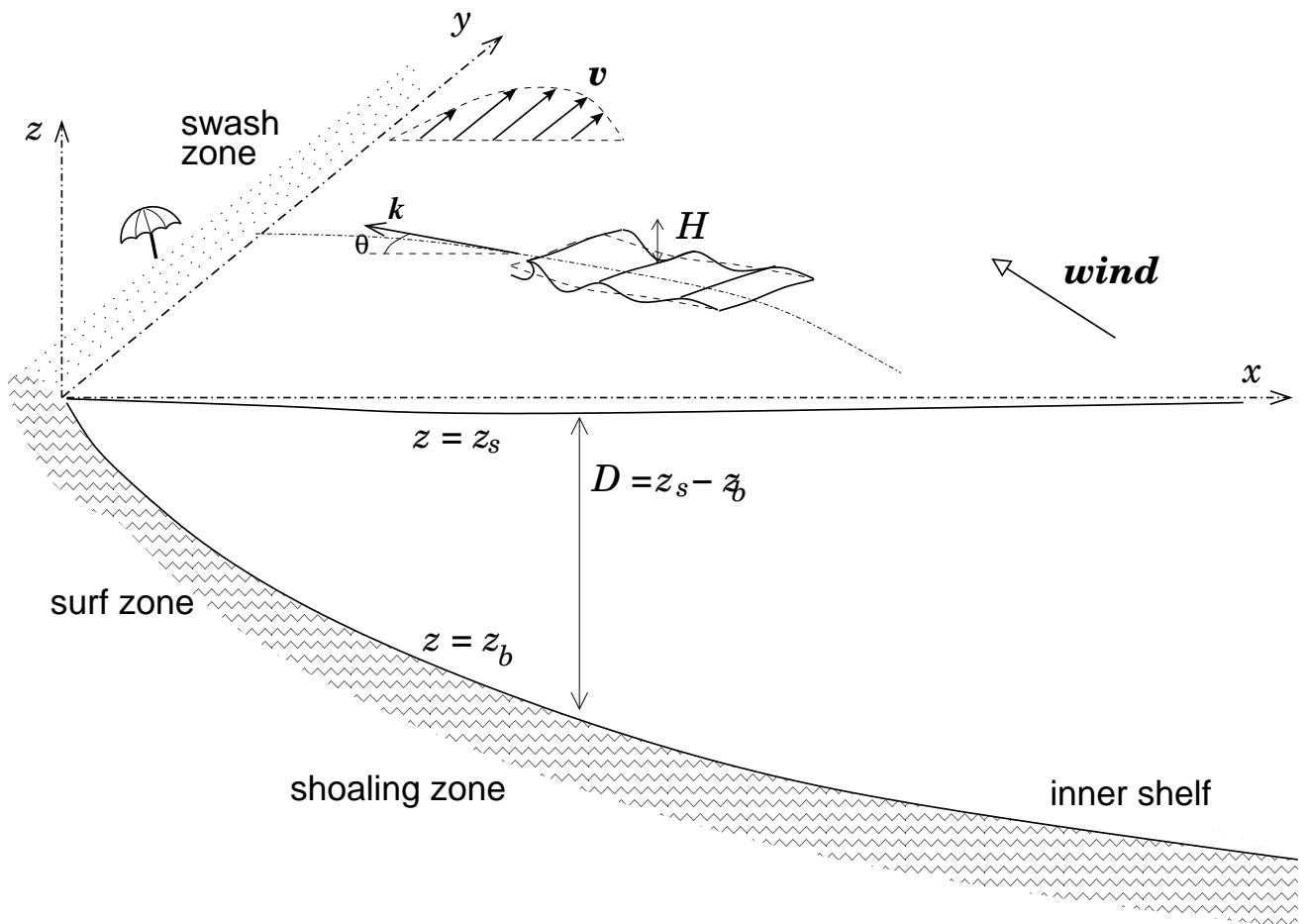


**Figure 2.** Illustration of the incoming waves (panel a), depth-averaged sediment concentration profile (DASC, panel b), and bed level (panel c) on the coastal zone. Satellite image (panel d) of the coastal zone in front of Duck, North Carolina, USA (panel d, source: Google Earth, image from Terrametrics and DigitalGlobe). Superimposed to the satellite image, examples of coastal features described in the manuscript: beach cusps (panel e, with a bathymetry from a nearby island; adapted from *Coco et al.* [2004b]), surf zone transverse and crescentic bars (panels f and g, respectively, with time-averaged video images from the same Duck beach; source: Dr. N. Plant, from U. S. Geological Survey), and shoreface-connected sand ridges (panel h, with a bathymetry in front of Long Island, New York; source: NOAA National Geophysical Data Center, USA). Each feature figure has its own scale. The transverse bars and the ridges are up-current oriented.

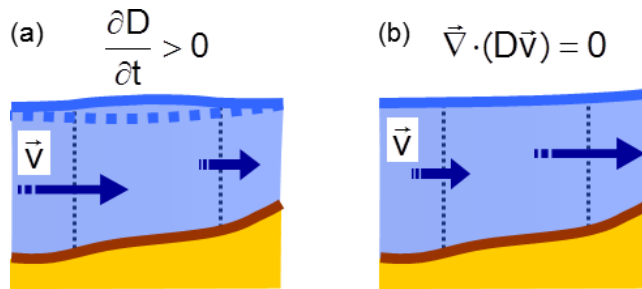




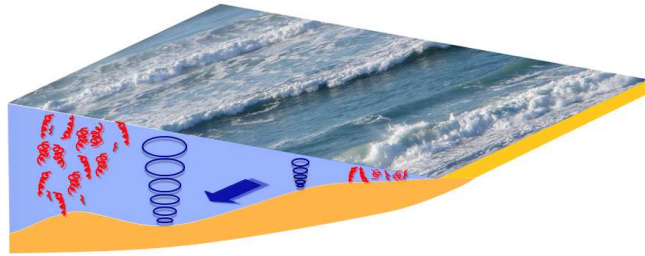
**Figure 3.** Sketch of the general framework of coastal morphodynamic models.



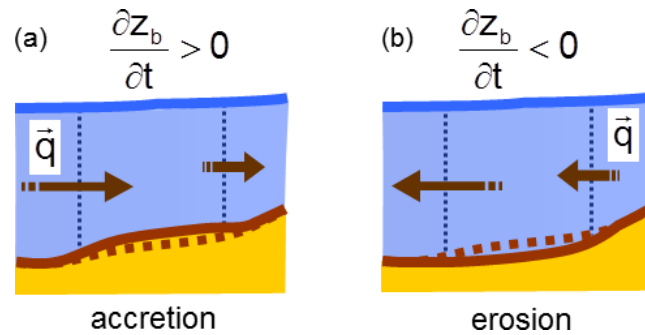
**Figure 4.** Schematic drawing of the coastal system and its four different zones (these zones are defined in the Glossary). The coordinate system and some important variables used in this contribution are also plotted (the meaning of the different symbols is described in the Notation).



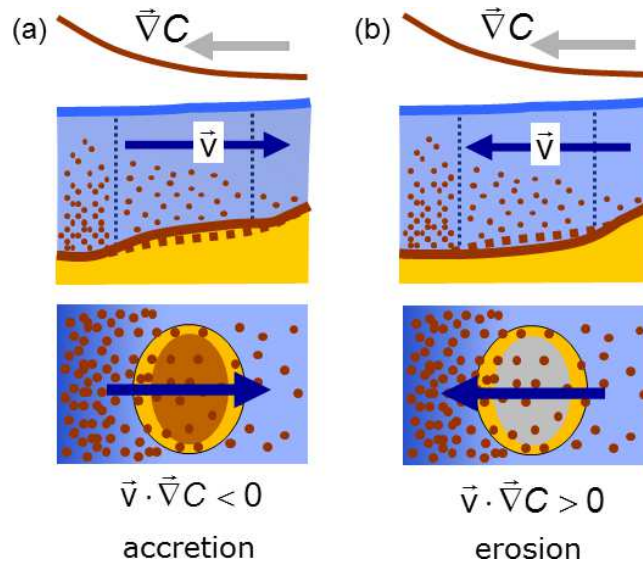
**Figure 5.** Sketch of two example processes that can be explained with the mass conservation equation: (a) case where a convergence of water flux leads to an increase in water depth, and (b) case where the quasi-steady hypothesis is assumed and a decrease in water depth leads to an acceleration of the flow.



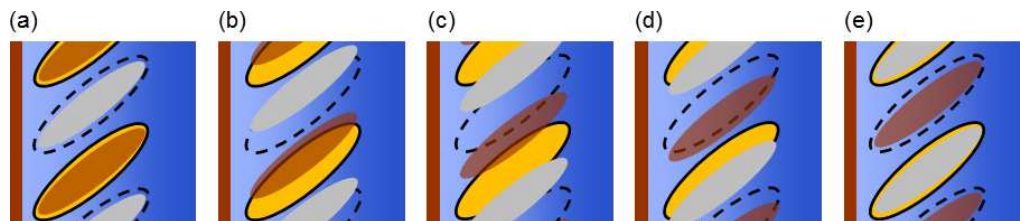
**Figure 6.** Sources of sediment stirring: wave orbital velocity at the bed (blue circles), depth-averaged current (blue wide arrow) and turbulent vortices (red swirls).



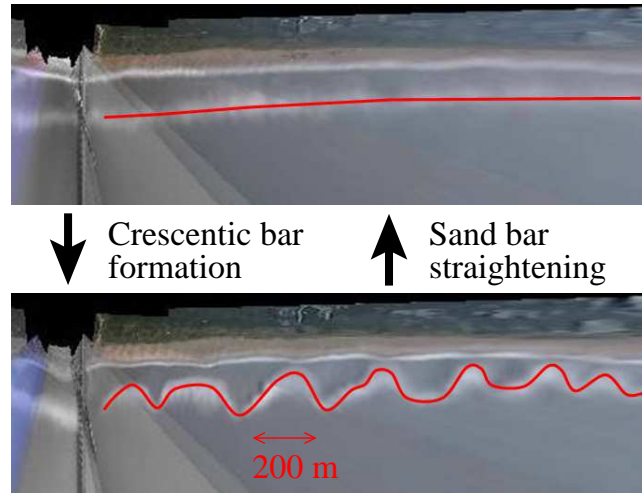
**Figure 7.** Sketches to interpret the sediment conservation equation in case of (a) bed accretion due to convergence of sediment transport, and (b) bed erosion due to divergence of sediment transport.



**Figure 8.** Sketch to interpret the erosion/deposition processes from the nonlinear BEE (8) in case of (a) bed accretion produced by a current with a component that opposes the gradient in  $C$  and (b) bed erosion produced by a current with a component in the direction of the gradient in  $C$ . The lower panels show a plan-view of the water column, with a bump on the bed plotted in yellow and brown/grey colors representing accretion/erosion of the bump.



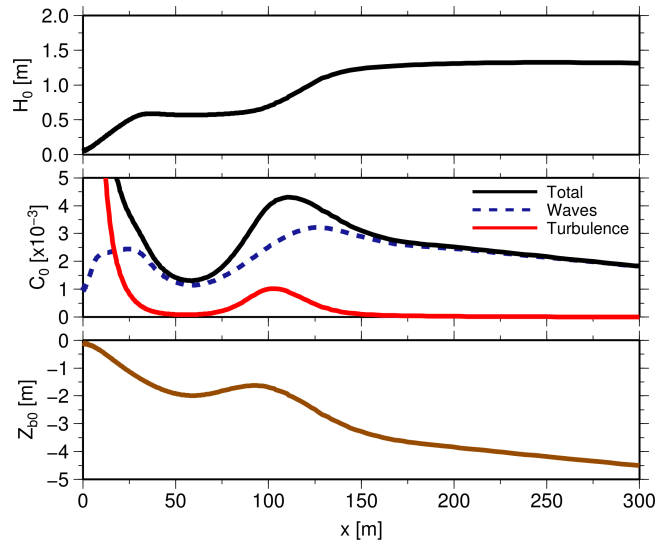
**Figure 9.** Erosion/deposition patterns producing (a) pure growth, (b) growth and down-drift migration, (c) pure down-drift migration, (d) decay and down-drift migration, and (e) pure decay of a morphologic feature consisting of an alongshore rhythmic system of bars and troughs. The bars are plotted in yellow and the brown (grey) colors represent the areas with accretion (erosion).



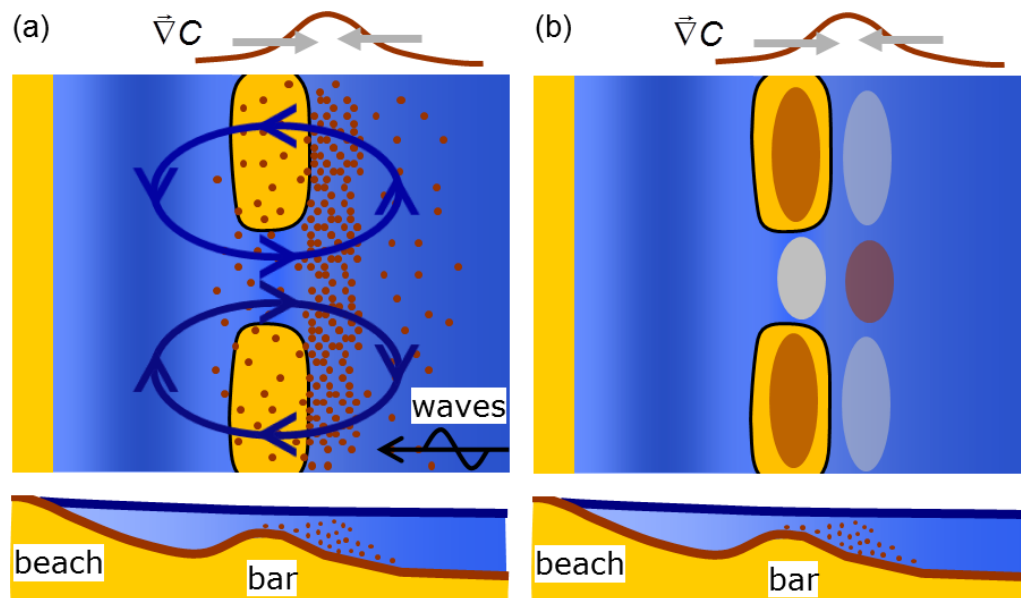
**Figure 10.** Time exposure images of a straight bar configuration (top) and a crescentic bar configuration (bottom), Duck, North Carolina, USA. The coast is at the top of the images. Courtesy of Prof. R. Holman, Oregon State University. Figure adapted from *Garnier et al.* [2013].

**TABLE 3.** Classification of observed transverse bars, following *Pellón et al.* [2014], depending on the wave energy environment, their length scales (wavelength and cross-shore span) and their aspect ratio (wavelength divided by cross-shore span).

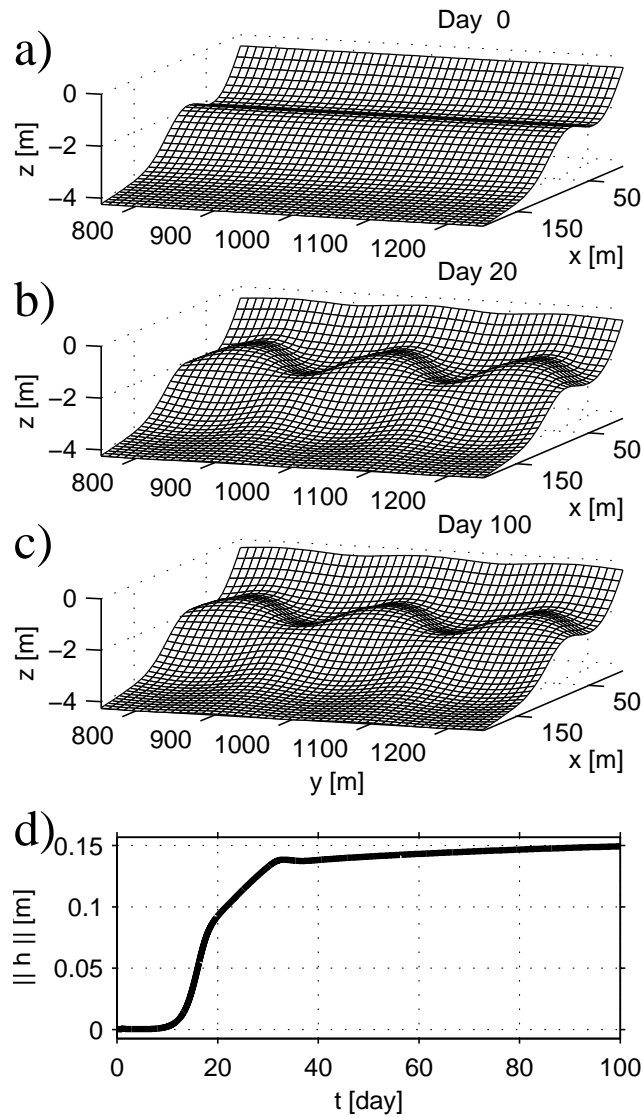
Type	Wave energy	Wavelength	Cross-shore span	Aspect ratio	References
(1) TBR bars	Medium-High	75–750 m	< 150 m	< 0.5	<i>Hunter et al.</i> [1979] <i>Wright et al.</i> [1979] <i>Lafon et al.</i> [2004] <i>MacMahan et al.</i> [2005] <i>Holman et al.</i> [2006] <i>Castelle et al.</i> [2006]
(2) Medium-energy finger bars	Medium	15–200 m	< 100 m	~ 1	<i>Konicki and Holman</i> [2000] <i>Ribas and Kroon</i> [2007] <i>Ribas et al.</i> [2014]
(3) Low-energy finger bars	Low	15–80 m	40–250 m	2–3	<i>Falqués</i> [1989] <i>Bruner and Smosna</i> [1989] <i>Eliot et al.</i> [2006] <i>Pellón et al.</i> [2014]
(4) Large-scale finger bars	Low-Medium	50–500 m	~ 1000 m	2–4	<i>Niederoda and Tanner</i> [1970] <i>Gelfenbaum and Brooks</i> [2003] <i>Levoy et al.</i> [2013]



**Figure 11.** Modeled wave height  $H_0$  (upper panel), depth-averaged sediment concentration  $C_0$  (middle panel), and bed level  $z_{b0}$  (lower panel). The variables have been computed with the model by *Ribas et al.* [2011] using normal wave incidence with an offshore wave height of 1.5 m and a wave period of 8 s. The  $C_0$  has been calculated with the Soulsby-van Rijn formula (Table 2), extended to include the stirring by turbulent vortices. In the middle panel, the dashed-blue line is the  $C_0$  due to the stirring by wave orbital velocities alone (the only stirring process in the original Soulsby-van Rijn formula), the red line is the  $C_0$  obtained due only to the turbulent vortices and the black line is the total  $C_0$  obtained when both stirring processes are accounted for.

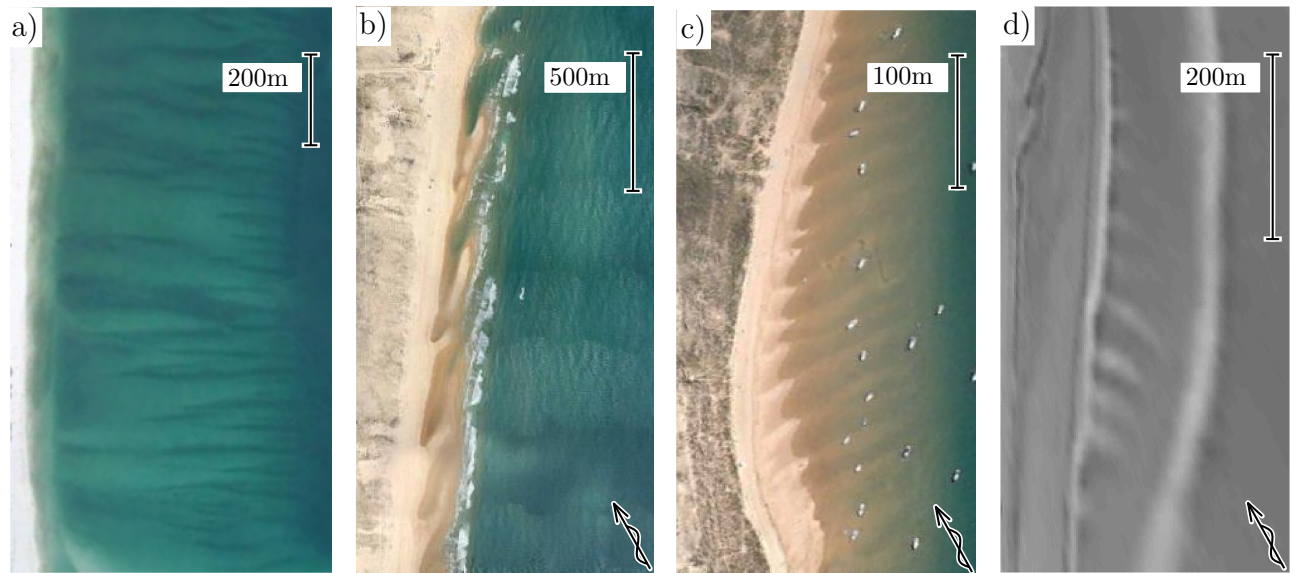


**Figure 12.** Sketch of the formation mechanism of crescentic bars from a shore-parallel bar for normal wave incidence. Left: Gradients in DASC in surf zones with a shore-parallel bar (as shown in Figure 11) and rip current circulation induced by an incipient crescentic bar. Right: Morphologic effect of the joint action of the gradients of DASC and the rip current circulation (brown are accretion areas and grey are erosion areas).



**Figure 13.** Modeled formation and finite-amplitude behavior of a crescentic bar system. Modeled bathymetry at (a)  $t = 0$ , (b)  $t = 20$  d and (c)  $t = 100$  d, and (d) time evolution of bar amplitude  $\|h\| = \bar{h}^2^{1/2}$ . Figure adapted from *Garnier et al.* [2008].



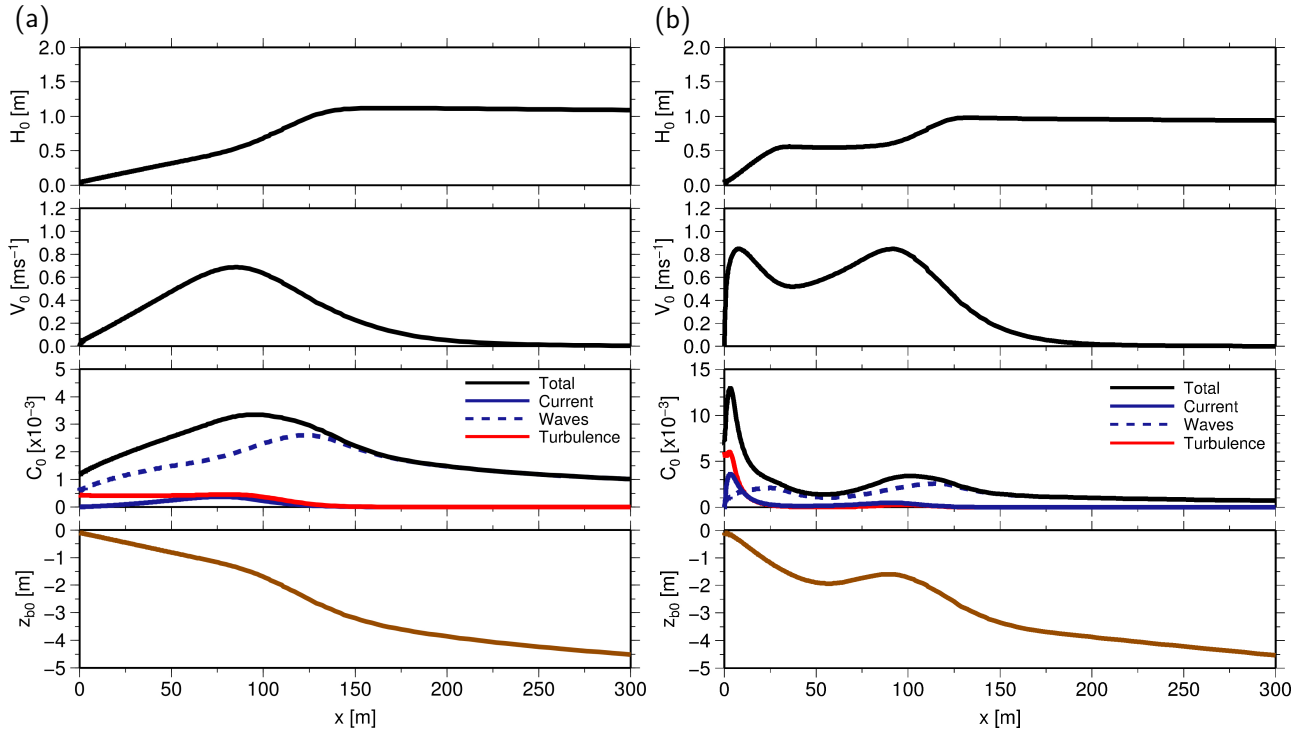


**Figure 14.** Examples of observed transverse bars with different orientations: (a) shore-normal large-scale finger bars at Anna Maria Island, USA (source: Google Earth, image from U.S. Geological Survey and USDA Farm Service Agency) [details in *Gelfenbaum and Brooks*, 2003], (b) down-current oriented TBR bars at the French Atlantic coast, France (source: Google Earth) [details in *Castelle et al.*, 2006], (c) down-current oriented low-energy finger bars at El Puntal, Santander, Spain (source: Google Earth) [details in *Pellón et al.*, 2014] and (d) up-current oriented medium-energy finger bars at Noordwijk, the Netherlands (time exposure video image) [details in *Ribas and Kroon*, 2007].

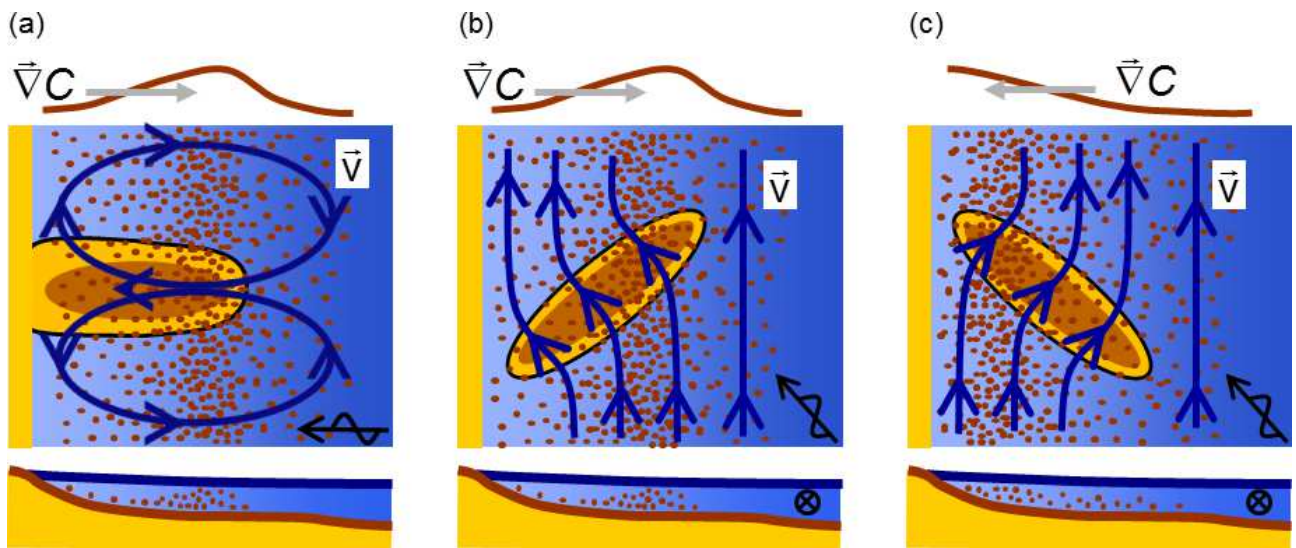
**TABLE 4.** Classification of observed transverse bars depending on their orientation, when the main driving processes (and the bar orientation) have been identified.

The mean beach slope below the bars is also indicated.

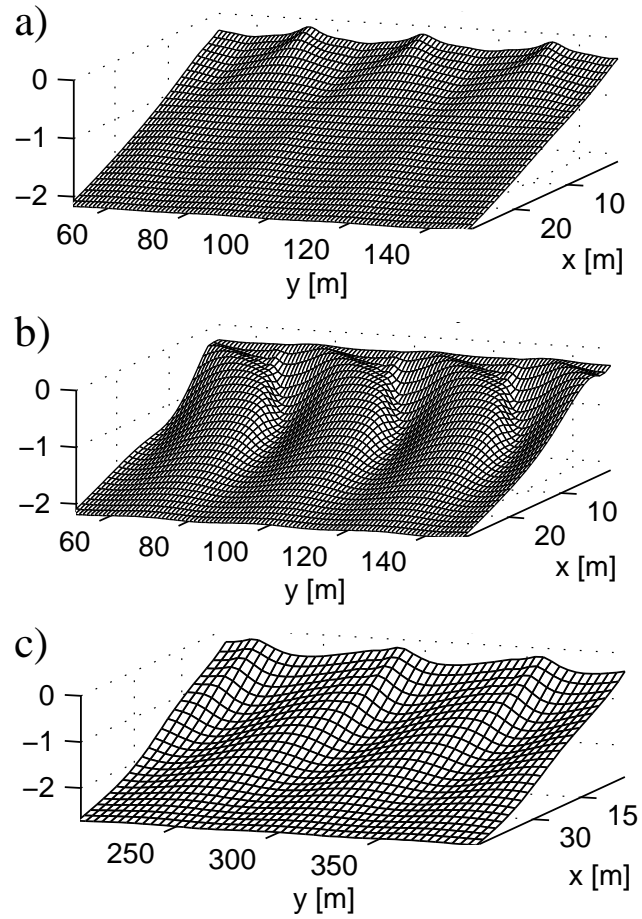
Orientation	Type	Beach slope	Main driving processes	References
Shore-normal	(1)	0.01	Wave breaking	<i>MacMahan et al.</i> [2005]
	(4)	0.003	Wave refraction	<i>Niederoda and Tanner</i> [1970]
Down-current	(1)	0.01	Wave breaking	<i>Castelle et al.</i> [2006]
			Wave-driven alongshore current	
	(3)	0.015	Wind-waves incoming obliquely	<i>Bruner and Smosna</i> [1989] <i>Pellón et al.</i> [2014]
Up-current	(2)	0.02–0.04	Wave breaking	<i>Ribas and Kroon</i> [2007]
			Wave-driven alongshore current	<i>Ribas et al.</i> [2014]



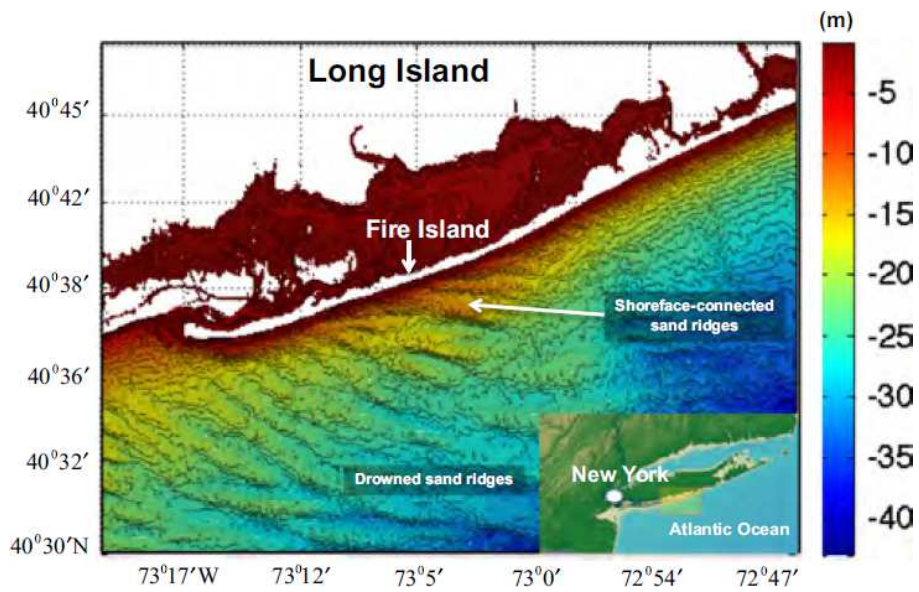
**Figure 15.** Modeled wave height  $H_0$  (upper panel), alongshore current,  $V_0$ , depth-averaged sediment concentration  $C_0$  (middle panel), and bed level  $z_{b0}$  (lower panel) in the case of (a) terraced profile and (b) profile with a shore-parallel bar. The variables have been computed with the model by Ribas *et al.* [2011] using oblique wave incidence with an offshore wave height of 1 m, an offshore wave angle of (a)  $20^\circ$  and (b)  $50^\circ$  (at 28 m depth), and a wave period of 8 s. The  $C_0$  has been calculated with the Soulsby-van Rijn formula (Table 2), extended to include the stirring by turbulent vortices. In the middle panels, the solid-blue line is the  $C_0$  due to the stirring by depth-averaged current alone, the dashed-blue line is the  $C_0$  due to the stirring by wave orbital velocities alone, the red line is the  $C_0$  obtained due only to the turbulent vortices and the black line is the total  $C_0$  with the three processes included.



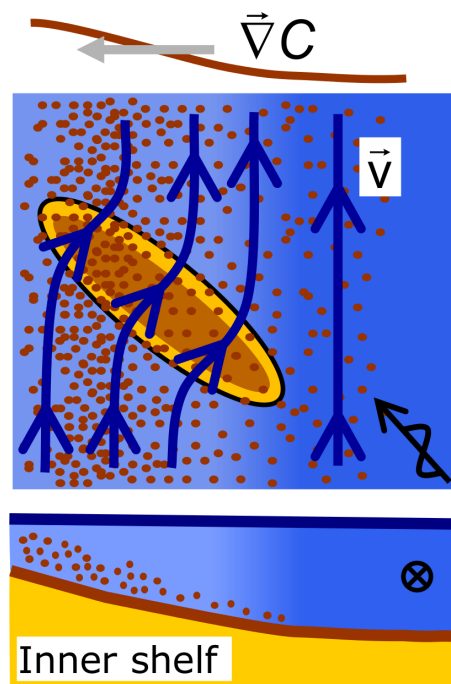
**Figure 16.** Sketch of the formation mechanism (gradients of DASC, horizontal circulation induced by the growing pattern in blue streamlines and accretion areas in brown) of transverse bars with (a) shore-normal, (b) down-current and (c) up-current orientations.



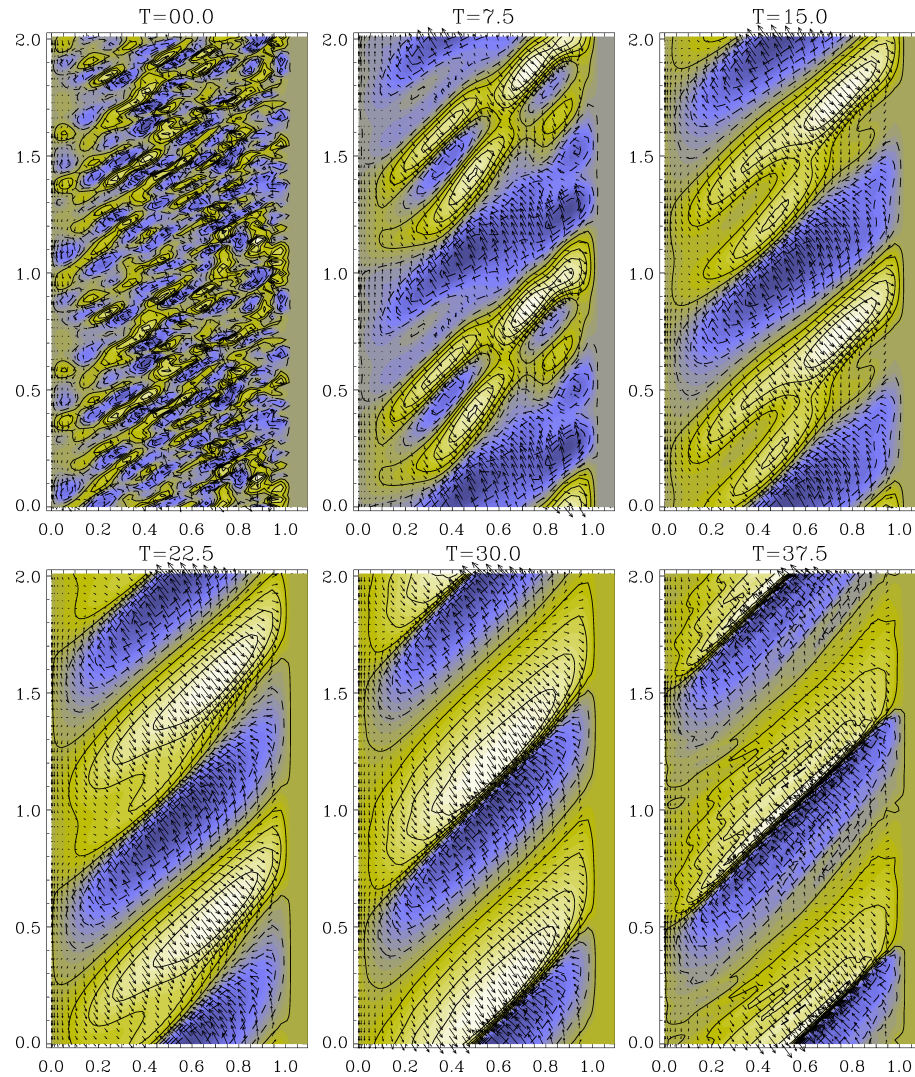
**Figure 17.** Simulations of (a) shore-normal (b) down-current (c) up-current oriented transverse bars obtained with a nonlinear model. Results of the bathymetry after several day of simulations, all the simulations have started with an alongshore uniform planar beach. Figures adapted from *Garnier et al.* [2006].



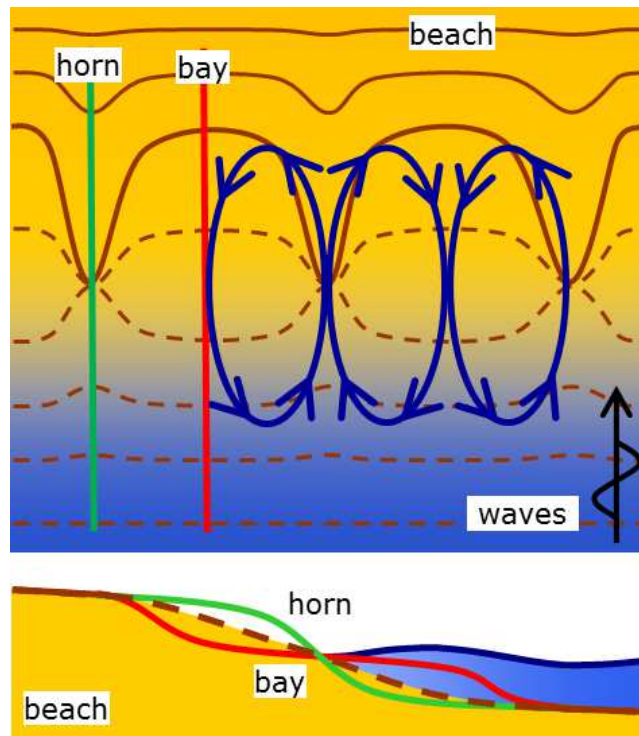
**Figure 18.** Bathymetric map of the Long Island continental shelf. Figure from *Nnafie et al.* [2014a].



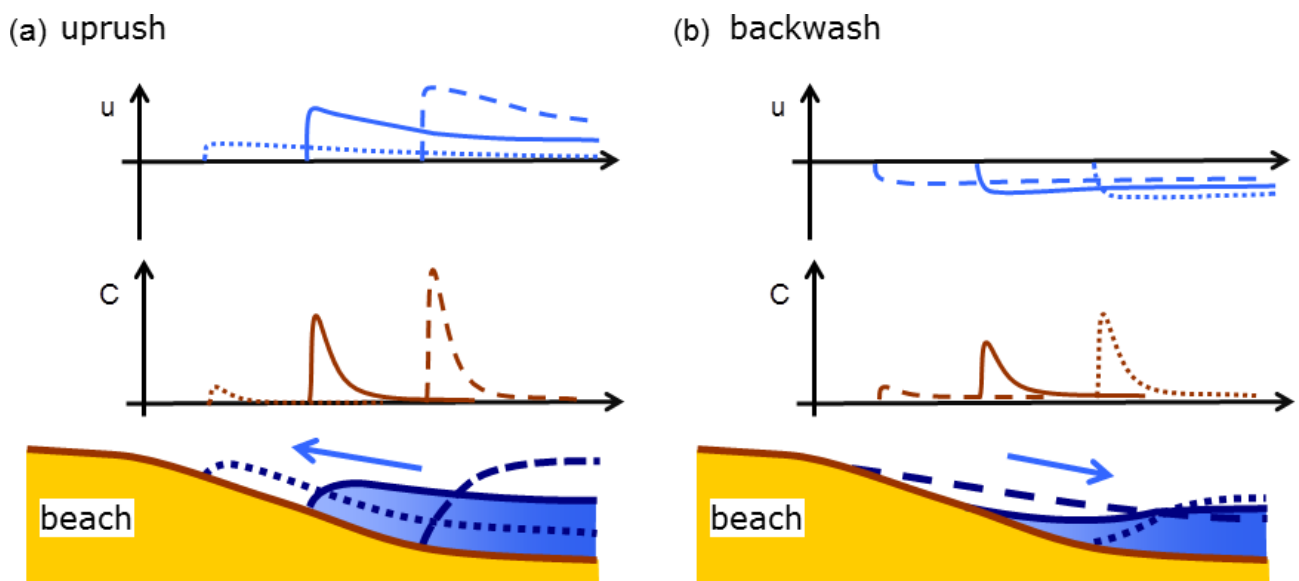
**Figure 19.** Sketch of the formation mechanism of shoreface-connected sand ridges with the gradients of DASC, the horizontal circulation induced by the growing pattern (blue streamlines), and the accretion areas (in brown).



**Figure 20.** Contour plots of perturbations in bed level at different times (dimensional times  $T = 0, 300, 600, 900, 1200, 1500$  yr) during a model simulation of shoreface-connected sand ridges. Figure adapted from *Calvete et al.* [2002]. The shoreline is located on the left of the panels.

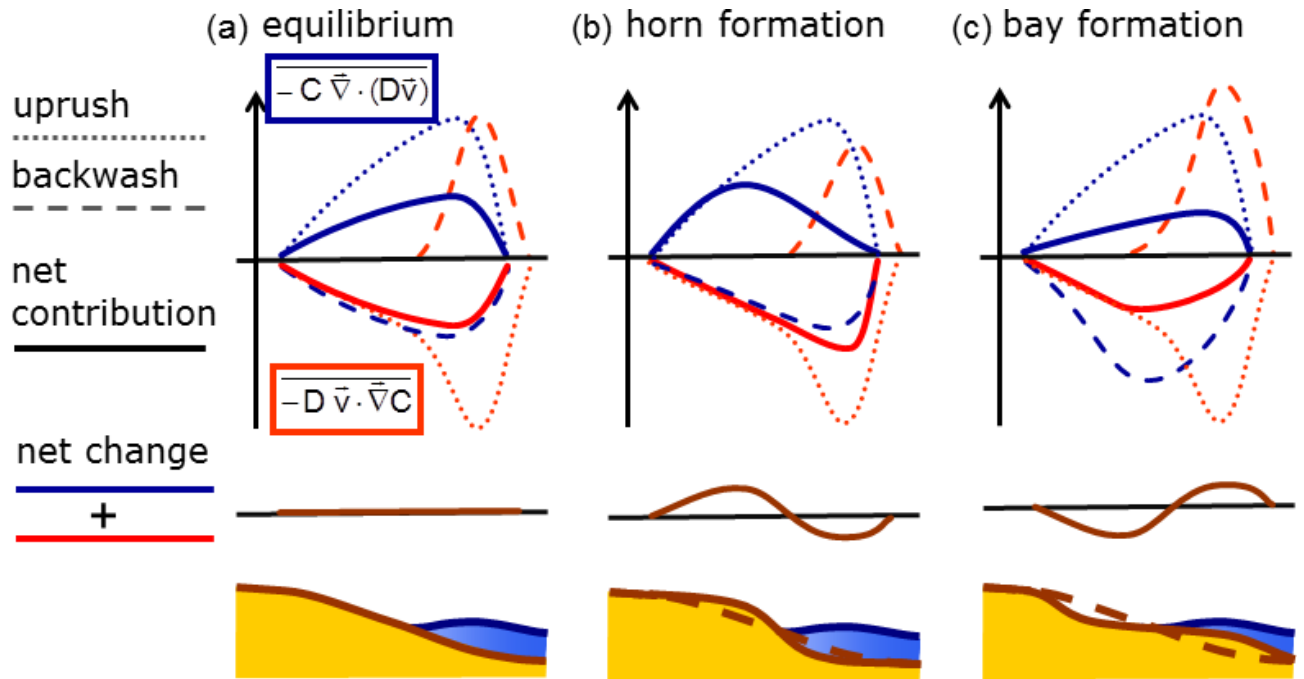


**Figure 21.** Schematic diagram of a cusped bathymetry, showing cusp horns and bays, together with the wave-averaged current circulation.

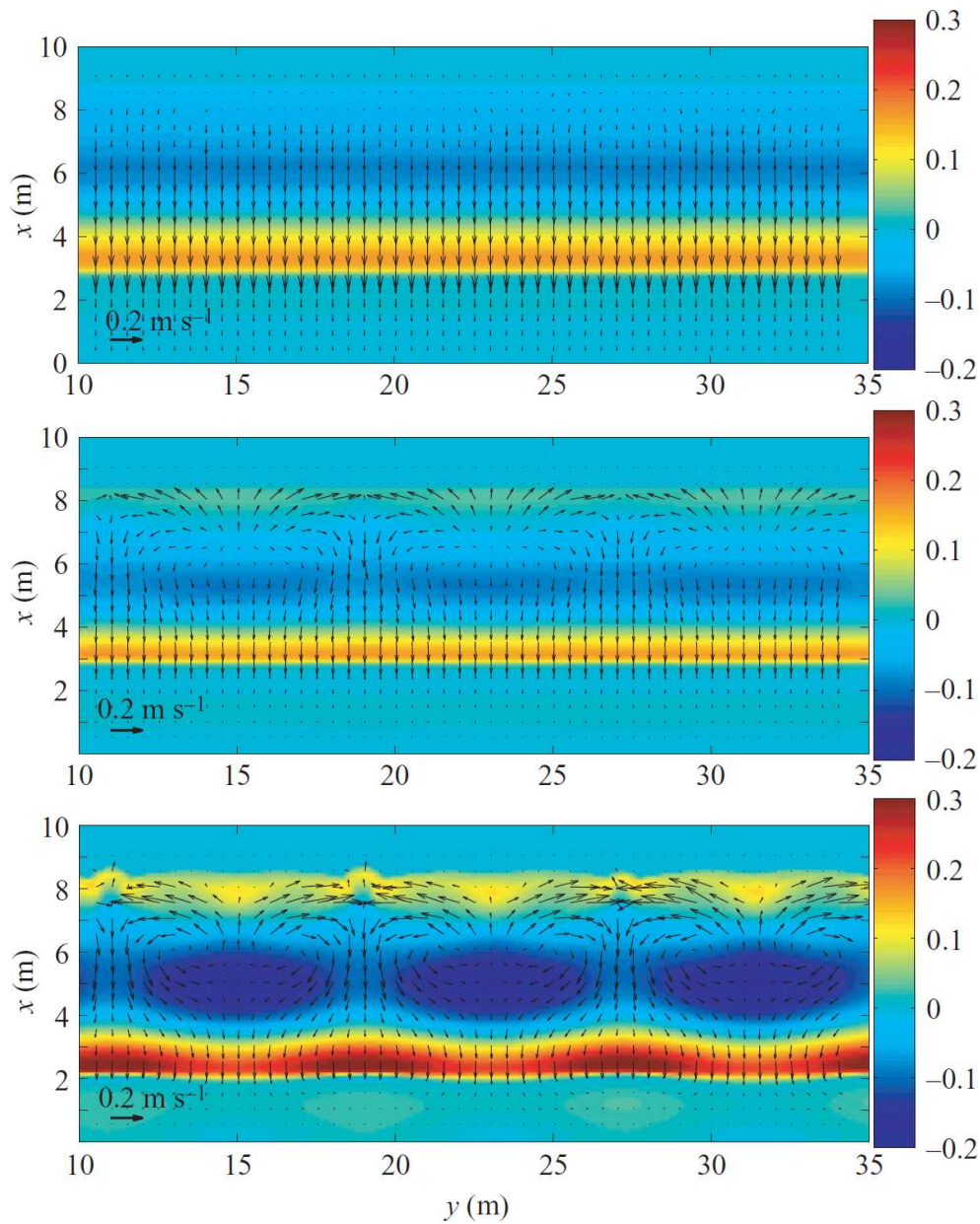


**Figure 22.** Schematic diagram showing cross-shore current (top), DASC (middle) and free surface (bottom) versus the cross-shore distance at different stages of the uprush (a) and backwash (b) in a swash excursion.

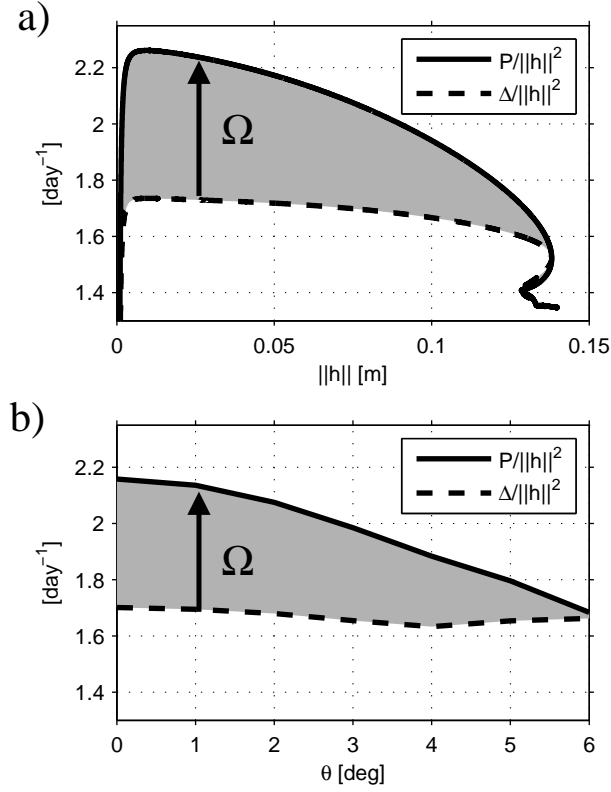




**Figure 23.** Schematic diagram showing the pattern of erosion and deposition in the swash zone in case of (a) alongshore uniform conditions, (b) horn location in presence of cusps and (c) bay location in presence of cusps. In b) and c), the dynamics is illustrated by altering backwash only: diminished at horns and enhanced at embayments. The upper plots show  $\overline{-C \vec{\nabla} \cdot (D\vec{v})}$  and  $\overline{-D\vec{v} \cdot \vec{\nabla} C}$ , terms of equation (11), for uprush and backwash (overbars here denote time averages over only those phases of the swash), and their net contribution during the whole wave cycle, versus the cross-shore distance. The pattern of erosion / deposition produced by the sum of the net contributions of these two terms is shown in the middle plots, and the corresponding change in bed level is shown in the lower plots.



**Figure 24.** Model simulation showing bed level change in colors (m), relative to a planar beach, and wave-averaged current vectors at  $t = 0$  (upper panel),  $t = 20$  wave periods (middle panel) and  $t = 100$  wave periods (lower panel). Figure adapted from *Dodd et al.* [2008]. The shoreline is located on the top of the panels.



**Figure 25.** Global analysis for finite amplitude behavior of crescentic bars. (a) Normal wave incidence, saturation mechanism. Production  $\mathcal{P}/\bar{h}^2$  and damping  $\Delta/\bar{h}^2$  as a function of  $\|h\| = \bar{h}^{1/2}$ . Figure adapted from *Garnier et al.* [2010]. (b) Inhibition of rip channel formation for oblique waves. Maximum production and maximum damping as a function of the wave incidence angle ( $\theta$ , at 4.5 m depth). Figure adapted from *Garnier et al.* [2013]. Note that the growth rate  $\Omega = \mathcal{P}/\bar{h}^2 - \Delta/\bar{h}^2$ .