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# 3D wave-resolving simulation of sandbar migration

Patrick Marchesiello<sup>a,\*</sup>, Julien Chauchat<sup>b</sup>, Hassan Shafiei<sup>b</sup>, Rafael Almar<sup>a</sup>, Rachid Benshila<sup>c</sup>, Franck Dumas<sup>d</sup>, Laurent Debreu<sup>e</sup>

<sup>a</sup>*IRD/LEGOS, Toulouse, France*

<sup>b</sup>*Univ. Grenoble Alpes, CNRS, Grenoble INP, LEGI, 38000 Grenoble, France*

<sup>c</sup>*CNRS/LEGOS, Toulouse, France*

<sup>d</sup>*SHOM, Brest, France*

<sup>e</sup>*Univ. Grenoble Alpes, Inria, CNRS, Grenoble INP, LJK, 38000 Grenoble, France*

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## Abstract

The problem of sandbar migration on the storm timescale is revisited with a 3D wave-resolving hydro-sedimentary model. The latter presents an intermediate approach between expensive wave-resolving two-phase flow models and highly parametrized wave-averaged models. Innovative features include the use of weakly compressible assumptions in the hydrodynamics and morphological acceleration of bed changes to speed up numerical simulations. The model accurately simulates the successive offshore and onshore bar migration observed in a large-scale flume experiment in response to wave forcing representing storm and post-storm (recovery) conditions. The diagnosis of sand transport and the analysis of an ensemble-averaged asymmetric wave cycle reveal the migration mechanisms in each phase. In all cases, sediment resuspension is impacted by breaker-induced turbulence, while sediment transport and bed evolution are primarily the result of the undertow distribution — the breaker-induced seaward undercurrent — across the sandbar. There is also a significant contribution from asymmetric wave-related onshore fluxes, due to greater mobilization and currents during the wave crest period.

*Keywords:* nearshore, sandbar migration, wave-resolving model, wave asymmetry, sediment transport

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\*Corresponding author

*Email address:* `patrick.marchesiello@ird.fr` (Patrick Marchesiello)

## 1. Introduction

The presence of nearshore sandbars are ubiquitous on natural beaches, which are prime areas for morphological changes, depending on wave and sediment characteristics (Roelvink and Stive, 1989; Thornton et al., 1996; Elgar et al., 2001; Almar et al., 2010; Grasso et al., 2011). However, they present a challenge to our understanding of sediment transport processes, which is often based on the study of non-breaking waves (Scott et al., 2009; van der Zanden et al., 2017; Lim et al., 2020). Numerical modeling can often be used to study these processes in a timely and cost-effective manner, but it typically relies on many assumptions and unknown parameters, so many aspects of bar dynamics remain unclear (Roelvink et al., 2012).

In particular, the role of wave asymmetry on bar migration is actively debated (Grasso et al., 2011). Velocity skewness (sharp, high crests and broad, shallow troughs) and asymmetry (saw tooth-type waves) are considered to be responsible for sediment transport in the direction of wave propagation (onshore). The basic idea is that fast crest velocities in the onshore direction would mobilize and transport more sediment than the offshore-directed trough velocities (Hsu and Hanes, 2004). Wave asymmetry with its steep front would more efficiently mobilize sediments than skewness (Drake and Calantoni, 2001), possibly also reducing the phase lag between mobilization and transport (Dohmen-Janssen et al., 2002). Breaking waves may also contribute to bed shear stress and thus mobilization asymmetry due to higher turbulence intensity during the wave crest period (Ting and Kirby, 1994), although this process is disputed (Scott et al., 2009). In addition, in contrast to the surface elevation, the low-frequency components of the orbital velocity (subharmonics) can contribute to a negative subsurface skewness in the most energetic wave groups (Alberello et al., 2016). Interestingly, it is generally assumed that bar migration results from a trade-off between onshore and offshore fluxes (e.g., Grasso et al. 2011), thus confusing the notions of convergent and confluent fluxes. A comprehensive conceptual model that could describe the interaction between onshore and offshore transport processes in bar migration, allowing for variable wave forcing, is still needed. The present work is a step in that direction.

29 It will confirm the existence of wave-related onshore transport (especially in the accretion  
30 phase) but will show that sediment transport around the sandbar depends primarily on the  
31 cross-shore distribution of the undertow, which varies with wave height.

32 Two-phase (water and sediment) flow models applied to individual wave breaking can be  
33 used to describe the hydro-morphodynamics in great detail (Cheng et al., 2017; Kim et al.,  
34 2019). This approach does not require parameterization of sand resuspension or solving the  
35 Exner equation for bed evolution, but it does require grid resolution on the order of grain  
36 size, and constitutive laws for particle-particle and fluid-particle interactions. As a result,  
37 these models are theoretically difficult, algorithmically complex, and computationally very  
38 expensive (Chauchat et al., 2017). Therefore, the study of sandbar formation and migra-  
39 tion has been done in the past with much simpler depth-averaged and/or wave-averaged  
40 models (Watanabe, 1982; Stive, 1986; Roelvink and Stive, 1989; van Rijn et al., 2003; Long  
41 et al.), which are used operationally today — XBeach being a classic example (Rafati et al.,  
42 2021). These models rely heavily on parametrizations to account for both onshore and off-  
43 shore sandbar migration, i.e., a balance between offshore transport by the undertow and  
44 onshore transport by skewed/asymmetric waves and wave streaming. Depth-dependent  
45 phase-averaged nearshore models (McWilliams et al., 2004) have improved the represen-  
46 tation of the undertow, but the effects of wave asymmetry/skewness remain parametrized  
47 as bedload transport in the wave boundary layer (e.g., van der Werf et al. 2015 for Delft3D;  
48 Kalra et al. 2019 for ROMS; Shafiei et al. 2022 for CROCO). Recently, empirical formula-  
49 tions have improved, using the concept of half-wave cycles (Dibajnia and Watanabe, 1992;  
50 van der A et al., 2013), and are implemented with mixed success (Veen, 2014; Schnitzler,  
51 2015).

52 As an intermediate approach between expensive wave-resolving two-phase flow models  
53 and highly parametrized wave-averaged models, we propose to apply a wave-resolving, free-  
54 surface, coupled hydro-sedimentary model. CROCO (Coastal and Regional Ocean Commu-  
55 nity model) has recently been adapted to nearshore wave dynamics and circulation problems  
56 (Marchesiello et al., 2021). This cost-effective 3D wave-resolving model can explicitly simu-  
57 lated the vertical profiles of oscillatory cross-shore sediment transport, allowing assessment

58 of the combined effects of mean undertow and asymmetric waves around a sandbar. This  
59 paper presents a brief overview of the CROCO coupled hydro-sedimentary model, before  
60 discussing its application to a large-scale flume experiment (LIP11D), and an analysis of off-  
61 shore and onshore sandbar migration under wave forcing representing storm and post-storm  
62 (recovery) conditions.

## 63 **2. Model description**

64 The development of 3D wave-resolving models to study nearshore dynamics in realistic  
65 environments has been limited by computing resources and the cost-effective use of these  
66 resources. Early applications starting in the 1990s used the volume-of-fluid (VOF) method  
67 for free-surface tracking (e.g., Lin and Liu 1998; Watanabe and Saeki 1999; Derakhti and  
68 Kirby 2014; Larsen et al. 2020). This model type with Cartesian coordinate, where the  
69 free surface crosses computational cells arbitrarily, is computationally intensive and does  
70 not allow the pressure boundary condition to be precisely applied on the free surface, which  
71 affects the overall accuracy of the model (Chen et al., 2018). More recently, several 3D wave-  
72 resolving, free-surface and terrain-following models based on Reynolds-Averaged Navier-  
73 Stokes (RANS) equations have emerged for the nearshore zone, e.g., SWASH (Zijlema et al.,  
74 2011), NHWAVE (Derakhti et al., 2016), and CROCO (Marchesiello et al., 2021). In this  
75 case, the explicit overturning of the free surface is excluded and the breaking wave is modeled  
76 with a single-valued free surface which follows a shock process and resembles a dissipating  
77 bore. Despite the absence of explicit overturning (replaced by parametrized turbulence),  
78 these models can be accurate as well as computationally efficient (orders of magnitude  
79 cheaper) in the study of waves and wave-driven mean and transient circulation.

80 CROCO ([www.croco-ocean.org](http://www.croco-ocean.org)) belongs to this class of models but, contrary to the  
81 other attempts, it solves the compressible Navier-Stokes equations (Auclair et al., 2018).  
82 A compressible approach preserves the hyperbolic nature of Navier-Stokes equations and  
83 does not require a global elliptic solver with incremental pressure corrections to ensure the  
84 incompressible mass balance. As a result, it avoids splitting errors between pressure and  
85 velocity and approximations made on free-surface conditions (Zijlema et al., 2011; Derakhti

86 et al., 2016), thereby preserving amplitude and nonlinear dispersive properties of surface  
 87 waves. In the same time, the absence of global computations by an elliptic solver makes  
 88 optimization and parallelization procedures much more efficient (excellent scalability even  
 89 on massively parallel computers). The cost of solving acoustic waves is managed with a  
 90 time-splitting technique and semi-implicit time discretization.

### 91 2.1. Model equations

92 The full set of Navier-Stokes equations for a free-surface ocean model is explicitly in-  
 93 tegrated in the nonhydrostatic, non-Boussinesq (compressible) version of CROCO, built  
 94 on the code structure of ROMS primitive equation solver (Shchepetkin and McWilliams,  
 95 2005; Debreu et al., 2012). Non-Boussinesq equations include the momentum and continu-  
 96 ity equations, the free-surface kinematic condition, conservation equations for heat, salt or  
 97 other tracer  $C$  (such as sediment concentration), which reads in Cartesian coordinates:

$$\frac{\partial \rho u}{\partial t} = -\vec{\nabla} \cdot (\rho \vec{v} u) + \rho f v - \rho \tilde{f} w - \frac{\partial P}{\partial x} + \mathcal{F}_u + \mathcal{D}_u \quad (1)$$

$$\frac{\partial \rho v}{\partial t} = -\vec{\nabla} \cdot (\rho \vec{v} v) - \rho f u - \frac{\partial P}{\partial y} + \mathcal{F}_v + \mathcal{D}_v \quad (2)$$

$$\frac{\partial \rho w}{\partial t} = -\vec{\nabla} \cdot (\rho \vec{v} w) + \rho \tilde{f} u - \frac{\partial P}{\partial z} - \rho g + \mathcal{F}_w + \mathcal{D}_w \quad (3)$$

$$\frac{\partial \rho}{\partial t} = -\vec{\nabla} \cdot (\rho \vec{v}) \quad (4)$$

$$\frac{\partial \eta}{\partial t} = w|_{z=\eta} - \vec{v}|_{z=\eta} \cdot \vec{\nabla} \eta \quad (5)$$

$$\frac{\partial \rho C}{\partial t} = -\vec{\nabla} \cdot (\rho \vec{v} C) + \mathcal{F}_C + \mathcal{D}_C \quad (6)$$

98  $(u, v, w)$  are the Cartesian  $(x, y, z)$  components of vector velocity  $\vec{v}$ ;  $\eta$  is the free surface;  $P$   
 99 the total pressure;  $\rho$  the density;  $f(x, y)$  and  $\tilde{f}(x, y)$  are the traditional and non-traditional

100 Coriolis parameters, function of latitude;  $g$  is acceleration of gravity;  $\mathcal{D}_u, \mathcal{D}_v, \mathcal{D}_C$  are eddy-  
 101 diffusion terms requiring second-moment turbulence closure models (bulk viscosity terms for  
 102 acoustic wave damping are not used here);  $\mathcal{F}_u, \mathcal{F}_v, \mathcal{F}_C$  are forcing terms.

103 In this study, movements are produced along a channel by a wave generator in a homo-  
 104 geneous non-rotating fluid. In this case, the longitudinal flow equation  $v$  disappears and the  
 105 Coriolis force, the baroclinic pressure force, and all surface fluxes (forcing terms) are zero.  
 106 There is no temperature or salinity stratification so that the associated density is constant in  
 107 time and space (but not the dynamic density associated with acoustic waves). The resulting  
 108 equations are:

$$\frac{\partial \rho u}{\partial t} = -\vec{\nabla} \cdot (\rho \vec{v} u) - \frac{\partial P}{\partial x} + \mathcal{D}_u \quad (7)$$

$$\frac{\partial \rho w}{\partial t} = -\vec{\nabla} \cdot (\rho \vec{v} w) - \frac{\partial P}{\partial z} - \rho g + \mathcal{D}_w \quad (8)$$

$$\frac{\partial \rho}{\partial t} = -\vec{\nabla} \cdot (\rho \vec{v}) \quad (9)$$

$$\frac{\partial \eta}{\partial t} = w|_{z=\eta} - \vec{v}|_{z=\eta} \cdot \vec{\nabla} \eta \quad (10)$$

$$\frac{\partial \rho C}{\partial t} = -\vec{\nabla} \cdot (\rho \vec{v} C) + \mathcal{D}_C \quad (11)$$

## 109 2.2. Computational procedure

110 In the above set of equations, a relation between  $\rho$  and  $P$  is required. To that end, and  
 111 as part of a time-splitting approach, density is decomposed into a slow component and a fast  
 112 component based on a first-order linear decomposition with respect to total pressure. The  
 113 Navier-Stokes equations are then integrated with two different time steps within the time-  
 114 splitting approach inherited from ROMS. The slow mode integration includes the slow part  
 115 of the vertical momentum equation, while the fast mode integration is in 3D and includes

116 the compressible terms of the momentum and continuity equations. More details can be  
117 found in Appendix A.

118 Note that acoustic waves can become pseudo-acoustic if their phase speed  $c_s = \sqrt{\frac{\partial P}{\partial \rho}}$   
119 is artificially reduced (but not lower than the speed of the shallow water phase). In this  
120 case, high-frequency processes associated with bulk compressibility (acoustic waves) may be  
121 degraded, but an accurate solution for slower nonhydrostatic dynamics (gravity waves) can  
122 be preserved, while relaxing CFL constraints. In our nearshore applications, a  $c_s$  value of  
123 100 m/s instead of 1500 m/s makes no difference for the physical solution but allows a great  
124 reduction in the computational time (by half).

125 CROCO is discretized on a C-grid with finite-difference methods for slow and fast modes  
126 (Shchepetkin and McWilliams, 2005; Soufflet et al., 2016). In short, the slow-mode time-  
127 stepping algorithm is a Leapfrog Adams-Moulton predictor-corrector scheme, that is third-  
128 order accurate for integrating advective terms. The fast mode is integrated with a horizon-  
129 tally explicit vertically implicit (HEVI) scheme, i.e. a generalized forward-backward scheme  
130 on the horizontal and a backward Euler scheme on the vertical. The implicit backward Eu-  
131 ler method solves the terms responsible for vertical acoustic propagation (the terms of the  
132 fast mode  $w$  equation). It allows for an extended stability range and, since it is a diffusive  
133 scheme, it has a damping effect on the acoustic waves, while preserving the gravity waves  
134 (Klemp et al., 2018).

135 The spatial discretization of the horizontal and vertical advection terms uses the im-  
136 proved WENO5-Z version of the 5th-order weighted essentially non-oscillatory scheme (Borges  
137 et al., 2008), which is popular for hyperbolic problems containing both shocks and smooth  
138 structures. WENO5-Z naturally copes with dispersive numerical modes as well as shocks  
139 caused by breaking waves, with no need for *ad hoc* criteria.

### 140 2.3. Turbulence closure

141 Along with the numerical treatment of breaking waves, a  $k$ - $\epsilon$  or  $k$ - $\omega$  model, solving  
142 the closure equations for turbulent kinetic energy  $k$  and dissipation  $\epsilon$  or dissipation rate  
143  $\omega \propto \epsilon k^{-1}$ , is used as part of a Generic Length Scale (GLS) method (Warner et al., 2005).



144 In the absence of buoyancy forcing, the turbulence equations express a balance between  
 145 transport, diffusion, shear production and dissipation:

$$\frac{\partial \rho k}{\partial t} = -\vec{\nabla} \cdot (\rho \vec{v} k) + \mathcal{D}_k + \rho(P - \epsilon) \quad (12)$$

146

$$\frac{\partial \rho \epsilon}{\partial t} = -\vec{\nabla} \cdot (\rho \vec{v} \epsilon) + \mathcal{D}_\epsilon + \rho \frac{\epsilon}{k} (c_{\epsilon 1} P - c_{\epsilon 2} \epsilon) \quad (13)$$

147 OR

$$\frac{\partial \rho \omega}{\partial t} = -\vec{\nabla} \cdot (\rho \vec{v} \omega) + \mathcal{D}_\omega + \rho \frac{\omega}{k} (c_{\omega 1} P - c_{\omega 2} \epsilon) \quad (14)$$

148 The eddy viscosity  $\nu_t = c_\mu l k^{\frac{1}{2}}$  is derived from these equations, with coefficient  $c_\mu$  de-  
 149 pendent on stability functions, and mixing length  $l \propto k^{\frac{3}{2}} \epsilon^{-1}$ .  $l$  is resolution independent,  
 150 which is consistent with a RANS rather than LES approach. The shear production term  
 151 for  $k$  is  $P = 2\nu_t S_{ij} S_{ij}$ , with the mean strain rate tensor  $S_{ij} = \frac{1}{2}(\frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i})$  (using Einstein  
 152 notation). All turbulence model parameters are given in Warner et al. (2005), based on  
 153 Burchard et al. (1998) for  $k$ - $\epsilon$  and Wilcox (1988) for  $k$ - $\omega$ . The only present modification  
 154 concerns the surface and bottom mixing lengths, which are model boundary conditions given  
 155 by:  $l_s = \kappa z_{0s}$  and  $l_b = \kappa z_{0b}$ . We found that a relatively high value of  $z_{0b}$  (5 cm) is needed to  
 156 match the observed sand concentration profiles near the bottom. With the  $k$ - $\epsilon$  model, the  
 157 momentum mixing is also sensitive to  $z_{0s}$ , which needs to be increased to 0.2 m in order to  
 158 match the observed velocity profiles (see validation section 3.3). For this reason, and for its  
 159 robustness through resolutions, the  $k$ - $\omega$  model will be our standard turbulence model.

#### 160 2.4. Wave maker at offshore boundary

161 The wave maker forces a spectrum of linear waves at the offshore boundary, as in Zijlema  
 162 et al. (2011). In full 3D applications, the spectrum has frequency and directional spreading  
 163 Marchesiello et al. (2021) but in this flume experiment, waves are shore normal:

$$\eta_{bc}(t) = \sum_i a_i \cos(\omega_i t + \phi_i) \quad (15)$$

164

$$u_{bc}(z, t) = \sum_i a_i \omega_i \frac{\cosh(k_i(z + h))}{\sinh(k_i h)} \cos(\omega_i t + \phi_i) \quad (16)$$

165 where  $(x,y,z)$  are cross-shore, alongshore and vertical directions respectively;  $(i)$  is the index  
 166 of spectral distribution;  $a_i$  is the amplitude at each frequency  $\omega_i$ , from a given statistical  
 167 distribution, e.g., JONSWAP (Sec. 3.2);  $\omega_i$  and  $k_i$  are related by the linear dispersion  
 168 relation:  $\omega_i^2 = g k_i \tanh(k_i h)$  with  $h$  the water depth;  $\phi_i$  is a uniformly distributed random  
 169 phase.

170  $w_{bc}$  is set to zero and our tests show only weak sensitivity to this choice. Depth-averaged  
 171 (barotropic) velocities  $(\bar{u}, \bar{v})$  must be provided as well in the wave maker because they are  
 172 prognostic variables, advanced together with the fast acoustic mode. Normal depth-averaged  
 173 velocity  $\bar{u}$  is complemented at the boundary by an anti-Stokes "compensation flow", opposite  
 174 to Stokes drift and thus closing the volume budget. We do not impose the depth-averaged  
 175 value of  $u_{bc}$  directly but through the incoming characteristic of the shallow water system as  
 176 in Flather-type conditions (Marchesiello et al., 2001; Blayo and Debreu, 2005):

$$\bar{u} = \bar{u}_{bc} - \sqrt{\frac{g}{h}}(\eta - \eta_{bc}) \quad (17)$$

177 This allows infragravity waves generated inside the domain to propagate out as long waves,  
 178 while ensuring a near conservation of mass and energy through the open boundary. Likewise,  
 179 the baroclinic components  $(u_{bc}, w_{bc})$  are applied via an adaptive radiation condition which  
 180 helps short waves and 3D flow perturbations to leave the domain with only a small effect  
 181 on the interior solution (Marchesiello et al., 2001).

## 182 2.5. Sediment transport model

183 CROCO comes with capabilities for water quality, marine ecosystem and sediment mod-  
 184 eling. They are built upon Eq. 6 for the transport of tracer concentration but, in the case  
 185 of sediment modeling, additional sources and sinks are needed to simulate the exchange  
 186 between the water column and sediment bed (Blaas et al., 2007; Warner et al., 2008). Ne-  
 187 glecting compressibility ( $\rho \sim \rho_0$ ), the sediment concentration follows:

$$\underbrace{\frac{\partial C}{\partial t}}_{\text{RATE}} = - \underbrace{\vec{\nabla} \cdot \vec{v} C}_{\text{ADVECTION}} + \underbrace{\mathcal{D}_C}_{\text{MIXING}} - \underbrace{\frac{\partial w_s C}{\partial z}}_{\text{SETTLING}} + \underbrace{\frac{E}{\delta z_b}}_{\text{EROSION}} \Big|_{z=z_b} \quad (18)$$

188  $w_s$  is the settling velocity, dependent on sediment grain size, but not on flow conditions and  
 189 concentrations (Soulsby, 1997). Settling is computed via a semi-Lagrangian advective flux  
 190 algorithm, which is unconditionally stable (Durrán, 2010). It uses a piece-wise parabolic  
 191 vertical reconstruction of the suspended sediment for high-order interpolation, with WENO  
 192 constraints to avoid oscillations.  $E$  is the erosion flux at the sea floor and is only applied to  
 193 the first grid level of height  $z_b$  and cell size  $\delta z_b$ .

194 In the present study, this suspended sediment model is used with simple settings. It is  
 195 composed of a single fine sand class with settling velocity  $w_s=2.5 \text{ cm s}^{-1}$  (grain size  $d_{50}=0.22$   
 196 mm). For resuspension, taking one sediment bed layer for simplicity, only two parameters  
 197 are needed: the critical shear stress  $\tau_{cr}$ , i.e., the threshold for initiation of sediment motion;  
 198 and the erosion rate  $E_0$  at the seafloor, expressed in the erosion flux formulation (Blaas  
 199 et al., 2007):

$$E = E_0(1 - p) \frac{\tau_b - \tau_{cr}}{\tau_{cr}} \quad \text{for } \tau_b > \tau_{cr} \quad (19)$$

200  $E_0$  and  $\tau_{cr}$  are used for calibration but constrained by known empirical relationships:  $E_0$   
 201 is set to  $5.10^{-3} \text{ kg m}^{-2} \text{ s}^{-1}$  (Smith and McLean, 1977);  $\tau_{cr}$  to  $0.11 \text{ N m}^{-2}$  (Soulsby and  
 202 Whitehouse, 1997).  $p$  is the sediment porosity (0.41).  $\tau_b$  is the bottom shear stress estimated  
 203 from the logarithmic law of the wall:

$$\vec{\tau}_b = \frac{\kappa^2}{\log^2\left(\frac{z_b}{z_{0b}}\right)} |u_b| \vec{u}_b \quad (20)$$

204  $|u_b|$  and  $\vec{u}_b$  are the magnitude and vector of oceanic bottom currents in the log layer at a  
 205 height  $z_b$  above bed;  $\kappa = 0.41$  is the Von Karman constant; and  $z_{0b}$  is the bottom roughness  
 206 length.

207 In the surfzone, the effect of wave-breaking, especially plunging breakers over sandbars,  
 208 on sediment resuspension needs to be addressed (Voulgaris and Collins, 2000; van der Zanden  
 209 et al., 2017; Otsuka et al., 2017; Lim et al., 2020). Lim et al. (2020) suggest a strong  
 210 correlation between the maximum sediment concentration observed in the LIP experiment  
 211 and the plunging point (about 10 m downstream of the breaking point) over the sandbar.  
 212 We adopt here the approach proposed by Reniers et al. (2004) consisting in replacing  $|u_b|$

213 by an stirring velocity reinforced by the turbulence motion induced by wave breaking:

$$\begin{aligned} u_{stir} &= \sqrt{u_b^2 + u_t^2} \\ u_t &= \phi(x) \alpha_{br} \sqrt{k_b} \end{aligned} \tag{21}$$

214 The subgrid bottom turbulent kinetic energy  $k_b$  is given by the turbulence model and  $\alpha_{br}$  is a  
215 factor explaining the effectiveness of breaking waves in the entrainment of bottom sediments.  
216 We used a common value of  $\alpha_{br} = 10$  (Ribas et al., 2011), which produces sand concentrations  
217 close to observations in both the erosive and accretive phases.  $\phi(x)$  is a Gaussian function  
218 centered around the plunging point in the bar trough (the standard deviation of about 10m  
219 corresponds to the size of the bar trough) in order to enhance the turbulence effect at this  
220 position. This increased efficiency in the trough of the bar may also reflect the effect of  
221 the large ripples observed in this region. In the absence of an appropriate parametrization  
222 for this effect,  $\phi$  represents the only non-generic aspect of the present model application,  
223 but it is essential for reproducing the correct profiles of sediment concentrations — and in  
224 turn a condition for accurate bar migration. We have tried more generic parametrization  
225 of resuspension by plunging breakers with some success (see Discussion section) but a more  
226 dedicated study is needed for what appears to be a real challenge.

227 For bedload transport, we do not rely on parametrization as skewed-asymmetric waves  
228 are resolved explicitly, but we make sure that the wave-boundary layer is resolved, and  
229 that the first vertical level is in a sheet flow layer (about 10 times the grain size). This  
230 is particularly important for the onshore bar migration phase. However, some bedload  
231 processes are not resolved by the model, most notably the formation of small bedforms such  
232 as the observed centimeter-scale ripples in the bar trough. Estimating  $z_{ob}$  is not trivial  
233 with moving sediment and various empirical formulations exist that account for roughness  
234 over a mobile bed (Wiberg and Rubin, 1989). They generally assume that grain roughness  
235 is the most important predictor for the onset of sand suspension, but bedload and ripple  
236 form roughness are also considered (Li and Amos, 2001). We use grain roughness here, but  
237 assume that the effect of large ripples formed in the bar trough of LIP-1B and LIP-1C are  
238 included in the  $\phi$  function and participates in the increased resuspension efficiency due to

239 turbulence.

## 240 2.6. Bed model and morphological acceleration

241 The bed model accounts for changes in sea floor elevation resulting from convergence or  
242 divergence in sediment fluxes (Exner equation):

$$\frac{\partial h}{\partial t} = -\frac{f_m}{\rho_s(1-p)}(w_s C - E) \quad (22)$$

243  $\rho_s$  is sediment density and  $f_m$  is a morphological acceleration factor (Roelvink, 2006) de-  
244 scribed below. Morphological changes can have a significant influence on flow and transport  
245 when they are greater than a few percent of the water depth. Morphological updating strate-  
246 gies are described by Roelvink (2006) and implemented in CROCO following Warner et al.  
247 (2008). For dynamical consistency, the vertical velocity is modified by the rate of change of  
248 vertical grid levels  $dz/dt$ , adjusting to the moving sea floor and free surface (grid "breath-  
249 ing" component; Shchepetkin and McWilliams 2005). This method is mass conserving and  
250 retains tracer constancy preservation.

251 Morphological acceleration can be achieved with no constraint on the model time step  
252 by multiplying erosion and deposition rates by the scale factor  $f_m$  at the bed-water inter-  
253 face (Eq. 22). Only the fluxes to and from the bed are changed, not the magnitude of  
254 the sediment concentrations in the water column. To our knowledge, this paper presents  
255 the first use of a morphological acceleration technique with a wave-resolving model. We  
256 have found that the technique is effective since we obtain the same result with or without  
257 morphological acceleration. We were able to use a large morphological factor  $f_m$  (at least  
258 up to 72; Appendix B) because the timescale of morphodynamics (here bar migration) is  
259 slow compared to that of wave dynamics.

## 260 3. Simulation of a large-scale flume experiment

261 We now present our model solution of sandbar migration, applied to the European Large  
262 Installation Plan (LIP) experiments, which was carried out at full scale in Delft Hydraulics's

263 Delta Flume (Roelvink and Reniers, 1995). We first present the model setup for these ex-  
264 periments, the model validation of current profiles, then the comparative analysis of offshore  
265 and onshore bar migration.

### 266 *3.1. LIP experiment*

267 The Flume is 225 m long, 5 m wide and 7 m deep. In LIP, three types of experiments  
268 were carried out under different types of irregular waves, which subsequently resulted in a  
269 stable (1A), erosive (1B), and accretive (1C) beach state (see Table 1 and Roelvink and  
270 Reniers 1995 for details). The initial profile is linear in LIP-1A, with a slope of 1:30 and  
271 consisting of a median grain size of 0.22 mm. The final profile of LIP-1A was used as the  
272 initial profile of LIP-1B and the final profile of LIP-1B as the initial profile of LIP-1C. The  
273 wave conditions were a JONSWAP narrow-banded random wave spectrum generated by a  
274 wave paddle, with a peak enhancement factor  $\gamma = 3.3$  and characteristic wave height and  
275 period:  $H_s = 1.4$  m,  $T_p = 5$  s (LIP-1B) and  $H_s = 0.6$  m,  $T_p = 8$  s (LIP-1C). Under this  
276 wave forcing, the sandbar developed during LIP-1B, increasing in height and migrating in  
277 the offshore direction. Under the accretive conditions of LIP-1C, the bar migration reversed  
278 to the onshore direction. The bed profile was measured with a profile follower that used an  
279 automated sounding system. Wave-averaged current profiles were captured by a movable  
280 carriage with attached current meters starting 10 cm above the bed and at 10 locations along  
281 the flume (2 cm/s accuracy). Similar profiles of wave-averaged concentrations of suspended  
282 sediment were measured by suction tubes mounted on a carriage starting 5 cm above the  
283 bed (10% accuracy). For validation of currents and sand concentration, we consider the  
284 time 8 hours after initialization in experiment 1B and 7 hours in 1C (Veen, 2014).

### 285 *3.2. Model setup*

286 The model setup is adapted from the LIP experiment. A JONSWAP wave spectrum  
287 similar to the experiment is generated with shore normal direction and zero directional  
288 spread. A no-slip condition is imposed on the lateral wall boundaries of the canal so that  
289 transverse modes are precluded. The grid spacing is  $dx=25$  cm with 20 vertical levels with

290 refinement at the bottom (stretching parameter  $\theta_b=8$ ; see the documentation in [www.croco-](http://www.croco-ocean.org)  
291 [ocean.org](http://www.croco-ocean.org)). This vertical refinement is important for wave-related transport (Appendix B),  
292 while the results were poorly sensitive to horizontal resolution (a test with  $dx=50$  cm and 1  
293 m is presented below). The model time step is  $dt = 25$  ms. The minimum depth is 1 mm on  
294 the shore, the position of which varies with the swash oscillation, relying on a wetting-drying  
295 scheme (Warner et al., 2013). For bed skin shear stress (setting sediment in motion), the  
296 logarithmic law of the wall is used with grain roughness  $z_{0s} \sim D50/12 = 0.02$  mm. To  
297 account for increased momentum friction due to near-bed sediment suspension (e.g., Suarez  
298 et al. 2014), we increase the roughness length for momentum to 0.2 mm.

299 The LIP-1B and LIP-1C experiments lasted 18 and 13 hours, respectively. In both cases,  
300 the model was run for one hour with a morphological acceleration factor  $f_m$  equal to 18 and  
301 13 respectively. Since the results were weakly sensitive to  $f_m$  (Appendix B), the acceleration  
302 technique allowed us to perform many sensitivity tests with even larger values.

### 303 *3.3. Undertow's sensitivity to the turbulence model*

304 An important process for offshore bar migration is the suspended load transport by  
305 the undertow. To assess the model hydrodynamics in the LIP configuration, we present a  
306 comparison of current profiles taken after 8 hours in experiment 1B and averaged over one  
307 hour. It will also be a test for our numerical wave maker in its ability to generate a spectrum  
308 of random waves.

309 Figure 1 shows a comparison of the model with data, using our standard configuration.  
310 The match with measured currents is very good throughout the complex morphology of the  
311 beach. The waves start to break before reaching the sandbar, but the breaking is more  
312 intense on the sandbar. The resulting velocity profile on the lee-side of the bar has a high  
313 shear and the undertow has an intensity above 30 cm/s. The horizontal resolution test (25  
314 cm, 50 cm and 1 m) shows a root mean square error of about 3 cm/s at all resolutions,  
315 close to the measurement error of 2 cm/s. The results are thus consistent at all resolutions  
316 despite no adjustment of any parameter. It confirms the validity of a RANS approach for  
317 estimating the mixing length of breaking-induced turbulence.

318 Turbulent kinetic energy and eddy viscosity estimated by the  $k$ - $\omega$  model in the breaker  
 319 zone have the expected structure (Fig. 1; top) and magnitude ( $\nu_t \sim 0.01h\sqrt{gh}$ ; Svendsen  
 320 1987; Cox et al. 1994). Interestingly, the transport terms in the closure equations tend  
 321 to reduce mixing at the break point by redistributing the turbulent energy, thus allowing  
 322 more intense shear to be maintained (not shown). The  $k$ - $\epsilon$  model works almost as well as  
 323 the  $k$ - $\omega$  model, with respect to mean current profiles, but the comparison is improved by  
 324 imposing a high value on the surface mixing length ( $z_{0s}=0.2$  m), as in wave-averaged models  
 325 (Feddersen and Trowbridge, 2005). The  $k$ - $\omega$  model may thus be a better choice for surface  
 326 wave breaking, possibly due to a more accurate near wall treatment (Mayer and Madsen,  
 327 2000; Brown et al., 2016; Devolder et al., 2018; Larsen et al., 2020). Note, however, that  
 328 this model produces a greater amount of mixing in potential flow regions outside the surf  
 329 zone (innershelf), mainly due to the divergence part of the mean strain rate tensor (Mayer  
 330 and Madsen, 2000).

### 331 3.4. Currents and suspended sediments

332 To analyse the model for different conditions, two experiments are selected, namely LIP-  
 333 1B and the LIP-1C cases (Table 1). Figure 2 shows a model-data comparison of wave height  
 334  $H_{rms}$  for the two experiments. The data is reproduced accurately. High waves in LIP-1B  
 335 ( $H_s = 1.4$  m,  $T_p = 5$  s) caused beach erosion while, in LIP-1C, moderate waves ( $H_s = 0.6$   
 336 m,  $T_p = 8$  s) caused an accretive beach.

337 We are first interested in the erosion phase where the sandbar moves offshore under  
 338 the effect of high waves. The LIP dataset is used for comparison purposes, containing  
 339 velocity and concentration profiles as well as bed level evolution (see next section). Figure  
 340 3 shows a comparison of wave-mean velocity and concentration profiles for LIP-1B. As for  
 341 velocity profiles presented in the validation section, sediment concentrations also show a very  
 342 good match between model and data. The high concentrations over the sandbar crest and  
 343 trough are triggered by a large bed shear stress associated with a combination of wave-mean  
 344 undertow, wave oscillatory flow and turbulent velocity.

345 Note that waves and associated orbital velocities are greatest off the sandbar. However,



346 the maximum suspended sand concentration is not found offshore but coincides instead with  
347 a maximum in turbulence intensity and undertow speed inshore of the bar crest. This is in  
348 contrast to parametrizations more suited to shelf dynamics, where sediment erosion responds  
349 to the addition of bed shear stresses by currents and waves. (e.g., Soulsby 1995).

350 The model simulation of the accretive phase is also reproduced accurately (Figure 4).  
351 In this case, wave breaking occurs right on the sand bar. The currents are thus weaker  
352 off the bar, but the undertow inside the bar remains quite strong. The suspended sand  
353 concentration is also significant there, although weaker than in the erosive phase.

### 354 3.5. High-order velocity moments

355 In Marchesiello et al. (2021), the model’s ability to simulate surface gravity wave prop-  
356 agation, nearshore breaking and the resulting circulation was validated against laboratory  
357 experiments. In particular, the GLOBEX scaled experiment (Michallet et al., 2014) pro-  
358 vided high-resolution measurements of surface wave height  $\eta$  and bottom currents  $u_b$ . A  
359 comparison was made of continuous profiles of wave statistics for  $\eta$  and  $u_b$  across the beach.  
360 The model statistics closely resembled the measurement data (in structure and magnitude  
361 for both  $\eta$  and  $u_b$ ), including high-order moments, showing the transition from skewness to  
362 asymmetry across the surfzone. More details can be found in Marchesiello et al. (2021).

363 The LIP experiment does not provide a high-resolution cross-shore profile but does pro-  
364 vide access to vertical profiles at different positions around the sandbar. We can therefore  
365 make model-data comparisons of these vertical profiles for high-order velocity moments.  
366 Figure 5 shows the standard deviation  $u_{rms} = \sqrt{\langle u'^2 \rangle}$ , which is a measure of the intensity of  
367 the orbital wave velocity and the wave-mean flow variability in the surfzone. In experiments  
368 1B and 1C, the model can faithfully reproduce the  $u_{rms}$  profiles. In experiment LIP-1B,  
369  $u_{rms}$  is more intense than in experiment LIP-1C, particularly seaward of the sandbar, and  
370 in both cases there is a marked surface maximum of  $\sim 1$  m/s in the surfzone (which starts  
371 much further offshore in LIP-1B).

372 Then, we compare the third moment of the velocity, more precisely its normalized skew-  
373 ness:  $S_u = \langle u'^3 \rangle / u_{rms}^3$ . This is a measure of the nonlinear effects on the wave cycle, which

374 allows for residual transports. Figure 1 shows a model-data comparison of  $S_u$  for vertical  
 375 profiles. Again, the ability of the model to reproduce the measured skewness is evident. In  
 376 both the model and the data, there is an increase in skewness associated with shoaling and  
 377 maximum values are near the surface ( $\sim 2$  or more). Note that the normalized skewness ap-  
 378 pears higher in the LIP-1C experiments. A more comprehensive analysis of the normalized  
 379 and absolute skewness, its transition to asymmetry, and its impact on sediment transport  
 380 is presented in Sections 4.2 and 4.3.

### 381 *3.6. Offshore and onshore sandbar migration*

382 Figure 7 (top) shows the bed evolution 18 hours after the start of experiment 1B, with  
 383 the corresponding section of wave-mean cross-shore currents. Some small discrepancies are  
 384 visible but the offshore sandbar migration is correctly reproduced in terms of distance of  
 385 migration and height of the bar. The maximum undertow magnitude ( $-0.3$  m/s) follows the  
 386 lee-side slope of the bar and appears as the driver of its migration. A secondary undertow  
 387 maximum is visible on the terrace of the inner surfzone and appears to coincide with an  
 388 offshore spreading of the terrace. In the outer surfzone (off of the main sandbar), a second  
 389 smaller bar forms both in the model and in the measurements during LIP-1B (at a position  
 390 of about 70 m), corresponding to a weakened undertow.

391 The bottom panel of Figure 7 shows the result of the accretive phase LIP-1C when  
 392 waves have lower amplitude, with an opposite migration directed towards the shore. In this  
 393 case, the flow structure is very similar but weaker and the bar seems to migrate against the  
 394 undertow. The latter is still able to bring sediment to the leeward slope of the bar, but the  
 395 offshore slope and crest are moving inshore, creating an asymmetry in the bar shape.

## 396 **4. Sediment transport mechanisms**

### 397 *4.1. Sediment budget*

398 The wave-mean sediment transport term  $-\frac{1}{\rho_0} \overline{\nabla \cdot (\rho \vec{v} C)}$  —  $\rho_0$  is a mean density, so that the  
 399 transport term is in units of  $\text{kg}/\text{m}^3/\text{s}$  — is presented in Figure 8 for the time halfway between  
 400 the erosive (top) or accretive (bottom) phase (averaged over a period of about 10 min). This

401 confirms for LIP-1B that during high waves and strong undertow, a large amount of sand  
402 is removed from the lee-side slope and trough area – after being first suspended and stirred  
403 upward – and then transported to the offshore slope, where it settles. Part of this process is  
404 attenuated by surface transport, which brings back some sediment to the bar trough region.  
405 For LIP-1C, however, sand is removed from both the trough of the bar and the offshore  
406 slope (including the crest), and it converges to the inner slope. This transport may qualify  
407 as bedload or sheet-flow transport as it occurs primarily within the wave boundary layer of  
408 about 10 cm.

409 For further analysis, Figure 9 presents the depth-integrated suspended sediment budget,  
410 again for the erosive (top) or accretive (bottom) phase. The different terms of Eq. 18 are  
411 computed online, averaged in time and integrated along the vertical axis. The budget comes  
412 down to three main tendency terms: erosion-settling residual, advection, and rate of change  
413 (vertical mixing cancels out when integrated). Suspended sediments are close to equilibrium  
414 since the rate of change is generally much smaller than the other two terms, which tend to be  
415 opposite. This analysis confirms the difference between LIP-1B and LIP-1C, with transport  
416 limited to the lee side of the bar in LIP-1C. In LIP-1B, there is a dominant erosion process  
417 in the bar trough and offshore transport and deposition beyond the bar crest. In LIP-1C,  
418 the process is similar but transport and deposition are confined to the inshore side of the  
419 bar crest, while a secondary zone of net erosion appears on the offshore side (this secondary  
420 erosion also appears in LIP-1B but much further offshore).

421 In the next section, we distinguish between wave and mean current transport in order  
422 to identify and quantify the role of asymmetric waves for bar migration, particularly in the  
423 accretion phase of LIP-1C.

#### 424 *4.2. Wave and current related fluxes*

425 For each experiment (erosive or accretive), we construct an ensemble-averaged wave by  
426 superposing instantaneous data over 50 wave cycles. To that end, the duration of all wave  
427 cycles must be fitted to the peak period, so that data averaging can be made over the  
428 different phases of a generic cycle. This composite wave is made at three locations around

429 the sand bar: offshore slope, crest and trough (corresponding to positions:  $x=131$ , 139 and  
430 148 m in LIP-1B; and  $x=129$ , 136 and 144 m in LIP-1C).

431 Starting with the more complex accretive case, Figure 10 (left panels) shows the currents  
432 and sand concentrations at the first model level (1-2 cm above bed) for the ensemble-averaged  
433 LIP-1C wave cycle. Here, the mean values are subtracted to retain the oscillatory flow and  
434 associated perturbed concentrations ( $u' = u - \bar{u}$  and  $C' = C - \bar{C}$ ). In addition, the figure  
435 shows the undertow  $u_b$  and effective turbulent velocity  $u_t$  entering the bed shear stress  
436 formulation. The currents have a typical skewed/asymmetric structure across the wave  
437 cycle. They are largely skewed over the offshore slope and become less skewed but more  
438 asymmetric as the wave progresses toward the shore. The largest sand concentrations are  
439 over the bar crest where it peaks about half a second after the peak velocity. The lag is  
440 similar for skewed velocity over the offshore slope, although the peak there is much weaker.  
441 The main reason is that the effective turbulent velocity is more than twice as large over the  
442 crest and more significantly contributes to the bed skin stress. The sand concentration cycle  
443 is different in the trough region, where the undertow becomes comparable in magnitude  
444 to the orbital velocities. The total onshore-directed flow is reduced during the crest half-  
445 cycle while the offshore-directed flow is increased during the trough half-cycle. As a result,  
446 positive resuspension anomalies are present during both phases of negative and positive wave  
447 velocities.

448 The right panels of Figure 10 show the profiles of suspended sand flux. The averaged  
449 flux over the composite wave cycle is separated into wave and mean components (Reynolds  
450 decomposition):  $\overline{uC} = \overline{u'C'} + \bar{u}\bar{C}$ . The undertow being weak over the offshore slope and crest  
451 (Fig. 7, bottom panel), the total flux there is dominated by the residual wave flux, which  
452 is onshore (at least in the wave boundary layer, but the flux is slightly negative above).  
453 The maximum flux over the crest appears to coincide with a combination of high sand  
454 concentration (1 g/L) and positive velocities (1 m/s) during the wave crest period (middle-  
455 right panel of Fig. 10). Over the offshore slope, the velocity is more skewed than asymmetric  
456 and the concentration is limited by a weaker bed stress (due to weaker turbulence). The  
457 phase lag between mobilization and transport of sediment (Dohmen-Janssen et al., 2002)

458 does not vary between the two positions (as expected from fast settling that gives a weak  
459 phase lag parameter) and therefore cannot explain the differences in sand flux. In the bar  
460 trough, the wave-related flux remains positive but its magnitude is much reduced. The  
461 reason is the second peak of sand resuspension during the wave trough phase (negative half-  
462 cycle), due to a strong undertow. An even stronger undertow could eventually reverse the  
463 sign of this flux (Scott et al., 2009).

464 The result of opposite onshore and offshore sand fluxes is a convergence of sediment  
465 between the trough and crest regions, which promotes an onshore bar migration. However,  
466 the bar takes on an asymmetric shape because the undertow plays a dominant role in the  
467 displacement of the bar crest.

468 In our simulations, wave-breaking turbulence contributes little to the onshore flux asym-  
469 metry, unlike in Ting and Kirby (1994). This is because, although more turbulent energy  
470 is generated at the surface during the wave crest period, a significant portion reaches the  
471 bottom during the trough of the next wave (Figure 1) and the bottom turbulent velocity  
472 in Figure 10 exhibits only small variations during the wave cycle. This tends to disprove a  
473 strong asymmetric effect of breaking turbulence, although we are aware that the turbulence  
474 model may misrepresent the role of plunging waves (i.e. the timing and efficiency with which  
475 they mobilize sediment). However, observations suggest that wave-breaking turbulence is  
476 sufficiently intermittent to lose some of its correlation with wave phase (Ruessink et al.,  
477 2011). Improvements could be made in this area.

478 Looking now at the LIP-1B erosive experiment (Fig. 11), we see a different dynamic,  
479 which generalizes the mechanism described earlier for the trough region. Since the undertow  
480 is present at all locations, the total flux is also negative at all locations. The wave-related flux  
481 tends to be weaker than in LIP-1C because the negative undertow decreases resuspension  
482 in the positive orbital velocity phase (wave crest) and increases it in the negative phase.  
483 However, over the crest, a significant wave-related flux remains. Note that waves over the  
484 sandbar in LIP-1B are actually weaker than in the accretive phase, despite larger offshore  
485 waves. The reason is that, in this case, the waves break offshore of the bar, where they lose  
486 much of their energy. However, the undertow extends across the sandbar and overwhelms

487 the wave effect everywhere. Note also the strong sediment flux convergence on the offshore  
488 side of the bar despite a consistently negative sand flux. In other words, there need not be  
489 a confluence of onshore and offshore fluxes for flux convergence to occur.

#### 490 4.3. Wave asymmetry

491 Figure 12 presents the profiles of velocity skewness and asymmetry for the three sandbar  
492 locations. Normalized values (right panel) are computed as  $\langle u'^3 \rangle / \langle u'^2 \rangle^{1.5}$  for skewness and  
493  $\langle \mathcal{H}(u')^3 \rangle / \langle u'^2 \rangle^{1.5}$  for asymmetry ( $\mathcal{H}$  is Hilbert transform). Since a logarithmic scale is used,  
494 the asymmetry (which is negative) is presented with an inverse sign. We also present the  
495 dimensional values (right panel), calculated without normalization by the cube of standard  
496 deviations (on a normal scale with the correct signs). It confirms the relatively high skewness  
497 and low asymmetry on the outer slope, relative to the bar crest. The dimensional values  
498 reveal a much higher absolute asymmetry on the bar crest, compared to the surrounding  
499 areas. This result is hidden in the dimensionless calculation. Bottom skewness on the slope  
500 is larger than asymmetry on the crest, but with less effect on the sand flux, as seen earlier.  
501 This would confirm the efficiency of asymmetry versus skewness, but in our case this is  
502 coincidental: the difference in sand flux is related to the varying intensity of turbulence across  
503 the bar, rather than a varying phase lag between mobilization and transport — sometimes  
504 associated with the transformation from skewness to asymmetry (Dohmen-Janssen et al.,  
505 2002). Our interpretation of the respective role of asymmetry and skewness thus involves a  
506 spatial correlation between wave asymmetry and breaking turbulence. Finally, the trough  
507 region is characterized by low absolute skewness and asymmetry near the bed and the sand  
508 flux there is always driven by the undertow.

509 The dimensional skewness and asymmetry also reveal large vertical variations from sur-  
510 face to bottom. Wave skewness/asymmetry is intensified at the surface because wave ve-  
511 locities are greater, but the intensification is much larger for broken waves inshore of the  
512 bar. In this case, the wave velocities are also much stronger at the surface (1.2 m/s, as  
513 for shoaling waves) than at the bottom. Another noticeable point about these profiles is  
514 a slight decrease of dimensional skewness and asymmetry in the wave boundary layer, and

515 a decrease of dimensionless asymmetry in favor of skewness, consistent with observations  
 516 (Suarez et al., 2014). However, here, this transformation is weak and does not seem to play  
 517 a significant role.

#### 518 4.4. Bed shear stress

519 It is interesting at this point to test the validity of formulations used for bed skin stress in  
 520 wave-averaged models (e.g., Blaas et al. 2007). The friction is generally parametrized as the  
 521 addition of wave-induced ( $\tau_w$ ) and current-induced ( $\tau_{cw}$ ) bottom stresses (Soulsby, 1995).  
 522 In the nearshore region (as confirmed here), the current-related friction has a lesser role and  
 523 we will focus on the friction induced by wave orbital velocities and breaking turbulence:

$$\tau_w = 0.5 f_w (u_w + u_t)^2 \quad (23)$$

524 with the friction factor of (Soulsby, 1995), but modulated by a constant  $r_f$ :

$$f_w = 1.4 r_f \left( \frac{u_w}{\sigma z_{0b}} \right)^{-0.52} \quad (24)$$

525  $u_w$  is the maximum bottom orbital velocity, determined from wave-mean wave character-  
 526 istics.  $u_w$  accounts for skewness as a correction to the linear theory (Isobe and Horikawa,  
 527 1982; Abreu et al., 2010; Malarkey and Davies, 2012; Nam et al., 2020):  $u_w = r_s u_w^{lin}$ , where  
 528  $r_s$  is a skewness factor ( $r_s = 1$  for linear waves).  $r_s$  can be obtained using empirical formulas  
 529 involving a logarithmic relation with the Ursell number  $Ur = H_s \lambda^2 / h^3$ , a measure of wave  
 530 nonlinearity (Nam et al., 2020). Here, we use a slightly modified formula compared to Nam  
 531 et al. (2020):  $r_s = 0.051 \log Ur + 0.84$ . The turbulence stirring velocity  $u_t$  is computed as  
 532 for the wave-resolving simulation (Eq. 21) but with wave-averaged  $k_b$ .

533 Figure 13 shows our attempt of reconstructing the simulated bed shear stress with  
 534 parametrizations. At offshore positions, where waves are nearly linear, a value of  $r_f = 0.4$  is  
 535 required to fit the simulated bed shear stress reconstructed from Soulsby (1995). Then, in  
 536 the nearshore region from about  $x = 40$  m, the parametrized bed shear stress is underesti-  
 537 mated with linear waves and no turbulence ( $r_s = 1$  and  $u_t = 0$ ). Adding skewness correction,  
 538 the bed shear stress is recovered more accurately around the sandbar, particularly on the

539 seaward side. The most striking feature is that turbulence stirring contributes most of the  
540 simulated stress on the sandbar (in the roller zone, downstream of the sandbar, the turbu-  
541 lence effect seems underestimated). This analysis thus confirms that consideration of wave  
542 asymmetry may be important but that turbulent mixing by plunging breakers is essential  
543 to mobilize significant sediment loads on the sandbar, allowing its realistic migration.

## 544 **5. Discussion and conclusion**

545 Analysing the interplay between wave-mean undertow and wave skewness/asymmetry in  
546 field measurements or even in physical models is a difficult task. Numerical modeling offers  
547 a complementary approach as long as their precision is not too compromised by unknown  
548 parameters. In the present study, we present the first application of a 3D wave-resolving  
549 hydro-sedimentary model to the case of bar migration. The simulation is that of a large-scale  
550 flume experiment comprising two phases: erosion (offshore bar migration) during high-wave  
551 conditions and accretion (onshore bar migration) during post-storm, moderate wave forcing.  
552 The model can faithfully reproduce in both cases the wave statistics, the profiles of wave-  
553 mean currents and sand concentrations, as well as the morphological evolution of the sand  
554 bed.

555 The analysis of the transport mechanisms shows the importance of the undertow distri-  
556 bution around the sandbar. Its convergence and divergence patterns appear to have more  
557 impact on submerged beach morphology than the confluence of wave- and current-related  
558 fluxes. In the erosion phase, in particular, the waves begin to break well ahead of the sand-  
559 bar and a strong undertow causes the mobilized sediments to move in a near morphological  
560 translation. In this case, the wave-related transport is weak because sand mobilization,  
561 dominated by the undertow, is distributed almost uniformly between the two phases of the  
562 wave cycle. Much of the sediment is carried in suspension as the intensity of turbulence  
563 within the water column is high. It is interesting to note that there is a moderate surface  
564 flux of sediment towards the shore, carried by the breaker-induced onshore flow.

565 Under moderate wave action, the sandbar becomes a breaking point and the undertow  
566 is confined to the inshore area of the bar crest with a maximum on the inner slope. Because



567 the bed shear stress and associated mixing is weaker in this case, the sand concentration is  
568 almost confined to a sheet flow layer. In comparison to more energetic wave conditions, the  
569 sandbar migrates onshore at a slower rate and with an asymmetric shape. This is the result  
570 of a combination of two mechanisms: the deposition of sand carried by the undertow on the  
571 inshore side and, to a lesser extent, onshore fluxes due to asymmetric waves on the offshore  
572 side. Note that these mechanisms are not opposed here, as is often suggested (Roelvink and  
573 Stive, 1989; Grasso et al., 2011), but complementary.

574 The ensemble-averaged analysis of wave skewness/asymmetry reveals that the onshore  
575 sediment flux is strongest over the bar crest where asymmetry produces a steep front with  
576 strong currents during the onshore half-cycle. Associated with these currents, the sand is  
577 resuspended within only half a second with much contribution by breaking turbulence. Over  
578 the offshore slope, where skewness is greater than asymmetry, a similar process is at work  
579 but is weaker due to weaker resuspension, and not because of a greater phase lag between  
580 mobilization and transport.

581 These results thus confirm a moderate role of wave-related onshore flow in the evolution  
582 of sandbars, and only under post-storm wave conditions, when the sandbar becomes a break-  
583 point. The role of bottom streaming is unclear. By increasing bottom friction (roughness  
584 increased to 1 mm), it becomes a player, although not a dominant one in our simulations  
585 (it contributes a few cm/s and hardly counteracts the undertow of about 30 cm/s over the  
586 bar, even in the accretive case).

587 The largest uncertainty in the model comes from the estimation of bed roughness and  
588 bed shear stress due ripple formation and wave-breaking turbulence. Sand mobilization (and  
589 transport) does not appear realistic in the model without turbulence, as noted earlier by  
590 van der Zanden et al. (2017) and Lim et al. (2020) for the same LIP experiments. Lim  
591 et al. (2020) suggested that a large local increase of suspended sand concentration in the  
592 LIP experiment could only be attributed to breaker-induced turbulence and more precisely  
593 to plunging breaker vortices invading the wave bottom boundary layer. We followed the  
594 idea of Reniers et al. 2004 of adding a breaker-induced turbulent bottom current (from the  
595 turbulence model) to the bed shear stress formulation. Using an enhancement factor and

596 positioning function for this turbulent motion leads to realistic results. The positioning  
597 function, however, lacks a clear generic type mechanism and efforts should be made in this  
598 direction. This positioning coincide with the presence of large ripples in the bar trough and a  
599 parametrization of ripple formation may be a way to account for the increased resuspension  
600 efficiency in this area. Among the attempts we made, we found good results by setting the  
601 bottom roughness as a function of the undertow (the undertow seems to be a good proxy  
602 due to its correlation with breaker dissipation; Faria et al. 2000). Additional work will likely  
603 be required, including explicit consideration of plunging breakers.

604 Another feature of the present simulations is the absence of a bedload transport model.  
605 Our simulations suggest that the suspended load is the dominant mechanism driving the  
606 LIP sandbar migration, including sand fluxes associated with wave asymmetry in the wave  
607 boundary layer. In this sense, the bedload transport is at least partially solved but this  
608 would deserve further analysis. In particular, our simulations are limited to medium sands  
609 which have a rather low sedimentation velocity compared to the local bed friction velocity  
610 and are therefore easily suspended by waves. Application of the model to coarse sand beds  
611 may require a dedicated formula for bedload transport, including an avalanche process.

612 With these caveats in mind, our results nonetheless demonstrate the reliability of compu-  
613 tationally efficient 3D wave-resolving models such as CROCO – intermediate between two-  
614 phase DNS and wave-averaged approaches – for addressing nearshore hydro-morphodynamic  
615 problems. They point out the deficiencies of the depth-averaged approach, with its hyper-  
616 sensitivity to bottom friction and rough estimation of the undertow, when its cross-shore dis-  
617 tribution seems so important. Our wave-resolving model can also help improve parametriza-  
618 tions in 3D wave-averaged models, in particular wave-related transport through the concept  
619 of wave half-cycles (Shafiei et al., 2022).

620 An interesting new result also comes from the morphological acceleration method, which  
621 is used here for the first time with a wave-resolving model. We demonstrated its applicability  
622 and were even able to use a factor as large as 72, because the timescale of morphodynamics  
623 (here bar migration) is slow compared to that of wave dynamics. This method is particularly  
624 interesting in terms of computational cost because it allows us to consider long period

625 simulations despite the choice to explicitly solve for waves. Therefore, the possibility of  
 626 using strong morphological acceleration in a wave-resolving approach makes it a suitable  
 627 alternative to wave-averaged models for realistic 3D applications.

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 634 these, all data were acquired by the authors and the CROCO source code is freely available  
 635 at [www.croco-ocean.org](http://www.croco-ocean.org). Both laboratory and modeling data are available upon request.

## 636 **Appendix A. Time-splitting procedure**

637 The compressible Navier-Stokes equations require a relation between  $\rho$  and  $P$ . To that  
 638 end, and as part of a time-splitting approach, density is decomposed into slow and fast com-  
 639 ponents based on a first-order linear decomposition with respect to total pressure (Auclair  
 640 et al., 2018; Marchesiello et al., 2021). In the following,  $s$  and  $f$  subscripts refer to slow and  
 641 fast-mode components respectively:

$$\rho = \rho_s(T, S, P_s) + \overbrace{\frac{\partial \rho}{\partial P} \Big|_{T,S} \delta P + O(\delta P^2)}^{\rho_f = c_s^{-2} P_f} \quad (\text{A.1})$$

$$P = \underbrace{P_{atm} + \int_z^\eta (\rho_s - \rho_0) g \, dz'}_{SLOW} + \underbrace{\rho_0 g (\eta - z) + \overbrace{\delta P}^{P_f}}_{FAST} \quad (\text{A.2})$$

642  $c_s$  is the speed of sound and  $\delta P = P_f$  is the nonhydrostatic pressure.

643 The Navier-Stokes equations are then integrated with two different time steps within  
 644 the time-splitting approach inherited from ROMS. The slow-mode integration is similar to

645 ROMS, with the addition of the slow part of vertical momentum equation, while the fast-  
646 mode integration is in 3D and includes the compressible terms of momentum and continuity  
647 equations. In vector form:

$$\frac{\partial \rho \vec{v}}{\partial t} = \underbrace{-\vec{\nabla} \cdot (\rho \vec{v} \otimes \vec{v}) - 2\rho \vec{\Omega} \times \vec{v} - \vec{\nabla} \left( \int_z^{\eta_f} (\rho_s - \rho_0) g \, dz' \right) + \vec{\mathcal{F}}_{\vec{v}} + \vec{\mathcal{D}}_{\vec{v}}}_{SLOW} + \underbrace{-\rho_0 g \vec{\nabla} \eta_f - \vec{\nabla} P_f + \rho_f \vec{g} + \lambda \vec{\nabla} (\vec{\nabla} \cdot \vec{v})}_{FAST} \quad (A.3)$$

$$\frac{\partial \rho_f}{\partial t} = -\frac{\partial \rho_s}{\partial t} - \vec{\nabla} \cdot (\rho \vec{v}) \quad (A.4)$$

$$P_f = c_s^2 \rho_f \quad (A.5)$$

$$\frac{\partial \eta_f}{\partial t} = w_f|_{z=\eta} - \vec{v}_f|_{z=\eta} \cdot \vec{\nabla} \eta_f \quad (A.6)$$

$$\frac{\partial \rho C_s}{\partial t} = -\vec{\nabla} \cdot (\rho \vec{v} C_s) + \mathcal{F}_C + \mathcal{D}_C \quad (A.7)$$

$$\rho_s = \rho(T_s, S_s, \eta_f) \quad (A.8)$$

$$\rho = \rho_s + \rho_f \quad (A.9)$$

648 The momentum is integrated both in slow and fast modes but the right-hand-side of  
649 the equation is split in two parts: a slow part, made of slowly varying terms (advection,  
650 Coriolis force, baroclinic pressure force and viscous dissipation), and a fast part, made of fast-  
651 varying terms (the surface-induced and compressible pressure force, weight, and dissipation  
652 associated with bulk-viscosity). This momentum equation is numerically integrated twice,  
653 once with a large time-step keeping the fast part constant, and once with a smaller time-step  
654 keeping the slow part constant. This is much more computationally efficient than integrating  
655 the whole set of equations at the same fast time step.

## 656 Appendix B. Model sensitivity to morphological acceleration and vertical res- 657 olution

658 We briefly present some of the sensitivities of the morphological evolution to the model  
659 parameters during the erosion phase 1B (Fig. 14). First, the reliability of the morphological  
660 factor is assessed by testing a value of  $f_m$  four times larger than in the standard run: 72  
661 instead of 18. In this case, the model is run for 20 minutes (about 10 min wall-clock time  
662 on a 4 processor machine) instead of the actual 18-hour experiment that would be required  
663 if no acceleration was performed ( $f_m=1$ ). After 18 hours, the difference in morphological  
664 evolution between the simulations with  $f_m$  of 18 and 72 is very small, even in the sandbar  
665 area, which is the most active. We did not increase  $f_m$  further because it would start  
666 affecting the statistics of wave spectrum.

667 A more sensitive parameter is the vertical resolution. We tested a simulation with 10  
668 vertical levels instead of 20 (Fig. 14). In this case, the current-related offshore transport is  
669 less affected than the wave-related onshore transport. This is because the latter occurs in  
670 a wave boundary layer of about 10 cm width that requires at least a few grid points to be  
671 resolved (see Sect. 4.2). Therefore, with weaker onshore transport, the bar migrates a few  
672 meters farther offshore than in the laboratory experiment and in the 20-level simulation.

## 673 References

- 674 Abreu, T., Silva, P.A., Sancho, F., Temperville, A., 2010. Analytical approximate wave form for asymmetric  
675 waves. *Coastal Engineering* 57, 656–667.
- 676 Alberello, A., Chabchoub, A., Gramstad, O., Babanin, A.V., Toffoli, A., 2016. Non-gaussian properties of  
677 second-order wave orbital velocity. *Coastal Engineering* 110, 42–49.
- 678 Almar, R., Castelle, B., Ruessink, B.G., Sénéchal, N., Bonneton, P., Marieu, V., 2010. Two- and three-  
679 dimensional double-sandbar system behaviour under intense wave forcing and a meso–macro tidal range.  
680 *Continental Shelf Research* 30, 781–792.
- 681 Auclair, F., Bordois, L., Dossmann, Y., Duhaut, T., Paci, A., Ulses, C., Nguyen, C., 2018. A non-hydrostatic  
682 non-boussinesq algorithm for free-surface ocean modelling. *Ocean Modelling* 132.
- 683 Blaas, M., Dong, C., Marchesiello, P., McWilliams, J.C., Stolzenbach, K.D., 2007. Sediment-transport  
684 modeling on southern californian shelves: A ROMS case study. *Continental Shelf Research* 27, 832 – 853.

685 Blayo, E., Debreu, L., 2005. Revisiting open boundary conditions from the point of view of characteristic  
686 variables. *Ocean Modelling* 9, 231–252.

687 Borges, R., Carmona, M., Costa, B., Don, W.S., 2008. An improved weighted essentially non-oscillatory  
688 scheme for hyperbolic conservation laws. *Journal of Computational Physics* 227, 3191 – 3211.

689 Brown, S., Greaves, D.M., Magar, V., Conley, D.C., 2016. Evaluation of turbulence closure models under  
690 spilling and plunging breakers in the surf zone. *Coastal Engineering* 114, 177–193.

691 Burchard, H., Petersen, O., Rippeth, T.P., 1998. Comparing the performance of the mellor-yamada and the  
692  $k - \epsilon$  two-equation turbulence models. *Journal of Geophysical Research: Oceans* 103, 10543–10554.

693 Chauchat, J., Cheng, Z., Nagel, T., Bonamy, C., Hsu, T.J., 2017. Sedfoam-2.0: a 3-d two-phase flow  
694 numerical model for sediment transport. *Geoscientific Model Development* 10, 4367–4392.

695 Chen, Q., Wu, W., Zhu, L., 2018. Recent Developments in Numerical Modeling of Coastal Hydrodynamics  
696 and Sediment Transport. chapter Chapter 5. pp. 145–197.

697 Cheng, Z., Hsu, T.J., Calantoni, J., 2017. Sedfoam: A multi-dimensional eulerian two-phase model for  
698 sediment transport and its application to momentary bed failure. *Coastal Engineering* 119, 32–50.

699 Cox, D.T., Kobayashi, N., Okayasu, A., 1994. Vertical Variations of Fluid Velocities and Shear Stress in  
700 Surf Zones. pp. 98–112.

701 Debreu, L., Marchesiello, P., Penven, P., Cambon, G., 2012. Two-way nesting in split-explicit ocean models:  
702 Algorithms, implementation and validation. *Ocean Modelling* 49–50, 1 – 21.

703 Derakhti, M., Kirby, J.T., 2014. Bubble entrainment and liquid–bubble interaction under unsteady breaking  
704 waves. *Journal of Fluid Mechanics* 761, 464–506.

705 Derakhti, M., Kirby, J.T., Shi, F., Ma, G., 2016. Nhwave: Consistent boundary conditions and turbulence  
706 modeling. *Ocean Modelling* 106, 121 – 130.

707 Devolder, B., Troch, P., Rauwoens, P., 2018. Performance of a buoyancy-modified  $k - \omega$  and  $k - \omega$  sst  
708 turbulence model for simulating wave breaking under regular waves using openfoam. *Coastal Engineering*  
709 138, 49–65.

710 Dibajnia, M., Watanabe, A., 1992. Sheet flow under nonlinear waves and currents. *Coastal Engineering*  
711 Proceedings 1.

712 Dohmen-Janssen, C., Kroekenstoel, D.F., Hassan, W.N., Ribberink, J.S., 2002. Phase lags in oscillatory  
713 sheet flow: experiments and bed load modelling. *Coastal Engineering* 46, 61–87.

714 Drake, T.G., Calantoni, J., 2001. Discrete particle model for sheet flow sediment transport in the nearshore.  
715 *Journal of Geophysical Research: Oceans* 106, 19859–19868.

716 Durran, D.R., 2010. Numerical Methods for Fluid Dynamics: With Applications to Geophysics. volume 32  
717 of Texts in Applied Mathematics. Springer.

718 Elgar, S., Gallagher, E.L., Guza, R.T., 2001. Nearshore sandbar migration. *Journal of Geophysical Research:*

719 Oceans 106, 11623–11627.

720 Faria, A., Thornton, E., Lippmann, T., Stanton, T., 2000. Undertow over a barred beach. *Journal of*  
721 *Geophysical Research* 105, 16999–17010.

722 Feddersen, F., Trowbridge, J., 2005. The effect of wave breaking on surf-zone turbulence and alongshore  
723 currents: A modeling study. *Journal of Physical Oceanography* 35.

724 Grasso, F., Michallet, H., Barthélemy, E., 2011. Sediment transport associated with morphological beach  
725 changes forced by irregular asymmetric, skewed waves. *Journal of Geophysical Research: Oceans* 116.

726 Hsu, T.J., Hanes, D.M., 2004. Effects of wave shape on sheet flow sediment transport. *Journal of Geophysical*  
727 *Research: Oceans* 109.

728 Isobe, M., Horikawa, K., 1982. Study on water particle velocities of shoaling and breaking waves. *Coastal*  
729 *Engineering in Japan* 25, 109–123.

730 Kalra, T., Sherwood, C., Warner, J., Rafati, Y., Hsu, T.J., 2019. Investigating bedload transport under  
731 asymmetrical waves using a coupled ocean-wave model. pp. 591–604.

732 Kim, Y., Mieras, R.S., Cheng, Z., Anderson, D., Hsu, T.J., Puleo, J.A., Cox, D., 2019. A numerical  
733 study of sheet flow driven by velocity and acceleration skewed near-breaking waves on a sandbar using  
734 sedwavefoam. *Coastal Engineering* 152, 103526.

735 Klemp, J.B., Skamarock, W.C., Ha, S., 2018. Damping acoustic modes in compressible horizontally explicit  
736 vertically implicit (hevi) and split-explicit time integration schemes. *Monthly Weather Review* 146, 1911  
737 – 1923.

738 Larsen, B.E., van der A, D.A., van der Zanden, J., Ruessink, G., Fuhrman, D.R., 2020. Stabilized rans  
739 simulation of surf zone kinematics and boundary layer processes beneath large-scale plunging waves over  
740 a breaker bar. *Ocean Modelling* 155, 101705.

741 Li, M.Z., Amos, C.L., 2001. Sedtrans96: the upgraded and better calibrated sediment-transport model for  
742 continental shelves. *Computers and Geosciences* 27, 619–645. *Numerical Models of Marine Sediment*  
743 *Transport and Deposition*.

744 Lim, G., Jayaratne, R., Shibayama, T., 2020. Suspended sand concentration models under breaking waves:  
745 Evaluation of new and existing formulations. *Marine Geology* 426, 106197.

746 Lin, P., Liu, P.L.F., 1998. A numerical study of breaking waves in the surf zone. *Journal of Fluid Mechanics*  
747 359, 239–264.

748 Long, W., Kirby, J.T., Hsu, T.J., . Cross shore sandbar migration predicted by a time domain boussinesq  
749 model incorporating undertow. pp. 2655–2667.

750 Malarkey, J., Davies, A., 2012. Free-stream velocity descriptions under waves with skewness and asymmetry.  
751 *Coastal Engineering* 68, 78–95.

752 Marchesiello, P., Auclair, F., Debreu, L., McWilliams, J., Almar, R., Benschila, R., Dumas, F., 2021. Tridi-

753      mensional nonhydrostatic transient rip currents in a wave-resolving model. *Ocean Modelling* 163, 101816.

754 Marchesiello, P., McWilliams, J.C., Shchepetkin, A., 2001. Open boundary conditions for long-term inte-  
755      gration of regional oceanic models. *Ocean modelling* 3, 1–20.

756 Mayer, S., Madsen, P.A., 2000. Simulation of Breaking Waves in the Surf Zone using a Navier-Stokes Solver.  
757      pp. 928–941.

758 McWilliams, J.C., Restrepo, J.M., Lane, E.M., 2004. An asymptotic theory for the interaction of waves and  
759      currents in coastal waters. *Journal of Fluid Mechanics* 511, 135–178.

760 Michallet, H., Ruessink, B.G., Vieira Lima Matias da Rocha, M., De Bakker, A., Van Der A, D.A., Ruju,  
761      A., Silva, P.A., Sénéchal, N., Marieu, V., Tissier, M., Almar, R., Abreu, T., Birrien, F., Vignal, L.,  
762      Barthélemy, E., Mouazé, D., Cienfuegos, R., Wellens, P., 2014. GLOBEX: Wave dynamics on a shallow  
763      sloping beach, in: *HYDRALAB IV Joint User Meeting, Lisbon, July 2014, Lisbonne, Portugal*. pp. 1–12.

764 Nam, P.T., Staneva, J., Thao, N.T., Larson, M., 2020. Improved calculation of nonlinear near-bed wave  
765      orbital velocity in shallow water: Validation against laboratory and field data. *Journal of Marine Science*  
766      and Engineering 8.

767 Otsuka, J., Saruwatari, A., Watanabe, Y., 2017. Vortex-induced suspension of sediment in the surf zone.  
768      *Advances in Water Resources* 110, 59–76.

769 Rafati, Y., Hsu, T.J., Elgar, S., Raubenheimer, B., Quataert, E., van Dongeren, A., 2021. Modeling the  
770      hydrodynamics and morphodynamics of sandbar migration events. *Coastal Engineering* 166, 103885.

771 Reniers, A.J.H.M., Roelvink, J.A., Thornton, E.B., 2004. Morphodynamic modeling of an embayed beach  
772      under wave group forcing. *Journal of Geophysical Research: Oceans* 109.

773 Ribas, F., de Swart, H., Calvete, D., Falqués, A., 2011. Modeling waves, currents and sandbars on natural  
774      beaches: The effect of surface rollers. *Journal of Marine Systems* 88, 90–101. 41st International Liege  
775      Colloquium on Ocean Dynamics”Science based management of the coastal waters.

776 Roelvink, D., Roelvink, J., Reniers, A., 2012. *A Guide to Modeling Coastal Morphology*. *Advances in*  
777      *coastal and ocean engineering*, World Scientific.

778 Roelvink, J., 2006. Coastal morphodynamic evolution techniques. *Coastal Engineering* 53, 277–287. *Coastal*  
779      *Hydrodynamics and Morphodynamics*.

780 Roelvink, J.A., Reniers, 1995. IP 11D delta flume experiments : a dataset for profile model validation. WL  
781      / Delft Hydraulics.

782 Roelvink, J.A., Stive, M.J.F., 1989. Bar-generating cross-shore flow mechanisms on a beach. *Journal of*  
783      *Geophysical Research: Oceans* 94, 4785–4800.

784 Ruessink, B.G., Michallet, H., Abreu, T., Sancho, F., Van der A, D.A., Van der Werf, J.J., Silva, P.A., 2011.  
785      Observations of velocities, sand concentrations, and fluxes under velocity-asymmetric oscillatory flows.  
786      *Journal of Geophysical Research: Oceans* 116.



787 Schnitzler, B., 2015. Modeling sand transport under breaking waves.

788 Scott, N.V., Hsu, T.J., Cox, D., 2009. Steep wave, turbulence, and sediment concentration statistics beneath  
789 a breaking wave field and their implications for sediment transport. *Continental Shelf Research* 29, 2303–  
790 2317.

791 Shafiei, H., Chauchat, J., Bonamy, C., Marchesiello, P., 2022. Adaptation of the santoss transport formula  
792 for 3d nearshore models: application to cross-shore sandbar migration. *Ocean Modelling in revision*.

793 Shchepetkin, A.F., McWilliams, J.C., 2005. The regional oceanic modeling system (roms): a split-explicit,  
794 free-surface, topography-following-coordinate oceanic model. *Ocean Modelling* 9, 347–404.

795 Smith, J.D., McLean, S.R., 1977. Spatially averaged flow over a wavy surface. *Journal of Geophysical*  
796 *Research* (1896-1977) 82, 1735–1746.

797 Soufflet, Y., Marchesiello, P., Lemarié, F., Jouanno, J., Capet, X., Debreu, L., Benshila, R., 2016. On  
798 effective resolution in ocean models. *Ocean Modelling* 98, 36–50.

799 Soulsby, R., 1995. Bed shear stresses due to combined waves and currents, in: Stive, M., Fredsøe, J.,  
800 Hamm, L., Soulsby, R., Teisson, C., Winterwerp, J. (Eds.), *Advances in Coastal Morphodynamics*, Delft  
801 *Hydraulics*, Delft, The Netherlands. pp. 420–423.

802 Soulsby, R., 1997. *Dynamics of marine sands*. Thomas Telford Publishing.

803 Soulsby, R., Whitehouse, R., 1997. Threshold of sediment motion in coastal environments.

804 Stive, M., 1986. A model for cross-shore sediment transport, in: *Proceedings of the 20th International*  
805 *Conference on Coastal Engineering*, American Society of Civil Engineers, New York.

806 Suarez, L., Barthelemy, E., Berni, C., Chauchat, J., Michallet, H., Cienfuegos, R., 2014. Vertical distribution  
807 of skewness and asymmetry in a boundary layer on a mobile bed. experiment and  $k-\omega$  comparison. *La*  
808 *Houille Blanche* , 88–94.

809 Svendsen, I., 1987. Analysis of surf zone turbulence. *Journal of Geophysical Research: Oceans* 92, 5115–5124.

810 Thornton, E.B., Humiston, R.T., Birkemeier, W., 1996. Bar/trough generation on a natural beach. *Journal*  
811 *of Geophysical Research: Oceans* 101, 12097–12110.

812 Ting, F.C., Kirby, J.T., 1994. Observation of undertow and turbulence in a laboratory surf zone. *Coastal*  
813 *Engineering* 24, 51–80.

814 van der A, D.A., Ribberink, J.S., van der Werf, J.J., O’Donoghue, T., Buijsrogge, R.H., Kranenburg, W.M.,  
815 2013. Practical sand transport formula for non-breaking waves and currents. *Coastal Engineering* 76,  
816 26–42.

817 van der Zanden, J., van der A, D., Hurther, D., Cáceres, I., O’Donoghue, T., Ribberink, J., 2017. Suspended  
818 sediment transport around a large-scale laboratory breaker bar. *Coastal Engineering* 125, 51–69.

819 van Rijn, L., Walstra, D., Grasmeijer, B., Sutherland, J., Pan, S., Sierra, J., 2003. The predictability of  
820 cross-shore bed evolution of sandy beaches at the time scale of storms and seasons using process-based

821 profile models. *Coastal Engineering* 47, 295–327.

822 Veen, R., 2014. The implementation and testing of the santoss sand transport model in delft3d.

823 Voulgaris, G., Collins, M., 2000. Sediment resuspension on beaches: response to breaking waves. *Marine*  
824 *Geology* 167, 167–187.

825 Warner, J., Sherwood, C., Arango, H., Signell, R., 2005. Performance of four turbulence closure methods  
826 implemented using a generic length scale method. *Ocean Modelling* 8, 81–113.

827 Warner, J.C., Defne, Z., Haas, K., Arango, H.G., 2013. A wetting and drying scheme for ROMS. *Computers*  
828 *and Geosciences* 58, 54 – 61.

829 Warner, J.C., Sherwood, C.R., Signell, R.P., Harris, C.K., Arango, H.G., 2008. Development of a three-  
830 dimensional, regional, coupled wave, current, and sediment-transport model. *Computers & Geosciences*  
831 34, 1284–1306.

832 Watanabe, A., 1982. Numerical models of nearshore currents and beach deformation. *Coastal Engineering*  
833 *in Japan* 25, 147–161.

834 Watanabe, Y., Saeki, H., 1999. Three-dimensional large eddy simulation of breaking waves. *Coastal Engi-*  
835 *neering Journal* 41, 281–301.

836 van der Werf, J., Veen, R., Ribberink, J., van der Zanden, J., 2015. Testing of the new santoss transport  
837 formula in the Delft3d morphological modeling system.

838 Wiberg, P.L., Rubin, D.M., 1989. Bed roughness produced by saltating sediment. *Journal of Geophysical*  
839 *Research: Oceans* 94, 5011–5016.

840 Wilcox, D.C., 1988. Reassessment of the scale-determining equation for advanced turbulence models. *AIAA*  
841 *Journal* 26, 1299–1310.

842 Zijlema, M., Stelling, G., Smit, P., 2011. Swash: An operational public domain code for simulating wave  
843 fields and rapidly varied flows in coastal waters. *Coastal Engineering* 58, 992 – 1012.

Table 1: LIP experiment conditions from Roelvink and Reniers (1995)

Experiment	Initial Geometry	$H_S$ [m]	$T_P$ [s]	Duration [h]	Collection [h]
LIP11-1A	Initial beach profile	0.9	5	12	
LIP11-1B	Result of 1A	1.4	5	18	8
LIP11-1C	Result of 1B	0.6	8	13	7

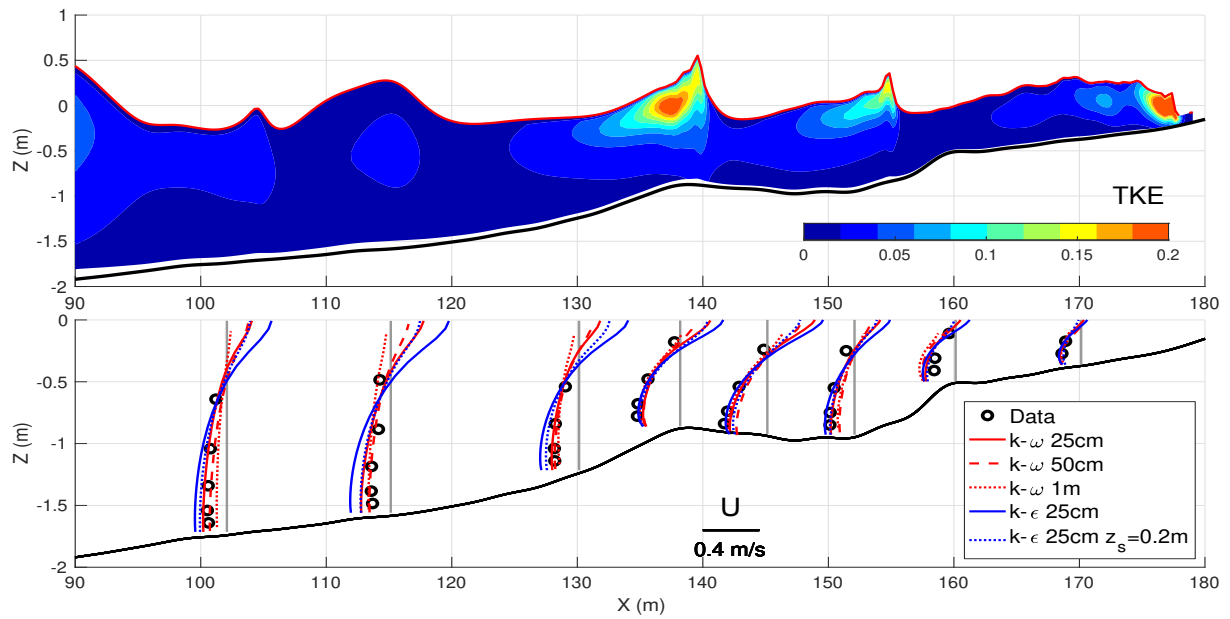


Figure 1: Model comparison with the large-scale LIP11-1B Flume experiment. Top) Snapshot of wave height and turbulent kinetic energy  $k$  [ $m^2/s^2$ ] from the reference model simulation (25 cm resolution;  $k$ - $\omega$  turbulence model). Bottom) Comparison of simulated and measured cross-shore current profiles: sensitivity to resolution (25 cm, 50 cm and 1 m) and turbulence models ( $k$ - $\omega$  in red;  $k$ - $\epsilon$  in blue).

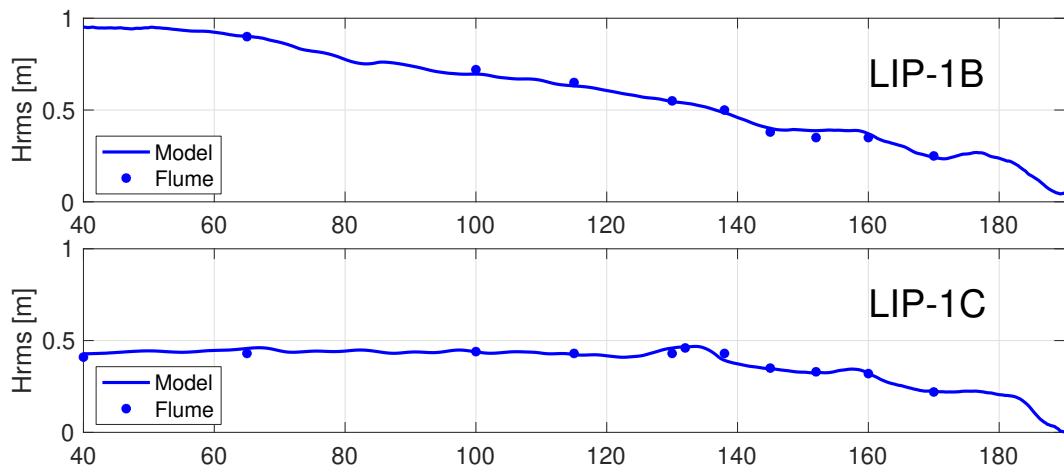


Figure 2: Model-data comparison of  $H_{rms}$  in LIP Flume experiment (top: 1B; bottom: 1C).

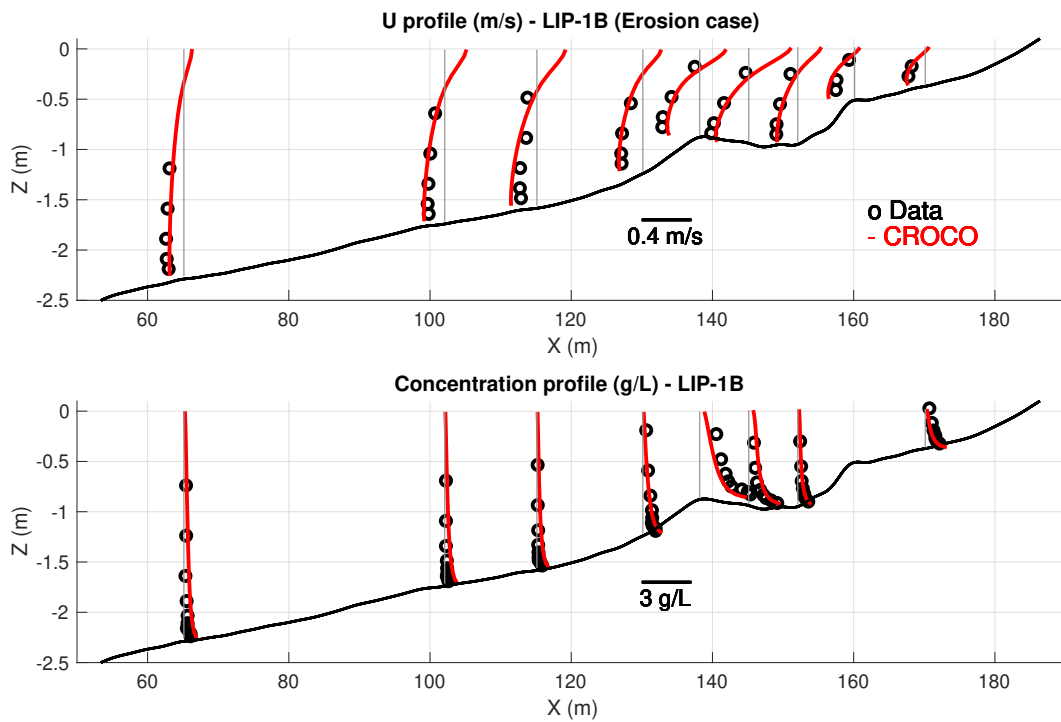


Figure 3: Model-data comparison of wave-mean cross-shore velocity and sand concentration profiles in the erosive LIP11-1B Flume experiment.

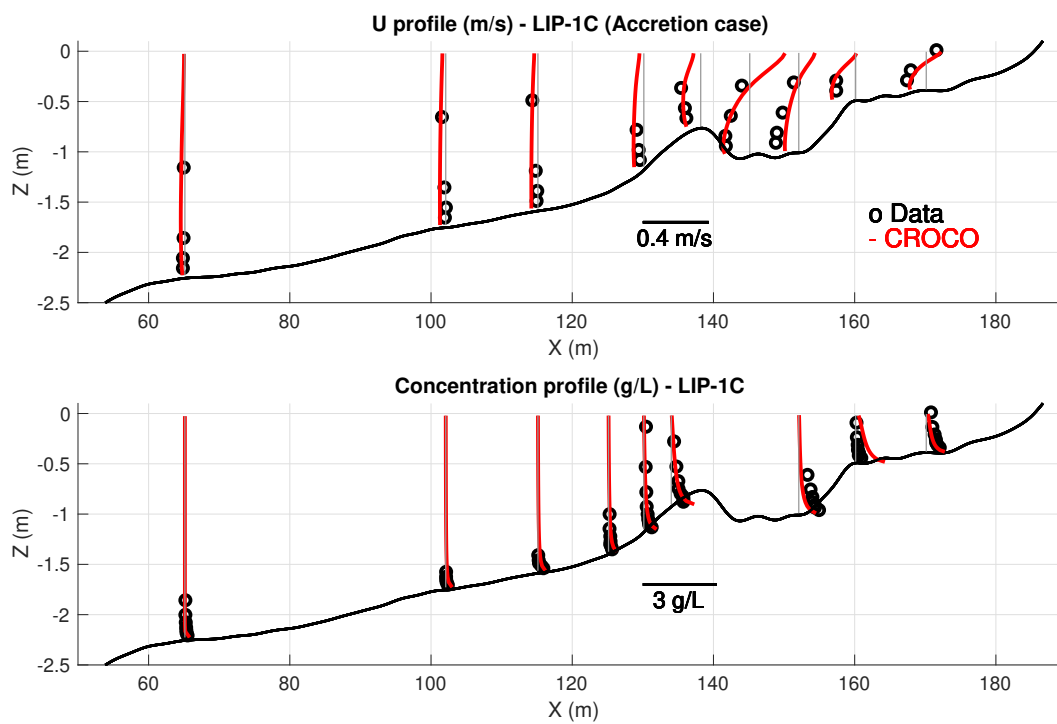


Figure 4: Model-data comparison of wave-mean cross-shore velocity and sand concentration profiles in the accretive LIP11-1C Flume experiment.

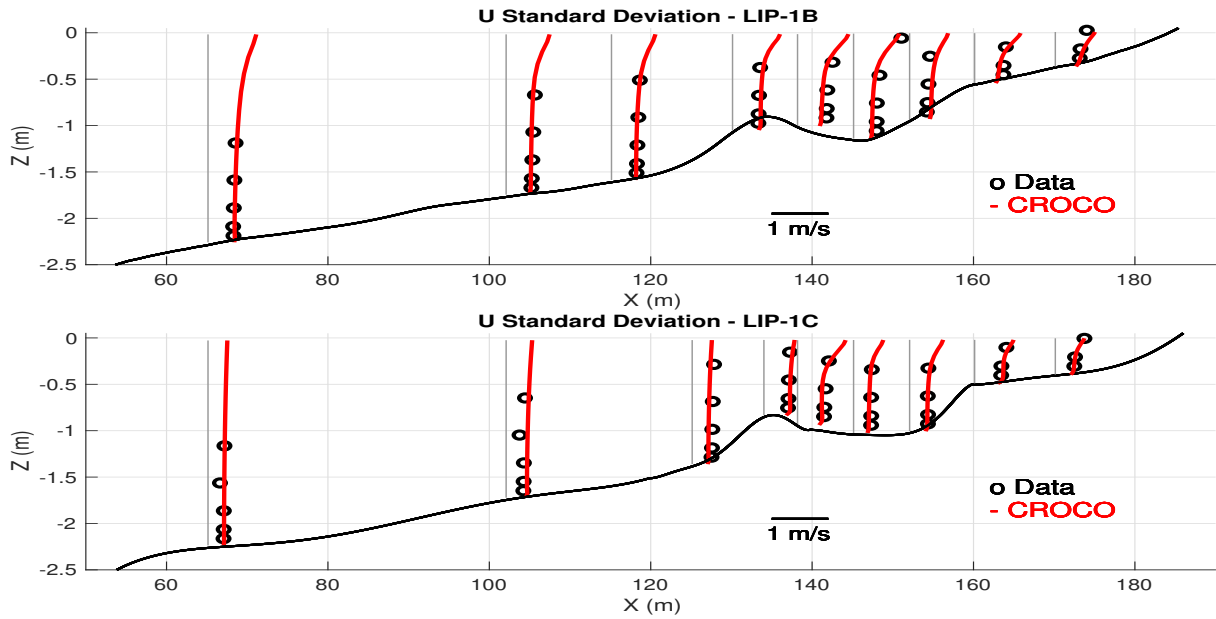


Figure 5: Model-data comparison of velocity standard deviation profiles in the LIP11-1B and LIP11-1C experiments.

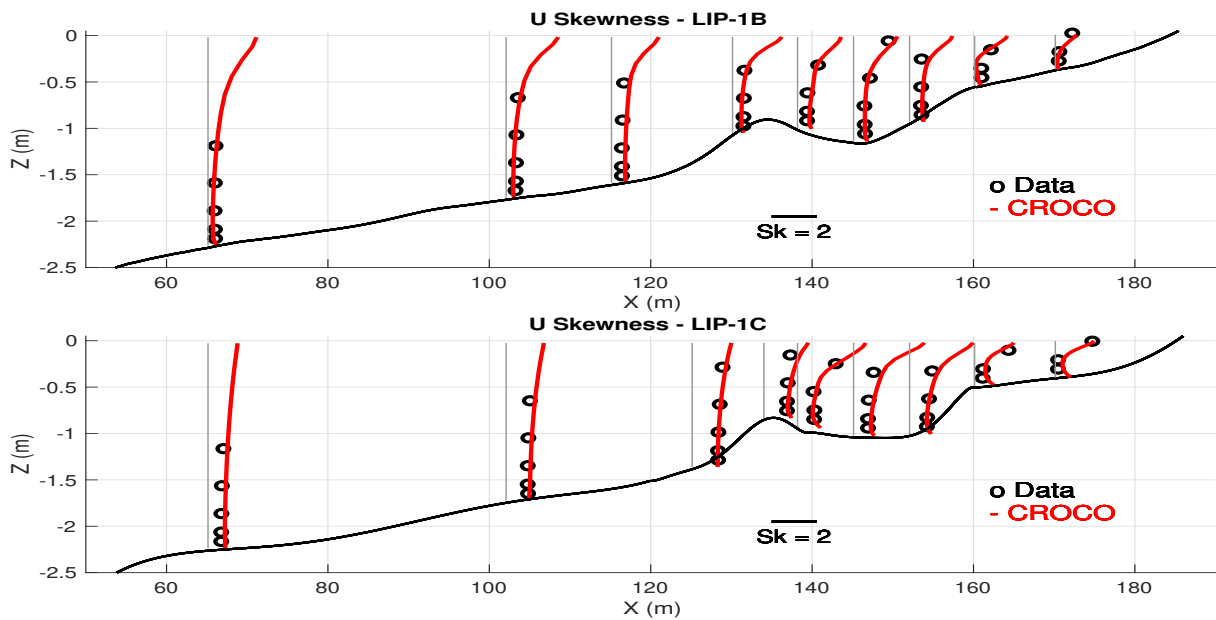


Figure 6: Model-data comparison of velocity skewness profiles in the LIP11-1B and LIP11-1C experiments.

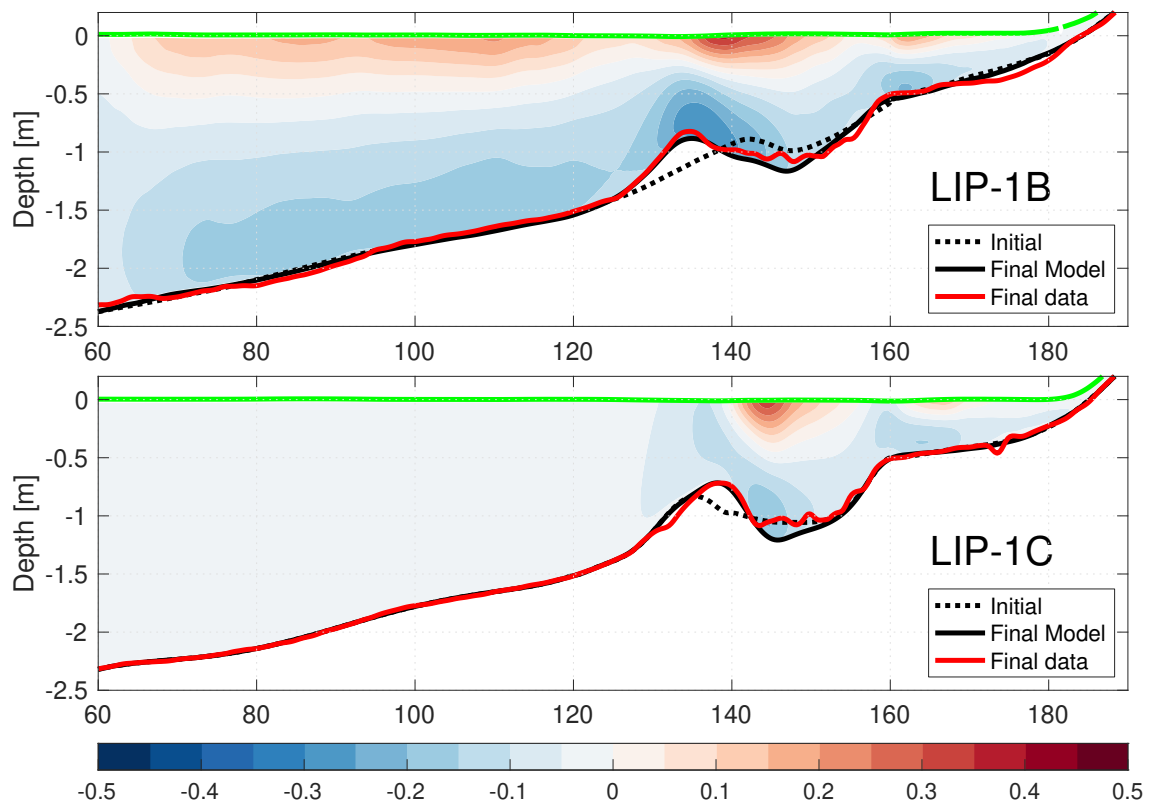


Figure 7: Model comparison of morphological changes with LIP measurements during the erosion phase (LIP-1B) and accretion phase (LIP-1C). Wave-mean cross-shore velocities are also shown in filled contours [m/s].



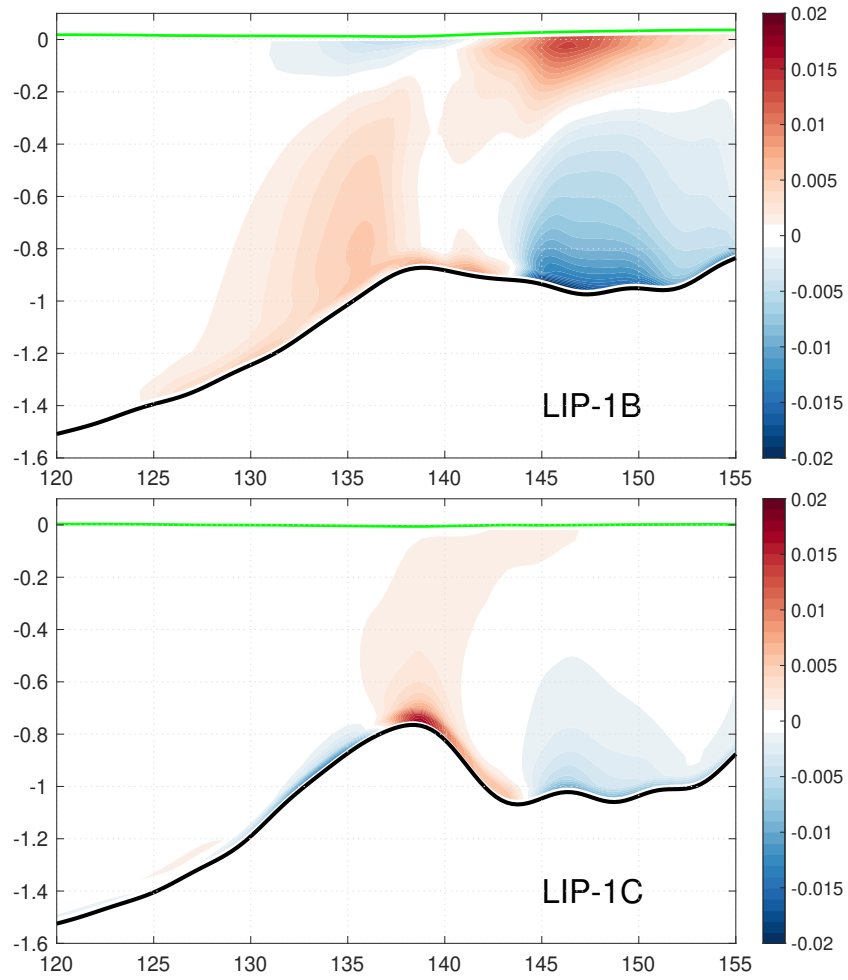


Figure 8: Wave-mean sand advection [ $\text{kg}/\text{m}^3/\text{s}$ ] halfway through the erosion phase (LIP-1B) and accretion phase (LIP-1C). Positive (negative) values indicate a gain (loss) in sand concentration due to transport.

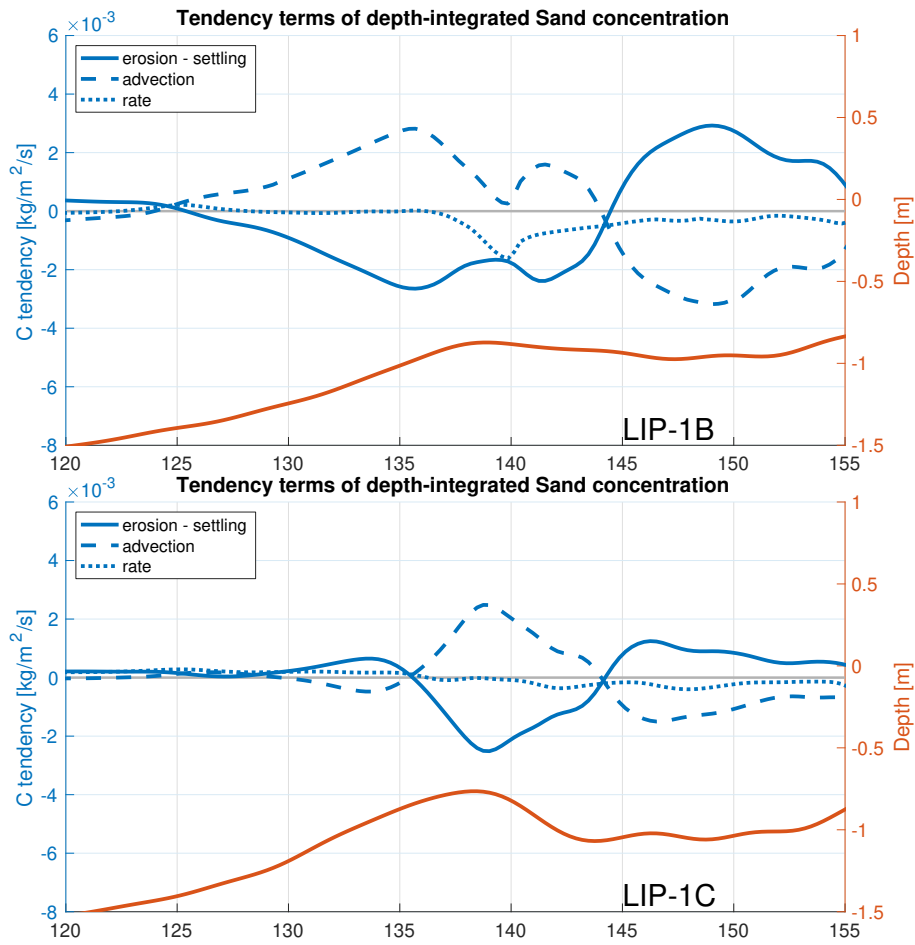


Figure 9: Depth-integrated budget of sand concentration halfway through the erosion phase (LIP-1B) and accretion phase (LIP-1C).

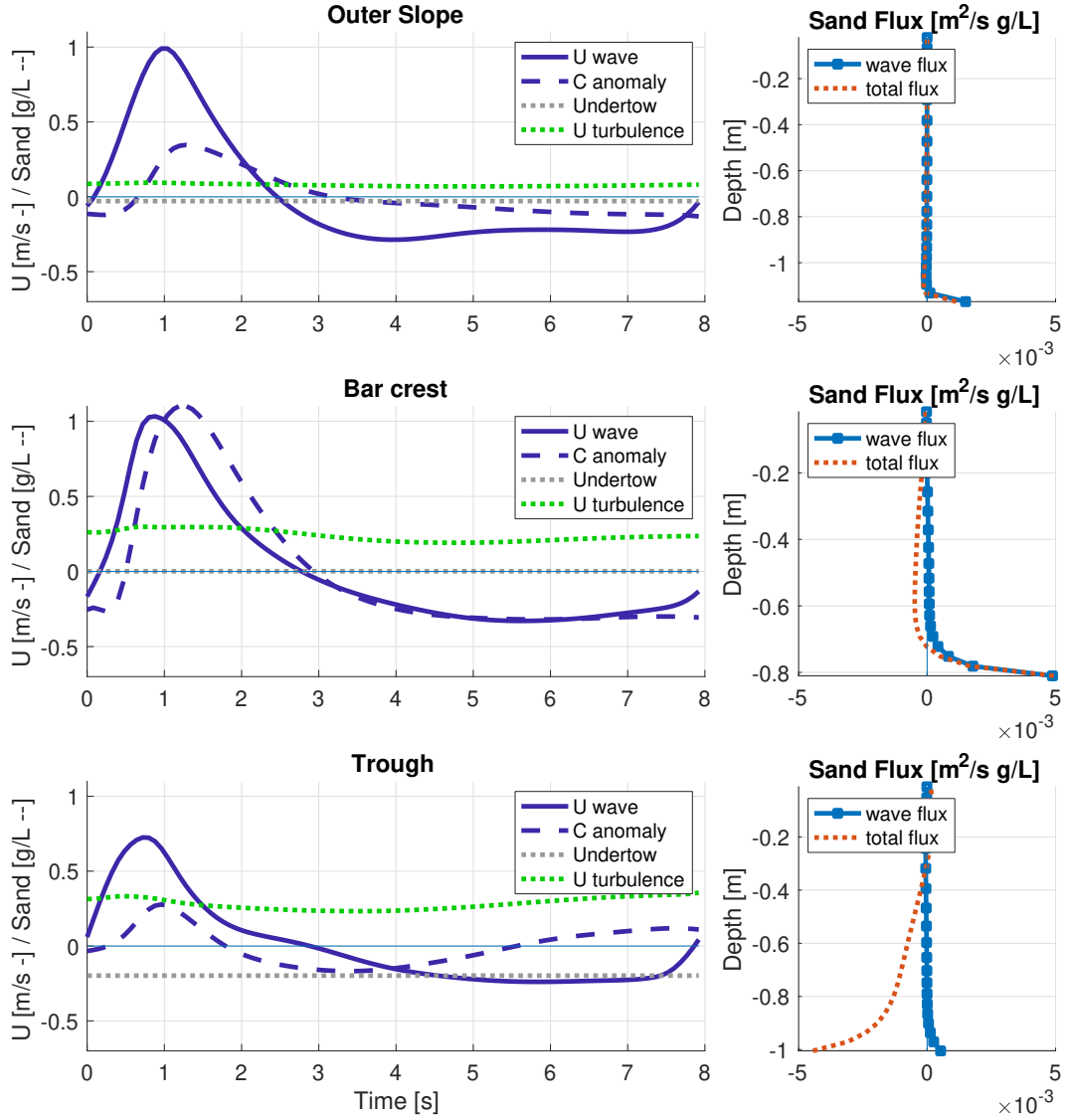


Figure 10: Left panels: ensemble-averaged wave cycle of current and sand concentration anomaly ( $u' = u - \bar{u}$  and  $C' = C - \bar{C}$ ) at three positions over the sandbar during the accretive LIP-1C Flume experiment. The wave-mean undertow  $u_b$  and mean effective turbulent velocity  $u_t$  at the bottom level  $z_b$  are also represented. Right panels: profiles of suspended sand flux (total flux:  $\overline{uC}$ ; and wave flux:  $\overline{u'C'}$ ).

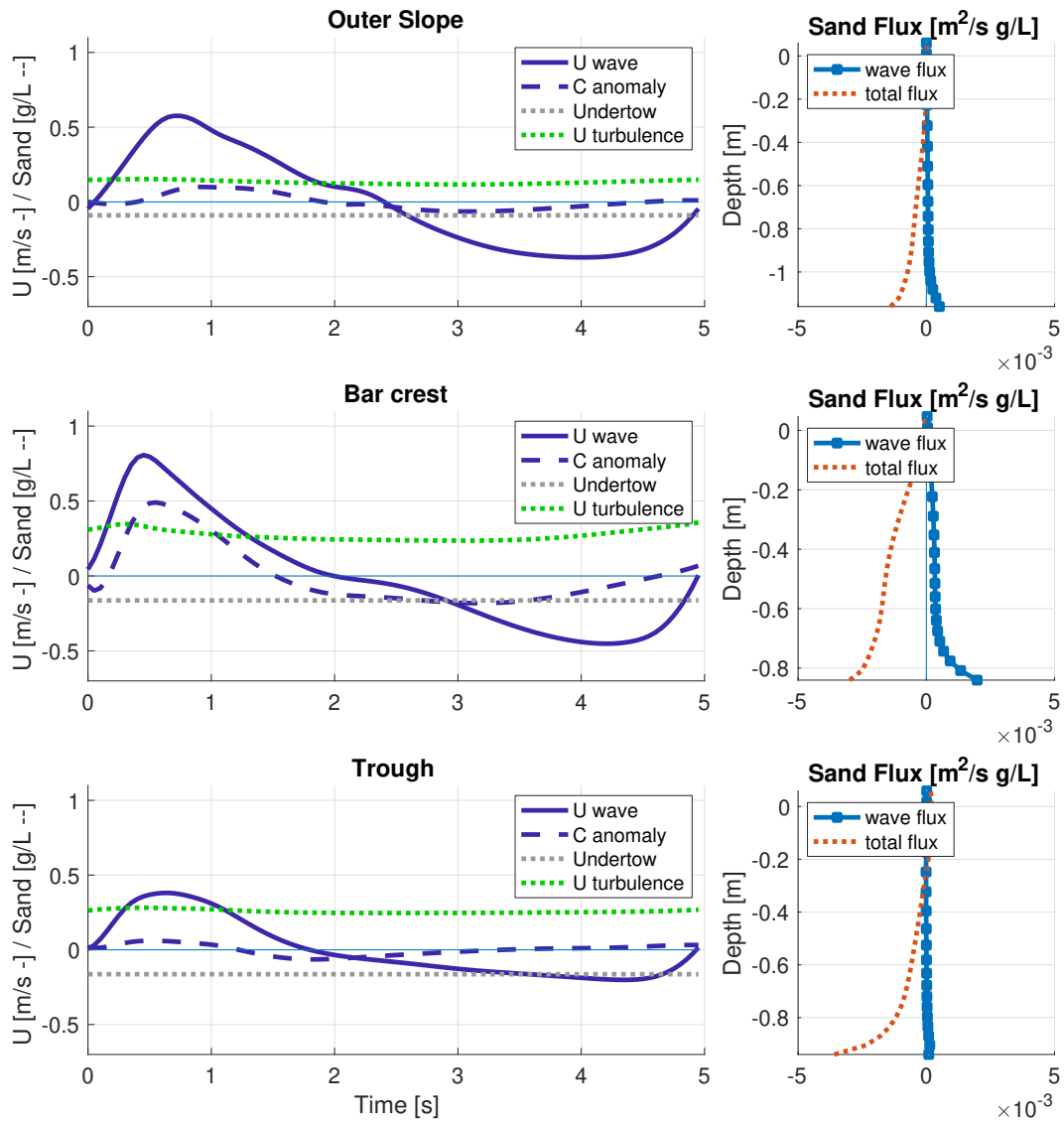


Figure 11: Same as Figure 10 but for the LIP-1B Flume experiment.

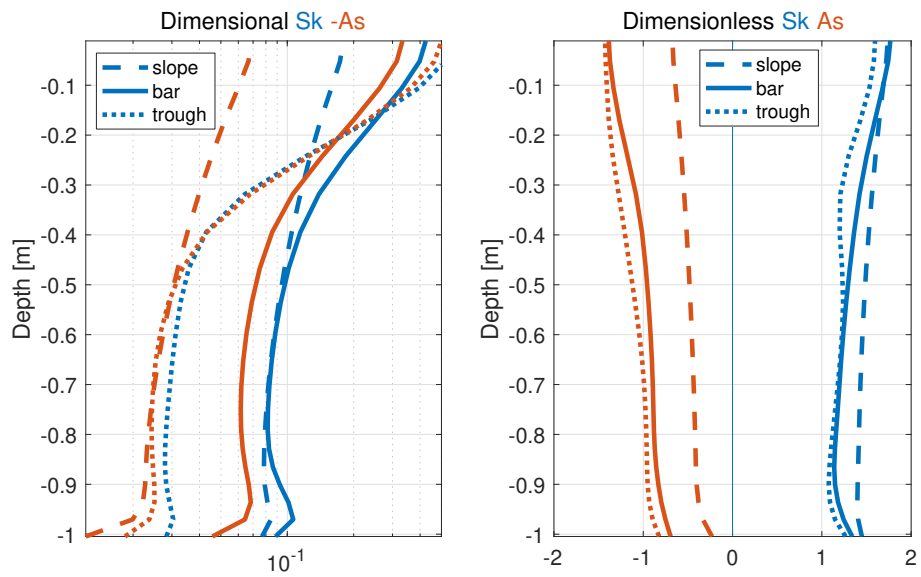


Figure 12: Ensemble-averaged analysis of wave asymmetry and skewness profiles (absolute and normalized) at three locations around the sandbar, during the accretive LIP-1C Flume experiment.

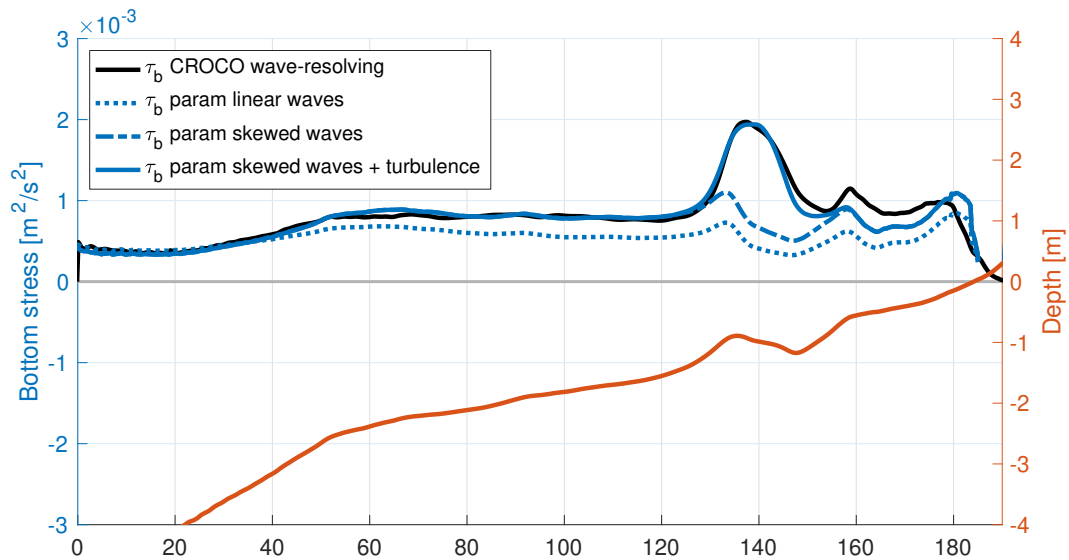


Figure 13: Bed shear stress profiles from the wave-resolving model and reconstructed from empirical formulations using wave-averaged values and linear wave theory with or without skewness correction.

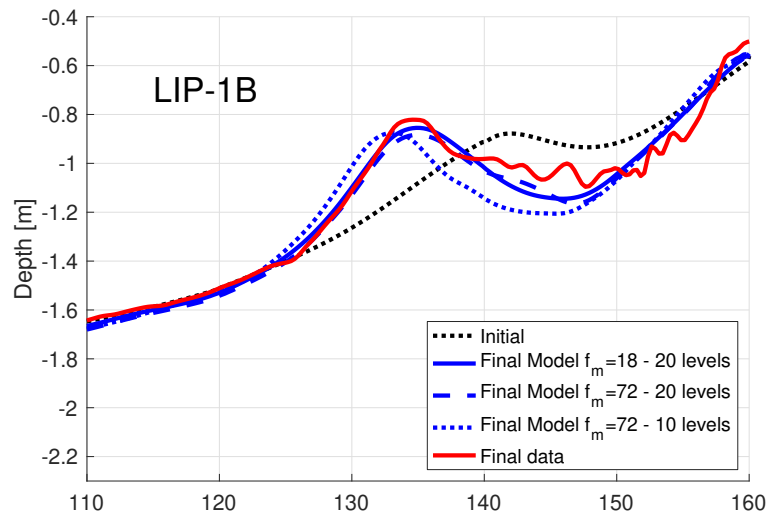


Figure 14: Model sensitivity to morphological acceleration parameter  $f_m$  and vertical grid resolution (number of vertical levels) during the erosion phase (LIP-1B).