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3D wave-resolving simulation of sandbar migration

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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 model accurately simulates the successive offshore and onshore bar migration observed in a large-scale flume experiment (LIP11D) in response to wave forcing representing storm and post-storm (recovery) conditions. The diagnosis of sand transport and the analysis of a composite asymmetric wave cycle reveal the migration mechanisms in each phase. In all cases, sediment resuspension is dominated by breaking-induced turbulence, while net sediment transport and bed profile evolution are primarily the result of undertow distribution across the sandbar, rather than a trade-off between onshore and offshore fluxes. In the erosion phase, a strong undertow carries the mobilized sediment seaward of the bar crest. In the accretion phase, the sandbar becomes the breaking point to more moderate waves and the undertow is limited to the lee-side of the bar, causing an counterflow migration of the bar crest. The contribution of wave-related onshore fluxes is significant in this case – although secondary in magnitude – and coincide with higher mobilization and currents during the wave crest period. We conclude that computationally efficient 3D wave-resolving models (including morphological acceleration) can be used to improve our understanding of nearshore morphodynamic problems in realistic applications.

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

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). The advantage of numerical models is their ease of application, but they usually rely 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 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 skewed waves (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 asymmetry due to possibly higher turbulence intensity during the wave crest period (Ting and Kirby, 1994), although this process is disputed (Scott et al., 2009). 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. It will confirm the existence of wave-related onshore transport (especially in the accretion phase) but will show that sediment transport around the sandbar depends primarily on the cross-shore distribution of the undertow, which varies with wave height.

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28 Two-phase flow models (water and sediment) applied to individual wave breaking can
29 be used to describe hydro-morphodynamics in much details (Cheng et al., 2017; Kim et al.,
30 2019), but are computationally very expensive. Therefore, the study of sandbar formation
31 and migration was made in the past with much simpler, wave-averaged models (Watan-
32 abe, 1982; Stive, 1986; Roelvink and Stive, 1989; van Rijn et al., 2003), which are still in
33 use — XBeach being a classic example (Rafati et al., 2021). These models rely heavily on
34 parametrizations to account for both onshore and offshore sandbar migration, i.e., a balance
35 between offshore transport by the undertow and onshore transport by skewed/asymmetric
36 waves and wave streaming. Depth-dependent nearshore models (McWilliams et al., 2004)
37 have improved the representation of the undertow, but the effects of wave asymmetry/skewness
38 remain parametrized as bedload transport in the wave boundary layer (e.g., van der Werf
39 et al. 2015 for Delft3D; Kalra et al. 2019 for ROMS). Recently, empirical formulations have
40 improved, using the concept of half-wave cycles (Dibajnia and Watanabe, 1992; van der A
41 et al., 2013), and are implemented with mixed success (Veen, 2014; Schnitzler, 2015). As an
42 intermediate approach between expensive wave-resolving two-phase flow models and highly
43 parametrized wave-averaged models, we propose to apply a wave-resolving, free-surface, cou-
44 pled hydro-sedimentary model. CROCO (Coastal and Regional Ocean Community model)
45 has recently been adapted to nearshore wave dynamics and circulation problems (March-
46 esiello et al., 2021). This cost-effective 3D wave-resolving model can explicitly simulated
47 the vertical profiles of oscillatory cross-shore sediment transport, allowing assessment of the
48 combined effects of mean undertow and asymmetric waves around a sandbar.

49 This paper presents a brief overview of the CROCO coupled hydro-sedimentary model,
50 before discussing its application to a large-scale flume experiment (LIP11D), and an analysis
51 of offshore and onshore sandbar migration under wave forcing representing storm and post-
52 storm (recovery) conditions.

53 **2. Model description**

54 Due to limited computational resources, 3D wave-resolving models are still rarely used to
55 study nearshore dynamics in realistic environments. Early applications starting in the 1990s

56 used the volume-of-fluid (VOF) method for free-surface tracking (e.g., Lin and Liu 1998;
57 Watanabe and Saeki 1999; Derakhti and Kirby 2014; Larsen et al. 2020). This model type
58 with Cartesian coordinate, where the free surface crosses computational cells arbitrarily,
59 fails to precisely apply the pressure boundary condition on the free surface, affecting the
60 model accuracy.

61 More recently, several 3D wave-resolving, free-surface and terrain-following RANS mod-
62 els have emerged for the nearshore zone, e.g., SWASH (Zijlema et al., 2011), NHWAVE
63 (Derakhti et al., 2016), and CROCO (Marchesiello et al., 2021). In this case, the explicit
64 overturning of the free surface is excluded and the breaking wave is modeled with a single-
65 valued free surface which follows a shock process and resembles a dissipating bore. Despite
66 the absence of explicit overturning (replaced by parametrized turbulence), these models can
67 be accurate as well as computationally efficient (orders of magnitude cheaper) in the study
68 of waves and wave-driven mean and transient circulation.

69 CROCO (www.croco-ocean.org) belongs to this class of models but, contrary to the
70 other attempts, it solves the compressible Navier-Stokes equations (Auclair et al., 2018).
71 A compressible approach preserves the hyperbolic nature of Navier-Stokes equations and
72 does not require a global elliptic solver with incremental pressure corrections to ensure the
73 incompressible mass balance. As a result, it avoids splitting errors between pressure and
74 velocity and approximations made on free-surface conditions (Zijlema et al., 2011; Derakhti
75 et al., 2016), thereby preserving amplitude and nonlinear dispersive properties of surface
76 waves. In the same time, the absence of global computations by an elliptic solver makes
77 optimization and parallelization procedures much more efficient (excellent scalability even
78 on massively parallel computers). The cost of solving acoustic waves is managed with a
79 time-splitting technique and semi-implicit time discretization.

80 *2.1. Model equations*

81 The full set of Navier-Stokes equations for a free-surface ocean is explicitly integrated
82 in the nonhydrostatic, non-Boussinesq (compressible) version of CROCO, built on the code
83 structure of ROMS primitive equation solver (Shchepetkin and McWilliams, 2005; Debreu

84 et al., 2012). Non-Boussinesq equations include the momentum and continuity equations,
 85 the surface kinematic relation (for free surface), heat, salt or other tracer C (such as sediment
 86 concentration) conservation equations, and the equation of state, which reads in Cartesian
 87 coordinates:

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

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

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

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

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

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

88 (u, v, w) are the (x, y, z) components of vector velocity \vec{v} ; η is the free surface; P the total
 89 pressure; ρ the density; $f(x, y)$ and $\tilde{f}(x, y)$ are the traditional and non-traditional Coriolis
 90 parameters, function of latitude; g is acceleration of gravity; $\mathcal{D}_u, \mathcal{D}_v, \mathcal{D}_C$ are eddy-diffusion
 91 terms requiring second-moment turbulence closure models; $\mathcal{F}_u, \mathcal{F}_v, \mathcal{F}_C$ are forcing terms; λ is
 92 the second (bulk) viscosity, associated with compressibility (used to damp acoustic waves).

93 In this study, movements are produced along a channel by a wave generator in a ho-
 94 mogeneous non-rotating fluid. In this case, the longitudinal flow equation v disappears and
 95 the Coriolis force, the baroclinic pressure force, and all surface fluxes are zero. There is no
 96 temperature or salinity stratification so that the associated density is constant in time and
 97 space (but not the dynamic density associated with acoustic waves).

98 *2.2. Computational procedure*

99 In the above set of equations, a relation between ρ and P is required. To that end, and
100 as part of a time-splitting approach, density is decomposed into slow and fast components
101 based on a first-order linear decomposition with respect to total pressure. The Navier-
102 Stokes equations are then integrated with two different time steps within the time-splitting
103 approach inherited from ROMS. The slow-mode integration is similar to ROMS, with the
104 addition of the slow part of vertical momentum equation, while fast-mode integration is
105 in 3D and includes the compressible terms of momentum and continuity equations. More
106 details can be found in Auclair et al. (2018) and Marchesiello et al. (2021).

107 Note that acoustic waves can become pseudo-acoustic if their phase speed $c_s = \frac{\partial P}{\partial \rho}$ is
108 artificially reduced (to be safe, no less than about 5 times the shallow-water phase speed).
109 In this case, high-frequency processes associated with bulk compressibility may be non-
110 physical, but an accurate solution for slower nonhydrostatic dynamics can be preserved,
111 while relaxing CFL constraints. In our nearshore applications, a c_s value of 100 m/s instead
112 of 1500 m/s makes almost no difference for the physical solution but allows a great reduction
113 in the computational time (by half).

114 CROCO is discretized on a C-grid with finite-difference methods for slow and fast modes
115 (Shchepetkin and McWilliams, 2005; Soufflet et al., 2016). In short, the slow-mode time-
116 stepping algorithm is a Leapfrog Adams-Moulton predictor-corrector scheme, that is third-
117 order accurate for integrating advective terms. The fast mode is integrated with a generalized
118 forward-backward scheme, which is also third-order accurate. Vertical flux terms that do
119 not require accuracy (vertical diffusion term in the slow mode and all acoustic terms of
120 the w equation in the fast mode) are computed with an implicit time stepping to increase
121 computational stability.

122 Horizontal and vertical advection terms are discretized using the WENO5-Z improved
123 version of the 5th-order weighted essentially non-oscillatory scheme (Borges et al., 2008),
124 which is popular for hyperbolic problems containing both shocks and smooth structures.
125 WENO5-Z naturally copes with dispersive numerical modes as well as shocks caused by
126 breaking waves, with no need for *ad hoc* criteria.

127 *2.3. Turbulence closure*

128 Along with the numerical treatment of breaking waves, a k - ϵ or k - ω model, solving
 129 the closure equations for turbulent kinetic energy k and dissipation ϵ or dissipation rate
 130 $\omega \propto \epsilon k^{-1}$, is used as part of a Generic Length Scale (GLS) method (Warner et al., 2005).
 131 In the absence of buoyancy forcing, the turbulence equations express a balance between
 132 transport, diffusion, shear production and dissipation:

$$\frac{\partial \rho k}{\partial t} = -\vec{\nabla} \cdot (\rho \vec{v} k) + \mathcal{D}_k + \rho(P - \epsilon) \quad (7)$$

133

$$\frac{\partial \rho \epsilon}{\partial t} = -\vec{\nabla} \cdot (\rho \vec{v} \epsilon) + \mathcal{D}_\epsilon + \rho \frac{\epsilon}{k} (c_{\epsilon 1} P - c_{\epsilon 2} \epsilon) \quad (8)$$

134 OR

$$\frac{\partial \rho \omega}{\partial t} = -\vec{\nabla} \cdot (\rho \vec{v} \omega) + \mathcal{D}_\omega + \rho \frac{\omega}{k} (c_{\omega 1} P - c_{\omega 2} \epsilon) \quad (9)$$

135 The eddy viscosity $\nu_t = c_\mu l k^{\frac{1}{2}}$ is derived from these equations, with coefficient c_μ de-
 136 pendent on stability functions, and mixing length $l \propto k^{\frac{3}{2}} \epsilon^{-1}$. l is resolution independent,
 137 which is consistent with a RANS rather than LES approach. The shear production term
 138 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
 139 notation). All turbulence model parameters are given in Warner et al. (2005), based on
 140 Burchard et al. (1998) for k - ϵ and Wilcox (1988) for k - ω . The only present modification
 141 concerns the surface and bottom mixing lengths, which are model boundary conditions given
 142 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
 143 match the observed sand concentration profiles near the bottom. With the k - ϵ model, the
 144 momentum mixing is also sensitive to z_{0s} , which needs to be increased to 0.2 m in order to
 145 match the observed velocity profiles (see validation section 3.3). For this reason, and for its
 146 robustness through resolutions, the k - ω model will be our standard turbulence model.

147 *2.4. Wave maker at offshore boundary*

148 The wave maker forces a spectrum of linear waves at the offshore boundary, as in Zijlema
 149 et al. (2011). In full 3D applications, the spectrum has frequency and directional spreading

150 Marchesiello et al. (2021) but in this flume experiment, waves are shore normal:

$$\eta_{bc}(t) = \sum_i a_i \cos(\omega_i t + \phi_i) \quad (10)$$

151

$$u_{bc}(z, t) = \eta_{bc}(y, t) \omega_p \frac{\cosh(k_p(z + h))}{\sinh(k_p h)} \quad (11)$$

152 where (x,y,z) are cross-shore, alongshore and vertical directions respectively; (i) is the index
 153 of spectral distribution; a_i is the amplitude at each frequency ω_i , from a given statistical
 154 distribution, e.g., JONSWAP (Sec. 3.2); ω_p and k_p are peak frequency and wavenumber
 155 related by the linear dispersion relation: $\omega_p^2 = g k_p \tanh(k_p h)$ with h the water depth; ϕ_i is
 156 a uniformly distributed random phase.

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

$$\bar{u} = \bar{u}_{bc} - \sqrt{\frac{g}{h}}(\eta - \eta_{bc}) \quad (12)$$

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

169 2.5. Sediment and bed model

170 CROCO comes with capabilities for water quality, marine ecosystem and sediment mod-
 171 eling. They are built upon Eq. 6 for the transport of tracer concentration but, in the case
 172 of sediment modeling, additional sources and sinks are needed to simulate the exchange

173 between the water column and sediment bed (Blaas et al., 2007; Warner et al., 2008). Ne-
 174 glecting compressibility ($\rho \sim \rho_0$), the sediment concentration follows:

$$\underbrace{\frac{\partial C}{\partial t}}_{\text{RATE}} = - \underbrace{\vec{\nabla} \cdot \vec{v} C}_{\text{ADVECTION}} + \underbrace{D_C}_{\text{MIXING}} - \underbrace{\frac{\partial w_s C}{\partial z}}_{\text{SETTLING}} + \underbrace{\frac{E}{\delta z_b}}_{\text{EROSION}} \Big|_{z=z_b} \quad (13)$$

175 w_s is the settling velocity, dependent on sediment grain size, but not on flow conditions and
 176 concentrations (Soulsby, 1997). Settling is computed via a semi-Lagrangian advective flux
 177 algorithm, which is unconditionally stable (Durrant, 2010). It uses a piece-wise parabolic
 178 vertical reconstruction of the suspended sediment for high-order interpolation, with WENO
 179 constraints to avoid oscillations. E is the erosion flux at the sea floor and is only applied to
 180 the first grid level of height z_b and cell size δz_b .

181 In the present study, this suspended sediment model is used with simple settings. It is
 182 composed of a single fine sand class with settling velocity $w_s=2.5 \text{ cm s}^{-1}$ (grain size $d_{50}=0.22$
 183 mm). For resuspension, taking one sediment bed layer for simplicity, only two parameters
 184 are needed: critical shear stress τ_{cr} and erosion rate E_0 at the seafloor, expressed in the
 185 erosion flux (Blaas et al., 2007):

$$E = E_0(1 - p) \frac{\tau_b - \tau_{cr}}{\tau_{cr}} \quad \text{for } \tau_b > \tau_{cr} \quad (14)$$

186 E_0 is set empirically to $5 \cdot 10^{-3} \text{ kg m}^{-2} \text{ s}^{-1}$ (Smith and McLean, 1977); p is the sediment
 187 porosity (0.41); τ_{cr} , the critical shear stress, i.e., the threshold for initiation of sediment
 188 motion, is set to 0.11 N m^{-2} (Soulsby and Whitehouse, 1997). τ_b is the bottom shear stress
 189 estimated from the logarithmic law of the wall:

$$\vec{\tau}_b = \frac{\kappa^2}{\log^2\left(\frac{z_b}{z_{0b}}\right)} |u_b| \vec{u}_b \quad (15)$$

190 $|u_b|$ and \vec{u}_b are the magnitude and vector of oceanic bottom currents in the log layer at a
 191 height z_b above bed; $\kappa = 0.41$ is the Von Karman constant; and z_{0b} is the bottom roughness
 192 length.

193 In the surfzone, the effect of wave-breaking, especially plunging breakers over sandbars,
 194 on sediment resuspension needs to be addressed (Voulgaris and Collins, 2000; van der Zanden

195 et al., 2017; Otsuka et al., 2017; Lim et al., 2020). Lim et al. (2020) suggest a strong
 196 correlation between the maximum sediment concentration observed in the LIP experiment
 197 and the plunging point (about 10 m downstream of the breaking point) over the sandbar.
 198 We adopt here the approach proposed by Reniers et al. (2004) consisting in replacing $|u_b|$
 199 by an stirring velocity reinforced by the turbulence motion induced by wave breaking:

$$\begin{aligned}
 u_{stir} &= \sqrt{u_b^2 + u_t^2} \\
 u_t &= \phi(x) \alpha_{br} \sqrt{k_b}
 \end{aligned}
 \tag{16}$$

200 The subgrid bottom turbulent kinetic energy k_b is given by the turbulence model and α_{br}
 201 is a factor explaining the effectiveness of breaking waves in the entrainment of bottom
 202 sediments. We used a common value of $\alpha_{br} = 10$ (Ribas et al., 2011), which produces sand
 203 concentrations close to observations in both the erosive and accretive phases. $\phi(x)$ is a
 204 Gaussian function centered around the plunging point in order to enhance the turbulence
 205 effect at this position. This part is the only non-generic aspect of the model presented here
 206 but is essential to reproduce the correct profiles of sediment concentrations — and in turn a
 207 requirement for accurate bar migration. We have tried alternative and more generic methods
 208 of resuspension by plunging breakers with some success (see Discussion section) but a more
 209 dedicated study is needed for what appears to be a real challenge.

210 Estimating z_{0b} is not trivial with moving sediment and various empirical formulations
 211 exist that account for roughness over a mobile bed (Wiberg and Rubin, 1989). They generally
 212 assume that grain roughness is the most important predictor for the onset of sand suspension,
 213 but bedload and ripple form roughness are also considered (Li and Amos, 2001). Here we
 214 use grain roughness.

215 For bedload transport, we do not rely on parametrization as skewed-asymmetric waves
 216 are resolved explicitly, but we make sure that the wave-boundary layer is resolved, and
 217 that the first vertical level is in a sheet flow layer (about 10 times the grain size). This is
 218 particularly important for the onshore bar migration phase. Note that in our formulation, the
 219 turbulence intensity (calculated with the closure model) only affects sediment resuspension,
 220 not deposition, as in, for example, Kawanisi and Shiozaki (2008)).

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

$$\frac{\partial h}{\partial t} = -\frac{f_m}{\rho_s(1-p)}(w_s C - E) \quad (17)$$

223 ρ_s is sediment density and f_m is a morphological acceleration factor (Roelvink, 2006) de-
224 scribed below. Morphological changes can have a significant influence on flow and transport
225 when they are greater than a few percent of the water depth. Morphological updating strate-
226 gies are described by Roelvink (2006) and implemented in CROCO following Warner et al.
227 (2008). For dynamical consistency, the vertical velocity is modified by the rate of change of
228 vertical grid levels dz/dt , adjusting to the moving sea floor and free surface (grid "breath-
229 ing" component; Shchepetkin and McWilliams 2005). This method is mass conserving and
230 retains tracer constancy preservation.

231 Morphological acceleration can be achieved with no constraint on the model time step by
232 multiplying erosion and deposition rates by the scale factor f_m at the bed-water interface (Eq.
233 17). Only the fluxes to and from the bed are changed, not the magnitude of the sediment
234 concentrations in the water column. To our knowledge, this paper presents the first use of
235 a morphological acceleration technique with a wave-resolving model. We have found that
236 the technique is effective since we obtain the same result with or without morphological
237 acceleration. We were able to use a large morphological factor f_m (at least up to 64)
238 because the timescale of morphodynamics (here bar migration) is slow compared to that of
239 wave dynamics.

240 **3. Simulation of a large-scale flume experiment**

241 We now present our model solution of sandbar migration, applied to the European Large
242 Installation Plan (LIP) experiments, which was carried out at full scale in Delft Hydraulics's
243 Delta Flume (Roelvink and Reniers, 1995). We first present the model setup for these ex-
244 periments, the model validation of current profiles, then the comparative analysis of offshore
245 and onshore bar migration.

246 In Marchesiello et al. (2021), the model’s ability to simulate surface gravity wave prop-
247 agation, nearshore breaking and the resulting circulation was validated against laboratory
248 experiments. In particular, the GLOBEX experiment (Michallet et al., 2014) provided high-
249 resolution measurements of surface wave height η and bottom currents u_b . A comparison
250 was made of continuous profiles of wave statistics for η and u_b across the beach. The model
251 statistics closely resembled the measurement data (in structure and magnitude), including
252 high-order moments, showing the transition from skewness to asymmetry across the surf-
253 zone. We will not repeat this comparative study here but wish to build on it to emphasize
254 the model’s good representation of wave asymmetry, which is important for the onshore bar
255 migration.

256 3.1. LIP experiment

257 The Flume is 225 m long, 5 m wide and 7 m deep. In LIP, three types of experiments
258 were carried out under different types of irregular waves, which subsequently resulted in a
259 stable (1A), erosive (1B), and accretive (1C) beach state (see Table 1 and Roelvink and
260 Reniers 1995 for details). The initial profile is linear in LIP-1A, with a slope of 1:30 and
261 consisting of a median grain size of 0.22 mm. The final profile of LIP-1A was used as the
262 initial profile of LIP-1B and the final profile of LIP-1B as the initial profile of LIP-1C. The
263 wave conditions were a JONSWAP narrow-banded random wave spectrum generated by a
264 wave paddle, with a peak enhancement factor $\gamma = 3.3$ and characteristic wave height and
265 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
266 wave forcing, the sandbar developed during LIP-1B, increasing in height and migrating in
267 the offshore direction. Under the accretive conditions of LIP-1C, the bar migration reversed
268 to the onshore direction. The bed profile was measured with a profile follower that used an
269 automated sounding system. Wave-averaged current profiles were captured by a movable
270 carriage with attached current meters starting 10 cm above the bed and at 10 locations along
271 the flume (2 cm/s accuracy). Similar profiles of wave-averaged concentrations of suspended
272 sediment were measured by suction tubes mounted on a carriage starting 5 cm above the
273 bed (10% accuracy). For validation of currents and sand concentration, we consider the

274 time 8 hours after initialization in experiment 1B and 7 hours in 1C (Veen, 2014).

275 *3.2. Model setup*

276 The model setup is adapted from the LIP experiment. A JONSWAP wave spectrum
277 similar to the experiment is generated with shore normal direction and zero directional
278 spread. A no-slip condition is imposed on the lateral wall boundaries of the canal so that
279 transverse modes are precluded. The grid spacing is $dx=25$ cm with 10 vertical levels with
280 refinement at the bottom. This vertical refinement is important for wave-related transport,
281 while the results were poorly sensitive to horizontal resolution (a test with $dx=50$ cm and 1
282 m is presented below). The model time step is $dt = 25$ ms. The minimum depth is 1 mm on
283 the shore, the position of which varies with the swash oscillation, relying on a wetting-drying
284 scheme (Warner et al., 2013). For bed skin shear stress (setting sediment in motion), the
285 logarithmic law of the wall is used with grain roughness $z_{0s} \sim D50/12 = 0.02$ mm. To
286 account for increased momentum friction due to near-bed sediment suspension (e.g., Suarez
287 et al. 2014), we increase the roughness length for momentum to 0.2 mm.

288 The LIP-1B and LIP-1C experiments lasted 18 and 13 hours, respectively. In both
289 cases, the model was run for one hour with a morphological acceleration factor f_m equal to
290 18 and 13 respectively. Since the results were similar without acceleration, the acceleration
291 technique allowed us to perform many sensitivity tests.

292 *3.3. Turbulence model validation*

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

298 Figure 1 shows a comparison of the model with data, using our standard configuration.
299 The match with measured currents is very good throughout the complex morphology of the
300 beach. The waves start to break before reaching the sandbar, but the breaking is more

301 intense on the sandbar. The resulting velocity profile on the lee-side of the bar has a high
302 shear and the undertow has an intensity above 30 cm/s. The horizontal resolution test (25
303 cm, 50 cm and 1 m) shows a mean error of about 3 cm/s at all resolutions, close to the
304 measurement error of 2 cm/s. The results are thus consistent at all resolutions despite no
305 adjustment of any parameter. It confirms the validity of a RANS approach for estimating
306 the mixing length of breaking-induced turbulence.

307 Turbulent kinetic energy and eddy viscosity estimated by the k - ω model in the breaker
308 zone have the expected structure (Fig. 1; top) and magnitude ($\nu_t \sim 0.01h\sqrt{gh}$; Svendsen
309 1987; Cox et al. 1994). Interestingly, the transport terms in the closure equations tend
310 to reduce mixing at the break point by redistributing the turbulent energy, thus allowing
311 more intense shear to be maintained (not shown). The k - ϵ model works almost as well as
312 the k - ω model, with respect to mean current profiles, but the comparison is improved by
313 imposing a high value on the surface mixing length ($z_{0s}=0.2$ m), as in wave-averaged models
314 (Feddersen and Trowbridge, 2005). The k - ω model may thus be a better choice for surface
315 wave breaking, possibly due to a more accurate near wall treatment (Mayer and Madsen,
316 2000; Devolder et al., 2018; Larsen et al., 2020). Note, however, that this model produces a
317 greater amount of mixing in potential flow regions outside the surf zone (innershelf), mainly
318 due to the divergence part of the mean strain rate tensor (Mayer and Madsen, 2000).

319 3.4. Currents and suspended sediments

320 To analyse the model for different conditions, two experiments are selected, namely LIP-
321 1B and the LIP-1C cases (Table 1). Figure 2 shows a model-data comparison of wave height
322 H_{rms} for the two experiments. The data is reproduced accurately. High waves in LIP-1B
323 ($H_s = 1.4$ m, $T_p = 5$ s) caused beach erosion while, in LIP-1C, moderate waves ($H_s = 0.6$
324 m, $T_p = 8$ s) caused an accretive beach.

325 We are first interested in the erosion phase where the sandbar moves offshore under
326 the effect of high waves. The LIP dataset is used for comparison purposes, containing
327 velocity and concentration profiles as well as bed level evolution (see next section). Figure
328 3 shows a comparison of wave-mean velocity and concentration profiles for LIP-1B. As for

329 velocity profiles presented in the validation section, sediment concentrations also show a very
330 good match between model and data. The high concentrations over the sandbar crest and
331 trough are triggered by a large bed shear stress associated with a combination of wave-mean
332 undertow, wave oscillatory flow and turbulent velocity.

333 Note that waves and associated orbital velocities are greatest off the sandbar. However,
334 the maximum suspended sand concentration is not found offshore but coincides instead with
335 a maximum in turbulence intensity and undertow speed inshore of the bar crest. This is in
336 contrast to parametrizations more suited to shelf dynamics, where sediment erosion responds
337 to the addition of bed shear stresses by currents and waves. (e.g., Soulsby 1995).

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

342 *3.5. Offshore and onshore sandbar migration*

343 Figure 5 (top) shows the bed evolution 18 hours after the start of experiment 1B, with
344 the corresponding section of wave-mean cross-shore currents. Some small discrepancies are
345 visible but the offshore sandbar migration is correctly reproduced in terms of distance of
346 migration and height of the bar. The maximum undertow magnitude (-0.3 m/s) follows the
347 lee-side slope of the bar and appears as the driver of its migration. A secondary undertow
348 maximum is visible on the terrace of the inner surfzone and appears to coincide with an
349 offshore spreading of the terrace. In the outer surfzone (off of the main sandbar), a second
350 smaller bar forms both in the model and in the measurements during LIP-1B (at a position
351 of about 70 m), corresponding to a weakened undertow.

352 The bottom panel of Figure 5 shows the result of the accretive phase LIP-1C when
353 waves have lower amplitude, with an opposite migration directed towards the shore. In this
354 case, the flow structure is very similar but weaker and the bar seems to migrate against the
355 undertow. The latter is still able to bring sediment to the leeward slope of the bar, but
356 the offshore slope and crest are moving inshore, creating an asymmetry in the bar shape.

357 Sediment transport processes are analyzed in more detail in the next section.

358 4. Sediment transport mechanisms

359 4.1. Sediment budget

360 The wave-mean sediment transport term $-\frac{1}{\rho_0} \overline{\nabla \cdot (\rho \vec{v} C)}$ — ρ_0 is a mean density, so that the
361 transport term is in units of $\text{kg}/\text{m}^3/\text{s}$ — is presented in Figure 6 for the time halfway between
362 the erosive (top) or accretive (bottom) phase (averaged over a period of about 10 min). This
363 confirms for LIP-1B that during high waves and strong undertow, a large amount of sand
364 is removed from the lee-side slope and trough area — after being first suspended and stirred
365 upward — and then transported to the offshore slope, where it settles. Part of this process is
366 attenuated by surface transport, which brings back some sediment to the bar trough region.
367 For LIP-1C, however, sand is removed from both the trough of the bar and the offshore
368 slope (including the crest), and it converges to the inner slope. This transport may qualify
369 as bedload or sheet-flow transport as it occurs primarily within the wave boundary layer of
370 about 10 cm.

371 For further analysis, Figure 7 presents the depth-integrated suspended sediment budget,
372 again for the erosive (top) or accretive (bottom) phase. The different terms of Eq. 13 are
373 computed online, averaged in time and integrated along the vertical axis. The budget comes
374 down to three main tendency terms: erosion-settling residual, advection, and rate of change
375 (vertical mixing cancels out when integrated). Suspended sediments are close to equilibrium
376 since the rate of change is generally much smaller than the other two terms, which tend to be
377 opposite. This analysis confirms the difference between LIP-1B and LIP-1C, with transport
378 limited to the lee side of the bar in LIP-1C. In LIP-1B, there is a dominant erosion process
379 in the bar trough and offshore transport and deposition beyond the bar crest. In LIP-1C,
380 the process is similar but transport and deposition are confined to the inshore side of the
381 bar crest, while a secondary zone of net erosion appears on the offshore side (this secondary
382 erosion also appears in LIP-1B but much further offshore).

383 In the next section, we distinguish between wave and mean current transport in order

384 to identify and quantify the role of asymmetric waves for bar migration, particularly in the
385 accretion phase of LIP-1C.

386 *4.2. Wave-related and current-related fluxes*

387 For each experiment (erosive or accretive), we construct a wave composite by superposing
388 instantaneous data over many wave cycles (about 50). To that end, the duration of all wave
389 cycles must be fitted to the peak period, so that data averaging can be made over the
390 different phases of a generic cycle. This composite is made at three locations around the
391 sand bar: offshore slope, crest and trough (corresponding to positions: $x=131, 139$ and 148
392 m in LIP-1B; and $x=129, 136$ and 144 m in LIP-1C).

393 Starting with the more complex accretive case, Figure 8 (left panels) shows the currents
394 and sand concentrations at the first model level (1-2 cm above bed) for the composited
395 LIP-1C wave cycle. Here, the mean values are subtracted to retain the oscillatory flow
396 and associated perturbed concentrations ($u' = u - \bar{u}$ and $C' = C - \bar{C}$). In addition, the
397 figure shows the undertow u_b and effective turbulent velocity u_t entering the bed shear
398 stress formulation. The currents have a typical skewed/asymmetric structure across the
399 wave cycle. They are largely skewed over the offshore slope and become less skewed but
400 more asymmetric as the wave progresses toward the shore. The largest sand concentrations
401 are over the bar crest where it peaks about half a second after the peak velocity. The lag is
402 similar for skewed velocity over the offshore slope, although the peak there is much weaker.
403 The main reason is that the effective turbulent velocity is more than twice as large over the
404 crest and more significantly contributes to the bed skin stress. The sand concentration cycle
405 is different in the trough region, where the undertow becomes comparable in magnitude
406 to the orbital velocities. The total onshore-directed flow is reduced during the crest half-
407 cycle while the offshore-directed flow is increased during the trough half-cycle. As a result,
408 positive resuspension anomalies are present during both phases of negative and positive wave
409 velocities.

410 The right panels of Figure 8 show the profiles of suspended sand flux. The averaged
411 flux over the composited wave cycle is separated into wave and mean components (Reynolds

412 decomposition): $\overline{uC} = \overline{u'C'} + \overline{u}\overline{C}$. The undertow being weak over the offshore slope and crest
413 (Fig. 5, bottom panel), the total flux there is dominated by the residual wave flux, which
414 is onshore (at least in the wave boundary layer, but the flux is slightly negative above).
415 The maximum flux over the crest appears to coincide with a combination of high sand
416 concentration (1 g/L) and positive velocities (1 m/s) during the wave crest period (middle-
417 right panel of Fig. 8). Over the offshore slope, the velocity is more skewed than asymmetric
418 and the concentration is limited by a weaker bed stress (due to weaker turbulence). The
419 phase lag between mobilization and transport of sediment (Dohmen-Janssen et al., 2002)
420 does not vary between the two positions (as expected from fast settling that gives a weak
421 phase lag parameter) and therefore cannot explain the differences in sand flux. In the bar
422 trough, the wave-related flux remains positive but its magnitude is much reduced. The
423 reason is the second peak of sand resuspension during the wave trough phase (negative half-
424 cycle), due to a strong undertow. An even stronger undertow could eventually reverse the
425 sign of this flux (Scott et al., 2009).

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

430 In our simulations, wave-breaking turbulence contributes little to the onshore flux asym-
431 metry, unlike in Ting and Kirby (1994). This is because, although more turbulent energy
432 is generated at the surface during the wave crest period, a significant portion reaches the
433 bottom during the trough of the next wave (Figure 1) and the bottom turbulent velocity in
434 Figure 8 exhibits only small variations during the wave cycle. This tends to argue against a
435 strong asymmetric effect of breaking turbulence, although we are aware that the turbulence
436 model may misrepresent the role of plunging waves (i.e. the timing and efficiency with which
437 they mobilize sediment). However, observations suggest that wave-breaking turbulence is
438 sufficiently intermittent to lose some of its correlation with wave phase (Ruessink et al.,
439 2011). Improvements could be made in this area.

440 Looking now at the LIP-1B erosive experiment (Fig. 9), we see a different dynamic,

441 which generalizes the mechanism described earlier for the trough region. Since the undertow
 442 is present at all locations, the total flux is also negative at all locations. The wave-related flux
 443 tends to be weaker than in LIP-1C because the negative undertow decreases resuspension
 444 in the positive orbital velocity phase (wave crest) and increases it in the negative phase.
 445 However, over the crest, a significant wave-related flux remains. Note that waves over the
 446 sandbar in LIP-1B (and their asymmetry; see next section and Fig. 11) are actually weaker
 447 than in the accretive phase, despite larger offshore waves. The reason is that, in this case,
 448 the waves break off the bar, where they lose much of their energy. However, the undertow
 449 extends across the sandbar and overwhelms the wave effect everywhere. Note also the strong
 450 sediment flux convergence on the offshore side of the bar despite a consistently negative sand
 451 flux. In other words, there need not be a confluence of onshore and offshore fluxes for flux
 452 convergence to occur.

453 *4.3. Wave asymmetry*

454 Figure 10 presents the profiles of current skewness and asymmetry for the three sandbar
 455 locations. Normalized values (right panel) are computed as $\langle u'^3 \rangle / \langle u'^2 \rangle^{1.5}$ for skewness and
 456 $\langle \mathcal{H}(u')^3 \rangle / \langle u'^2 \rangle^{1.5}$ for asymmetry. We also present the dimensional values (left panels), com-
 457 puted without normalization by the cube of standard deviations. It confirms the relatively
 458 high skewness and low asymmetry on the outer slope, relative to the bar crest. The dimen-
 459 sional values reveal a much higher absolute asymmetry on the bar crest, compared to the
 460 surrounding areas. This result is hidden in the dimensionless calculation. Bottom skewness
 461 on the slope is larger than asymmetry on the crest, but with less effect on the sand flux,
 462 as seen earlier. This would confirm the efficiency of asymmetry versus skewness, but in our
 463 case this is coincidental: the difference in sand flux is related to the varying intensity of tur-
 464 bulence across the bar, rather than a varying phase lag between mobilization and transport
 465 — sometimes associated with the transformation from skewness to asymmetry (Dohmen-
 466 Janssen et al., 2002). Our interpretation of the respective role of asymmetry and skewness
 467 thus involves a spatial correlation between wave asymmetry and breaking turbulence. Fi-
 468 nally, the trough region is characterized by low absolute skewness and asymmetry near the

469 bed and the sand flux there is always driven by the undertow.

470 The dimensional skewness and asymmetry also reveal large vertical variations from sur-
471 face to bottom. Wave skewness/asymmetry is intensified at the surface because wave ve-
472 locities are greater, but the intensification is much larger for broken waves inshore of the
473 bar. In this case, the wave velocities are also much stronger at the surface (1.2 m/s, as
474 for shoaling waves) than at the bottom. Another noticeable point about these profiles is
475 a slight decrease of dimensional skewness and asymmetry in the wave boundary layer, and
476 a decrease of dimensionless asymmetry in favor of skewness, consistent with observations
477 (Suarez et al., 2014). However, here, this transformation is weak and does not seem to play
478 a significant role.

479 4.4. Bed shear stress

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

$$\tau_w = 0.5 f_w (u_w + u_t)^2 \quad (18)$$

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

$$f_w = 1.4 r_f \left(\frac{u_w}{\sigma z_{0b}} \right)^{-0.52} \quad (19)$$

486 u_w is the maximum bottom orbital velocity, determined from wave-mean wave character-
487 istics. u_w accounts for skewness as a correction to the linear theory (Isobe and Horikawa,
488 1982; Abreu et al., 2010; Malarkey and Davies, 2012; Nam et al., 2020): $u_w = r_s u_w^{lin}$, where
489 r_s is a skewness factor ($r_s = 1$ for linear waves). r_s can be obtained using empirical formulas
490 involving a logarithmic relation with the Ursell number $Ur = H_s \lambda^2 / h^3$, a measure of wave
491 nonlinearity (Nam et al., 2020). Here, we use a slightly modified formula compared to Nam

492 et al. (2020): $r_s = 0.051 \log Ur + 0.84$. The turbulence stirring velocity u_t is computed as
493 for the wave-resolving simulation (Eq. 16) but with wave-averaged k_b .

494 Figure 12 shows our attempt of reconstructing the simulated bed shear stress with
495 parametrizations. At offshore positions, where waves are nearly linear, a value of $r_f = 0.4$ is
496 required to fit the simulated bed shear stress reconstructed from Soulsby (1995). Then, in
497 the nearshore region from about $x = 40$ m, the parametrized bed shear stress is underesti-
498 mated with linear waves and no turbulence ($r_s = 1$ and $u_t = 0$). Adding skewness correction,
499 the bed shear stress is recovered more accurately around the sandbar, particularly on the
500 seaward side. The most striking feature is that turbulence stirring contributes most of the
501 simulated stress on the sandbar (in the roller zone, downstream of the sandbar, the turbu-
502 lence effect seems underestimated). This analysis thus confirms that consideration of wave
503 asymmetry may be important but that turbulent mixing by plunging breakers is essential
504 to mobilize significant sediment loads on the sandbar, allowing its realistic migration.

505 5. Discussion and conclusion

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

516 The Analysis of the transport mechanisms shows the importance of the undertow distri-
517 bution around the sandbar. Its convergence and divergence patterns appear to have more
518 impact on submerged beach morphology than the confluence of wave- and current-related

519 fluxes. In the erosion phase, in particular, the waves begin to break well ahead of the sand-
520 bar and a strong undertow causes the mobilized sediments to move in a near morphological
521 translation. In this case, the wave-related transport is weak because sand mobilization,
522 dominated by the undertow, is distributed almost uniformly between the two phases of the
523 wave cycle. Much of the sediment is carried in suspension as the intensity of turbulence
524 within the water column is high. It is interesting to note that there is a moderate surface
525 flux of sediment towards the shore, carried by the breaker-induced onshore flow.

526 Under moderate wave action, the sandbar becomes a breaking point and the undertow
527 is confined to the inshore area of the bar crest with a maximum on the inner slope. Because
528 the bed shear stress and associated mixing is weaker in this case, the sand concentration is
529 almost confined to a sheet flow layer. In comparison to more energetic wave conditions, the
530 sandbar migrates onshore at a slower rate and with an asymmetric shape. This is the result
531 of a combination of two mechanisms: the deposition of sand carried by the undertow on the
532 inshore side and, to a lesser extent, onshore fluxes due to asymmetric waves on the offshore
533 side. Note that these mechanisms are not opposed here, as is often suggested (Roelvink and
534 Stive, 1989; Grasso et al., 2011), but complementary.

535 The composite analysis of wave skewness/asymmetry reveals that the onshore sediment
536 flux is strongest over the bar crest where asymmetry produces a steep front with strong cur-
537 rents during the onshore half-cycle. Associated with these currents, the sand is resuspended
538 within only half a second with much contribution by breaking turbulence. Over the offshore
539 slope, where skewness is greater than asymmetry, a similar process is at work but is weaker
540 due to weaker resuspension, and not because of a greater phase lag between mobilization
541 and transport.

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

548 The largest uncertainty in the model comes from the estimation of bed roughness and
549 bed shear stress due wave-breaking turbulence. Clearly, sand mobilization (and transport)
550 is unrealistic without turbulence, as already noted by van der Zanden et al. (2017) and Lim
551 et al. (2020) for the same LIP experiments. Lim et al. (2020) suggested that a large local
552 increase of suspended sand concentration in the LIP experiment could only be attributed
553 to breaking wave turbulence and more precisely to plunging breaker vortices invading the
554 wave bottom boundary layer. We followed the idea of Reniers et al. 2004 of adding a
555 breaking-induced turbulent bottom current (from the turbulence model) to the bed shear
556 stress formulation. Using an enhancement factor and positioning function for this turbulent
557 motion leads to realistic results. The positioning function, however, lacks a generic type
558 mechanism and efforts should be made in this direction. Among the alternatives we tried,
559 we found promising results using power laws to locally enhance the effect of subgrid-scale
560 turbulent energy. We also obtained similar results by setting the bottom stress as a function
561 of the undertow (the undertow seems to be a good proxy due to its correlation with breaker
562 dissipation; Faria et al. 2000). Additional work will likely be required, including explicit
563 consideration of plunging breakers.

564 With these caveats in mind, our results nonetheless demonstrate the reliability of compu-
565 tationally efficient 3D wave-resolving models such as CROCO – intermediate between two-
566 phase DNS and wave-averaged approaches – for addressing nearshore hydro-morphodynamic
567 problems. They point out the deficiencies of the depth-averaged approach, with its hypersen-
568 sitivity to bottom friction and rough estimation of the undertow, when its cross-shore distri-
569 bution seems so important. Our wave resolution model can also help improve parametriza-
570 tions in 3D wave-averaged models, in particular wave-related transport through the concept
571 of wave half-cycles.

572 An interesting new result also comes from the morphological acceleration method, which
573 is used here for the first time with a wave-resolving model. We demonstrated its applicability
574 and were even able to use a factor as large as 64, because the timescale of morphodynamics
575 (here bar migration) is slow compared to that of wave dynamics. This method is particularly
576 interesting in terms of computational cost because it allows us to consider long period

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

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587 at www.croco-ocean.org. Both laboratory and modeling data are available upon request.

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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

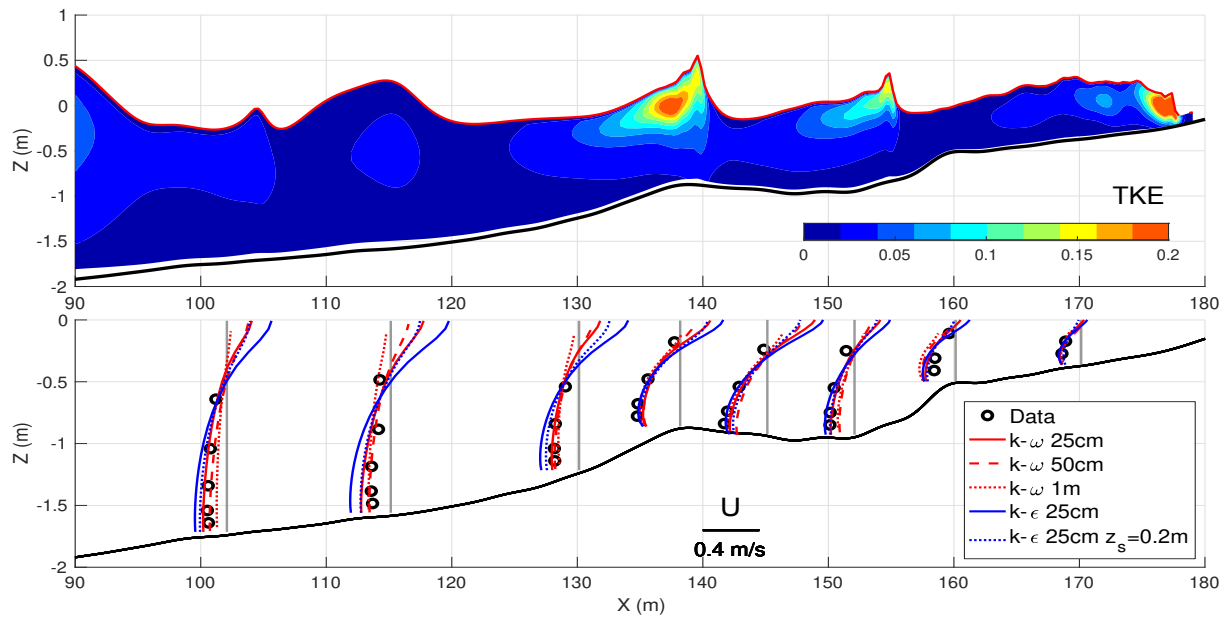


Figure 1: Model comparison with the large-scale LIP11-1B Flume experiment. Top) Snapshot of wave height and turbulent kinetic energy k [m^2/s^2] from the reference model simulation (25 cm resolution; $k-\omega$ turbulence model). Bottom) Comparison of simulated and measured cross-shore current profiles: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

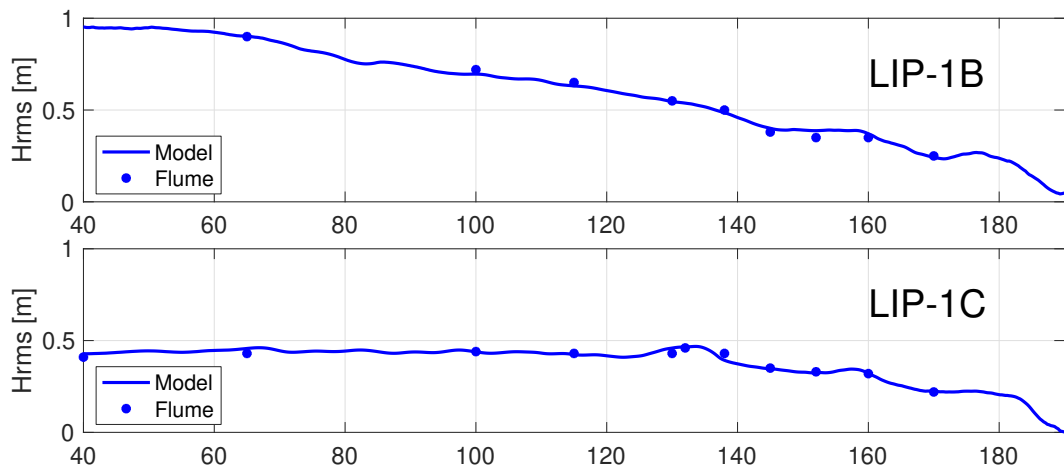


Figure 2: Model-data comparison of H_{rms} in LIP Flume experiment (top: 1B; bottom: 1C).

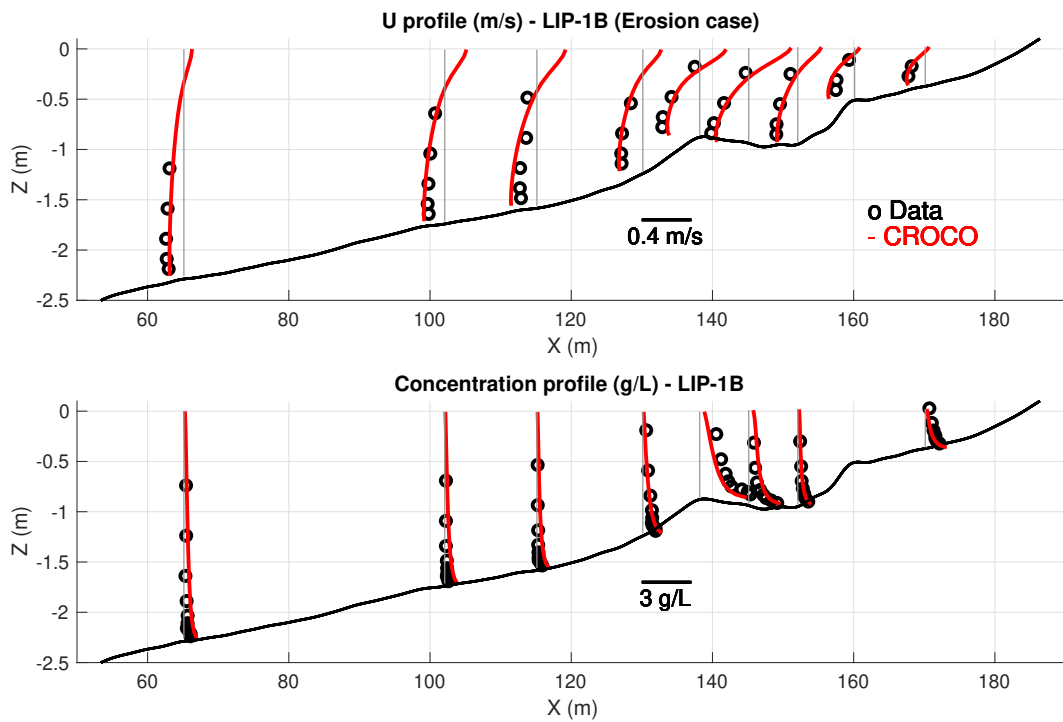


Figure 3: Model-data comparison of wave-mean cross-shore velocity and sand concentration profiles in the erosive LIP11-1B Flume experiment.

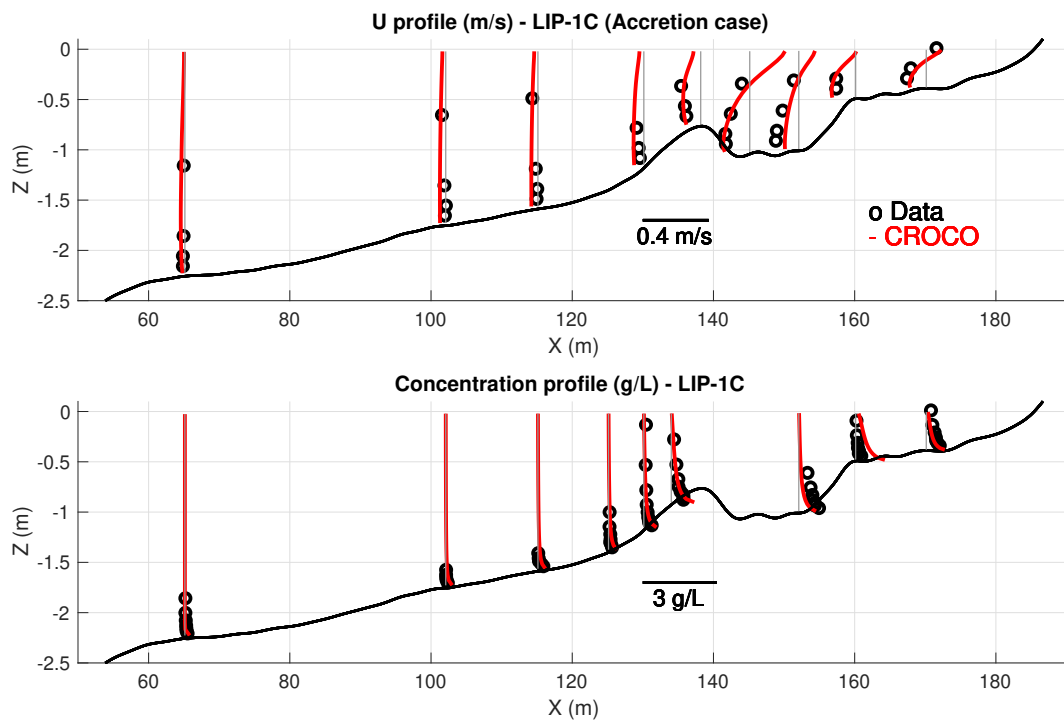


Figure 4: Model-data comparison of wave-mean cross-shore velocity and sand concentration profiles in the accretive LIP11-1C Flume experiment.

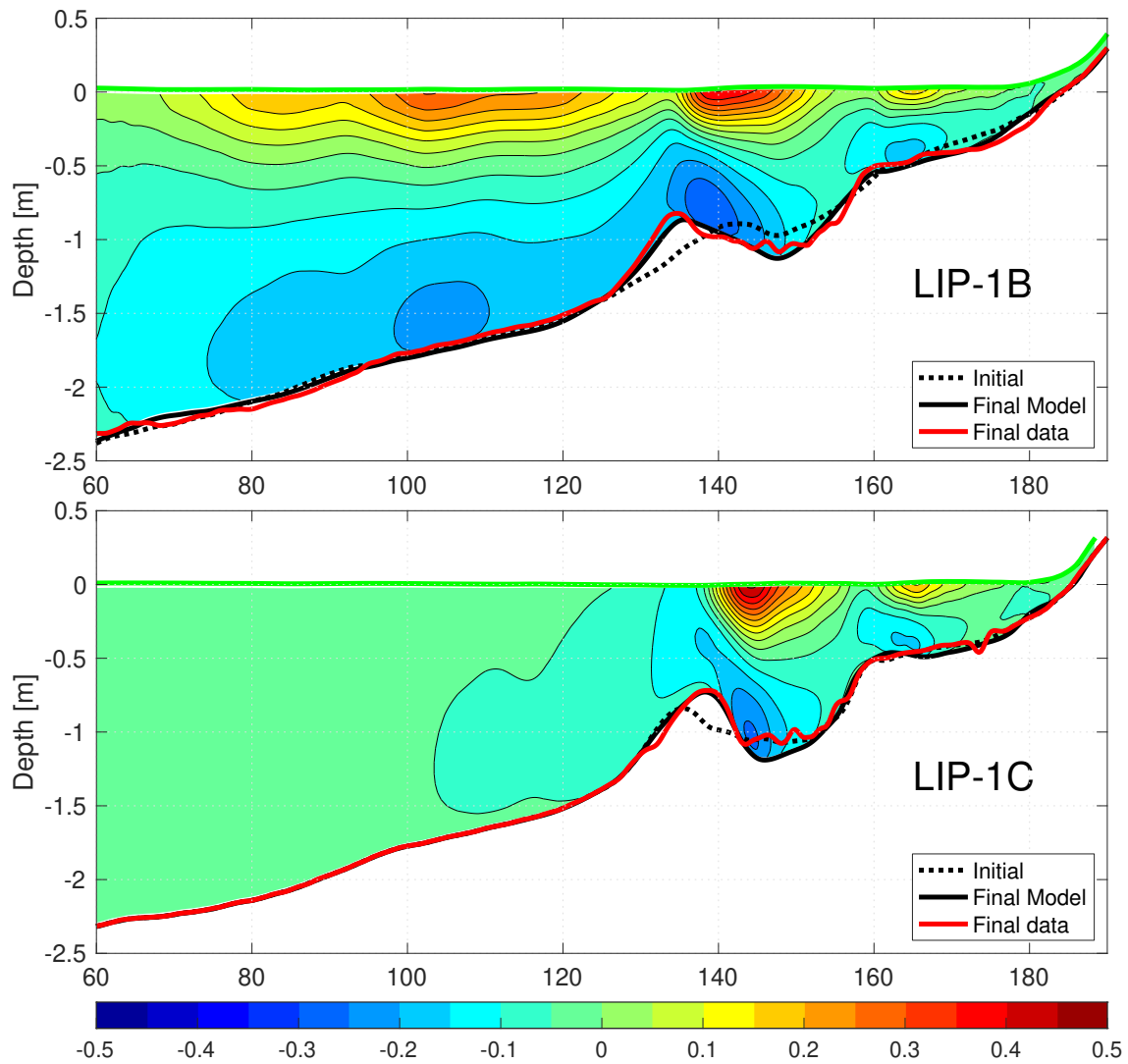


Figure 5: Model comparison of morphological changes with LIP measurements during the erosion phase (LIP-1B) and accretion phase (LIP-1C). Wave-mean cross-shore velocities are also shown in filled contours [m/s].

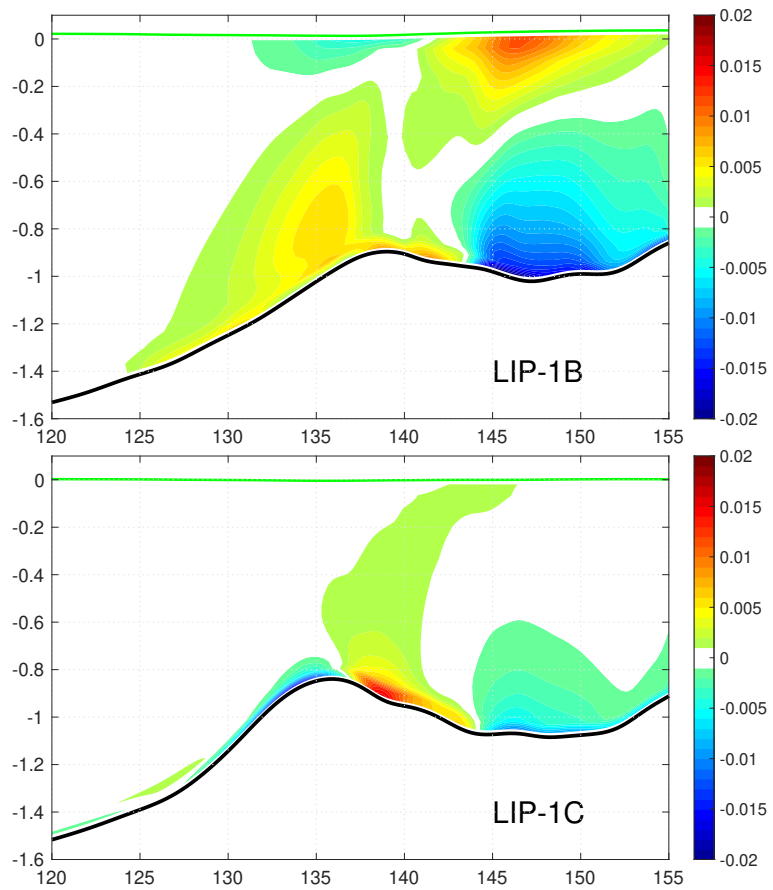


Figure 6: Wave-mean sand advection [$\text{kg/m}^3/\text{s}$] halfway through the erosion phase (LIP-1B) and accretion phase (LIP-1C).

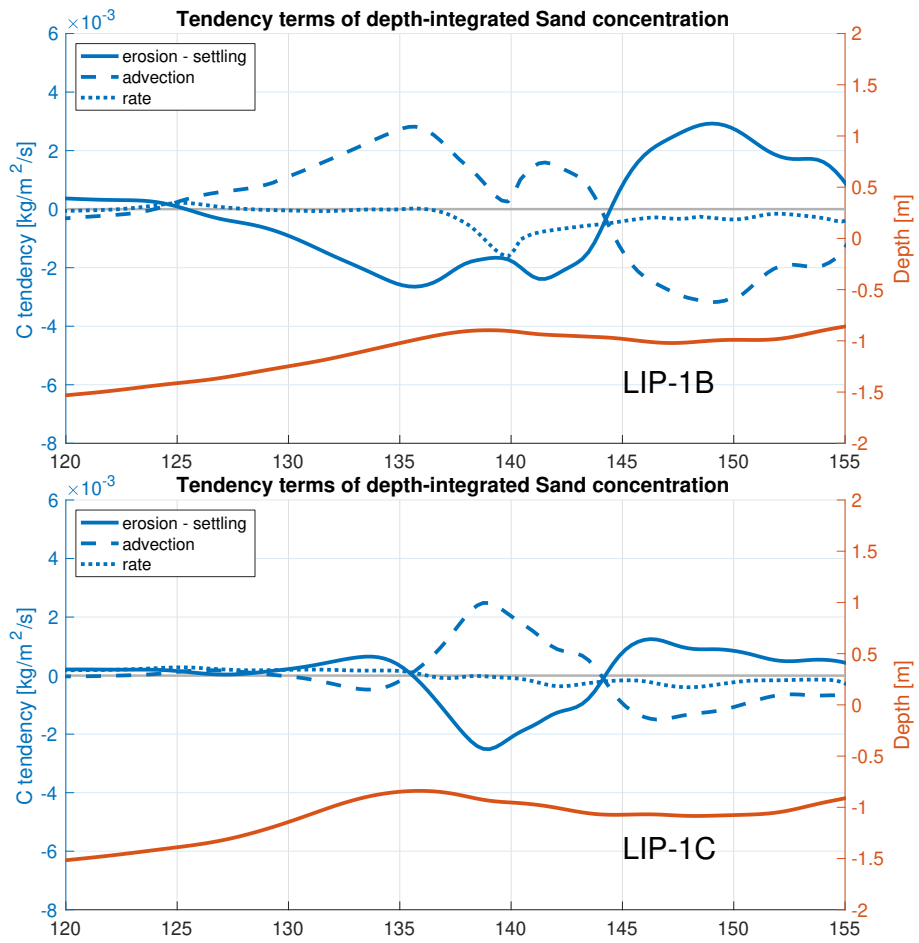


Figure 7: Depth-integrated budget of sand concentration halfway through the erosion phase (LIP-1B) and accretion phase (LIP-1C).

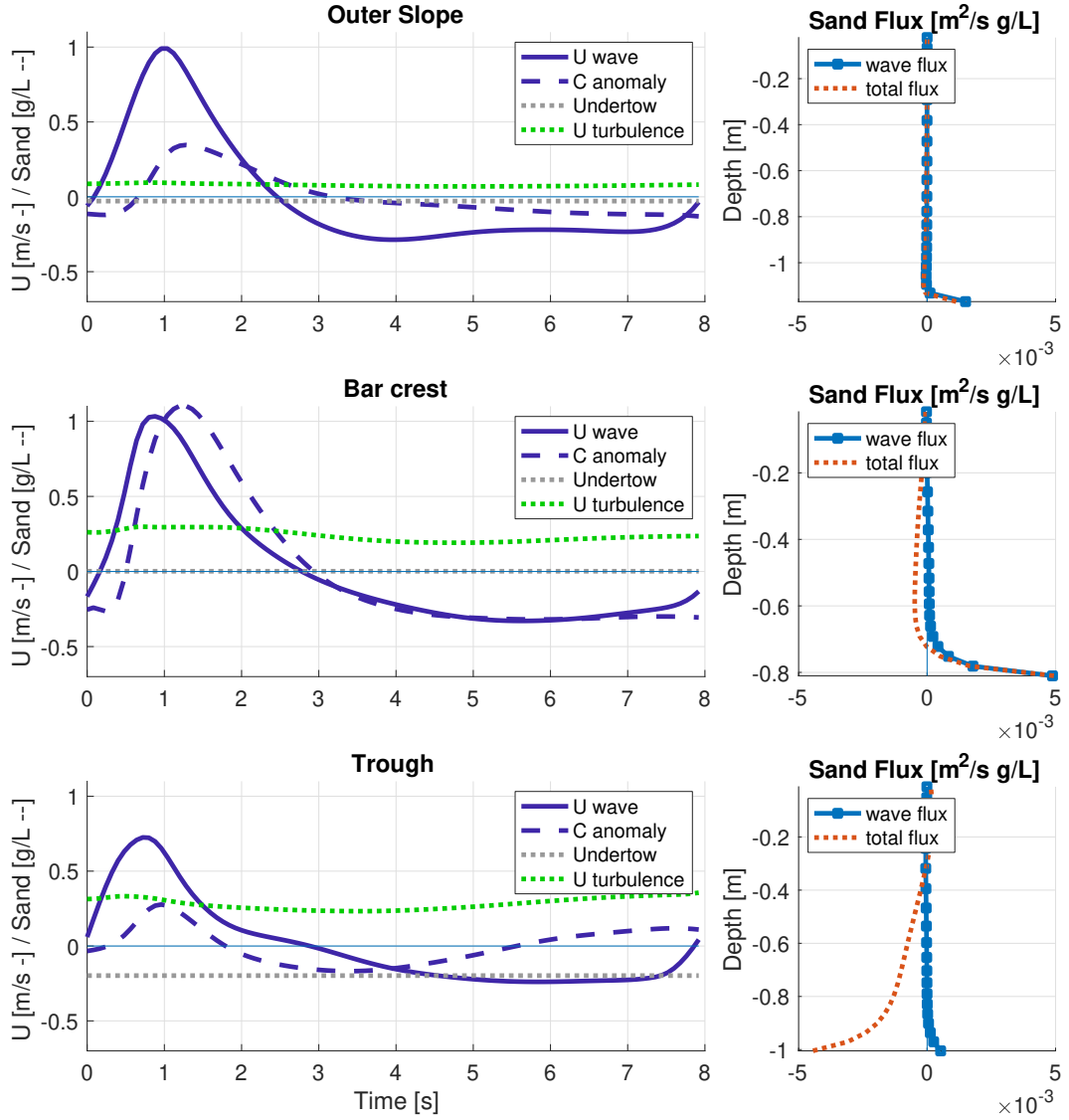


Figure 8: Left panels: composite wave cycle of current and sand concentration anomaly ($u' = u - \bar{u}$ and $C' = C - \bar{C}$) at three positions over the sandbar during the accretive LIP-1C Flume experiment. The wave-mean undertow u_b and mean effective turbulent velocity u_t at the bottom level z_b are also represented. Right panels: profiles of suspended sand flux (total flux: $\overline{u'C'}$; and wave flux: $\overline{u'C'}$).

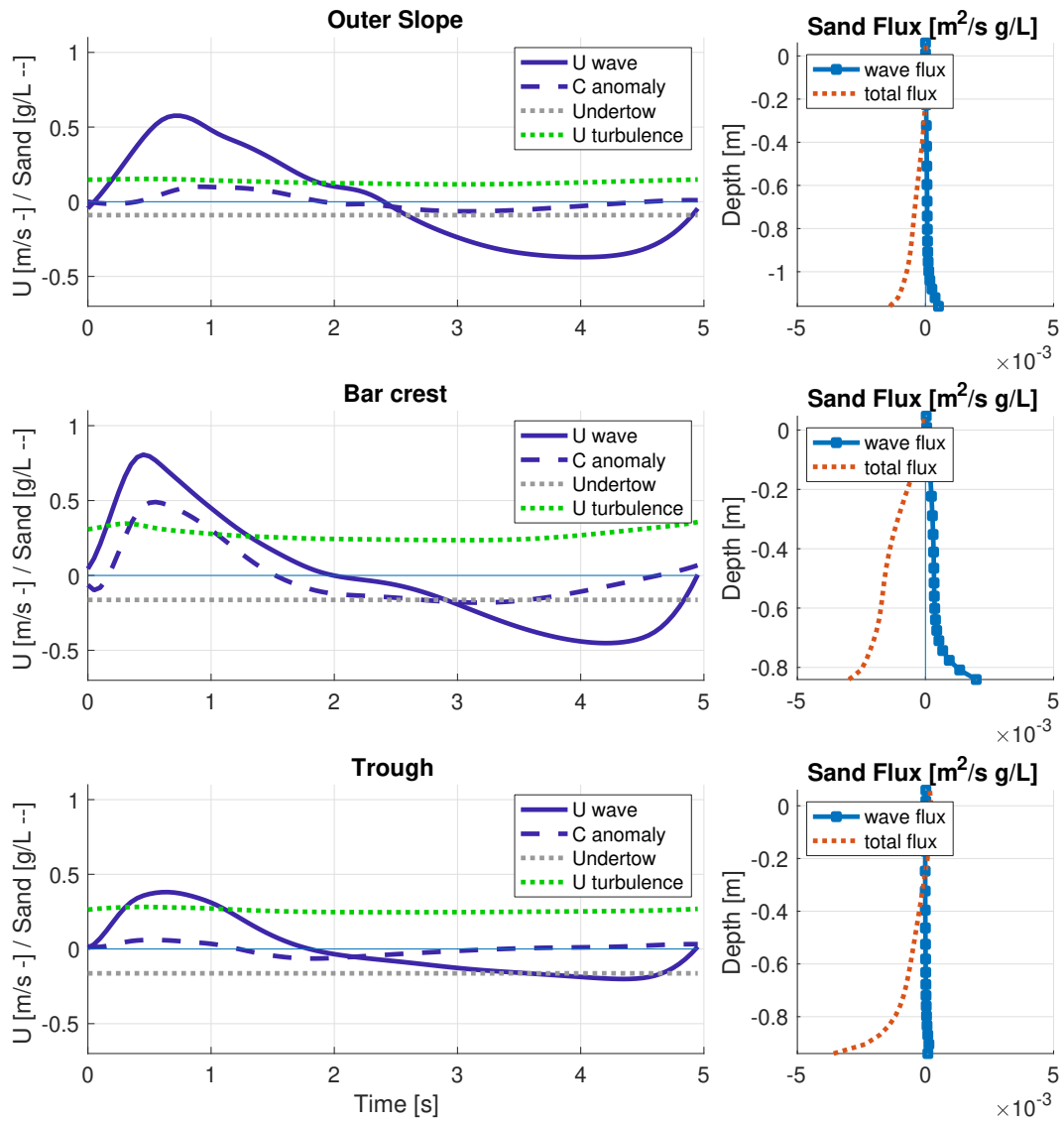


Figure 9: Same as Figure 8 but for the LIP-1B Flume experiment.

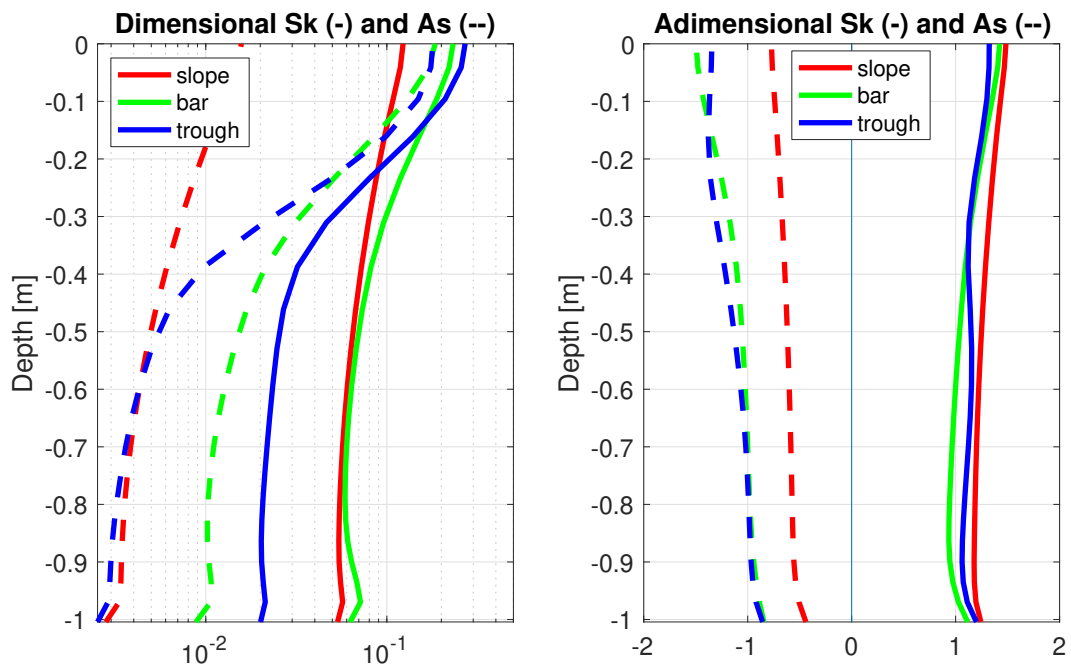


Figure 10: Composite analysis of wave asymmetry and skewness profiles (absolute and normalized) at three locations around the sandbar, during the accretive LIP-1C Flume experiment. Asymmetry is represented by dashed lines, skewness by solid lines.

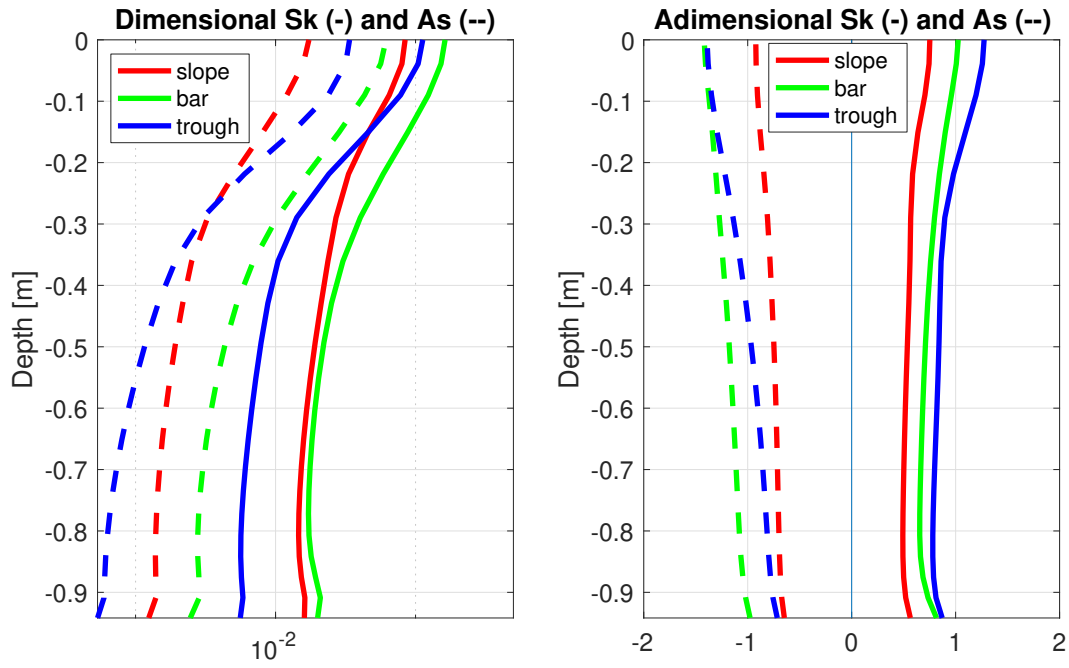


Figure 11: Composite analysis of wave asymmetry and skewness profiles (dimensional and dimensionless) at three locations around the sandbar, during the erosive LIP-1B Flume experiment.

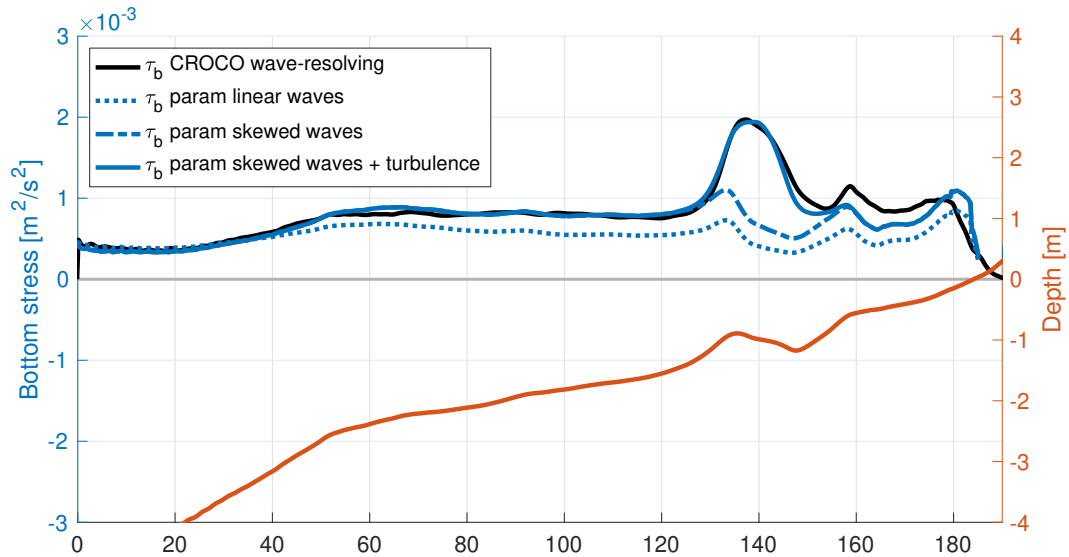


Figure 12: Bed shear stress profiles from the wave-resolving model and reconstructed from empirical formulations using wave-averaged values and linear wave theory with or without skewness correction.