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► **To cite this version:**

Patrick Marchesiello, Julien Chauchat, Hassan Shafiei, Rafael Almar, Rachid Benshila, et al.. 3D wave-resolving simulation of sandbar migration. *Ocean Modelling*, 2022, 180, pp.102127. 10.1016/j.ocemod.2022.102127 . hal-03345439v2

HAL Id: hal-03345439

<https://hal.science/hal-03345439v2>

Submitted on 10 Oct 2022

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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 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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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) 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} ; η is the free surface; P
 99 the total pressure; ρ 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 ρ 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 - ϵ or k - ω model, solving
142 the closure equations for turbulent kinetic energy k and dissipation ϵ 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_μ 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 - ϵ and Wilcox (1988) for k - ω . 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 - ϵ 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 - ω 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 ω_i , from a given statistical
 167 distribution, e.g., JONSWAP (Sec. 3.2); ω_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; ϕ_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 (Durrant, 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 δz_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 τ_{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 τ_{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); τ_{cr} to 0.11 N m^{-2} (Soulsby and
 202 Whitehouse, 1997). p is the sediment porosity (0.41). τ_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 α_{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, ϕ 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 ϕ 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 ρ_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 η and bottom currents u_b . A
359 comparison was made of continuous profiles of wave statistics for η and u_b across the beach.
360 The model statistics closely resembled the measurement data (in structure and magnitude
361 for both η 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 ~ 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 (~ 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 $-\frac{1}{\rho_0} \overline{\nabla \cdot (\rho \vec{v} C)}$ — ρ_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 (τ_w) and current-induced (τ_{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.

628 **Acknowledgement**

629 This research has received support from a consortium of French research agencies, as part
 630 of CROCO's development project (Insu GdR n°2014 named CROCO) and from the French
 631 Naval Hydrographic and Oceanographic Service (SHOM DGA-Protevs II and MEPELS
 632 projects). It was granted access to the HPC resources of CALMIP supercomputing center
 633 under allocation P19069. We thank Dano Roelvink for sharing the LIP data. Apart from
 634 these, all data were acquired by the authors and the CROCO source code is freely available
 635 at 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 ρ 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.

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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 |

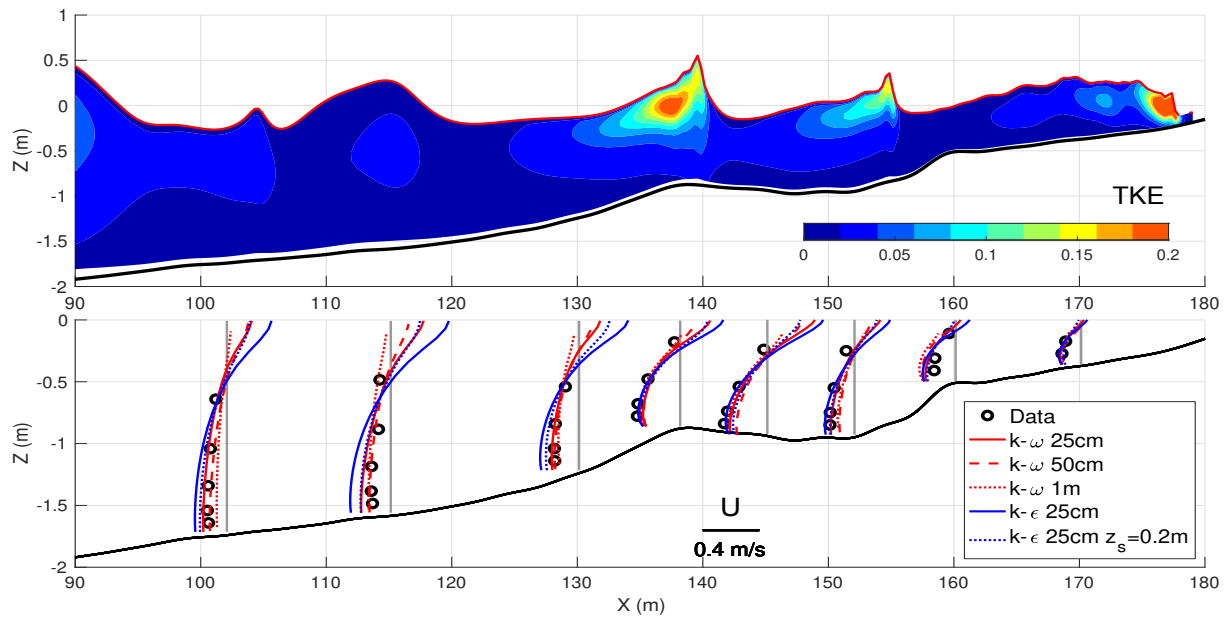


Figure 1: Model comparison with the large-scale LIP11-1B Flume experiment. Top) Snapshot of wave height and turbulent kinetic energy k [m^2/s^2] from the reference model simulation (25 cm resolution; k - ω turbulence model). Bottom) Comparison of simulated and measured cross-shore current profiles: sensitivity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

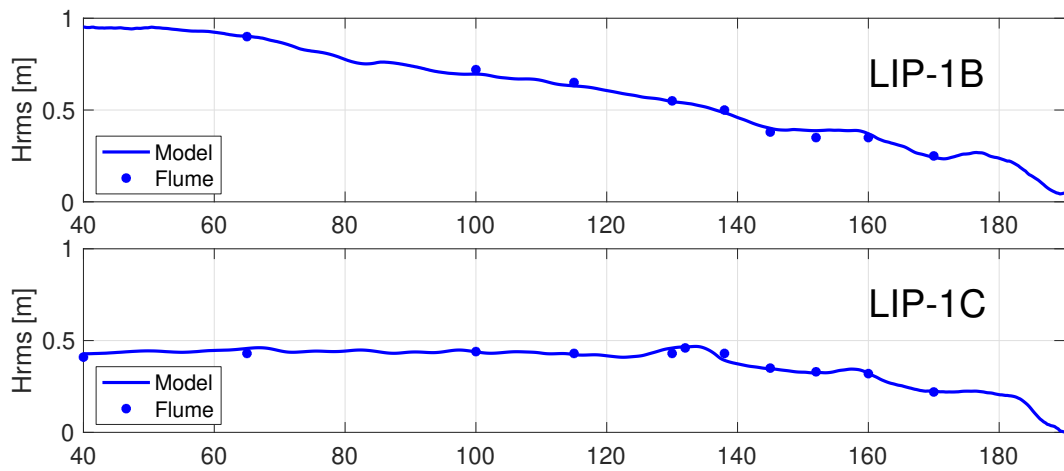


Figure 2: Model-data comparison of H_{rms} in LIP Flume experiment (top: 1B; bottom: 1C).

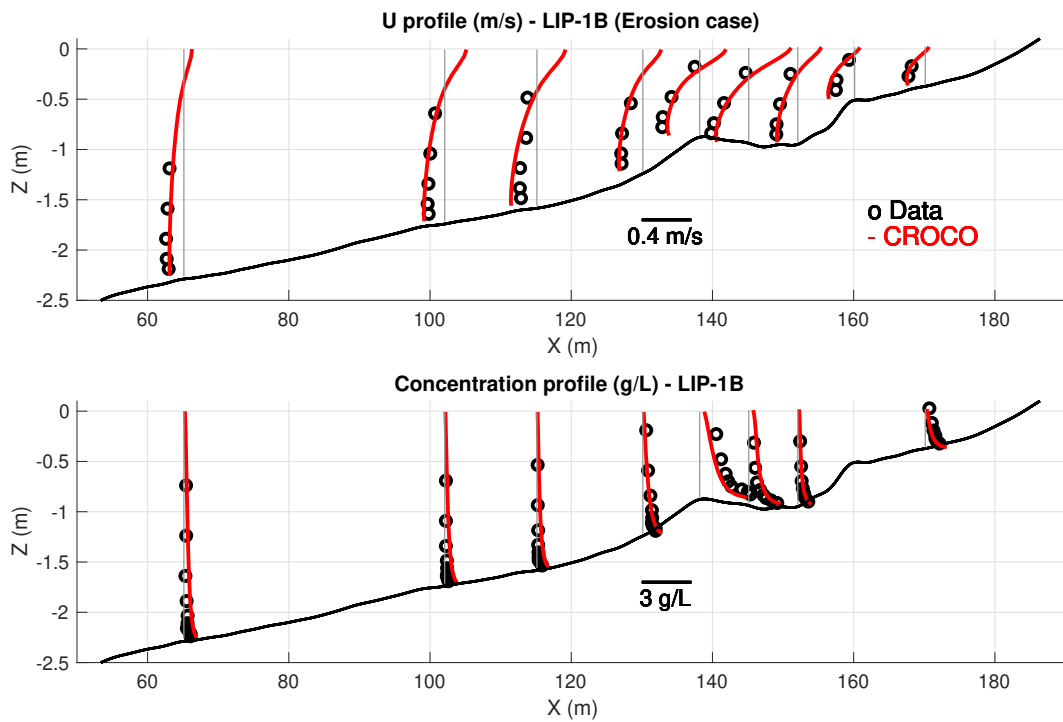


Figure 3: Model-data comparison of wave-mean cross-shore velocity and sand concentration profiles in the erosive LIP11-1B Flume experiment.

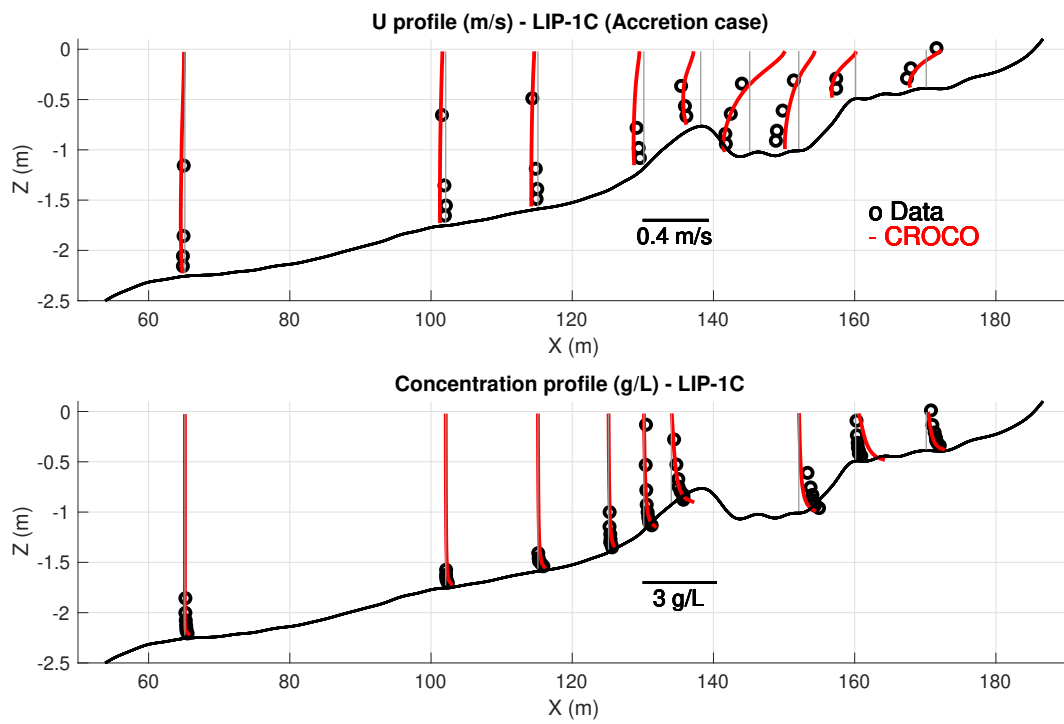


Figure 4: Model-data comparison of wave-mean cross-shore velocity and sand concentration profiles in the accretive LIP11-1C Flume experiment.

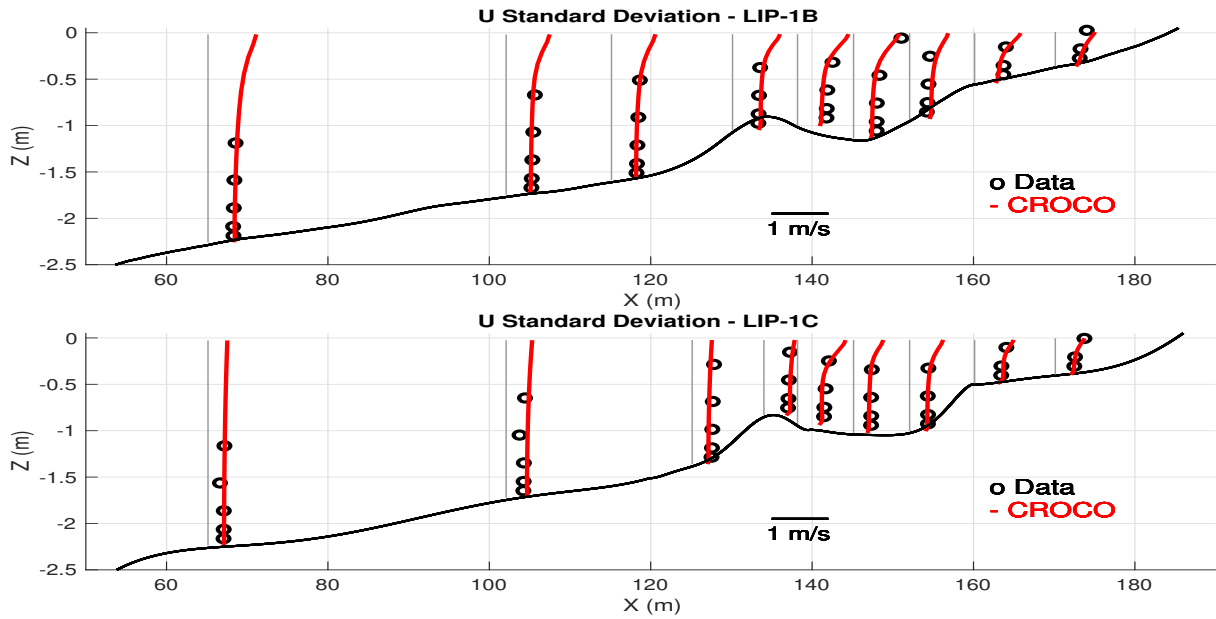


Figure 5: Model-data comparison of velocity standard deviation profiles in the LIP11-1B and LIP11-1C experiments.

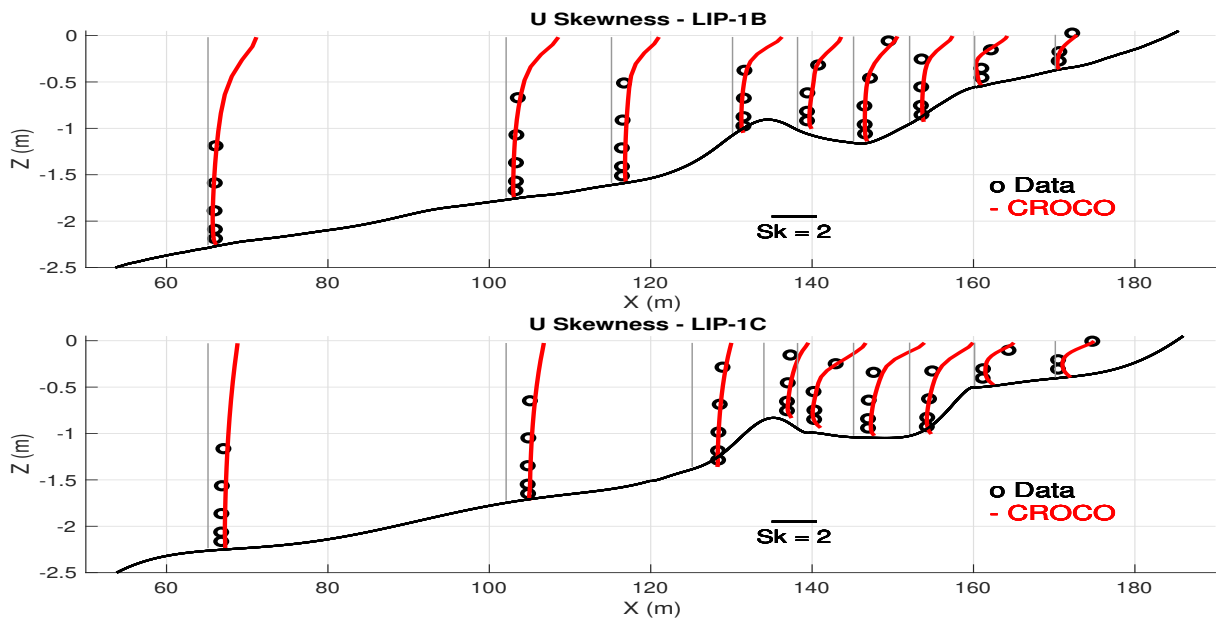


Figure 6: Model-data comparison of velocity skewness profiles in the LIP11-1B and LIP11-1C experiments.

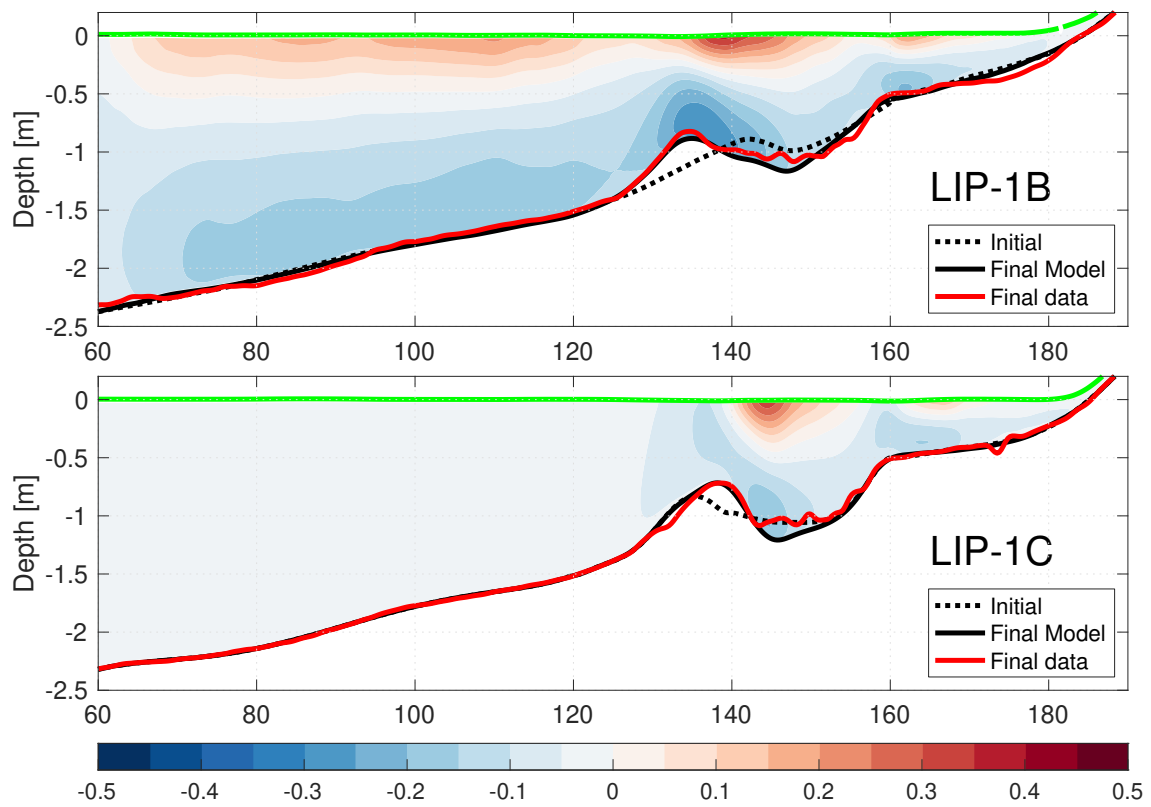


Figure 7: Model comparison of morphological changes with LIP measurements during the erosion phase (LIP-1B) and accretion phase (LIP-1C). Wave-mean cross-shore velocities are also shown in filled contours [m/s].

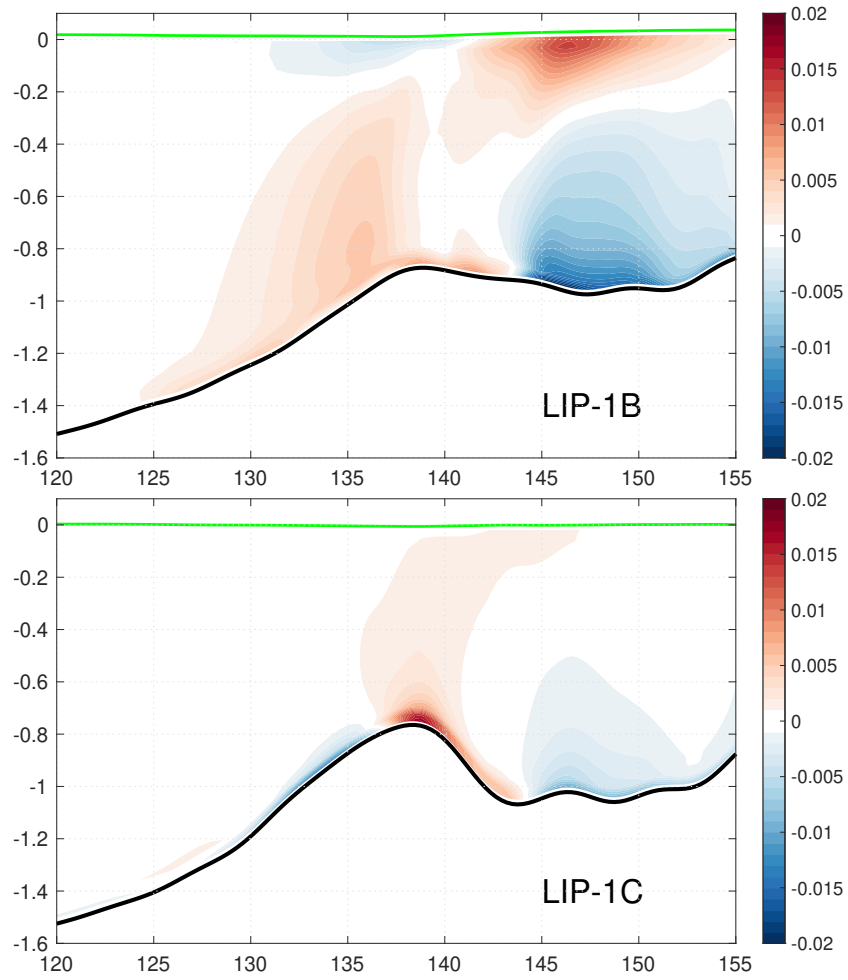


Figure 8: Wave-mean sand advection [$\text{kg}/\text{m}^3/\text{s}$] halfway through the erosion phase (LIP-1B) and accretion phase (LIP-1C). Positive (negative) values indicate a gain (loss) in sand concentration due to transport.

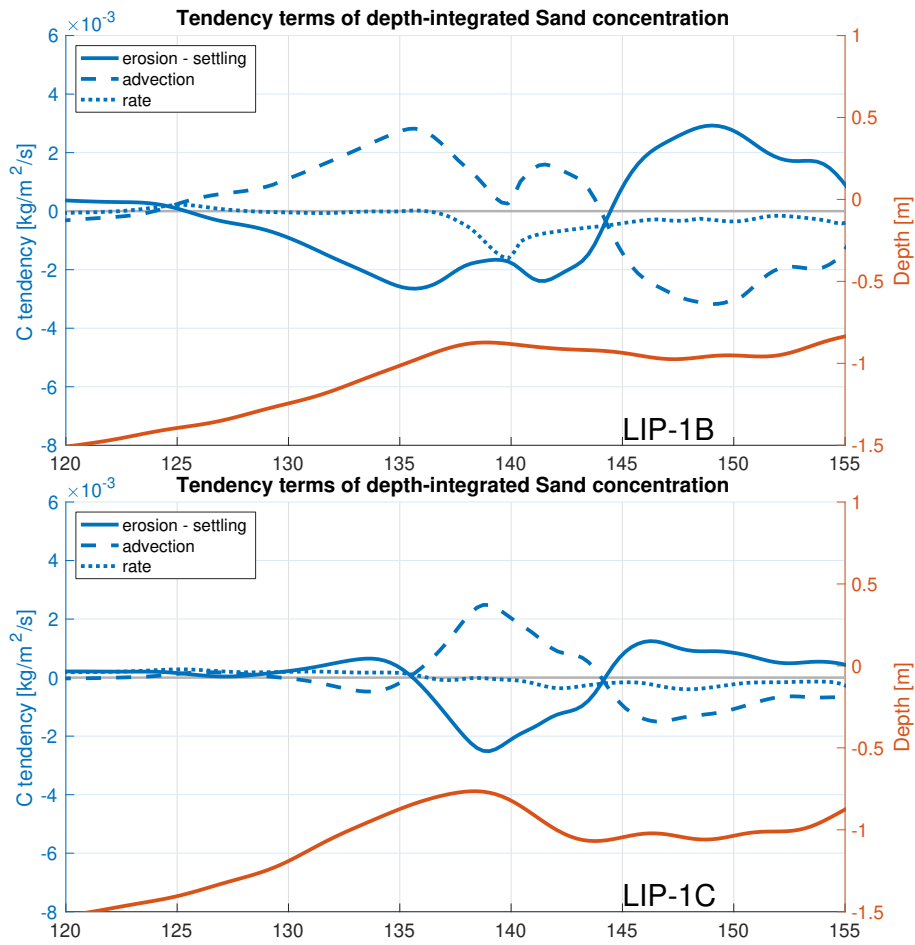


Figure 9: Depth-integrated budget of sand concentration halfway through the erosion phase (LIP-1B) and accretion phase (LIP-1C).

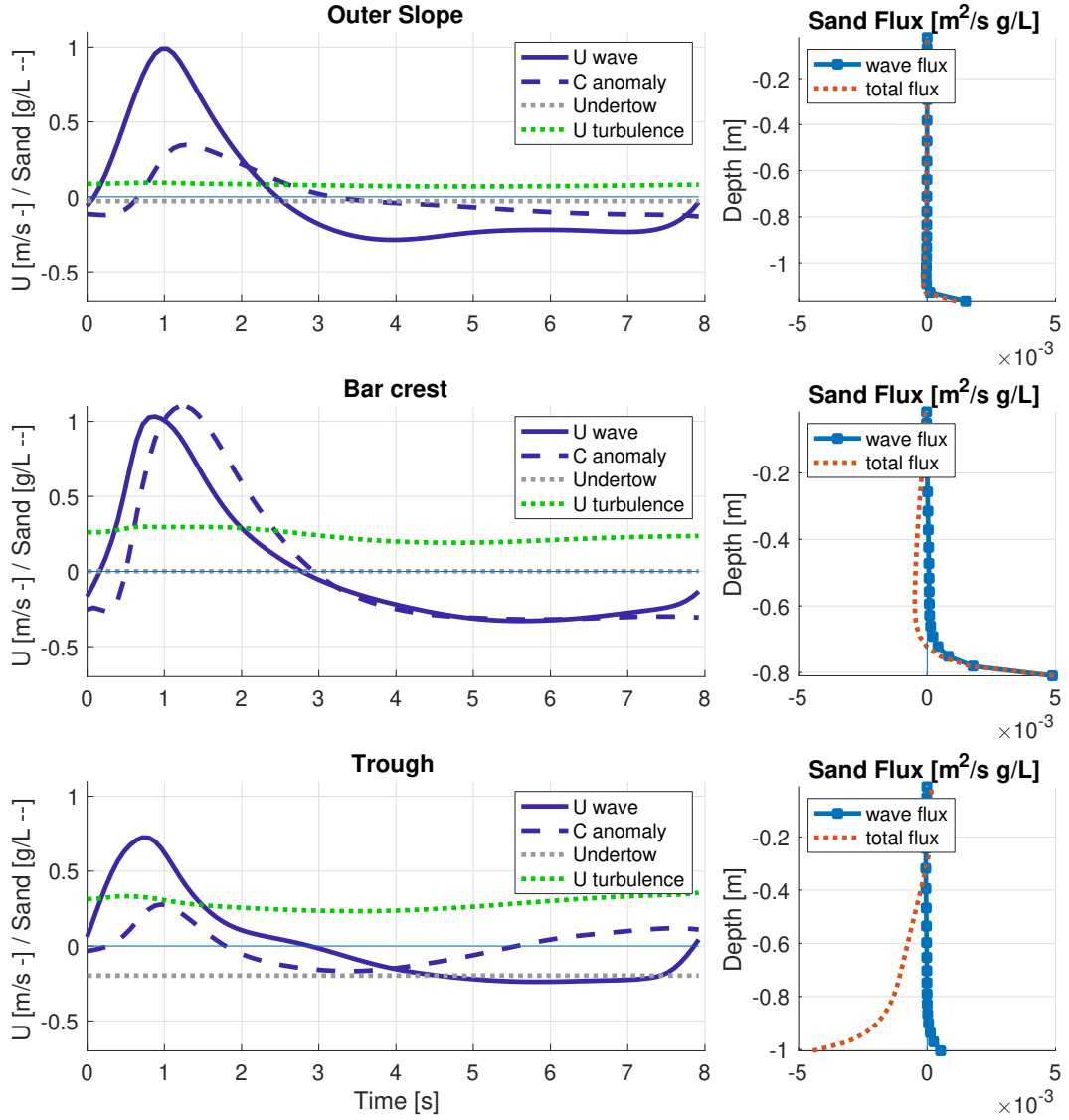


Figure 10: Left panels: ensemble-averaged wave cycle of current and sand concentration anomaly ($u' = u - \bar{u}$ and $C' = C - \bar{C}$) at three positions over the sandbar during the accretive LIP-1C Flume experiment. The wave-mean undertow u_b and mean effective turbulent velocity u_t at the bottom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

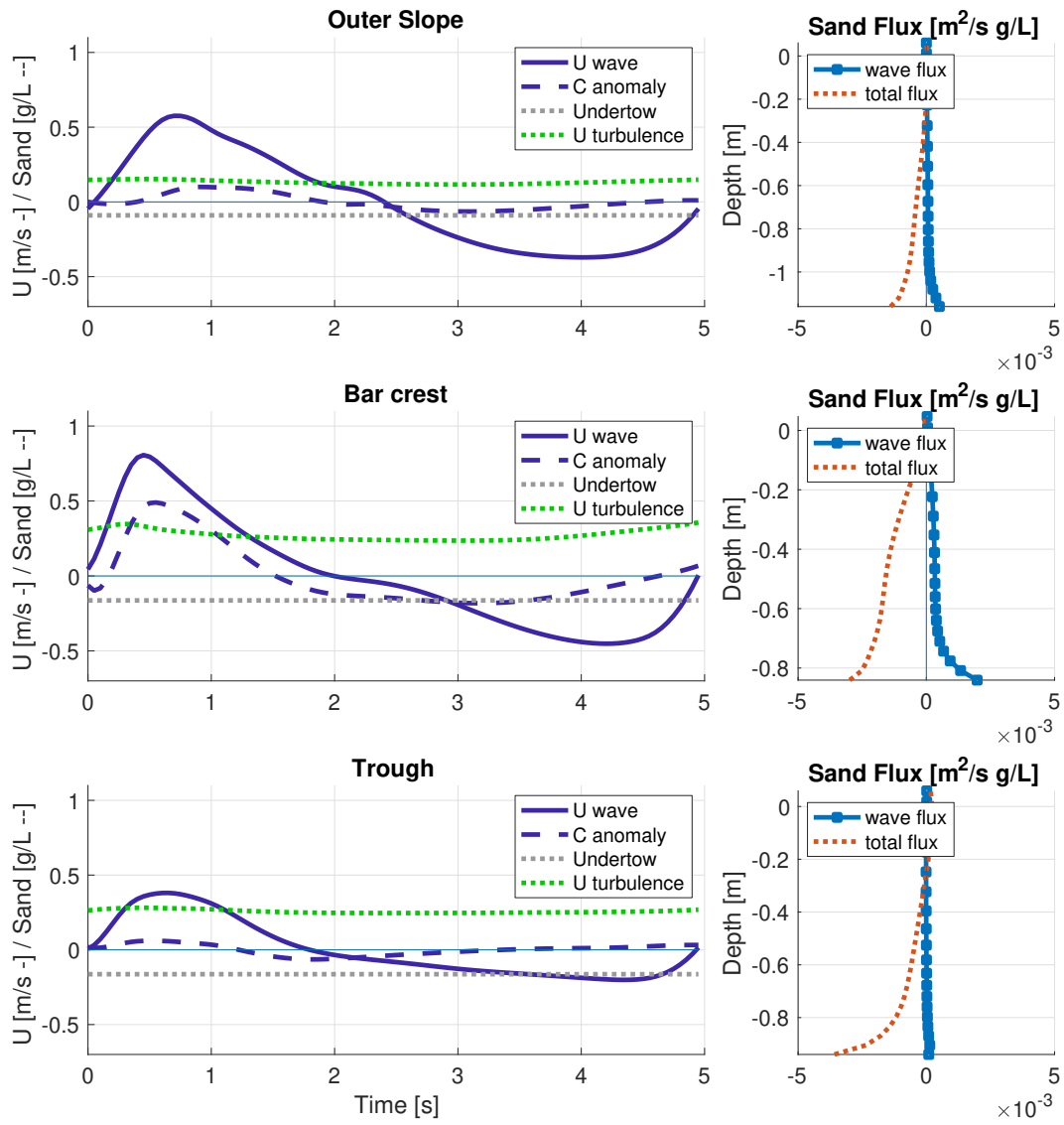


Figure 11: Same as Figure 10 but for the LIP-1B Flume experiment.

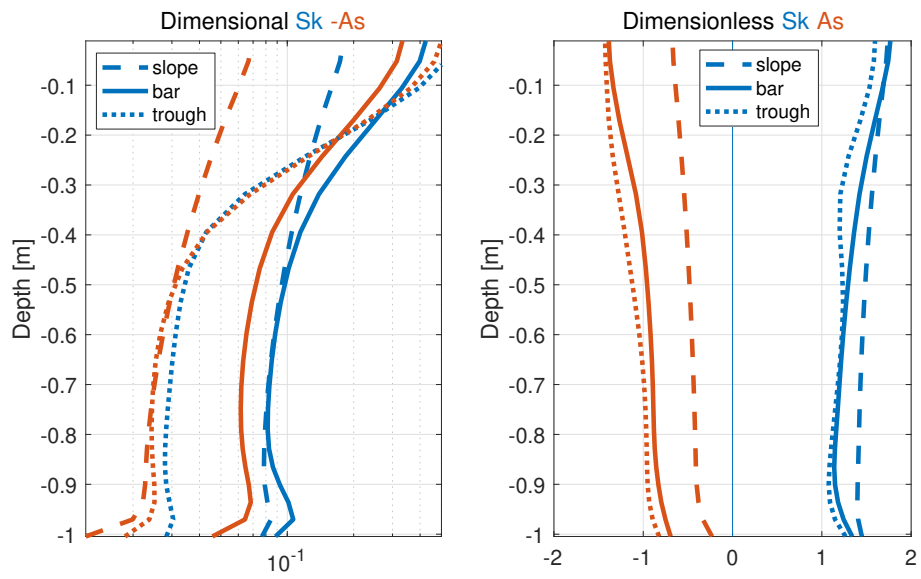


Figure 12: Ensemble-averaged analysis of wave asymmetry and skewness profiles (absolute and normalized) at three locations around the sandbar, during the accretive LIP-1C Flume experiment.

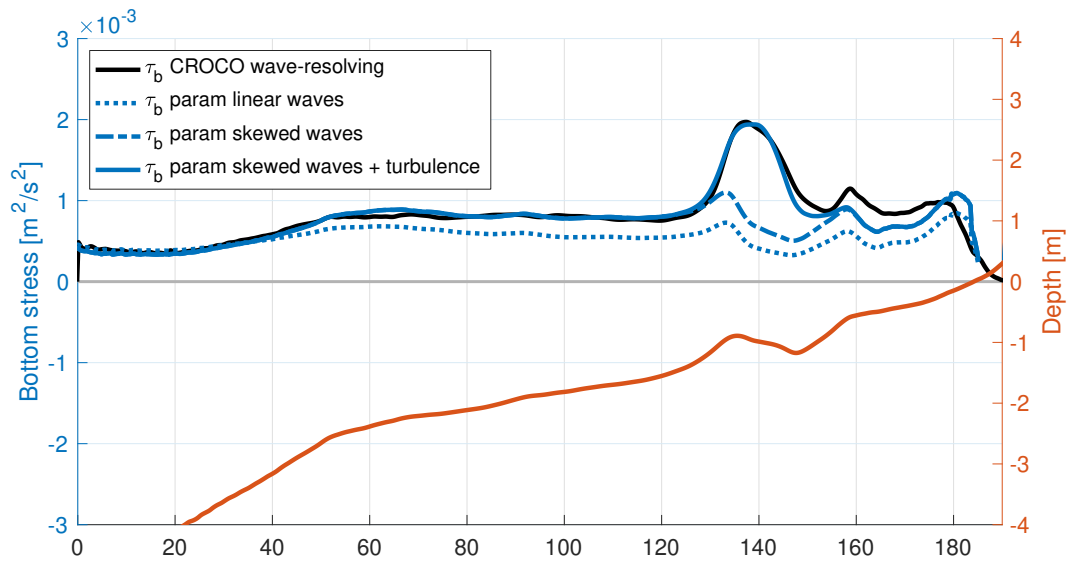


Figure 13: Bed shear stress profiles from the wave-resolving model and reconstructed from empirical formulations using wave-averaged values and linear wave theory with or without skewness correction.

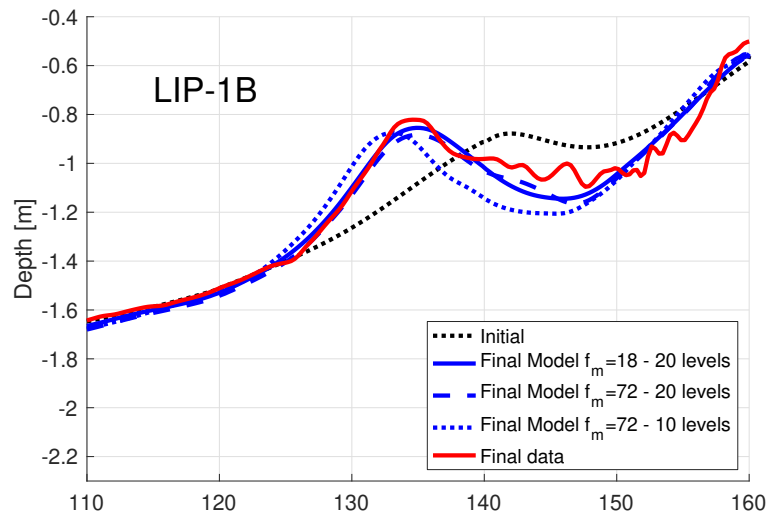


Figure 14: Model sensitivity to morphological acceleration parameter f_m and vertical grid resolution (number of vertical levels) during the erosion phase (LIP-1B).