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Tridimensional nonhydrostatic transient rip currents in a wave-resolving model

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Abstract

Flash rips and surf eddies are transient horizontal structures of the order of 10 to 100 m, which can be generated in the surfzone in the absence of bathymetric irregularities. They are traditionally evaluated in a depth averaged setting which involves intrinsic horizontal shear instabilities and the direct generation of vorticity by short-crested waves. In this article, we revisit the processes of surf eddy generation with a new three-dimensional wave resolution model (CROCO) and provide a plausible demonstration of new 3D non-hydrostatic instability and turbulent cascade. We first present a quick overview of a compressible free surface approach suitable for nearshore dynamics. Its ability to simulate the propagation of surface gravity waves and nearshore wave-driven circulation is validated by two laboratory experiments. Next, we present a real world application from Grand Popo Beach, Benin, forced by waves with frequency and directional spreading. The generation of surf eddies by the 3D model differs from depth-averaged models, due to the vertical shear associated with shallow breaking waves. In this case, the generation of eddies from both horizontal shear instability and the breaking of short-crested waves is hampered, the former by stretching the alongshore current and the latter by inhibiting the inverse energy cascade. Instead, the vertical shear flow is subjected to forced wave group variability and Kelvin-Helmholtz type instability at an inflection point. Primary and secondary instabilities generate spanwise and streamwise vorticity connecting small-scale eddies to larger horizontal surfzone structures. Streamwise

filaments, appearing as 5 m wide ribs or mini-rips, can extend beyond the surfzone but with moderate energy. These results appear consistent with the velocity spectra and observed patterns of tracers and suspended sediments at Grand Popo Beach. The timescale associated with the mean shear-induced turbulence is several times the wave period and suggests an intermediate range between breaker-induced turbulence and large-scale surf eddies.

Keywords: Surfzone, Rip currents, 3D instability, Turbulent cascade, Wave-resolving RANS model

1. Introduction

Flash rips and surf eddies are generally defined as transient horizontal structures of size ranging between water depth and surfzone width, i.e., of order 10-100 m, which are generated in the surfzone in the absence of bathymetric irregularities. They are studied separately from stationary rip currents confined to deeper channels between sandbars (MacMahan et al., 2006; Marchesiello et al., 2015). They are also separated from breaker-induced rollers that scale with wave height (Cox and Anderson, 2001), and even smaller vortices of the fully developed turbulent bore (Svendsen and Madsen, 1984). However, the separation between surf eddies and turbulence is uncertain (Longo et al., 2002) and the possibility of intermediate scales and processes linking horizontal and vertical vorticity generation has been suggested — e.g., Short et al. (1993) describing ephemeral and shallow *mini rips* over Australian low tide terrace beaches — but not clearly demonstrated.

Because it is difficult to sample transient rip currents with sufficient spatial resolution (Lippmann et al., 2016; Henderson et al., 2017), our concepts largely rely on numerical models. Three types of processes stand out:

- Horizontal (2D) shear instability of longshore currents
- Short-crested wave vorticity generation (here called Peregrine process)

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18 • Tridimensional (3D) shear instability

19 The horizontal shear instability of longshore currents was the earliest process proposed
20 for eddy generation, describing the intrinsic variability of wave-induced currents. (Bowen
21 and Holman, 1989; Dodd et al., 1992; Allen et al., 1996; Slinn et al., 1998; Özkan-Haller
22 and Kirby, 1999; Dodd et al., 2000; Uchiyama et al., 2009). This process has generally been
23 studied with wave-averaged shallow water models, in which the momentum transfer from
24 waves to currents is fully parametrized. Its importance has faded over the last decade due
25 to the prevalence of the Peregrine process, but also to the contradictory results given by
26 three-dimensional wave-averaged models. (Newberger and Allen, 2007; Splinter and Slinn,
27 2003).

28 The second process, largely due to Peregrine (1998), is the current nearshore community
29 views of driving mechanisms for wave-averaged circulation in the surfzone (Kirby and De-
30 rakhti, 2019). Boussinesq equations for weakly dispersive intermediate and shallow water
31 waves provides a conceptual model for the action of spatially varying wave breaking, i.e.,
32 short-crested waves (Johnson and Pattiaratchi, 2006; Bonneton et al., 2010; Feddersen et al.,
33 2011; Clark et al., 2012; Feddersen, 2014). In this model, small vortices result from gener-
34 ation by differential breaking, and combine over time into larger eddies through an inverse
35 cascade mechanism, consistent with 2D turbulence. The surfzone is thus a production center
36 for eddies with scales roughly ranging from 10 to 100 m. In addition, the coastal boundary
37 imposes that eddies and associated filaments can only go offshore, providing a mechanism
38 for enhanced cross-shore dispersion of various tracers.

39 The theoretical framework from depth-integrated models neglect the effect of verti-
40 cal shear. Following the advent of robust 3D formulations of wave-averaged equations
41 (McWilliams et al., 2004; Ardhuin et al., 2008), a number of 3D modeling studies have
42 emerged in the last decade (Newberger and Allen, 2007; Uchiyama et al., 2010; Kumar
43 et al., 2012; Marchesiello et al., 2015; Uchiyama et al., 2017; McWilliams et al., 2018; Akan
44 et al., 2020). They show a modulation of nearshore circulation when wave breaking occurs in
45 a shallow surface layer. However, short-crested wave breaking is generally neglected in these

46 wave-averaged studies (or addressed in ad-hoc manners) and all real-scale applications to
47 date are performed using hydrostatic assumption, thus underestimating horizontal vorticity
48 motions.

49 Nonhydrostatic dynamics are essential in our third listed process of surf eddy generation.
50 They are mostly studied in laboratory experiment (Nadaoka et al., 1989) and laboratory-
51 scale Large-Eddy Simulations (LES) using 2.5D CFD models applied to individual wave
52 breaking (Lin and Liu, 1998; Li and Dalrymple, 1998; Watanabe and Saeki, 1999; Watanabe
53 et al., 2005; Lubin and Glockner, 2015). These previous studies show that the spanwise
54 (mostly alongshore) component of vorticity is an important aspect of the breaking process.
55 Surface breaking produces traveling rolls through a primary instability, which can evolve
56 through secondary instability to produce streamwise vorticity, transitioning toward fully
57 tridimensional turbulence.

58 However, 2.5D CFD models are computationally very expensive and applied to individual
59 breaking waves with only few alongshore wavelengths of the secondary instability, precluding
60 any evaluation of eddy statistics. In addition, these studies do not always clearly distinguish
61 whether the instability is associated with the instantaneous plunging and rebounding jet
62 produced by breakers or with the mean shear flow caused by momentum transfer. Yet, the
63 two processes may be sorted by their timescale, i.e., smaller than the wave period for the
64 rebounding jet (Watanabe et al., 2005) and longer for the mean shear turbulence (Li and
65 Dalrymple, 1998). If confirmed, the latter could therefore be an intermediate phenomenon
66 between breaker-induced turbulence and large-scale surf eddies.

67 3D nonhydrostatic processes are usually studied independently of the two others (by
68 separate research communities) and rarely compared in terms of scales, magnitude and
69 interaction. The only attempt was made by Splinter and Slinn (2003) in a proceeding
70 report. Using a 3D nonhydrostatic model where breaking acceleration is introduced as a
71 body force, they show that a simulation with deep breaking reproduces 2D model solutions,
72 while the more realistic shallow breaking process seems to disrupt the formation of horizontal
73 shear instability at the expense of vertical shear instability. However, their domain size does
74 not allow statistical comparisons and the profile of breaking acceleration is imposed, not

75 computed from a wave-resolving model. The present study is a step forward compared with
76 this first work, also addressing the case of short-crested wave generation. Note that Kumar
77 and Feddersen (2017) studied transient eddies produced by a 3D nearshore circulation model,
78 forced by short-crested waves computed beforehand with a Boussinesq model. However, wave
79 forcing is prescribed as a depth-uniform body force, i.e., as deep breaking, and could not
80 produce vertical shear of the cross-shore flow. Their hydrostatic assumption also precluded
81 the model from vertical shear instabilities.

82 CROCO (Coastal and Regional Ocean Community model) is a new oceanic modeling
83 system built upon ROMS (Shchepetkin and McWilliams, 2005; Debreu et al., 2012), with
84 an added non-Boussinesq kernel (Auclair et al., 2018). It solves Reynolds-averaged Navier-
85 Stokes Equations (RANS) on a free surface and terrain-following grid and is designed to
86 study realistic, fine-scale processes from the regional ocean to the littoral zone. Particular
87 attention is paid to numerical accuracy, high performance computing (optimization, scalabil-
88 ity), portability and ease of access (url www.croco-ocean.org). This paper presents a quick
89 overview of the nonhydrostatic CROCO solver with a non-Boussinesq (compressible) ap-
90 proach, before embarking in its application to nearshore dynamics. First, its ability to simu-
91 late the propagation of surface gravity waves, near-shore breaking and the resulting vertical
92 circulation is validated against small and large-scale laboratory experiments. Second, we
93 present a 3D, wave-resolving, real-case simulation of transient rips, in the presence (or not)
94 of short-crested waves and strong alongshore currents. We discuss fundamental differences
95 in the generation of surf eddies by 3D wave-resolving models compared with depth-averaged
96 models, with a focus on the vertical structure of currents produced by shallow breaking
97 and associated turbulence. We conclude on the limitation of simplified vorticity evolution
98 equations in which only the vertical part is considered when so much activity resides in the
99 horizontal vorticity, governed by 3D non-hydrostatic equations.

100 **2. Model description**

101 Because of limited computational resources, 3D wave-resolving models are still rarely
102 used to study nearshore dynamics in realistic environments. LES applications appeared in

103 the 1990s and are generally restricted to 2.5D laboratory-scale experiments of individual
104 wave breaking. Early applications used the volume-of-fluid (VOF) method for free-surface
105 tracking (e.g., Lin and Liu 1998; Watanabe and Saeki 1999; Watanabe et al. 2005; Derakhti
106 and Kirby 2014; Larsen et al. 2020). This model type with Cartesian coordinate, where
107 the free surface crosses computational cells arbitrarily, fails to precisely apply the pressure
108 boundary condition on the free surface, affecting the model accuracy.

109 More recently, several 3D wave-resolving, free-surface and terrain-following RANS models
110 have emerged for the nearshore zone, e.g., SWASH (Zijlema et al., 2011) and NHWAVE (Ma
111 et al., 2012; Derakhti et al., 2016), based on earlier attempts (e.g., Lin and Li 2002). In
112 this case, the explicit overturning of the free surface is excluded and the breaking wave is
113 modeled instead with a single-valued free surface which follows a shock process and resembles
114 a dissipating bore. Despite the absence of explicit overturning (replaced by parametrized
115 turbulence), these models can be accurate as well as computationally efficient (orders of
116 magnitude cheaper) in the study of waves and wave-driven mean and transient circulation.

117 CROCO belongs to this class of models but, unlike other attempts, resolves the com-
118 pressible Navier-Stokes equations (Auclair et al., 2018). A compressible approach preserves
119 the hyperbolic nature of Navier-Stokes equations and does not require a global elliptic solver
120 with incremental pressure corrections to ensure the incompressible mass balance. As a re-
121 sult, it avoids splitting errors between pressure and velocity and approximations made on
122 free-surface conditions (Zijlema et al., 2011; Derakhti et al., 2016), thereby preserving ampli-
123 tude and nonlinear dispersive properties of surface waves. In the same time, the absence of
124 global computations by an elliptic solver makes parallelization and optimization procedures
125 much more efficient. The cost of solving acoustic waves is managed with a time-splitting
126 technique and semi-implicit time discretization, introduced below.

127 The development of CROCO around the Regional Oceanic Modeling System (ROMS)
128 has advantages for realistic applications. It benefits from capabilities long developed in
129 oceanic models – high-performance computing; high-order discretization; coupling with bio-
130 geochemistry and sediment models; pre-processing tools for rapid generation of model input;
131 various online and offline diagnostics. The nonhydrostatic model version can thus be ap-

132 plied without much effort to realistic, highly nonlinear regimes, e.g., large internal solitons
 133 and hydraulic jumps (Hilt et al., 2020), Kelvin-Helmholtz instabilities (Penney et al., 2020),
 134 Langmuir turbulence (Herman et al., 2020) or wave-induced nearshore circulation as in the
 135 present study. It is naturally suited for bridging ocean and coastal sciences, e.g., addressing
 136 surf-shelf exchange processes in a 3D, rotating and stratified framework. In addition, both
 137 wave-resolving and wave-averaged (Uchiyama et al., 2010; Marchesiello et al., 2015) model
 138 equations are available within the same code, which has potential advantages for evaluating
 139 the parametrizations of wave-current interactions in wave-averaged models.

140 *2.1. Free-surface, compressible ocean model equations*

141 The full set of Navier-Stokes equations for a free-surface ocean is explicitly integrated
 142 in the nonhydrostatic, non-Boussinesq (compressible) version of CROCO, built on the code
 143 structure of ROMS primitive equation solver. In the compressible approach (Auclair et al.,
 144 2018), acoustic waves are solved explicitly to avoid Boussinesq degeneracy, which inevitably
 145 leads to a 3D Poisson system in nonhydrostatic incompressible methods – detrimental to
 146 computational costs and accuracy of free-surface model implementation.

147 Non-Boussinesq equations include the momentum and continuity equations, the surface
 148 kinematic relation (for free surface), heat, salt or other tracer C conservation equations, and
 149 the equation of state, which reads in Cartesian coordinates:

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

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

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

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

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

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

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

155 2.2. Time-splitting principle

156 In the above set of equations, a relation between ρ and P is required. To that end, and
 157 as part of a time-splitting approach, density is decomposed into slow and fast components
 158 based on a first-order linear decomposition with respect to total pressure. In the following,
 159 s and f subscripts refer to slow and fast-mode components respectively:

$$\rho = \rho_s(T, S, P_s) + \overbrace{\frac{\partial \rho}{\partial P} \Big|_{T,S} \delta P + O(\delta P^2)}^{\rho_f = c_s^{-2} P_f} \quad (7)$$

$$P = \underbrace{P_{atm} + \int_z^\eta (\rho_s - \rho_0) g \, dz'}_{SLOW} + \underbrace{\rho_0 g (\eta - z) + \overbrace{\delta P}^{P_f}}_{FAST} \quad (8)$$

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

161 The Navier-Stokes equations are then integrated with two different time steps within
 162 the time-splitting approach inherited from ROMS. The slow-mode integration is similar
 163 to ROMS, with the addition of the slow part of vertical momentum equation, while fast-
 164 mode integration is in 3D and includes the compressible terms of momentum and continuity
 165 equations. In vector form:

$$\frac{\partial \rho \vec{v}}{\partial t} = \underbrace{-\vec{\nabla} \cdot (\rho \vec{v} \otimes \vec{v}) - 2\rho \vec{\Omega} \times \vec{v} - \vec{\nabla} \left(\int_z^{\eta_f} (\rho_s - \rho_0) g \, dz' \right) + \vec{\mathcal{F}}_{\vec{v}} + \vec{\mathcal{D}}_{\vec{v}}}_{SLOW} + \underbrace{-\rho_0 g \vec{\nabla} \eta_f - \vec{\nabla} P_f + \rho_f \vec{g} + \lambda \vec{\nabla} (\vec{\nabla} \cdot \vec{v})}_{FAST} \quad (9)$$

$$\frac{\partial \rho_f}{\partial t} = -\frac{\partial \rho_s}{\partial t} - \vec{\nabla} \cdot (\rho \vec{v}) \quad (10)$$

$$P_f = c_s^2 \rho_f \quad (11)$$

$$\frac{\partial \eta_f}{\partial t} = w_f|_{z=\eta} - \vec{v}_f|_{z=\eta} \cdot \vec{\nabla} \eta_f \quad (12)$$

$$\frac{\partial \rho C_s}{\partial t} = -\vec{\nabla} \cdot (\rho \vec{v} C_s) + \mathcal{F}_C + \mathcal{D}_C \quad (13)$$

$$\rho_s = \rho(T_s, S_s, \eta_f) \quad (14)$$

$$\rho = \rho_s + \rho_f \quad (15)$$

166 The momentum is integrated both in slow and fast modes but the right-hand-side of
 167 the equation is split in two parts: a slow part, made of slowly varying terms (advection,
 168 Coriolis force, baroclinic pressure force and viscous dissipation), and a fast part, made of fast-
 169 varying terms (the surface-induced and compressible pressure force, weight, and dissipation
 170 associated with bulk-viscosity). This momentum equation is numerically integrated twice,
 171 once with a large time-step keeping the fast part constant, and once with a smaller time-step
 172 keeping the slow part constant. This is much more computationally efficient than integrating

173 the whole set of equations at the same fast time step. More details can be found in Auclair
174 et al. (2018)¹.

175 Note that acoustic waves can become pseudo-acoustic if their phase speed c_s is artificially
176 reduced (c_s is a model parameter). In this case, high-frequency processes associated with
177 bulk compressibility may be unphysical, but an accurate solution for slower nonhydrostatic
178 dynamics can be preserved, while relaxing CFL constraints. In our nearshore applications, a
179 c_s value of 200 m/s instead of 1500 m/s makes almost no difference for the physical solution
180 but allows a great reduction in the computation time (by almost half).

181 *2.3. Discretized equations for nearshore applications*

182 In this study, motions are produced by an offshore wave-maker in a non-rotating, homo-
183 geneous fluid. In this case, the Coriolis force, baroclinic pressure force and all surface fluxes
184 are null. There is no temperature or salinity stratification so that slow density ρ_s is constant
185 in time and space.

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

194 Horizontal and vertical advection terms are discretized using the WENO5-Z improved
195 version of the 5th-order weighted essentially non-oscillatory scheme (Borges et al., 2008),
196 which is popular for hyperbolic problems containing both shocks and smooth structures.

¹Auclair et al. (2018) presented a first implementation of the compressible approach involving a 3-level time splitting (internal, external and acoustic). CROCO was simplified to only retain a slow and a fast time level, where acoustic waves are solved together with the external (depth-averaged) mode. This procedure is more computationally efficient

197 WENO5-Z naturally copes with dispersive numerical modes as well as shocks caused by
 198 breaking waves, with no need for ad hoc criteria.

199 2.4. Turbulence closure

200 Along with the numerical treatment of breaking waves, a k - ϵ or k - ω model, solving
 201 the closure equations for turbulent kinetic energy k and dissipation ϵ or dissipation rate
 202 $\omega \propto \epsilon k^{-1}$, is used as part of a Generic Length Scale (GLS) method (Warner et al., 2005).
 203 In the absence of buoyancy forcing, the turbulence equations express a balance between
 204 transport, diffusion, shear production and dissipation:

$$\frac{\partial \rho k}{\partial t} = -\vec{\nabla} \cdot (\rho \vec{v} k) + \mathcal{D}_k + \rho(P - \epsilon) \quad (16)$$

205

$$\frac{\partial \rho \epsilon}{\partial t} = -\vec{\nabla} \cdot (\rho \vec{v} \epsilon) + \mathcal{D}_\epsilon + \rho \frac{\epsilon}{k} (c_{\epsilon 1} P - c_{\epsilon 2} \epsilon) \quad (17)$$

206 OR

$$\frac{\partial \rho \omega}{\partial t} = -\vec{\nabla} \cdot (\rho \vec{v} \omega) + \mathcal{D}_\omega + \rho \frac{\omega}{k} (c_{\omega 1} P - c_{\omega 2} \epsilon) \quad (18)$$

207 The eddy viscosity $\nu_t = c_\mu l k^{\frac{1}{2}}$ is derived from these equations, with coefficient c_μ depen-
 208 dent on stability functions, and mixing length $l \propto k^{\frac{3}{2}} \epsilon^{-1}$. l is resolution independent, which
 209 is consistent with a RANS rather than LES approach. The shear production term for k is
 210 $P = 2\nu_t S_{ij} S_{ij}$, 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 notation).
 211 All turbulence model parameters are given in Warner et al. (2005), based on Burchard et al.
 212 (1998) for k - ϵ and Wilcox (1988) for k - ω . The only present modification in the k - ϵ model
 213 concerns the surface mixing length (a model boundary condition), which is briefly discussed
 214 in the validation section 3.2. For this reason, and for its robustness through resolutions and
 215 benchmarks, the k - ω model will be our standard turbulence model. However, we note as
 216 Mayer and Madsen (2000) and Larsen and Fuhrman (2018) that this model tends to pro-
 217 duce excessive mixing in potential flow regions, i.e., on the innershelf. This problem will be
 218 addressed in further studies, including realistic conditions of stratification and wind forcing.

219 *2.5. Wave maker at offshore boundary*

220 The wave maker forces a spectrum of 3D linear waves at the offshore boundary, as in
 221 Zijlema et al. (2011), rather than as an interior source term (Wei et al., 1999). The spectrum
 222 has frequency and directional spreading similar to Feddersen et al. (2011):

$$\eta_{bc}(y, t) = \sum_i a_i \sum_j d_j \cos(k_{y,ij}y - \omega_i t - \phi_{ij}) \quad (19)$$

223

$$\text{with } d_j = e^{-\left(\frac{\theta_j - \theta_m}{\sigma_\theta}\right)^2} \text{ and } \sum d_j = 1 \quad (20)$$

224

$$u_{bc}(y, t) = \eta_{bc}(y, t) \omega_p \cos(\theta_m) \frac{\cosh(k_p(z+h))}{\sinh(k_p h)} \quad (21)$$

225

$$v_{bc}(y, t) = \eta_{bc}(y, t) \omega_p \sin(\theta_m) \frac{\cosh(k_p(z+h))}{\sinh(k_p h)} \quad (22)$$

226 where (x, y, z) are cross-shore, alongshore and vertical directions respectively; (i, j) are in-
 227 dices of spectral distribution in frequency and direction respectively; a_i is the amplitude
 228 at each frequency ω_i , from a given statistical distribution, e.g., JONSWAP (Sec. 4.1);
 229 $k_{y,ij} = k_i \sin(\theta_j)$ is the alongshore wavenumber, where k_i is the linear theory wavenumber:
 230 $\omega_i^2 = g k_i \tanh(k_i h)$ with h the mean water depth; θ_j is wave angle, θ_m is the mean wave
 231 direction and σ_θ the directional spread around the mean; ω_p and k_p are peak frequency
 232 and wavenumber; d_j is a normalized frequency-dependent directional distribution; ϕ_{ij} is a
 233 uniformly distributed random phase.

234 Here w_{bc} is set to zero and our tests show only weak sensitivity to this choice. Depth-
 235 averaged (barotropic) velocities (\bar{u}, \bar{v}) must be provided as well in the wave maker because
 236 they are prognostic variables of our split-explicit model, advanced together with the fast
 237 acoustic mode. Normal depth-averaged velocity \bar{u} is complemented at the boundary by an
 238 anti-Stokes "compensation flow", opposite to Stokes drift and thus closing the volume bud-
 239 get. We do not impose the depth-averaged value of u_{bc} directly but the value of the incoming
 240 characteristic of the shallow water system as in Flather-type conditions (Marchesiello et al.,

241 2001; Blayo and Debreu, 2005):

$$\bar{u} = \bar{u}_{bc} - \sqrt{\frac{g}{h}}(\eta - \eta_{bc}) \quad (23)$$

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

247 3. Validation in Flume experiments

248 3.1. GLOBEX experiment

249 As a first step towards 3D modeling, we present here a validation of wave propagation
250 and breaking using a wave flume experiment. The Gently sLOping Beach EXperiments
251 (GLOBEX²) were performed in the Scheldt flume of Deltares (Delft, the Netherlands) in
252 2012, and described in Michallet et al. (2014). The project objective was to collect high-
253 resolution space-time data of the cross-shore evolution of short and infragravity waves on a
254 gentle slope for a range of wave conditions.

255 The flume is 110 m long, 1 m wide and 1.2 m high. The waves were generated with a
256 piston-type wave maker equipped to minimize reflections from the wave paddle. A concrete
257 beach with a weakly reflexive 1:80 slope was constructed, with its toe at 16.57 m from the
258 wave maker. All experiments were run with a still water depth of 0.85 m and shoreline at
259 $x = 84.57$ m. The material that was laying loose on the concrete bed before the flume was
260 filled with water had a median grain size $D_{50} = 0.75$ mm.

261 Sea-surface elevation measurements were taken at 190 locations (repeating an experiment
262 ten times while relocating the 21 wave gauges), together with velocity measurements at 43
263 locations, mostly (but not always) at 1 cm above bed to focus on the undertow. The
264 sampling frequency of the instruments during these experiments is 128 Hz. Here, we focus

²Globex data is freely available at zenodo.org/record/4009405

265 on experiment B3, corresponding to second-order (Stokes) wave generation of bichromatic
266 frequencies (simulated with a Boussinesq model in Michallet et al. 2014). The characteristics
267 are as follows: $a_1 = 0.09$ m; $a_2 = 0.01$ m; $f_1 = 0.420$ Hz; $f_2 = 0.462$ Hz (short-wave peak
268 period $T_p = 2/(f_1 + f_2) = 2.27$ s; and group period $T_g = 1/(f_2 - f_1) = 23.81$ s). The signal
269 had a total duration of 75 minutes.

270 The model is set-up with the same conditions as the wave flume experiment. Second-
271 order bichromatic waves are generated at the offshore boundary, with shore normal direction
272 and zero directional spread. A no-slip condition is imposed on the lateral wall boundaries
273 of the canal so that transverse modes are precluded. The grid spacing is $dx = 1$ cm with 10
274 vertical levels evenly spaced between the free surface and bottom. A simulation with 20 levels
275 gave similar results, while the solution is moderately degraded (mostly in higher moments)
276 with coarser horizontal resolution ($dx=3, 6$ and 12 cm), which shows good convergence
277 properties. The model time step is $dt = 0.15$ ms. The minimum depth is 1 mm on the
278 shore, the position of which varies with the swash oscillation, relying on a wetting-drying
279 scheme (Warner et al., 2013). For bottom drag, the logarithmic law of the wall is used with
280 roughness length $z_0 \sim D50/12 = 0.0625$ mm.

281 Figure 1 compares an Hovmuller plot (x,time) of data and model sea level η and un-
282 dertow u_b . u_b is interpolated at the measurement depth. When some data is missing in the
283 measurements, it is also removed from the model output. The general structure reflecting
284 wave speed and frequencies, wave packets, surf and swash zones are all very similar. Model-
285 data correlations are high with 0.85 and 0.80 respectively for η and u_b , and root mean square
286 errors are 2.7 cm and 12.2 cm/s. Some scattering in the undertow data is noticeable. Ac-
287 cording to Michallet et al. (2014), it may be attributed to the presence of secondary motions
288 generated by transverse waves at the break point where the transverse mode-1 seiche can
289 be excited at frequency $f_1 + f_2$. It may also be due to variations in the depth level of flow
290 measurements.

291 A snapshot of wave field across the flume during runup (Figure 2) highlights the main
292 processes of propagation, nonlinear wave-wave interactions, shoaling, breaking, roller prop-
293 agation and runup. Model-data correlation is high, as already mentioned, and non-linearity

294 is apparent in both cases in the increasingly non-sinusoidal shape of short waves as they
295 approach the shore (Elgar and Guza, 1985). They first develop short, high wave crests with
296 increasing skewness (asymmetry about the horizontal axis, measuring crest/trough shape),
297 and as they break, transition into the characteristic saw-tooth shape with asymmetry about
298 the vertical axis.

299 Wave statistics (first, second and third moments) for η and u_b are shown in Figure 3:
300 mean, standard deviation (or H_s for η), skewness $\langle\phi'^3\rangle/\langle\phi'^2\rangle^{1.5}$ and asymmetry $\langle\mathcal{H}(\phi')^3\rangle/\langle\phi'^2\rangle^{1.5}$
301 (\mathcal{H} is Hilbert transform). The model sea-level statistics (left of Fig. 3) closely resemble the
302 measurement data, including high-order moments, showing the transition from skewness to
303 asymmetry across the shoaling and surf zones (with two peaks in the asymmetry profile
304 corresponding to outer and inner surf zone evolution).

305 For the mean field of measured undertow, a few scattered points lying far outside the
306 standard deviation are corrected using a polynomial fit. The model appears to replicate
307 the observed cross-shore undertow profile (top-right of Fig. 3). The undertow is part of a
308 vertical recirculation associated with breaking-induced surface onshore flow. Here, we call
309 undertow the bottom return flow (10 cm/s in this experiment), which includes the Eulerian
310 anti-Stokes compensation flow (preserving Lagrangian flow continuity). The latter is the
311 only undertow component captured by depth-averaged models and is relatively weak (on
312 the order of 1 cm/s in this simulation). In Figure 3, the model also correctly represents
313 high-order moments, which show profiles similar to those of the sea level (skewness and
314 asymmetry).

315 From sensitivity tests, it appears that a realistic reproduction of H_s cross-shore evolution
316 in the surfzone benefits from using a shock capturing scheme (WENO5-Z). The results
317 are degraded (not shown) when replacing WENO5-Z with UP5, a non-monotonic linear
318 upstream-biased 5th-order advection scheme (Menesguen et al., 2018). This is in line with
319 analogies between breaking waves, bores and hydraulic jumps, that may be treated as a shock
320 (Cienfuegos et al., 2010; Tissier et al., 2012; Lubin and Chanson, 2017). The preservation
321 of steep wave fronts by shock-capturing schemes can help generate asymmetry, even to the
322 excess depending on resolution (Tissier et al., 2012). However, this is balanced here by

323 additional eddy viscosity from the turbulence model, which is otherwise required (below
324 wave trough level) to capture the right vertical shear, as shown next.

325 *3.2. Large-scale LIP flume experiment*

326 The undertow may be considered as a proxy for vertical shear, which will appear as
327 an essential parameter of surf eddy generation in the next section. To further confirm the
328 model's ability to simulate this shear, we now present a comparison with the European Large
329 Installation Plan (LIP) experiments, designed for profile validation and carried out at full
330 scale in Delft Hydraulics's Delta Flume (Roelvink and Reniers, 1995). It will also be a test
331 for our numerical wave maker in its ability to generate a spectrum of random waves.

332 The Flume is 225 m long, 5 m wide and 7 m deep. In LIP, three types of experiments were
333 designed with different wave conditions, which subsequently resulted in a stable (A), erosive
334 (B), and accretive (C) beach state. Here, we use the erosive experiment (LIP11D-1B). The
335 wave conditions were a JONSWAP narrow-banded random wave spectrum generated by
336 a wave paddle, with characteristics $H_s = 1.4$ m, $T_p = 5$ s, and peak enhancement factor
337 $\gamma = 3.3$. Under this wave forcing, a sandbar formed and slowly migrated across the initial
338 beach profile of slope 1:30, consisting of a median grain size of 0.22 mm ($z_0 \sim D50/12 = 0.02$
339 mm). A movable carriage was placed 10 cm above the bed to capture the depth-varying
340 structure of the currents at 10 locations along the flume with a given accuracy of 2 cm/s.
341 We use measurements taken after 8 hours in experiment 1B and averaged over one hour.

342 The model setup is adapted from the GLOBEX experiment to the LIP experiment. In
343 particular, a JONSWAP wave spectrum similar to the experiment is generated with shore
344 normal direction and zero directional spread. The grid spacing is $dx=25$ cm with 10 vertical
345 levels evenly spaced between the free surface and bottom. Here again, a simulation with 20
346 levels gave similar results and a test of coarser horizontal resolutions ($dx=50$ cm and 1 m)
347 is presented below. The model time step is $dt = 25$ ms.

348 Figure 4 shows a comparison of the model with data, using our standard configuration.
349 The match with measured currents is very good throughout the complex morphology of the
350 beach. The waves start to break before the sandbar, but the breaking is more intense on

351 the sandbar, where the surf is strongest on the onshore side. The resulting undertow has a
352 strong shear and maximum intensity of about 30 cm/s. The resolution test (25 cm, 50 cm
353 and 1 m) shows a mean error of about 3 cm/s at all resolution, close to the measurement
354 error of 2 cm/s. The results are thus consistent at all resolution despite no adjustment of
355 any parameter. With the small-scale GLOBEX experiment, it confirms the validity of a
356 RANS approach for estimating the mixing length of breaking-induced turbulence — which
357 will be distinguished from mean shear-induced turbulence in the next section.

358 Turbulent kinetic energy and eddy viscosity estimated by the k - ω model in the breaker
359 zone have the expected structure (Fig. 4; top) and magnitude ($\nu_t \sim 0.01h\sqrt{gh}$; Svendsen
360 1987; Cox et al. 1994). Interestingly, the transport terms in the closure equations tend to
361 reduce mixing at break point by redistributing the turbulent energy, thus allowing a more
362 intense shear to be maintained (not shown). The k - ϵ model works almost as well as the k - ω
363 model, with respect to mean current profiles, but the comparison is improved by imposing
364 a high value on the surface mixing length (0.2 m), as in wave-averaged models (Feddersen
365 and Trowbridge, 2005; Kumar et al., 2012)). The k - ω model may thus be a better choice
366 for surface wave breaking, possibly due to a more accurate near wall treatment (Mayer
367 and Madsen, 2000; Devolder et al., 2018; Larsen et al., 2020). Note, however, that this
368 model produces a greater amount of mixing in potential flow regions outside the surf zone
369 (innershelf), mainly due to the divergence part of the mean strain rate tensor Mayer and
370 Madsen (2000).

371 While an extensive study of surfzone turbulence is beyond the scope of this paper, we
372 conclude that our combination of numerical and physical closures with off-the-box paramet-
373 ers, although perfectible, provides a realistic and robust framework for the horizontal and
374 vertical circulation in the surfzone.

375 4. Natural beach application

376 We now turn to a full 3D experiment with longshore-uniform bathymetry. The configu-
377 ration is derived from Grand Popo Beach (6.2°N, 1.7°E, Benin, in the Gulf of Guinea; Fig.
378 5). This stretch of coast presents a longshore-uniform, low tide terrace and steep upper

379 shoreface (Almar et al., 2014, 2016) and a sandy wave-dominated and microtidal environ-
380 ment, exposed to S-SW long period swells generated at high latitudes in the South Atlantic
381 (Almar et al., 2015a). A field experiment was conducted at Grand Popo Beach from 10
382 to 18 March 2014 (Almar et al., 2014; Derian and Almar, 2017). For our setup, we focus
383 on conditions in the middle of the afternoon of March 13 2014. The weather, tides and
384 wave conditions were ideal: weak winds and wind waves well separated from a narrow-band
385 swell with significant wave height $H_s = 1.15$ m, peak period $T_p = 11$ s, and wave incidence
386 $D = 10^\circ$ from shore normal direction (measured from an Acoustic Doppler Current Profiler
387 moored in 10-m depth). The water was at mid neap tide level (low-tide terrace at about 1 m
388 depth), promoting a narrow surfzone less than 50 m wide. A Nortek high-frequency acoustic
389 Doppler velocimeter (ADV with sampling rate of 8 Hz) was deployed in the surfzone, in the
390 middle of the terrace, measuring currents about 0.5 m from the bottom. A dye release was
391 conducted to monitor the dispersion induced by flash rips, coupled with UAV flights (STB
392 DS6 hexacopter) at an elevation of 100 m. The Drone camera (NIKON D700) was looking
393 down, with a vertical angle, and recorded 4256×2832 -px scenes at 1 Hz (Almar et al., 2014;
394 Derian and Almar, 2017).

395 *4.1. Model Setup*

396 The domain is 542 m alongshore by 240 m across-shore, with periodic alongshore bound-
397 ary conditions. In order to prevent distortion when oblique waves are used with periodic
398 conditions, the alongshore size is adjusted according to peak wavelength and mean wave di-
399 rection. This method proved to perform well even with long-crested waves. The grid spacing
400 is $dx = dy = 0.5$ m. There are 10 vertical levels evenly spaced between the free surface and
401 bottom (a simulation with 20 levels did not change the results much). The model time step
402 is $dt = 0.01$ s. The bathymetry is longshore-uniform and built with continuous functions
403 to smoothly fit the low-tide terrace structure observed during the survey of (Almar et al.,
404 2014, 2018). The depth h is 8 m offshore to 1 mm at shore level, the position of which
405 varies with the swash oscillation (wetting-drying scheme of Warner et al. 2013). The depth
406 on the terrace is about 1 m, which corresponds to the mid-tide conditions of the afternoon

407 of March 13, 2014.

408 The wave maker is set with following parameters: $H_s=1.15$ m, $Tp=11$ s, $D = 0^\circ, 10^\circ$, and
409 directional spread $\sigma_\theta = 30^\circ$. A JONSWAP spectrum is constructed with these parameters
410 and a peak enhancement factor γ of 3.3. The WENO5-Z scheme is used again with the $k-\omega$
411 turbulence model. Bottom roughness is $z_0 = 0.01$ mm, which may seem low but gives a drag
412 coefficient $C_D \sim 0.002$ in the surfzone ($C_D = (\kappa/\log \frac{z_1}{z_0})^2$, with $\kappa = 0.41$ and $z_1 \sim H/10$
413 the first level height above bed), a usual value in depth-averaged models (Chen et al., 2003;
414 Feddersen et al., 2011). We follow the practice of these models here to reproduce their
415 results within the pseudo-2D approach described below. It is of little consequence for 3D
416 simulations because, as will be shown, they are much less sensitive to bottom drag than 2D
417 models (Sec. 5). The model is run for an hour starting from rest and adjusting through a
418 rapid spin-up phase. Figure 6 presents a snapshot of sea level that shows realistic features:
419 short-crested waves generated at an angle, refract and break, producing rollers, swash and
420 some reflection.

421 CROCO comes with capabilities for water quality, marine ecosystem and sediment mod-
422 eling. In the present study, they are used with simple settings. First, we introduce a passive
423 tracer in the swash zone for comparison with dye releases made during the beach survey.
424 Second, a suspended sediment model (Blaas et al., 2007; Warner et al., 2008) allows the rip
425 patterns to be compared with aerial photos taken during the survey. We use a single fine
426 sand class with settling velocity of 1 cm/s. For resuspension, taking one sediment bed layer
427 for simplicity, only two parameters are needed: critical shear stress and erosion rate at the
428 seafloor, expressed in the erosion flux (Blaas et al., 2007):

$$E = E_0(1 - p) \frac{\tau_b - \tau_{cr}}{\tau_{cr}} \text{ for } \tau_b > \tau_{cr} \quad (24)$$

429 τ_b is the bottom shear stress computed by the model. E_0 is an empirical erosion rate set
430 to 10^{-5} kg/m²/s; p is the sediment porosity (0.41); τ_{cr} is the critical shear stress, i.e., the
431 threshold for initiation of sediment motion, set to 0.01 N/m².

432 *4.2. Shallow vs. deep breaking (and a Boussinesq model)*

433 The Peregrine vorticity generation process only requires short crested waves, with no
434 need for unstable longshore currents generated by oblique waves. Boussinesq models are very
435 efficient in this process, but an important question for us is whether a 3D model will remain
436 so. An essential difference between the two types of model is the depth penetration of wave
437 breaking. In a 2D Boussinesq model, deep breaking is implicitly assumed as momentum is
438 transferred instantaneously to the depth-averaged flow. However, this is a rough assumption
439 as the breaking-induced flow is produced essentially above trough level (where the onshore
440 flow is located), while turbulence generated at the surface spreads downward by diffusion
441 with a limited mixing length of 10 to 30 percent of the water depth (Svendsen, 1987; Cox
442 et al., 1994; Mocke, 2001; Longo et al., 2002). (Uchiyama et al., 2010) show that the deep
443 breaking assumption is inconsistent with the cross-shore velocity profiles measured during
444 Duck94. Splinter and Slinn (2003) also suggest that transient dynamics produced by deep
445 breaking may collapse in the more realistic case of shallow breaking. The results presented
446 here are in agreement with these studies and we propose to assess the role of 3D dynamics on
447 surf eddies by comparing simulations forced by shallow and deep breaking. Deep breaking
448 will constitute a pseudo-2D model whose results can be compared with a Boussinesq model
449 solution (Section 4.5).

450 Boussinesq-type models (see Barthelemy 2004, for a review) are common tools to simulate
451 weakly dispersive waves and their transformations from the ocean to the swash zone. Several
452 developments allowed their application to a wide range of scales, from surfzone processes to
453 ocean basin-scale tsunami propagation (Kirby, 2016). Here, for a verification of our pseudo-
454 2D CROCO version, we use FUNWAVE-TVD (Shi et al., 2012). It solves the fully nonlinear
455 Boussinesq equations using a hybrid finite-volume finite-difference scheme. Parametrizations
456 are similar to CROCO, with wave breaking handled by a shock capturing TVD scheme
457 (making the need of explicit criterion unnecessary) and a quadratic drag formulation with
458 $Cd = 0.002$ for bottom friction. Note that newer generation models than FUNWAVE are
459 available with better dispersive properties — using Green-Naghdi equations (Lannes and
460 Bonneton, 2009) or incorporating an additional enstrophy equation (Kazakova and Richard,

461 2019) — but FUNWAVE is widely used and share with this class of models the essential
462 depth-averaged assumption.

463 To force deep breaking in a 3D wave-resolving model, we can artificially enforce strong,
464 deep vertical mixing around the breaker zone by multiplying the vertical eddy viscosity by
465 10 (from values of $\nu_t \sim 0.01-0.05 \text{ m}^2/\text{s}$). Figure 7 shows a time and longshore average
466 of cross-shore and alongshore currents in the case of shallow and deep breaking. Shallow
467 breaking is computed by the 3D model with no explicit constrain on penetration scale (but
468 a parametrization of eddy viscosity induced by breakers and currents). It drives a shallow
469 onshore flow about 40 cm deep and an offshore near-bottom undertow, resulting in strong
470 vertical shear of about 1 s^{-1} . With artificially strong vertical viscosity, momentum in the
471 breaker zone is almost instantly mixed to the bottom and the cross-shore flow is reduced to
472 the part required by mass-conservation (the anti-Stokes compensation flow), and consistent
473 with depth-averaged models. The longshore flow driven by oblique waves and a number of
474 other simulations with shallow and deep breaking (Tab. 2) will be analyzed and compared
475 in the following sections.

476 4.3. Reference simulations and comparison with data

477 To introduce the 3D processes of flash rip generation, we present simulations with shal-
478 low and deep breaking (3D_SC.D10 and 2D_SC.D10), representing mid-tides conditions on
479 March 13 2014 in Grand Popo Beach. Figure 8 compares the vertical vorticity of surface
480 flow: $\omega_z = \frac{\partial v_s}{\partial x} - \frac{\partial u_s}{\partial y}$, phase-averaged over two peak periods (22 s). With deep breaking (left
481 panel), the vortical field is rich with large filaments and surf eddies of 50-100 m scale that
482 are generated from short-crested waves, similar to Boussinesq model solutions. However,
483 the full 3D model with shallow breaking offers a radically new solution (Fig. 8, right panel).
484 Some of the large-scale fluctuations are present but over-shadowed by shorter scales. This
485 mode presents itself as rib structures (or mini rips following a relevant observation by Short
486 et al. 1993, already mentioned) with short longshore wavelength of about 5 m and period
487 about 1 min.

488 We now compare the two simulations with data collected during the survey of March

489 2014. Flash rips did not appear particularly large to the survey team at high or mid tide,
490 but a higher frequency signal was present and interpreted as swash rips (Castelle et al., 2014;
491 Scott et al., 2018; Floc'h et al., 2018). Dye experiments revealed filament generation, but
492 of relatively short scales. Although the survey was not extensive enough to draw definite
493 conclusions, the ADV data as well as the released dye and suspended sediment patterns
494 suggest a dynamical regime closer to 3D shallow-breaking than pseudo-2D deep-breaking
495 simulations, as will be seen.

496 The dye experiment presented in Figure 9 illustrates both the structure of the alongshore
497 flow and scales of flash rips emerging from the surfzone. Careful analysis of observed versus
498 modeled dye evolution, as in Hally-Rosendahl and Feddersen (2016), is beyond the scope of
499 this paper, but useful information can be gained from a simpler analysis. Figure 9 presents
500 two consecutive aerial photos (at 116s interval), and the corresponding snapshots of tracer
501 simulations with the full 3D model. We do not expect an exact match between observed and
502 modeled rips, considering the chaotic nature of these phenomena³, but scales and structures
503 are meaningful. In both cases, a thin filament of about 5 m expands quickly seaward at a
504 speed of about 0.5 m/s, reaching about 70 m from shore. Using sequential photos of the
505 tracer from the Drone camera, it is also possible to extract a simplified cross-shore profile
506 of longshore drift velocities (see Derian and Almar 2017 for more extensive Lagrangian
507 calculations). The result is presented in Figure 10 together with the ADV measurement
508 of mean longshore current over the terrace and model solutions with deep and shallow
509 breaking. The estimated longshore flow has an asymmetric V-shape similar to the full 3D
510 model solution with a peak velocity of about 0.5 m/s in the inner surfzone, and error bar of
511 about 0.1 m/s (Derian and Almar, 2017). Deep breaking solutions have a more symmetric
512 profile centered in the outer surfzone. The profiles in both simulations are a result of cross-

³During the survey, a few attempts of dye release were made before obtaining a clear filament patch. Similarly for the model, we selected one occurrence among few tracer patches initialized at regular interval along the coast (in the swash zone). Even though all tracer patches eventually ended up with similar V-shape and similar scales of evolution, there was variability in the evolution and we selected the most visually comparable filament with Drone photos.

513 shore advection. With deep breaking, advection is weak and the longshore flow remains
514 centered in the breaker zone (Fig. 7). This is a common bias of depth-averaged models
515 (Larson et al., 2002).

516 Another qualitative comparison of patterns can be made, looking at surfzone suspended
517 sediments in the aerial photo (Fig. 11, left panel). The contrasts in the photo is enhanced
518 to better expose suspended sediments (brown color), which is seen weakly extending beyond
519 the surfzone. Snapshots of the model's surface sediment concentration is also shown after
520 15 min of simulation. With shallow breaking (center panel), sediments tend to resuspend
521 in the breaker zone and mix efficiently within the surf zone, but only weakly extend to the
522 innershelf. The rib structure is apparent at the seaward front of sediment concentration. It
523 is also apparent in white streaks representing alongshore surface current convergence, that
524 have a structure similar to the foam lines in the aerial photo. Overall, the patterns are
525 similar to the observations, particularly in the upper part of the photo, where there is less
526 foam or sunglint. The same suspended sediment simulation with deep breaking gives very
527 different results (Fig. 11, right panel). Resuspension is now maximum in the inner surfzone
528 (as for eddy energy; Sec. 4.5). Filaments and eddies are more coherent, larger, their growth
529 slower, but extend further seaward. Mixing in the surfzone is less efficient than for the
530 shallow-breaking case, but shelf-surf exchange is more intense, due to filament extension.

531 For a more quantitative local comparison, we now turn to ADV measurements of hor-
532 izontal velocities $u_h = \sqrt{u^2 + v^2}$. Figure 12 presents the Power Spectral Density (PSD) of
533 velocity fluctuations, using Welch's noise reduction method, for the model and ADV data
534 in the middle of the terrace (left panel) and for the model alone over the outer terrace slope
535 (right panel). The short-wave spectrum around the peak period (11 s) is well represented
536 given the JONSWAP approximation made for the model wave maker. At a lower frequency,
537 a good fit with the data is also given by the full 3D solution, while the deep-breaking simu-
538 lation exhibits two opposite biases in successive frequency ranges (valid at 95% confidence
539 level), which are even more pronounced near the terrace slope. These biases are consistent
540 with those noted in Feddersen et al. (2011). The first is an underestimation of energy by
541 the pseudo-2D model in the 30-100s period range. This band is consistent with visual in-

542 spection of rip structure oscillation in animated vorticity fields. The 3D model seems to
543 correct the deficit, particularly in the lower frequencies. Note that this energy range for 3D
544 eddies overlaps that of infragravity waves, making it difficult to separate the two phenom-
545 ena from observations alone. At very low frequency (VLF), for timescales between 2 and
546 15 min, a second bias of opposite sign is observed in the pseudo-2D model solution. In this
547 range, the eddies produced by short-crested waves have more energy in the deep-breaking
548 case than in the 3D case, as predicted by inspection of vorticity and suspended sediment.
549 The comparison with the data therefore suggests that the VLF energy is overestimated by
550 depth-averaged models, consistent with Feddersen et al. (2011) and Newberger and Allen
551 (2007).

552 We conclude from this section that the observations at Grand Popo Beach are in better
553 agreement with a complete 3D solution of surf eddies that includes the presence of 3D rib
554 structures. Next, we analyze their generation process.

555 4.4. *Structure and production of vertical shear instability*

556 The surfzone eddy variability seen in the 3D model solution is truly three-dimensional.
557 Vertical vorticity is only one manifestation, but horizontal vorticity is the main player. Fig-
558 ure 13 presents the Q field defined by $Q = -\frac{1}{2} \frac{\partial u_i}{\partial x_j} \frac{\partial u_j}{\partial x_i}$, using Einstein summation convention
559 over the three dimensions. Q is commonly used to enhance detection of vortical flows.
560 Here, we split cross-shore and alongshore Q components ($Q_y = -\frac{\partial u}{\partial z} \frac{\partial w}{\partial x} - \frac{1}{2} \frac{\partial u^2}{\partial x}$ in red and
561 $Q_x = -\frac{\partial v}{\partial z} \frac{\partial w}{\partial y} - \frac{1}{2} \frac{\partial v^2}{\partial y}$ in green), then normalize them and only plot positive isosurface values
562 0.02 for clarity (negative values give counter-rotating features).

563 The result is strikingly consistent with instabilities of a transitional mixing layer (Met-
564 calfe et al., 1987; Lesieur, 1990), the transition being constrained by surfzone width. Q_y
565 shows spanwise "rolls" created from the primary instability while Q_x identify streamwise
566 "ribs" that are transverse (counter-rotating) vortices from secondary instability (assumingly
567 growing from perturbations generated between the rolls, in the braid region). Note that
568 streamwise designate the shear direction, which is cross-shore, rather than the oblique wave
569 direction. The reason is that the bottom flow aligns with the surface flow in the alongshore

570 direction, forming only a weak mean vertical shear (Fig. 7). However, the rips can take an
 571 oblique direction when advected by the mean longshore flow. They extend seaward beyond
 572 the surfzone while stretching in the vertical direction. Filaments of vertical vorticity (or Q_z ,
 573 not shown) also have similar rib structures to Q_x , recognisable in the surface vorticity plot
 574 of Figure 8.

575 The mean shear flow is composed of the wave-mean surface onshore flow and associated
 576 seaward undertow (Fig. 7). The inflected velocity profile is inviscidly unstable to small per-
 577 turbations and unstable modes of Kelvin-Helmholtz type can emerge. The spanwise rolls are
 578 large vertical eddies that rapidly evolve into transverse (streamwise) rib vortices connected
 579 by “braid” regions, and stretched seaward and downward. This picture is reminiscent of de-
 580 scriptions based on measurements and simulations at the laboratory scale (Nadaoka et al.,
 581 1989; Watanabe et al., 2005; Lubin and Glockner, 2015) but, here, the instabilities are gen-
 582 erated by the wave-mean shear flow rather than by direct breaking — breaking-induced
 583 turbulence is parametrized. It is therefore more consistent with the instability of the under-
 584 tow profile described by Li and Dalrymple (1998).

585 According to linear stability analyses, the wavelength of primary shear instability setting
 586 the distance between spanwise rolls is an order of magnitude larger than the mixing layer
 587 width δ — in free shear layers (Michalke, 1964, 1965), wavelength, frequency and growth
 588 rate of the most unstable modes are 14δ , $0.015 U/\delta$ and $0.1 U/\delta$ respectively. The sec-
 589 ondary instability wavelength is of the same order as that of the primary instability ($2/3$
 590 in Pierrehumbert and Widnall 1982). If the mixing layer width is taken as the vorticity
 591 thickness $\delta = \Delta U / [\frac{\partial u}{\partial z}]_{max} \sim 50$ cm, then the wavelength of both roll and rib structures is
 592 about 5 m, consistent with our simulation. Note that given an effective resolution of 5-10
 593 dx for CROCO (Soufflet et al., 2016), a grid resolution of 0.5 m can be considered “eddy
 594 resolving” for 3D instabilities under current conditions. As for frequency, the mixing layer
 595 size would be associated with modes around 0.015 Hz (60 s period), a range usually reserved
 596 to surf beat. The model with 3D instability has energy in this range that is lacking in
 597 the deep breaking case (see previous section and Fig. 12), but we will see that nonlinear
 598 interactions can spread this energy around the injection scale (Sec. 4.6). Note that the

599 ribs sometimes develop localized pairing, which may be evidence of subharmonic resonance,
600 another instability associated with shear layers (Craik, 1971; Pierrehumbert and Widnall,
601 1982; Herbert, 1983).

602 In order to isolate the mechanism of eddy-mean flow interaction, we analyze a solution
603 forced with monochromatic, shore-normal, long-crested waves (3D_MONO_D0 in Tab. 2).
604 This simulation has constant wave forcing in both space and time when averaging over
605 the wave period (11 s). In this case, the same rib structure is generated (Fig. 14) but
606 without the large-scale alongshore variation seen in the full solution (Fig. 8, right panel).
607 A comparison of this simulation with other test cases will be presented in the next section.
608 Here, we analyse the mechanism and patterns of shear production. The EKE source terms
609 $-\overline{u'_i u'_j \frac{\partial \bar{u}_i}{\partial x_j}}$ represent the energy spent by the mean flow to feed the instability leading to rolls
610 and ribs. The largest of these terms is the vertical shear production $-\overline{u' w' \frac{\partial \bar{u}}{\partial z}}$ (Metcalf et al.,
611 1987), which is shown in Fig. 15. There is a lesser contribution from cross-shore convergence
612 $-\overline{u' u' \frac{\partial \bar{u}}{\partial x}}$ (all other 7 combinations are negligible). The main site of shear production is in
613 the breaker zone (Fig. 15) and the maximum values are located at the inflection point
614 in the mean velocity profile (represented by a magenta line), in agreement with mixing
615 layer instability theory. The figure also shows a cross-section of mean eddy kinetic energy
616 ($EKE = \frac{1}{2}(u'^2 + v'^2 + w'^2)$, where u', v', w' are fluctuations of phase-averaged velocities,
617 with respect to the time-mean flow presented in Fig. 7). EKE and shear production have a
618 similar spatial pattern, although high EKE values extend from the production center in all
619 directions. The primary rolls are thus produced in the outer surfzone but turbulent energy
620 is diffused by the mean and eddy flow across the water depth and towards both the inner
621 surfzone and innershelf. EKE transport is stronger at the surface and streamwise filaments
622 extend farther offshore at the surface than at the bottom (despite some amount of vertical
623 stretching as they leave the terrace).

624 For comparison, the mean subgrid-scale turbulent kinetic energy (TKE) produced by
625 the k - ω closure equations is also shown in Figure 15 (bottom). EKE and TKE have a
626 similar structure, while EKE amplitude is about a quarter of that of TKE. The mean shear
627 turbulence intensity is thus a significant part of total 3D turbulence generated by breaking

628 waves. We expect that part of TKE (parametrized by the closure model) could be transferred
629 to EKE if a higher resolution was used.

630 *4.5. Short-crested waves and the Peregrine process*

631 An important question of our study concerns the effect of wave variations (frequency and
632 directional spreading) on flash rip generation in a full 3D model. To address this question, it
633 is useful to simplify the problem and progressively add the multiple conditions of variability.
634 In this section, we analyse shore-normal, short-crested wave simulations with shallow or deep
635 breaking (3D_SC_D0 or 2D_SC_D0). Shore-normal conditions prevents the formation of a
636 longshore current and associated horizontal shear instability. In addition, we look at long-
637 crested wave solutions to isolate the effect of 3D instabilities, i.e., the monochromatic solution
638 (3D_MONO_D0), presented in the previous section, and a similar case with JONSWAP
639 frequency spectrum (3D_LC_D0). This latter comparison will help evaluate the effect of
640 frequency spreading on eddy variability before addressing the effect of directional spreading.

641 Figure 16 compares vertical vorticity for shore-normal, short-crested wave cases. As for
642 oblique waves, deep-breaking leads to a rich vortical field with large filaments extending
643 far offshore, similarly to Boussinesq models. This is confirmed here with a comparison
644 between FUNWAVE-TVD and pseudo-2D CROCO applied to the same configuration. A
645 difference between Boussinesq and pseudo-2D solutions is the effect of 3D dynamics over the
646 innershelf in the latter case, where surface-intensified offshore eddies and filaments present a
647 more fragmented aspect due to a forward energy cascade (Uchiyama et al., 2017; McWilliams
648 et al., 2018). However, the full 3D nonhydrostatic model with shallow breaking (Fig. 16,
649 right panel) shows again different patterns from both Boussinesq and pseudo-2D solutions
650 with regular rib structures having a shorter alongshore scale and a more limited cross-shore
651 extension. Figure 17 presents EKE cross-sections for all shore-normal wave experiments.
652 A striking element of these figures is the presence of large surface and bottom EKE in the
653 shallow breaking cases. This pattern is not a result of shear production because it is absent
654 from the monochromatic case (see previous section and Fig. 15). Therefore, it can only result
655 from wave groups associated with frequency spreading. Through wave height modulation,

656 wave groups produce variability in the surface onshore flow and associated undertow. The
657 variability amounts to about half the integrated mean EKE. However, it is much smaller
658 in the deep-breaking case, consistent with depth-averaged model results (de Schipper et al.,
659 2014).

660 We now turn to the effect of directional spreading. In Figure 17, short-crested waves
661 (produced by directional spreading) extend EKE production over a wider surf zone than long-
662 crested waves, where EKE is confined to the breaker zone. However, the seaward extension is
663 significantly larger in the pseudo-2D model, confirming the impression made from vorticity
664 inspection. Further confirmation is given by vertical EKE integration (and normalization
665 by mean depth; Fig. 18). It highlights 3 distinct regions: the inner and outer surfzone and
666 innershelf. The top panel presents unfiltered data. In this case, 3D instability and wave
667 group forcing dominates eddy production in the surfzone. The deep-breaking solution has a
668 larger cross-shore expansion, extending out to the innershelf, where EKE levels are twice as
669 high as in the shallow breaking case. This is even clearer using a low-pass filter on velocity
670 fluctuations, removing a large part of variability from 3D instability and wave group forcing
671 (bottom panel of Fig. 18). What remains is closer to the usual definition of surfzone eddies
672 as Very Low Frequency features. There is now a maximum in the inner surfzone, consistent
673 with findings from previous Boussinesq model studies that filaments forced by short-crested
674 waves originate in the inner surfzone, then spread offshore, forming eddies that grow in
675 scale (Johnson and Pattiaratchi, 2006). However, the difference of EKE profiles between
676 deep and shallow breaking cases is reminiscent of the overestimation of shelf-surf exchange
677 by Boussinesq models (see profiles of dye concentration in Fig. 10 of Hally-Rosendahl and
678 Feddersen 2016).

679 *4.6. 2D and 3D surfzone turbulent cascade*

680 If there is energy produced by short-crested waves in the inner surfzone of the shallow
681 breaking case, the question is why does it produce fewer large filaments than the deep break-
682 ing case. We found an answer in computing the spectrum energy flux. In 2D turbulence, the
683 flux of energy is negative and small fluctuations can grow into larger coherent structures.

684 This process is usually involved to explain the growth of filaments and eddies from variable
685 wave forcing (e.g., Johnson and Pattiaratchi 2006; Feddersen 2014). An inverse cascade in
686 the surfzone has recently been confirmed by observations (Elgar and Raubenheimer, 2020),
687 but how efficient is it exactly?

688 To answer this question, we performed a wavenumber spectral flux analysis for pseudo-
689 2D and 3D simulations 2D_SC_D0 and 3D_SC_D0 (Fig. 19). The spectral flux is computed
690 as in Marchesiello et al. (2011) by spectral integration of v advection term. Consistently
691 with 2D turbulence there is a strong inverse cascade of kinetic energy (negative flux) in the
692 pseudo-2D model starting from the scale of injection corresponding to short-crested wave
693 forcing (wavelength of ~ 30 m here), and there is no direct cascade towards smaller scales.
694 In the 3D case, the turbulent regime is different. The negative flux of energy produced
695 by variable wave forcing is present but significantly reduced. In addition, there is a second
696 injection at smaller scales that corresponds to the most unstable mode of 3D instability
697 (wavelength of about 5 m). This small-scale energy travels both backward and forward
698 across the spectrum and thus widens the range of variability associated with 3D instability.
699 In the 3D long-crested wave case 3D_LC_D0, a similar spectral flux is produced at small
700 scales but there is no large-scale inverse cascade due to missing injection by short-crested
701 waves.

702 This analysis confirms that the growth of filaments and eddies associated with the Pere-
703 grine process heavily relies on a 2D inverse cascade, but this cascade is impaired by 3D
704 dynamics. In this case, vorticity fluctuations generated by waves with finite alongshore
705 extent produce less coherent horizontal structures than in the 2D paradigm built on depth-
706 averaged models (Johnson and Pattiaratchi, 2006). The extent of this inhibition process
707 depends on the intensity of turbulent mixing. Sensitivity analysis using reduced viscosity
708 shows that lower viscosity leads to a reduction of the 2D inverse cascade at VLF, but to
709 an increase of energy production and fluxes at smaller scales. The opposite is true when
710 increasing viscosity. Therefore, breaking waves can transfer energy into both 2D and 3D
711 transient circulations but the distribution of energy between the two regimes is regulated
712 by turbulent mixing which occurs at a higher frequency. These regimes are not mutually

713 exclusive, at least within some range of turbulence intensity, the realism of which still needs
714 to be better evaluated.

715 *4.7. Oblique waves and horizontal shear instability*

716 The precedent experiments with shore-normal waves show that shallow breaking tends
717 to hinder the generation of large 2D eddies by short-crested waves, while sustaining forced
718 and intrinsic 3D surfzone eddy dynamics. We now address the case of oblique waves with
719 direction $D = 10^\circ$. From linear stability analysis of a 2D problem (Bowen and Holman,
720 1989), the wavelength, frequency and growth rate of the most unstable shear waves are
721 $\lambda_h = 2.5L$ and $f_h = 0.07\frac{V}{L}$, $\sigma_h = 0.15\frac{V}{L}$ where V is the longshore current magnitude
722 and L is the longshore current half-width (outer shear). For a narrow, shoreline-intensified
723 jet typical of Grand Popo at mid-tide (Almar et al., 2014, 2015b), shear can be strong
724 ($\sim 0.05 \text{ s}^{-1}$), implying a minimum shear wave period of 5 min, wavelength of 80 m, and
725 growth time $\sigma_h^{-1} = 3 \text{ min}$.

726 When forcing long-crested waves ($\sigma_\theta = 0$) with deep breaking (2D_LC_D10), CROCO
727 recovers results that are typical of wave-averaged shallow-water models (or 3D models with
728 deep breaking as in Kumar and Feddersen 2017). Figure 20 (left panel) shows an active
729 horizontal shear instability producing shear waves with wavelength consistent with the lin-
730 ear theory. Shear waves propagate with the longshore current as they become nonlinearly
731 unstable, generating filaments and eddies that extend offshore. When both horizontal shear
732 instability and short-crested wave vorticity generation are active (with deep breaking), eddies
733 and filaments are more prominent (3rd panel of Fig. 20). However, with shallow breaking,
734 the horizontal shear instability appears weaker and is again replaced by rib structures. Both
735 horizontal processes are thus reduced by the vertical shear.

736 Inspection of the cross-shore profile of mean longshore currents (Fig. 7) may help to
737 understand how shallow breaking undermines horizontal shear instability. With deep break-
738 ing, cross-shore advection is inactive and the longshore current remains trapped over the
739 terrace slope (which is steep in Grand Popo, around 1/10), and its outer shear is strong.
740 With shallow breaking, however, the longshore current is advected by the cross-shore circu-

741 lation, stretching its profile across the terrace so as to minimize the outer shear. Then, the
742 instability growth rate becomes too weak to overcome friction (from turbulence or bottom
743 drag). This may explain, for example, the puzzled observation by Newberger and Allen
744 (2007) that their 3D wave-averaged model produced no horizontal shear instability, unlike
745 many previous 2D modeling studies (Allen et al., 1996; Slinn et al., 1998; Uchiyama et al.,
746 2009).

747 This inhibitory process is verified in Figure 21, showing cross-shore profiles of the mean
748 and eddy flow averaged over time and the alongshore direction for the four cases of oblique
749 waves. Horizontal shear instability is best assessed with the term for horizontal shear pro-
750 duction of turbulent kinetic energy: $-\overline{u'v'}\frac{\partial \bar{v}}{\partial x}$. In all cases, the shear production is clearly
751 correlated with the outer and inner slopes of the mean longshore current, with higher produc-
752 tion in the outer shear. Deep-breaking cases show higher shear production and greater EKE
753 centered on the outer slope of the terrace, where the shear is greatest. In the case of horizon-
754 tal shear instability alone (2D_LC_D10), the magnitude is lower than that of 2D_SC_D10,
755 despite similar shear intensity. This indicates an amplification of shear instability by short
756 crested waves as they drive transient intensification of longshore currents. In this case also,
757 The EKE maximum extends farther offshore than expected from shear production, possibly
758 due to mean and eddy advection, but short-crested waves provide the most efficient process
759 for innershelf eddy activity.

760 The shallow-breaking cases (3D_LC_D10 and 3D_SC_D10) also features horizontal shear
761 production, but weaker and in shallower water, where friction is more prevalent. As a
762 result, EKE is significantly reduced. Short-crested waves (3D_SC_D10) appear to amplify
763 the inner surfzone energy (compared with 3D_LC_D10), but in both cases the offshore energy
764 is considerably reduced.

765 5. Discussion and conclusion

766 Flash rips and surfzone eddies are traditionally conceived within a depth-averaged frame-
767 work that involves intrinsic horizontal shear instabilities or/and direct short-crested wave

768 vorticity generation. They are revisited in this study using a 3D nonhydrostatic wave-
769 resolving model applied to a natural beach with ideal longshore-uniform topography (Grand
770 Popo Beach, Benin). We first presented a quick overview of a new free-surface, compressible
771 approach adapted to wave-resolved nearshore dynamics. Its ability to simulate surface grav-
772 ity wave propagation, nearshore breaking and the resulting circulation is validated against
773 small- and large-scale laboratory experiments. Then, the model is applied to the neashore
774 circulation generated at Grand Popo Beach by waves with frequency and directional spread-
775 ing. We assume on the basis of the comparison with Boussinesq solutions that the essential
776 difference between 2D and 3D models is reduced to the vertical profile of breaking-induced
777 acceleration, i.e., deep or shallow breaking. This allows a direct comparison of 2D and 3D
778 frameworks within the same model equations and setup.

779 The generation of transient ribs by the 3D model is shown to differ from that produced
780 by depth-averaged models, owing to the vertical structure of currents produced by surface-
781 intensified acceleration. Processes of both horizontal shear instability and short-crested wave
782 breaking are limited in our 3D model by the cross-shore vertical recirculation, which can
783 restrict an otherwise strong inverse cascade. Variable wave forcing (in space and time) tends
784 to increase flow variability in the surfzone (especially at the surface and bottom), but it does
785 not fully translate into large-scale ribs streaming far offshore. Usual 2D mechanisms may
786 thus be weaker than expected but complemented by a Kelvin-Helmholtz-type instability
787 generated at the inflection point of the mean vertical shear flow. The latter generates rib
788 structures with spanwise and streamwise (alongshore and cross-shore) vorticity of intermedi-
789 ate scale between turbulence and large horizontal eddies — timescale several times the wave
790 period encroaching on the infragravity wave range and wavelength around 5 m. Streamwise
791 filaments extend beyond the surfzone but with lower intensity than usual VLF ribs (the
792 offshore mean EKE can be halved). Comforted by observed energy spectra and patterns of
793 tracer and sediment concentrations at Grand Popo Beach, our study may call into question
794 the accuracy of nearshore depth-averaged models. It may explain, in particular, the evidence
795 of overestimation by these models of the shelf-surf exchange (Spydell and Feddersen, 2009;
796 Hally-Rosendahl and Feddersen, 2016) or VLF variability (Feddersen et al., 2011).

797 Our results are representative of mid-tide conditions of a low tide terrace with moderate
798 wave heights. Interestingly, the rib structures that are described here are comparable to
799 the *mini rips* described for similar conditions by Short et al. (1993): *Under typical mid-*
800 *tide conditions, with waves breaking across the bar, a low 'friendly' surf zone is produced.*
801 *Waves are less than 1 m and most water appears to head toward the shore. In fact it is*
802 *return seaward also, both by reflection of the beach face and via the mini rips, even if no rip*
803 *channel are present. The rips, however, are usually weak, ephemeral and shallow.*

804 In order to assess the ubiquity of mini rips in the nearshore zone, future 3D studies
805 should explore different nearshore conditions. Several sensitivity tests were performed in
806 this study, which we only briefly report, pending further exploration of the model parameter
807 space in the future. Nevertheless, they provide useful material for discussion. We first
808 tested the effect of wave amplitude as it affects the breaking-induced flow and turbulence. A
809 simulation forced with twice as large waves (H_s from 1.15 to 2.30 m) did not fundamentally
810 change the results. In case of higher waves, breaking-induced turbulence can reach deeper
811 depths but the cross-shore flow acceleration is also stronger, so the result on vertical shear
812 is uncertain. Next, we tested a different beach profile, from the steep slope of Grand Popo
813 to a more gentle slope (smaller Iribarren number) similar to Duck Beach (as in Noyes et al.
814 2005). Here again, the results were similar, as the intensity of the mean vertical shear does
815 not appear to be too sensitive to the beach slope. The bottom roughness length z_0 was
816 another relevant parameter. As z_0 is increased from 0.01 to 1 mm, the drag coefficient goes
817 from about 0.002 to 0.008. Simulations with deep breaking were very sensitive to these values
818 and the largest roughness value can completely shutdown horizontal shear instability due to
819 shorter frictional time (decreased from 5 to 2 min, i.e., shorter than growth time of about
820 3 min), and also damp vortical generation by short-crested waves. On the contrary, full 3D
821 solutions with shallow breaking are only weakly sensitive to bottom roughness, because of
822 surface intensified currents and shorter growth time of 3D instability. These tests inspire
823 greater confidence in our results, while highlighting the overestimated importance given to
824 bottom drag in studies using depth-averaged models (e.g., Allen et al. 1996).

825 Eddy viscosity provides the largest source of sensitivity in our results. Artificially chang-

826 ing the eddy viscosity coefficient lead to qualitatively similar results but with significant
827 variation in the intensity of processes. An increase of eddy viscosity in the surfzone reduces
828 the vertical shear, damps vertical shear instability — as the frictional time h^2/ν_t becomes
829 closer to the growth time of about 1 minute — and intensifies the inverse turbulent cascade.
830 In the same time, cross-shore advection is reduced so that the wave-averaged longshore
831 flow is sharper and in deeper waters, therefore more sensitive to horizontal shear instability.
832 Therefore, there is a relationship between the intensity of turbulence and sorting of energy
833 that enters 2D and 3D dynamical regimes. Given the present model uncertainty, we can
834 only infer that these regimes are not mutually exclusive and should coexist. More validation
835 in different settings will be needed to assess their relative importance.

836 Finally, the present modeling study shed light on the mechanistic process that could
837 inhibit the generation of surf eddies by short-crested waves. Vertical vorticity generation by
838 the Peregrine process can be written as:

$$\frac{\partial \omega_z}{\partial t} = -\frac{\partial F_{br}}{\partial y_c} \quad (25)$$

839 where F_{br} is the breaking force, extending to the bottom in a depth-averaged model, and y_c
840 is the along-crest direction. Using a parametrization for breaker acceleration, Clark et al.
841 (2012) propose a scaling relation for vorticity generation of a single wave as $H_s^3 h^{-2.5}$, at a
842 maximum in the outer surfzone. However, flash rip generation originates in the inner surfzone
843 in depth-averaged models (Johnson and Pattiaratchi, 2006) as well as in our simulations.
844 Therefore, a transient vorticity source in the outer surfzone is not sufficient to generate the
845 expected local horizontal recirculation (with offshore filament). It needs a coastal boundary
846 and an inverse energy cascade that transform vorticity fluctuations into larger-scale coherent
847 structures, as shown by our spectral flux analysis. In a 3D regime, this cascade can be
848 reduced (due to vortex tilting by the shear flow; McWilliams et al. 2018) and the variability
849 generated in the inner surfzone does not fully translate into large rips jetting offshore.

850 In conclusion, our results suggest that nearshore dynamics and transport processes may
851 be affected by nonhydrostatic dynamics, not only for surface gravity waves and small-scale
852 turbulence as is well known, but also for larger-scale vortical motions. We expect this

853 conclusion to be qualitatively valid in other applications, but further studies should explore
854 the range of parameters encountered in the global coastal ocean. Most importantly, they
855 should pay special attention to how these parameters affect the mean cross-shore current
856 profiles as a key to 3D transient dynamics.

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Table 1: Model setup of CROCO's Grand Popo Beach configuration

Model parameters	Values
Grid type	Horizontal: rectangular structured; vertical: free-surface, terrain-following
Domain size (L_x, L_y)	240 x 542 m
Horizontal resolution (dx/dy)	0.5 m
Vertical grid (Nz)	10 levels (dz=10 cm over the terrace)
Time step	dt=0.01 s, $c_{\text{sound}}=200$ m/s (pseudo-acoustic)
Bathymetry (h)	longshore-uniform low tide terrace derived from March 2014 survey
Tides	mid-tide: 1-m depth over the terrace
Wave forcing (u,v,w at offshore boundary)	linear wave solutions reconstructed from JONSWAP spectrum with $H_s = 1.15$ m and $T_p = 11$ s
Air-sea fluxes (u,v,T,S)	No
Sediment model ($D_{50}, w_s, E_0, \tau_{cr}$)	$D_{50} = 100 \mu\text{m}$; $w_s=5$ mm/s; $E_0 = 10^{-5}$ kg/m ² /s; $\tau_{cr} = 0.01$ N/m ²
Breaking/turbulence	WENO5-Z shock-capturing advection scheme $k-\omega$ (or $k-\epsilon$) turbulence model
Bottom roughness (z_0)	0.01-1 mm

Table 2: List of CROCO simulations for Grand Popo Beach.

Simulation	Breaking	Wave type	Direction	Test
3D_SC_D10	shallow	short-crested	10°	Data vs. 3D instab. vs. Peregrine process
2D_SC_D10	deep	short-crested	10°	
3D_SC_D0	shallow	short-crested	0°	3D instab. vs. Peregrine process
2D_SC_D0	deep	short-crested	0°	
3D_LC_D0	shallow	long-crested	0°	
3D_MONO_D0	shallow	monochromatic	0°	
3D_LC_D10	shallow	long-crested	10°	3D instab. vs. 2DH instab.
2D_LC_D10	deep	long-crested	10°	
FUNWAVE-TVD	deep	short-crested	0°	Test of pseudo-2D CROCO

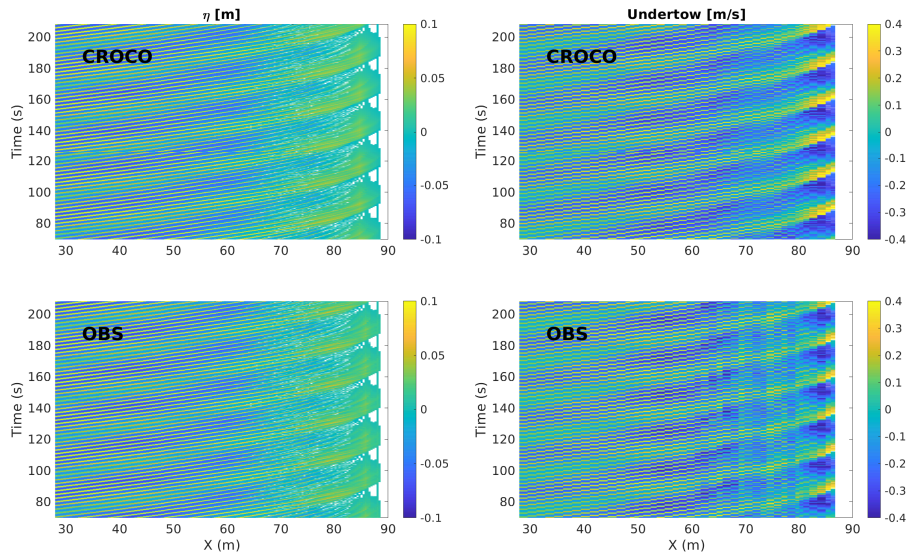


Figure 1: Hovmuller plot (x,time) of data and model sea level η and undertow u_b for the GLOBEX B3 flume experiment. When data is missing in the measurements, it is also removed from the model output.

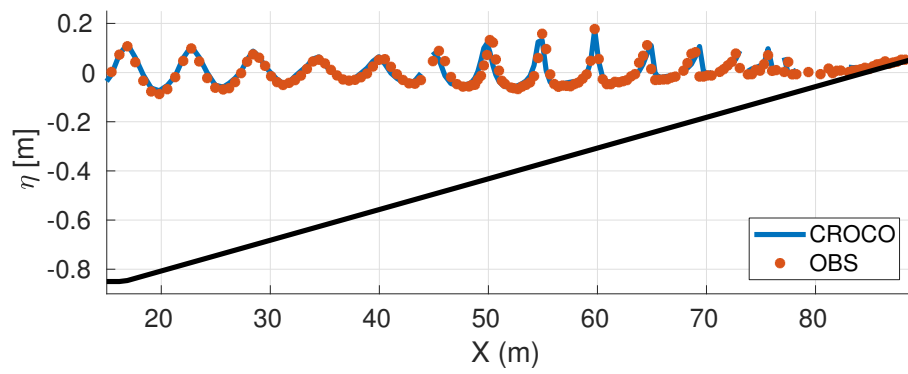


Figure 2: Snapshot of wave field across the GLOBEX B3 flume experiment during runup conditions for model (blue line) and data (red dots).

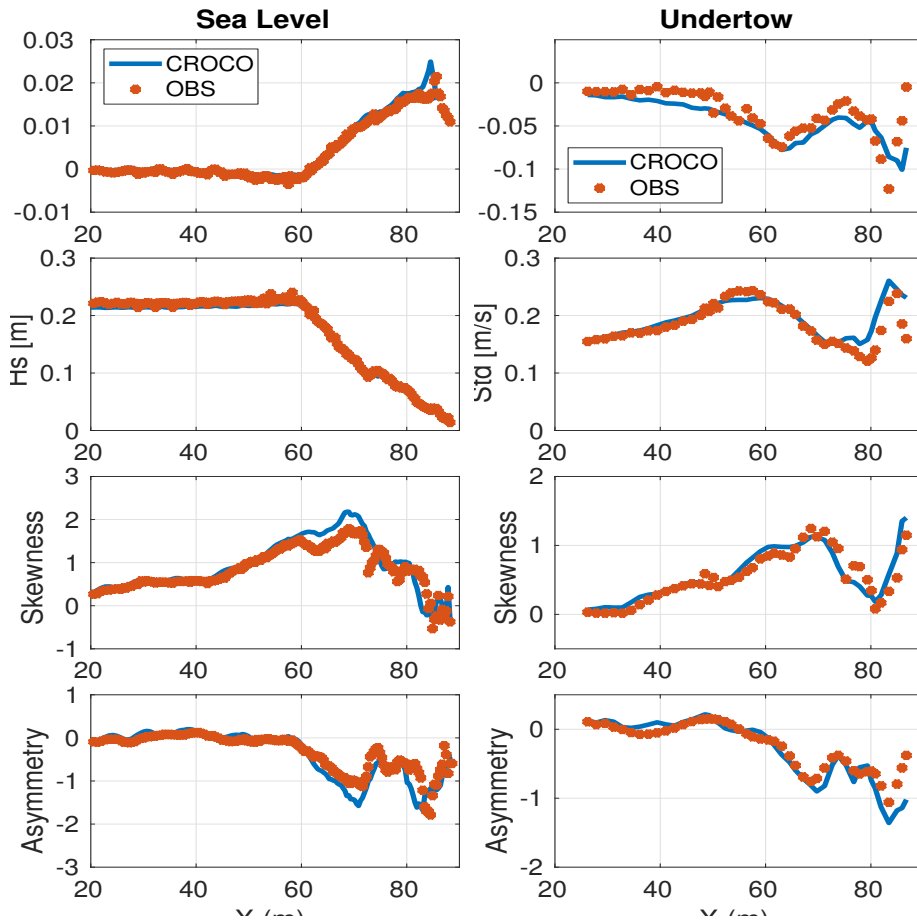


Figure 3: Wave statistics from GLOBEX B3 experiment for sea level η (right) and undertow u_b (left) in the model (blue line) and data (red line or dots). From top to bottom: mean, standard deviation (or H_s for η), skewness and asymmetry.

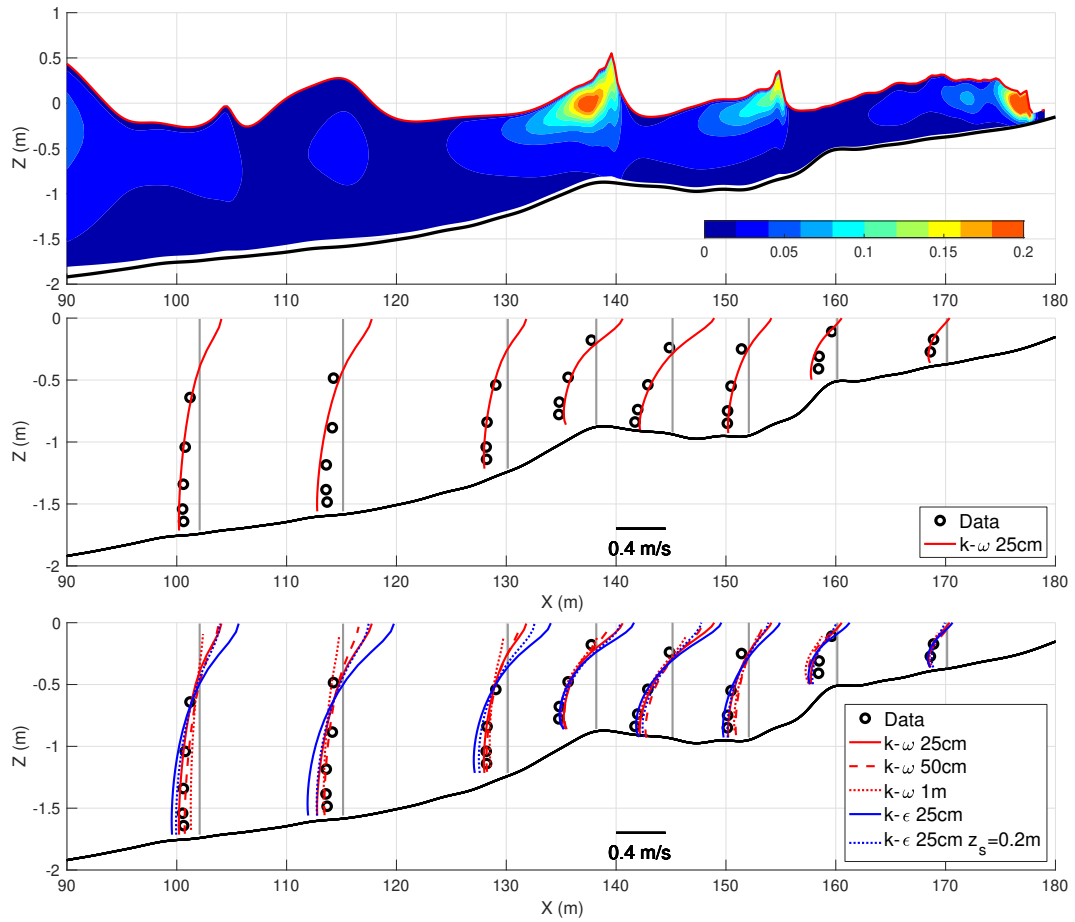


Figure 4: Model comparison with the large-scale LIP11-1B Flume experiment. Top) Snapshot of wave height and turbulent kinetic energy k from the reference model simulation (25 cm resolution; $k-\omega$ turbulence model). Middle) Comparison of simulated and measured current profiles. Bottom) Sensitivity to resolutions (25 cm, 50 cm and 1 m) and turbulence models ($k-\omega$ in red; $k-\epsilon$ in blue).

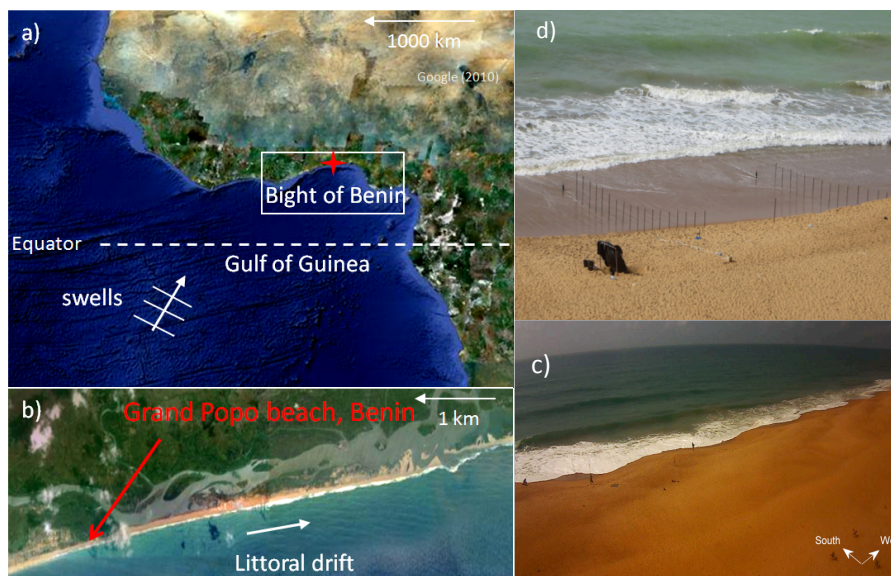


Figure 5: Grand Popo Beach (6.2°N , 1.7°E) is a longshore uniform beach located off Benin in the Gulf of Guinea (a-b-c). It is representative of an open, wave-dominated and microtidal environment exposed to S-SW long period swells generated at high latitudes in the South Atlantic. Panel (d) shows Grand Popo's low-tide terrace where instruments were deployed in March 2014 (the terrace is exposed here during low tide).

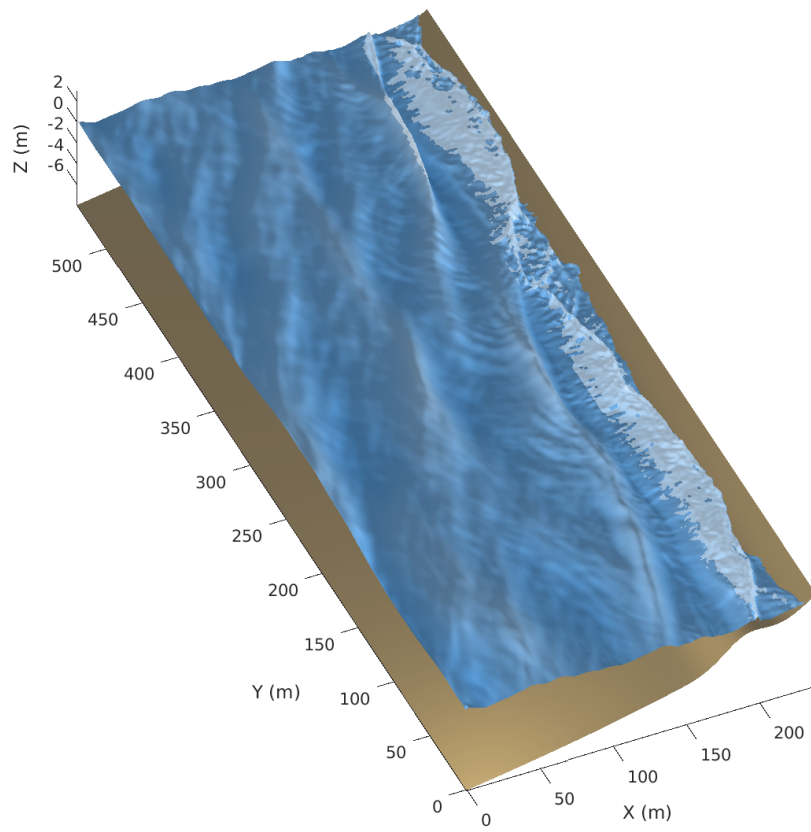


Figure 6: 3D representation of surface waves in CROCO. Short-crested waves are generated at the offshore boundary and propagate shoreward at an angle of 10° , with refraction and breaking through the surf and swash zones. Foam in the surfzone is approximated by white patches in locations of high turbulent kinetic energy of the subgrid-scale model.

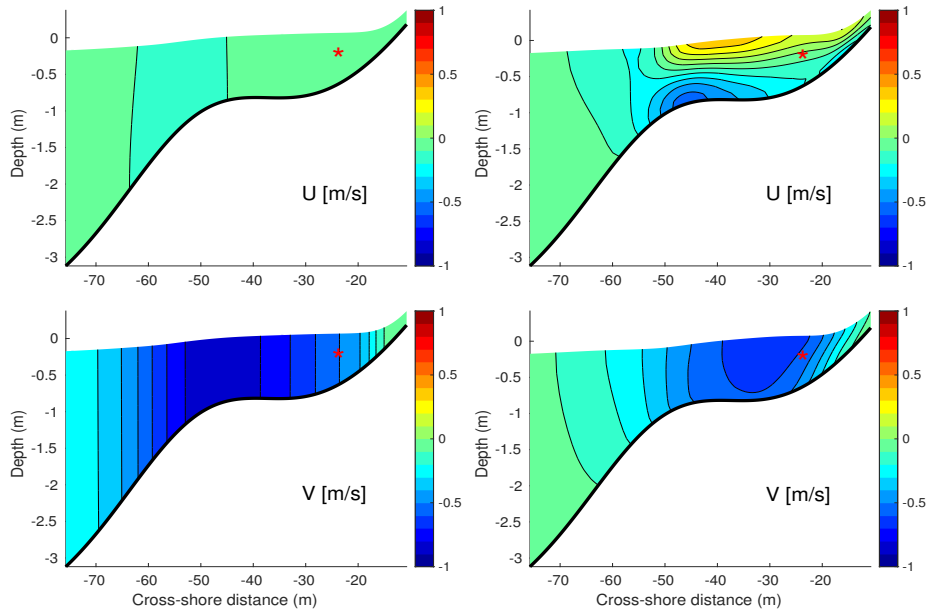


Figure 7: Cross-section of time-mean and longshore-mean, cross-shore (top) and longshore (bottom) currents in the case of shallow (right) and deep (left) breaking. The location of ADV measurement is indicated by a red star.

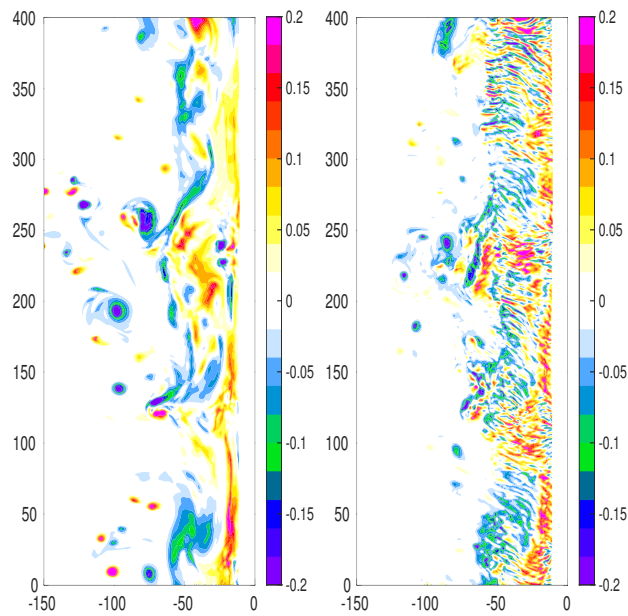


Figure 8: Surface vertical vorticity snapshot in the case of shallow (right) and deep (left) breaking.

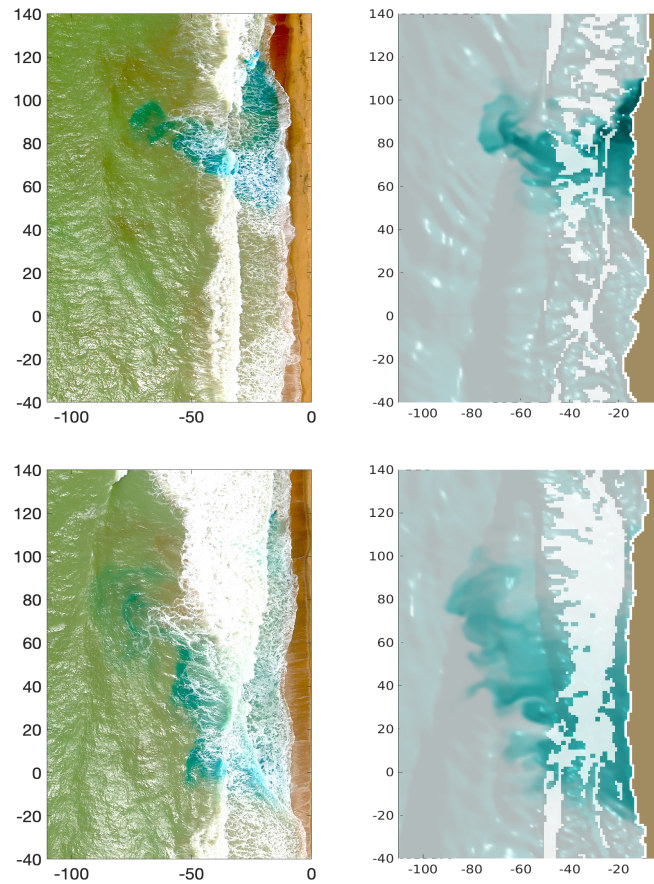


Figure 9: Drone photos (left) and CROCO simulation (right) of dye release at 116 s interval during the Grand Popo survey of March 13 2014 at afternoon mid-tide. In the model, wave surfaces are presented in the background with light from the left (as in the photos), and foam is approximated by high turbulent energy levels as in Fig. 6

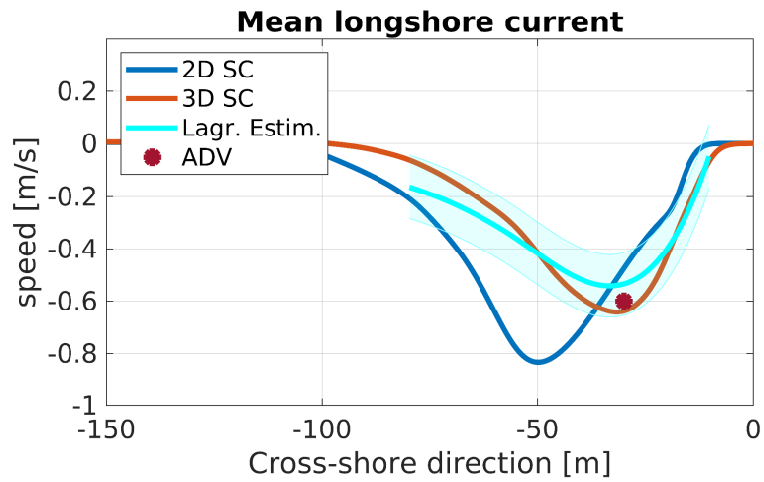


Figure 10: Cross-shore profile of time-mean and longshore-mean surface longshore flow for the full 3D simulation with wave direction $D = 10^\circ$. ADV measurement (extrapolated to the surface using the model profiles) and an estimation of Lagrangian velocities from drone photos (Fig. 9) is added for comparison.

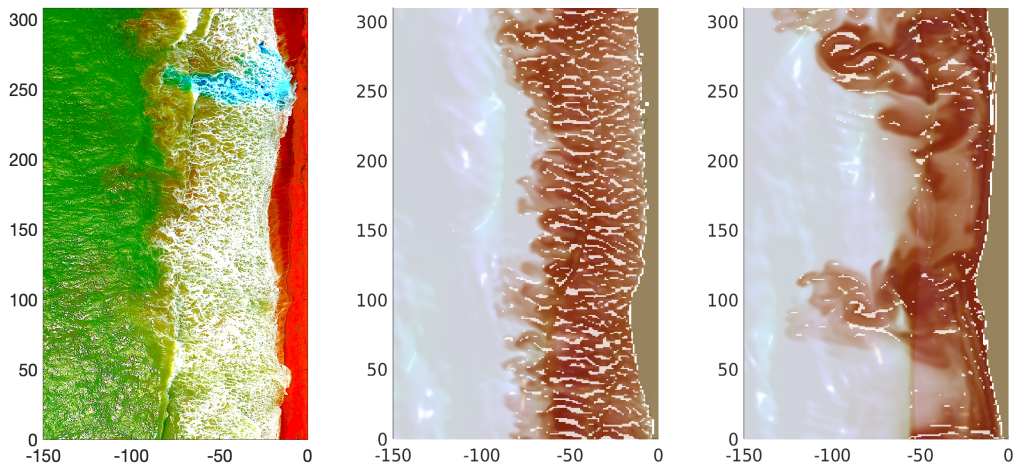


Figure 11: Aerial photo (left) and CROCO simulations of surfzone suspended sediments for shallow-breaking (center) and deep-breaking (right) cases. The contrasts in the photo is enhanced to better expose suspended sediments (brown color). The model suspended sediments correspond to snapshots at 15 min of simulation. Lines of strong alongshore convergence of surface currents are represented with white patches, to compare with foam structures observed in the photo.

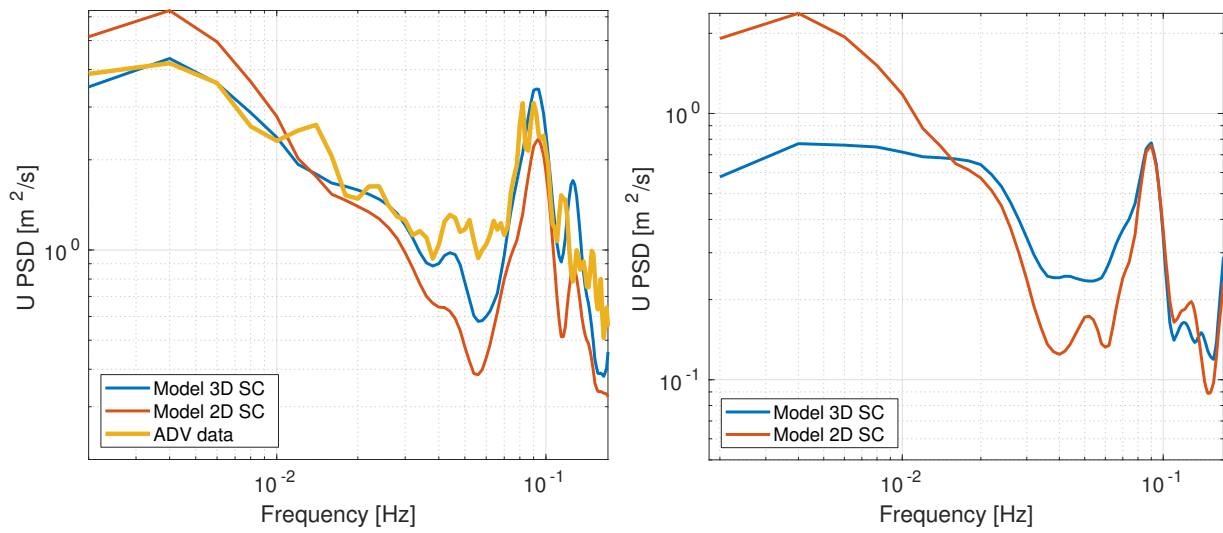


Figure 12: Power Spectral Density of horizontal velocity fluctuations compared with ADV measurements in the middle of the terrace (left) and over the outer terrace slope (right), in March 13 2014 at afternoon mid-tide.

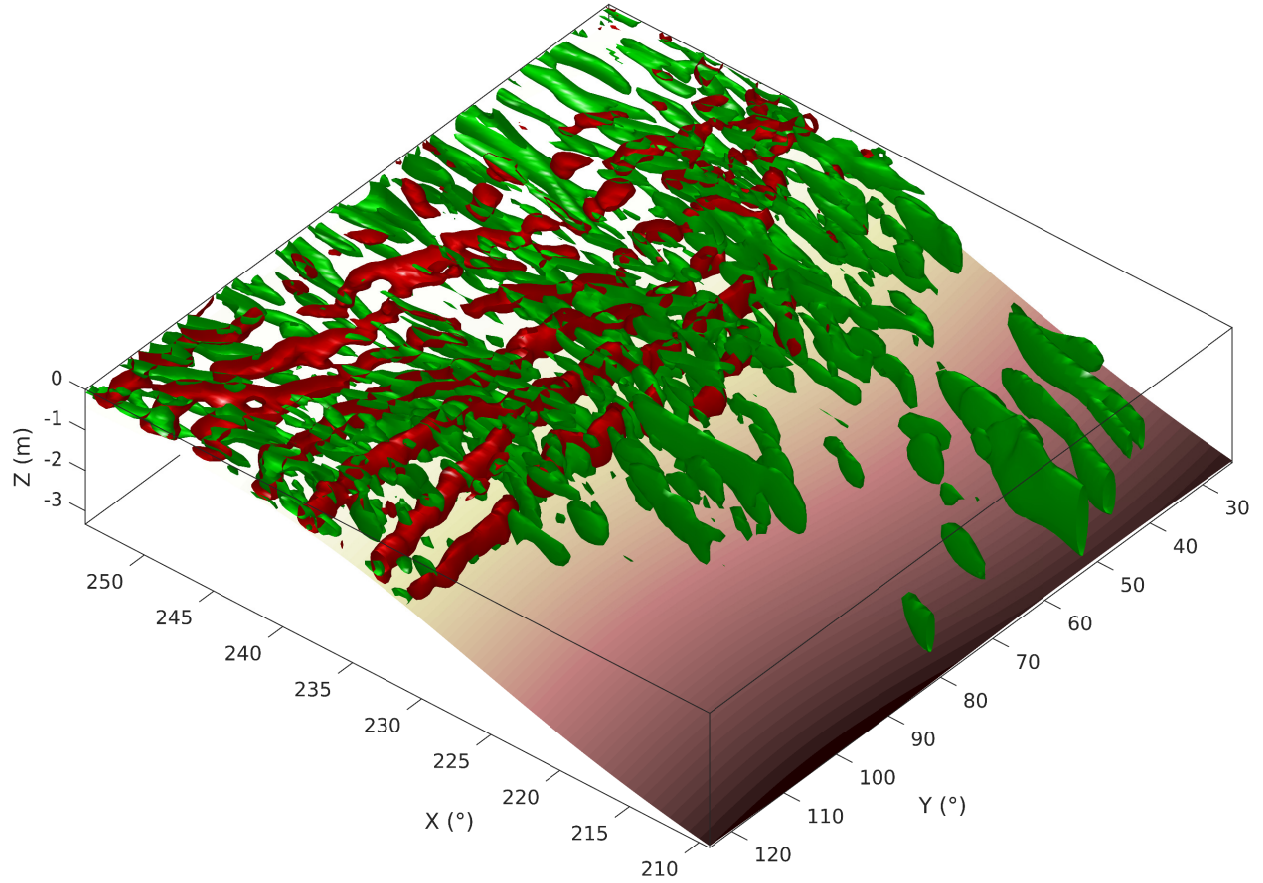


Figure 13: Q field defined by $Q = -\frac{1}{2} \frac{\partial u_i}{\partial x_j} \frac{\partial u_j}{\partial x_i}$, showing coherent structures similar to rolls and ribs in a transitional mixing layer. Cross-shore and alongshore Q terms are split: spanwise rolls (aligned across shear direction) are identified by $Q_y = -\frac{\partial u}{\partial z} \frac{\partial w}{\partial x} - \frac{1}{2} \frac{\partial u^2}{\partial x}$ in red; and streamwise ribs (along shear direction) are identified by $Q_x = -\frac{\partial v}{\partial z} \frac{\partial w}{\partial y} - \frac{1}{2} \frac{\partial v^2}{\partial y}$ in green. The fields are normalized and only positive isosurface values (0.02) are plotted.

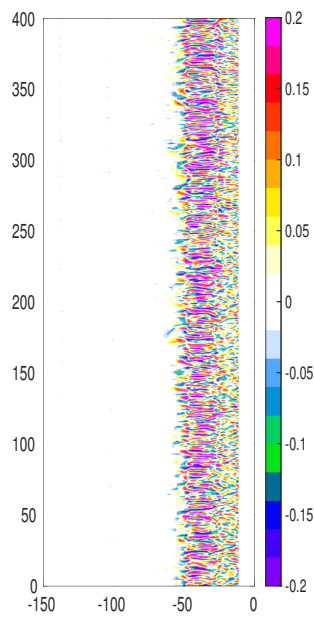


Figure 14: Surface vertical vorticity snapshot for the case with monochromatic shore-normal wave forcing (3D_MONO_D0)

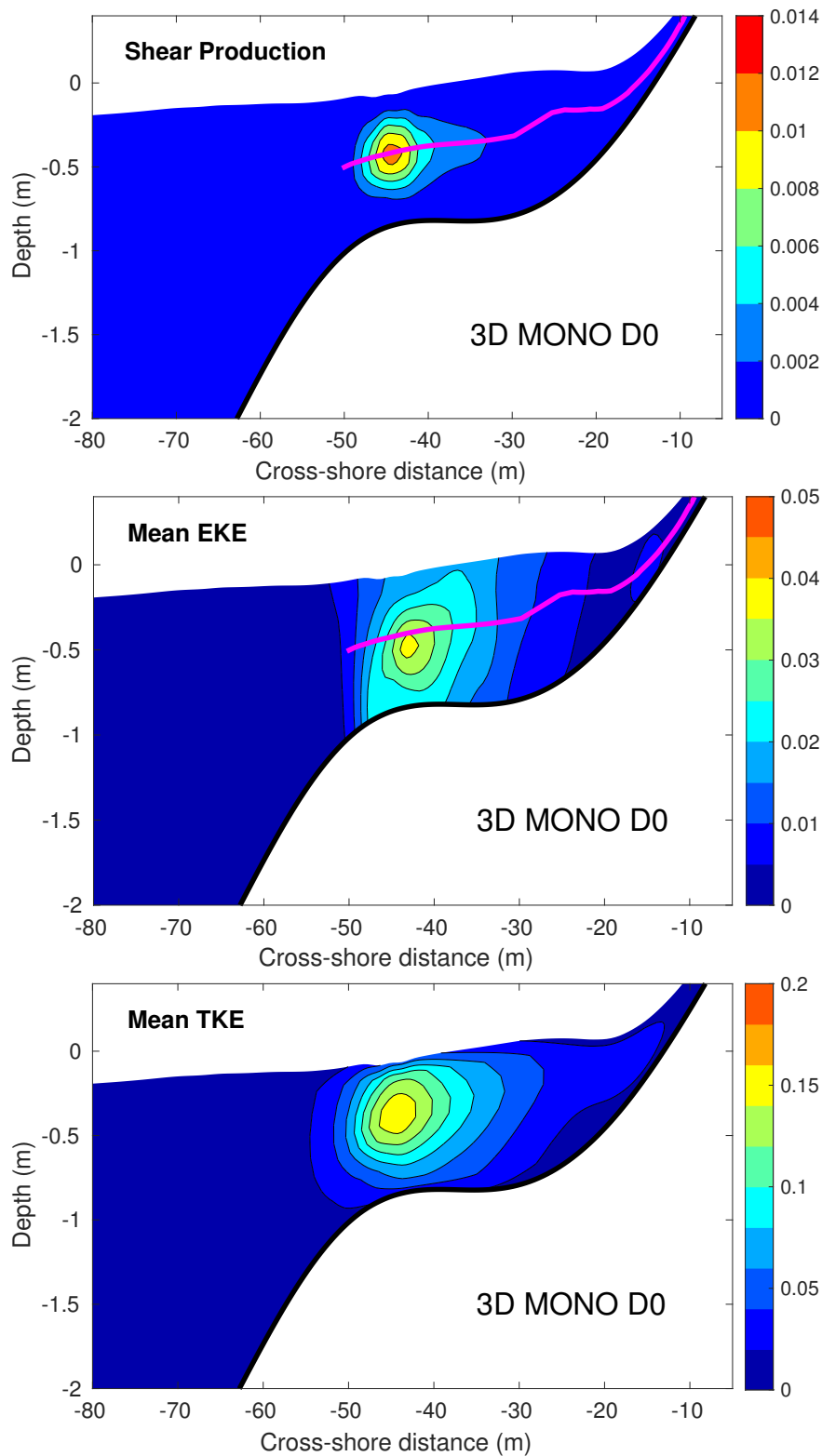


Figure 15: Cross-section of EKE production by the mean vertical shear flow $-\overline{u'w'} \frac{\partial \bar{u}}{\partial z}$ [m^2/s^3] (top); Mean EKE [m^2/s^2] (middle); Mean turbulent kinetic energy k [m^2/s^2] given by the k - ω model (bottom). All fields are for the case with 3D instabilities only, forced by monochromatic long-crested waves (3D_MONO_D0). The presence of an inflection point in the velocity profile where $\frac{\partial^2 \bar{u}}{\partial z^2} = 0$ is shown in a magenta line

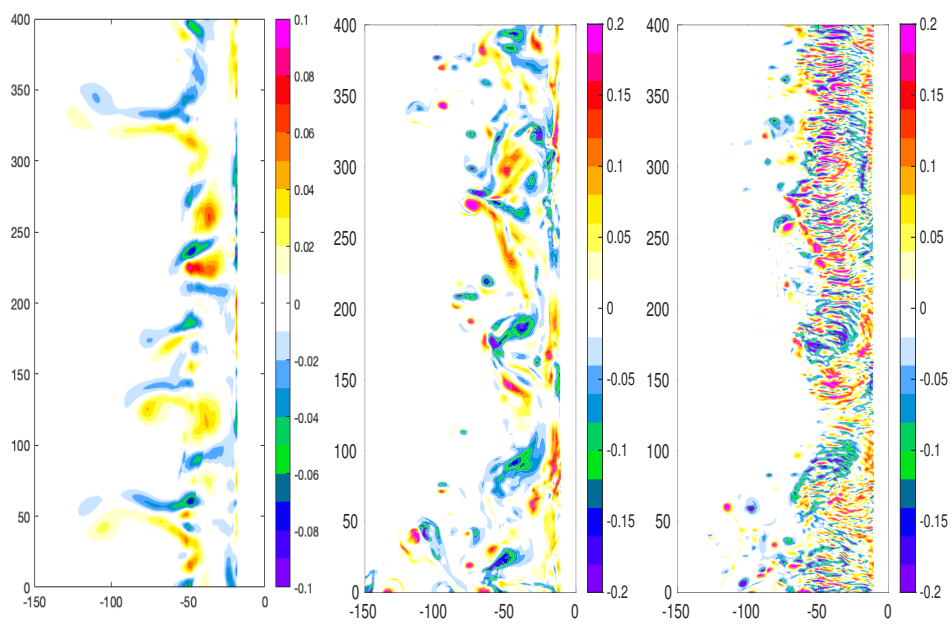


Figure 16: Vertical vorticity snapshot for 3 cases with shore-normal wave forcing: FUNWAVE-TVD (2D Boussinesq) with short-crested waves (left); CROCO with short-crested waves and deep breaking (center); short-crested waves and shallow breaking (right).

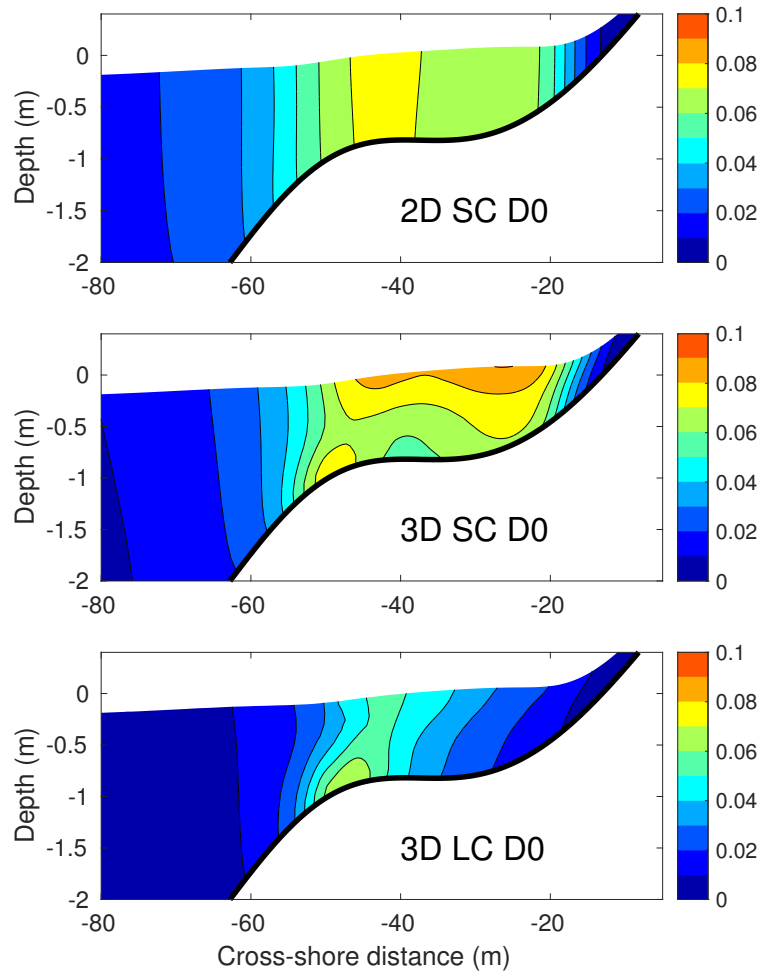


Figure 17: Cross-section of time-mean and longshore-mean EKE for the 3 cases of Figure 16

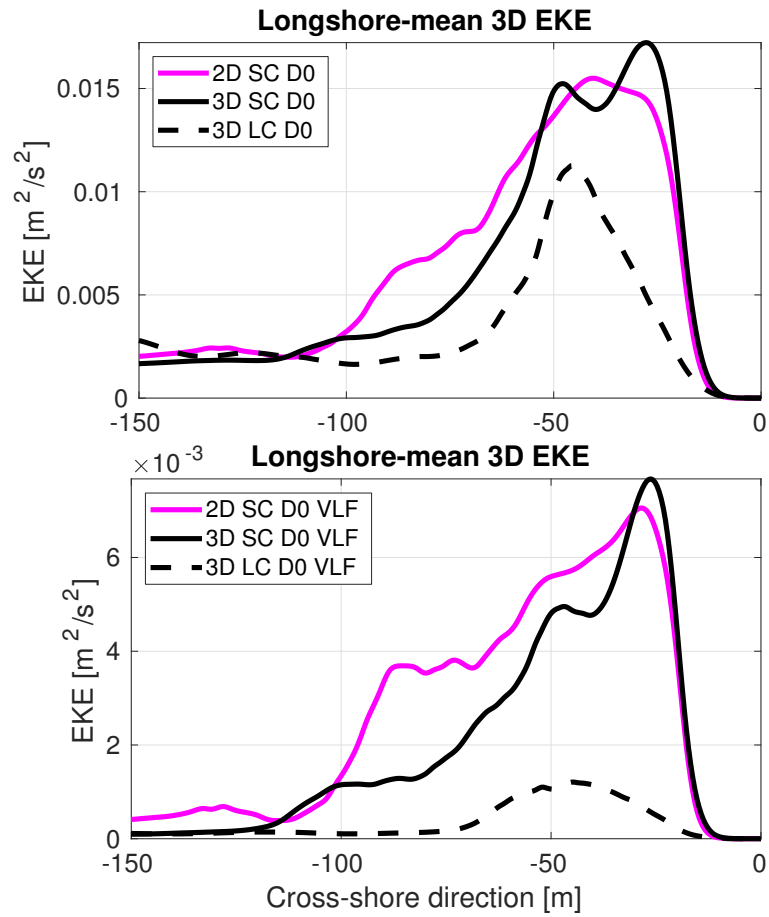


Figure 18: Cross-shore profile of depth-integrated (normalized), time and longshore-mean EKE for the 3 cases in Figure 16. Top: EKE of wave-averaged flow; bottom: EKE of low-pass filtered flow to remove 3D instability modes.

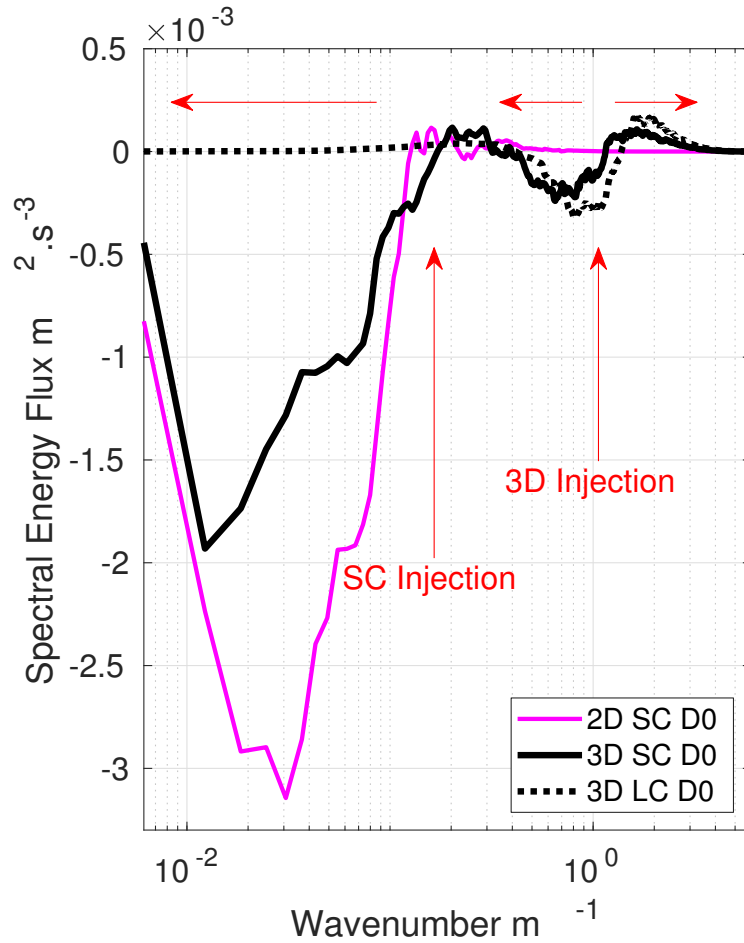


Figure 19: 3D and 2D model wavenumber spectral flux in the surfzone. Vertical red arrows point to the wavenumbers of energy injection from short-crested waves (~ 30 m wavelength) and from 3D shear instability (~ 5 m). The spectral flux is computed by spectral integration of v advection term (Marchesiello et al., 2011). Horizontal red arrows indicate positive/negative fluxes, i.e., direct/inverse energy cascade toward smaller/larger scales.

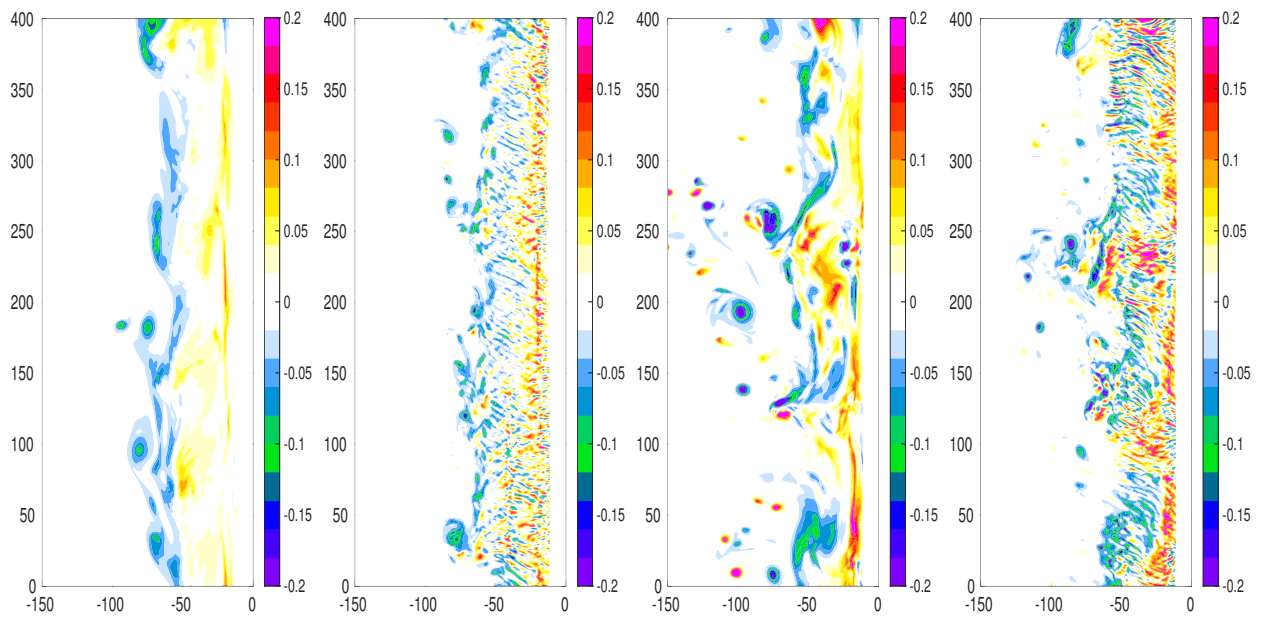


Figure 20: Snapshot of vorticity fields for 4 cases with oblique waves: long-crested waves with deep breaking (2D_LC_D10: first left); long-crested waves with shallow breaking (3D_LC_D10: second left); short-crested waves with deep breaking (2D_SC_D10: second right); and short-crested waves with shallow breaking (3D_SC_D10: last right).

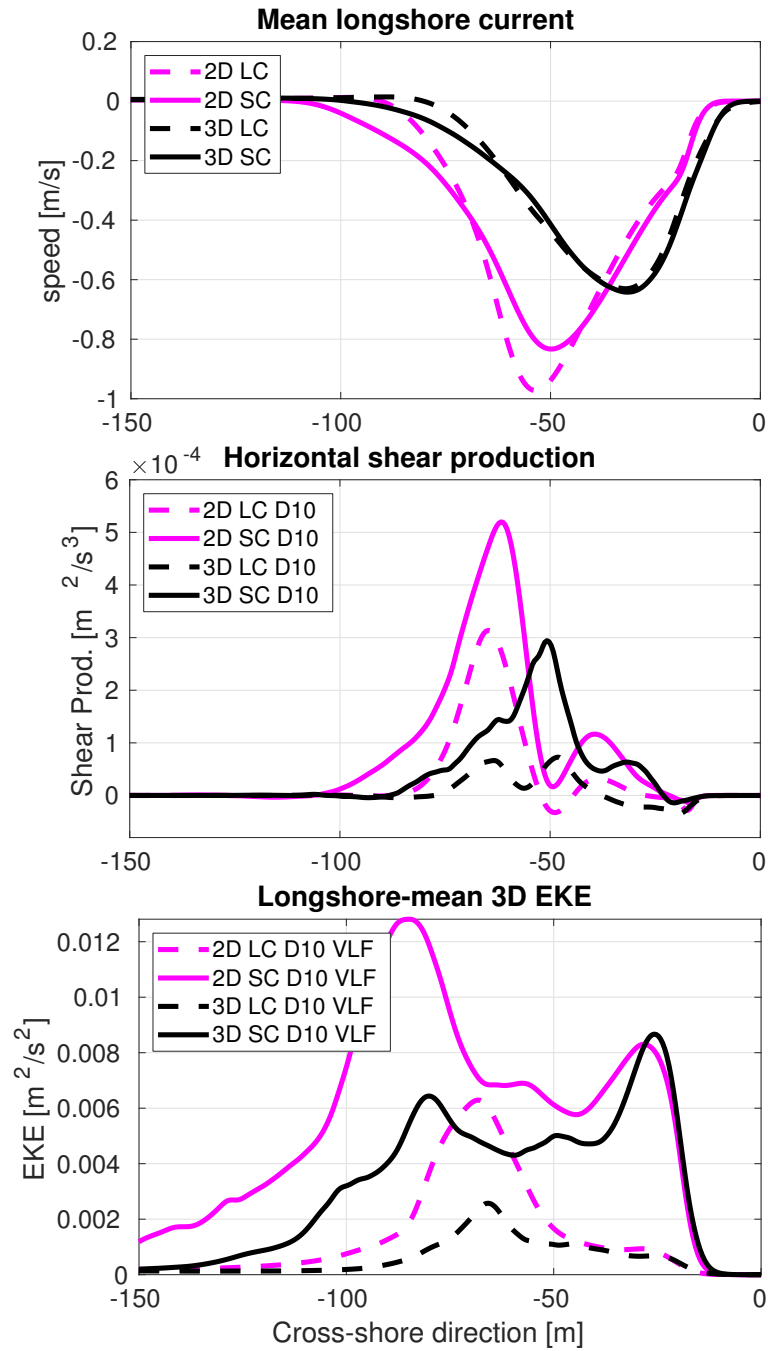


Figure 21: Cross-shore profile of time-mean and longshore-mean flow and eddy statistics for the 4 cases in Figure 16; top: surface longshore flow; center: horizontal shear production $-\overline{u'v'} \frac{\partial \bar{v}}{\partial x}$; bottom: depth-integrated EKE of low-frequency modes.