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# Tridimensional nonhydrostatic transient rip currents in a wave-resolving model

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## Abstract

Flash rips and surfzone eddies are transient horizontal structures of order 10-100 m scale, that can be generated in the surfzone in the absence of bathymetric irregularities. They are traditionally assessed within a depth-averaged framework that involves intrinsic horizontal shear instabilities or/and direct short-crested wave vorticity generation. In this paper, we revisit surf eddy generation processes with a new tridimensional, wave-resolving model (CROCO). We first present a quick overview of a free-surface, compressible approach adapted to nearshore dynamics. Its ability to simulate surface gravity wave propagation and nearshore breaking is validated against laboratory experiments. Next, we present a real-world application of the 3D wave-resolving model to Grand Popo Beach, Benin, forced by waves with frequency and directional spreading. The generation of surfzone eddies by the 3D model is shown to differ considerably from that produced by depth-averaged models, owing to the vertical structure of currents produced by shallow breaking that prevents a strong inverse energy cascade. Processes of both horizontal shear instability and short-crested wave breaking are restricted. Instead, the vertical shear flow is prone to forced variability by wave groups and to a Kelvin-Helmholtz-type instability. The latter shows primary and secondary instabilities, that generate spanwise and streamwise vorticity connecting breaking-induced rolls to larger horizontal surfzone structures. Streamwise filaments, appearing as ribs or mini rips, can extend beyond the surfzone but with lower energy than depth-averaged model rips

(halving offshore mean EKE). Velocity spectra and visual inspection of tracer and sediment concentrations in Grand Popo Beach show much better fit between the 3D model patterns and in-situ and aerial observations. Our results thus propose a plausible demonstration of new 3D nonhydrostatic instability modes and turbulent cascades for the nearshore zone, which will need to be reassessed in a wider range of coastal settings.

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

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## 1. Introduction

Flash rips and surfzone 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 breaking-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 – depth-averaged, mostly. Three types of processes stand out:

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

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18 • Tridimensional (3D) shear instability

19 Horizontal shear instability of longshore currents is the earliest process of eddy gener-  
20 ation, accounting for the intrinsic variability of wave-driven currents (Bowen and Holman,  
21 1989; Dodd et al., 1992; Allen et al., 1996; Slinn et al., 1998; Özkan-Haller and Kirby, 1999;  
22 Dodd et al., 2000; Uchiyama et al., 2009). This process was typically studied with wave-  
23 averaged shallow-water models, where momentum transfer from wave to current fields is  
24 entirely parametrized. Its importance has faded during the last decade due to prevalence of  
25 the Peregrine process, but also to conflicting results given by newly developed wave-averaged  
26 tridimensional models (Newberger and Allen, 2007; Splinter and Slinn, 2003).

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

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

46 date are performed using hydrostatic assumption, thus underestimating horizontal vorticity  
47 motions.

48 Nonhydrostatic dynamics are essential in our third listed process of surf eddy generation.  
49 They are mostly studied in laboratory experiment (Nadaoka et al., 1989) and laboratory-  
50 scale Large-Eddy Simulations (LES) using 2.5D CFD models applied to individual wave  
51 breaking (Lin and Liu, 1998; Watanabe and Saeki, 1999; Watanabe et al., 2005; Kirby and  
52 Derakhti, 2019). These previous studies show that the spanwise (mostly alongshore) com-  
53 ponent of vorticity is an important aspect of the breaking process. It is associated with  
54 vertical shear generated by shallow breaking, which can produce traveling rolls through a  
55 primary instability (Watanabe et al., 2005). The rolls, typical of unsteady parallel shear  
56 flows, can subsequently evolve through a secondary instability to yield streamwise vorticity,  
57 transitioning toward fully tridimensional turbulence. However, 2.5D CFD models are ap-  
58 plied to individual breaking waves with only few alongshore wavelengths of the secondary  
59 instability, precluding any evaluation of eddy statistics.

60 3D nonhydrostatic processes are usually studied independently of the two others (by  
61 separate research communities) and rarely compared in terms of scales, magnitude and in-  
62 teraction. The only attempt was made by Splinter and Slinn (2003) in a proceeding report.  
63 Using a 3D nonhydrostatic model where breaking acceleration is introduced as a body force,  
64 they show that deep breaking reproduces 2D model solutions, while the more realistic shal-  
65 low breaking process seems to disrupt the formation of horizontal shear instability at the  
66 expense of vertical shear instability. However, their domain size do not allow statistical  
67 comparisons and the profile of breaking acceleration is imposed, not computed from a wave-  
68 resolving model. The present study is a step forward from this early work, also addressing  
69 the case of short-crested wave generation. Note that (Kumar and Feddersen, 2017) stud-  
70 ied transient eddies produced by a 3D nearshore circulation model, forced by short-crested  
71 waves computed beforehand with a Boussinesq model. However, wave forcing is prescribed  
72 as a depth-uniform body force, i.e., deep breaking, and could not produce vertical shear of  
73 the cross-shore flow. Their hydrostatic assumption also precluded the model from vertical  
74 shear instabilities.

75 CROCO (Coastal and Regional Ocean Community model) is a new oceanic modeling  
76 system built upon ROMS (Shchepetkin and McWilliams, 2005; Debreu et al., 2012), with  
77 added non-Boussinesq kernel (Auclair et al., 2018). CROCO is designed for solving fine-scale  
78 processes from the regional ocean to littoral zone with capabilities for LES in realistic config-  
79 urations. Particular attention is given to numerical accuracy, high-performance computing  
80 (optimization, scalability), portability and easy access ([www.croco-ocean.org](http://www.croco-ocean.org)). This pa-  
81 per presents a quick overview of the nonhydrostatic CROCO solver with a non-Boussinesq  
82 (compressible) approach, before addressing its application to nearshore dynamics. First,  
83 its ability to simulate surface gravity wave propagation and nearshore breaking is validated  
84 against laboratory experiment. Second, we analyse and validate a full 3D, wave-resolving,  
85 real-case simulation of transient rips, in the presence (or not) of short-crested waves and  
86 strong alongshore currents. We discuss fundamental differences in the generation of surf  
87 eddies by 3D wave-resolving models compared with depth-averaged models, with a focus on  
88 the vertical structure of currents produced by shallow breaking and associated turbulence.  
89 We conclude on the limitation of simplified vorticity evolution equations in which only the  
90 vertical part is considered when so much activity lies in horizontal vorticity, governed by 3D  
91 nonhydrostatic equations.

## 92 **2. Model description**

93 Because of limited computational resources, 3D wave-resolving models are still rarely  
94 used to study nearshore dynamics in realistic environments. LES applications appeared in  
95 the 1990s and are generally restricted to 2.5D laboratory-scale experiments of individual  
96 wave breaking. Early applications used the volume-of-fluid (VOF) method for free-surface  
97 tracking (e.g., Lin and Liu 1998; Watanabe and Saeki 1999; Watanabe et al. 2005; Derakhti  
98 and Kirby 2014). This model type with Cartesian coordinate, where the free surface crosses  
99 computational cells arbitrarily, fails to precisely apply the pressure boundary condition on  
100 the free surface, affecting the model accuracy.

101 More recently, several 3D wave-resolving, free-surface and terrain-following models have  
102 emerged for the nearshore zone, e.g., SWASH (Zijlema et al., 2011) and NHWAVE (Ma

103 et al., 2012; Derakhti et al., 2016), based on earlier attempts (e.g., Lin and Li 2002). In this  
104 case, explicit free-surface overturning, occurring during breaking, is precluded and must be  
105 parametrized as a turbulent or a shock process. However, these models can be accurate as  
106 well as computational efficient in the study of waves and wave-driven mean and transient  
107 circulation.

108 CROCO belongs to this class of models but, unlike other attempts, resolves the full com-  
109 pressible Navier-Stokes equations (Auclair et al., 2018). A compressible approach preserves  
110 the hyperbolic nature of Navier-Stokes equations and does not require a global elliptic solver  
111 with incremental pressure corrections to ensure the incompressible mass balance. As a re-  
112 sult, it avoids splitting errors between pressure and velocity and approximations made on  
113 free-surface conditions (Zijlema et al., 2011), thereby preserving amplitude and dispersion  
114 properties of nonhydrostatic surface waves. In the same time, the absence of global compu-  
115 tations by an elliptic solver makes parallelization and optimization procedures much more  
116 efficient. The cost of solving acoustic waves is managed with a time-splitting technique and  
117 semi-implicit time discretization, introduced below.

118 The development of CROCO around the Regional Oceanic Modeling System (ROMS)  
119 has advantages for realistic LES applications. It benefits from capabilities long developed  
120 in oceanic models – high-performance computing; high-order discretization; coupling with  
121 biogeochemistry and sediment models; pre-processing tools for fast model input generation.  
122 The nonhydrostatic model version can thus be applied without much effort to realistic, highly  
123 nonlinear regimes, e.g., large internal solitons and hydraulic jumps (Hilt et al., 2020), Kelvin-  
124 Helmholtz instabilities (Penney et al., 2020), or wave-induced nearshore circulation as in the  
125 present study. It is naturally suited for bridging ocean and coastal sciences, e.g., addressing  
126 surf-shelf exchange processes in a 3D, rotating and stratified framework. In addition, both  
127 wave-resolving and wave-averaged (Uchiyama et al., 2010; Marchesiello et al., 2015) model  
128 equations are available within the same code, which has potential benefits for evaluating  
129 parametrizations of wave-current interactions in wave-averaged models.

130 *2.1. Free-surface, compressible ocean model equations*

131 The full set of Navier-Stokes equations for a free-surface ocean is explicitly integrated in  
 132 the nonhydrostatic, non-Boussinesq (compressible) version of CROCO, built upon the code  
 133 structure of ROMS primitive equation solver. In the compressible approach (Auclair et al.,  
 134 2018), acoustic waves are solved explicitly to avoid Boussinesq degeneracy, which inevitably  
 135 leads to a 3D Poisson system in nonhydrostatic incompressible methods – detrimental to  
 136 computational costs and accuracy of free-surface model implementation.

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

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

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

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

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

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

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

140  $(u, v, w)$  are the  $(x, y, z)$  components of vector velocity  $\vec{v}$ ;  $\eta$  is the free surface;  $\phi(x, y, z, t)$   
 141 is dynamic pressure  $\phi = P/\rho_0$ , with  $P$  the total pressure;  $\rho_0 + \rho(x, y, z, t)$  is total *in situ*  
 142 density;  $f(x, y)$  and  $\tilde{f}(x, y)$  are the traditional and non-traditional Coriolis parameters,  
 143 function of latitude;  $g$  is acceleration of gravity;  $\mathcal{D}_u, \mathcal{D}_v, \mathcal{D}_C$  are diffusive terms;  $\mathcal{F}_u, \mathcal{F}_v, \mathcal{F}_C$   
 144 are forcing terms;  $\lambda$  is the second (bulk) viscosity, associated with compressibility (used to  
 145 damp acoustic waves).



146 *2.2. Time-splitting principle*

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

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

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

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

152 The Navier-Stokes equations are then integrated with two different time steps within  
 153 the time-splitting approach inherited from ROMS. The slow-mode integration is similar to  
 154 ROMS with the addition of the slow part of vertical momentum equation, whereas fast-  
 155 mode integration is 3D and includes the compressible terms of momentum and continuity  
 156 equations. In vector form:

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

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

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

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

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

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

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

157 The momentum is integrated both in slow and fast modes but the right-hand-side of  
 158 the equation is split in two parts: a slow part, made of slowly varying terms (advection,  
 159 Coriolis force, baroclinic pressure force and viscous dissipation), and a fast part, made of fast-  
 160 varying terms (the surface-induced and compressible pressure force, weight, and dissipation  
 161 associated with bulk-viscosity). This momentum equation is numerically integrated twice,  
 162 once with a large time-step keeping the fast part constant, and once with a smaller time-step  
 163 keeping the slow part constant. This is much more computationally efficient than integrating  
 164 the whole set of equations at the same fast time step. More details can be found in Auclair  
 165 et al. (2018)<sup>1</sup>.

166 Note that acoustic waves can become pseudo-acoustic if their phase speed  $c_s$  is artificially  
 167 reduced ( $c_s$  is a model parameter). In this case, high-frequency processes associated with  
 168 bulk compressibility may be unphysical, but an accurate solution for slower nonhydrostatic  
 169 dynamics can be preserved, while relaxing CFL constraints.

### 170 *2.3. Discretized equations for nearshore application*

171 In this study, motions are produced by an offshore wave-maker in a non-rotating, homo-  
 172 geneous fluid. In this case, the Coriolis force, baroclinic pressure force and all surface fluxes  
 173 are null. There is no temperature or salinity stratification so that slow density  $\rho_s$  is constant  
 174 in time and space.

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<sup>1</sup>Auclair et al. (2018) presents 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

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

183 Horizontal and vertical advection terms are discretized using the WENO5-Z improved  
184 version of the 5th-order weighted essentially non-oscillatory scheme (Borges et al., 2008),  
185 which is popular for hyperbolic problems containing both shocks and smooth structures.  
186 WENO5-Z naturally copes with dispersive numerical modes as well as shocks caused by  
187 breaking waves, with no need for ad hoc criteria. Additionally, a 3D Smagorinsky model is  
188 used for turbulent closure on anisotropic grids:

$$\nu_t = (C_S \Delta)^2 |S| \quad (16)$$

189 where the 3D strain rate  $|S| = \sqrt{2S_{ij}S_{ij}}$ , with  $S_{ij} = \frac{1}{2}(\frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i})$  (using Einstein notation);  $\Delta$   
190 the grid-characteristic length scale, i.e. the grid size:  $\sqrt{\delta_x \delta_y}$  for horizontal viscosity and  $\delta_z$  for  
191 vertical viscosity, where  $(\delta_x, \delta_y, \delta_z)$  are mesh sizes in (x,y,z) directions.  $C_S$  is the Smagorin-  
192 sky coefficient, taking a standard value of 0.1 (Canuto and Cheng, 1997; Christensen and  
193 Deigaard, 2001), although the sensitivity of model results to this value is discussed along the  
194 paper. Alternatively for some tests, a  $k - \epsilon$  model, solving closure equations for turbulent  
195 kinetic energy  $k$  and dissipation  $\epsilon$ , was used as part of a Generic Length Scale (GLS) method  
196 (Warner et al., 2005). In this case, turbulent viscosity  $\nu_t = c_\mu l k^{\frac{1}{2}}$ , with coefficient  $c_\mu$  de-  
197 pendent on stability functions, and mixing length scale  $l \propto k^{\frac{3}{2}} \epsilon^{-1}$ . Here,  $l$  is independent of  
198 resolution, which is more strictly associated with RANS (Reynolds-Averaged Navier-Stokes  
199 equations) than LES. All  $k - \epsilon$  model parameters are standard (Warner et al., 2005).

200 *2.4. Wave maker at offshore boundary*

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

$$\eta_{bc}(y, t) = \sum_i a_i \sum_j d_j \cos(k_{y,ij}y - \omega_i t - \phi_{ij}) \quad (17)$$

204

$$\text{with } d_j = e^{-\frac{\theta_j - \theta_m}{\sigma_\theta^2}} \text{ and } \sum d_j = 1 \quad (18)$$

205

$$u_{bc}(y, t) = \eta_{bc}(y, t) \omega_p \cos(\theta_m) \frac{\cosh(k_p(z+h))}{\sinh(k_p h)} \quad (19)$$

206

$$v_{bc}(y, t) = \eta_{bc}(y, t) \omega_p \sin(\theta_m) \frac{\cosh(k_p(z+h))}{\sinh(k_p h)} \quad (20)$$

207 where  $(x, y, z)$  are cross-shore, alongshore and vertical directions respectively;  $(i, j)$  are in-  
 208 dices of spectral distribution in frequency and direction respectively;  $a_i$  is the amplitude  
 209 at each frequency  $\omega_i$ , from a given statistical distribution, e.g., JONSWAP (Sec. 4.1);  
 210  $k_{y,ij} = k_i \sin(\theta_j)$  is the alongshore wavenumber, where  $k_i$  is the linear theory wavenum-  
 211 ber:  $\omega_i^2 = g k_i \tanh(k_i h)$ , and  $\theta_j$  is wave angle;  $\theta_m$  is the mean wave direction and  $\sigma_\theta$  the  
 212 directional spread around the mean;  $\omega_p$  and  $k_p$  are peak frequency and wavenumber;  $d_j$   
 213 is a normalized frequency-dependent directional distribution;  $\phi_{ij}$  is a uniformly distributed  
 214 random phase.

215 Here  $w_{bc}$  is set to zero and our tests show only weak sensitivity to this choice. Depth-  
 216 averaged (barotropic) velocities  $(\bar{u}, \bar{v})$  must be provided as well in the wave maker because  
 217 they are prognostic variables of our split-explicit model, advanced together with the fast  
 218 acoustic mode. Normal depth-averaged velocity  $\bar{u}$  is complemented at the boundary by a  
 219 "compensation flow", opposite to Stokes drift and thus closing the volume budget. We do not  
 220 impose the depth-averaged value of  $u_{bc}$  directly but the value of the incoming characteristic  
 221 of the shallow water system as in Flather-type conditions (Marchesiello et al., 2001; Blayo

222 and Debreu, 2005):

$$\bar{u} = \bar{u}_{bc} - \sqrt{\frac{g}{h}}(\eta - \eta_{bc}) \quad (21)$$

223 This allows infragravity waves generated inside the domain to propagate out as long waves,  
224 while ensuring a near conservation of mass and energy through the open boundary. Similarly,  
225 the baroclinic components  $(u_{bc}, v_{bc}, w_{bc})$  are applied through an adaptive radiation condition  
226 that allows 3D flow perturbations to leave the domain without deteriorating the interior  
227 solution (Marchesiello et al., 2001).

### 228 3. Validation in Flume experiment

#### 229 3.1. GLOBEX experiment

230 As a first step towards 3D modeling, we present here a validation of wave propagation  
231 and breaking using a wave flume experiment. The Gently sLOping Beach EXperiments  
232 (GLOBEX<sup>2</sup>) were performed in the Scheldt flume of Deltares (Delft, the Netherlands) in  
233 2012, and described in Michallet et al. (2014). The project objective was to collect high-  
234 resolution space-time data of the cross-shore evolution of short and infragravity waves on a  
235 gentle slope for a range of wave conditions.

236 The flume is 110 m long, 1 m wide and 1.2 m high. The waves were generated with  
237 a piston-type wave maker equipped with Active Reflection Compensation to minimize re-  
238 flections from the wave paddle. A concrete beach with a weakly reflexive 1:80 slope was  
239 constructed, with its toe at 16.57 m from the wave maker. All experiments were run with a  
240 still water depth of 0.85 m and still-water shoreline at  $x = 84.57$  m. The material that was  
241 laying loose on the concrete bed before the flume was filled with water had a median grain  
242 size  $D50=0.75$  mm (roughness length  $z_0 \sim D50/12 = 0.0625$  mm).

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

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<sup>2</sup>Globex data is freely available at [zenodo.org/record/4009405](https://zenodo.org/record/4009405)

247 on one of the experiments, B3, corresponding to second-order (Stokes) wave generation of  
248 bichromatic frequencies (simulated with a Boussinesq model in Michallet et al. 2014). The  
249 characteristics are as follows:  $a_1 = 0.09$  m;  $a_2 = 0.01$  m;  $f_1 = 0.420$  Hz;  $f_2 = 0.462$  Hz  
250 (short-wave peak period  $T_p = 2/(f_1 + f_2) = 2.27$  s; and group period  $T_g = 1/(f_2 - f_1) =$   
251 23.81 s). The signal had a total duration of 75 minutes.

### 252 3.2. Model comparison

253 The model is set-up with the same conditions as the wave flume experiment. Second-  
254 order bichromatic waves are generated at the offshore boundary, with shore normal direction  
255 and zero directional spread. A no-slip condition is imposed on the lateral wall boundaries  
256 of the canal so that transverse modes are precluded.

257 Figure 1 compares an Hovmuller plot (x,time) of data and model sea level  $\eta$  and undertow  
258  $u_b$ .  $u_b$  is interpolated at the measurement depth assuming a logarithmic profile. When data  
259 is missing in the measurements, it is also removed from the model output. The general  
260 structure reflecting wave speed and frequencies, wave packets, surf and swash zones are all  
261 very similar. Model-data correlations are high with 0.85 and 0.77 respectively for  $\eta$  and  $u_b$ ,  
262 and root mean square errors are 2 cm and 0.13 m/s. Some scattering in the undertow data  
263 is noticeable. According to Michallet et al. (2014), it may be attributed to the presence of  
264 secondary motions generated by transverse waves at the break point where the transverse  
265 mode-1 seiche can be excited at frequency  $f_1 + f_2$ . It may also be due to variations in depth  
266 level of flow measurements in the surf zone.

267 A snapshot of wave field across the flume during runup condition is presented in Figure  
268 2. It highlights the main processes of propagation, nonlinear interaction, shoaling, breaking,  
269 roller propagation and runup. Model-data correlation is high, as already mentioned, and non-  
270 linearity is apparent in both cases in the increasingly non-sinusoidal shape of short waves as  
271 they approach the shore (Elgar and Guza, 1985). They first develop short, high wave crests  
272 with increasing skewness (asymmetry about the horizontal axis, measuring crest/trough  
273 shape), and as they break, transition into the characteristic saw-tooth shape with asymmetry  
274 about the vertical axis.

275 Wave statistics (first, second and third moments) for  $\eta$  and  $u_b$  are shown in Figure 3:  
 276 mean, standard deviation (or  $H_s$  for  $\eta$ ), skewness  $\langle\phi'^3\rangle/\langle\phi'^2\rangle^{1.5}$  and asymmetry  $\langle\mathcal{H}(\phi')^3\rangle/\langle\phi'^2\rangle^{1.5}$   
 277 ( $\mathcal{H}$  is Hilbert transform). For the mean field of measured undertow, a few scattered points  
 278 lying far outside the standard deviation are removed using a polynomial fit. With this, the  
 279 model appears to replicate the observed cross-shore undertow profile (top-left of Fig. 3).  
 280 The undertow is part of a vertical recirculation associated with breaking-induced surface  
 281 onshore flow. It may be considered a proxy of the vertical shear, which will appear as an  
 282 essential parameter of surfzone eddy generation in the next section. Here, we call undertow  
 283 the bottom return flow (10 cm/s in this experiment). There is also an Eulerian compensation  
 284 flow for the onshore surface Stokes drift (continuity of total flow must be preserved), which  
 285 is sometimes referred to as undertow. This compensation flow is the only cross-shore flow  
 286 captured by depth-averaged models and is relatively weak (on the order of 1 cm/s in this  
 287 simulation). In Figure 3, the model also properly represents the envelop of high-order  
 288 moments. Details of skewness and asymmetry may differ in model and data, due in part  
 289 to irregular current measurement depth and transverse modes in surfzone measurement,  
 290 but the evolution and transition from high skewness to high asymmetry is similar. Apart  
 291 from somewhat higher skewness and lower asymmetry than measured in the breaker zone, we  
 292 conclude that nonlinear shoaling, breaking-induced wave transformations and – maybe more  
 293 importantly for the present study – the surfzone recirculation are generally well reproduced.

294 From sensitivity tests, it appears that a realistic reproduction of  $H_s$  cross-shore evolution  
 295 in the breaker zone ( $x = 60 - 70$  m) is conditioned to using the shock capturing scheme  
 296 WENO5-Z. The results are degraded (not shown) when replacing WENO5-Z with UP5, a  
 297 non-monotonic linear upstream-biased 5th-order advection scheme (Menesguen et al., 2018).  
 298 The Smagorinsky model alone (or alternatively a  $k - \epsilon$  model) was less efficient in dissipating  
 299 the breaking wave front. This is in line with analogies between breaking waves, bores and  
 300 hydraulic jumps, that may be treated as a shock (Cienfuegos et al., 2010; Tissier et al.,  
 301 2012; Lubin and Chanson, 2017). On the other end, the WENO scheme is less efficient in  
 302 the roller zone ( $x > 70$  m), where turbulent dissipation dominates over production (van der  
 303 A et al., 2017). There, the turbulent closure scheme takes over as main dissipation process.

304 Therefore, our combination of numerical and physical closure with off-the-box parameters  
305 provides a realistic framework for surfzone dynamics. The fact that no tuning was needed  
306 for these schemes and that there is no ad hoc detection and treatment of wave breaking  
307 provides confidence in applying the model to other configurations.

#### 308 **4. Natural beach application**

309 We now turn to a full 3D experiment with longshore-uniform bathymetry. The configu-  
310 ration is derived from Grand Popo Beach (6.2°N, 1.7°E, Benin, in the Gulf of Guinea; Fig.  
311 4). This stretch of coast presents a longshore-uniform, low tide terrace and steep upper  
312 shoreface (Almar et al., 2014, 2016) and a sandy wave-dominated and microtidal environ-  
313 ment, exposed to S-SW long period swells generated at high latitudes in the South Atlantic  
314 (Almar et al., 2015a). A field experiment was conducted at Grand Popo Beach from 10  
315 to 18 March 2014 (Almar et al., 2014; Derian and Almar, 2017). For our setup, we focus  
316 on conditions in the middle of the afternoon of March 13 2014. The weather, tides and  
317 wave conditions were ideal: weak winds and wind waves well separated from a narrow-band  
318 swell with significant wave height  $H_s = 1.15$  m, peak period  $Tp = 11$  s, and wave incidence  
319  $D = 10^\circ$  from shore normal direction (measured from an Acoustic Doppler Current Profiler  
320 moored in 10-m depth). The water was at mid neap tide level (low-tide terrace at about 1 m  
321 depth), promoting a narrow surfzone less than 50 m wide. A Nortek high-frequency acoustic  
322 Doppler velocimeter (ADV with sampling rate of 8 Hz) was deployed in the surfzone, in the  
323 middle of the terrace, measuring currents about 0.5 m from the bottom. A dye release was  
324 conducted to monitor the dispersion induced by flash rips, coupled with UAV flights (STB  
325 DS6 hexacopter) at an elevation of 100 m. The Drone camera (NIKON D700) was looking  
326 down, with a vertical angle, and recorded  $4256 \times 2832$ -px scenes at 1 Hz (Almar et al., 2014;  
327 Derian and Almar, 2017).

##### 328 *4.1. Model Setup*

329 The domain is 542 m alongshore by 240 m across-shore, with periodic alongshore bound-  
330 ary conditions. In order to prevent distortion when oblique waves are used with periodic



331 conditions, the alongshore size is adjusted according to peak wavelength and mean wave di-  
332 rection. This method proved to perform well even with long-crested waves. The grid spacing  
333 is generally  $dx = dy = 1$  m but one simulation is run with higher resolution of 0.5 m. There  
334 are 10 vertical levels evenly spaced between the free surface and bottom (a simulation with  
335 20 levels did not change the results much). The model time step is  $dt = 0.025$  s (0.0125 s at  
336 0.5 m resolution). The model bathymetry is longshore-uniform and built with continuous  
337 functions to smoothly fit the low-tide terrace structure observed during the survey of (Almar  
338 et al., 2014, 2018). Depth  $H$  is 8 m offshore to 1 cm at the shoreline, whose position varies  
339 with swash oscillation, relying on a wetting-drying scheme (Warner et al., 2013). Depth over  
340 the terrace is about 1 m according to the afternoon mid-tide conditions of March 13 2014.

341 The wave maker is set with following parameters:  $H_s=1.15$  m,  $T_p=11$  s,  $D = 0^\circ$ ,  $10^\circ$   
342 or  $20^\circ$  ( $D = 10^\circ$  for comparison with data in Sec 4.3), and directional spread  $\sigma_\theta = 30^\circ$ . A  
343 JONSWAP spectrum is constructed with these parameters and a peak enhancement factor  
344  $\gamma$  of 3.3. The quasi-monotonic and shock-capturing WENO5-Z scheme is used again with  
345 Smagorinsky turbulence closure and  $C_S = 0.1$ . Bottom roughness is  $z_0 = 0.01$  mm, which  
346 may seem low but gives a drag coefficient  $C_D \sim 0.002$  in the surfzone ( $C_D = \kappa^2/\ln(z_1/z_0)^2$ ,  
347 with  $\kappa = 0.41$  and  $z_1 \sim H/10$  the first level height above bed), which is a usual value in  
348 depth-averaged models (Chen et al., 2003; Feddersen et al., 2011). We follow the practice  
349 of these models here to reproduce their results within the pseudo-2D approach described  
350 below. It is of little consequence for 3D simulations because, as will be shown, they are  
351 much less sensitive to bottom drag than 2D models (Sec. 5).

352 The model is run for an hour starting from rest and adjusting through a rapid spin-up  
353 phase. Figure 5 shows a snapshot of sea level that shows realistic features: short-crested  
354 waves generated at an angle refract and break, generating rollers, swash and some reflection.

355 CROCO comes with capabilities for water quality, marine ecosystem and sediment mod-  
356 eling. In the present study, some of these capabilities are used with simple settings. First, we  
357 introduce a passive tracer in the swash zone for comparison with dye releases made during  
358 the beach survey. Second, a suspended sediment model (Blaas et al., 2007; Warner et al.,  
359 2008) allows comparison of rip patterns with aerial photos taken during the survey. We

360 use a single fine sand class with settling velocity of 1 cm/s. For resuspension, taking one  
 361 sediment bed layer for simplicity, only two parameters are needed: critical shear stress and  
 362 erosion rate at the seafloor, expressed in the erosion flux (Blaas et al., 2007):

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

$$E = 0 \quad \text{otherwise}$$

363  $\tau_b$  is the bottom shear stress computed by the model.  $E_0$  is an empirical erosion rate set  
 364 to  $10^{-5}$  kg/m<sup>2</sup>/s;  $p$  is the sediment porosity (0.41);  $\tau_{cr}$  is the critical shear stress, i.e., the  
 365 threshold for initiation of sediment motion, set to 0.01 N/m<sup>2</sup>.

#### 366 4.2. Shallow vs. deep breaking (and a Boussinesq model)

367 The Peregrine vorticity generation process only requires short crested waves, with no  
 368 need for unstable longshore currents generated by oblique waves. Boussinesq models are very  
 369 efficient in this process, but an important question for us is whether a 3D model will remain  
 370 so. An essential difference between the two types of model is the depth penetration of wave  
 371 breaking. In a 2D Boussinesq model, deep breaking is implicitly assumed as momentum is  
 372 transferred instantaneously to the depth-averaged flow. However, this is a crude assumption  
 373 as breaking-induced flow and turbulence is produced essentially above trough level (where  
 374 the onshore flow is located), while turbulence at deeper levels depends more on vertical  
 375 diffusion and less on local production (Grasso et al., 2012; van der A et al., 2017). (Uchiyama  
 376 et al., 2010) show that the deep breaking assumption is inconsistent with high-resolution  
 377 cross-shore velocity profiles measured during Duck94. Splinter and Slinn (2003) also suggest  
 378 that transient dynamics produced by deep breaking may collapse in the more realistic case  
 379 of shallow breaking. The results presented here are in agreement with these studies and we  
 380 propose to assess the role of 3D dynamics on surf eddies by comparing simulations forced by  
 381 shallow and deep breaking. Deep breaking will constitute a pseudo-2D model whose results  
 382 can be compared with a Boussinesq model solution (Section 4.5).

383 Boussinesq-type models (see Barthelemy 2004, for a review) are common tools to simulate  
 384 weakly dispersive waves and their transformations from the ocean to the swash zone. Several

385 developments allowed their application to a wide range of scales, from surfzone processes to  
386 ocean basin-scale tsunami propagation (Kirby, 2016). Here, for a verification of our pseudo-  
387 2D CROCO version, we use FUNWAVE-TVD (Shi et al., 2012). It solves the fully nonlinear  
388 Boussinesq equations using a hybrid finite-volume finite-difference scheme. Parametrizations  
389 are similar to CROCO, with wave breaking handled by a shock capturing TVD scheme  
390 (making the need of explicit criterion unnecessary) and a quadratic drag formulation with  
391  $Cd = 0.002$  for bottom friction. Note that newer generation models than FUNWAVE are  
392 available with better dispersive properties — using Green-Naghdi equations (Lannes and  
393 Bonneton, 2009) or incorporating an additional enstrophy equation (Kazakova and Richard,  
394 2019) — but FUNWAVE is widely used and share with this class of models the essential  
395 depth-averaged assumption.

396 To force deep breaking in a 3D wave-resolving model, we can artificially enforce strong,  
397 deep vertical mixing around the breaker zone by multiplying the vertical turbulent viscosity  
398 by 500 ( $\nu_t \sim 1 \text{ m}^2/\text{s}$ , on the vertical only). Figure 6 shows a time and longshore average  
399 of cross-shore and alongshore currents in the case of shallow and deep breaking. Shallow  
400 breaking is computed by the 3D model with no explicit constrain on penetration scale (but  
401 a parametrization of turbulence induced by breakers and currents). It drives a shallow  
402 onshore flow about 40 cm deep and an offshore near-bottom undertow, resulting in strong  
403 vertical shear of about 2 m/s within 1 m depth. With artificially strong vertical viscosity,  
404 momentum in the breaker zone is almost instantly mixed to the bottom and the cross-shore  
405 flow is reduced to nearly zero as required by mass-conservation (the residual flow is the  
406 Stokes drift compensation flow), and consistent with depth-averaged models. The longshore  
407 flow driven by oblique waves and a number of other simulations with shallow and deep  
408 breaking (Tab. 2) will be analyzed and compared in the following sections.

#### 409 *4.3. Reference simulations and comparison with data*

410 To introduce the 3D processes of flash rip generation, we present simulations with shal-  
411 low and deep breaking (3D\_SC\_D10 and 2D\_SC\_D10), representing mid-tides conditions on  
412 March 13 2014 in Grand Popo Beach. Figure 7 compares the vertical vorticity of surface

413 flow:  $\omega_z = \frac{\partial v_s}{\partial x} - \frac{\partial u_s}{\partial y}$ , averaged over two peak periods (22 s). With deep breaking (left  
414 panel), the vortical field is rich with large filaments and surf eddies of 50-100 m scale that  
415 are generated from short-crested waves, similar to Boussinesq model solutions. However,  
416 the full 3D model with shallow breaking offers a radically new solution (Fig. 7, right panel).  
417 Some of the large-scale fluctuations are present but over-shadowed by shorter scales. This  
418 mode presents itself as rib structures (or mini ribs following a relevant observation by Short  
419 et al. 1993, already mentioned) with short longshore wavelength of order 5-10 m and periods  
420 less than 5 min.

421 We now compare the two simulations with data collected during the survey of March  
422 2014. Flash ribs did not strike the survey team as particularly imposing (Castelle et al.,  
423 2014; Scott et al., 2018; Floc'h et al., 2018). Dye experiments revealed filament generation,  
424 but of relatively short scales. The survey was not extensive in terms of spatial resolution but  
425 ADV and dye observations suggested a dynamical regime much closer to 3D shallow-breaking  
426 than pseudo-2D deep-breaking simulations, as will be seen.

427 The dye experiment presented in Figure 8 illustrates both the structure of the alongshore  
428 flow and scales of flash ribs emerging from the surfzone. Careful analysis of observed versus  
429 modeled dye evolution, as in Hally-Rosendahl and Feddersen (2016), is beyond the scope of  
430 this paper, but useful information can be gained from a simpler analysis. Figure 8 presents  
431 two consecutive aerial photos (at 154 s interval), and the corresponding snapshots of tracer  
432 simulations with the full 3D model. We do not expect an exact match between observed and  
433 modeled ribs, considering the chaotic nature of these phenomena, but scales and structures  
434 are meaningful. During the survey, a few attempts of dye release were made before obtaining  
435 a clear filament patch. Similarly for the model, we selected one occurrence among few tracer  
436 patches initialized at regular interval along the coast (in the swash zone). Even though all  
437 tracer patches eventually ended up with similar V-shape and similar scales of evolution, there  
438 was variability in the evolution and we selected the most visually comparable filament with  
439 Drone photos. In both cases, a thin filament of 5-10 m expands quickly seaward at a speed of  
440 about 0.5-1 m/s, reaching about 70 m from shore. Using sequential photos of the tracer from  
441 the Drone camera, it is also possible to extract a simplified cross-shore profile of longshore

442 drift velocities (see Derian and Almar 2017 for more extensive Lagrangian calculations).  
443 The result is presented in Figure 9 together with the ADV measurement of mean longshore  
444 current over the terrace and model solutions with deep and shallow breaking. The estimated  
445 longshore flow has an asymmetric V-shape very similar to the full 3D model solution with a  
446 peak velocity of about 0.4 m/s in the inner surfzone, and error bar of about 0.1 m/s (Derian  
447 and Almar, 2017). Deep breaking solutions have a more symmetric profile centered in the  
448 outer surfzone. The profiles in both simulations are a result of cross-shore advection. With  
449 deep breaking, advection is weak and the longshore flow remains centered in the breaker  
450 zone (Fig. 6). This is also a common bias of depth-averaged models (Larson et al., 2002).

451 Another qualitative comparison of patterns can be made, looking at surfzone suspended  
452 sediments in the aerial photo (Fig. 10, left panel). The contrasts in the photo is enhanced  
453 to better expose suspended sediments (brown color), which is seen weakly extending beyond  
454 the surfzone. Snapshots of the model's surface sediment concentration is also shown after  
455 15 min of simulation. With shallow breaking (center panel), sediments tend to resuspend  
456 in the breaker zone and mix efficiently within the surf zone, but only weakly extend to the  
457 innershelf. The rib structure is apparent at the seaward front of sediment concentration.  
458 It is also apparent in alongshore convergence lines that have a structure similar to the  
459 foam lines in the aerial photo. Overall, the patterns are very similar to the observations,  
460 particularly in the upper part of the photo, where there is less foam or sunglint. The same  
461 suspended sediment simulation with deep breaking gives very different results (Fig. 10,  
462 right panel). Resuspension is now maximum in the inner surfzone (as for eddy energy; Sec.  
463 4.5). Filaments and eddies are more coherent, larger, their growth slower, but they extend  
464 much further seaward. Mixing in the surfzone is less efficient than for the shallow-breaking  
465 case, but shelf-surf exchange is more intense, due to filament extension. This more coherent  
466 pattern may have aesthetic values but does not fit visual observations of Grand Popo Beach.

467 For a more quantitative local comparison, we used ADV measurements, looking at hor-  
468 izontal  $u_h = \sqrt{u^2 + v^2}$  and vertical  $w$  velocities, the latter being a good indicator of 3D  
469 motions. Figure 11 presents Power Spectral Density (PSD), using Welch's method of noise  
470 reduction, of  $u_h$  and  $w$  velocity fluctuations for the model and the ADV data at a similar

471 location in the middle of the terrace. The short-wave spectrum around the peak period (11  
472 s) is well represented considering that the model wave maker uses a JONSWAP approxima-  
473 tion. At lower frequency in both  $u_h$  and  $w$  PSD, a good fit with data is given by the full 3D  
474 model, while the deep-breaking simulation has a clear energy deficit (valid at 95% confidence  
475 level). Looking at  $u_h$  PSD, there is a band of high energy around 30-100s period that is  
476 clearly missing in the pseudo-2D case. This frequency is consistent with visual inspection  
477 of rip structure oscillation in animated vorticity fields. However, the difference between 3D  
478 and pseudo-2D cases can be seen at even lower frequency. This is best illustrated by  $w$  PSD  
479 and will be shown to result from nonlinear upscaling (Sec. 4.6). Interestingly, the energy  
480 range of 3D eddies encroaches on that of infragravity waves, making it hard to separate the  
481 two dynamics from observations alone.

482 On the innershelf, about 80 m offshore where unfortunately no data is available, the  
483 comparison between model simulations shows a different relation at very low frequency, for  
484 time scales between 3 and 15 min (Fig. 12). In this range, eddies produced by short-crested  
485 waves have more energy in the deep-breaking case, as expected from visual inspection of  
486 vorticity and suspended sediments. We conclude from this section that observations in  
487 Grand Popo Beach are in better agreement with a full 3D solution of surf eddies, which  
488 presents a radically different dynamical regime. We will now analyze more closely the 3D  
489 structure and generation process of rib structures.

#### 490 4.4. Structure and production of vertical shear instability

491 The surfzone eddy variability seen in the 3D model solution is truly three-dimensional.  
492 Vertical vorticity is only one manifestation, but horizontal vorticity is the key player. For  
493 better detection, we use a higher-resolution simulation ( $dx = dy = 0.5$  m; run for 15 minutes)  
494 that gives a very similar solution to the lower resolution case. Figure 13 presents the  $Q$  field  
495 defined by  $Q = -\frac{1}{2} \frac{\partial u_i}{\partial x_j} \frac{\partial u_j}{\partial x_i}$ , using Einstein summation convention over the three dimensions.  
496  $Q$  is commonly used to enhance detection of vortical flows. Here, we split cross-shore and  
497 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  $Q_x = -\frac{\partial v}{\partial z} \frac{\partial w}{\partial y} - \frac{1}{2} \frac{\partial v^2}{\partial y}$  in green),  
498 then normalize them and only plot positive isosurface values 0.02 for clarity (negative values

499 give counter-rotating features).

500 The result is strikingly consistent with instabilities of a transitional mixing layer (Met-  
501 calfe et al., 1987; Lesieur, 1990), the transition being constrained by surfzone width.  $Q_y$   
502 shows spanwise "rolls" created from the primary instability while  $Q_x$  identify streamwise  
503 "ribs" that are transverse (counter-rotating) vortices from secondary instability (assumingly  
504 growing from perturbations generated between the rolls, in the braid region). Note that  
505 streamwise designate the shear direction, which is cross-shore rather than (oblique) wave  
506 direction. The reason is that the undertow aligns with surface flow in the alongshore di-  
507 rection, forming only weak mean longshore vertical shear. (Fig. 6). However, the rips can  
508 take an oblique direction as they are advected by the mean longshore flow. They extend  
509 seaward beyond the surfzone while stretching in the vertical direction. Filaments of verti-  
510 cal vorticity (or  $Q_z$ , not shown) also have similar rib structures to  $Q_x$ , recognisable in the  
511 surface vorticity plot of Figure 7.

512 The mean shear flow is composed of breaking-induced mean onshore flow and associated  
513 seaward undertow (Fig. 6). The inflected velocity profile is inviscidly unstable to small per-  
514 turbations and unstable modes of Kelvin-Helmholtz type can emerge. The spanwise rolls are  
515 large vertical eddies that rapidly evolve into transverse (streamwise) rib vortices connected  
516 by "braid" regions, and stretched seaward and downward. This picture is reminiscent of  
517 descriptions based on laboratory-scale measurements and simulations (Nadaoka et al., 1989;  
518 Watanabe et al., 2005; Kirby and Derakhti, 2019) — although here no air mixing by bubbles  
519 is required for the secondary instability to occur.

520 According to linear stability analyses, the wavelength of primary shear instability setting  
521 the distance between spanwise rolls is an order of magnitude larger than the mixing layer  
522 width  $\delta$  — in free shear layers (Michalke, 1964, 1965), wavelength, frequency and growth  
523 rate of the most unstable modes are  $14 \delta$ ,  $0.015 U/\delta$  and  $0.1 U/\delta$  respectively). The sec-  
524 ondary instability wavelength is of the same order as that of the primary instability ( $2/3$   
525 in Pierrehumbert and Widnall 1982). If the mixing layer width is taken as the vorticity  
526 thickness  $\delta = \Delta U / [\frac{\partial u}{\partial z}]_{max} \sim 60$  cm, then the wavelength of both roll and rib structures is  
527 5-10 m, consistent with our simulation. Note that given an effective resolution of 5-10  $dx$  for

528 CROCO (Soufflet et al., 2016), a simulation with 1-m grid resolution can only be considered  
 529 "eddy-permitting" for 3D instability in the present conditions. As for frequency, the mixing  
 530 layer size would be associated with modes of 0.01-0.04 Hz (25-100 s period), a range usually  
 531 reserved to surf beat. The model with 3D instability has energy in this range that is lacking  
 532 in the deep breaking case (see previous section and Fig. 11), but it has excess energy at  
 533 lower frequency as well, which may result from upscaling through nonlinear interactions (the  
 534 same spectrum is produced with monochromatic wave forcing; see also Sec. 4.6).

535 In order to isolate the mechanism of eddy-mean flow interaction, we analyze a solution  
 536 forced with monochromatic, shore-normal, long-crested waves (3D\_MONO\_D0 in Tab. 2).  
 537 This simulation has constant wave forcing in both space and time when averaging over  
 538 the wave period (11 s). In this case, the same rib structure is generated (Fig. 14) but  
 539 without the large-scale alongshore variation seen in the full solution (Fig. 7, right panel).  
 540 A comparison of this simulation with other test cases will be presented in the next section.  
 541 Here, we analyse the mechanism and patterns of shear production. Figure 15 shows EKE  
 542 source terms  $-\overline{u'_i u'_j \frac{\partial \bar{u}_i}{\partial x_j}}$ , that results from eddy-mean flow interaction, i.e., the energy spent  
 543 by the mean flow to feed rolls and ribs. The largest term is vertical shear production  
 544  $-\overline{u'w' \frac{\partial \bar{u}}{\partial z}}$  (Metcalf et al., 1987), with some lesser contribution from cross-shore convergence  
 545  $-\overline{u'u' \frac{\partial \bar{u}}{\partial x}}$  (all other 7 combinations are negligible). The main production site is in the breaker  
 546 zone. A maximum value is located just beneath the inflection point in the mean velocity  
 547 profile (maximum shear represented in Fig. 15 by a thick blue line), consistent with mixing  
 548 layer instability. The figure also presents a cross-section of mean 3D eddy kinetic energy  
 549 ( $EKE = \frac{1}{2}(u'^2 + v'^2 + w'^2)$ , where  $u', v', w'$  are fluctuation velocities to the time-mean  
 550 flow presented in Fig. 6). EKE and shear production have a very similar spatial pattern,  
 551 although high EKE values extend from the production center in all directions: surface,  
 552 bottom, shoreward and seaward. The primary rolls are thus produced in the outer surfzone  
 553 and advected by the mean and eddy flow towards both the inner surfzone and innershelf.  
 554 EKE transport is stronger at the surface, so that the streamwise filaments extend farther  
 555 offshore at the surface than at the bottom (despite some amount of vertical stretching as  
 556 they leave the terrace).



557 *4.5. Short-crested waves and the Peregrine process*

558 An important question of our study concerns the effect of wave variations (frequency and  
559 directional spreading) on flash rip generation in a full 3D model. To address this question, it  
560 is useful to simplify the problem and progressively add the multiple conditions of variability.  
561 In this section, we analyse shore-normal, short-crested wave simulations with shallow or deep  
562 breaking (3D\_SC\_D0 or 2D\_SC\_D0). Shore-normal conditions prevents the formation of a  
563 longshore current and associated horizontal shear instability. In addition, we look at long-  
564 crested wave solutions to isolate the effect of 3D instabilities, i.e., the monochromatic solution  
565 (3D\_MONO\_D0), presented in the previous section, and a similar case with JONSWAP  
566 frequency spectrum (3D\_LC\_D0). This latter comparison will help evaluate the effect of  
567 frequency spreading on eddy variability before addressing the effect of directional spreading.

568 Figure 16 compares vertical vorticity for shore-normal, short-crested wave cases. As for  
569 oblique waves, deep-breaking leads to a rich vortical field with large filaments extending  
570 far offshore, similarly to Boussinesq models. This is confirmed here with a comparison  
571 between FUNWAVE-TVD and pseudo-2D CROCO applied to the same configuration. A  
572 difference between Boussinesq and pseudo-2D solutions is the effect of 3D dynamics over the  
573 innershelf in the latter case, where surface-intensified offshore eddies and filaments present a  
574 more fragmented aspect due to forward energy cascade (Uchiyama et al., 2017; McWilliams  
575 et al., 2018). However, the full 3D nonhydrostatic model with shallow breaking (Fig. 16,  
576 right panel) shows patterns radically different from both Boussinesq and pseudo-2D solutions  
577 with regular shorter-scale rib structures that present a more limited cross-shore extension.

578 Figure 17 presents EKE cross-sections for all shore-normal wave experiments. A strik-  
579 ing element of these figures is the presence of large surface and bottom EKE in the shal-  
580 low breaking cases. This pattern is not a result of shear production and is absent from  
581 the monochromatic case (see previous section and Fig. 15). Therefore, it can only result  
582 from wave groups associated with frequency spreading. Through wave height modulation,  
583 wave groups produce variability in the breaking-induced surface onshore flow and associated  
584 undertow. The variability amounts to about half the integrated mean EKE. However, it  
585 is much reduced in the deep-breaking case, consistent with depth-averaged model studies

586 (de Schipper et al., 2014).

587 We now turn to the effect of directional spreading. In Figure 17, short-crested waves  
588 (produced by directional spreading) extend EKE production over a wider surf zone than long-  
589 crested waves, where EKE is confined to the breaker zone. However, the seaward extension  
590 is much larger in the pseudo-2D model, confirming the impression made from vorticity  
591 inspection. Further confirmation is given by vertical EKE integration (and normalization  
592 by mean depth; Fig. 18). It highlights 3 regions of eddy energy: the inner and outer  
593 surfzone and innershelf. The top panel presents unfiltered data. In this case, 3D instability  
594 and wave group forcing dominates eddy production in the outer surfzone. The most striking  
595 observation is that the deep-breaking solution has a much larger cross-shore expansion,  
596 stretching shoreward with a maximum in the inner surfzone and also 100 m seaward over  
597 the innershelf. The difference of EKE profiles between deep and shallow breaking cases is  
598 reminiscent of reported dye mixing overestimation by Boussinesq models (e.g., see profiles  
599 of dye concentration in Fig. 10 of Hally-Rosendahl and Feddersen 2016).

600 Short-crested waves do not seem to affect 3D instability, while, on the contrary, shallow-  
601 breaking largely affects eddy production associated with the Peregrine process. This is  
602 particularly true for seaward extension. This is even clearer using a low-pass filter on velocity  
603 fluctuations, removing a large part of variability from 3D instability and wave group forcing  
604 (bottom panel of Fig. 18). What remains is closer to the usual definition of surfzone eddies  
605 as Very Low Frequency features. They are much more energetic in the deep breaking case  
606 (up to 3 times more than shallow breaking), showing a clearer EKE maximum in the inner  
607 surfzone and a large patch over the innershelf. This result is consistent with the findings from  
608 previous Boussinesq model studies that filaments forced by short-crested waves originate in  
609 the inner surfzone, then spread offshore, forming eddies that grow in scale (Johnson and  
610 Pattiaratchi, 2006).

#### 611 *4.6. 2D and 3D surfzone turbulent cascade*

612 If there is energy produced by short-crested waves in the inner surfzone of the shallow  
613 breaking case, the question is why does it not produce large filaments, as in the deep break-

614 ing case. We found an answer in computing the spectrum energy flux. In 2D turbulence, the  
615 flux of energy is negative and small fluctuations can grow into larger coherent structures.  
616 This process is usually involved to explain the growth of filaments and eddies from variable  
617 wave forcing (e.g., Johnson and Pattiaratchi 2006; Feddersen 2014).

618 To check its validity in a 3D model, we performed a wavenumber spectral flux analysis  
619 for pseudo-2D and 3D simulations 2D\_SC\_D0 and 3D\_SC\_D0 (Fig. 19). The spectral flux  
620 is computed as in Marchesiello et al. (2011) by spectral integration of  $v$  advection term.  
621 Consistently with 2D turbulence there is a strong inverse cascade of kinetic energy (negative  
622 flux) in the pseudo-2D model starting from the scale of injection corresponding to short-  
623 crested wave forcing (wavelength of 30-40 m here), and there is no direct cascade towards  
624 smaller scales. In the 3D case, the turbulent regime is very different. The negative flux of  
625 energy produced by variable wave forcing is present but considerably reduced. In addition,  
626 there is a second injection at smaller scales that corresponds to the most unstable mode of  
627 3D instability (wavelength of 7-10 m). This small-scale energy travels both backward and  
628 forward across the spectrum and thus widens the range of variability associated with 3D  
629 instability. In the 3D long-crested wave case 3D\_LC\_D0, the same spectral flux is produced  
630 at small scales but there is no large-scale inverse cascade due to missing injection by short-  
631 crested waves.

632 This analysis confirms that the growth of filaments and eddies associated with the Pere-  
633 grine process heavily relies on a 2D inverse cascade, but this cascade nearly collapses when  
634 extending to 3D dynamics. In this case, vorticity fluctuations generated by waves with  
635 a finite alongshore extent can hardly turn into coherent structures, as they do in the 2D  
636 paradigm built on depth-averaged models (Johnson and Pattiaratchi, 2006). The shallow  
637 water approximation is therefore too restrictive an assumption for nearshore flows.

#### 638 *4.7. Oblique waves and horizontal shear instability*

639 The precedent experiments with shore-normal waves show that shallow breaking tends to  
640 hinder the generation of large 2D eddies by short-crested waves, while sustaining forced and  
641 intrinsic 3D surfzone eddy dynamics. We now address the case of oblique waves by increasing

642 obliquity to  $D = 20^\circ$  in order to generate longshore currents with larger horizontal shear.  
643 From linear stability analysis of a 2D problem (Bowen and Holman, 1989), the wavelength,  
644 frequency and growth rate of the most unstable shear waves are  $\lambda_h = 2.5L$  and  $f_h = 0.07\frac{V}{L}$ ,  
645  $\sigma_h = 0.15\frac{V}{L}$  where  $V$  is the longshore current magnitude and  $L$  is the longshore current  
646 half-width (outer shear). For a narrow, shoreline-intensified jet typical of Grand Popo at  
647 mid-tide (Almar et al., 2014, 2015b), shear can be strong ( $\sim 0.05 \text{ s}^{-1}$ ), implying a minimum  
648 shear wave period of 5 min, wavelength of 80 m, and growth time  $\sigma_h^{-1} = 3 \text{ min}$ .

649 Forcing long-crested waves ( $\sigma_\theta = 0$ ) with imposed deep breaking (2D\_LC\_D20), CROCO  
650 recovers results that are typical of wave-averaged shallow-water models (or 3D models with  
651 deep breaking as in Marchesiello et al. 2016, or Kumar and Feddersen 2017). Figure 20 (left  
652 panel) shows an active horizontal shear instability producing shear waves with wavelength  
653 consistent with linear theory. Shear waves propagate with the longshore current as they  
654 become nonlinearly unstable, generating filaments and eddies that extend offshore. When  
655 both horizontal shear instability and short-crested wave vorticity generation are active (with  
656 deep breaking), the solution has even more energy (Fig. 20, center panel). However, with  
657 shallow breaking, the horizontal shear instability appears much weaker and is replaced again  
658 by rib structures associated with vertical shear instability. Both processes appear reduced  
659 by vertical shear.

660 Inspection of the cross-shore profile of mean longshore currents (Fig. 6) may help to un-  
661 derstand how shallow breaking undermines horizontal shear instability. With deep breaking,  
662 cross-shore advection is inactive and the longshore current remains trapped over the terrace  
663 slope (which is steep in Grand Popo, around 1/10), and his outer shear is strong. With shal-  
664 low breaking, however, the longshore current is advected by the cross-shore recirculation,  
665 stretching its profile across the terrace in a way that minimizes the outer shear. Then, the  
666 instability growth rate becomes too weak to overcome friction (from turbulence or bottom  
667 drag). This process may explain why, for example, Newberger and Allen (2007) did not  
668 found horizontal shear instability in their 3D wave-averaged model solutions as opposed to  
669 many previous 2D modeling studies (Allen et al., 1996; Slinn et al., 1998; Uchiyama et al.,  
670 2009).

671 This process is verified in Figure 21, showing cross-sections of time-mean and longshore-  
672 mean eddy and mean flow features for the four cases with oblique waves. Horizontal shear  
673 instability is best assessed with the eddy-mean flow interaction term  $-\overline{u'v'}\frac{\partial \bar{v}}{\partial x}$ . In all cases,  
674 shear production is clearly correlated with outer and inner slopes of the mean longshore  
675 current, with stronger production in the outer shear. Deep-breaking cases show strong  
676 shear production and large EKE patches centered over the terrace outer slope, where the  
677 current shear is strongest. In the case of horizontal shear instability alone (2D\_LC\_D20), the  
678 magnitude is lower than that of 2D\_SC\_D20, despite similar shear intensity. This indicates an  
679 amplification of shear instability by short crested waves as they drive transient intensification  
680 of longshore currents.

681 When comparing with shore-normal wave simulations of deep breaking cases, offshore  
682 EKE seems nearly a linear combination of effects from short-crested waves and horizontal  
683 shear instability. Offshore EKE extends further than predicted from shear production due to  
684 mean and eddy advection, but short-crested waves clearly provide the most efficient process  
685 for innershelf eddy activity.

686 The shallow-breaking cases (3D\_LC\_D20 and 3D\_SC\_D20) also features shear production,  
687 but within a narrower zone and in shallower water, better controlled by friction. As a result,  
688 EKE is significantly reduced. Short-crested waves (3D\_SC\_D20) appear to amplify the inner  
689 surfzone energy (compared with 3D\_LC\_D20), but in both cases offshore energy is weak.  
690 Over the innershelf, the EKE magnitude can be reduced by a factor 4-5.

## 691 5. Discussion and conclusion

692 Flash rips and surfzone eddies are traditionally conceived within a depth-averaged frame-  
693 work that involves intrinsic horizontal shear instabilities or/and direct short-crested wave  
694 vorticity generation. They are revisited in this study using a 3D nonhydrostatic wave-  
695 resolving model applied to a natural beach with ideal longshore-uniform topography (Grand  
696 Popo Beach, Benin). We first presented a quick overview of a new free-surface, compressible  
697 approach adapted to wave-resolved nearshore dynamics. Its ability to simulate surface grav-  
698 ity wave propagation and nearshore breaking is validated against laboratory experiment.

699 Then, the model is applied to Grand Popo Beach nearshore circulation generated by waves  
700 with frequency and directional spreading. We assume based on comparison with Boussinesq  
701 solutions that the essential difference between 2D and 3D models reduces to the vertical pro-  
702 file of breaking acceleration, i.e., deep or shallow breaking. This allows a direct comparison  
703 of 2D and 3D frameworks within the same model equations and model setup.

704 The generation of transient rips by the 3D model is shown to differ considerably from  
705 that produced by depth-averaged models, owing to the vertical structure of currents pro-  
706 duced by shallow breaking. Processes of both horizontal shear instability and short-crested  
707 wave breaking are restricted in a 3D model by the cross-shore vertical recirculation that  
708 prevent a strong inverse cascade. Variable wave forcing (in space and time) tends to in-  
709 crease flow variability in the surfzone (especially at the surface and bottom), but it does not  
710 translate into large-scale rips streaming far offshore. Usual 2D mechanisms are replaced by  
711 a Kelvin-Helmholtz-type instability, generating rib structures with spanwise and streamwise  
712 (alongshore and cross-shore) vorticity of intermediate scale between turbulence and large  
713 horizontal eddies — periods of 30 s to a few minutes encroaching on the infragravity wave  
714 range and wavelength around 5-10 m. Streamwise filaments extend beyond the surfzone but  
715 with lower intensity than usual depth-averaged model rips (offshore mean EKE is reduced  
716 by 50 percent). Energy spectra and visual inspections of tracer and sediment concentration  
717 shows a much better fit of the 3D model patterns to observational data in Grand Popo Beach.  
718 Our results thus question the qualitative and quantitative relevance of depth-averaged mod-  
719 els for nearshore dynamics. They may explain in particular some overestimation by these  
720 models of surf-shelf exchange (Spydell and Feddersen, 2009; Hally-Rosendahl and Feddersen,  
721 2016) and very low frequency (VLF) variability (Feddersen et al., 2011).

722 These results are representative of mid-tide conditions of a low tide terrace with moderate  
723 wave heights. Interestingly, the rib structures that are described here are comparable to the  
724 *mini rips* described for similar conditions by Short et al. (1993): *Under typical mid-tide*  
725 *conditions, with waves breaking across the bar, a low 'friendly' surf zone is produced. Waves*  
726 *are less than 1 m and most water appears to head toward the shore. In fact it is return*  
727 *seaward also, both by reflection of the beach face and via the mini rips, even if no rip*

728 *channel are present. The rips, however, are usually weak, ephemeral and shallow.*

729 In order to assess the ubiquity of mini rips in the nearshore zone, future 3D studies  
730 should address different nearshore conditions. We assume that the result would be partic-  
731 ularly sensitive to the penetration depth of wave breaking. As breaking-induced flow and  
732 turbulence are mostly produced above trough level, the breaking vertical length-scale would  
733 vary as the root-mean-square wave height  $H_{rms}$  (Uchiyama et al., 2010). In our model, it  
734 is about half of  $H_{rms}$ . Therefore, in case of higher waves, higher penetration would reduce  
735 vertical shear and increase the contribution of horizontal shear instability and short-crested  
736 waves in surfzone eddy generation — as seen in overpublicized photos of large roll-up vortices  
737 developing during high wave events. The penetration depth in our simulation is a function of  
738 subgrid scale turbulence. Changing from Smagorinsky to  $k-\epsilon$  models does not fundamentally  
739 change the results, while increasing the Smagorinsky coefficient from 0.1 to 0.5 (which is an  
740 unrealistic limit, e.g., Canuto and Cheng 1997; Meyers and Sagaut 2006) reduces the vertical  
741 shear and largely damps vertical shear instability — the frictional time  $h^2/\nu_t$  is then close  
742 to the growth time of about 1 minute. In this case, cross-shore advection is reduced and the  
743 wave-averaged longshore flow stays closer to the breaker zone. Horizontal shear instability  
744 is thus stronger, but the variability associated with short-crested waves, while increasing in  
745 the inner surfzone, remains low on the innershelf. Therefore, even with high turbulence,  
746 conventional 2D surf-shelf exchange processes are restrained.

747 Sensitivity to bottom roughness length  $z_0$  is another interesting aspect of our tests. We  
748 increased  $z_0$  from 0.01 to 1 mm. In this case, the drag coefficient goes from about 0.002 to  
749 0.008. Simulations with deep breaking were very sensitive to these values and the largest  
750 roughness would completely shutdown horizontal shear instability due to shorter frictional  
751 time (decreased from 5 to 2 min, i.e., shorter than growth time of about 3 min), and damp  
752 short-crested wave vortical generation. On the contrary, full 3D solutions with shallow  
753 breaking are weakly sensitive to bottom roughness, because of surface intensified currents  
754 and shorter growth time of 3D instability. These tests inspire further confidence in our  
755 results, while they point to an overemphasized importance given to bottom drag in studies  
756 using depth-averaged models (e.g., Allen et al. 1996).

757 Finally, an important question is what exactly in the 3D model counteracts the generation  
 758 of surf eddies by short-crested waves. Vertical vorticity generation by the Peregrine process  
 759 can be written as:

$$\frac{\partial \omega_z}{\partial t} = -\frac{\partial F_{br}}{\partial y_c} \quad (22)$$

760 where  $F_{br}$  is the breaking force, extending to the bottom in a depth-averaged model, and  $y_c$   
 761 is the along-crest direction. Using a breaking parametrization, Clark et al. (2012) propose  
 762 a scaling relation for vorticity generation of a single wave as  $\frac{H_s^3}{h^{2.5}}$ , at a maximum in the  
 763 outer surfzone. However, flash rip generation originates in the inner surfzone in depth-  
 764 averaged models (Johnson and Pattiaratchi, 2006) and in our deep breaking simulations. The  
 765 reason is that adding a transient source of vorticity is not enough to generate a horizontal  
 766 recirculation (with offshore filament). It needs a coastal boundary and an inverse energy  
 767 cascade that transform vorticity fluctuations into larger-scale coherent structures, as shown  
 768 by our spectral flux analysis (Sec. 4.6 and Fig. 19). In a 3D turbulent regime, this cascade  
 769 is much reduced (due to vortex tilting by the shear flow; McWilliams et al. 2018) and the  
 770 variability generated in the inner surfzone does not translate into large rips jetting offshore.  
 771 The inhibition of inverse cascade persist with higher viscosity ( $C_s = 0.5$ ), which however  
 772 damps 3D instabilities. Therefore, tridimensionality is responsible for both counteracting  
 773 the generation of large horizontal eddies, and favoring that of intermediate-scale 3D mini  
 774 rips.

775 In conclusion, our results suggest that nearshore dynamics and transport processes may  
 776 be dominated by nonhydrostatic dynamics, not only for surface gravity waves as is known  
 777 since the 19<sup>th</sup> century, but also for vortical motions. This conclusion should hold for mod-  
 778 erate waves in other applications but further studies should explore the range of waves and  
 779 morphological conditions encountered in the global coastal ocean.

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

Model parameters	Values
Grid type	Horizontal: rectangular, regular mesh; vertical: terrain following grid with free surface
Domain size ( $L_x, L_y$ )	240 x 542 m
Horizontal resolution (dx/dy)	1 m
Vertical grid (Nz)	10 levels (dz=10 cm over the terrace)
Time step	dt=0.025 s, dt <sub>f</sub> =0.001 s, c <sub>sound</sub> =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 forcing (u,v,T,S fluxes)	0
Sediment model (D50, Ws, $E_0$ , $\tau_{cr}$ )	$D50 = 100 \mu\text{m}$ ; $Ws=5$ mm/s; $E_0 = 10^{-5}$ kg/m <sup>2</sup> /s; $\tau_{cr} = 0.01$ N/m <sup>2</sup>
Breaking/turbulence	WENO5-Z shock-capturing advection scheme Smagorinsky SGS model ( $C_S = 0.1$ )
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_SC_D20	shallow	short-crested	20°	3D instab. vs. 2DH instab. vs. Peregrine process
3D_LC_D20	shallow	long-crested	20°	
2D_SC_D20	deep	short-crested	20°	
2D_LC_D20	deep	long-crested	20°	
FUNWAVE-TVD	deep	short-crested	0°	Test of pseudo-2D CROCO

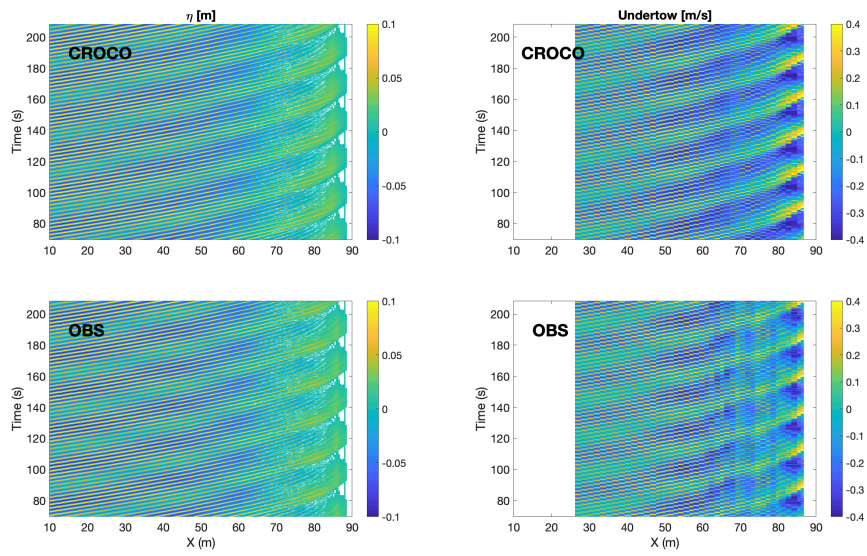


Figure 1: Hovmuller plot (x,time) of data and model sea level  $\eta$  and undertow  $u_b$  for the GLOBEX B3 flume experiment. When data is missing in the measurements, it is also removed from the model output.

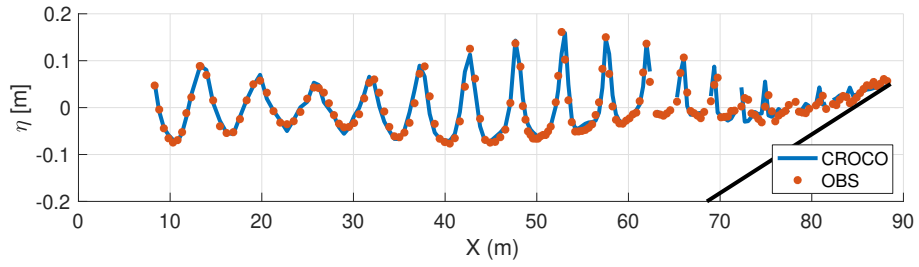


Figure 2: Snapshot of wave field across the GLOBEX B3 flume experiment during runup conditions for model (blue line) and data (red dots).

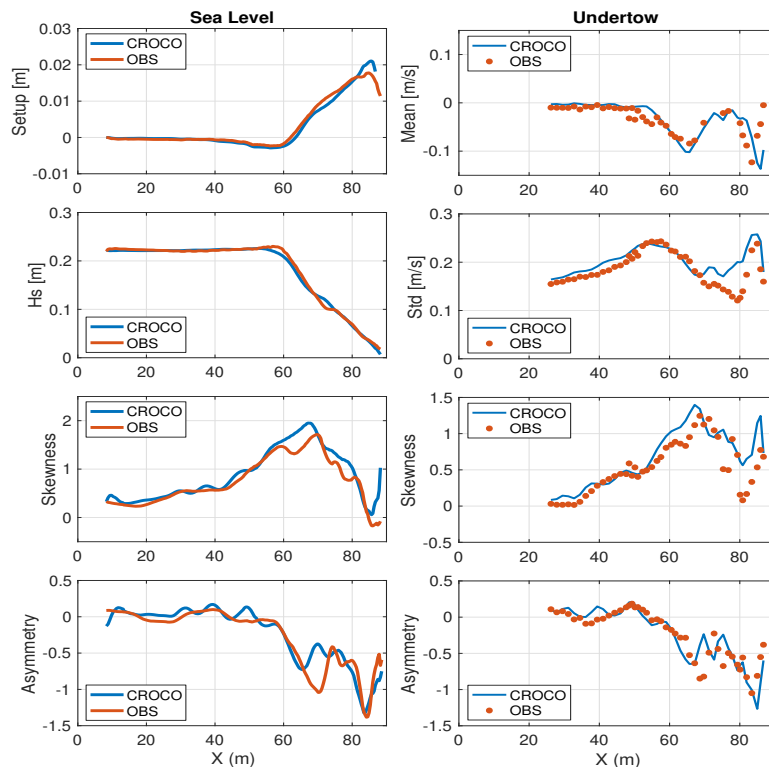


Figure 3: Wave statistics from GLOBEX B3 experiment for sea level  $\eta$  (right) and undertow  $u_b$  (left) in the model (blue line) and data (red line or dots). From top to bottom: mean, standard deviation (or  $H_s$  for  $\eta$ ), skewness and asymmetry.

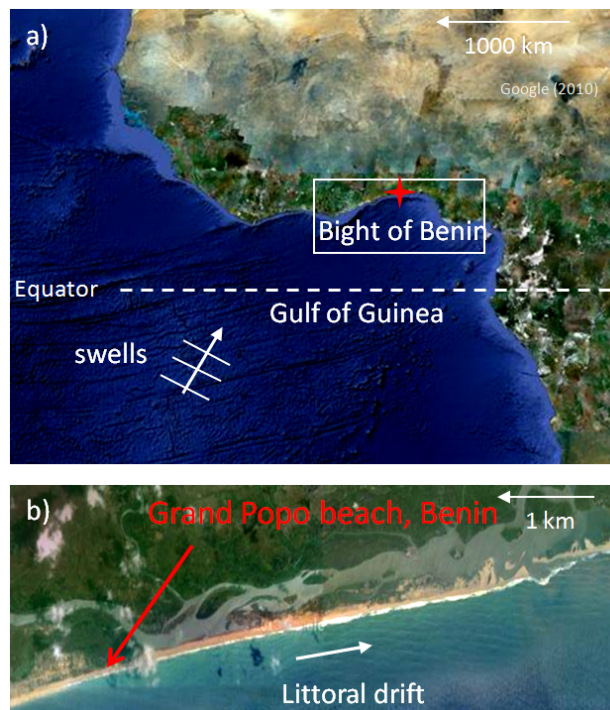


Figure 4: Grand Popo Beach, Benin, is a longshore uniform beach located in the Gulf of Guinea. It is representative of an open, wave-dominated and microtidal environment exposed to S-SW long period swells generated at high latitudes in the South Atlantic.

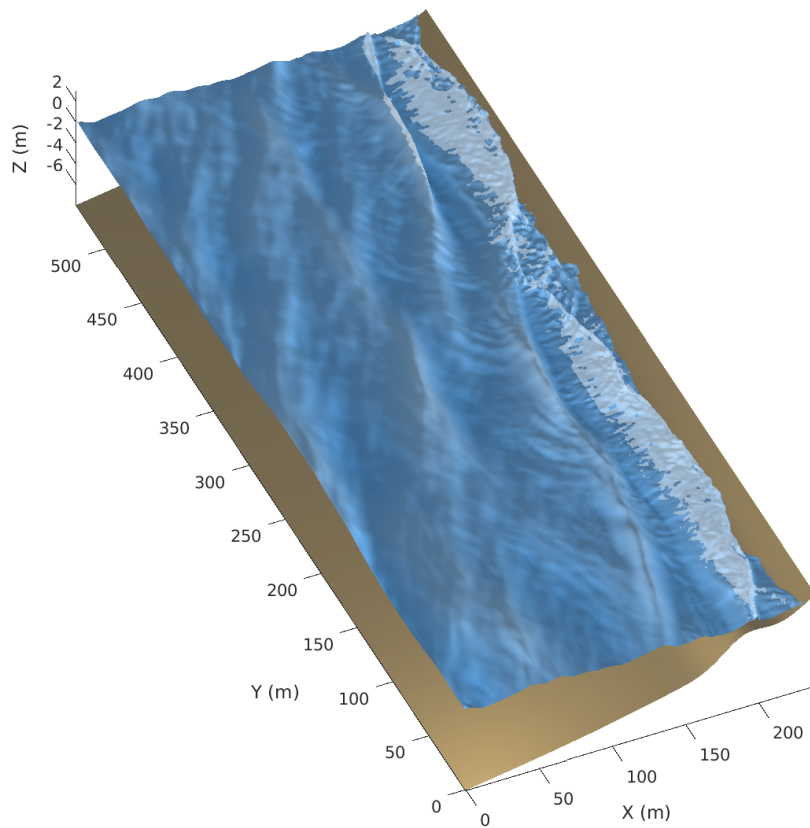


Figure 5: 3D representation of surface waves in CROCO. Short-crested waves are generated at the offshore boundary and propagate shoreward at an angle of  $10^\circ$ , with refraction and breaking through the surf and swash zones. Foam in the surfzone is approximated by white patches in locations of high turbulent kinetic energy of the subgrid-scale model.

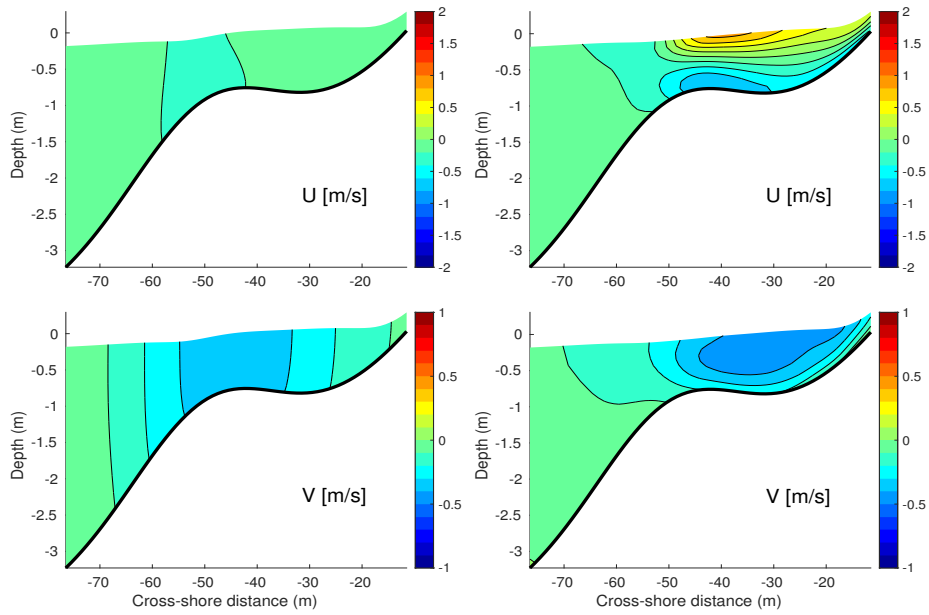


Figure 6: Cross-section of time-mean and longshore-mean, cross-shore (top) and longshore (bottom) currents in the case of shallow (right) and deep (left) breaking.

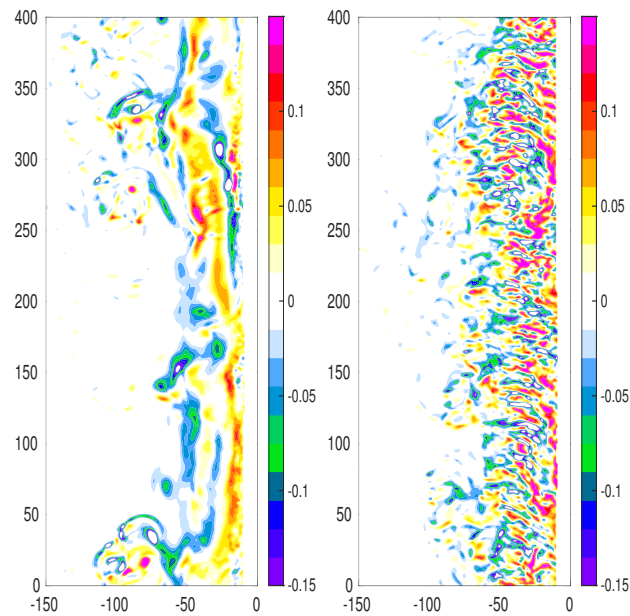


Figure 7: Surface vertical vorticity snapshot in the case of shallow (right) and deep (left) breaking.

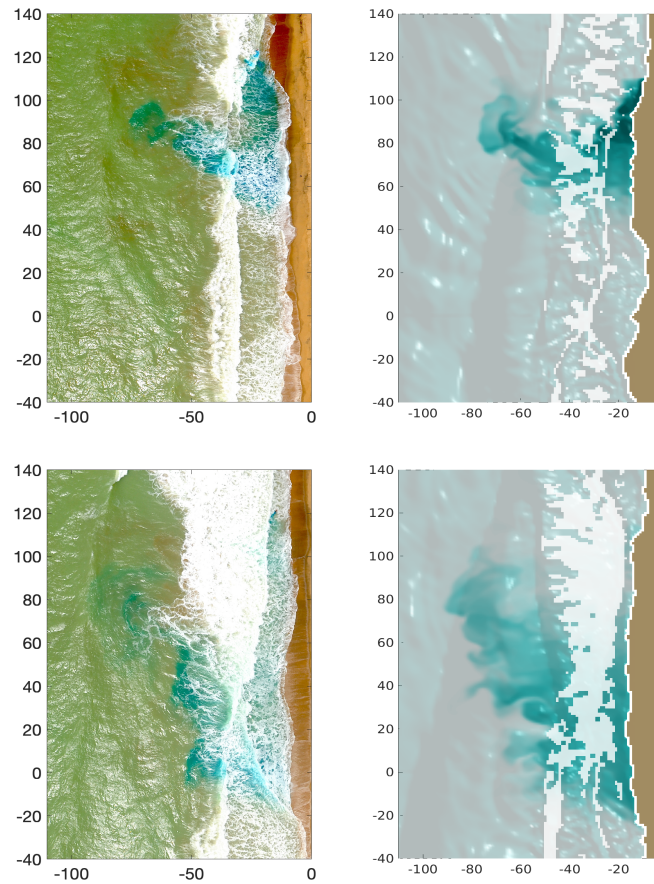


Figure 8: Drone photos (left) and CROCO simulation (right) of dye release at 154 s interval during the Grand Popo survey of March 13 2014 at afternoon mid-tide. In the model, wave surfaces are presented in the background with light from the left (as in the photos), and foam is approximated by high turbulent energy levels as in Fig. 5

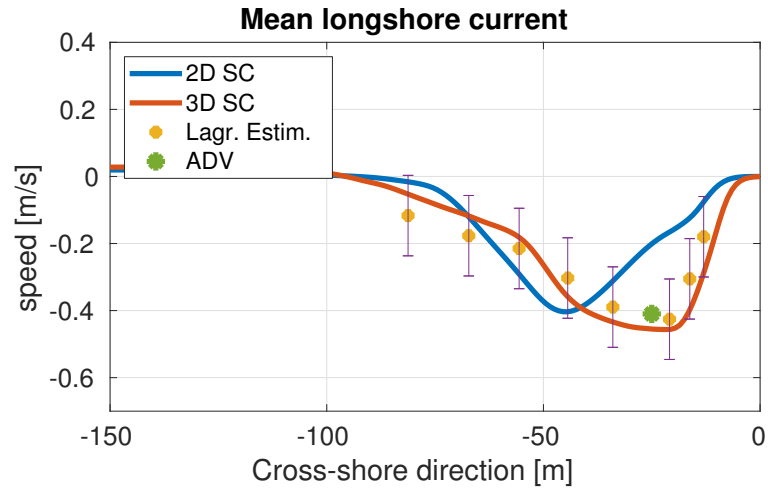


Figure 9: Cross-shore profile of time-mean and longshore-mean surface longshore flow for the full 3D simulation with wave direction  $D = 10^\circ$ . ADV measurement and an estimation of Lagrangian velocities from drone photos (Fig. 8) is added for comparison.

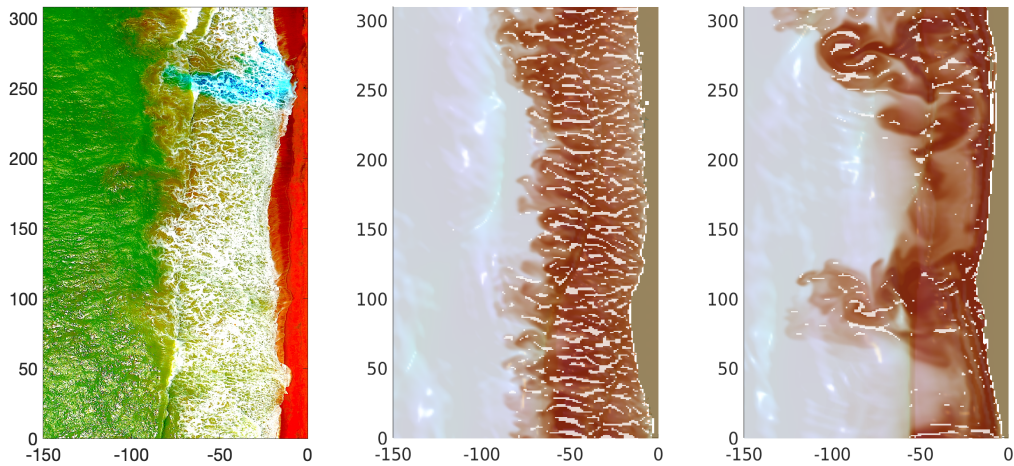


Figure 10: Aerial photo (left) and CROCO simulations of surfzone suspended sediments for shallow-breaking (center) and deep-breaking (right) cases. The contrasts in the photo is enhanced to better expose suspended sediments (brown color). The model suspended sediments correspond to snapshots at 15 min of simulation. Lines of strong alongshore convergence are represented with white patches, to compare with foam structures observed in the photo.



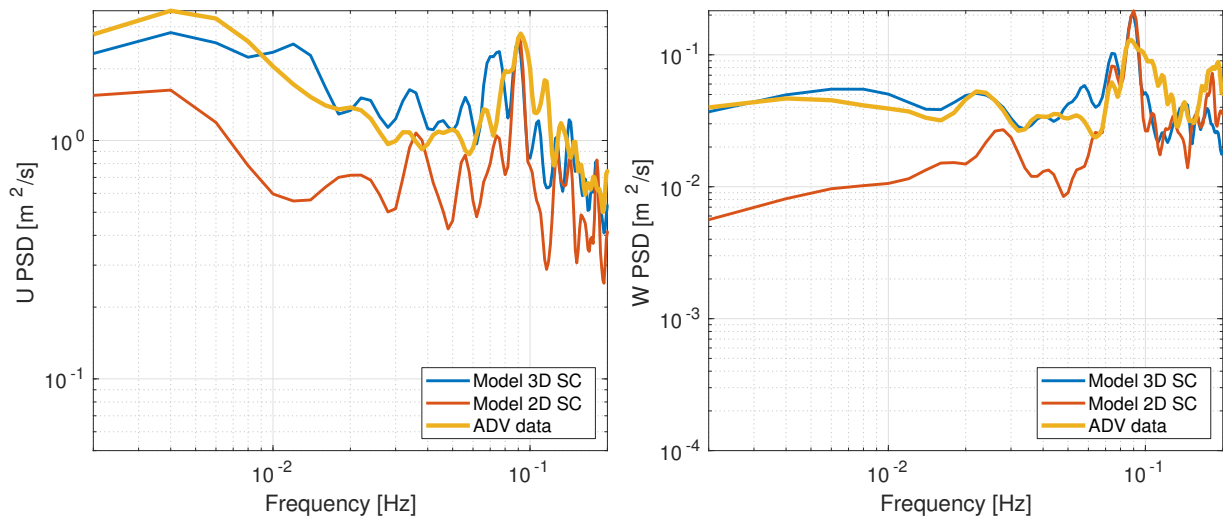


Figure 11: Power Spectral Density of horizontal (left) and vertical (right) velocity fluctuations compared with ADV measurements in the middle of the terrace, in March 13 2014 at afternoon mid-tide.

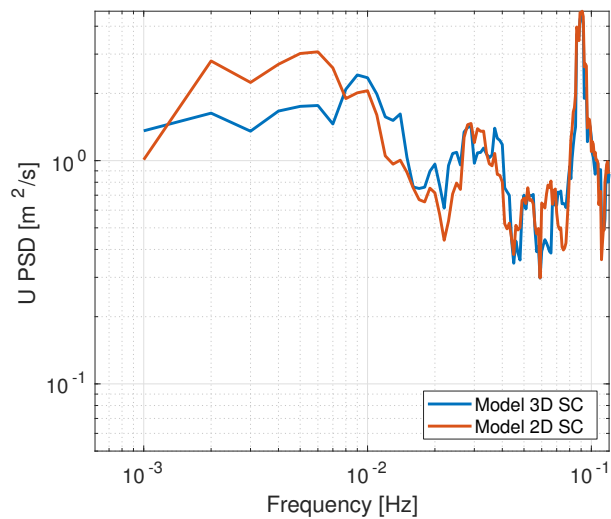


Figure 12: Power Spectral Density of offshore horizontal velocity fluctuations in March 13 2014 at afternoon mid-tide.

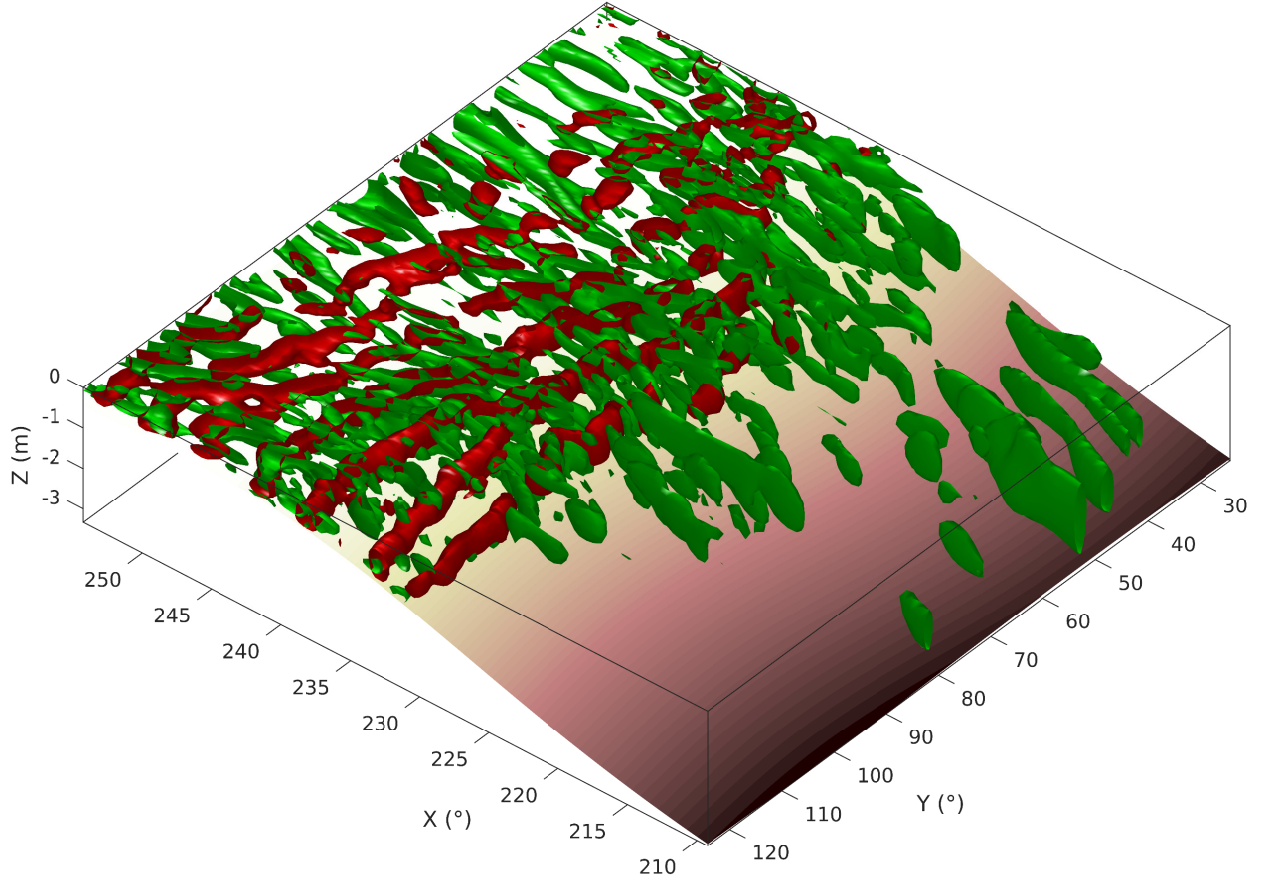


Figure 13:  $Q$  field defined by  $Q = -\frac{1}{2} \frac{\partial u_i}{\partial x_j} \frac{\partial u_j}{\partial x_i}$ , showing coherent structures similar to rolls and ribs in a transitional mixing layer. Cross-shore and alongshore  $Q$  terms are split: spanwise rolls (aligned across shear direction) 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.

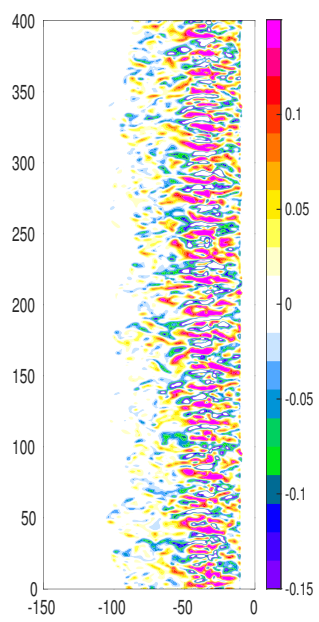


Figure 14: Surface vertical vorticity snapshot for the case with monochromatic shore-normal wave forcing (3D\_MONO\_D0)

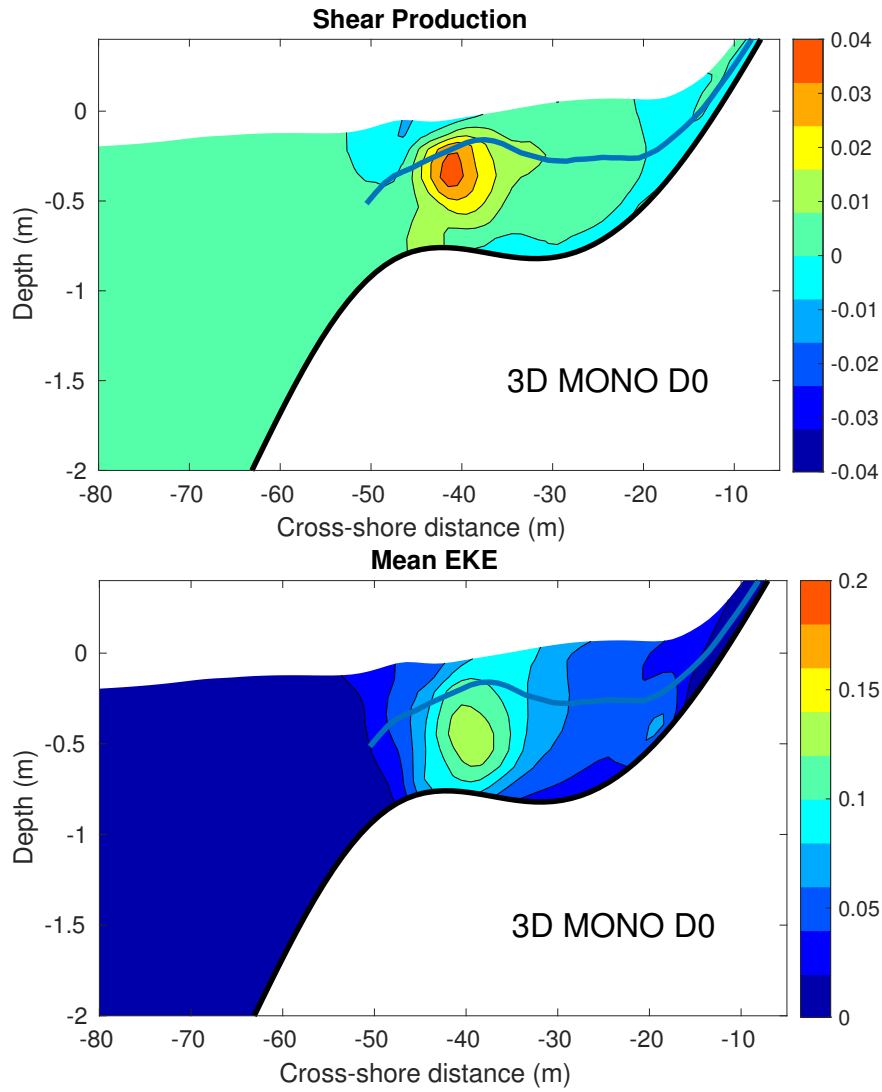


Figure 15: Cross-section of EKE [ $m^2/s^3$ ] (bottom) and EKE production by the mean shear flow (top) [ $m^2/s^3$ ] for the case with 3D instabilities only, forced with monochromatic long-crested waves (3DMONOD0). The position of shear flow inflection point is drawn in a blue line

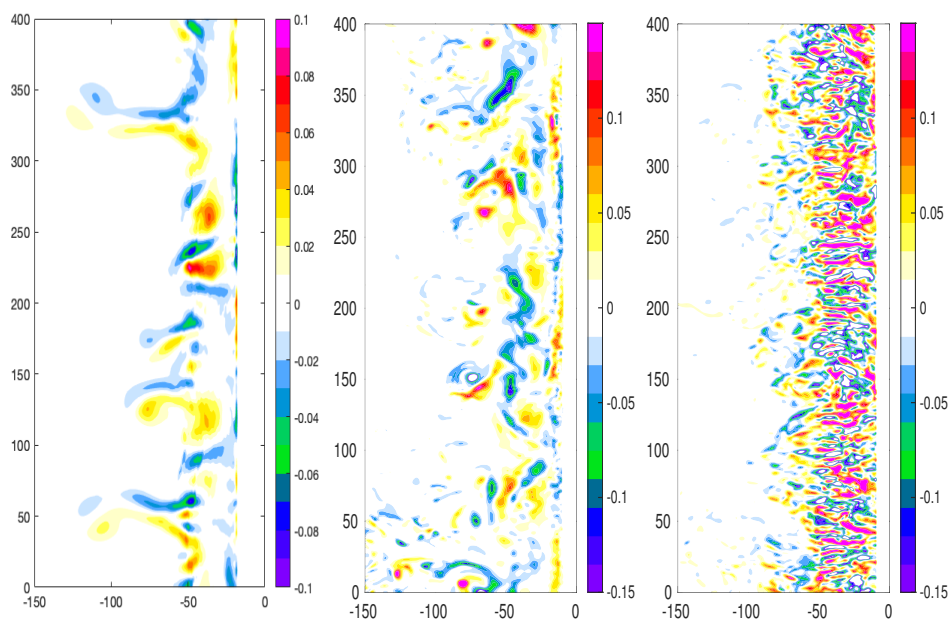


Figure 16: Vertical vorticity snapshot for 3 cases with shore-normal wave forcing: FUNWAVE-TVD (2D Boussinesq) with short-crested waves (left); CROCO with short-crested waves and deep breaking (center); short-crested waves and shallow breaking (right).

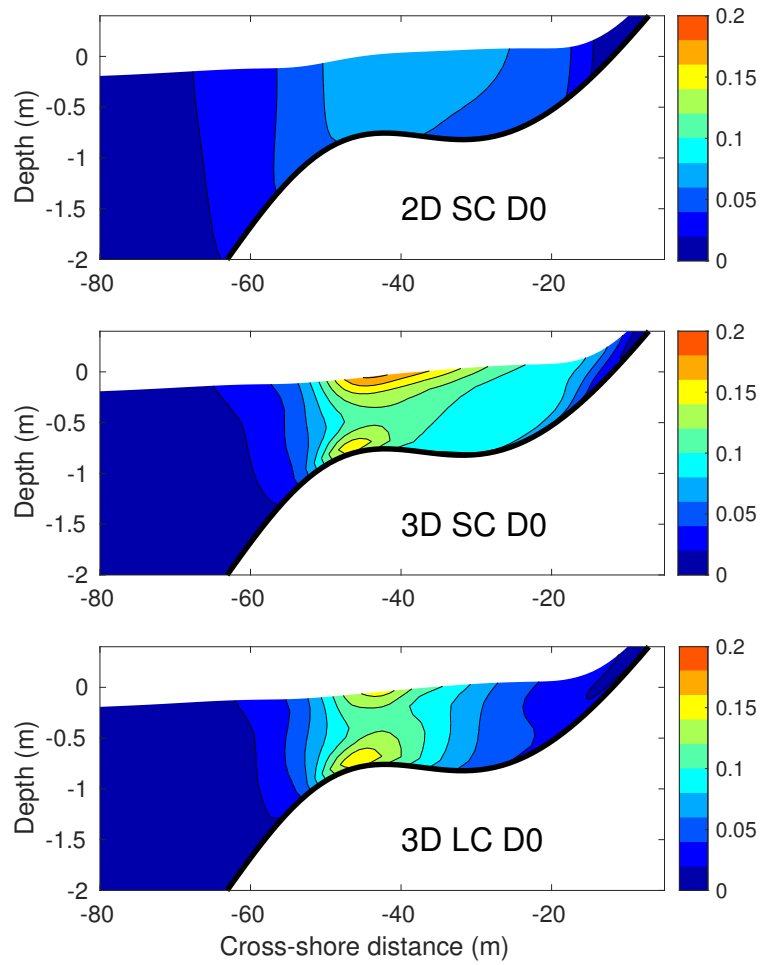


Figure 17: Cross-section of time-mean and longshore-mean EKE for the 3 cases of Figure 16

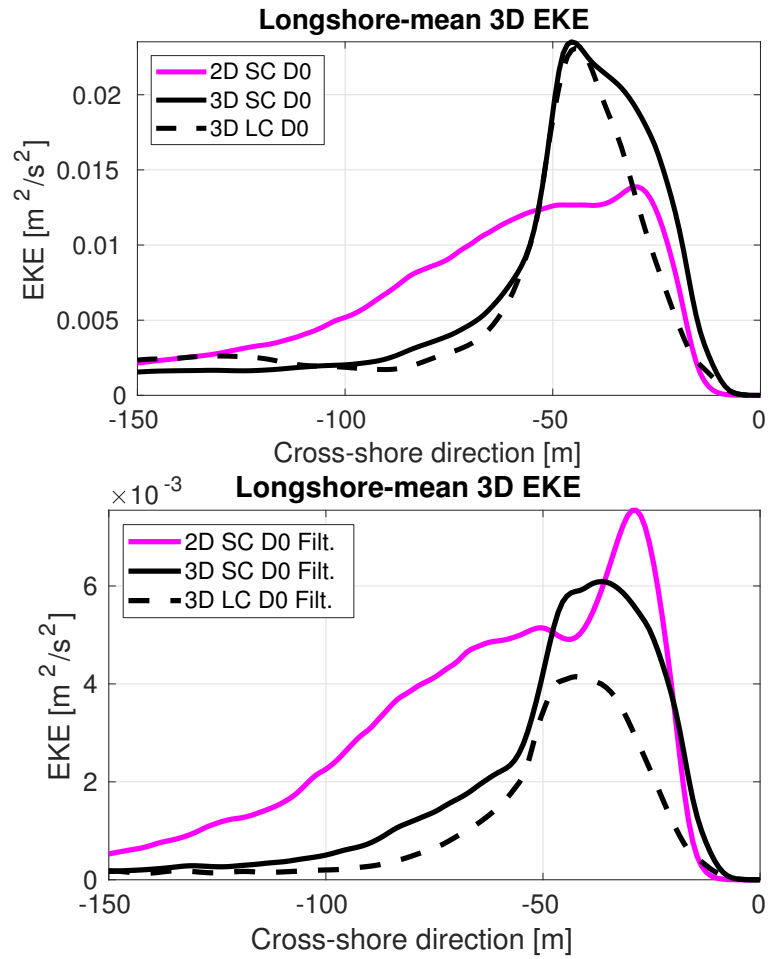


Figure 18: Cross-shore profile of depth-integrated (normalized), time and longshore-mean EKE for the 3 cases in Figure 16. Top: EKE of wave-averaged flow; bottom: EKE of low-pass filtered flow to remove 3D instability modes.

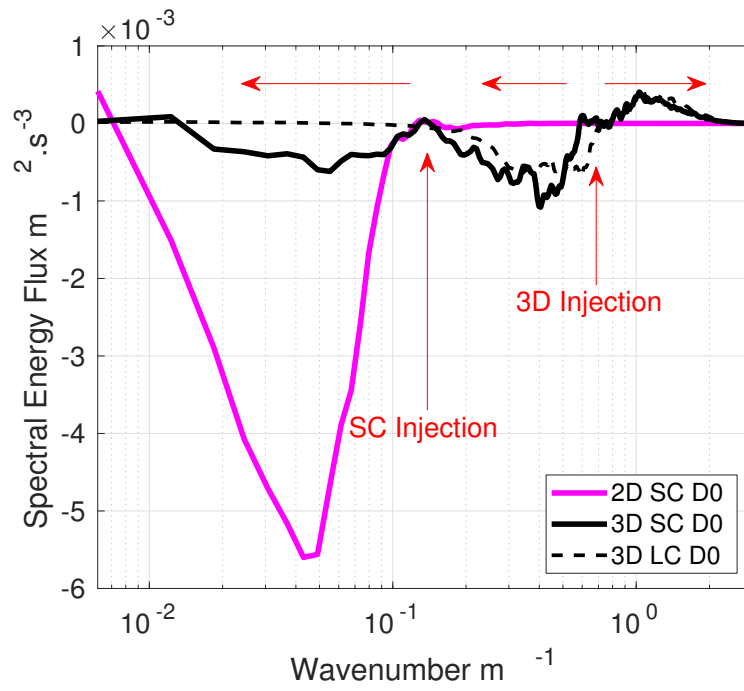


Figure 19: 3D and 2D model wavenumber spectral flux in the surfzone. Vertical red arrows point to the wavenumbers of energy injection from short-crested waves ( $\sim 40$  m wavelength) and from 3D shear instability ( $\sim 10$  m). The spectral flux is computed by spectral integration of  $v$  advection term (Marchesiello et al., 2011). Horizontal red arrows indicate positive/negative fluxes, i.e., direct/inverse energy cascade toward smaller/larger scales.



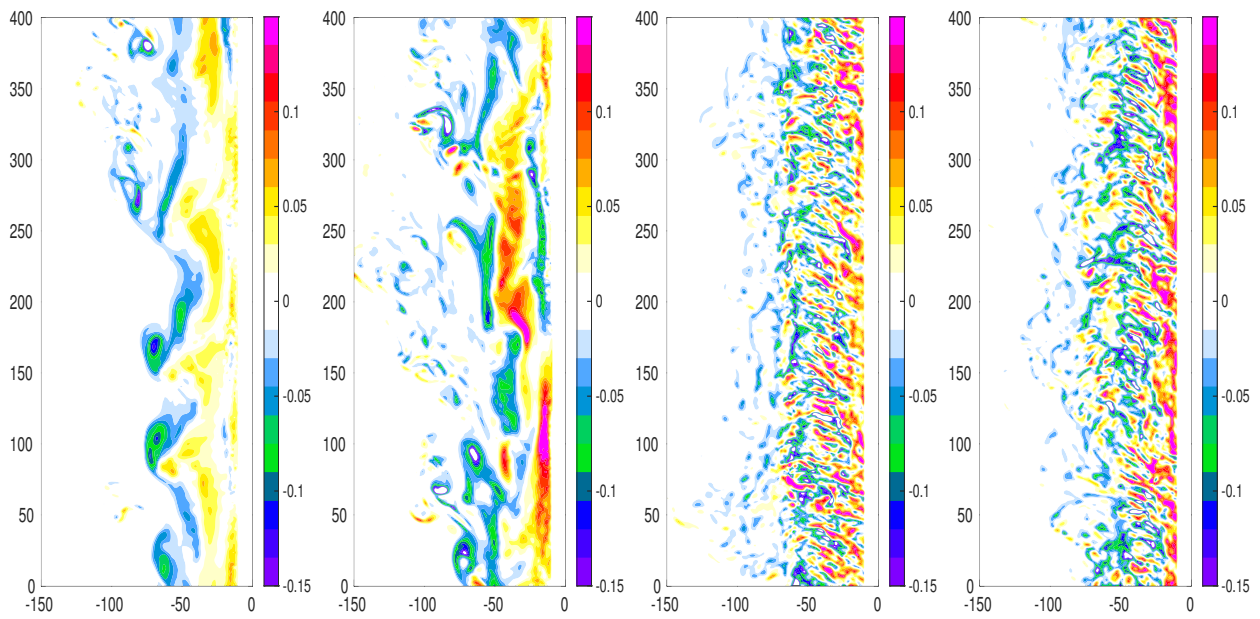


Figure 20: Snapshot of vorticity fields for 4 cases with oblique waves: long-crested waves with deep breaking (2D\_LC\_D20: first left); short-crested waves with deep breaking (2D\_SC\_D20: second left); long-crested waves with shallow breaking (3D\_LC\_D20: first right) and short-crested waves with shallow breaking (3D\_SC\_D20: last right).

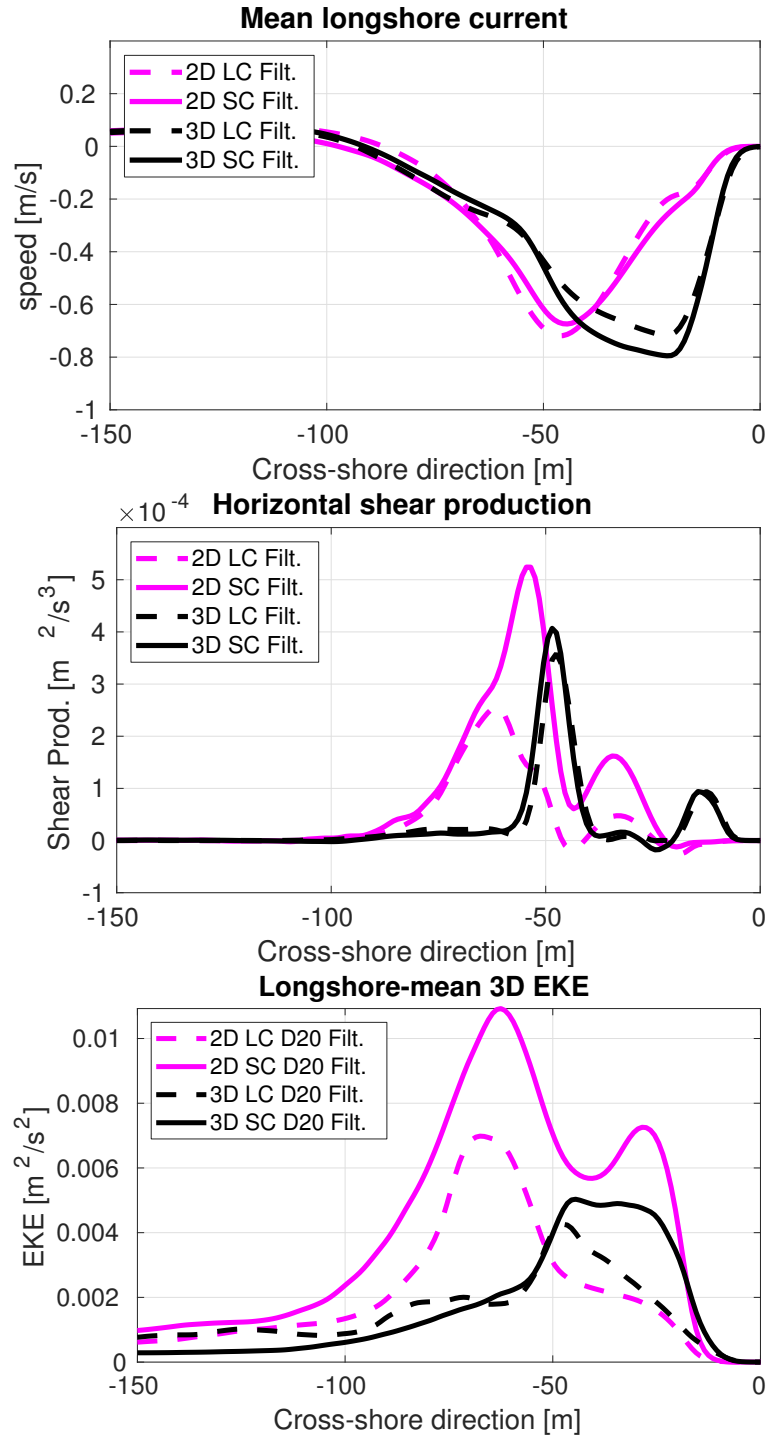


Figure 21: Cross-shore profile of time-mean and longshore-mean flow and eddy statistics for the 4 cases in Figure 16; top: surface longshore flow; center: horizontal shear production  $-\overline{u'v'} \frac{\partial \bar{v}}{\partial x}$ ; bottom: depth-integrated EKE of low-frequency modes.