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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

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). 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} ; η 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; λ 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 ρ 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)¹.

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 ρ_s is constant
 174 in time and space.

¹Auclair 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); Δ
190 the grid-characteristic length scale, i.e. the grid size: $\sqrt{\delta_x \delta_y}$ for horizontal viscosity and δ_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 ϵ , 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_μ 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 ω_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 θ_j is wave angle; θ_m is the mean wave direction and σ_θ the
 212 directional spread around the mean; ω_p and k_p are peak frequency and wavenumber; d_j
 213 is a normalized frequency-dependent directional distribution; ϕ_{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 (GLOBEX²) 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

²Globex data is freely available at 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 η 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 η 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 η and u_b are shown in Figure 3:
 276 mean, standard deviation (or H_s for η), skewness $\langle\phi'^3\rangle/\langle\phi'^2\rangle^{1.5}$ and asymmetry $\langle\mathcal{H}(\phi')^3\rangle/\langle\phi'^2\rangle^{1.5}$
 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 4256×2832 -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°
342 or 20° ($D = 10^\circ$ 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 γ 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 τ_b is the bottom shear stress computed by the model. E_0 is an empirical erosion rate set
 364 to 10^{-5} kg/m²/s; p is the sediment porosity (0.41); τ_{cr} is the critical shear stress, i.e., the
 365 threshold for initiation of sediment motion, set to 0.01 N/m².

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 δ — in free shear layers (Michalke, 1964, 1965), wavelength, frequency and growth
523 rate of the most unstable modes are 14δ , $0.015 U/\delta$ 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- ϵ 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/ν_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 19th 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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789 References

- 790 van der A, D.A., van der Zanden, J., O'Donoghue, T., Hurther, D., Cáceres, I., McLelland, S.J., Ribberink,
791 J.S., 2017. Large-scale laboratory study of breaking wave hydrodynamics over a fixed bar. *Journal of*
792 *Geophysical Research: Oceans* 122, 3287–3310.
- 793 Akan, C., McWilliams, J.C., Uchiyama, Y., 2020. Topographic and coastline influences on surf eddies. *Ocean*
794 *Modelling* 147, 101565.
- 795 Allen, J.S., Newberger, P.A., Holman, R.A., 1996. Nonlinear shear instabilities of alongshore currents on
796 plane beaches. *Journal of Fluid Mechanics* 310, 181–213.
- 797 Almar, R., Almeida, P., Blenkinsopp, C., Catalan, P., 2016. Surf-swash interactions on a low-tide terraced
798 beach. *Journal of Coastal Research* 75, 348–352.
- 799 Almar, R., Du Penhoat, Y., Honkonnou, N., Castelle, B., Laibi, R., Anthony, E.J., Sénéchal, N., Degbe, G.,
800 Chuchla, R., Sohou, Z., Dorel, M., Mensah-Senoo, T., Quenum, M., Addo K., A., Ibaceta, R., Kestenare,
801 E., Zodehougan, G., Laryea W., S., Lefebvre, J., 2014. The Grand Popo experiment, Benin. *Journal of*
802 *Coastal Research SI* 70, pp. 651–656.
- 803 Almar, R., Kestenare, E., Reyns, J., Jouanno, J., Anthony, E., Laibi, R., Hemer, M., Du Penhoat, Y.,
804 R., R., 2015a. Response of the bight of benin (gulf of guinea, west africa) coastline to anthropogenic
805 and natural forcing, part1: Wave climate variability and impacts on the longshore sediment transport.
806 *Continental Shelf Research* 110, 48 – 59.
- 807 Almar, R., Larnier, S., Castelle, B., Scott, T., Floc'h, F., Detandt, G., 2015b. On the use of the radon
808 transform to estimate longshore currents from video imagery. *Coastal Engineering* 114, 301–308.
- 809 Almar, R., Lerma, A.N., Castelle, B., Scott, T., 2018. On the influence of reflection over a rhythmic swash
810 zone on surf zone dynamics. *Ocean Dynamics* , 1–11.
- 811 Arduin, F., Rascle, N., Belibassakis, K., 2008. Explicit wave-averaged primitive equations using a general-
812 ized lagrangian mean. *Ocean Modelling* 20, 35.
- 813 Auclair, F., Bordois, L., Dossmann, Y., Duhaut, T., Paci, A., Ulses, C., Nguyen, C., 2018. A non-hydrostatic
814 non-boussinesq algorithm for free-surface ocean modelling. *Ocean Modelling* 132.

815 Barthelemy, E., 2004. Nonlinear shallow water theories for coastal waves. *Surveys in Geophysics* 25, 315–337.

816 Blaas, M., Dong, C., Marchesiello, P., McWilliams, J.C., Stolzenbach, K.D., 2007. Sediment-transport
817 modeling on southern californian shelves: A ROMS case study. *Continental Shelf Research* 27, 832 – 853.

818 Blayo, E., Debreu, L., 2005. Revisiting open boundary conditions from the point of view of characteristic
819 variables. *Ocean Modelling* .

820 Bonneton, P., Bruneau, N., Castelle, B., Marche, F., 2010. Large scale vorticity generation due to dissipating
821 waves in the surf zone. *Discrete and Continuous Dynamical Systems - series B* 13, 729–738.

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

824 Bowen, A.J., Holman, R.A., 1989. Shear instabilities of the mean longshore current: 1. theory. *Journal of*
825 *Geophysical Research: Oceans* 94, 18023–18030.

826 Canuto, V.M., Cheng, Y., 1997. Determination of the smagorinsky–lilly constant cs . *Physics of Fluids* 9,
827 1368–1378.

828 Castelle, B., du Penhoat, Y., Almar, R., Anthony, E., Lefebvre, J.P., Laibi, R., Chuchla, R., Dorel, M.,
829 Senechal, N., 2014. Flash rip dynamics on a high-energy low-tide-terraced beach (grand popo, benin,
830 west africa). *Journal of Coastal Research* 70, 633–638.

831 Chen, Q., Kirby, J.T., Dalrymple, R.A., Shi, F., Thornton, E.B., 2003. Boussinesq modeling of longshore
832 currents. *Journal of Geophysical Research: Oceans* 108.

833 Christensen, E.D., Deigaard, R., 2001. Large eddy simulation of breaking waves. *Coastal Engineering* 42,
834 53 – 86.

835 Cienfuegos, R., Barthélemy, E., Bonneton, P., 2010. Wave-breaking model for boussinesq-type equations
836 including roller effects in the mass conservation equation. *Journal of Waterway, Port, Coastal, and Ocean*
837 *Engineering* 136, 10–26.

838 Clark, D.B., Elgar, S., Raubenheimer, B., 2012. Vorticity generation by short-crested wave breaking.
839 *Geophysical Research Letters* 39.

840 Cox, D.T., Anderson, S.L., 2001. Statistics of intermittent surf zone turbulence and observations of large
841 eddies using piv. *Coastal Engineering Journal* 43, 121–131.

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

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

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

848 Derian, P., Almar, R., 2017. Wavelet-based optical flow estimation of instant surface currents from shore-

849 based and uav videos. *IEEE Transactions on Geoscience and Remote Sensing* 55, 5790–5797.

850 Dodd, N., Iranzo, V., Reniers, A., 2000. Shear instabilities of wave-driven alongshore currents. *Reviews of*
851 *Geophysics* 38, 437–463.

852 Dodd, N., Oltman-Shay, J., Thornton, E.B., 1992. Shear instabilities in the longshore current: A comparison
853 of observation and theory. *Journal of Physical Oceanography* 22, 62–82.

854 Elgar, S., Guza, R.T., 1985. Observations of bispectra of shoaling surface gravity waves. *Journal of Fluid*
855 *Mechanics* 161, 425–448.

856 Feddersen, F., 2014. The generation of surfzone eddies in a strong alongshore current. *Journal of Physical*
857 *Oceanography* 44, 600–617.

858 Feddersen, F., Clark, D.B., Guza, R.T., 2011. Modeling surf zone tracer plumes: 2. transport and dispersion.
859 *Journal of Geophysical Research: Oceans* 116.

860 Floc'h, F., Mabiala, G., Almar, R., Castelle, B., Hall, N., Du Penhoat, Y., Scott, T., Delacourt, C., 2018.
861 Flash rip statistics from video images. *Journal of Coastal Research* 81, 100–106.

862 Grasso, F., Castelle, B., Ruessink, G., 2012. Turbulence dissipation under breaking waves and bores in a
863 natural surf zone. *Continental Shelf Research* 43, 133–141.

864 Hally-Rosendahl, K., Feddersen, F., 2016. Modeling surfzone to inner-shelf tracer exchange. *Journal of*
865 *Geophysical Research: Oceans* 121, 4007–4025.

866 Henderson, S., Arnold, J., Özkan Haller, H., Solovitz, S., 2017. Depth dependence of nearshore currents and
867 eddies. *Journal of Geophysical Research: Oceans* 122.

868 Hilt, M., Auclair, F., Benschila, R., Bordoio, L., Capet, X., Debreu, L., Dumas, F., Jullien, S., Lemarié, F.,
869 Marchesiello, P., Nguyen, C., Roblou, L., 2020. Numerical modelling of hydraulic control, solitary waves
870 and primary instabilities in the strait of gibraltar. *Ocean Modelling* 151, 101642.

871 Johnson, D., Pattiaratchi, C., 2006. Boussinesq modelling of transient rip currents. *Coastal Engineering* 53,
872 419 – 439.

873 Kazakova, M., Richard, G.L., 2019. A new model of shoaling and breaking waves: one-dimensional solitary
874 wave on a mild sloping beach. *Journal of Fluid Mechanics* 862, 552–591.

875 Kirby, J.T., 2016. Boussinesq models and their application to coastal processes across a wide range of scales.
876 *Journal of Waterway, Port, Coastal, and Ocean Engineering* 142, 03116005.

877 Kirby, J.T., Derakhti, M., 2019. Short-crested wave breaking. *European Journal of Mechanics - B/Fluids*
878 73, 100–111.

879 Kumar, N., Feddersen, F., 2017. The effect of stokes drift and transient rip currents on the inner shelf. part
880 i: No stratification. *Journal of Physical Oceanography* 47, 227–241.

881 Kumar, N., Voulgaris, G., Warner, J.C., Olabarrieta, M., 2012. Implementation of the vortex force formalism
882 in the coupled ocean-atmosphere-wave-sediment transport (coawst) modeling system for inner shelf and

883 surf zone applications. *Ocean Modelling* 47, 65 – 95.

884 Lannes, D., Bonneton, P., 2009. Derivation of asymptotic two-dimensional time-dependent equations for
885 surface water wave propagation. *Physics of Fluids* 21, 016601.

886 Larson, M., , Kraus, N., 2002. NMLONG: Numerical model for simulating longshore current; Report 2:
887 Wave-current interaction, roller modeling, and validation of model enhancements. Technical Report. Tech-
888 nical Report ERDC/CHL TR-02- 22, US Army Engineer Research and Development Center, Vicksburg,
889 MS.

890 Lesieur, M., 1990. *Turbulence in Fluids. Fluid Mechanics and its Applications*, Springer Netherlands.

891 Lin, P., Li, C.W., 2002. A σ -coordinate three-dimensional numerical model for surface wave propagation.
892 *International Journal for Numerical Methods in Fluids* 38, 1045–1068.

893 Lin, P., Liu, P.L.F., 1998. A numerical study of breaking waves in the surf zone. *Journal of Fluid Mechanics*
894 359, 239–264.

895 Lippmann, T.C., Thornton, E.B., Stanton, T.P., 2016. The Vertical Structure of Low-Frequency Motions
896 in the Nearshore. Part I: Observations. *Journal of Physical Oceanography* 46, 3695–3711.

897 Longo, S., Petti, M., Losada, I.J., 2002. Turbulence in the swash and surf zones: a review. *Coastal*
898 *Engineering* 45, 129 – 147. *Surface and Swash Zone Mechanics*.

899 Lubin, P., Chanson, H., 2017. Are breaking waves, bores, surges and jumps the same flow? *Environmental*
900 *Fluid Mechanics* 17, 47–77.

901 Ma, G., Shi, F., Kirby, J.T., 2012. Shock-capturing non-hydrostatic model for fully dispersive surface wave
902 processes. *Ocean Modelling* 43-44, 22 – 35.

903 MacMahan, J.H., Thornton, E.B., Reniers, A.J., 2006. Rip current review. *Coastal Engineering* 53, 191 –
904 208. *Coastal Hydrodynamics and Morphodynamics Symposium celebrating the academic closing address*
905 *of Jurjen A. Battjes*.

906 Marchesiello, P., Almar, R., Benschila, R., Larnier, S., Castelle, B., McWilliams, J.C., 2016. On eddy-mixed
907 longshore currents: Video observation and 3d modeling off grand popo beach, benin. *Journal of Coastal*
908 *Research* 75, 408–412.

909 Marchesiello, P., Benschila, R., Almar, R., Uchiyama, Y., McWilliams, J.C., Shchepetkin, A., 2015. On
910 tridimensional rip current modeling. *Ocean Modelling* 96, 36 – 48. *Waves and coastal, regional and global*
911 *processes*.

912 Marchesiello, P., Capet, X., Menkes, C., Kennan, S.C., 2011. Submesoscale dynamics in tropical instability
913 waves. *Ocean Modelling* 39, 31–46.

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

916 McWilliams, J.C., Akan, C., Uchiyama, Y., 2018. Robustness of nearshore vortices. *Journal of Fluid*

917 Mechanics 850, R2.

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

920 Menesguen, C., Le Gentil, S., Marchesiello, P., Ducouso, N., 2018. Destabilization of an oceanic meddy-like
921 vortex: energy transfers and significance of numerical settings. *Journal Of Physical Oceanography* 48,
922 1151–1168.

923 Metcalfe, R.W., Orszag, S.A., Brachet, M.E., Menon, S., Riley, J.J., 1987. Secondary instability of a
924 temporally growing mixing layer. *Journal of Fluid Mechanics* 184, 207–243.

925 Meyers, J., Sagaut, P., 2006. On the model coefficients for the standard and the variational multi-scale
926 smagorinsky model. *Journal of Fluid Mechanics* 569, 287–319.

927 Michalke, A., 1964. On the inviscid instability of the hyperbolic tangent velocity profile. *Journal of Fluid*
928 *Mechanics* 19, 543–556.

929 Michalke, A., 1965. On spatially growing disturbances in an inviscid shear layer. *Journal of Fluid Mechanics*
930 23, 521–544.

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

935 Nadaoka, K., Hino, M., Koyano, Y., 1989. Structure of the turbulent flow field under breaking waves in the
936 surf zone. *Journal of Fluid Mechanics* 204, 359–387.

937 Newberger, P.A., Allen, J.S., 2007. Forcing a three-dimensional, hydrostatic, primitive-equation model for
938 application in the surf zone: 2. application to duck94. *Journal of Geophysical Research: Oceans* 112.

939 Özkan-Haller, H.T., Kirby, J.T., 1999. Nonlinear evolution of shear instabilities of the longshore current:
940 A comparison of observations and computations. *Journal of Geophysical Research: Oceans* 104, 25953–
941 25984.

942 Penney, J., Morel, Y., Haynes, P., Auclair, F., Nguyen, C., 2020. Diapycnal mixing of passive tracers by
943 kelvin–helmholtz instabilities. *Journal of Fluid Mechanics* 900, A26.

944 Peregrine, D.H., 1998. Surf zone currents. *Theor. Comput. Fluid Dyn.* 10, 295–309.

945 Pierrehumbert, R., Widnall, S., 1982. The two- and three-dimensional instabilities of a spatially periodic
946 shear layer. *Journal of Fluid Mechanics* 114, 59 – 82.

947 de Schipper, M.A., Reniers, A.J., Ranasinghe, R., Stive, M.J., 2014. The influence of sea state on formation
948 speed of alongshore variability in surf zone sand bars. *Coastal Engineering* 91, 45 – 59.

949 Scott, T., Castelle, B., Almar, R., Senechal, N., Floch, F., Detandt, G., 2018. Controls on flash rip current
950 hazard on low-tide terraced tropical beaches in west africa. *Journal of Coastal Research* 81, 92–99.

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

953 Shi, F., Kirby, J.T., Harris, J.C., Geiman, J.D., Grilli, S.T., 2012. A high-order adaptive time-stepping tvd
954 solver for boussinesq modeling of breaking waves and coastal inundation. *Ocean Modelling* 43-44, 36 –
955 51.

956 Short, A.D., Hogan, C.L., Safety, A.B., Program., M., 1993. *Beaches of the New South Wales coast : a*
957 *guide to their nature, characteristics, surf and safety*. Australian Beach Safety and Management Program
958 Beaconsfield, N.S.W.

959 Slinn, D.N., Allen, J.S., Newberger, P.A., Holman, R.A., 1998. Nonlinear shear instabilities of alongshore
960 currents over barred beaches. *Journal of Geophysical Research* 103, 18,357 – 18,379.

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

963 Splinter, K.D., Slinn, D.N., 2003. Three-dimensional modeling of alongshore current dynamics, in: *Canadian*
964 *Coastal Conference*, Kingston, Ontario, pp. 1–14.

965 Spydell, M., Feddersen, F., 2009. Lagrangian drifter dispersion in the surf zone: Directionally spread,
966 normally incident waves. *Journal of Physical Oceanography* 39, 809–830.

967 Svendsen, I.A., Madsen, P.A., 1984. A turbulent bore on a beach. *Journal of Fluid Mechanics* 148, 73–96.

968 Tissier, M., Bonneton, P., Marche, F., Chazel, F., Lannes, D., 2012. A new approach to handle wave
969 breaking in fully non-linear boussinesq models. *Coastal Engineering* 67, 54 – 66.

970 Uchiyama, Y., McWilliams, J., Shchepetkin, A., 2010. Wave-current interaction in an oceanic circulation
971 model with a vortex-force formalism: Application to the surf zone. *Ocean Modelling* 34, 16–35.

972 Uchiyama, Y., McWilliams, J.C., Akan, C., 2017. Three-dimensional transient rip currents: Bathymetric
973 excitation of low-frequency intrinsic variability. *Journal of Geophysical Research: Oceans* 122, 5826–5849.

974 Uchiyama, Y., McWilliams, J.C., Restrepo, J.M., 2009. Wave-current interaction in nearshore shear insta-
975 bility analyzed with a vortex force formalism. *Journal of Geophysical Research: Oceans* 114.

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

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

980 Warner, J.C., Sherwood, C.R., Signell, R.P., Harris, C.K., Arango, H.G., 2008. Development of a three-
981 dimensional, regional, coupled wave, current, and sediment-transport model. *Computers & Geosciences*
982 34, 1284–1306.

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

- 985 Watanabe, Y., Saeki, H., Hosking, R.J., 2005. Three-dimensional vortex structures under breaking waves.
986 *Journal of Fluid Mechanics* 545, 291–328.
- 987 Wei, G., Kirby, J.T., Sinha, A., 1999. Generation of waves in boussinesq models using a source function
988 method. *Coastal Engineering* 36, 271 – 299.
- 989 Zijlema, M., Stelling, G., Smit, P., 2011. Swash: An operational public domain code for simulating wave
990 fields and rapidly varied flows in coastal waters. *Coastal Engineering* 58, 992 – 1012.

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 _f =0.001 s, c _{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 forcing (u,v,T,S fluxes)	0
Sediment model (D50, Ws, E_0 , τ_{cr})	$D50 = 100 \mu\text{m}$; $Ws=5$ 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 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

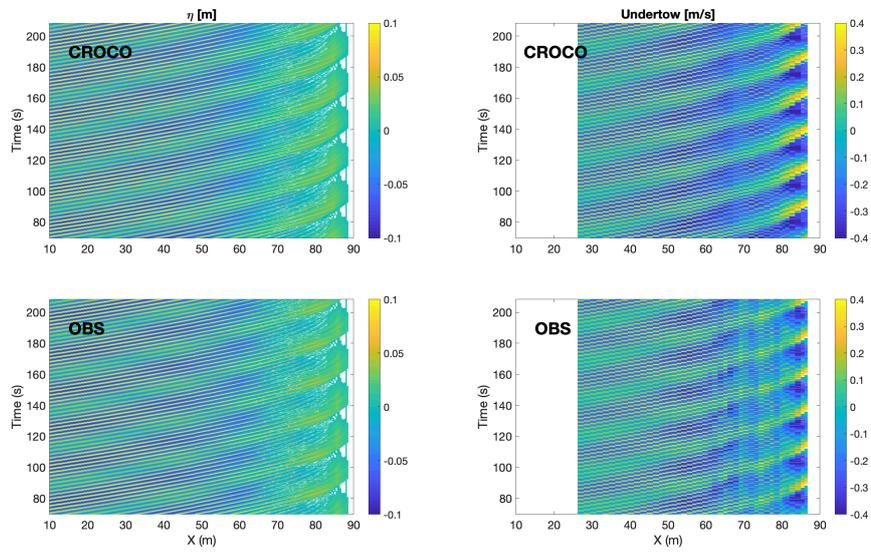


Figure 1: Hovmuller plot (x,time) of data and model sea level η and undertow u_b for the GLOBEX B3 flume experiment. When data is missing in the measurements, it is also removed from the model output.

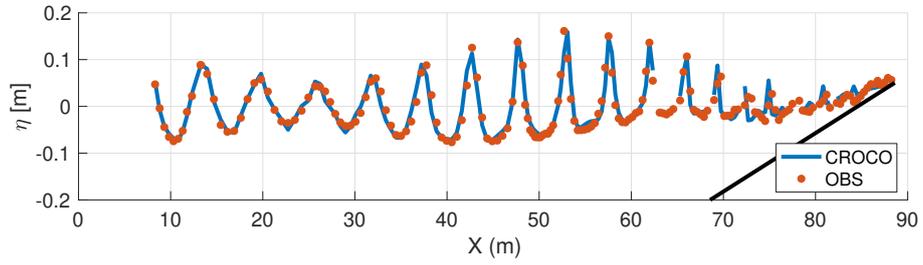


Figure 2: Snapshot of wave field across the GLOBEX B3 flume experiment during runup conditions for model (blue line) and data (red dots).

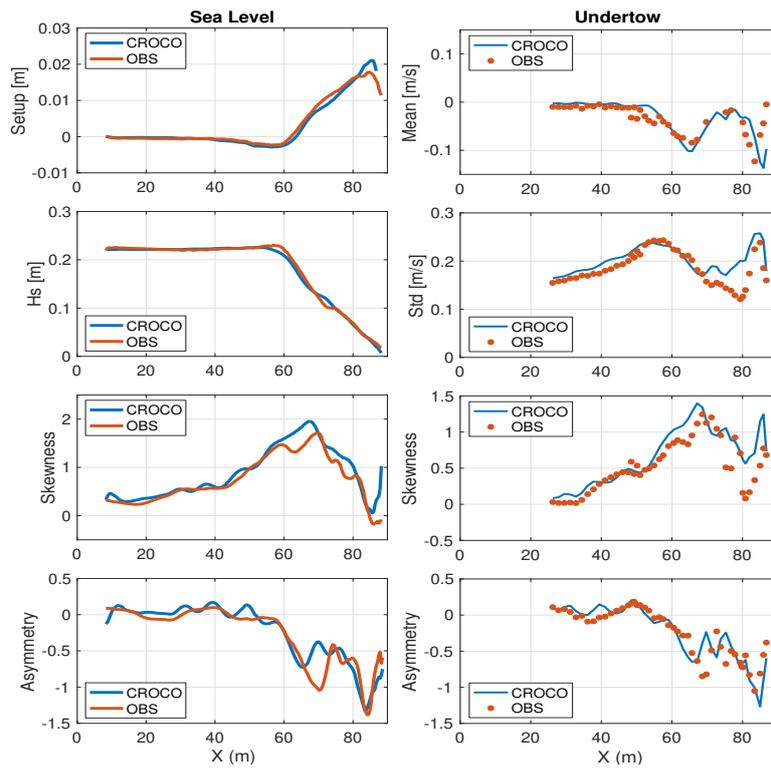


Figure 3: Wave statistics from GLOBEX B3 experiment for sea level η (right) and undertow u_b (left) in the model (blue line) and data (red line or dots). From top to bottom: mean, standard deviation (or H_s for η), skewness and asymmetry.

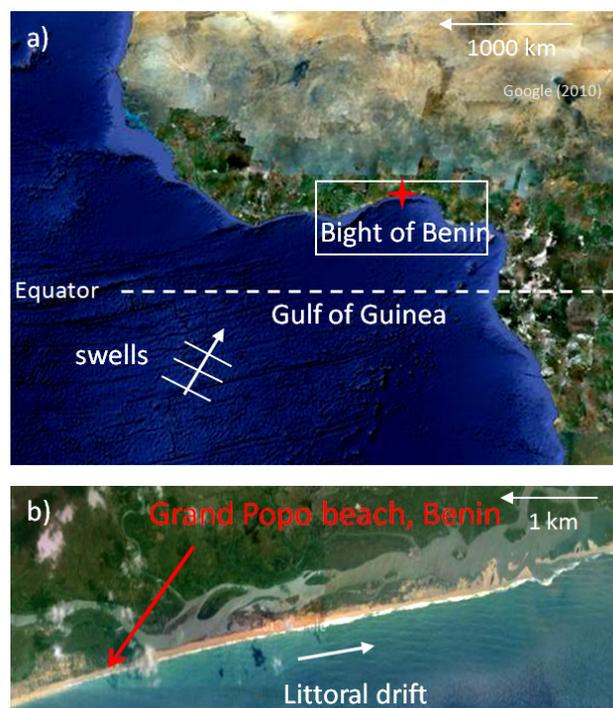


Figure 4: Grand Popo Beach, Benin, is a longshore uniform beach located in the Gulf of Guinea. It is representative of an open, wave-dominated and microtidal environment exposed to S-SW long period swells generated at high latitudes in the South Atlantic.

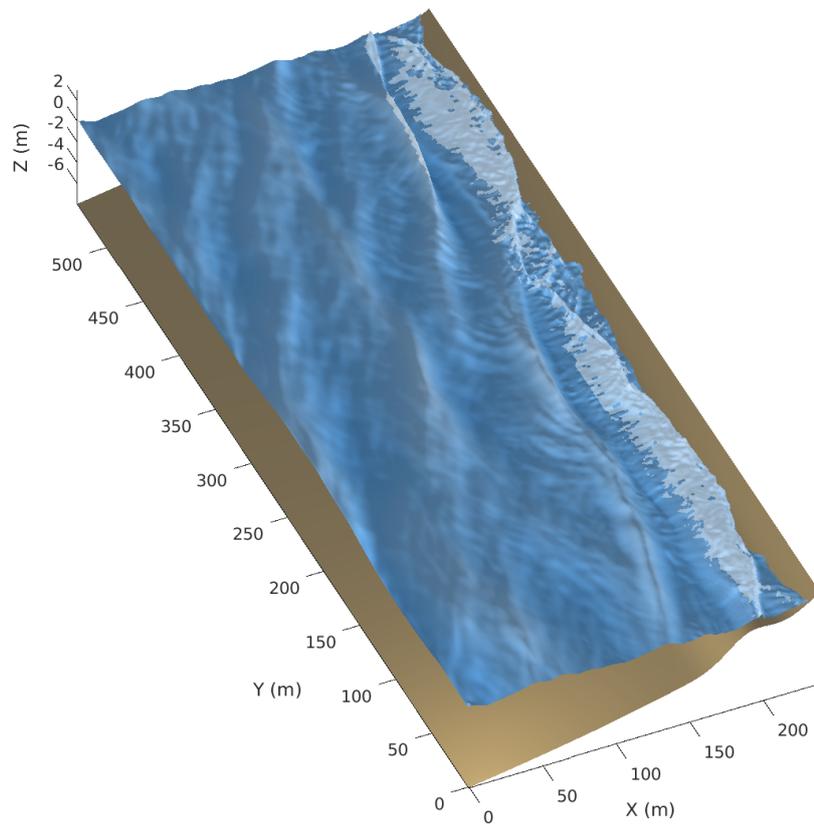


Figure 5: 3D representation of surface waves in CROCO. Short-crested waves are generated at the offshore boundary and propagate shoreward at an angle of 10° , with refraction and breaking through the surf and swash zones. Foam in the surfzone is approximated by white patches in locations of high turbulent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

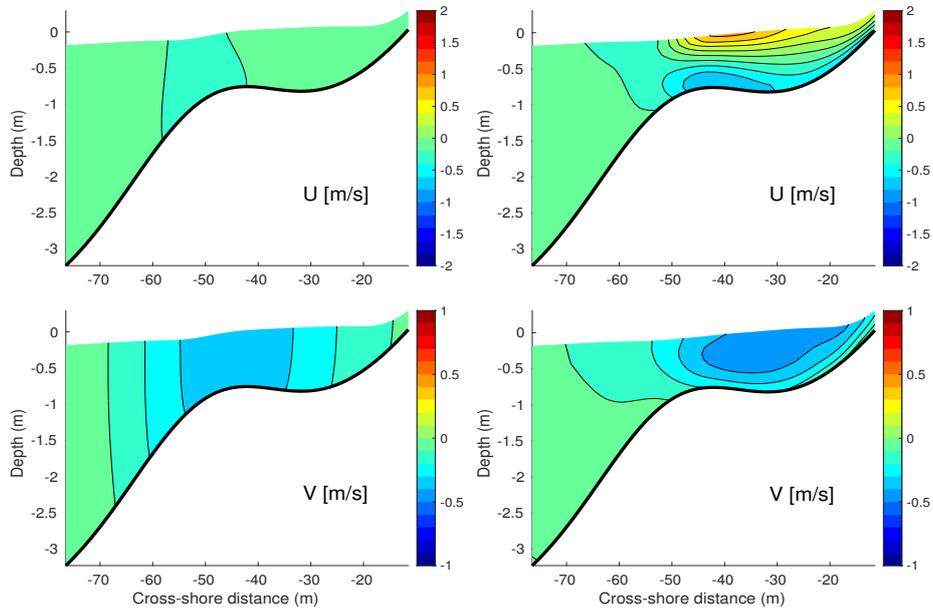


Figure 6: Cross-section of time-mean and longshore-mean, cross-shore (top) and longshore (bottom) currents in the case of shallow (right) and deep (left) breaking.

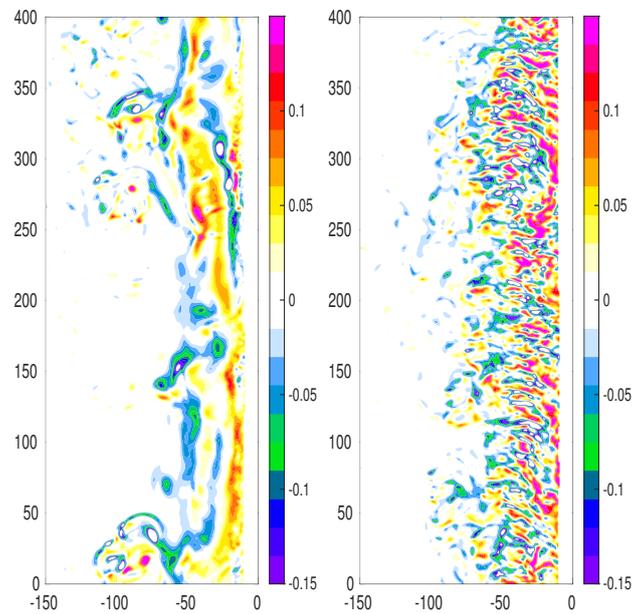


Figure 7: Surface vertical vorticity snapshot in the case of shallow (right) and deep (left) breaking.

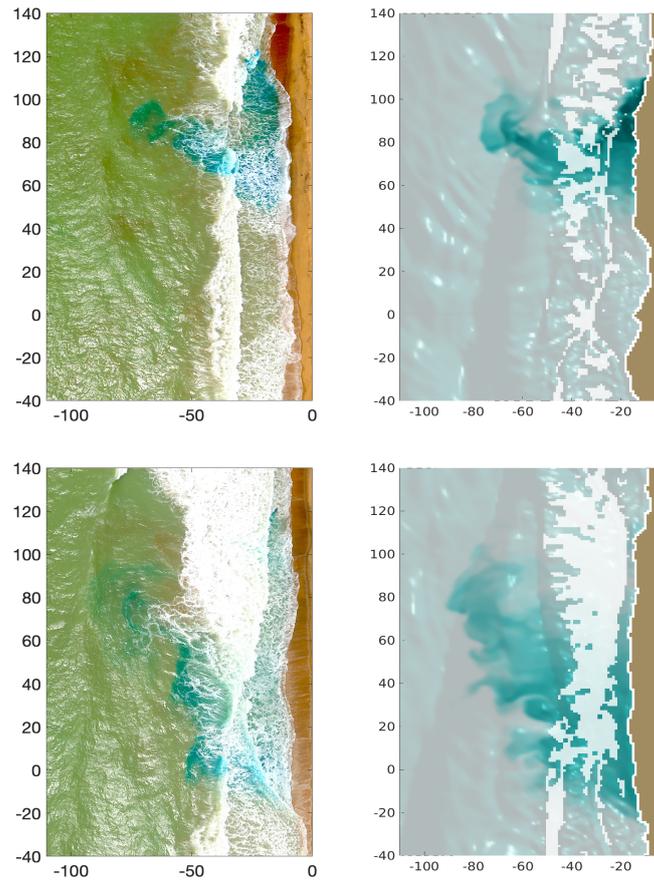


Figure 8: Drone photos (left) and CROCO simulation (right) of dye release at 154 s interval during the Grand Popo survey of March 13 2014 at afternoon mid-tide. In the model, wave surfaces are presented in the background with light from the left (as in the photos), and foam is approximated by high turbulent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

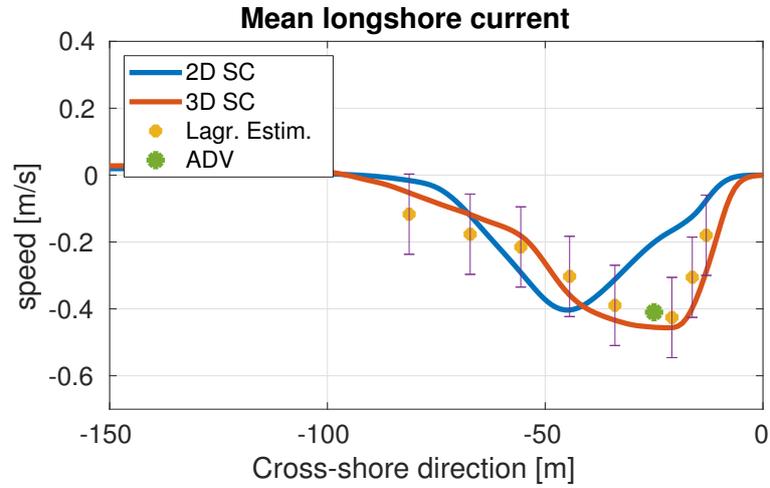


Figure 9: Cross-shore profile of time-mean and longshore-mean surface longshore flow for the full 3D simulation with wave direction $D = 10^\circ$. ADV measurement and an estimation of Lagrangian velocities from drone photos (Fig. 8) is added for comparison.

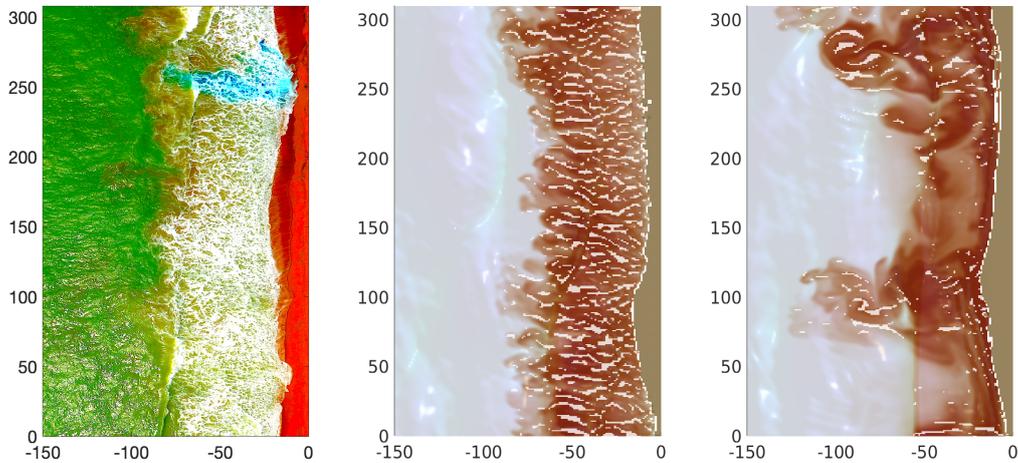


Figure 10: Aerial photo (left) and CROCO simulations of surfzone suspended sediments for shallow-breaking (center) and deep-breaking (right) cases. The contrasts in the photo is enhanced to better expose suspended sediments (brown color). The model suspended sediments correspond to snapshots at 15 min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

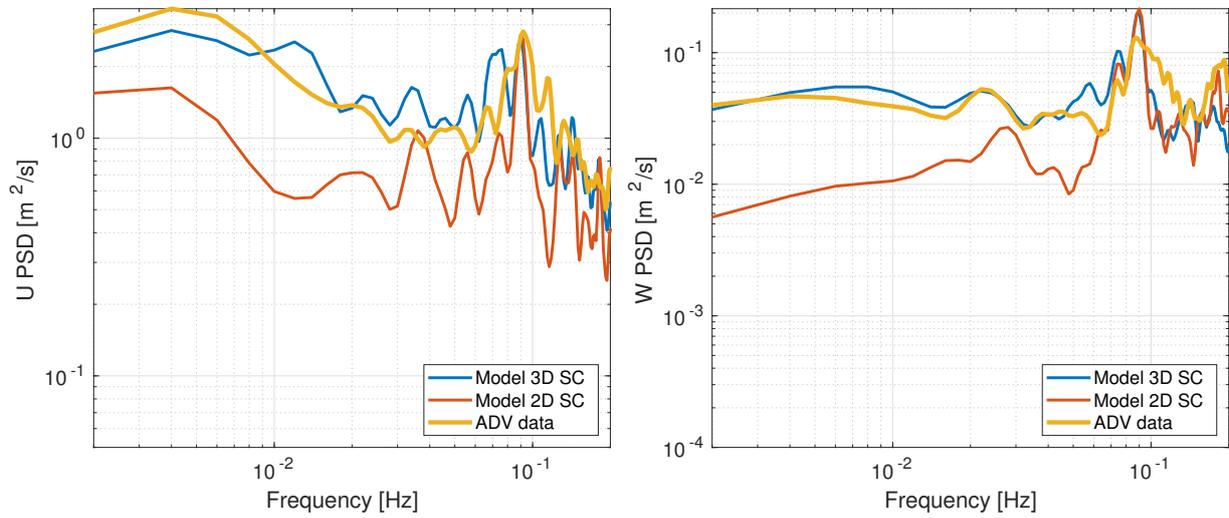


Figure 11: Power Spectral Density of horizontal (left) and vertical (right) velocity fluctuations compared with ADV measurements in the middle of the terrace, in March 13 2014 at afternoon mid-tide.

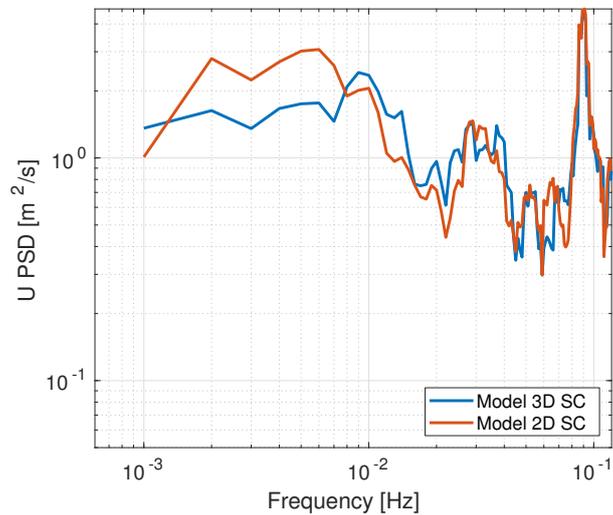


Figure 12: Power Spectral Density of offshore horizontal velocity fluctuations in March 13 2014 at afternoon mid-tide.

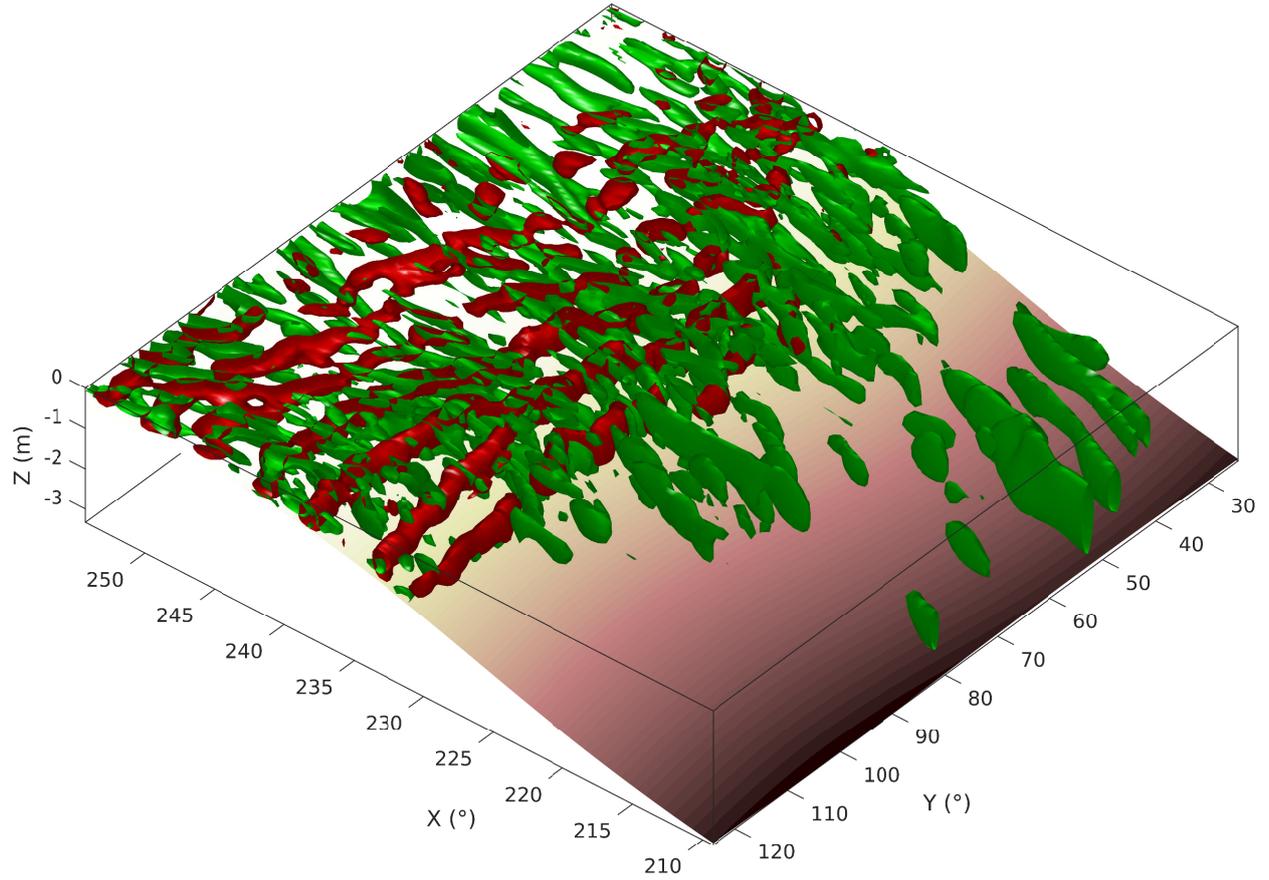


Figure 13: Q field defined by $Q = -\frac{1}{2} \frac{\partial u_i}{\partial x_j} \frac{\partial u_j}{\partial x_i}$, showing coherent structures similar to rolls and ribs in a transitional mixing layer. Cross-shore and alongshore Q terms are split: spanwise rolls (aligned across shear direction) are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

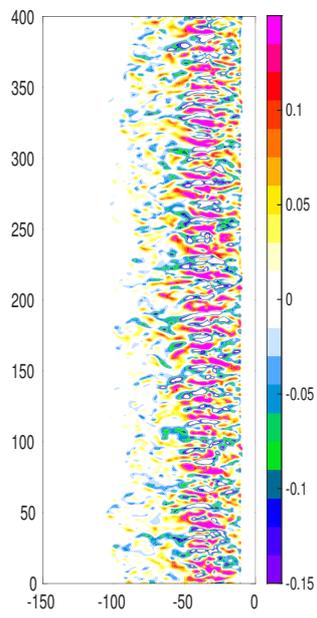


Figure 14: Surface vertical vorticity snapshot for the case with monochromatic shore-normal wave forcing (3D_MONO_D0)

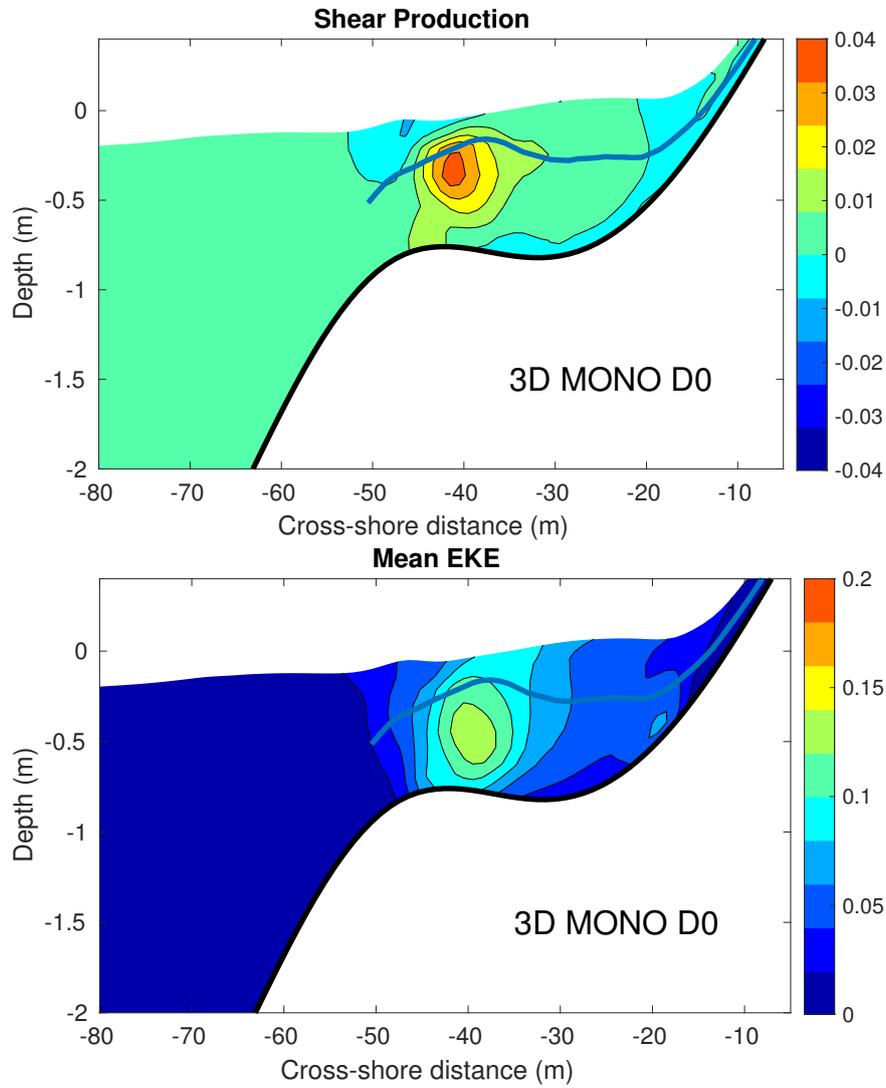


Figure 15: Cross-section of EKE [m^2/s^3] (bottom) and EKE production by the mean shear flow (top) [m^2/s^3] for the case with 3D instabilities only, forced with monochromatic long-crested waves (3DMONOD0). The position of shear flow inflection point is drawn in a blue line

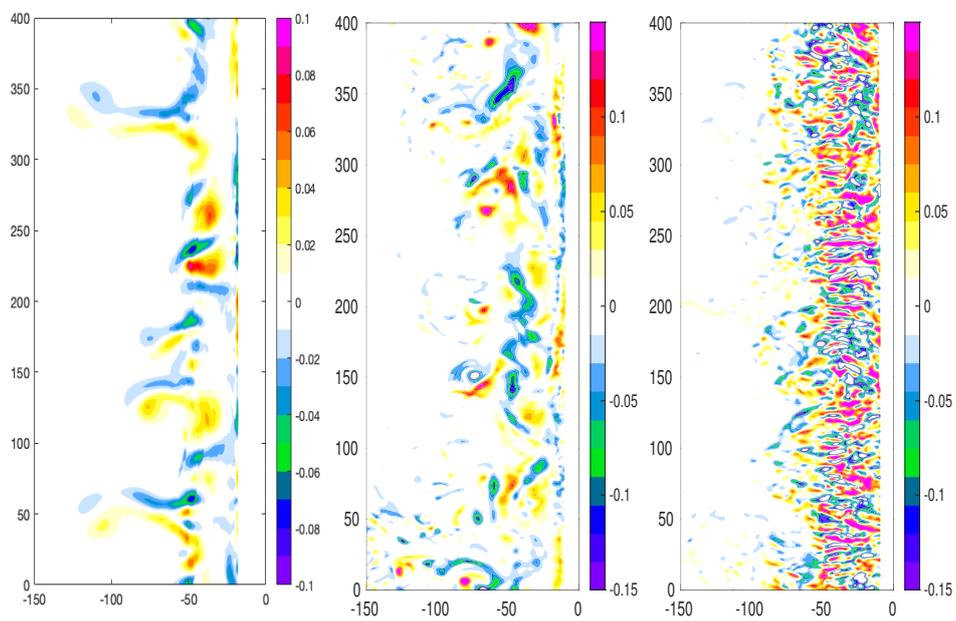


Figure 16: Vertical vorticity snapshot for 3 cases with shore-normal wave forcing: FUNWAVE-TVD (2D Boussinesq) with short-crested waves (left); CROCO with short-crested waves and deep breaking (center); short-crested waves and shallow breaking (right).

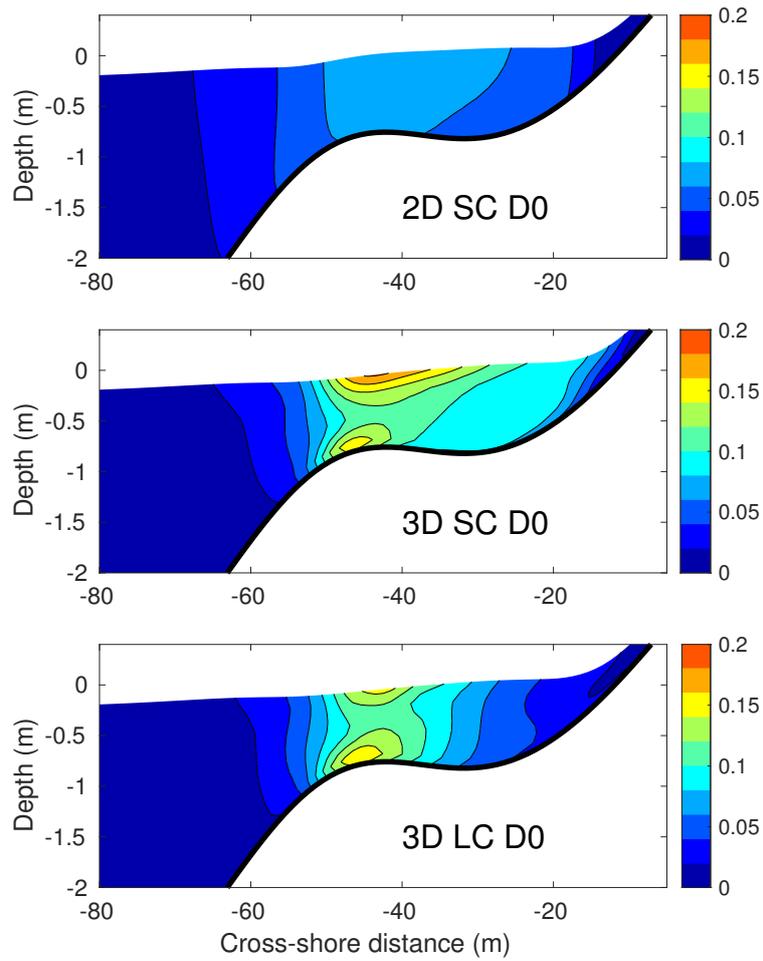


Figure 17: Cross-section of time-mean and longshore-mean EKE for the 3 cases of Figure 16

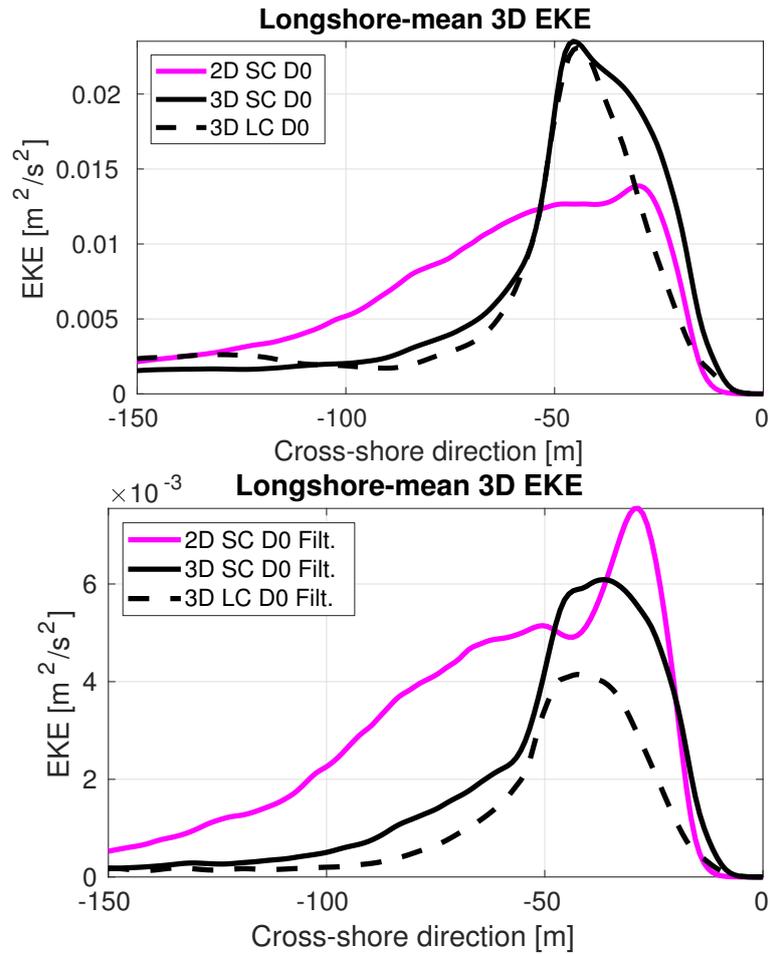


Figure 18: Cross-shore profile of depth-integrated (normalized), time and longshore-mean EKE for the 3 cases in Figure 16. Top: EKE of wave-averaged flow; bottom: EKE of low-pass filtered flow to remove 3D instability modes.

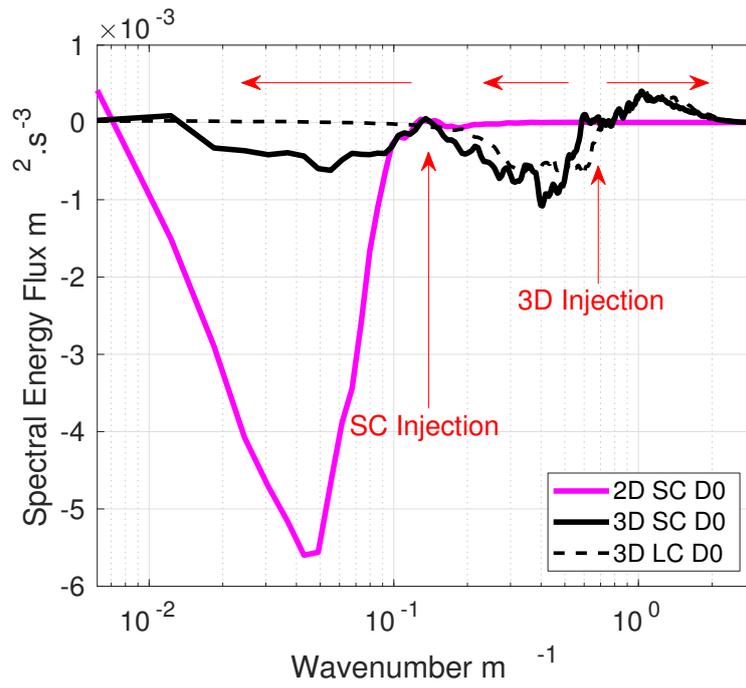


Figure 19: 3D and 2D model wavenumber spectral flux in the surfzone. Vertical red arrows point to the wavenumbers of energy injection from short-crested waves (~ 40 m wavelength) and from 3D shear instability (~ 10 m). The spectral flux is computed by spectral integration of v advection term (Marchesiello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

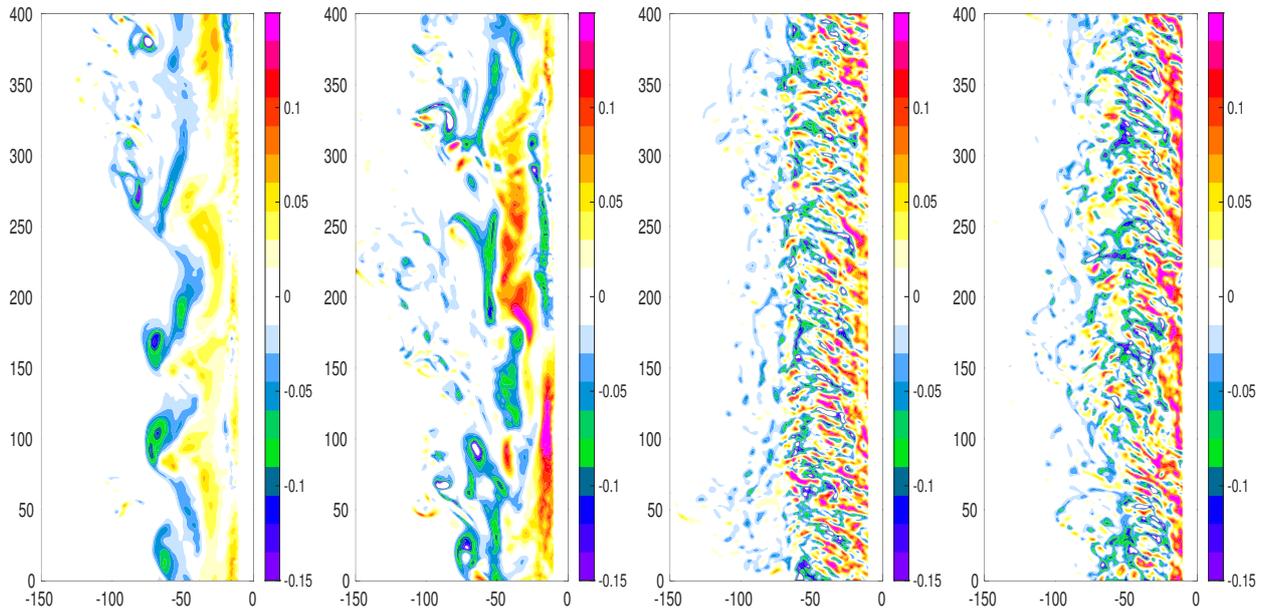


Figure 20: Snapshot of vorticity fields for 4 cases with oblique waves: long-crested waves with deep breaking (2D_LC_D20: first left); short-crested waves with deep breaking (2D_SC_D20: second left); long-crested waves with shallow breaking (3D_LC_D20: first right) and short-crested waves with shallow breaking (3D_SC_D20: last right).

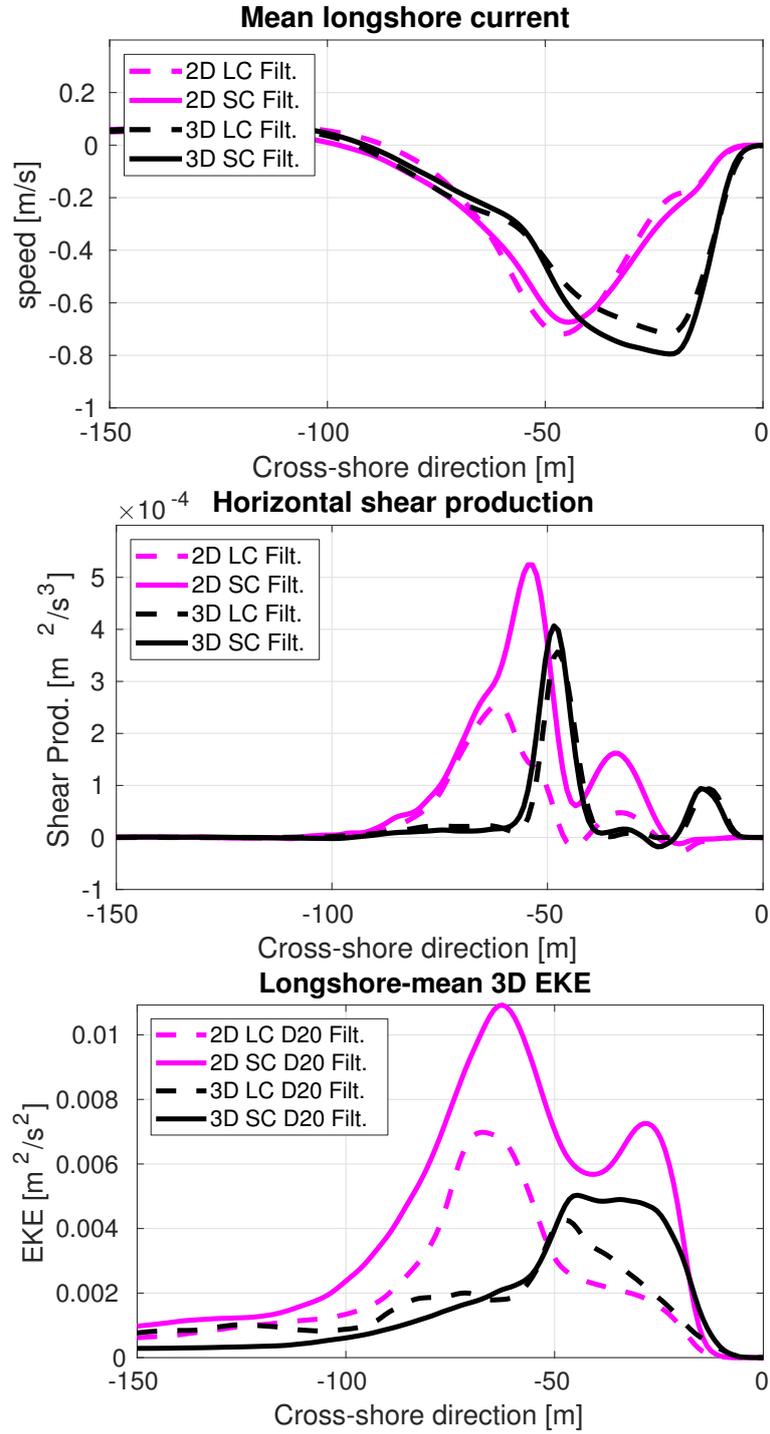


Figure 21: Cross-shore profile of time-mean and longshore-mean flow and eddy statistics for the 4 cases in Figure 16; top: surface longshore flow; center: horizontal shear production $-\overline{u'v'} \frac{\partial \bar{v}}{\partial x}$; bottom: depth-integrated EKE of low-frequency modes.