



**HAL**  
open science

# Cross-Shore Distribution of the Wave-Induced Circulation Over a Dissipative Beach Under Storm Wave Conditions

Marc Pezerat, Xavier Bertin, Kévin Martins, Laura Lavaud

► **To cite this version:**

Marc Pezerat, Xavier Bertin, Kévin Martins, Laura Lavaud. Cross-Shore Distribution of the Wave-Induced Circulation Over a Dissipative Beach Under Storm Wave Conditions. *Journal of Geophysical Research. Oceans*, 2022, 127 (3), pp.e2021JC018108. 10.1029/2021JC018108 . hal-03818795

**HAL Id: hal-03818795**

**<https://univ-rochelle.hal.science/hal-03818795v1>**

Submitted on 18 Oct 2022

**HAL** is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.

# Cross-shore distribution of the wave-induced circulation over a dissipative beach under storm wave conditions

Marc Pezerat<sup>1</sup>, Xavier Bertin<sup>1</sup>, Kévin Martins<sup>2</sup>, Laura Lavaud<sup>1</sup>

<sup>1</sup>UMR 7266 LIENSs, CNRS/La Rochelle Université, 2 rue Olympe de Gouges, 17000 La Rochelle, France

<sup>2</sup>UMR 5805 EPOC, CNRS/Université de Bordeaux, Allée Geoffroy Saint-Hilaire, F-33615 Pessac, France

## Key Points:

- Field experiment at a dissipative beach with 6 m  $H_{m0}$  at breaking and undertows reaching 0.2 m/s as far as 4 km from the shoreline.
- Accurate reproduction of the cross-shore hydrodynamics using a phase-averaged 3D circulation model.
- Strong sensitivity of the vertical mixing to breaking wave parameterization.
- Wave dissipation by breaking locally increases seaward-directed flows by over 100% compared to the surface Stokes drift velocity.

---

Corresponding author: Marc Pezerat, [marc.pezerat@univ-lr.fr](mailto:marc.pezerat@univ-lr.fr)

**Abstract**

This study explores the spatial distribution and the driving mechanisms of the wave-induced cross-shore flow within the shoreface and surf zone of a dissipative beach. Unpublished results from a field campaign carried out in early 2021 under storm wave conditions are presented and compared with the predictions from a state-of-the-art phase-averaged three-dimensional circulation modelling system based on the vortex force formalism. Under storm wave conditions, the cross-shore flow is dominated by a strong seaward-directed current in the lower part of the water column. The largest current velocities of this return current are located in the surf zone, where the dissipation by depth-induced breaking is most intense, but offshore-directed velocities up to 0.2 m/s are observed as far as 4 km from the shoreline ( $\simeq 12$  m-depth). Numerical experiments further highlight the key control exerted by non-conservative wave forces and wave-enhanced mixing on the cross-shore flow across a transition zone, where depth-induced breaking, whitecapping and bottom friction all significantly contribute to the wave energy dissipation. Under storm conditions, this transition zone extended almost 6 km offshore and the cross-shore Lagrangian circulation shows a strong seaward-directed jet in the lower part of the water column, whose intensity progressively decreases offshore. In contrast, the surf zone edge appears clearly delimited under fair weather conditions and the seaward-directed current is weakened by a near bottom shoreward-directed current associated with wave bottom streaming in the shoaling region, such that the clockwise Lagrangian overturning circulation is constrained by an additional anti-clockwise overturning cell at the surf zone edge.

**Plain Language Summary**

As waves propagate toward the shore fluid parcels experience a net transport in the direction of wave propagation. This onshore mass transport is compensated by a near bed return flow, which dynamics remain poorly understood. This study combines measurements from a field campaign carried out in early 2021 in front of a gentle sloping beach and numerical modelling to explore the spatial distribution and the driving mechanisms of this wave-induced cross-shore flow. Both observations and model results show that the largest current velocities of this return current are located very close to the shoreline, where the wave breaking is the most intense, but values up to 0.2 m/s are observed as far as 4 km from the shoreline under storm conditions. Numerical experiments further highlight the key control exerted by the wave forces and the wave-enhanced mixing, which induce very

46 contrasted circulation patterns under fair weather or storm conditions and strongly constrain  
47 the vertical structure of the cross shore flow.

## 48 **1 Introduction**

49 The nearshore circulation driven by breaking waves contributes to the cross-shelf trans-  
50 port of material, especially in the vicinity of the surf zone, such as the transport of nutrient  
51 (e.g. Morgan et al., 2018) or sediment, which can result in large morphological changes  
52 under storm conditions (e.g. Wright & Short, 1984; Coco et al., 2014; Castelle et al., 2015).

53 Considering a weak along-shore variability of the topography and a shore normal inci-  
54 dence of waves, the interplay between waves and currents most notably drives the so-called  
55 undertow. Based on the depth-integrated continuity equation, the undertow commonly des-  
56 ignates the time- and depth-averaged Eulerian offshore flow compensating for the onshore  
57 mass transport associated with the Stokes drift. The onshore-directed mass transport is  
58 further enhanced within the surf zone due to contribution from surface wave rollers (e.g.  
59 Svendsen, 1984a). The underlying dynamics were further investigated both theoretically  
60 and experimentally, providing some insights onto the vertical structure of the (Eulerian)  
61 cross-shore flow. Within the surf zone, several pioneering studies (e.g. Svendsen, 1984a;  
62 Stive & Wind, 1986; Deigaard et al., 1991; Haines & Sallenger Jr, 1994; Garcez Faria et  
63 al., 2000, among many others) proposed theoretical models, which all conceptually rely on  
64 the local imbalance between the depth-uniform barotropic pressure gradient associated with  
65 the wave setup and the depth-varying gradient of the wave radiation stresses. These models  
66 adequately predict parabolic velocity profile, whose curvature is a function of local wave  
67 quantities and the vertical eddy viscosity (e.g. Garcez Faria et al., 2000).

68 In contrast, the wave induced dynamics seaward of the surf zone received much less  
69 attention, particularly under storm waves. Most notably, Lentz et al. (2008) combined long  
70 term observations with a one-dimensional vertical model adapted from Xu and Bowen (1994)  
71 to study the vertical structure of the cross-shore flow up to the inner-shelf. These authors  
72 showed that the cross-shore velocity profiles seaward of the surf zone do not resemble the  
73 parabolic profiles observed within the surf zone, but exhibit a maximum near the surface,  
74 which is consistent with a balance between the Coriolis force associated with the offshore flow  
75 and the Stokes-Coriolis force, also referred to as the Hasselmann wave stress (K. Hasselmann,  
76 1970). As a result, the offshore flow tends to be equal in magnitude but opposite in direction



77 to the onshore Stokes drift velocity all along the water column, which implies a nearly depth-  
78 uniform zero cross-shore Lagrangian flow seaward of the surf zone.

79 In recent years, wave-averaged three-dimensional (3D) circulation models have been  
80 developed aiming to represent consistently the effect of short waves on the mean circulation  
81 for a wide range of nearshore, coastal and open-ocean applications. Several theoretical ap-  
82 proaches were proposed on the form of the wave-modified primitive equations that would  
83 be suitable for such models (e.g. see Bennis et al., 2011). The wave-averaged vortex force  
84 formalism, which separates conservative and non-conservative wave forcing on the 3D quasi-  
85 Eulerian mean circulation, constitutes a theoretically-robust framework employed within  
86 several of these modelling systems (e.g. Uchiyama et al., 2010; Kumar et al., 2012; Michaud  
87 et al., 2012; Delpey et al., 2014; Zheng et al., 2017; Gu erin et al., 2018). For nearshore appli-  
88 cations, non-conservative effects associated with wave energy dissipation processes through  
89 depth-induced breaking, whitecapping and bottom friction are expected to play a crucial  
90 role. While Smith (2006) consistently derived the contribution of these processes to the  
91 depth-integrated momentum equations, no definite theory exists to express these terms for  
92 the depth-resolving equations. In particular, it is assumed that the dissipation of wave en-  
93 ergy by breaking acts either like a surface stress on the mean flow or like a body force, in  
94 which case one can thus impose an empirical vertical distribution such that the breaking  
95 contribution applies at appropriate depths near the surface (e.g. Uchiyama et al., 2010). As  
96 pointed out by Raschle (2007), the wave-enhanced vertical mixing associated with the pro-  
97 duction of turbulence by breaking waves mostly controls the vertical shear of the horizontal  
98 current velocity so that the near-surface distribution of the momentum source sparsely mat-  
99 ters. In this regard, wave-averaged 3D circulation models are usually supplemented by a  
100 two-equation turbulence closure model, which allows to approximate the wave-enhanced  
101 turbulent kinetic energy (TKE) budget across the water column. There is a consensus in  
102 the literature to model the TKE injection with a flux-type boundary condition at the water  
103 surface assuming a power law for the decay of TKE (Umlauf & Burchard, 2003). Craig  
104 and Banner (1994) proposed to express the surface flux of energy injected into the water  
105 column in proportion to the surface wind friction velocity cubed. Following this approach,  
106 it is assumed that the energy flux from the wind to the wave field very closely matches  
107 that transferred from the wave field to the water column, which appears especially relevant  
108 for the deep ocean where breaking processes (whitecapping) significantly impacts the at-  
109 mospheric drag coefficient. For nearshore applications however, observations support the

110 fact that the surface flux of TKE scales with the energy dissipated through depth-induced  
111 breaking (e.g. Feddersen & Trowbridge, 2005). It is also interesting to note that the decay  
112 of TKE near the surface is particularly sensitive to the surface mixing length, which remains  
113 an empirically parameterized quantity (e.g. see Moghimi et al., 2016).

114 Among the above mentioned studies, Uchiyama et al. (2010), Kumar et al. (2012),  
115 Michaud et al. (2012) and Zheng et al. (2017) essentially detailed the implementation of the  
116 vortex force formalism within various modelling framework and further aimed to demon-  
117 strate the general applicability of this approach to study surf zone dynamics over commonly  
118 used study cases, including applications at Duck, N.C., which serves as a reference bench-  
119 mark. In addition, Kumar et al. (2012) also reproduced the results from Lentz et al. (2008)  
120 seaward of the surf zone using the same dataset. In a recent model-based study following  
121 Uchiyama et al. (2010), Wang et al. (2020) further discussed the effect of the bottom wave  
122 streaming, which is the stress along the direction of wave propagation that accompanies the  
123 wave energy dissipation by bottom friction. Most notably, their results tended to show that  
124 the Lagrangian overturning circulation within the surf zone could be substantially weakened  
125 by an opposite overturning cell arising seaward of the surf zone and extending within it,  
126 associated with the bottom wave streaming. Realistic applications of state-of-the-art, fully  
127 coupled, wave-averaged 3D circulation models in the nearshore region remain very scarce  
128 (Michaud et al., 2012; Delpey et al., 2014; Gu erin et al., 2018), especially under storm  
129 conditions, such that our comprehensive understanding of the wave-induced hydrodynamics  
130 remains somehow limited and the predictive skills of these models uncertain.

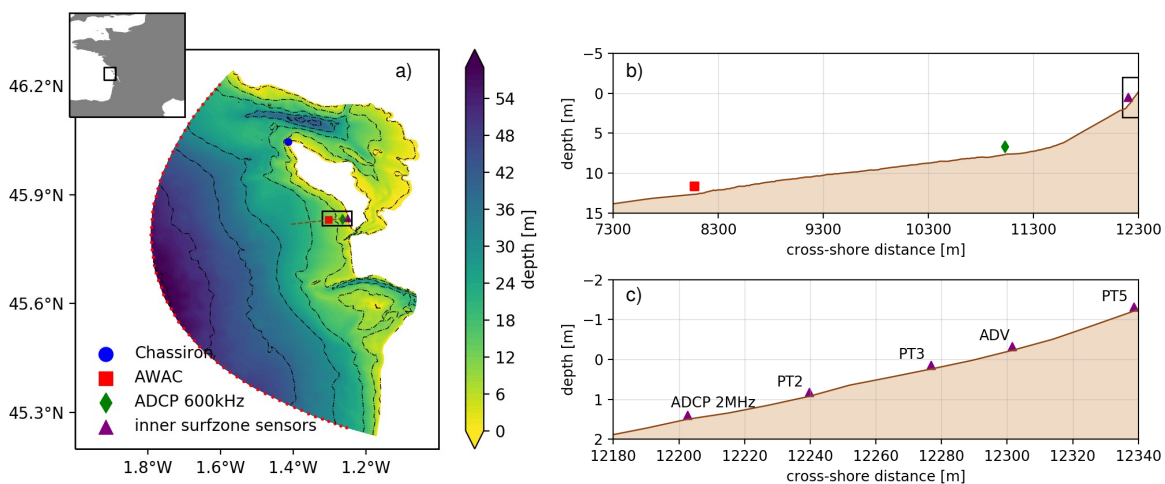
131 This study aims to explore the cross-shore distribution and the driving mechanisms  
132 of the wave-induced cross-shore circulation within the shoreface and the surf zone of a  
133 dissipative beach. We present unpublished results from a field campaign carried out in  
134 early 2021 in the central part of the French Atlantic coast under storm wave conditions,  
135 complemented with predictions from the state-of-the-art 3D circulation model SCHISM  
136 (Y. J. Zhang et al., 2016), fully coupled with the spectral wave model WWM (Roland et  
137 al., 2012). The manuscript is organized as follows. The study area, the field campaign and  
138 the processing of in-situ measurements are presented in Section 2. The parameterization  
139 of the modelling system is detailed in Section 3 and its predictive skills are assessed in  
140 Section 4 for the case study considered here. The section 5 discusses the contrasted wave-  
141 induced circulation patterns and associated driving mechanisms under high and moderate

142 wave energy conditions based on further numerical experiments. Finally, concluding remarks  
143 are provided in Section 6.

## 144 **2 Study area and field campaign**

### 145 **2.1 Study area**

146 The study area is located along the South-Western coast of the Oléron Island in the  
147 central part of the French Atlantic coast (see Fig. 1a), in front of Saint-Trojan Beach. This  
148 beach corresponds to a 8 km-long sandspit bounded to the South by the Maumusson Inlet  
149 and to the North by a rocky shore platform (Lavaud et al., 2020). In this region, tides  
150 are semi-diurnal and range from 1.5 m to 5.5 m, which corresponds to a macrotidal regime.  
151 Yearly mean wave conditions along the 30 m isobath are characterized by a significant wave  
152 height of 1.6 m, a mean wave period of 5.9 s and a direction of  $285^\circ$  from the true North  
153 (Dodet et al., 2019), but the offshore significant wave height can exceed 10 m with peak  
154 periods over 20 s (Bertin et al., 2015). This area is characterized by a very gently-sloping  
155 shoreface (the isobath 20 m being found approximately 10 km offshore) and a non-barred dis-  
156 sipative beach composed of fine sandy sediments and exposed to an energetic wave climate.  
157 Although this stretch of coast is relatively along-shore uniform, small amplitude inter-tidal  
158 bars can develop after the persistence of fair weather conditions (see Bertin et al., 2008;  
159 Guérin et al., 2018, for supplementary studies in this area).



**Figure 1.** a) Location of the study area in the Bay of Biscay, bathymetric map covering the computational domain (the open boundary is symbolised with the red dotted line) with isobaths reduced to the mean sea level displayed every 10 m (black dash-dotted lines), and position of the Chassiron meteorological station, AWAC, ADCP 600kHz and inter-tidal area sensors. A cross-shore profile extending from the isobath 25 m to PT5 sensor is also displayed (brown dashed line). b) Cross-shore profile from AWAC location to the first sensor deployed within the inter-tidal area. c) Zoom on inter-tidal area sensors.

## 160 2.2 Field campaign and data processing

161 The field campaign was carried out between January and February 2021 in two steps.  
 162 First, two Acoustic Doppler Current Profilers (ADCP) were mounted on structures anchored  
 163 in the seabed, at approximately 12.5 m- and 7.5 m-depth locations below Mean Sea Level  
 164 (MSL) for a long-term deployment between the 19<sup>th</sup> of January and the 26<sup>th</sup> of February (see  
 165 Fig. 1). The most offshore instrument is a high resolution ADCP (1 MHz) integrating an  
 166 Acoustic Surface Tracker (AST) and is hereafter referred to as the AWAC, whereas the other  
 167 one is a medium-resolution ADCP (600 kHz). Both instruments alternated a "current cycle"  
 168 and a "wave cycle" each hour. During the current cycle, 10 min-averaged velocity profile  
 169 measurements were collected along the vertical axis, whereas during the wave cycle velocity  
 170 measurements within a fixed 2 m-high cell and pressure measurements were performed at

171 2 Hz during 20 minutes. Second, a set of sensors was deployed in the inter-tidal area (Fig.  
 172 1c) between the 29<sup>th</sup> and the 31<sup>st</sup> of January during spring tides so as to capture a highly  
 173 energetic event associated with the storm Justine. The offshore significant wave height at the  
 174 Biscay Buoy location (5°W, 45.23°N) reached 10 m, which corresponds to a return period  
 175 of the order of one year (Nicolae-Lerma et al., 2015). The swell associated with the storm  
 176 reached the study area during the night of the 30<sup>th</sup> from a westward direction, while local  
 177 winds reached 15 m/s at the storm peak. This set of sensors was deployed along a cross-shore  
 178 profile and included one 2 MHz ADCP (with a similar data collection scheme than for the  
 179 two offshore ADCPs), three pressure transducers with a 2 Hz sampling frequency (PT) and  
 180 one Acoustic Doppler Velocimeter deployed 20 cm above the seabed, with a 16 Hz sampling  
 181 frequency (ADV). The PTs and ADV all performed continuous measurements.

182 For each sensor, sea-bottom pressure timeseries were split into 20 min-long bursts  
 183 (consistent with ADCPs "wave cycle"), corrected for sea level atmospheric pressure using  
 184 data collected at the nearby meteorological station of Chassiron (Fig. 1a), detrended and  
 185 converted into a sea-surface elevation signal assuming a hydrostatic pressure. For the sen-  
 186 sors deployed in the inter-tidal area, measurements below a burst-averaged water depth of  
 187 0.5 m were discarded due to the presence of substantial infra-gravity waves, which caused  
 188 the sensors to be intermittently dry. Then, pressure attenuation with depth due to non-  
 189 hydrostatic effects was corrected using the Transfer Function Method based on the linear  
 190 wave theory (TFM, e.g. Bishop & Donelan, 1987). This method requires an upper cutoff  
 191 frequency to remove high frequency noise that is amplified by the TFM correction, and to  
 192 prevent the over-amplification of high-frequency energy levels due to non-linear interactions  
 193 in intermediate and shallow-water depths (Mouragues et al., 2019). The cutoff frequency  
 194 was set to 0.2 Hz for the two offshore sensors and 0.4 Hz for the sensors in the inter-tidal  
 195 area. Finally, the sea surface elevation density spectra  $E(f)$  were computed by means of a  
 196 Fast Fourier Transform on 10 Hanning-windowed segments with a 50% overlap, which al-  
 197 lows a good compromise between statistical stability (20 degrees of freedom) and frequency  
 198 resolution (8.3 mHz). The processing of pressure measurements was further verified at the  
 199 AWAC location by comparing the sea-surface elevation spectra with those obtained from  
 200 the AST measurements. These measurements were not used directly due to spurious sig-  
 201 nals during the most energetic events, whose are probably explained by the presence of air  
 202 bubbles in the water column induced by the breaking of storm waves (not shown). Wave  
 203 bulk parameters (significant wave height,  $H_{m0}$ , mean and continuous peak periods,  $T_{m02}$

204 and  $T_{pc}$ ) were computed using the  $p^{\text{th}}$  moments of the spectra:

$$205 \quad m_p = \int_{f_{min}}^{f_c} f^p E(f) df \quad (1)$$

206 such that:

$$\begin{aligned} H_{m0} &= 4\sqrt{m_0} \\ T_{m02} &= \sqrt{\frac{m_0}{m_2}} \\ T_{pc} &= \frac{m_{-2}m_1}{m_0^2} \end{aligned} \quad (2)$$

208 For each sensor, an adaptive  $f_{min}$  value, defined as half the continuous peak frequency  
209 computed at the AWAC location, was used in order to separate the gravity from the infra-  
210 gravity bands (e.g. Hamm & Peronnard, 1997; Bertin et al., 2020).

211 For the two offshore ADCPs, 10 min-averaged vertical profiles of current velocities were  
212 acquired along bins spanned respectively every 1 m (AWAC) and 0.5 m (ADCP 600 kHz).  
213 The measurements above a distance equal to the water depth minus half the significant wave  
214 height were discarded due to contamination by surface reflections from the sidelobes of the  
215 ADCP acoustic pulses (Appell et al., 1991). Current velocity profile measurements from the  
216 ADCP 2MHz deployed in the inter-tidal area were discarded because of spurious bin-to-bin  
217 velocity differences. Finally, continuous velocity measurements from the ADV, were split  
218 into 30 min-long bursts and filtered from spikes using the phase-space thresholding method  
219 of Goring and Nikora (2002).

220 The inter-tidal topography was surveyed at low tide during the deployment (29/01/2021)  
221 and the recovery (31/01/2021) of the instruments with PPK GNSS over an area centered on  
222 the instrumented transect and extending 1 km along-shore. The comparison between both  
223 datasets showed very limited morphological changes (with a root mean square difference of  
224 0.10 m along the instrumented profile), a behaviour already reported by Guérin et al. (2018)  
225 under similar storm wave conditions. The subtidal bathymetry was surveyed at the location  
226 of the instrumented profile up to a water depth of 11 m below MSL four weeks after the  
227 deployment in the inter-tidal area by means of a Norbit multi-beam echo-sounder. This  
228 bathymetric dataset was merged with an extensive single-beam echo-sounder survey carried  
229 out in April 2013. Both datasets in the region where they overlap show a smooth transition  
230 in the subtidal zone with changes of the order of 0.4 m.

### 3 Modelling system

The modelling system used in this study couples the 3D circulation model SCHISM (Y. J. Zhang et al., 2016) and the third generation spectral wave model WWM (Roland et al., 2012). This modelling system offers the flexibility to cover large geographic areas with unstructured grid and very robust numerical schemes for both models. The 3D wave-induced circulation is modelled through the vortex force formalism, such as presented by Bennis et al. (2011). Its detailed implementation in SCHISM can be found in Gu erin et al. (2018) and is recalled in Appendix A. In the following, only the parameterization of the relevant part of the model and further improvements since Gu erin et al. (2018) are described.

#### 3.1 Parameterization of WWM

##### 3.1.1 Depth-induced breaking parameterization

The wave energy dissipation rate by depth-induced breaking is computed according to the model of Battjes and Janssen (1978) with an adaptive breaking coefficient ( $B$ ) as proposed by Pezerat et al. (2021). The local mean (phase-averaged) rate of energy dissipation per unit area  $D_{db}$  in ( $\text{W}/\text{m}^2$ ) reads:

$$D_{db} = \frac{B}{4} \rho g f_{mean} Q_b H_m^2 \quad (3)$$

where  $B = 40 \tan \beta$ , with  $\tan \beta$  the local bottom slope;  $f_{mean}$  is a mean frequency, usually computed as the ratio  $m_1/m_0$  (see Eq. 1);  $Q_b$  is the local fraction of breaking (and broken) waves and  $H_m$  is the local maximum possible wave height estimated by means of a parameterized Miche-type breaking criterion. Under the shallow water assumption,  $H_m$  reads:

$$H_m = \gamma h \quad (4)$$

where  $\gamma$  is the breaking index, an adjustable coefficient, usually kept constant at 0.73 following the calibration performed by Battjes and Stive (1985). However, as pointed out by Pezerat et al. (2021), the introduction of the adaptive breaking coefficient requires a newly calibrated breaking index. Based on sensitivity tests performed with the model on  $H_{m0}$  results considering the entire dataset (not shown), a constant value of  $\gamma = 0.60$  was considered for this study. It is worth noting that this value might show some site- or wave conditions-specificity, but it is not the purpose of this study to propose an extensive calibration of the breaking index. Finally, following the approach of Eldeberky and Battjes (1996),

261 the corresponding source term in WWM is computed by distributing  $D_{db}$  over frequencies  
 262 and directions in proportion to the spectral action density:

$$263 \quad S_{db} = -\frac{D_{db}}{E_{tot}}N \text{ where } E_{tot} = \rho g \int_{\sigma} \int_{\theta} E d\sigma' d\theta' \quad (5)$$

### 264 **3.1.2 Other source terms**

265 The wind input ( $S_{in}$ ) and dissipation by whitecapping ( $S_{ds}$ ) are formulated according  
 266 to the parameterization of Ardhuin et al. (2010). Non-linear quadruplet interactions ( $S_{nl,4}$ )  
 267 are taken into account following the approach of S. Hasselmann et al. (1985), whereas the  
 268 non-linear triad interactions ( $S_{nl,3}$ ) are parameterized following the approach of Eldeberky  
 269 (1997). Finally, the formulation of the energy dissipation by bottom friction ( $S_{bf}$ ) is adapted  
 270 from Tolman (1994). This formulation is based on the model of Madsen et al. (1988)  
 271 with a parameterization of the roughness issued from Grant and Madsen (1982), and was  
 272 subsequently adapted by Ardhuin et al. (2003) based on the SHOWEX experiment.

## 273 **3.2 Parameterization of SCHISM**

### 274 **3.2.1 Non-conservative wave forces**

275 The non-conservative wave forcing due to wave breaking [ $F_{br,x}, F_{br,y}$ ] includes the  
 276 effect of whitecapping and depth-induced breaking further modified due to the action of  
 277 wave rollers (cf. Appendix A for the description of the roller model):

$$278 \quad [\hat{F}_{br,x}, \hat{F}_{br,y}] = \frac{\delta_{z,\bar{\eta}}}{\rho} \left( - \int_{\sigma} \int_{\theta} k ((1 - \alpha_R) S_{db} + S_{ds}) [\cos \theta', \sin \theta'] d\sigma' d\theta' + \frac{D_R}{\sigma_p} k_p [\cos \theta_m, \sin \theta_m] \right) \quad (6)$$

279 where the vertical distribution function is given by a delta function  $\delta_{z,\bar{\eta}}$ , such that the  
 280 breaking-induced momentum is injected within the first cell below the free surface (Bennis  
 281 et al., 2011). The roller model shows barely any effect at both offshore locations, while the  
 282 current intensity slightly increases with  $\alpha_R$  at the ADV location (not shown). The model  
 283 best-fit results were obtained with  $\alpha_R = 0.5$ .

284 The bottom streaming corresponds to the stress along the direction of wave propaga-  
 285 tion that accompanies the dissipation of wave energy by bottom friction within the wave  
 286 boundary layer (Longuet-Higgins, 1953). The corresponding body force [ $F_{ws,x}, F_{ws,y}$ ] is for-  
 287 mulated by means of an upward decaying vertical distribution of the wave energy dissipation



288 by bottom friction (Uchiyama et al., 2010):

$$289 \quad [\hat{F}_{ws,x}, \hat{F}_{ws,y}] = -\frac{f^{ws}(z)}{\rho} \int_{\sigma} \int_{\theta} k S_{bf} [\cos \theta', \sin \theta'] d\sigma' d\theta' \quad (7)$$

290 where  $f^{ws}$  is a vertical distribution function given by:

$$291 \quad f^{ws}(z) = \frac{1 - \tanh(k_{wd}(d+z))^2}{\int_{-d}^{\bar{\eta}} 1 - \tanh(k_{wd}(d+z))^2 dz} \quad (8)$$

292 where  $1/k_{wd} = a_{wd}\delta_{wbl}$  is a decay length proportional to the wave boundary layer thickness  
 293 ( $\delta_{wbl}$ ). Within SCHISM,  $\delta_{wbl}$  is derived from the wave boundary layer model of Madsen  
 294 (1995) that is used to compute the apparent roughness length for the parameterization of  
 295 the bottom friction within the circulation model (cf. section 3.2.3). The proportionality  
 296 coefficient  $a_{wd}$  is taken equal to unity, such that the decay length matches the theoretical  
 297 wave boundary layer thickness for monochromatic waves although laboratory measurements  
 298 of the bottom boundary layer under random waves suggest a significant increase in the  
 299 thickness, i.e.  $a_{wd} > 1$  (Klopman, 1994).

### 300 **3.2.2 Wave-enhanced turbulence at the surface**

301 The circulation model is supplemented by a  $K - \omega$  turbulence closure model retrieved  
 302 from the Generic Length Scale (GLS) two-equation turbulence closure model within the  
 303 framework of the General Ocean Turbulence Model (GOTM, Burchard et al., 1999; Umlauf  
 304 et al., 2005). At the water surface, the turbulence closure model accounts for a TKE injection  
 305 by breaking waves ( $K$  in  $\text{m}^2/\text{s}^2$ ) through a flux-type boundary condition assuming a power  
 306 law for the vertical decay of  $K$  (Umlauf & Burchard, 2003), which reads:

$$307 \quad \frac{\nu}{\sigma_K} \frac{\partial K}{\partial z} = F_K \left( \frac{z_0^s - z'}{z_0^s} \right)^{\frac{3}{2}\alpha} \quad \text{at } z = \bar{\eta} \quad (9)$$

308 where  $F_K$  (in  $\text{m}^3/\text{s}^3$ ) is the surface flux of energy injected into the water column,  $\nu$  is the  
 309 vertical eddy viscosity,  $\sigma_K$  is the turbulent Schmidt number for  $K$ ,  $\alpha$  is the spatial decay  
 310 rate of TKE in the wave enhanced layer,  $z_0^s$  is the surface mixing length and  $z'$  is the distance  
 311 below the surface at which the flux is imposed. For numerical reason,  $z'$  is prescribed as  
 312 half the height of the top cell, such that the boundary condition for  $K$  requires a refined  
 313 discretization of the vertical grid near the surface. The surface mixing length ( $z_0^s$ ) controls  
 314 the depth of penetration for the injected TKE. There are strong uncertainties over this  
 315 quantity, which has been either parameterized as a constant (e.g.  $z_0^s = 0.2\text{ m}$ , Feddersen  
 316 & Trowbridge, 2005) or as a function of the significant wave height:  $z_0^s = \alpha_w H_{m0}$ , with  
 317  $\alpha_w = \mathcal{O}(1)$  (see Moghimi et al., 2016, for a short review). Following the approach of

318 Feddersen and Trowbridge (2005), the surface flux of TKE injected at the surface scales  
 319 with the energy dissipated through wave-related processes at the surface:

$$320 \quad F_K = \frac{c_{db}}{\rho} \left( -(1 - \alpha_R) \int_{\sigma} \int_{\theta} \sigma' S_{db} d\sigma' d\theta' + D_R \right) - \frac{c_{ds}}{\rho} \int_{\sigma} \int_{\theta} \sigma' S_{ds} d\sigma' d\theta' \quad (10)$$

321 where the coefficients  $c_{db}$  and  $c_{ds}$  control the amount of energy injected to the water column.  
 322 The range of value 0.01 to 0.25 has been proposed in the literature for  $c_{db}$  (e.g. Feddersen  
 323 & Trowbridge, 2005; Huang et al., 2009; Feddersen, 2012a, 2012b), while Paskyabi et al.  
 324 (2012) suggest  $c_{ds} \simeq 1$ . The model best-fit results were obtained with  $z_0^s = \alpha_w H_{m0}$ ,  $\alpha_w = 1$ ;  
 325  $c_{db} = 0.15$  and  $c_{ds} = 1$ .

### 326 **3.2.3 Bottom boundary condition**

327 The bottom boundary condition imposes a balance between the internal Reynolds  
 328 stress and the bottom frictional stress  $[\tau_{b,x}, \tau_{b,y}]$ :

$$329 \quad \nu \frac{\partial[\hat{u}, \hat{v}]}{\partial z} = \frac{[\tau_{b,x}, \tau_{b,y}]}{\rho} \text{ at } z = -d \quad (11)$$

330 The law of the wall is then assumed, leading to a logarithmic profile for the velocity within  
 331 a constant stress layer that presumably contains the bottom cell while the bottom stress is  
 332 formulated with a quadratic bottom drag parameterization. The bottom stress finally reads  
 333 (e.g. Blumberg & Mellor, 1987):

$$334 \quad [\tau_{b,x}, \tau_{b,y}] = \rho C_d \sqrt{\hat{u}_b^2 + \hat{v}_b^2} [\hat{u}_b, \hat{v}_b] \text{ with } C_d = \left( \frac{\kappa}{\ln\left(\frac{\delta_b}{z_0}\right)} \right)^2 \quad (12)$$

335 where  $C_d$  is the friction factor,  $[\hat{u}_b, \hat{v}_b]$  is the velocity at the top of the bottom computational  
 336 cell,  $\kappa$  is the von Kármán's constant,  $\delta_b$  is the thickness of the bottom cell (in m) and  $z_0$  is  
 337 the bottom roughness length (in m). In the presence of waves, the wave-current interaction  
 338 theory by Madsen (1995) as modified by Mathisen and Madsen (1999) is applied to compute  
 339 an apparent roughness length  $z_0^a$ , which further replaces  $z_0$  in the expression of  $C_d$ , and  
 340 thus allows to account for the enhanced roughness experienced by the current in presence of  
 341 waves. This approach takes as input a (physical) bottom roughness length ( $z_0$ ) to provide  
 342 an expression of the wave boundary layer thickness and  $z_0^a$  following a numerical procedure  
 343 described in H. Zhang et al. (2004). In this study,  $z_0$  varies spatially between 0.1 mm and  
 344 5 mm to account for the seabed granulometric variability based on the sea-bottom nature  
 345 map provided by the Hydrographic and Oceanographic French Office.

### 3.3 Model implementation

The coupling between SCHISM and WWM is made at the source code level. Both models share the same unstructured grid and domain decomposition. The horizontal spatial resolution ranges from 2 km at the offshore boundary down to 20 m in the surf zone. The vertical grid for the circulation model is discretized using 25 S-levels stretched near the surface and the bottom. The time step for the circulation model is set to 10 s whereas WWM is running in implicit mode (Booij et al., 1999; Abdolali et al., 2020). This allows to relax the constraint for the time step of the wave module, which was set to 300 s. Finally, the spectral space in WWM was discretized in 36 directions covering the entire trigonometric circle and 24 frequencies ranging from 0.03 to 0.4 Hz.

At the domain offshore boundaries, the tidal forcing was computed considering the 16 main constituents linearly interpolated from the regional tidal model of Bertin et al. (2012), whereas WWM was forced with timeseries of energy spectra obtained from a North Atlantic application of the spectral wave model WaveWatch III (WW3, Tolman, 1991). For both SCHISM and WWM, the atmospheric forcing consisted of MSL pressure and wind speed at 10 m issued from the meteorological operational model ARPEGE (e.g. Déqué et al., 1994), interpolated onto a  $0.1^\circ$  regular grid. WW3 was forced with wind fields at 10 m originating from the Climate Forecast System Reanalysis (CFSR, Saha et al., 2011) extracted from a  $0.2^\circ$  regular grid covering the entire North Atlantic basin. ARPEGE was preferred to CFSR for our local application of the modelling system considering its slightly improved predictive skills as compared to measurements of wind speed and direction at the nearby meteorological station of Chassiron (Fig. 1).

## 4 Observations and predictive skills of the model

A comparison between measurements and model best-fit results (run Rref) is presented. The overall model performance for each quantity is assessed with the normalized bias (NB) and the normalized root mean square error (NRMSE):

$$\text{NB}(X) = \begin{cases} \frac{\overline{\hat{X}-X}}{(\max X - \min X)} \times 100 & \text{if } \overline{X} \simeq 0 \\ \frac{\overline{\hat{X}-X}}{\overline{X}} \times 100 & \end{cases} \quad (13)$$

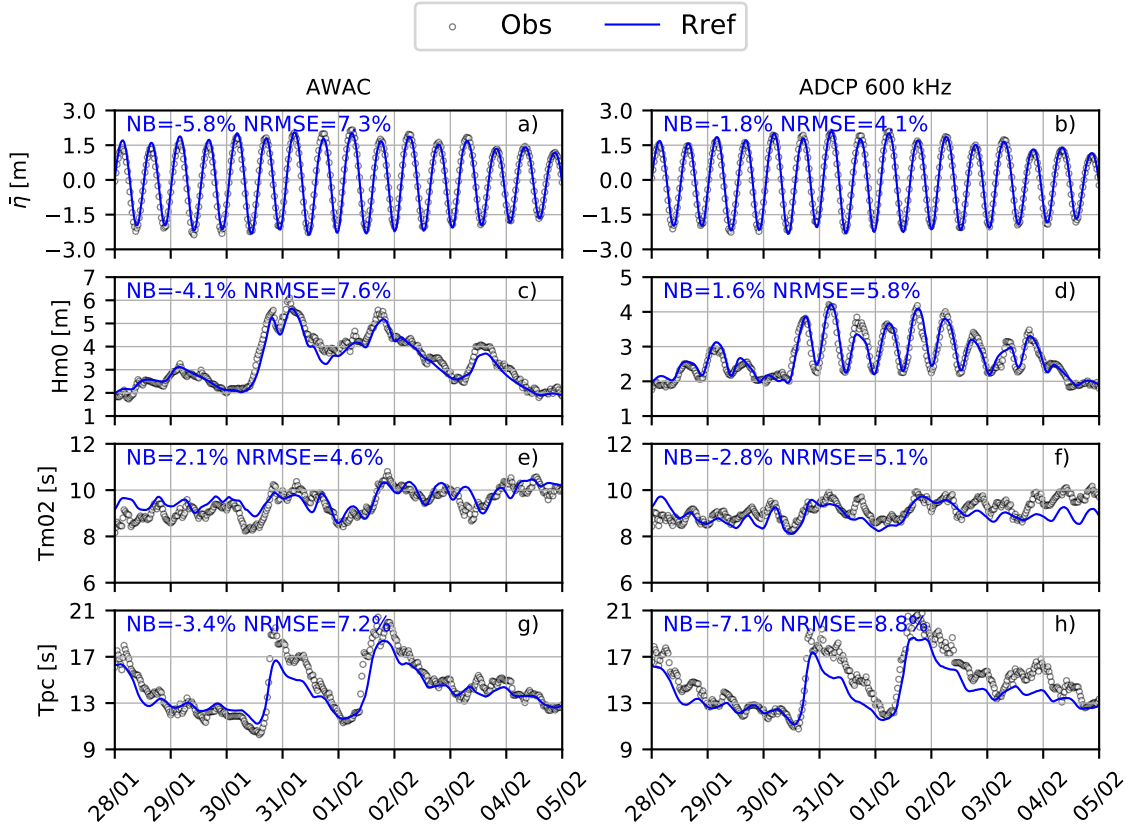
$$\text{NRMSE}(X) = \begin{cases} \frac{\sqrt{\overline{(\hat{X}-X)^2}}}{(\max X - \min X)} \times 100 & \text{if } \overline{X} \simeq 0 \\ \sqrt{\frac{\overline{(\hat{X}-X)^2}}{\overline{X^2}}} \times 100 & \end{cases} \quad (14)$$

375 where  $X$  and  $\hat{X}$  respectively correspond to the measured and modelled quantity and the  
 376 overbar denotes the average over the timeseries.

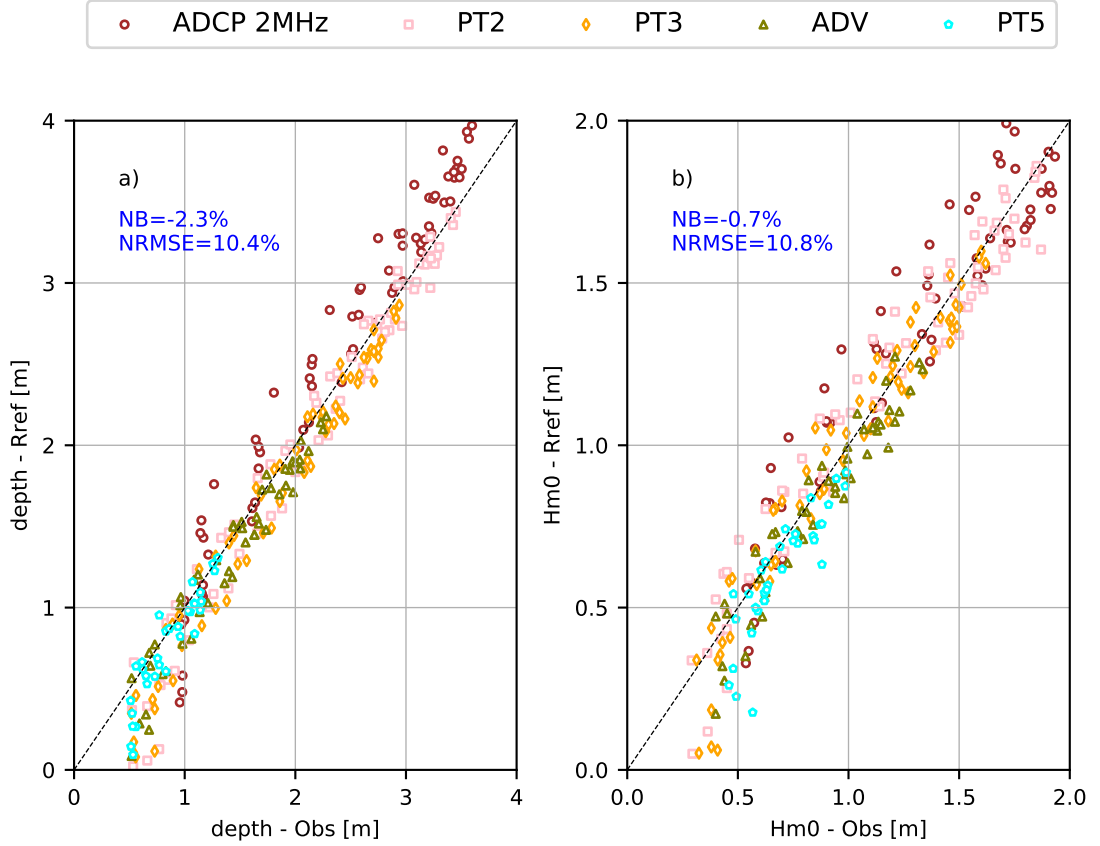
#### 377 4.1 Water levels and short waves

378 Phase-averaged surface elevation variations are very well reproduced by the model at  
 379 the AWAC and ADCP 600 kHz locations (Fig. 2a-b), with a NRMSE lower than 6% obtained  
 380 at both locations. The negative bias observed at the AWAC location (NB = -4.1%) could  
 381 be attributed to pressure sensor drift, as it increases in time while such bias is not observed  
 382 at the nearby ADCP 600 kHz location. Model results on short waves are compared to  
 383 observations by means of bulk parameters computed with moments integrated over the same  
 384 frequency range as the data. These are very well reproduced by the model at both locations  
 385 (Fig. 2c-h) with a NRMSE on both  $H_{m0}$  and  $T_{m02}$  below 7%, only the continuous peak  
 386 period is slightly underestimated at the storm peak, resulting in a negative NB of -7.3% at  
 387 the ADCP 600 kHz location. This problem might be explained by the fact that spectral wave  
 388 models allow to represent energy transfers towards higher harmonics by sum self interactions  
 389 with the LTA formulation of (Eldeberky, 1997) but not by difference interactions. Thus,  
 390 the model cannot represent transfers from the gravity band towards the IG band and the  
 391 subsequent transfers back to the gravity band by the generation of IG wave higher harmonics  
 392 (e.g. Bertin et al., 2020), nor direct transfers towards low frequencies in the gravity band  
 393 (e.g. De Bakker et al., 2015). Both field measurements at the ADCP 600 kHz location and  
 394 model results show that  $H_{m0}$  is tidally modulated when exceeding approximately 2 m at the  
 395 AWAC location (Fig. 2d). Such modulation also occurs at the storm peak at the AWAC  
 396 location ( $H_{m0} \simeq 6$  m, Fig. 2c). This clearly suggests that depth-induced breaking starts  
 397 being substantial as far as 4 km away from the shoreline ( $\simeq 12.5$  m-depth) under energetic  
 398 conditions. In addition  $T_{m02}$  period is also tidally modulated, highlighting the significant  
 399 contribution of the non-linear triad interactions, whose intensity varies with the water-depth.  
 400 As the water-depth decreases, more energy is transferred towards the higher harmonics by  
 401 non-linear triad interactions, such that  $T_{m02}$  decreases. Within the inter-tidal area most  
 402 of the differences between model results and measurements are observed for water-depths  
 403 below 1 m (see Fig. 3). Water levels and significant wave heights are well reproduced by  
 404 the model with a NRMSE computed over all sensors that respectively reaches 10.4% and  
 405 10.8%. The  $T_{m02}$  period is not shown because a significant amount of energy is transferred  
 406 towards the IG band, which can partly go back to the gravity band through the generation

407 of IG wave higher harmonics (Bertin et al., 2020), resulting in an increase of  $T_{m02}$ , a process  
 408 that cannot be reproduced by the phase-averaged model.



**Figure 2.** Measured and simulated phase-averaged free surface elevation ( $\bar{\eta}$ ), significant wave height ( $H_{m0}$ ), mean period ( $T_{m02}$ ) and continuous peak period ( $T_{pc}$ ) timeseries at the AWAC (left panels) and ADCP 600 kHz (right panels) locations using the configuration of reference (Rref).



**Figure 3.** Measured and simulated water depth (a) and  $H_{m0}$  (b) scatter plot at the inner surfzone sensors locations using the configuration of reference (Rref).

409

## 4.2 Cross-shore circulation

410

411

412

413

414

415

416

417

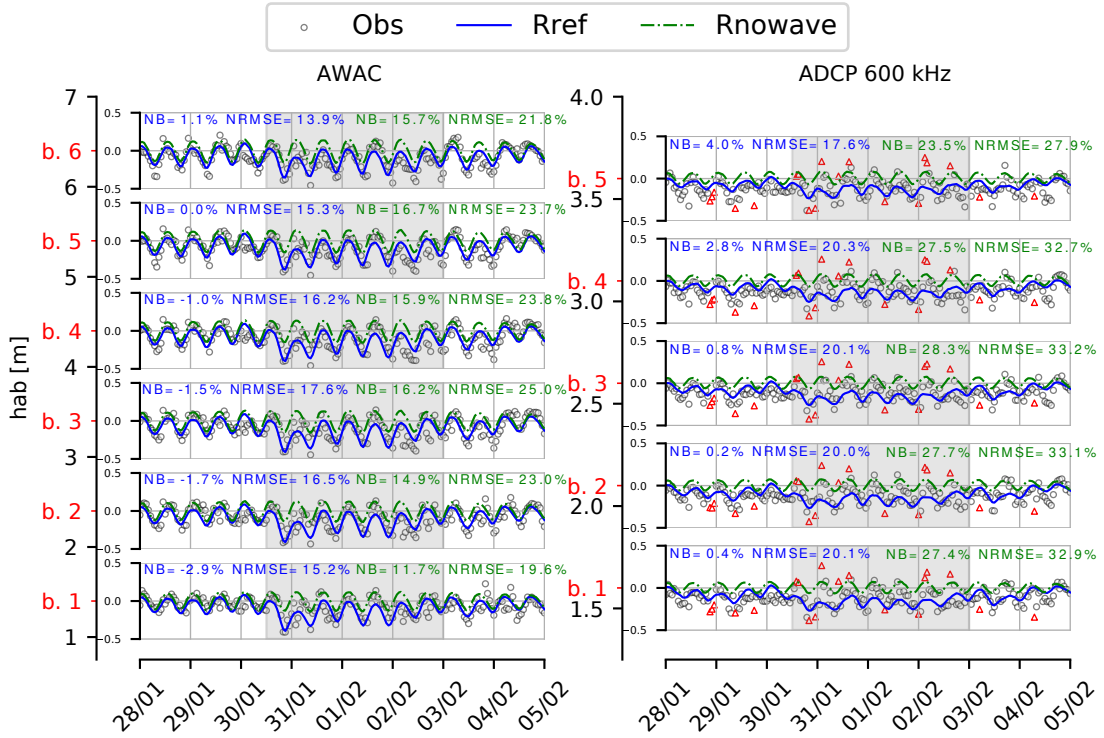
418

419

420

The model results for Rref on the cross-shore velocity component show a fairly good agreement with measurements from the AWAC and the ADCP 600 kHz as shown Fig. 4 below the lowest sea surface tidal level. Overall, the NRMSE varies between 13.9% and 20%, while the NB fluctuates between  $-2.9\%$  and  $5.6\%$ . Note, however, that measurements from the ADCP 600 kHz are more scattered under energetic wave conditions. First, this might be partly due to measurement artefacts associated with the generation of bubbles in the water column by depth-induced breaking, which is particularly active at this location (cf. section 4.1). Second, as vertical current profile measurements are 10 min-averaged, there may be an aliasing of the signal associated with currents induced by IG waves, whose period can exceed 300s at this beach under storm conditions (Bertin et al., 2020). As averaging currents over a longer period within a thicker cell results in a more accurate measurement

421 of the current velocity, we arbitrarily discarded current profile velocity measurement that  
 422 departed by more than 0.15 m/s from the 20 min-averaged velocity measurement performed  
 423 during the wave cycle within the fixed 2 m-high cell (see the red triangles in Fig. 4). Both  
 424 model results and measurements clearly show that the cross-shore velocity is mostly offshore-  
 425 directed under energetic wave conditions reaching almost -0.5 m/s at the AWAC location  
 426 around the 31/01/2021. The model results for a run performed without waves (Rnowave),  
 427 which only accounts for the tides and the atmospheric forcing, show a strong positive bias  
 428 at both locations (e.g. the NB reaches 27.3% at the ADCP 600 kHz location), so that the  
 429 comparison between Rref and Rnowave highlights the significant contribution of the wave-  
 430 induced current to the cross-shore flow. Based on this comparison, the contribution of the  
 431 wave-induced current at the peak of the storm reaches as much as 0.25 m/s.

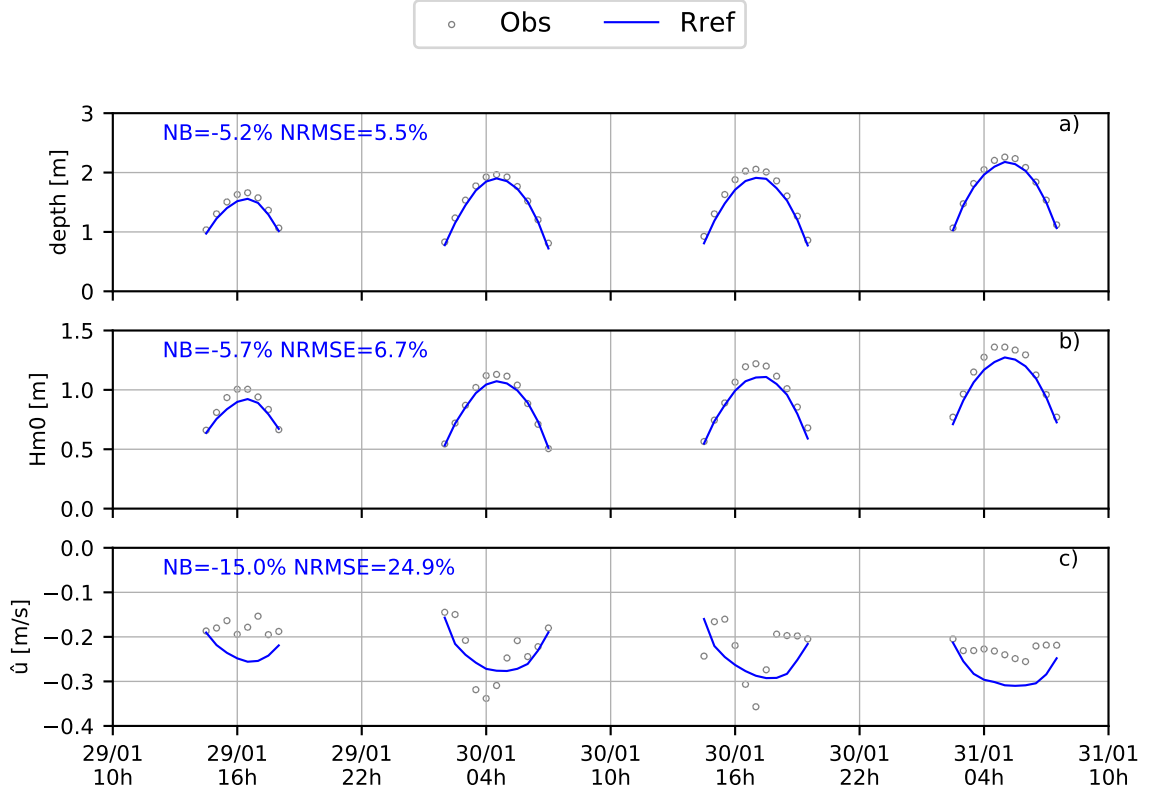


**Figure 4.** Measured and simulated cross-shore velocity component timeseries at the AWAC (left panel) and ADCP 600 kHz (right panel) locations displayed for each vertical bin below the sea surface lowest level. Model results are presented for the run of reference (Rref) and for a run performed without waves (Rnowave). Red triangles tag the discarded measurements. The vertical position of each bin (denoted b) is measured from the bottom and hab stands for height above the sea-bed. The time range of the storm event is delimited by the grey background.

432 At the ADV location, burst-averaged cross-shore velocity measurements are quite scat-  
433 tered, especially during the second and third tidal cycle with fluctuations reaching 0.2 m/s.  
434 The 30-min averaged current velocities suggest the presence of Very Low Frequency oscilla-  
435 tions (VLF, frequencies below 4 mHz), while IG waves contribution was presumably filtered.  
436 The analysis of the whole dataset revealed that waves were mostly normally incident during  
437 the storm event, which is quite common during energetic event at this site (Bertin et al.,  
438 2008). As a consequence, mean longshore currents remained weak during the field campaign  
439 (ranging from -0.10 to 0.15 m/s) and were alternatively northward and southward-directed  
440 within a very wide surf zone (not shown). Shear instabilities of mean longshore currents,  
441 which require the presence of a strong shear (e.g., associated with highly-oblique large  
442 waves breaking over a bar, see Oltman-Shay et al., 1989; Noyes et al., 2004) cannot there-  
443 fore explain such VLF motions. Instead, it could be attributed to the breaking of energetic  
444 wave groups that has been identified as a mechanism for the generation of surf zone eddies  
445 through the generation of vorticity at the scale of individual waves or wave groups (Long  
446 & Özkan-Haller, 2009; Feddersen, 2014), which is then transferred to VLF frequencies and  
447 larger spatial scales through non-linear inverse energy cascades (Feddersen, 2014; Elgar &  
448 Raubenheimer, 2020).

449 These complex dynamical features cannot be reproduced by the present phase-averaged  
450 modelling approach, which undermines the comparison with the field observations (the  
451 NRMSE on the cross-shore velocity component reaches 24.9%). Both model results and  
452 observations qualitatively show that the cross-shore velocity component is dominated by a  
453 wave-induced seaward-oriented current, whose intensity increases with the significant wave  
454 height (Fig. 5b-c). As more energy is dissipated by depth-induced breaking when the sig-  
455 nificant wave height increases, it further suggests that non-conservative breaking wave force  
456 strengthens locally this seaward-oriented current.





**Figure 5.** Measured and simulated water depth,  $H_{m0}$  and cross-shore velocity component time-series at ADV location.

## 5 Discussion

The results clearly show that the wave-induced circulation plays a crucial role on the cross-shore flow. Circulation patterns show a strong seaward-directed current in the lower part of the water column as far as 4 km from the shoreline, which cannot be reproduced by solely accounting for the wind and the tidal forcing. This unsteady cross-shore circulation is quite well reproduced by the model, which shows excellent predictive skills for short waves with errors on bulk parameters ranging from 4 to 9%. Switching off the wind does not significantly impact the results on short waves because the local wave growth remains weak (not shown). A new configuration of the model was thus setup with wind and tidal forcing switched off in order to investigate the driving mechanisms of the wave-induced circulation while bypassing the unsteadiness associated with tides and wind. This configuration was run for two distinct 24 h-periods associated with storm waves ( $H_{m0} = 5.3$  m - 30/01/2021) and moderate wave energy conditions ( $H_{m0} = 2.0$  m - 04/02/2021).

470 **5.1 Wave energy dissipation patterns**

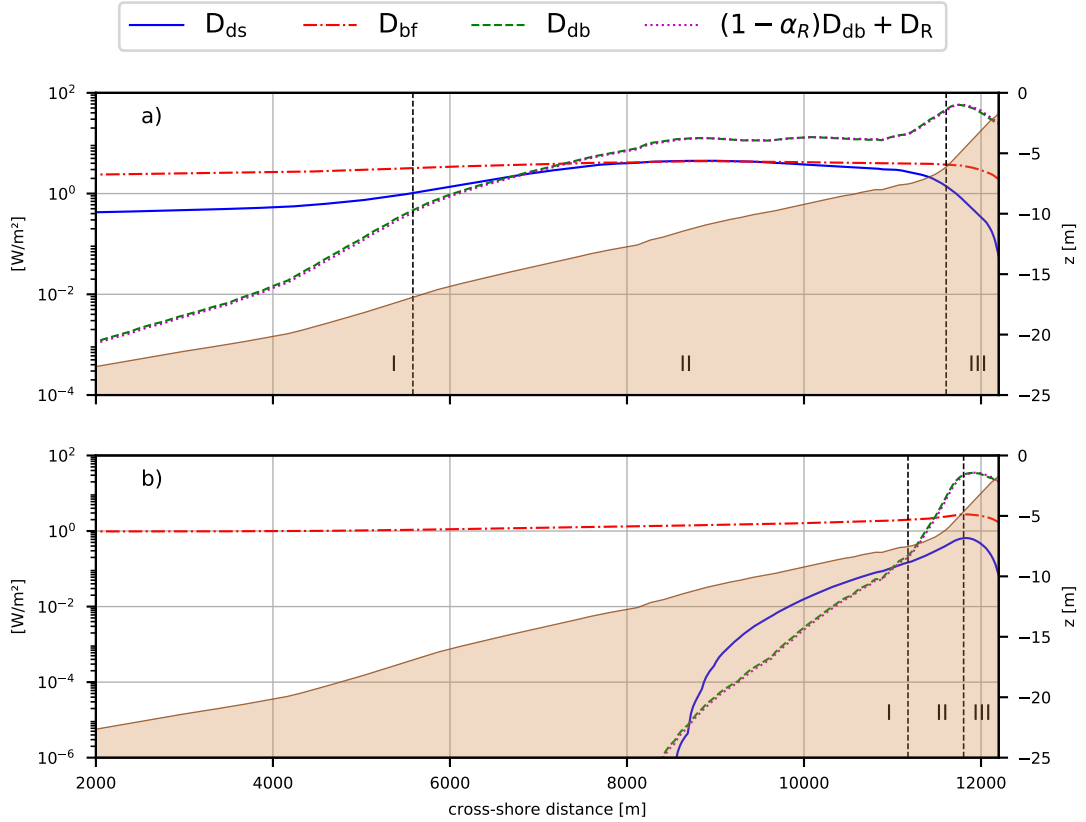
471 Considering the significant contribution of the wave energy dissipation processes either  
 472 on the TKE injection or through the non-conservative wave forces, the energy dissipation  
 473 rates associated with depth-induced breaking ( $D_{db}$ ), whitecapping ( $D_{ds}$ ) and bottom friction  
 474 ( $D_{bf}$ ) are first examined following the approach of Pezerat et al. (2021), based on the  
 475 empirical ratio  $R$  reading:

$$476 \quad R = \frac{D_{db}}{(D_{ds} + D_{bf} + D_{db})} \text{ with } D_x = \int_{\sigma} \int_{\theta} \sigma' S_x d\sigma' d\theta' \quad (15)$$

477 where the subscript  $x$  represents either of the "db", "ds" or "bf" subscripts. As shown  
 478 in Fig. 6 along a cross-shore profile (see Fig. 1), one can distinguish three typical areas.  
 479 Offshore (zone I,  $R < 0.1$ ), wave energy dissipation is mostly associated with bottom friction.  
 480 Closer to shore, a transition zone appears (zone II,  $0.1 < R < 0.9$ ), within which bottom  
 481 friction, whitecapping and depth-induced breaking together significantly contribute to the  
 482 incident wave energy dissipation, before depth-induced breaking becomes dominant (zone  
 483 III,  $R > 0.9$ ). Most notably, the transition zone is much wider under high wave energy  
 484 conditions (approximately 6 km-long spanning from approximately 15 m- to 5 m-depth) than  
 485 under moderate wave energy conditions (less than 1 km-long with water depth comprised  
 486 between 5 m and 7 m). The edge of the third zone is approximately located at the same  
 487 distance from the shoreline in both situations ( $\simeq 5$  m-depth). Interestingly, the wave energy  
 488 dissipation rates associated with whitecapping and bottom friction are of the same order  
 489 within the zones I and II under high wave energy conditions, and even dominate the depth-  
 490 induced breaking contribution over approximately 5 km up to a water-depth of the order  
 491 of 15 m (Fig. 6a). Although it is commonly understood that the surf zone is widening  
 492 under storm waves, it appears more appropriate to introduce this conceptual transition  
 493 zone, while the third zone more likely corresponds to the inner surf zone. It is worth noting  
 494 that the development of this wide transition zone is also presumably related to the gentle  
 495 and smoothly increasing bottom slope that characterizes the study area, while the transition  
 496 to a regime dominated by depth-induced breaking is much more abrupt over barred beach or  
 497 fringing environments. It would be thus interesting to perform similar analysis in contrasted  
 498 environments, with a diversity of beach profiles and bottom substrate.

499 The roller model shows a weak effect on the location where energy is dissipated. This  
 500 could be presumably attributed to the gentle bottom slope characterizing the study area,

501 while previous studies highlighted the significant roller effect over a barred and steeper beach  
 502 (e.g. Reniers et al., 2004; Uchiyama et al., 2010; Kumar et al., 2012; Zheng et al., 2017).



**Figure 6.** Wave energy dissipation rates profiles associated with bottom friction ( $D_{bf}$ ), whitecapping ( $D_{ds}$ ), depth-induced breaking ( $D_{db}$ ) and roller energy dissipation rates ( $(1 - \alpha_R)D_{db} + D_R$ ) under high (a) or moderate (b) wave energy conditions.

503 **5.2 Sensitivity to vertical mixing**

504 The results presented above show a substantial wave energy dissipation associated with  
 505 whitecapping and depth-induced breaking as far as 6 km from the shoreline under energetic  
 506 conditions, which further suggest that a substantial amount of TKE could be injected at the  
 507 surface. Three principal parameters control the injection of TKE at the surface ( $c_{db}$  and  $c_{ds}$ )  
 508 and its vertical decay ( $z_0^s$ ). If we consider the default value taken for  $c_{db}$  and  $c_{ds}$ , we note  
 509 that for equivalent energy dissipated through whitecapping and depth-induced breaking (i.e.  
 510  $D_{ds} \sim D_{db}$ ), more weight is given to the whitecapping contribution. Previous studies have

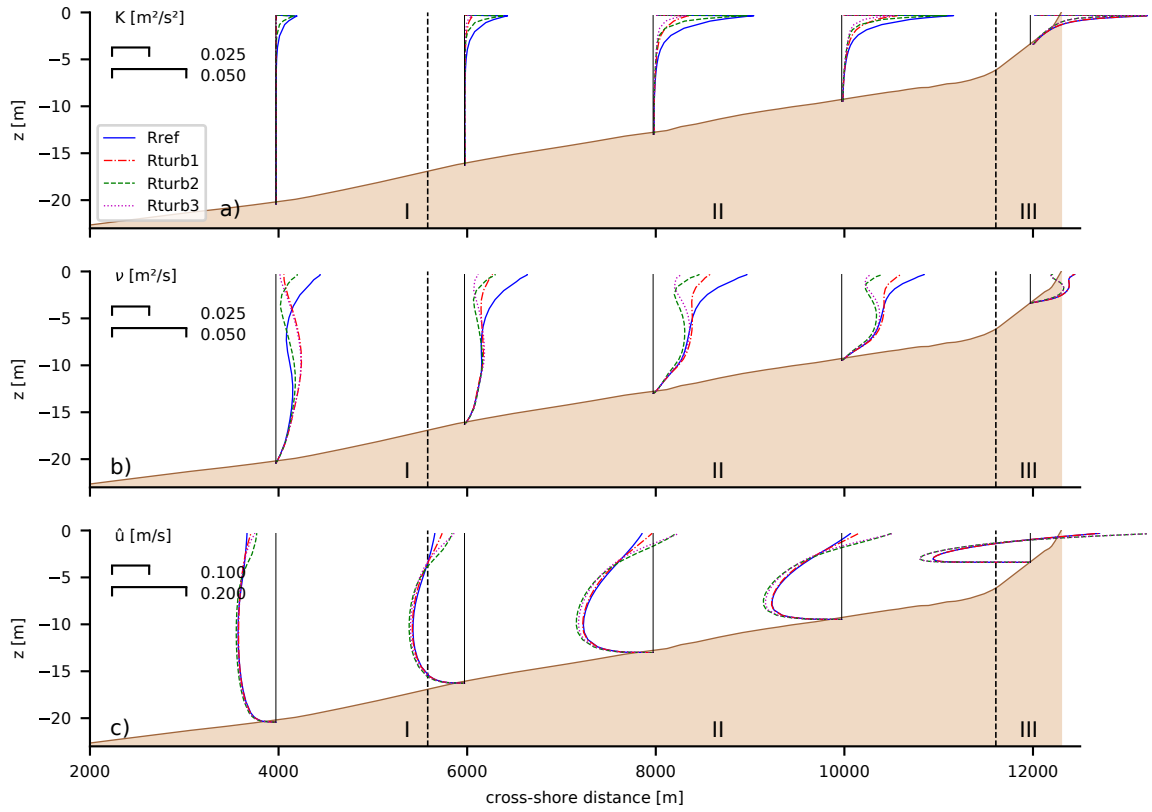
511 highlighted the effect of breaking-wave-generated turbulence on the mean circulation within  
 512 the surf zone, which results in a reduction of the vertical shear of the horizontal current  
 513 (e.g. Feddersen & Trowbridge, 2005; Kumar et al., 2012). In intermediate water depths,  
 514 Paskyabi et al. (2012) showed that surface currents associated with Ekman transport are  
 515 better reproduced when accounting for whitecapping contribution to the wave-enhanced  
 516 mixing, while Lentz et al. (2008) used a crude parameterization based on the wind stress  
 517 for the vertical eddy viscosity for their study over the inner continental shelf. Overall, the  
 518 contribution of the wave-enhanced turbulence to the mean circulation under the combined  
 519 effects of depth-induced breaking and whitecapping has thus never been evaluated across  
 520 the shoreface (note that Kumar et al., 2012, gathered both contributions to compute the  
 521 surface flux of TKE and used  $c_{db} = c_{ds}$ ). In addition, a sensitivity analysis of the vertical  
 522 shear of the mean cross-shore current to the parameterization of the surface mixing length  
 523  $z_0^s$  was performed to supplement the study of Moghimi et al. (2016), who highlighted the  
 524 sensitivity of turbulence closure models to  $z_0^s$  because of the power law for the decay of the  
 525 TKE. Three additional runs were thus retained (see Table 1) to assess the sensitivity of the  
 526 vertical mixing to the parameterization of the TKE injection and how it further impacts  
 527 the cross-shore circulation. The results are compared along vertical profiles of TKE ( $K$ ),  
 528 vertical eddy viscosity ( $\nu$ ) and cross-shore quasi-Eulerian velocity component ( $\hat{u}$ ) distributed  
 529 along the aforementioned cross-shore profile (see Fig. 7 and Table 2).

**Table 1.** Turbulence settings for the runs Rref, Rturb1, Rturb2 and Rturb3. The mixing length increases with  $\alpha_w$  while  $c_{ds}$  controls the TKE injection associated with wave energy dissipation through whitecapping.

		Mixing length	
		$\alpha_w = 1$	$\alpha_w = 0.5$
$F_K$	$c_{ds} = 1$	Rref	Rturb2
	$c_{ds} = 0$	Rturb1	Rturb3

530 Under storm conditions, whitecapping explains more than two-third of the TKE in-  
 531 jected at the surface in regions I and II. This greatly affects the mixing in the upper third of  
 532 the water column (see  $\nu$  almost doubled at some locations between Rref and Rturb1 in Fig.  
 533 7). Note, however, that it only slightly impacts the cross-shore circulation ( $RD(\hat{u}) \leq 7\%$ , see

534 the comparisons  $R_{\text{turb1}}/R_{\text{ref}}$  and  $R_{\text{turb3}}/R_{\text{turb2}}$  in Table 2), with the largest differences  
535 being found in the upper part of the water column. Furthermore, in both high or moderate  
536 wave energy conditions, the model shows a strong sensitivity to the parameterization of the  
537 surface mixing length, which strengthens as the injection of TKE at the surface increases  
538 closer to shore (within the zones II and III, see Table 2). With a shorter surface mixing  
539 length, the eddy viscosity at the surface is weaker, such that the cross-shore velocity profiles  
540 are more sheared (see Fig. 7 under high wave energy conditions, while similar results -  
541 not shown - are found under moderate wave energy conditions). As a result, the orienta-  
542 tion of the cross-shore flow near the surface even changes for one profile within the zone II  
543 under high energy condition with a shorter mixing length. The relative difference on the  
544 cross-shore velocity reaches 32% depending on the parameterization of  $z_0^s$ . Interestingly,  
545 comparing  $R_{\text{turb2}}/R_{\text{ref}}$  and  $R_{\text{turb3}}/R_{\text{turb1}}$  tends to show that the sensitivity to the pa-  
546 rameterization of the surface mixing length is stronger depending on the contribution of the  
547 whitecapping to the TKE injection (see Table 2). Overall, it appears that the parameteri-  
548 zation of the TKE injection impacts the vertical shear of the cross-shore velocity as far as  
549 6 km from the shoreline under energetic conditions.



**Figure 7.** Vertical profiles of turbulent kinetic energy ( $K$ , panel **a**), vertical eddy viscosity ( $\nu$ , panel **b**) and quasi-Eulerian cross-shore velocity ( $\hat{u}$ , panel **c**). The results are presented for the four configurations of the model (see Table 1) and are extracted under high wave energy conditions. The delimited zones are defined in section 5.1

**Table 2.** Relative difference (in %)  $RD_{1/2}(X) = \frac{|\hat{X}_1 - \hat{X}_2|}{|\hat{X}_2|}$  of modelled turbulent kinetic energy ( $K$ ), vertical eddy viscosity ( $\nu$ ) and cross-shore velocity ( $\hat{u}$ ) vertical profiles for the different configurations of the model within the three delimited area introduced in section 5.1, under high (HE) or moderate (LE) wave energy conditions. High RD values indicate a strong sensitivity to turbulence settings, either to the mixing length or the surface flux of TKE.

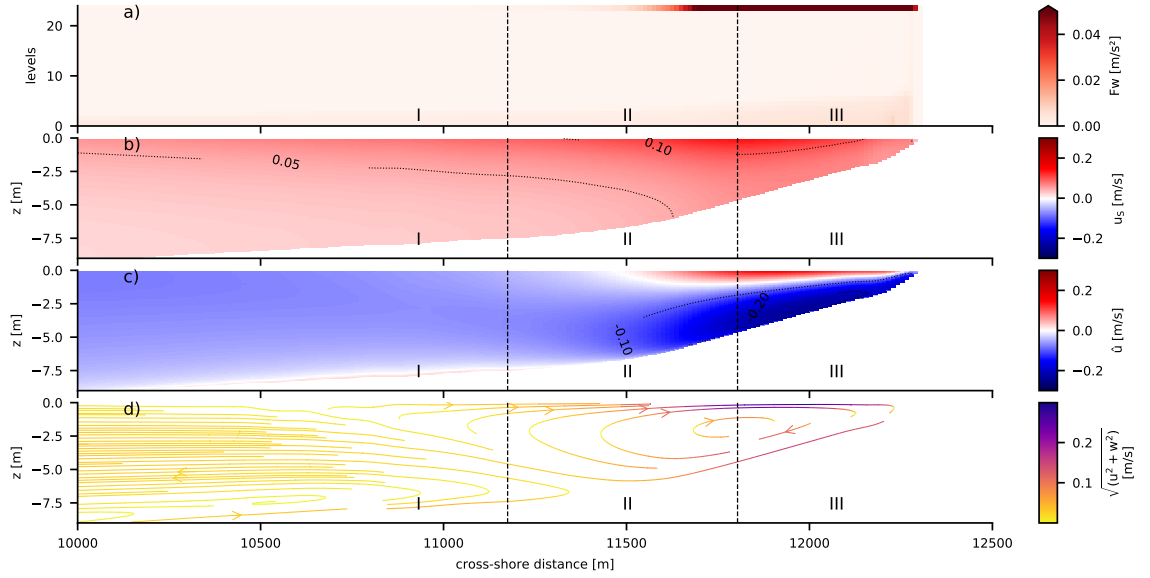
		Rturb1/Rref	Rturb3/Rturb2	Rturb2/Rref	Rturb3/Rturb1
<b>zone I</b>					
HE	RD(K)	92	88	30	5
	RD( $\nu$ )	62	49	42	8
	RD( $\hat{u}$ )	3	6	12	4
<b>zone II</b>					
	RD(K)	60	56	27	21
	RD( $\nu$ )	25	23	41	28
	RD( $\hat{u}$ )	7	4	28	19
<b>zone III</b>					
	RD(K)	2	2	16	16
	RD( $\nu$ )	<1	<1	29	29
	RD( $\hat{u}$ )	<1	<1	30	30
<b>zone I</b>					
LE	RD(K)	66	60	25	12
	RD( $\nu$ )	26	18	26	12
	RD( $\hat{u}$ )	2	2	10	7
<b>zone II</b>					
	RD(K)	30	28	24	21
	RD( $\nu$ )	10	9	29	25
	RD( $\hat{u}$ )	2	2	32	27
<b>zone III</b>					
	RD(K)	4	3	16	16
	RD( $\nu$ )	<1	<1	29	29
	RD( $\hat{u}$ )	<1	<1	30	29

### 5.3 Wave-induced circulation patterns and forcing mechanisms

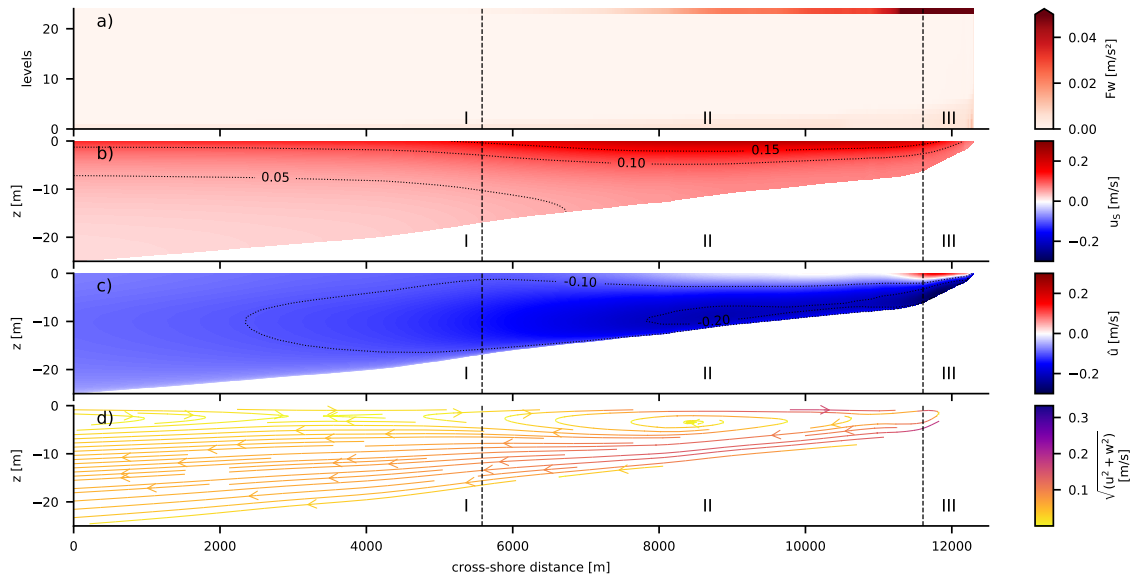
Under moderate wave energy conditions, non-conservative breaking wave forces arise relatively close to shore, up to 1 km from the shoreline (Fig. 8a) and are mostly associated with depth-induced breaking, which dominates the wave energy dissipation within the zones II and III (Fig. 6b). The quasi-Eulerian cross-shore flow therefore shows a strong seaward-directed current reaching 0.3 m/s in the lower part of the water column, whereas a shoreward-directed current arises near the surface (Fig. 8c). The depth-averaged quasi-Eulerian cross-shore flow (i.e. the undertow) nearly compensates for the depth-averaged cross-shore Stokes drift velocity component (the divergence of the alongshore velocity component is not exactly zero), while the depth-varying cross-shore flow is locally strengthened by the effect of the non-conservative breaking wave forces within the zones II and III, resulting in a 100% stronger seaward-directed current in the lower part of the water column than the surface Stokes drift velocity (Fig. 8b-c and 10b). In order to better evaluate the contribution of breaking, an additional run was performed with the non-conservative breaking wave forces uniformly distributed over the vertical ( $RF_{br}$ ). The resulting quasi-Eulerian cross-shore flow is less sheared near the surface within the zones II and III, yielding a weaker seaward-directed current in the lower part of the water column than that obtained with a near surface momentum source by up to a factor three (Fig. 10b). Within zone I, the quasi-Eulerian cross-shore flow is also mostly seaward-directed, with a maximum intensity in the upper part of the water column, while a weak shoreward-directed current associated with wave streaming arises near the bottom and extends inside the zone II (Fig. 8a,c and 10b). As a result, a strong clockwise Lagrangian overturning circulation develops within the zones II and III (the magnitude of the current locally reaches 0.3 m/s), while a weaker anti-clockwise overturning cell arises at the seaward edge of the zone II, which constrains the offshore flow (see the upward deflection of the streamlines of the clockwise cell within the zone II in Fig. 8d). A similar circulation pattern in the vicinity of the surf zone was found by Wang et al. (2020). The magnitude of the Lagrangian circulation then decreases relatively rapidly within zone I (Fig. 8d). Interestingly, the cross-shore quasi-Eulerian flow does not exactly compensate for the cross-shore Stokes drift velocity component, which would have been expected assuming a balance between the Coriolis force associated with the quasi-Eulerian flow and the Stokes-Coriolis force, as pointed out by Lentz et al. (2008). Presumably, this could be attributed to the shape of the coastline, which is not alongshore-uniform at the scale of several kilometers, the distance where the instruments were located.



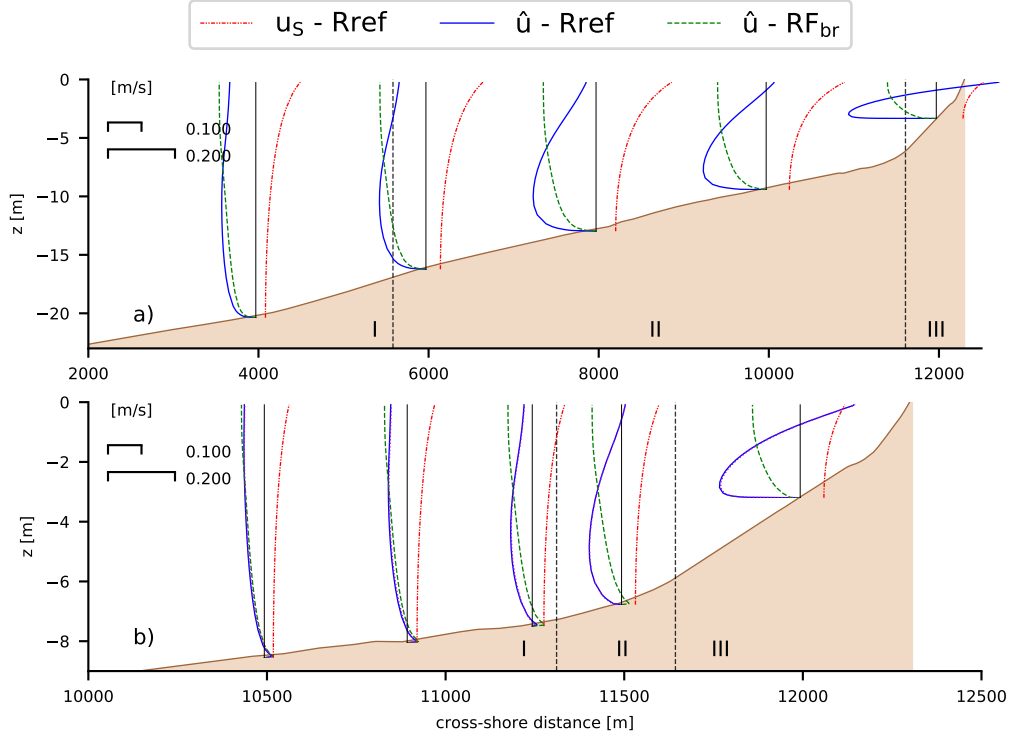
583 Under high wave energy conditions, non-conservative breaking wave forces arise at the  
584 surface as far as 6 km from the shoreline and strengthen shoreward (Fig. 9a), associated with  
585 a significant wave energy dissipation occurring through whitecapping and depth-induced  
586 breaking within the zones II and III (Fig. 6a). The quasi-Eulerian cross-shore flow shows a  
587 strong seaward-directed current of the order of 0.2 m/s up to 4 km from the shoreline, whose  
588 intensity progressively decreases offshore reaching 0.1 m/s as far as 10 km from the shoreline  
589 (Fig. 9c). Similar to moderate wave energy conditions, this current is strengthened locally  
590 by the effect of the non-conservative breaking wave forces within the zones II and III (Fig.  
591 10a). Interestingly, no near-bottom shoreward-directed current arises as the intensity of the  
592 bottom streaming decreases beyond 20 m-depth, while closer to shore the cross-shore flow  
593 in the lower part of the water column is dominated by the strong seaward-directed current.  
594 The Lagrangian circulation therefore shows a wide clockwise overturning cell extending over  
595 8 km, which generates a seaward-oriented jet in the lower part of the water column (Fig. 9d)  
596 that contrasts with the circulation pattern obtained under moderate wave energy conditions.



**Figure 8.** Cross-shore profiles under moderate wave energy conditions of the wave force cross-shore component, which includes the Stokes-Coriolis term, the vortex force, the wave-induced pressure term and the non-conservative wave forces (a), the Stokes drift velocity cross-shore component (b), the quasi-Eulerian velocity cross-shore component (c), and the 2DV Lagrangian circulation streamlines and magnitude (d). For readability, S-level indices are used for representing the wave force profile (the level 0 corresponds to the bottom and the level 24 to the free surface). Note that the cross-shore distance axis here extends over 2 km.



**Figure 9.** Same as Figure 8 under high wave energy conditions. Note that the cross-shore distance axis here extends over 12 km.



**Figure 10.** Vertical profiles of the Stokes drift velocity cross-shore component ( $u_S$ ) and the quasi-Eulerian cross-shore velocity ( $\hat{u}$ ) under high (a) or moderate (b) wave energy conditions for the baseline run (Rref) and for a run performed with the non-conservative breaking wave forces uniformly distributed (RF<sub>br</sub>). The delimited zones are defined in section 5.1.

## 597 6 Conclusions and perspectives

598 In this study, we presented a dataset comprising water levels, wave parameters and  
 599 currents collected under fair weather and storm conditions issued from a field campaign  
 600 carried out in early 2021 within the shoreface and the surf zone of the dissipative beach of  
 601 Saint-Trojan (France). These results were complemented with the predictions from a state-  
 602 of-the-art wave-averaged 3D circulation modelling system coupling the circulation model  
 603 SCHISM (Y. J. Zhang et al., 2016) with the spectral wave model WWM (Roland et al.,  
 604 2012) in order to study the wave-induced cross-shore circulation in the nearshore area up  
 605 to the surf zone. Both measurements and model results clearly showed that the cross-  
 606 shore flow is dominated by a strong seaward-directed current in the lower part of the water  
 607 column, which under storm wave conditions can reach 0.2 m/s as far as 4 km from the  
 608 shoreline. The model was further employed to study the driving mechanisms of the wave-

609 induced cross-shore circulation. It was shown that a wide ( $\simeq 6$  km) transition zone appears  
 610 under high wave energy conditions where depth-induced breaking, whitecapping and bottom  
 611 friction all contribute significantly to the wave energy dissipation, whereas the surf zone  
 612 edge appears clearly delimited under moderate wave energy conditions. Once wave energy  
 613 dissipation through breaking processes starts being substantial, sensitivity tests performed  
 614 with the model tended to show that the parameterization of the wave-enhanced mixing could  
 615 substantially impact the vertical shear of the cross-shore velocity close to the surface, while  
 616 non-conservative breaking wave forces strengthen locally the seaward-directed current in the  
 617 lower part of the water column. The wave-induced cross-shore circulation thus presents very  
 618 contrasting patterns. Under moderate wave energy conditions, the dominance of the wave  
 619 energy dissipation by bottom friction seaward of the surf zone allows the development of a  
 620 near-bottom onshore flow associated with bottom streaming of the order of a few centimetres  
 621 per second. Thus, the surf zone clockwise Lagrangian overturning circulation weakens, while  
 622 an additional anti-clockwise overturning cell arises seaward of the surf zone and extends  
 623 within it. In contrast, under high energy conditions the Lagrangian circulation shows a  
 624 seaward-directed jet in the lower part of the water column, whose intensity progressively  
 625 decreases offshore. Among different implications, this study opens perspectives for sediment  
 626 transport modelling using a similar 3D framework in order to produce realistic morphological  
 627 evolutions across the shoreface.

## 628 Code and Data Availability Statement

629 The processed field data presented in this paper are available through a Zenodo repos-  
 630 itory (Pezerat et al., 2022). The instructions to download and install the model used in  
 631 this study can be accessed freely at <https://github.com/schism-dev/schism>. Figures were  
 632 made with python module Matplotlib version 3.1.1, available under the Matplotlib licenses  
 633 at <https://matplotlib.org/>.

## 634 Appendix A Vortex force framework in the modelling system SCHISM

635 In the vortex force framework the continuity and momentum equations read:

$$636 \quad \frac{\partial \hat{u}}{\partial x} + \frac{\partial \hat{v}}{\partial y} + \frac{\partial \hat{w}}{\partial z} = 0 \quad (\text{A1})$$

$$637 \quad \frac{D\hat{u}}{Dt} - f\hat{v} = -\frac{1}{\rho} \frac{\partial P_H}{\partial x} + \frac{\partial}{\partial z} \left( \nu \frac{\partial \hat{u}}{\partial z} \right) + v_S \left( f_C + \left( \frac{\partial \hat{v}}{\partial x} - \frac{\partial \hat{u}}{\partial y} \right) \right) - w_S \frac{\partial \hat{u}}{\partial z} - \frac{\partial J}{\partial x} + \hat{F}_x \quad (\text{A2})$$

$$639 \quad \frac{D\hat{v}}{Dt} + f\hat{u} = -\frac{1}{\rho} \frac{\partial P_H}{\partial y} + \frac{\partial}{\partial z} \left( \nu \frac{\partial \hat{v}}{\partial z} \right) - u_S \left( f_C + \left( \frac{\partial \hat{v}}{\partial x} - \frac{\partial \hat{u}}{\partial y} \right) \right) - w_S \frac{\partial \hat{v}}{\partial z} - \frac{\partial J}{\partial y} + \hat{F}_y \quad (\text{A3})$$

641 where  $[\hat{u}, \hat{v}, \hat{w}]$  is the quasi-Eulerian velocity, equal to the mean Lagrangian velocity  $[u, v, w]$   
 642 minus the Stokes drift velocity  $[u_S, v_S, w_S]$ . In Eqs A2 and A3,  $f_C$  is the Coriolis parameter,  
 643  $\rho$  is the water density,  $g$  is the gravitational acceleration,  $\nu$  is the vertical eddy viscosity,  
 644  $P_H$  is the hydrostatic pressure,  $J$  is the wave-induced mean pressure and  $[\hat{F}_x, \hat{F}_y]$  gathers  
 645 the non-conservative wave forces.

646 The three components of the Stokes drift velocities, the wave-induced pressure term  
 647 and the non-conservative wave forces are all computed from local variables issued from  
 648 WWM, which simulates the generation, propagation and transformation of short waves by  
 649 solving the Wave Action Equation (e.g. Komen et al., 1994):

$$650 \quad \frac{\partial N}{\partial t} + \frac{\partial}{\partial x} ((C_{g,x} + \tilde{u})N) + \frac{\partial}{\partial y} ((C_{g,y} + \tilde{v})N) + \frac{\partial(\dot{\sigma}N)}{\partial \sigma} + \frac{\partial(\dot{\theta}N)}{\partial \theta} = S_{tot} \quad (\text{A4})$$

651 where  $\sigma$  is the relative wave frequency, which is related to the wavenumber  $k$  by the linear  
 652 dispersion relation;  $\theta$  corresponds to the wave direction;  $N = \rho g E / \sigma$  is the wave action  
 653 density spectrum, with  $E$  the sea surface elevation density spectrum;  $[C_{g,x}, C_{g,y}]$  is the  
 654 intrinsic group velocity;  $[\tilde{u}, \tilde{v}]$  is the advective current velocity, here equal to the depth-  
 655 averaged horizontal quasi-Eulerian velocity. Finally,  $S_{tot}$  is a sum of source terms that  
 656 account for the energy input due to wind, non linear wave-wave interactions, and energy  
 657 dissipation due to whitecapping, depth-induced breaking and bottom friction (cf. section  
 658 3.1).

659 The horizontal Stokes drift velocity vector is given by:

$$660 \quad [u_S, v_S] = \int_{\sigma} \int_{\theta} \sigma' k E \frac{\cosh(2k(z+d))}{\sinh^2(kh)} [\cos \theta', \sin \theta'] d\sigma' d\theta' \quad (\text{A5})$$

661 where  $h = d + \bar{\eta}$  is the (local) phase-averaged water depth (with  $d$ , the still water depth and  
 662  $\bar{\eta}$ , the phase-averaged free surface elevation). The vertical Stokes drift component is given  
 663 by the horizontal divergence of  $[u_S, v_S]$  as the full Stokes drift flow is non divergent at the  
 664 lowest order (Ardhuin et al., 2008). The wave-induced pressure term reads:

$$665 \quad J = \int_{\sigma} \int_{\theta} \frac{gkE}{\sinh(2kh)} d\sigma' d\theta' \quad (\text{A6})$$

666 Finally, the formulation of the non-conservative wave forces are detailed in section 3.2.1.

667 The wave model is supplemented by a roller model that helps representing the inertia of  
 668 depth-induced breaking processes by slightly advecting the location where energy is actually  
 669 dissipated towards the shoreline (Svendsen, 1984b). As compared to the implementation  
 670 detailed in Gu erin et al. (2018), the roller model solves a balance equation for the roller

671 kinetic energy  $E_R$  (e.g Reniers et al., 2004), slightly modified to account for the modification  
 672 of the wave phase velocity by the mean current within the advection term, such that:

$$673 \quad \frac{\partial E_R}{\partial t} + \frac{\partial}{\partial x} (2E_R(c_x + \bar{u})) + \frac{\partial}{\partial y} (2E_R(c_y + \bar{v})) = \alpha_R D_{db} - D_R \quad (\text{A7})$$

674 where  $[c_x, c_y]$  is the wave phase velocity computed by means of the short wave (continuous)  
 675 peak wavenumber ( $k_p$ ) and mean direction ( $\theta_m$ );  $D_{db}$  is the bulk wave energy dissipation  
 676 rate by depth-induced breaking;  $\alpha_R$  is the percentage of wave energy dissipation by depth-  
 677 induced breaking transferred to the rollers (Tajima & Madsen, 2006) and  $D_R$  is the roller  
 678 energy dissipation rate, which reads:

$$679 \quad D_R = \frac{2g \sin \beta_R E_R}{\sqrt{c_x^2 + c_y^2}} \quad (\text{A8})$$

680 where  $\sin \beta_R = 0.1$  is the roller angle (Nairn et al., 1991; Reniers et al., 2004). Surface rollers  
 681 contribute to the total mass flux in proportion to the roller energy. Although this transport  
 682 primarily occurs near the surface, above trough level, there is no consensus on its vertical  
 683 distribution. We here follow the choice to impose an homogeneous vertical distribution.  
 684 This contribution is accounted for through an additional term in the horizontal Stokes drift  
 685 velocity vector, which reads:

$$686 \quad [u_{S,R}, v_{S,R}] = \frac{2E_R}{\rho h \sqrt{c_x^2 + c_y^2}} [\cos \theta_m, \sin \theta_m] \quad \forall z \in [-d, \bar{\eta}] \quad (\text{A9})$$

687 Note, however, that  $w_S$  is still computed with  $u_S, v_S$  from Eq. A5 as it is assumed that  
 688 rollers do not contribute to the vertical transport.

## 689 Acknowledgments

690 M. Pezerat is supported by a PhD fellowship from CDA La Rochelle and from the FEDER  
 691 project DURALIT. K. Martins greatly acknowledges the financial support from the Uni-  
 692 versity of Bordeaux, through an International Postdoctoral Grant (Idex, nb. 1024R-5030).  
 693 L. Lavaud is supported by a PhD fellowship from the Region Nouvelle-Aquitaine and the  
 694 UNIMA engineering consulting company. The authors appreciate the administrative sup-  
 695 port of the DDTM to carry out long term deployment of sensors. The topo-bathymetric  
 696 data of the studied area were acquired in the scope of the National Observation System  
 697 DYNALIT (<https://www.dynalit.fr>), part of the research infrastructure ILICO. The Ob-  
 698 servatoire de la Cte Aquitaine (OCA) partly funded the field campaign presented in this  
 699 study. N. Lachausse built the structures to anchor the sensors, whereas T. Coulombier and  
 700 D. Dausse provided valuable assistance during the field campaign. Lastly, the authors want

701 to thank O. de Viron for his enlightening insights on data post-processing, and the two  
 702 anonymous reviewers for their constructive comments.

## 703 References

- 704 Abdolali, A., Roland, A., Van Der Westhuysen, A., Meixner, J., Chawla, A., Hesser, T. J.,  
 705 ... Sikiric, M. D. (2020). Large-scale hurricane modeling using domain decomposition  
 706 parallelization and implicit scheme implemented in wavewatch iii wave model. *Coastal*  
 707 *Engineering*, *157*, 103656.
- 708 Appell, G. F., Bass, P., & Metcalf, M. A. (1991). Acoustic doppler current profiler perfor-  
 709 mance in near surface and bottom boundaries. *IEEE journal of oceanic engineering*,  
 710 *16*(4), 390–396.
- 711 Ardhuin, F., O’Reilly, W. C., Herbers, T. H. C., & Jessen, P. F. (2003). Swell transformation  
 712 across the continental shelf. part 1: Attenuation and directional broadening. *Journal*  
 713 *of Physical Oceanography*, *33*, 1921–1939.
- 714 Ardhuin, F., Rasche, N., & Belibassakis, K. A. (2008). Explicit wave-averaged primitive  
 715 equations using a generalized lagrangian mean. *Ocean Modelling*, *20*(1), 35–60.
- 716 Ardhuin, F., Rogers, E., Babanin, A. V., Filipot, J-F., Magne, R., Roland, A., ... others  
 717 (2010). Semiempirical dissipation source functions for ocean waves. part i: Definition,  
 718 calibration, and validation. *Journal of Physical Oceanography*, *40*(9), 1917–1941.
- 719 Battjes, J. A., & Janssen, J. (1978). Energy loss and set-up due to breaking of random  
 720 waves. In *Coastal engineering 1978* (pp. 569–587).
- 721 Battjes, J. A., & Stive, M. (1985). Calibration and verification of a dissipation model for  
 722 random breaking waves. *Journal of Geophysical Research: Oceans*, *90*(C5), 9159–  
 723 9167.
- 724 Bennis, A.-C., Ardhuin, F., & Dumas, F. (2011). On the coupling of wave and three-  
 725 dimensional circulation models: Choice of theoretical framework, practical implemen-  
 726 tation and adiabatic tests. *Ocean Modelling*, *40*(3-4), 260–272.
- 727 Bertin, X., Bruneau, N., Breilh, J-F., Fortunato, A. B., & Karpytchev, M. (2012). Im-  
 728 portance of wave age and resonance in storm surges: The case xynthia, bay of biscay.  
 729 *Ocean Modelling*, *42*, 16–30.
- 730 Bertin, X., Castelle, B., Chaumillon, E., Butel, R., & Quique, R. (2008). Longshore  
 731 transport estimation and inter-annual variability at a high-energy dissipative beach:  
 732 St. trojan beach, sw oléron island, france. *Continental Shelf Research*, *28*(10-11),



- 733 1316–1332.
- 734 Bertin, X., Li, K., Roland, A., & Bidlot, J.-R. (2015). The contribution of short-waves in  
 735 storm surges: Two case studies in the bay of biscay. *Continental Shelf Research*, *96*,  
 736 1–15.
- 737 Bertin, X., Martins, K., de Bakker, A., Chataigner, T., Guérin, T., Coulombier, T., & de  
 738 Viron, O. (2020). Energy transfers and reflection of infragravity waves at a dissipa-  
 739 tive beach under storm waves. *Journal of Geophysical Research: Oceans*, *125*(5),  
 740 e2019JC015714.
- 741 Bishop, C. T., & Donelan, M. A. (1987). Measuring waves with pressure transducers.  
 742 *Coastal Engineering*, *11*(4), 309–328.
- 743 Blumberg, A. F., & Mellor, G. L. (1987). A description of a three-dimensional coastal ocean  
 744 circulation model. *Three-dimensional coastal ocean models*, *4*, 1–16.
- 745 Booij, N., Ris, R. C., & Holthuijsen, L. H. (1999). A third-generation wave model for coastal  
 746 regions: 1. model description and validation. *Journal of geophysical research: Oceans*,  
 747 *104*(C4), 7649–7666.
- 748 Burchard, H., Bolding, K., & Villarreal, M. (1999). *Gotm, a general ocean turbulence*  
 749 *model: Theory, implementation and test cases* (Tech. Rep. No. EUR18745). European  
 750 Commission.
- 751 Castelle, B., Marieu, V., Bujan, S., Splinter, K. D., Robinet, A., Sénéchal, N., & Ferreira,  
 752 S. (2015). Impact of the winter 2013–2014 series of severe western europe storms  
 753 on a double-barred sandy coast: Beach and dune erosion and megacusp embayments.  
 754 *Geomorphology*, *238*, 135–148.
- 755 Coco, G., Senechal, N., Rejas, A., Bryan, K. R., Capo, S., Parisot, J., ... MacMahan,  
 756 J. H. (2014). Beach response to a sequence of extreme storms. *Geomorphology*, *204*,  
 757 493–501.
- 758 Craig, P. D., & Banner, M. L. (1994). Modeling wave-enhanced turbulence in the ocean  
 759 surface layer. *Journal of Physical Oceanography*, *24*(12), 2546–2559.
- 760 De Bakker, A., Herbers, T., Smit, P., Tissier, M., & Ruessink, B. (2015). Nonlinear  
 761 infragravity–wave interactions on a gently sloping laboratory beach. *Journal of Phys-  
 762 ical Oceanography*, *45*(2), 589–605.
- 763 Deigaard, R., Justesen, P., & Fredsøe, J. (1991). Modelling of undertow by a one-equation  
 764 turbulence model. *Coastal Engineering*, *15*(5-6), 431–458.
- 765 Delpey, M., Ardhuin, F., Otheguy, P., & Jouon, A. (2014). Effects of waves on coastal water

- 766 dispersion in a small estuarine bay. *Journal of Geophysical Research: Oceans*, 119(1),  
767 70–86.
- 768 Déqué, M., Dreveton, C., Braun, A., & Cariolle, D. (1994). The arpege/ifs atmosphere  
769 model: a contribution to the french community climate modelling. *Climate Dynamics*,  
770 10(4), 249–266.
- 771 Dodet, G., Bertin, X., Bouchette, F., Gravelle, M., Testut, L., & Wöppelmann, G. (2019).  
772 Characterization of sea-level variations along the metropolitan coasts of france: waves,  
773 tides, storm surges and long-term changes. *Journal of Coastal Research*, 88(sp1), 10–  
774 24.
- 775 Eldeberky, Y. (1997). Nonlinear transformation of wave spectra in the nearshore zone.  
776 *Oceanographic Literature Review*, 4(44), 297.
- 777 Eldeberky, Y., & Battjes, J. A. (1996). Spectral modeling of wave breaking: Application  
778 to boussinesq equations. *Journal of Geophysical Research: Oceans*, 101(C1), 1253–  
779 1264.
- 780 Elgar, S., & Raubenheimer, B. (2020). Field evidence of inverse energy cascades in the  
781 surfzone. *Journal of Physical Oceanography*, 50(8), 2315–2321.
- 782 Feddersen, F. (2012a). Observations of the surf-zone turbulent dissipation rate. *Journal of*  
783 *Physical Oceanography*, 42(3), 386–399.
- 784 Feddersen, F. (2012b). Scaling surf zone turbulence. *Geophysical Research Letters*, 39(18).
- 785 Feddersen, F. (2014). The generation of surfzone eddies in a strong alongshore current.  
786 *Journal of Physical Oceanography*, 44(2), 600–617.
- 787 Feddersen, F., & Trowbridge, J. (2005). The effect of wave breaking on surf-zone turbulence  
788 and alongshore currents: A modeling study. *Journal of Physical Oceanography*, 35(11),  
789 2187–2203.
- 790 Garcez Faria, A., Thornton, E., Lippmann, T., & Stanton, T. (2000). Undertow over a  
791 barred beach. *Journal of Geophysical Research: Oceans*, 105(C7), 16999–17010.
- 792 Goring, D. G., & Nikora, V. I. (2002). Despiking acoustic doppler velocimeter data. *Journal*  
793 *of hydraulic engineering*, 128(1), 117–126.
- 794 Grant, W. D., & Madsen, O. S. (1982). Movable bed roughness in unsteady oscillatory flow.  
795 *Journal of Geophysical Research*, 87, 469–481.
- 796 Guérin, T., Bertin, X., Coulombier, T., & de Bakker, A. (2018). Impacts of wave-induced  
797 circulation in the surf zone on wave setup. *Ocean Modelling*, 123, 86–97.
- 798 Haines, J. W., & Sallenger Jr, A. H. (1994). Vertical structure of mean cross-shore currents

- 799 across a barred surf zone. *Journal of Geophysical Research: Oceans*, 99(C7), 14223–  
800 14242.
- 801 Hamm, L., & Peronnard, C. (1997). Wave parameters in the nearshore: A clarification.  
802 *Coastal Engineering*, 32(2-3), 119–135.
- 803 Hasselmann, K. (1970). Wave-driven inertial oscillations. *Geophysical and Astrophysical*  
804 *Fluid Dynamics*, 1(3-4), 463–502.
- 805 Hasselmann, S., Hasselmann, K., Allender, J., & Barnett, T. (1985). Computations and  
806 parameterizations of the nonlinear energy transfer in a gravity-wave spectrum. part  
807 ii: Parameterizations of the nonlinear energy transfer for application in wave models.  
808 *Journal of Physical Oceanography*, 15(11), 1378–1391.
- 809 Huang, Z.-C., Hsiao, S.-C., Hwung, H.-H., & Chang, K.-A. (2009). Turbulence and energy  
810 dissipations of surf-zone spilling breakers. *Coastal Engineering*, 56(7), 733–746.
- 811 Klopman, G. (1994). *Vertical structure of the flow due to waves and currents – laser-*  
812 *doppler flow measurements for waves following or opposing a current* (Tech. Rep. No.  
813 H840-30). Deltares (WL).
- 814 Komen, G. J., Cavaleri, L., Donelan, M., Hasselmann, K., Hasselmann, S., & Janssen,  
815 P. A. E. M. (1994). *Dynamics and modelling of ocean waves*. Cambridge University  
816 Press, Cambridge, U.K. doi: 10.1017/CBO9780511628955
- 817 Kumar, N., Voulgaris, G., Warner, J. C., & Olabarrieta, M. (2012). Implementation of  
818 the vortex force formalism in the coupled ocean-atmosphere-wave-sediment transport  
819 (coawst) modeling system for inner shelf and surf zone applications. *Ocean Modelling*,  
820 47, 65–95.
- 821 Lavaud, L., Pezerat, M., Coulombier, T., Bertin, X., & Martins, K. (2020). Hydrodynamics  
822 on a rocky shore under moderate-energy wave conditions. *Journal of Coastal Research*,  
823 95(SI), 1473–1479.
- 824 Lentz, S. J., Fewings, M., Howd, P., Fredericks, J., & Hathaway, K. (2008). Observa-  
825 tions and a model of undertow over the inner continental shelf. *Journal of Physical*  
826 *Oceanography*, 38(11), 2341–2357.
- 827 Long, J. W., & Özkan-Haller, H. T. (2009). Low-frequency characteristics of wave group-  
828 forced vortices. *Journal of Geophysical Research: Oceans*, 114(C8).
- 829 Longuet-Higgins, M. (1953). Mass transport in water waves. *Philosophical Transactions of*  
830 *the Royal Society*, 245, 535–581.
- 831 Madsen, O. S. (1995). Spectral wave-current bottom boundary layer flows. In *Coastal*

- 832           *engineering 1994* (pp. 384–398).
- 833 Madsen, O. S., Poon, Y. K., & Graber, H. C. (1988). Spectral wave attenuation by bottom  
834           friction: Theory. In *21st international conference on coastal engineering* (pp. 492–504).  
835           Malaga: ASCE.
- 836 Mathisen, P. P., & Madsen, O. S. (1999). Waves and currents over a fixed rippled bed: 3.  
837           bottom and apparent roughness for spectral waves and currents. *Journal of Geophys-*  
838           *ical Research: Oceans*, *104*(C8), 18447–18461.
- 839 Michaud, H., Marsaleix, P., Leredde, Y., Estournel, C., Bourrin, F., Lyard, F., . . . Ardhuin,  
840           F. (2012). Three-dimensional modelling of wave-induced current from the surf zone  
841           to the inner shelf. *Ocean Science*, *8*(4), 657–681.
- 842 Moghimi, S., Thomson, J., Özkan-Haller, T., Umlauf, L., & Zippel, S. (2016). On the  
843           modeling of wave-enhanced turbulence nearshore. *Ocean Modelling*, *103*, 118–132.
- 844 Morgan, S. G., Shanks, A. L., MacMahan, J. H., Reniers, A. J., & Feddersen, F. (2018).  
845           Planktonic subsidies to surf-zone and intertidal communities. *Annual review of marine*  
846           *science*, *10*, 345–369.
- 847 Mouragues, A., Bonneton, P., Lannes, D., Castelle, B., & Marieu, V. (2019). Field data-  
848           based evaluation of methods for recovering surface wave elevation from pressure mea-  
849           surements. *Coastal Engineering*, *150*, 147–159.
- 850 Nairn, R. B., Roelvink, J., & Southgate, H. N. (1991). Transition zone width and im-  
851           plications for modelling surfzone hydrodynamics. In *Coastal engineering 1990* (pp.  
852           68–81).
- 853 Nicolae-Lerma, A., Bulteau, T., Lecacheux, S., & Idier, D. (2015). Spatial variability of  
854           extreme wave height along the atlantic and channel french coast. *Ocean Engineering*,  
855           *97*, 175–185.
- 856 Noyes, T. J., Guza, R., Elgar, S., & Herbers, T. (2004). Field observations of shear waves  
857           in the surf zone. *Journal of Geophysical Research: Oceans*, *109*(C1).
- 858 Oltman-Shay, J., Howd, P., & Birkemeier, W. (1989). Shear instabilities of the mean  
859           longshore current: 2. field observations. *Journal of Geophysical Research: Oceans*,  
860           *94*(C12), 18031–18042.
- 861 Paskyabi, M. B., Fer, I., & Jenkins, A. D. (2012). Surface gravity wave effects on the  
862           upper ocean boundary layer: modification of a one-dimensional vertical mixing model.  
863           *Continental Shelf Research*, *38*, 63–78.
- 864 Pezerat, M., Bertin, X., Martins, K., & Lavaud, L. (2022). *Cross-shore distribution of*

- 865        *the wave-induced circulation over a dissipative beach under storm wave conditions:*  
866        *the dataset.* <https://doi.org/10.5281/zenodo.5878857>. Zenodo. doi: 10.5281/  
867        zenodo.5878857
- 868 Pezerat, M., Bertin, X., Martins, K., Mengual, B., & Hamm, L. (2021). Simulating storm  
869        waves in the nearshore area using spectral model: Current issues and a pragmatic  
870        solution. *Ocean Modelling*, *158*, 101737.
- 871 Rasclé, N. (2007). *Impact of waves on the ocean circulation* (Unpublished doctoral disserta-  
872        tion). Universit de Bretagne Occidentale.
- 873 Reniers, A. J., Roelvink, J., & Thornton, E. (2004). Morphodynamic modeling of an  
874        embayed beach under wave group forcing. *Journal of Geophysical Research: Oceans*,  
875        *109*(C1).
- 876 Roland, A., Zhang, Y. J., Wang, H. V., Meng, Y., Teng, Y.-C., Maderich, V., . . . Zanke,  
877        U. (2012). A fully coupled 3d wave-current interaction model on unstructured grids.  
878        *Journal of Geophysical Research: Oceans*, *117*(C11).
- 879 Saha, S., Moorthi, S., Wu, X., Wang, J., Nadiga, S., Tripp, P., . . . Becker, E. (2011). *Ncep*  
880        *climate forecast system version 2 (cfsv2) selected hourly time-series products*. Boulder  
881        CO: Research Data Archive at the National Center for Atmospheric Research, Com-  
882        putational and Information Systems Laboratory. Retrieved from 10.5065/D6N877VB
- 883 Smith, J. A. (2006). Wave–current interactions in finite depth. *Journal of Physical Oceanog-*  
884        *raphy*, *36*(7), 1403–1419.
- 885 Stive, M., & Wind, H. (1986). Cross-shore mean flow in the surf zone. *Coastal Engineering*,  
886        *10*(4), 325–340.
- 887 Svendsen, I. A. (1984a, February). Mass flux and undertow in a surf zone. *Coastal engi-*  
888        *neering*, *8*(4), 347–365.
- 889 Svendsen, I. A. (1984b, January). Wave heights and set-up in a surf zone. *Coastal engi-*  
890        *neering*, *8*(4), 303–329.
- 891 Tajima, Y., & Madsen, O. S. (2006). Modeling near-shore waves, surface rollers, and  
892        undertow velocity profiles. *Journal of waterway, port, coastal, and ocean engineering*,  
893        *132*(6), 429–438.
- 894 Tolman, H. L. (1991). A third-generation model for wind waves on slowly varying, unsteady,  
895        and inhomogeneous depths and currents. *Journal of Physical Oceanography*, *21*(6),  
896        782–797.
- 897 Tolman, H. L. (1994). Wind waves and movable-bed bottom friction. *Journal of Physical*

- 898 *Oceanography*, 24, 994–1009.
- 899 Uchiyama, Y., McWilliams, J. C., & Shchepetkin, A. F. (2010). Wave–current interaction  
900 in an oceanic circulation model with a vortex-force formalism: Application to the surf  
901 zone. *Ocean Modelling*, 34(1-2), 16–35.
- 902 Umlauf, L., & Burchard, H. (2003). A generic length-scale equation for geophysical turbu-  
903 lence models. *Journal of Marine Research*, 61(2), 235–265.
- 904 Umlauf, L., Burchard, H., & Bolding, K. (2005). *The general ocean turbulence model (gotm)*  
905 *scientific documentation:version 3.2* (Tech. Rep. No. 63). Warnemunde, Germany:  
906 Leibnitz Institute for Baltic Sea Research.
- 907 Wang, P., McWilliams, J. C., Uchiyama, Y., Chekroun, M. D., & Yi, D. L. (2020). Effects  
908 of wave streaming and wave variations on nearshore wave-driven circulation. *Journal*  
909 *of Physical Oceanography*, 50(10), 3025–3041.
- 910 Wright, L. D., & Short, A. D. (1984). Morphodynamic variability of surf zones and beaches:  
911 a synthesis. *Marine geology*, 56(1-4), 93–118.
- 912 Xu, Z., & Bowen, A. (1994). Wave-and wind-driven flow in water of finite depth. *Journal*  
913 *of physical oceanography*, 24(9), 1850–1866.
- 914 Zhang, H., Madsen, O. S., Sannasiraj, S., & Chan, E. S. (2004). Hydrodynamic model with  
915 wave–current interaction in coastal regions. *Estuarine, Coastal and Shelf Science*,  
916 61(2), 317–324.
- 917 Zhang, Y. J., Ye, F., Stanev, E. V., & Grashorn, S. (2016). Seamless cross-scale modeling  
918 with schism. *Ocean Modelling*, 102, 64–81.
- 919 Zheng, P., Li, M., van der Zanden, J., Wolf, J., Chen, X., Wang, C., et al. (2017). A 3d  
920 unstructured grid nearshore hydrodynamic model based on the vortex force formalism.  
921 *Ocean Modelling*, 116, 48–69.