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Energy transfers and reflection of infragravity waves at 1 a dissipative beach under storm waves 2

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Key Points:

in beach slope

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11	•	Very large IG waves are observed at a dissipative beach under a storm, long-period
12		swell and are driven by the bound wave mechanism
13	•	In the surf zone, IG wave energy is transferred not only towards superharmonic but
14		also subharmonic frequencies
15	•	IG wave reflection is tidally-modulated and almost full at high tide due to the increase

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17 Abstract

This study presents unpublished field observations of infragravity waves, collected at the dis-18 sipative beach of Saint-Trojan (Oléron Island, France) during the storm Kurt (03/02/2017), 19 characterized by incident short waves of significant heights reaching 9.5 m and peak peri-20 ods reaching 22 s. Data analysis reveals the development of exceptionally large infragravity 21 waves, with significant heights reaching 1.85 m close to shore. Field observations are com-22 plemented by numerical modelling with XBeach, which well reproduces the development 23 of such infragravity waves. Model results reveals that infragravity waves were generated 24 mainly through the bound wave mechanism, enhanced by the development of a phase lag 25 with the short wave energy envelope. Spectral analysis of the free surface elevation shows the 26 generation of superharmonic and subharmonic infragravity waves, the latter dominating the 27 free surface elevation variance close to shore. Modelling results suggest that subharmonic 28 infragravity waves result, at least partly, from infragravity-wave merging, promoted by the 29 combination of free and bound infragravity waves propagating across a several kilometer-30 wide surf zone. Due to the steeper slope of the upper part of the beach profile, observed and 31 modeled reflection coefficients under moderate-energy show a strong tidal modulation, with 32 a weak reflection at low tide $(R^2 < 0.2)$ and a full reflection at high tide $(R^2 \sim 1.0)$. Un-33 der storm waves, the observed reflection coefficients remain unusually high for a dissipative 34 beach $(R^2 \sim 0.5 - 1.0)$, which is explained by the development of subharmonic infragravity 35 waves with frequencies around 0.005 Hz, too long to suffer a substantial dissipation. 36

37 1 Introduction

Infragravity (hereafter IG) waves are surface gravity waves with frequencies typically 38 ranging from 0.004 to 0.04 Hz, related to the presence of groups in incident short waves see 39 Bertin et al., 2018, for a recent review]. IG waves are only a few mm to cm-high in the deep 40 ocean [e.g. Rawat et al., 2014; Crawford et al., 2015; Smit et al., 2018] but can exceed 1.0 m 41 close to shore under storm conditions [Ruessink, 2010; Fiedler et al., 2015; Inch et al., 2017]. 42 Although not validated against field observations, the application of XBeach [Roelvink et al., 43 2009] to the SW of Oléron Island (France) by Baumann et al. [2017] even suggests that IG 44 waves could exceed 2.0 m close to shore under very energetic, narrow-banded incident swells. 45 Therefore, IG waves have an essential contribution to sandy beach [e.g. Guza and Thornton, 46 1982; Elgar et al., 1992; Reniers et al., 2002; Ruessink et al., 1998a; Guedes et al., 2013] and 47 tidal inlet hydrodynamics [Bertin and Olabarrieta, 2016; Bertin et al., 2019; Mendes et al., 48 2020], sediment transport [e.g. Russell, 1993; Aagaard and Greenwood, 2008; De Bakker 49 et al., 2016a] and dune and barrier breaching [e.g. Roelvink et al., 2009], particularly under 50 storm waves. 51

At dissipative beaches, IG waves usually dominate the spectrum of the free surface 52 elevation close to shore [Guza and Thornton, 1982; Russell, 1993; Ruessink et al., 1998b] and 53 therefore control the development of large runup [Raubenheimer and Guza, 1996; Ruessink 54 et al., 1998b; Ruggiero et al., 2001 and overwash along the coast and subsequent washovers 55 in low-lying zones [e.g. Baumann et al., 2017]. At such gently-sloping beaches, IG waves 56 are mostly generated through the bound wave mechanism, which corresponds to second-57 order nonlinear wave-wave interactions that result in the development of long waves, out of 58 phase with the energy envelope of the short waves. This mechanism was first demonstrated 59 analytically in 1D by Biesel [1952] and Longuet-Higgins and Stewart [1962] and was then 60 extended to 2D random waves by *Hasselmann* [1962]. These analytical solutions imply 61 that the frequency of the resulting IG waves should decrease under narrow-banded and 62 long-period incident short waves. Yet, Bertin and Olabarrieta [2016] observed relatively 63 short-period IG waves (e.g. T < 60 s) under such incident short waves while other authors 64 reported IG wave periods in the range 120 - 300 s under shorter period incident waves [e.g. 65 De Bakker et al., 2014; Inch et al., 2017]. This apparent inconsistency suggests that the 66 relationship between the shape of the incident short wave spectra and the frequency of IG 67 waves is not straightforward. Also, these analytical solutions were derived considering a flat 68

bottom while for sloping bottoms, the bound waves increasingly lag behind the wave groups 69 as and when the water depth decreases. This phase lag was observed in both field [e.g. List, 70 1992; Masselink, 1995; Inch et al., 2017] and laboratory experiments [Battjes et al., 2004; 71 De Bakker et al., 2013; Padilla and Alsina, 2017] and was demonstrated semi-analytically 72 [Janssen et al., 2003; Guérin et al., 2019]. Van Dongeren et al. [2007] showed that this phase 73 lag enhances energy transfer from short waves to IG waves. This process is often referred to 74 as "bound wave shoaling", which is somehow misleading because the resulting growth of IG 75 waves is higher than the growth associated with conservative shoaling (i.e. Green's Law). 76

77 In the surf zone, depth-induced breaking of short waves results in a shoreward reduction of short wave groupiness, which allows for the release of IG waves that then propagate 78 as free waves [e.g. Janssen et al., 2003; Battjes et al., 2004]. Baldock [2012] proposed that 79 bound wave release can only occur if short waves are in the shallow-water regime when 80 breaking, which condition is not necessarily met under short period waves and/or storm 81 conditions. Throughout their propagation in the inner surf zone of gently sloping beaches, 82 IG waves transfer energy to higher frequencies and steepen, which promotes their dissipation 83 through depth-limited breaking [Battjes et al., 2004; Van Dongeren et al., 2007; Lin and Hwung, 2012; De Bakker et al., 2014]. Several authors reported the saturation of wave runup 85 in the IG band under storm waves [Ruessink et al., 1998b; Ruggiero et al., 2004; Sénéchal 86 et al., 2011], possibly owing to IG wave dissipation close to shore. However, Fiedler et al. 87 [2015] did not observe any runup saturation under comparable incident conditions and beach 88 slope, suggesting that further research on IG wave dissipation is needed. IG wave energy 89 that is not dissipated in the surf zone can be reflected along the shore, a process that was 90 observed and quantified both in the field [Huntley et al., 1981; Guza and Thornton, 1985; 91 Elgar et al., 1994; De Bakker et al., 2014; Inch et al., 2017] and in laboratory experiments 92 [Battjes et al., 2004; Van Dongeren et al., 2007]. The reflection coefficient R^2 , which cor-93 responds to the ratio between the outgoing and the incoming IG wave energy, was shown 94 to increase with the beach slope and with the period of incident IG waves [Van Dongeren 95 et al., 2007]. As sandy beaches often exhibit steeper slopes in their upper part, a tidal 96 modulation of IG wave reflection is expected and was already reported [Okihiro and Guza, 97 1995]. Recent observations of free IG waves across the shelf of Oregon even revealed that 98 such a tidal modulation can be strong in the deep ocean [Smit et al., 2018]. The frequency-99 dependence of IG wave dissipation and reflection processes produces large differences in the 100 cross-shore structure of surface elevation spectra. Indeed, higher frequency IG waves tend 101 to be progressive across the surf zone while lower frequencies tend to develop quasi-standing 102 patterns due to a stronger reflection [De Bakker et al., 2014; Inch et al., 2017]. 103

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In a recent review paper, Bertin et al. [2018] pointed out that further research on IG 105 waves was needed to better understand their generation mechanisms, their transformations 106 across the surf zone and their reflection along the coast. This study aims to contribute to 107 this effort and to answer more specifically the following questions: how large can IG waves 108 be in the nearshore? What is the relationship between the shape of the incident short-wave 109 spectra and the IG waves spectra along the coast? Do IG waves necessarily suffer strong 110 dissipation and subsequent low reflection at a dissipative beach? To address these questions, 111 we present the analysis of an unpublished dataset collected under storm-wave conditions at a 112 dissipative beach located to the SW of Oléron island (France), complemented with numerical 113 modelling using XBeach [Roelvink et al., 2009]. 114

115 2 Study area

This study takes place at Saint-Trojan Beach, a dissipative beach located in the central part of the French Atlantic coast (Figure 1), along the south-western coast of Oléron Island. This stretch of coast corresponds to a 8 km-long sandspit, bounded to the South by the Maumusson Inlet [*Bertin et al.*, 2005]. The continental shelf in front of the study area is about 150 km-wide, with a very gently sloping shoreface, the isobath 20 m being found

about 10 km away from the shoreline. The tidal regime in this region is semi-diurnal and 121 macrotidal, with a tidal range varying between about 1.5 m during neap tides and 5.5 m dur-122 ing spring tides. Tidal currents remain weak at the studied beach and the impact of tides on 123 short waves remains mostly restricted to water level variations. According to Bertin et al. 124 2008, yearly-mean deep water wave conditions are characterized by a significant height 125 (H_s) of 2 m, a peak period (T_p) of 10 s and a mean direction of 285°N. During storms, H_s 126 can reach episodically 10 m in deep water and T_p can exceed 20 s with a westerly direction 127 [Bertin et al., 2015]. Due to the very gently sloping shoreface and the wide continental shelf, 128 the most energetic waves suffer strong dissipation and numerical wave models suggest that 129 their H_s hardly exceeds 5 m at the breaking point [Bertin et al., 2008]. 130

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The studied beach is mainly composed of fine and well sorted sands ($d_{50} = 0.18 -$ 132 $0.22 \,\mathrm{mm}$), which together with the energetic wave climate and the macrotidal range cause 133 its morphology to be non-barred and dissipative (Figure 1-B). Small-amplitude intertidal 134 bars can only develop after the persistence of fair weather conditions. Its slope typically 135 ranges from about 0.0015 at the shoreface to 0.015 in the intertidal area [Bertin et al., 136 2008, although a berm usually develops during the course of the summer period, with a 137 slope reaching 0.04. Due to the persistence of low to moderate-energy wave conditions in 138 autumn 2016 and early winter 2017, such a berm was still present during our field campaign. 139 This gently sloping morphology and the presence of a shallow shoreface induce a strong wave 140 refraction, so that the wave angle at breaking is usually small, with typical values smaller 141 than 10° [Bertin et al., 2008]. 142

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¹⁵⁰ 3 Field campaign and data processing

The field campaign presented in this study was motivated by the development of the 151 storm Kurt, which started to deepen to the SE of the Newfoundland on the 31/01/2017 and 152 crossed the North Atlantic Ocean following an Eastward track, centered between 48° and 153 50°. Based on the CFSR reanalysis [Saha et al., 2010], Kurt reached a minimum sea-level 154 pressure around 955 mbar on the 02/02/2017 about 1000 km to the West of the Bay of Bis-155 cay, accompanied with a ~ 500 km-wide band of western winds in the range 25 - 35 m.s⁻¹. 156 This atmospheric setting drove a very large and long period swell, that reached the French 157 coasts in the night of the 02/02/2017 to the 03/02/2017. At the deep water buoy of Biscay 158 (Figure 1-A), the mean wave period increased from 8.0 s to 13.0 s and H_s rapidly increased 159 from 3.0 m to almost 10.0 m, which value corresponds to a return period on the order of 160 one year [Nicolae-Lerma et al., 2015]. The wave hindcast described in Guérin et al. [2018] 161 suggests that the wave peak period T_p exceeded 20 s. The measurement period encompassed 162 four tidal cycles, from the 01/02/2017 to the 03/02/2017, characterized by a tidal range of 163 3.5 to 4 m. An Acoustic Doppler Current Profiler (ADCP1) with a frequency of 600 kHz 164 and equipped with a pressure sensor was deployed about 3 km offshore (Figure 1-B). In the 165 intertidal zone, 9 pressure transducers (PT) sampling at 4 Hz were deployed (Figure 1-C) as 166 well as a second ADCP (ADCP2), with a head frequency of 2 MHz mounted with a pressure 167 sensor at the location of PT3. The PTs were buried between 0.05 to 0.10 m of sand, in 168 order to avoid dynamic pressure errors. The position of each sensor was measured with a 169 differential GNSS, using a post-processing technique with a base station settled on the dune 170 crest in front of the instrumented profile [Guérin et al., 2018]. This GNSS was also used to 171 survey the intertidal beach topography at every low tide. The analysis of this data revealed 172 surprisingly small morphological changes for storm wave conditions, typically lower than 173 0.1 m along the cross-shore profile of the instruments. 174

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For each sensor, bottom pressure measurements were first corrected for sea level atmospheric pressure measured at the nearby meteorological station of Chassiron (Figure 1-A). The entire record was split into consecutive bursts of 30 min and only the bursts in which the



Figure 1. (A) Location of the study area in the Bay of Biscay, with the Biscay buoy (*) and Chassiron Meteorological station (*). (B) Detailed bathymetry of the study area (m relative to mean sea-level), showing the location of the offshore ADCP1 (*) and the instrumented cross-shore profile (dashed line). The coordinates are in meter (Lambert-93 projection) (C) Zoom-in on the intertidal part of the instrumented cross-shore profile, showing the location of the instruments used in this study (+) with respect to the beach topography (blue line).

sensor was continuously submerged were considered. PT9 was never continuously submerged during more than 30 minutes and was therefore discarded for the present study. Bottom pressure power density spectra $E_p(f)$ were computed using Fast Fourier Transforms, with 10 Hanning-windowed, 50% overlapping segments (20 degrees of freedom). These pressure spectra were then converted into elevation spectra E(f) considering linear wave theory [see for instance Bishop and Donelan, 1987]. The spectral significant wave height (H_{m0}) was computed as:

$$H_{m0} = 4\sqrt{m_0} \tag{1}$$

with

$$m_0 = \int_{f_{min}}^{f_{max}} E(f) df \tag{2}$$

where f_{max} was set to 0.4 Hz, a value for which the pressure correction reaches about 13 by 3 m water depth, well below the threshold of 100 to 1000 recommended by *Bishop and Donelan* [1987]. f_{min} is time-varying and defined following *Roelvink and Stive* [1989] or Hamm and Peronnard [1997] as half the continuous peak frequency f_p , computed from the offshore ADCP data as:

$$f_p = \frac{m_0^2}{m_{-2} \ m_1} \tag{3}$$

where

$$m_k = \int_{f_{min}}^{f_{max}} f^k E(f) df \tag{4}$$

The continuous peak frequency was preferred to the discrete one because it is a more 176 stable parameter, particularly under multimodal sea states, as it was the case at the begin-177 ning of the field campaign. As the incident peak period doubled during the field campaign, 178 a fixed frequency cutoff would not have allowed for a proper separation between the gravity 179 and the IG bands. In order to characterize the spatio-temporal variations of incoming IG 180 wave frequencies, the mean IG wave period T_{m02,IG^+} was also computed based on the in-181 coming IG wave signal, separated from the outgoing one at the ADCP2 and the ADV using 182 the method of *Guza et al.* [1984]: 183

$$T_{m02,IG^+} = \sqrt{\frac{m_0}{m_2}}$$
(5)

¹⁸⁴ 4 Numerical modelling

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4.1 Model description

XBeach is a two-dimensional modelling system that couples a Saint-Venant solver with 186 a simplified wave-action model and a sediment transport model to simulate the generation 187 and propagation of IG waves and the associated dynamics [Roelvink et al., 2009]. Based on 188 directional wave spectra provided along the open boundary, XBeach applies random phase 189 summation to reconstruct time series of the free surface elevation. A Hilbert transform 190 is applied to derive time-series of short-wave energy varying at the scale of wave groups 191 and imposed as boundary conditions in the wave action model. IG waves are represented 192 explicitly in the Saint-Venant model and are forced by the gradient of wave radiation stress 193 computed from the wave model. The incoming bound-wave is computed following Herbers 194 et al. [1994] and is imposed along the open boundary of the Saint-Venant model. 195

¹⁹⁶ 4.2 Model implementation

A rectilinear grid centered on the instrumented cross-shore profile and extending 4000 m in the alonghore direction was implemented. In the cross-shore direction, this grid starts from the dune crest and extends 11000 m offshore, corresponding to a mean water depth of about 23 m (Figure 1-B). The spatial resolution ranges from 20 m along the open

boundary to 1 m in the beach upper part (Figure 1-B). A sensitivity analysis revealed that 201 such a fine resolution was required to adequately represent IG waves in shallow water and 202 their reflection along the shoreline. Along the offshore open boundary, XBeach was forced 203 with time series of water levels recorded at ADCP1 (Figure 1-B). The wave model was 204 forced with time-series of directional wave spectra, originating from a regional application 205 of the WaveWatchIII model [Tolman, 2009] forced by wind fields from the CFSR reanalysis 206 [Saha et al., 2010]. In deep water, Guérin et al. [2018] showed that this wave hindcast 207 resulted in a normalized root mean squared discrepancy (hereafter NRMSD) of 10% for 208 H_{m0} and 6% for the mean wave period. Bottom friction was represented by a quadratic 209 bottom shear stress with a constant Chezy coefficient set to $80 \,\mathrm{m}^{0.5} \,\mathrm{s}^{-1}$. The horizontal 210 eddy viscosity was assumed constant $(0.5 \,\mathrm{m^2.s^{-1}})$ and the minimum water depth was set 211 to 0.01 m. Depth-limited breaking in the wave model was represented using the model of 212 *Roelvink* [1993a] with a breaking index γ adjusted to 0.38 after calibration. The model was 213 run for the duration of the field experiment (3 days), and time series of short wave height, 214 surface elevation and current velocities were archived at 2 Hz. In order to get insights into 215 IG wave generation mechanisms, additional numerical experiments were performed, where 216 wave forces were switched off either inside or outside the surf zone. The breaking formulation 217 of Roelvink et al. [2009] was used to determine whether a computational node was located 218 inside or outside the surf zone, i.e. if the local wave height was not exceeding the threshold 219 H_{max} as defined in *Roelvink* [1993b]. In the bound wave case, wave forces were turned off 220 inside the surf zone while in the breakpoint case, wave forces were turned off outside the 221 surf zone and the incoming boundwave was not included along the open boundary, follow-222 ing Pomeroy et al. [2012] and Bertin and Olabarrieta [2016]. Lastly, additional 24 h-long 223 simulations were run with constant water levels and wave forcing at specific times of the 224 measurement period to provide sufficient resolution and number of degrees of freedom for 225 spectral and bispectral analysis used to investigate IG wave energy transfers through non-226 linear triad coupling. Spectral estimates were computed following the same methodology 227 as for the field observations (section 3). Due to the random phase summation technique, a 228 substantial scatter in terms of IG waves can be observed from one simulation to another. For 229 model/data comparisons, bulk parameters computed from field observations were compared 230 against the mean of a model ensemble of 10 realizations. 231

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4.3 Model post-processing

Incoming and outgoing IG waves along the studied cross-shore profile were separated 234 using a Radon transform [Radon, 1917]. This technique projects the two-dimensional wave 235 field (here space corresponds to the studied cross-shore profile and time to 1800 s-long sam-236 ples) into polar space [Yoo et al., 2011]. The Radon transform is a powerful tool to study 237 nearshore waves as it allows the separation of incoming and outgoing wave fields on high-238 resolution data sets without any hypotheses on the hydrodynamics [e.g. Almar et al., 2014; 239 Martins et al., 2017]. The energy conservation properties of the method were assessed by 240 examining the ratio of energy computed using the sum of the separated signals and that of 241 the original total signal. The relative difference never exceeded 4% and was 1.4% on average 242 over the simulated period. Reflection coefficients R^2 were computed as the ratio between 243 the outgoing and the incoming IG wave energy. 244

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In order to analyze IG-wave generation mechanisms, the phase lag between the short wave energy envelope A and incoming IG waves η^+ was estimated using cross-spectral analysis:

$$\phi_{f,x} = \arctan\left[\frac{\mathcal{I}\{Cr_{A,\eta^+}(f,x)\}}{\mathcal{R}\{Cr_{A,\eta^+}(f,x)\}}\right]$$
(6)

where Cr is the cross-spectrum of the short wave envelope A and the incoming IG waves η_+ , 249 separated along the studied cross-shore transect using the Radon transform described above 250 and \mathcal{I} and \mathcal{R} are the imaginary and the real parts of the cross-spectrum. As this method 251 provides a phase lag per frequency, the averaged values over the IG band $(0.005 - 0.04 \,\text{Hz})$ 252 were considered, which eliminates the problem of selecting arbitrarily a single frequency 253 while providing more stable phase lag estimates. 254

Possible energy transfers in the IG band were investigated through bispectral analysis, 255 a technique originally introduced by Hasselmann et al. [1963] and then used in numerous IG 256 wave studies [e.g. Elgar and Guza, 1985; De Bakker et al., 2015]. Source terms for nonlinear 257 energy transfers between frequencies S_{nl} were computed using the stochastic formulation of 258 the second order nonlinear wave interaction theory of *Herbers et al.* [2000]: 259

$$S_{nl,f} = \frac{3\pi f}{h} \mathcal{I} \left\{ \left(\sum_{f'=0}^{f} \Delta f B_{f',f-f'} - 2 \sum_{f'=0}^{f_N - f} \Delta f B_{f',f} \right) \right\}.$$
 (7)

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where the term $\sum_{f'=0}^{f} B_{f',f-f'}$ represents the sum interactions in the imaginary part of the bispectrum B, and the term $-2\sum_{f'=0}^{\infty} B_{f',f}$ represents the difference interactions, as a particular frequency can contribute simultaneously to both difference and sum interactions.

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The 24-h analyzed model samples allow computing these terms with 602 degrees of freedom 263 and a spectral resolution of 0.0019 Hz. Further details on these analysis techniques can be 264

found in Bertin et al. [2018]. 265

5 Results 266

5.1 Characterization of short waves and IG wave heights

As explained above, the field campaign presented in this study was motivated by the 268 development of the storm Kurt, which generated offshore H_s of the order of 10.0 m with T_p 269 over 20 s. At the ADCP1 (Figure 1-A), the spectral significant wave height in the gravity 270 band (hereafter $H_{m0,G}$) increased from 2.0 m at the beginning of the campaign to a max-271 imum of 6.0 m on the 03/02/2017 (Figure 2-B). Over the same period, T_p increased from 272 10 - 15 s to more than 20 s. The increase in $H_{m0,G}$ is well captured by XBeach, with a 273 RMSD of 0.3 m and a NRMSD below 10%. In more detail, $H_{m0,G}$ is tidally modulated 274 during the last tidal cycle, which suggests that the surf zone at low tide extended offshore 275 of the ADCP1, located more than 3000 m from the shoreline. At the sensors deployed in 276 the intertidal zone (PT1 to PT9), wave heights were depth-limited and therefore strongly 277 tidally-modulated, with $H_{m0,G}$ roughly equal to half of the water depth (Figure 3). Wave 278 heights are well reproduced by XBeach, with a mean RMSD of 0.12 m over the 8 intertidal 279 sensors, which results in a NRMSD of 15% (Figure 5-A). Guérin et al. [2018] obtained 280 slightly better predictions using an adaptive wave breaking parameterization, where the 281 breaking index γ increases with the bottom slope. Such a parameterization is not available 282 in XBeach and is in any case questionable for a spectral model resolving wave groups [e.g. 283 Roelvink, 1993a]. 284

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Very large IG waves developed during the last tidal cycle, which corresponds to the 293 arrival of the large swell generated by the storm Kurt. At the ADCP1 (Figure 1-B), $H_{m0,IG}$ 294 ranged from 0.25 to 0.5 m during the first three tidal cycles and increased to a maximum 295 of 1.2 m (Figure 2), a value undocumented for a water depth of 13 m. $H_{m0,IG}$ are well 296 reproduced by XBeach, except during the second tidal cycle, where an underestimation by 297 a factor of 2 can be observed (Figure 2). In the intertidal zone, IG waves also increased 298 from about $0.5 \,\mathrm{m}$ during the first three tidal cycles to a maximum of $1.85 \,\mathrm{m}$ at PT6, with 299



Figure 2. Time series of observed (circles) and modeled (solid lines) bulk parameters and spectra at the location of the ADCP1 (Figure 1-B): (A) water depth; (B) H_{m0} in the gravity band; (C) H_{m0} in the IG band; (D) continuous peak (red) and mean (blue) wave periods and (E) power density spectra of observed water levels computed, where the white dashed line corresponds to the adaptive frequency cutoff between gravity and IG bands as described in section 3.



Figure 3. Modelled (blue line) against observed (black circle) water depth (A), H_{m0} in the gravity (B) and in the IG (C) bands at PT3/ADCP2.

individual IG waves exceeding 2.5 m (not shown). Figure 4 shows an example of time series 300 of measured water depth at the ADCP2 (Figure 4-A), where the low-pass filtered signal 301 reveals that individual IG waves exceeded 2.0 m and induced cross-shore velocities ranging 302 from $-2.0 \,\mathrm{m.s^{-1}}$ (offshore) to $+2.0 \,\mathrm{m.s^{-1}}$ (onshore)(Figure 4-B). This figure also shows that 303 alongshore velocities were much weaker (Figure 4-C), with the ratio of the alongshore over 304 the cross-shore IG velocity variance being lower than 10%. The extension of this analysis to 305 the whole ADCP2 dataset reveals this ratio is about $19 \pm 8\%$ (where \pm corresponds to one 306 standard deviation), which means that the IG wave dynamics was dominated by cross-shore 307 motions. 308



Figure 4. Time series of water depth (A), near bottom cross-shore (B) and longshore velocity (C) at 2 Hz (blue) and low-pass filtered (black), measured at the ADCP2/PT3 (Figure 1-C) on the 03/02/2017 at 5h00.

 $H_{m0,IG}$ are well reproduced by XBeach, except during the second tidal cycle where a 315 $0.2 \,\mathrm{m}$ negative bias can be observed (Figure 3), a problem that is also found at the ADCP1 316 (Figure 2-C). Considering the whole set of intertidal sensors, $H_{m0,IG}$ are reproduced by 317 XBeach with a RMSD of $0.14 \,\mathrm{m}$, which corresponds to a NRMSD of 20% (Figure 5-B). To 318 be more specific, $H_{m0,IG}$ are particularly underestimated when the incident short waves 319 have a large directional spreading (i.e. $> 35^\circ$, Figure 5-B), which occurred mainly during 320 the second tidal cycle (Figure 3). While discarding model results for such conditions, the 321 RMSD drops to 0.08 m and the NRMSD to 11%, which corresponds to high predictive 322 skills compared to previously published studies that provide a detailed validation of XBeach 323 [e.g. Van Dongeren et al., 2013; Bertin and Olabarrieta, 2016; Lashley et al., 2018]. The 324 underestimation of IG waves for broad short wave spectra is related to the original surfbeat 325 approach of Xbeach, where the wave energy from different directional bins is simply added 326 up, without considering the interference of the different wave components [Roelvink et al., 327 2018]. To overcome this problem, Roelvink et al. [2018] recently proposed an alternative 328 approach, where the mean wave directions are calculated first and the short wave energy 329 is then propagated along these directions. However evaluating the recent improvements of 330



Figure 5. Modelled against observed H_{m0} in the gravity (A) and in the IG (B) bands, where symbol color corresponds to the directional spreading of the incident short waves at the open boundary of the modeled domain (Figure 1-C).

XBeach is outside the scope of this study, focused on the physical processes associated with
 IG waves.

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5.2 Frequency repartition of IG wave energy

In order to investigate the evolution of the IG wave energy frequency repartition with a sufficient spectral resolution, we analyzed 3 h-long time-series of modeled and observed water depth, centered on the high tide of the last tidal cycle (storm Kurt). Incoming and outgoing modeled IG waves were separated using a Radon transform (section 4.3) but the number of intertidal sensors was too limited to apply this technique to field measurements.

Figure 6-A and 6-B show similarities between the modeled and the observed spatio-342 frequential patterns of IG wave energy in the intertidal zone, with a main peak between 0.01343 and 0.015 Hz, a super-harmonic peak around 0.02 and 0.025 Hz and a subharmonic peak be-344 tween 0.001 and 0.005 Hz. From the beach lower part to the beach upper part, these three 345 peaks slide towards higher frequencies and the lower frequency one totally dominates the 346 power density spectra along the shoreline. Such intriguing patterns were already observed 347 in laboratory experiments of IG waves [Battjes et al., 2004; Buckley et al., 2018] and short 348 waves [Martins et al., 2017] and are explained by interference between incoming and out-349 going IG waves, creating nodes and antinodes, whose position with respect to the shoreline 350 varies with the IG wave frequency and the bottom slope. Figure 6-D and 6-E corroborate 351 this hypothesis and show that this pattern disappears in the modeled spectra corresponding 352 to the incoming and outgoing signals. The analysis of the modeled spectra for the incoming 353 signal (Figure 6-D) reveals that, at the shoreface, IG wave energy is mostly located around 354 a first peak at 0.01 Hz and a second broader peak centred around 0.005 Hz. A superhar-355 monic secondary peak is also present around 0.02 Hz. Approaching the shoreline, IG wave 356 energy at higher frequencies decreases while most of the energy is located between 0.002357 and 0.006 Hz. The spectra of the outgoing signal (Figure 6-E) suffer less transformation 358 and, compared to the incoming signal, display much less energy at frequencies above 0.015 359 to 0.02 Hz. 360

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This analysis was extended to the whole studied period based on the mean incoming IG wave period T_{m02,IG^+} (Figure 7-A), computed from the incoming IG wave signal. For moderate-energy wave conditions, this figure shows that T_{m02,IG^+} first decreases from about



Figure 6. Spatio-frequential repartition of IG wave power spectral density modeled (A) and observed (B) for the total signals in the intertidal zone and modeled along the studied cross-shore transect for the total (C), the incoming (D) and the outgoing (E) signals.



Figure 7. (A) Spatio-temporal repartition of incoming mean IG wave period T_{m02,IG^+} along the studied cross-shore transect. Modeled against observed time-series of T_{m02,IG^+} and the ADCP2 (B) and the ADV (C).

60 to 70 s in 10 m water depth to about 50 s in 2 m water depth. For a given location along 368 the studied cross-shore profile, T_{m02,IG^+} exhibits therefore a substantial tidal modulation. 369 In water depth less than 1 m, T_{m02,IG^+} increases back to 60-70 and even over $80 \,\mathrm{s}$ at low 370 tide (Figure 7). Under storm wave conditions (3/02/2017), a different behaviour is observed, 371 with a continuous increase in T_{m02,IG^+} from about 70 s in 10 m water depth to 90 - 100 s 372 in 2 m water depth. In water depth less than 1 m, T_{m02,IG^+} further increases up to 120 s at 373 low tide. The modeled temporal evolution of T_{m02,IG^+} fairly matches the observations at 374 the location of the ADCP2 and ADV, with a RMSD of 8s, corresponding to a 12% error. 375

5.3 Reflection

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The comparison between observed and modeled reflection coefficient R^2 shows a good agreement in the intertidal zone, with a RMSD of 0.12, corresponding to a normalized error of about 15% (Figure 8-B and -C). Under moderate-energy incident short waves, both modeled and observed R^2 show a strong tidal modulation, with values below 0.2 at low tide and close to 1.0 at high tide, suggesting full reflection along the waterline. This tidal modulation substantially weakens under storm wave conditions, with R^2 above 0.5 at low tide along the waterline.



Figure 8. (A) Spatio-temporal variations of reflection coefficients R^2 computed over the studied period using a Radon transform [*Radon*, 1917] and modeled (blue) against observed (circles) timeseries of R^2 at the ADCP2 (B) and the ADV (C).

Such a tidal modulation of R^2 was already reported by *Okihiro and Guza* [1995], who explained the higher reflection at high tide by the steeper slope of the beach upper part. In the present study, the beach slope increases from about 1:100 in the lower part of the intertidal zone to 1:25, which could directly explain the observed tidal modulation of IG wave reflection. *Battjes et al.* [2004] proposed that IG wave reflection is related to a normalized beach slope parameter β_H , defined as:

$$\beta_H = \frac{\beta}{\omega} \sqrt{\frac{g}{H}},\tag{8}$$

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where β is the beach slope, ω is the IG wave angular frequency, g is the gravitational 394 acceleration and H corresponds to the height of the incoming IG wave. According to 395 these authors, a large reflection would occur for β_H substantially above unity. Considering 396 incoming IG waves of height 0.5 m and frequency 0.01 Hz, β_H would increase from about 397 0.7 at low tide to 3.0 at high tide, which explains the observed tidal modulation of R^2 . 398 Outside the surf zone, model results suggest R^2 well above unity at high tide, which was 399 already observed in the field by Elgar et al. [1994] and Sheremet et al. [2002]. These authors 400 interpreted this behaviour by the fact that IG waves were gaining energy between their 401 sensors and the shoreline, before being reflected from a steep beach face. 402

403 6 Discussion

6.1 Generation mechanisms

The IG waves observed during our field campaign correspond to the largest values reported in the literature [*Ruessink*, 2010; *Fiedler et al.*, 2015; *Inch et al.*, 2017], while our adaptive frequency cutoff to separate gravity and IG bands is more conservative than the fixed frequency cutoff used in these studies. At gently sloping beaches, it is usually admitted that IG waves are generated through the bound wave mechanism [*Battjes et al.*, 2004; *Inch et al.*, 2017; *Bertin et al.*, 2018].



Figure 9. Modeled incoming IG wave height H_{m0,IG^+} in the baseline simulation (A), the bound wave case (B) and the breakpoint case (C). The black dotted line corresponds to the offshore limit of the surf zone, as defined in section 4.2

⁴¹⁴ Considering the analytical solution proposed by *Hasselmann* [1962] to compute the ⁴¹⁵ bound wave based on directional wave spectra, energy transfers from the gravity band to ⁴¹⁶ the IG band increase when the wave energy is large and distributed over narrow spectra.

Therefore, the very energetic and narrow-banded swell associated with the storm Kurt 417 would directly explain the development of such large IG waves. In order to verify this 418 hypothesis, we compared the incoming IG wave heights H_{m0,IG^+} computed from our baseline 419 simulation with those computed from simulations where only either the bound wave or the 420 breakpoint mechanisms were represented (see section 4.2). This comparison reveals only 421 modest differences between the baseline simulation (Figure 9-A) and the simulation where 422 only the bound wave is considered (Figure 9-B). Conversely, the simulation where only 423 the breakpoint mechanism is considered (Figure 9-C) shows much smaller IG waves, with 424 H_{m0,IG^+} ranging from 0.1 to 0.2 m during first three tidal cycles and 0.4 m around the 425 peak of the storm. This analysis confirms that IG waves were mostly generated through 426 the bound wave mechanism, independently from the tidal phase or the incident short-wave 427 energy. 428

In accordance with this first finding, the modeled incoming IG waves are negatively 429 correlated with the short wave energy envelope (hereafter WEE, Figure 10-A). Strongest 430 correlations are found from the breaking point up to water depths of about 12 m, with 431 coefficients ranging from -0.4 to -0.9 over the studied period (always significant at 95%, 432 mean -0.54 ± 0.16). In more details, the strongest negative correlations are not found at 433 zero lag and incoming IG waves are lagging behind the WEE by a few seconds. Figure 10-B 434 shows the modeled spatio-temporal evolution of the normalized phase lag between incoming 435 IG waves and the WEE, shifted by π for the sake of readibility. For moderate-energy incident 436 conditions, this phase lag continuously increases from about $\pi/20$ by 10 m water depth to 437 about $\pi/5$ by 4 m water depth, which corresponds to the beginning of the surf zone. In 438 the inner surf zone, the correlation cancels out and then switches to positive (Figure 10-A), 439 due to the modulation of short wave heights by IG waves, a process that is well documented 440 [Guza et al., 1984; Tissier et al., 2015; Inch et al., 2017]. 441



Figure 10. (A) Modeled correlation coefficient between the WEE and incoming IG waves and (B) modeled normalized phase lag, shifted by π (i.e. phase lag in radians shifted by π and then normalized by π) between incoming IG waves and the WEE. The black solid lines correspond to isolines of the mean water depth.

For storm wave conditions, a phase lag of about $\pi/10$ is already present by 10 m water depth (i.e. 3000 m from the shoreline, Figure 10) and increases up to $\pi/5$ at the beginning of

the inner surf zone. According to Van Dongeren et al. [2007], this phase lag enhances energy 448 transfers from the short waves to the IG waves. Due to the gentle slope of the studied beach, 449 this process is active over several kilometers under storm waves, which would explain the 450 development of such large IG waves. However, our field dataset entails several limitations 451 that prevent from verifying these values. First, the ADCP1 was located more than 3000 m 452 from the intertidal profile, which implies travel times of IG waves ranging from 400 to 500 s, 453 depending on the tidal phase. Over such a long distance/duration, IG waves suffer multiple 454 transformations (see next sections) so that we were unable to correlate the offshore WEE 455 with the IG waves at the intertidal sensors. Second, the width of the surf zone ranged from 456 200 m to more than 3000 m, so that the intertidal sensors were always located inside the surf 457 zone, where the correlation between IG waves and the WEE is biased positively as explained 458 above. 459

6.2 Energy transfers

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In order to better explain the spatio-temporal repartition of IG wave energy, source 461 terms for nonlinear energy transfers between IG wave frequencies S_{nl} were computed using 462 bispectra following Herbers et al. [2000] based on 24 h-long runs corresponding to moderate-463 energy and storm conditions, both at low and high tide (Figure 11). This figure reveals 464 that, at high tide, substantial energy transfers from frequencies in the range 0.01-0.03 Hz 465 towards higher frequencies start to occur about 200 m from the shoreline (Figure 11-A). At 466 low tide, similar energy transfers occur but starting about 500 m from the waterline (Figure 467 11-B). Surprisingly, the overall patterns of energy transfers are not very different for storm 468 conditions, except that they affect lower IG wave frequencies (0.002 Hz to 0.02 Hz) (Figure 469 11-C and -D). Note that for these high-energy cases, the overall patterns are shifted onshore 470 because the mean water level is higher due to a larger wave setup. 471



Figure 11. Source terms for nonlinear energy transfers between IG wave frequencies along the studied cross-shore profile computed for each frequency following *Herbers et al.* [2000] over 24 h stationary runs for moderate and high-energy consistions, both at low and high tide.

For moderate-energy wave conditions, the cross-shore evolution of Snl (Figure 11-A 475 and -B) suggests that the decrease in mean incoming IG wave period T_{m02,IG^+} with the 476 water depth (Figure 7) and its subsequent tidal modulation is explained by energy transfers 477 from principal components to higher harmonics via non-linear coupling between triads. In 478 shallow depth (i.e. less than 1 m), the increase in T_{m02,IG^+} is explained by the breaking 479 of the higher-frequency IG waves, a process well supported at gently-sloping beaches by 480 field observations [De Bakker et al., 2014; Inch et al., 2017], laboratory experiments [Battjes 481 et al., 2004; Van Dongeren et al., 2007; Padilla and Alsina, 2017] and numerical modelling 482 [Ruju et al., 2012; De Bakker et al., 2016b; Mendes et al., 2018]. For storm wave conditions, 483 a different situation occurs, where T_{m02,IG^+} continuously increases across the surf zone, so 484 that T_{m02,IG^+} almost doubles from 10 m water depth to the shoreline (Figure 7). While 485 the increase in T_{m02,IG^+} close to the shoreline can also be attributed to the breaking of 486 the highest frequency IG waves, an additional mechanism should be active in intermediate 487 water depths where IG waves cannot break. Indeed, classical interactions between pairs of 488 frequencies could hardly explain this behaviour as the computation of S_{nl} reveals virtually 489 no energy transfers towards very low frequencies, as for moderate-energy conditions (Figure 490 11). Analyzing field observations collected under storm waves at a steep rocky shore in 491 western Brittany, Sheremet et al. [2014] reported the development of 300 s period IG waves 492 along the shore. These authors applied a phase-resolving triad numerical model, which failed 493 to reproduce such low frequency IG waves and concluded that, as in the present study, it 494 does not result from classical non-linear triad interactions. 495



Figure 12. Timestack sample of modeled incident surface elevation (η_{IG^+}) along the studied 496 cross-shore transect around high tide for (A) moderate incident energy and (B) high-energy (storm 497 Kurt). Numbers correspond to IG wave crests and the dashed black lines correspond to the troughs 498 of the WEE, automatically detected at x = 3000 m and then computed across the surf zone. 499

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While the development of IG wave super-harmonics at gently-sloping beaches is reasonably well documented in the literature and explained through IG-IG nonlinear interactions [Elgar and Guza, 1985; De Bakker et al., 2015, 2016b; Inch et al., 2017; Padilla and Alsina, 2017], the transfer of energy towards subharmonic frequencies received much less attention [De Bakker et al., 2016b]. To further look into this phenomenon, we analyzed timestack samples of modeled incoming free surface elevation, close to high tide under moderate energy (first tidal cycle) and storm waves (fourth tidal cycle, storm Kurt) (Figure 12-A and -B). On this figure, the troughs of the WEE were automatically detected at x = 3000 m and their location was then computed across the surf zone using the wave group velocity C_q , given by the linear wave theory and the representative frequency used in XBeach, which corresponds to $1/T_{m-1,0}$. For moderate energy, Figure 12-A shows that IG wave crests 1 to 12 follow the WEE troughs, which means that they propagate at C_g and shows that

they are bound to the wave groups. This behaviour illustrates the dominance of the bound 512 wave mechanism in IG wave generation, as explained in section 6.1. This situation strongly 513 contrasts with storm wave conditions, where IG wave crests depart from the WEE troughs 514 at different locations across the surf zone, meaning that they no longer travel at C_q but 515 behave as free waves (Figure 12-B). For instance, focusing on the IG wave crests number 1 516 to 4, crest 1 seems bound to the wave group until x = 500 m, crest 2 and 4 until x = 2300 m 517 and crest 3 until x = 1000 m. Although less impressive, this process can also be observed 518 for IG wave crests 6 and 10. Considering a mean water depth of 12 m at x = 3000 m and a 519 representative frequency of 0.064 Hz at the time of Figure 12-B, bound IG waves propagate 520 at $C_g = 9.8 \, m.s^{-1}$ while free IG waves propagate at $10.84 \, m.s^{-1}$. Whereas such differences 521 only represent 11 % and further less when approaching the coastline, they result in travel-522 time differences ranging from 25 to 30 s once integrated over such a large surf zone. This 523 allows the merging of individual crests into longer-period IG waves close to shore, as for 524 instance crest 2 with crest 1 and crests 4 with crest 3 (Figure 12-B). A closer look at the 525 model results however suggests that the behaviour of the near-shore IG wave-field is more 526 complex. For instance, new IG wave crests appear inside the surf zone, such as between 527 crests number 2 and 3 at $x = 1500 \,\mathrm{m}$ or between crests 11 and 12 at $x = 2300 \,\mathrm{m}$. As 528 these IG wave crests coincide with the crests of the WEE, they may correspond to free IG 529 waves generated through the breakpoint mechanism [Symonds et al., 1982]. This hypothesis 530 is supported by Figure 9-C, which shows that the breakpoint mechanism is active 2000 m 531 from the shoreline under storm waves. Future studies based on extensive field measurements 532 and/or phase resolving models will have to investigate the detailed processes responsible for 533 IG wave merging and the possible contribution of breakpoint-generated free IG waves. 534

Small beach slopes and resulting wide surf zones exacerbate this process as it provides 535 individual IG waves with more time to travel at slightly different speed and catch up with 536 one another. Although it could not be evidenced here, a modulation of individual IG wave 537 celerity by longer IG waves might also be possible in this region of the shoreface [*Tissier*] 538 et al., 2015], which would also promote the merging of individual waves. Ultimately, we 539 propose that the interactions of incident bound and free IG waves along the gently sloping 540 shoreface of Oléron Island explain the increase of the incoming IG wave period by a factor 541 of two across the surf zone under storm waves (Figure 7). This mechanism also suggests 542 that the frequency of IG waves close to shore is not only controlled by the shape of the 543 incident short wave spectra [i.e. Hasselmann, 1962] but also by the width of the surf zone 544 and hence the slope of the beach. This mechanism will have to be verified in the field, 545 although instrumenting a surf zone several kilometre-wide is very challenging. 546

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6.3 Implications for other studies

In this study, we observed reflexion coefficient R^2 to be very high for a dissipative 548 beach. For instance, De Bakker et al. [2014] and Inch et al. [2017] reported values for R^2 of 549 the order of 0.5 or below for beaches of comparable mean bottom slope. The presence of a 550 steeper upper beach in Saint Trojan can explain the higher reflexion observed at high tide, 551 but not at low tide under storm wave conditions, where R^2 hardly drops below 0.5. We pro-552 pose that the high reflection that we observe under storm waves is related to energy transfers 553 towards subharmonic frequencies (section 6.2), so that the resulting 200 to 300 s period IG 554 waves are too long to suffer a substantial dissipation, even when propagating over a mild 555 slope. Albeit not addressing the underlying mechanisms, a few other studies reported the 556 development of IG waves with frequencies around 0.005 Hz under storm waves and suggested 557 that they had a key role in runup [Sheremet et al., 2014; Ruggiero et al., 2004]. Indeed, 558 such low frequency IG waves suffer less dissipation when propagating in shallow depth and 559 560 may therefore propagate farther inland and induce larger damages in low-lying coastal zones compared to higher-frequency IG waves. The generation of low-frequency IG waves through 561 merging of bound and free waves across wide surf zones appears therefore as a process of 562 key importance and will have to be investigated at other sites, such as the very gently 563 sloping coastlines of Bangladesh, which are extremely vulnerable to flooding under tropical 564

hurricanes [*Krien et al.*, 2017]. This process should also be investigated at large estuaries and tidal inlets, where the ebb deltas can drive the development of kilometer-wide surf zones.

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Besides, it is nowadays well admitted that free IG waves in the ocean are the main 568 source of background-free oscillations of the solid earth, also referred to as "the hum of 569 the Earth" [e.g. Rhie and Romanowicz, 2006; Webb, 2007]. As better understanding the 570 generation of the hum could help refining the use of microseism background signals to 571 investigate the structure of the solid Earth, understanding IG wave reflection along the 572 573 coast is of key importance. Correlating deep ocean pressure measurements with short-wave bulk parameters over the North Atlantic Ocean, Crawford et al. [2015] proposed that the 574 main source for free IG wave generation was located between the Western part of the Iberic 575 Peninsula and NW Africa, namely because they display steep beaches. Our study suggests 576 that strong reflection can also occur along dissipative beaches under storm waves, which 577 will have to be considered in future studies on deep ocean IG waves. 578

579 7 Conclusions

This study aimed at improving our understanding of the generation and transfor-580 mation of IG waves, combining an unpublished data set with numerical modeling. The 581 development of exceptionally large IG waves was primarily explained by very energetic and 582 narrow-banded incident short waves associated with the storm Kurt. The analysis of model 583 results revealed that the bound wave mechanism was clearly dominant over the breakpoint 584 mechanisms in the generation of IG waves. Model results also showed that the gently sloping 585 shoreface of the study area leads to the presence of a phase lag between IG waves and WEE 586 over several kilometers under storm waves, which certainly enhanced IG wave generation 587 through a dynamic shoaling, although our dataset did not allow to further analyze this 588 process. The XBeach modelling system was found capable of reproducing the generation 589 and transformation of such IG waves, which is an important result as XBeach is becoming 590 more and more popular while detailed hydrodynamic validations are lacking under storm 591 conditions. Model results and field observations revealed that large energy transfers occur in the IG band across the surf zone, with the development of superharmonics but also the 593 transfers of energy towards subharmonic frequencies. While the former behaviour is reason-594 ably well documented in the literature and explained by non-linear coupling between triads, 595 the latter was reported in a few studies only and we propose that it is (at least partly) 596 related to the merging of IG waves across the surf zone. According to our modelling results, 597 IG wave merging seems to be related to the combination of free and bound IG waves in the 598 surf zone, traveling at different celerities over a long distance. This mechanism will have to 599 be verified on the field and is potentially of key importance under extreme storms, because 600 low-frequency IG waves suffer little dissipation and may therefore propagate farther inland 601 and induce larger damages compared to higher frequency IG waves. This mechanism also 602 results in IG-wave reflection surprisingly high for a dissipative beach, which has implications 603 for future studies on deep ocean IG waves. 604

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