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Cross-shore suspended sediment transport in a macro-tidal and low-sloping upper shoreface

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Abstract

The driving mechanisms of sediment transport across the upper shoreface have yet been poorly examined because of the scarcity of field measurements beyond the inter-tidal area, which is mainly due to instrumentation difficulties. Here, we investigate the cross-shore suspended sediment transport from synchronized timeseries of bottom pressure, current velocity and suspended sediment concentration measured by 13 m water depth, in a macro-tidal and low-sloping shoreface, under storm waves and fair weather conditions. Depending on the tidal phase and the incoming wave height, the measuring structure was alternatively located in the shoaling zone and in the outer surfzone. The analysis of field observations first revealed that steady currents dominate suspended transport, with a notable contribution of the wave driven return current under storm wave conditions. Cross-spectral analysis further indicates that short waves globally induce an offshore-directed transport under energetic conditions associated with a crest-to-trough phase lag effect. The contribution of infragravity (IG) waves, which reaches as much as 20% of the total transport, shows contrasting directions depending on whether IG waves are bound to the short wave groups or propagate as free waves.

Keywords

Suspended sediment transport, shoreface, undertow, short waves, phase lag, IG waves

1 Introduction

At seasonal to decadal timescales, the shoreface transitionning between the surf zone and the shelf plays a fundamental role in the morphological evolution of adjacent shorelines (Hamon-Kerivel et al., 2020). In coastal regions characterized by large tidal ranges and gently sloping shoreface, this transition region can shift by several kilometers in the cross-shore direction, depending on the tidal phase and the incoming wave height (e.g. see Pezerat et al., 2022).

It is acknowledged that sediment transport generally takes place as both bedload and suspended load, but fine sandy sediments are more likely transported in suspension across wave-exposed nearshore area (e.g. Jaffe et al., 1985; Osborne and Greenwood, 1992; Russell, 1993). Several studies further showed that suspended sediment transport on beaches is associated with both oscillatory wave motions and steady currents, i.e. time-averaged over several wave periods (e.g. Osborne and

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34 Greenwood, 1992; Russell, 1993; Ruessink et al., 1998). In situ observations support the fact that
35 the oscillatory transport within the gravity (G) band - in frequency (f) space - associated with short
36 waves ($0.04 \text{ Hz} \lesssim f \lesssim 0.4 \text{ Hz}$) is mainly onshore-directed and dominates the suspended load motion
37 under moderate wave energy conditions (Osborne and Greenwood, 1992; Russell, 1993). Based on
38 laboratory experiments, several studies further related the transport at short wave frequencies
39 under sheet flow regime to the wave shape that is determined by a certain degree of skewness
40 and asymmetry (e.g. Dibajnia and Watanabe, 1992; Dohmen-Janssen et al., 2002; Hsu and Hanes,
41 2004; Grasso et al., 2011). Skewed waves are characterized by stronger velocities in the direction
42 of propagation under their crests, hence (presumably) mobilizing a larger amount of sediment
43 than under their troughs, below which the velocity is oriented in the opposite direction. While
44 Hsu and Hanes (2004) found that the transport is oriented in the direction of wave propagation
45 under skewed waves, Dohmen-Janssen et al. (2002) and Grasso et al. (2011) actually observed
46 that the orientation of the transport depends on the phase lag between the wave passing and
47 the re-suspension event. For highly skewed waves, the peak of suspended sediment concentration
48 could be slightly shifted in time to be found concomitant with the wave's trough, which thus
49 results in a transport offshore-directed (crest-to-trough phase lag effect). Within the surf zone,
50 waves become more asymmetric, resulting in a skewed acceleration over the wave period as the
51 acceleration is stronger under the front face than under the rear face of the wave (see Elgar et al.,
52 2001, their Fig.1). Several studies reported that sediment resuspension is more likely correlated
53 with orbital acceleration (e.g. Hanes and Huntley, 1986; Jaffe and Rubin, 1996), such that for
54 asymmetric waves, a trough-to-crest phase lag is induced between the resuspension event and the
55 wave passing, resulting in a transport onshore-directed. Infragravity (IG) waves ($0.001 \text{ Hz} \lesssim f \lesssim$
56 0.04 Hz) have also a direct bearing on suspended sediment transport, although observations result
57 in contrasting direction and relative importance of the associated transport (see the pioneering
58 studies from Abdelrahman and Thornton, 1987; Beach and Sternberg, 1988; Roelvink and Stive,
59 1989; Osborne and Greenwood, 1992; Russell, 1993, among many others). de Bakker et al. (2016)
60 proposed a conceptual model relying on the local ratio of IG wave height to short wave height
61 to infer the direction of the transport and its relative importance based on observations collected
62 either on moderately or gently sloping beach to account for the different IG waves propagation
63 regime, depending on whether IG waves are bound to the short wave groups or propagate as free
64 waves. Finally, the wave-driven return current (the so-called undertow) was found to be the main
65 mechanism that induces an offshore directed transport within the surf zone (e.g. Russell, 1993;
66 Ruessink et al., 1998).

67 The driving mechanisms of the cross-shore sediment transport beyond the inner surf zone are
68 still poorly understood because of the scarcity of field measurements in relatively deeper water,
69 which is mainly due to instrumentation difficulties, especially under storm conditions. In this
70 context, the present study aims to investigate the relative contributions to cross-shore suspended
71 sediment transport over the macro-tidal and low sloping upper shoreface in front of a beach made
72 of fine sands, based on in situ measurements at a 12.8 m-depth location. The dataset covers storm
73 events as well as fair weather conditions, thus encompassing a variety of forcing regimes.

74 2 Methods

75 2.1 Study area

76 The study area is located in front of Saint-Trojan beach, a 8 km-long sandspit along the South-
77 Western coast of the Oléron Island in the central part of the French Atlantic coast (Fig. 1a). This
78 stretch of coast spans in a North-West to South-East direction forming an angle of approximately
79 5° from the true North at the median latitude. This area is characterized by a very gently sloping
80 shoreface (the isobath 20 m being found 10 km offshore, see Fig. 1b) and a non-barred dissipative
81 beach composed of fine sandy sediments with a median grain size of the order of 0.2 mm (Bertin
82 et al., 2008). In this region, tides are semi-diurnal and range from 1.5 to 5.5 m, which corresponds
83 to a macro-tidal regime. Yearly mean wave conditions along the 30 m isobath are characterized
84 by a significant wave height of 1.6 m, a mean wave period of 5.9 s and a direction of 285° from the
85 true North (Dodet et al., 2019), but the offshore significant wave height can exceed 10 m with peak
86 periods over 20 s (Bertin et al., 2015).

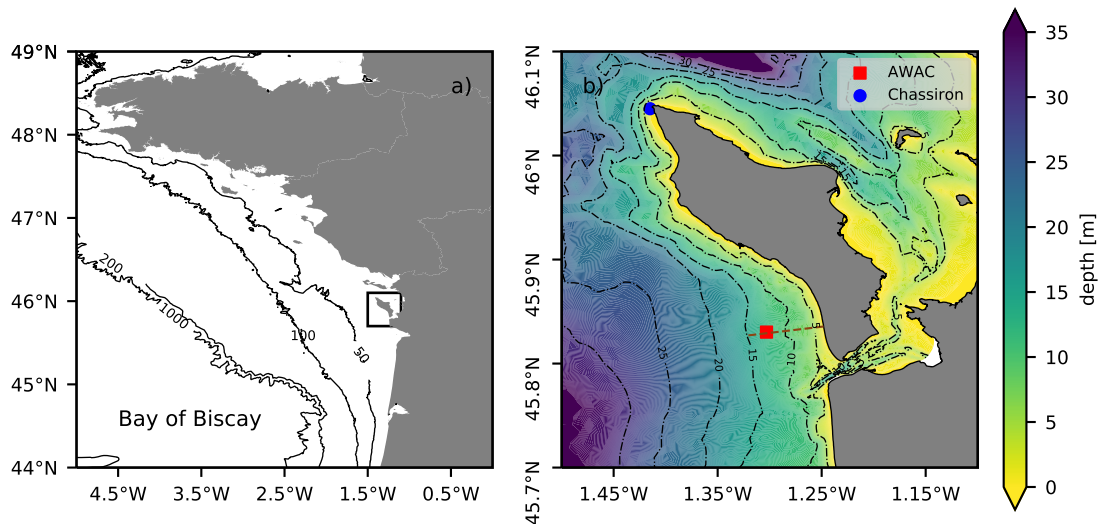


Figure 1: **a)** Location of the study area in the bay of Biscay. **b)** Bathymetric map of the study area with isobaths reduced to the Mean Sea Level (MSL) displayed every 5 m (black dash-dotted lines), sensors location and position of the Chassiron meteorological station while the dashed line symbolizes the cross-shore direction.

87 2.2 Field campaign and data processing

88 This study builds on the field campaign carried out between January and February 2021 already
89 presented by Pezerat et al. (2022). The measurements from a high resolution (1 MHz) Acoustic
90 Wave and Current Profiler (AWAC) deployed by 12.8 m-depth below MSL are here combined with
91 measurements from an Optical Backscatter Sensor (OBS) that was plugged to the AWAC and
92 deported 0.5 m above its head. The OBS signal intensity was converted into Suspended Sediment
93 Concentration (SSC) after a calibration in the laboratory, using sediment samples collected at the
94 location of the instruments. The measurements alternated a "current cycle" and a "wave cycle"
95 each hour. During the "current cycle", 10-min averaged velocity profile measurements were col-

96 lected along the vertical axis, while during the "wave cycle", velocity measurements within a fixed
 97 cell located 4.5 m above sensor's head, bottom pressure measurements and SSC measurements
 98 were performed at 2 Hz during 20 minutes. Pressure measurements were first corrected for sea
 99 level atmospheric pressure using data collected at the nearby meteorological station of Chassiron
 100 (Fig. 1b), detrended and converted into a sea surface elevation signal assuming a hydrostatic pres-
 101 sure. Pressure attenuation with depth due to non-hydrostatic effects was then corrected using the
 102 Transfer Function Method based on the linear theory (TFM, e.g. see Bishop and Donelan, 1987)
 103 to compute the short waves free surface elevation signal (η_L). This method requires an upper
 104 cutoff frequency (f_c , here set to 0.2 Hz) so as to remove high frequency noise that is amplified by
 105 the TFM correction and to prevent the over-amplification of high-frequency energy levels due to
 106 non-linear interactions in intermediate and shallow-water depths (Mouragues et al., 2019). Finally
 107 the sea surface elevation density spectra $E(f)$ were computed by means of a fast Fourier Trans-
 108 form on 5 Hanning-windowed segments with a 50% overlap which allows an *ad hoc* compromise
 109 between statistical stability (10 degrees of freedom) and frequency resolution (4.2 mHz) required
 110 to investigate the contributions to the elevation signal from both the G and the IG bands. The
 111 spectral significant wave height (H_{m0}) and the continuous peak period (T_{pc}) were computed using
 112 the p - th moments of the spectra:

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

113 such that:

$$H_{m0} = 4\sqrt{m_0} \quad (2)$$

114 and,

$$T_{pc} = \frac{m_{-2}m_1}{m_0^2} \quad (3)$$

115 where f_{min} was set to 0.001 Hz for the IG band and f_{max} was set to $f_c = 0.2$ Hz for the G band,
 116 while both spectral domains were separated based on the separation frequency $f_{min,G}$ defined as
 117 half the peak frequency (Hamm and Peronnard, 1997; Bertin et al., 2020).

118 During the studied period, the waves globally reached the study area from a Westward direction,
 119 almost normally incident to the shoreline, suggesting a prevailing contribution of the cross-shore
 120 transport. The data from the "wave cycle" were thus used to compute the time-averaged sediment
 121 transport rate in suspension along the cross-shore direction. It is worth pointing out that velocity
 122 and SSC measurements were not vertically co-located, such that the quantitative estimate of the
 123 sediment fluxes is presumably biased. It is nonetheless assumed that both measurements show
 124 consistent patterns allowing to identify the relative contributions of the different forcing processes
 125 involved. This strong assumption is found justified for three reasons: (i) considering the forcing
 126 conditions, the transport of fine sandy sediment likely takes place as suspended load (see Appendix
 127 C), (ii) the decrease of short wave orbital velocities between the wave cell and the OBS roughly
 128 estimated by means of the linear wave theory, is rather moderate, of the order of few percents (see
 129 Appendix B) and (iii) the steady current keep the same orientation along the vertical at this scale
 130 (e.g., see Pezerat et al., 2022). Following the seminal approach of Jaffe et al. (1985), instantaneous
 131 SSC (c) and cross-shore velocity (u) are decomposed into a mean ($\bar{\cdot}$) and an oscillating component

132 $(\bar{\cdot})$ such that the time-averaged sediment flux reads:

$$q_{tot} = \overline{u\tilde{c}} = \overbrace{\overline{u\tilde{c}}}^{q_{mean}} + \overbrace{\overline{u\tilde{c}}}^{q_{osc}} \quad (4)$$

133 where q_{mean} corresponds to the transport forced by the mean current, that integrates tidal current
 134 and the wave-driven mean current, and q_{osc} , also referred to as the flux coupling, measures the
 135 correlation between the fluctuations of velocity and concentration. This term can be readily esti-
 136 mated from the co-spectrum between \tilde{u} and \tilde{c} , which is the real part ($\Re\{\cdot\}$) of the cross-spectrum
 137 between both signals ($S_{\tilde{u}\tilde{c}}$):

$$q_{osc} = \int_f \Re\{S_{\tilde{u}\tilde{c}}\}df = \overbrace{\int_{f_{min,IG}}^{f_{min,G}} \Re\{S_{\tilde{u}\tilde{c}}\}df}^{q_{osc}^{IG}} + \overbrace{\int_{f_{min,G}}^{f_{max}} \Re\{S_{\tilde{u}\tilde{c}}\}df}^{q_{osc}^G} \quad (5)$$

138 where f_{max} was here set to 0.4 Hz as measurements were directly considered. For each 20 min-long
 139 burst, raw signals were first detrended to compute the oscillating components of both signals. The
 140 cross-spectrum was then computed from the product of the conjugate of the Fourier transform of
 141 \tilde{u} with the Fourier transform of \tilde{c} using again 5 Hanning-windowed segments with a 50% overlap.
 142 Finally, the relative absolute contributions of each term to the time-averaged sediment flux is given
 143 by:

$$R_X = \frac{|X|}{|q_{mean}| + |q_{osc}^{IG}| + |q_{osc}^G|} \quad (6)$$

144 where X represents either q_{mean} , q_{osc}^G or q_{osc}^{IG} .

145 3 Results

146 The local time-averaged sediment flux is alternatively of very low intensity (e.g. between the 05/02
 147 and 08/02), positive (i.e. onshore-directed) or negative (i.e. offshore-directed). Overall, the trans-
 148 port forced by the mean current (q_{mean}) dominates, with a relative contribution averaged over the
 149 entire time period reaching 88%. One can observe for instance the strong tidal modulation of q
 150 during spring tides between the 10/02 and the 20/02, with a transport alternatively onshore- or
 151 offshore-directed depending on the tidal phase (Fig. 2a,d). During the first spring tides, concomi-
 152 tant with the main storm episode (between the 28/01 and the 05/02, see the grey shaded area in
 153 Fig. 2), Pezerat et al. (2022) showed that the cross-shore mean current is mostly offshore-directed
 154 associated with the prevailing contribution of the wave driven return current generating a jet like
 155 circulation with a local current intensity reaching as much as 0.25 m/s at the peak of the storm. As
 156 a result, the local time-averaged cross-shore sediment transport is mostly offshore-directed, reach-
 157 ing up to $-7.3 \text{ kg/m}^2/\text{s}$, such that flood currents are not intense-enough to reverse the orientation
 158 of the transport (Fig. 2d). For some bursts, the flux coupling contributes predominantly to the
 159 sediment transport, especially under energetic conditions (Fig. 2e). The results show a prevailing
 160 contribution of the flux coupling in the G band as compared to the IG band, associated with a
 161 low local ratio ($\simeq 10\%$) of IG wave height to short wave height (Fig 2b,e), which was found to be
 162 a pertinent proxy for the contribution of IG waves to sediment transport (de Bakker et al., 2016).
 163 Focusing on the energetic episode, the co-spectra reveal contrasting dynamics with direction re-
 164 versals within both the IG and the G bands (Fig 3a). It is worth pointing out that the order of

165 magnitude for the time-averaged sediment flux is fairly consistent with the results of the literature
 166 ($q \simeq 5 \text{ kg/m}^2/\text{s}$ Russell, 1993; Osborne and Greenwood, 1992; de Bakker et al., 2016), although
 167 previous studies typically relied on in situ measurements collected in the inter-tidal area.

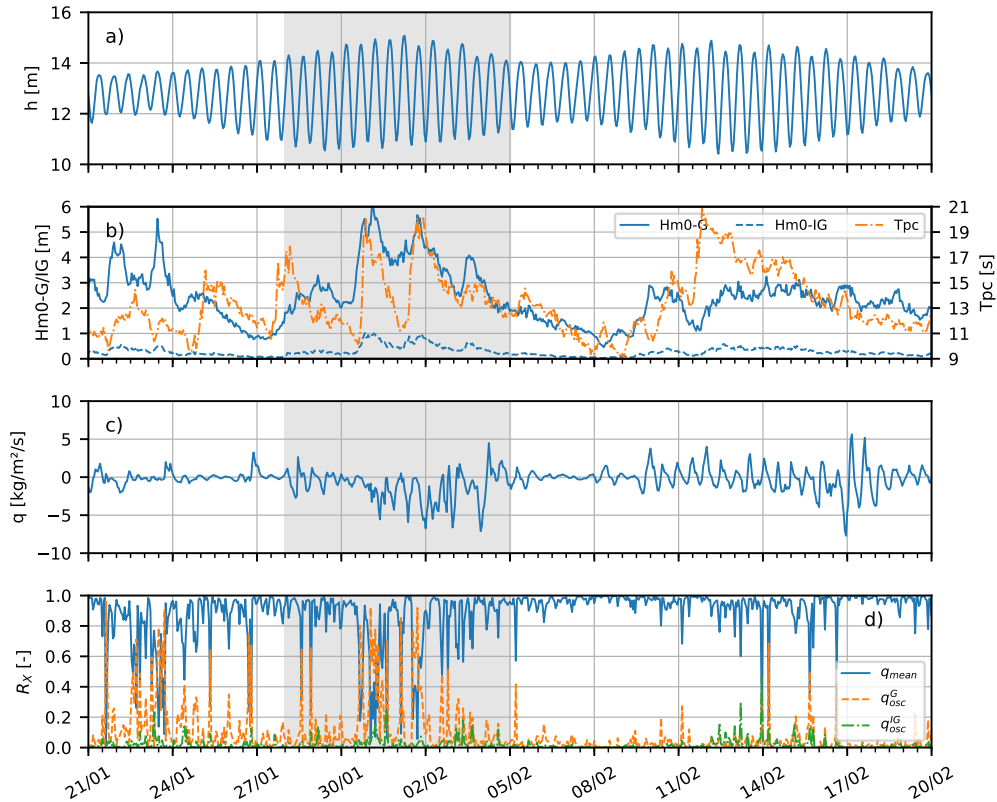


Figure 2: **a)** Burst-averaged water depth. **b)** Short waves and IG waves significant height and continuous peak period of short waves. **c)** Net time-averaged cross-shore sediment transport rate. **d)** Relative absolute contributions of each component of the transport. The grey shaded area in each panel covers the storm event studied by Pezerat et al. (2022).

168 4 Discussion

169 4.1 Suspended sediment transport by short waves

170 The contribution of short waves is further investigated regarding the phase-lag between \tilde{u} and \tilde{c}
 171 within the G band, which reads:

$$\Phi^G = \int_{f_{min,G}}^{f_{max}} \arctan \left(\frac{\Im\{S_{\tilde{u}\tilde{c}}\}}{\Re\{S_{\tilde{u}\tilde{c}}\}} \right) df \quad (7)$$

172 where $\Im\{\cdot\}$ designates the Imaginary part. The phase-lag is examined in relation to the develop-
 173 ment of the short wave skewness and asymmetry, respectively estimated through the two following

174 parameters (e.g. Elgar and Guza, 1985):

$$Sk = \frac{\overline{\tilde{u}_G^3}}{\overline{\tilde{u}_G^2}^{3/2}} \quad (8)$$

175 and,

$$As = \frac{\overline{\mathcal{H}\{\tilde{u}_G\}^3}}{\overline{\tilde{u}_G^2}^{3/2}} \quad (9)$$

176 where \tilde{u}_G is obtained by correcting \tilde{u} from the low-pass filtered signal with a cutoff frequency set
177 to $f_{min,G}$, while $\mathcal{H}\{\cdot\}$ designates the Hilbert's transform. The values of Sk are typically between
178 0 and 1 (the higher the value of Sk , the more the waves are skewed), whereas the values of As are
179 usually negative between 0 and -1.5 (the lower the value of As , the more the waves are asymmetric).
180 While the asymmetry remains relatively weak, the skewness is particularly developed at the storm
181 peaks (e.g. around the 31/01 and 02/02, see Fig. 3b). The results tend to show that the transport
182 is mostly offshore-directed when Φ^G is negative, concomitantly with high Sk values (Fig. 3a,b),
183 which is consistent with the crest-to-trough phase lag effect described by Grasso et al. (2011) from
184 their laboratory experiments. This phase lag is readily observed by comparing the timeseries of \tilde{u}
185 and c for a given burst (Fig. 3c,d).

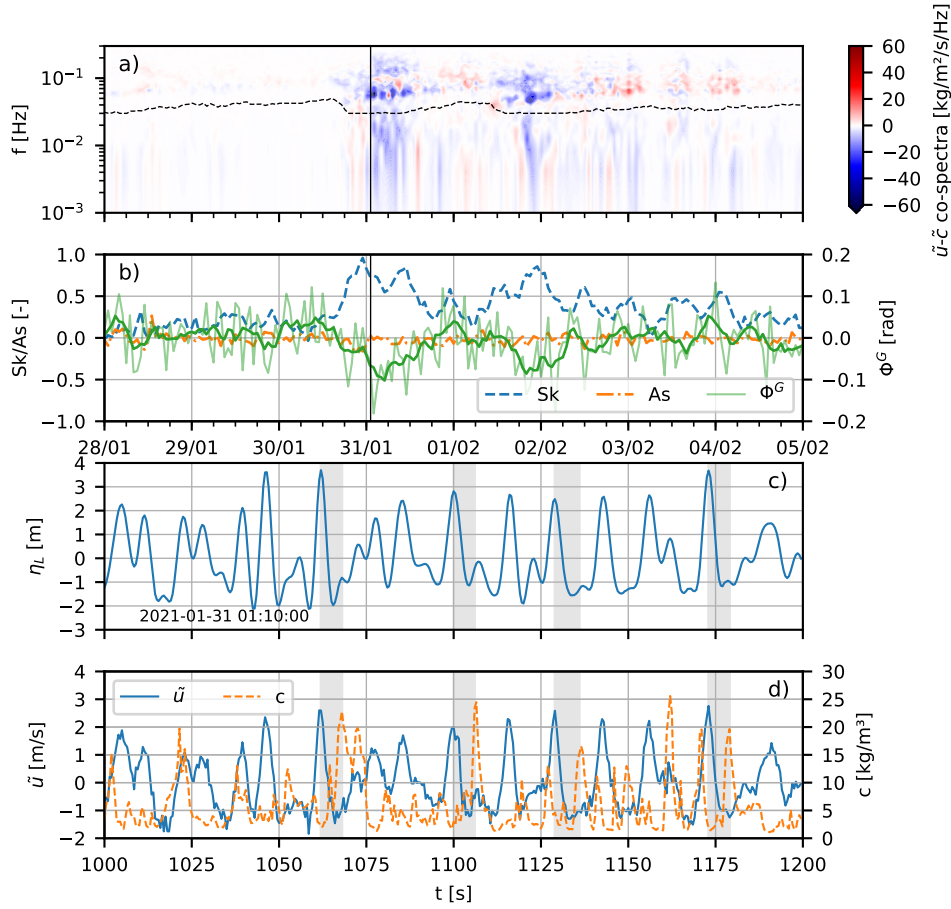


Figure 3: **a)** \tilde{u} - \tilde{c} co-spectra and separation frequency between the IG and G bands ($f_{min,G}$, see the black dashed line). **b)** Short waves skewness (Sk) and asymmetry (As) and phase lag between \tilde{u} and \tilde{c} at short wave frequencies (raw signal in light green and 6 h-running-averaged signal in dark green). **c)** Burst sample of the short waves free surface elevation signal (see the black vertical line in panels **a** and **b** for the burst considered). **d)** Burst sample of \tilde{u} and c . The grey-shaded area highlight the phase-lag between the maxima of short waves orbital velocity and the peak of SSC.

186 4.2 Suspended sediment transport by IG waves

187 Although the contribution of the flux coupling within the IG band is overall of second order, its
 188 relative contribution reaches as much as 20% at the peak of the storm (e.g. the 31/01, just before
 189 6 h, see Fig. 2e). The direction of the cross-shore transport that occurs within the IG band is
 190 further analysed. Following de Bakker et al. (2016), the correlation between the incoming IG waves
 191 and the short wave energy envelope is quantified to estimate whether the IG waves are bound to
 192 the short wave group or propagate as free waves. The low-pass filtered reconstructed free surface
 193 elevation (η_{IG}) is first partitioned into incoming (η_{IG}^+) and outgoing (η_{IG}^-) IG wave signals using
 194 the approach of Van Dongeren and Svendsen (1997) (their Eq. 4.30):

$$\begin{aligned} \eta_{IG}^+ &= \frac{\sqrt{gh}\eta_{IG} + Q}{c_g + \sqrt{gh}} \\ \eta_{IG}^- &= \frac{c_g\eta_{IG} - Q}{c_g + \sqrt{gh}} \end{aligned} \quad (10)$$

195 with:

$$Q = \tilde{u}_{IG}(h + \eta_{IG}) \quad (11)$$

196 where \tilde{u}_{IG} is obtained by low-pass filtering \tilde{u} with a cutoff frequency set to $f_{min,G}$. These equations
197 are modified versions of those given by Guza et al. (1985), extending their approach to the shoreface
198 by assuming that incoming waves travel at the speed of short wave groups and outgoing waves
199 propagate as free waves. The reflection coefficient (\mathcal{R}^2) is thus defined as the ratio between
200 the incoming and the outgoing energy. The envelope of the short wave elevation signal (A) is
201 computed following the procedure described by List (1991) and the Groupiness Factor (GF) is
202 finally introduced to quantify the modulation of the amplitude of the elevation signal corresponding
203 to the presence of groups, such as (List, 1991):

$$GF = \frac{\sqrt{2}\sigma_A}{A} \quad (12)$$

204 where σ_A is the standard deviation of A . A strong tidal modulation of the reflection coefficient
205 is observed under moderate wave energy conditions, with $\mathcal{R}^2 \lesssim 0.5$ at low tides and $\mathcal{R}^2 \gtrsim 1$ at
206 high tides (Fig. 4a,b), suggesting a total reflection at the shoreline. The values of \mathcal{R}^2 above unity
207 would indicate an increase of incoming IG waves energy between the sensor's location and the
208 shoreline. Interestingly, the values of \mathcal{R}^2 drop and the tidal modulation weakens under energetic
209 conditions (see Fig. 4b between the 30/01 and the 03/02). These results are consistent with those
210 observed by Bertin et al. (2020) during a field campaign carried out in February 2017 under similar
211 hydrodynamic conditions. The increase in \mathcal{R}^2 at high tide would be explained by a steeper slope of
212 the beach upper part, while lower values of \mathcal{R}^2 would be associated with the dissipation of IG waves
213 between the sensor and the shoreline. Bertin et al. (2020) shows indeed that, across this wide area
214 within which breaking is active, a large amount of IG wave energy could be transferred towards
215 higher frequency through non-linear triad interactions, which ultimately lead to IG wave breaking
216 in shallow depth. The results further show a negative correlation between A and the incoming IG
217 waves elevation signal associated with relatively high values of the GF ($GF \simeq 0.6$), which suggests
218 a bound wave pattern for the incoming IG waves (Fig. 4c). The strong re-suspension by short
219 waves at the scale of a group thus coincides with the trough of the bound IG wave associated
220 with an offshore-directed current (Fig. 5a,c,e). This bound wave mechanism, as conceptualised by
221 Abdelrahman and Thornton (1987) - see also Roelvink and Stive (1989) - and already reported by
222 few studies (e.g. Osborne and Greenwood, 1992; de Bakker et al., 2016), could explain the offshore-
223 directed transport within the IG band that is globally observed. Yet, the negative correlation
224 occasionally decreases along with the GF (the most notable example being observed around the
225 30/01, midnight, see Fig. 4c), which could be interpreted as the release of bound IG waves as
226 the highest short waves break. As the released IG waves propagate a bit faster than the weakly
227 dispersive short waves, the crest of the IG waves would be in phase with the partially remaining
228 groups (Fig. 5b,d,e), which would result in inverting the former mechanism.

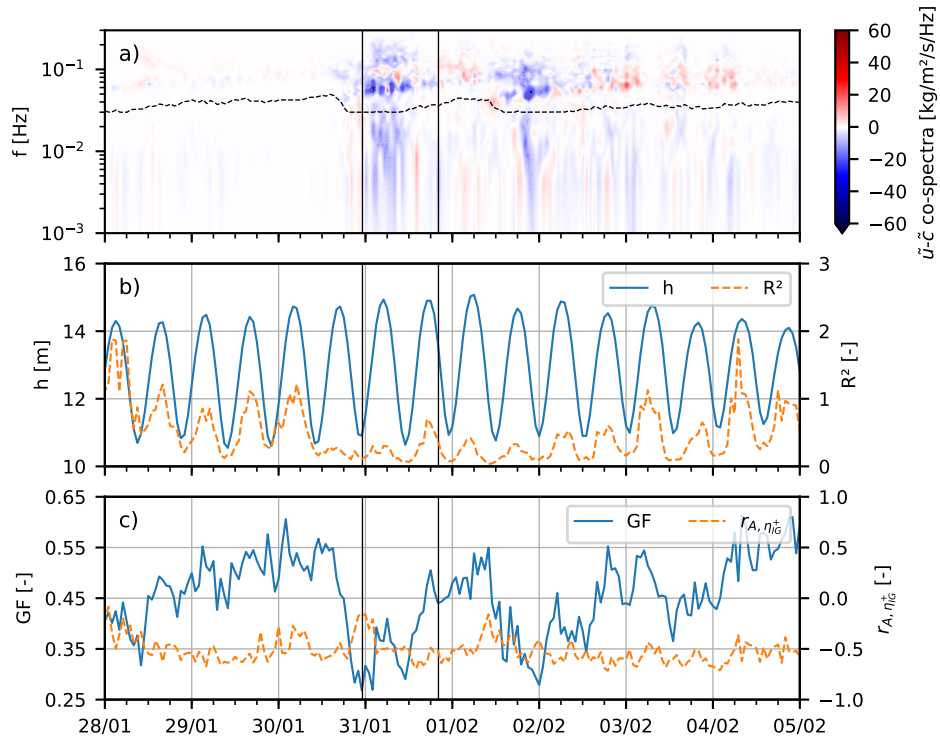


Figure 4: **a)** $\tilde{u}\text{-}\tilde{c}$ co-spectra and separation frequency between the IG and G bands ($f_{min,G}$, see the black dashed line). **b)** Burst-averaged water depth and squared reflection coefficient. **c)** Groupiness Factor and Pearson correlation coefficient with a confidence level of 95% between the envelope of short waves elevation signal and the incoming IG waves elevation signal.

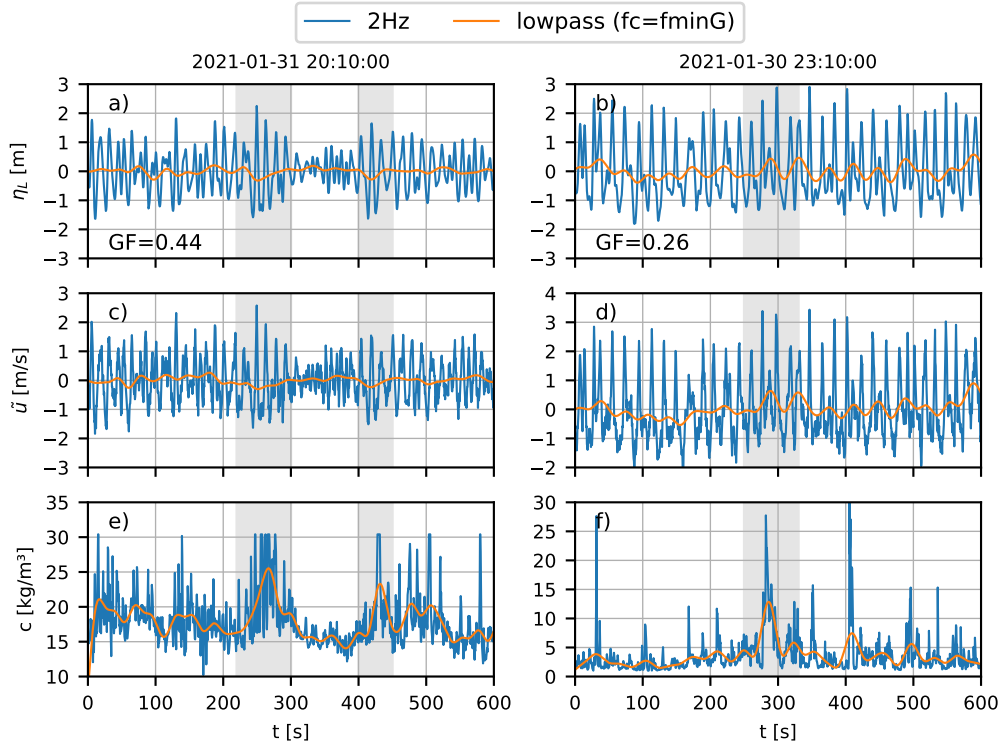


Figure 5: **a)** and **b)** Burst samples of the free surface elevation and associated IG waves free surface elevation for two bursts (see the vertical black lines in Fig. 4). **c)** and **d)** Burst samples of the oscillating component of the cross-shore velocity and associated IG waves signal. **e)** and **f)** Burst samples of SSC signal and associated signal within the IG band. The panels on the left illustrate the bound wave mechanism, whereas those on the right are representative of the transport mechanism with released incoming IG waves (see in particular the grey shaded area).

5 Conclusion

229 This study first shows that steady currents dominate suspended sediment transport within the
 230 upper shoreface. While tidal currents modulate the orientation of the transport under fair weather
 231 conditions, the wave driven return current is shown to force an offshore-directed sediment transport
 232 at the storm peak at the measurement location, as far as 4 km from the shoreline. Cross-spectral
 233 analysis of the oscillating transport further shows the key contribution of short wave motions
 234 especially under energetic conditions, which globally induce an offshore directed transport resulting
 235 from a crest-to-trough phase lag effect. Finally the relatively weak contribution of IG waves to
 236 suspended transport appears correctly described by the conceptual model proposed by de Bakker
 237 et al. (2016) with a transport mostly offshore directed, associated with incoming IG waves bound to
 238 the wave groups. Although based on measurements vertically not co-located, from a single location,
 239 these results highlight the processes governing the cross-shore sediment transport beyond the inner
 240 surf zone, and further advocate for a comprehensive description of the wave-driven dynamics all
 241 across the nearshore area. It thus appears crucial to repeat similar field campaigns, extending
 242 measurement locations across the shoreface and the surf zone in contrasted environment to further
 243 investigate the influence of the local bottom slope or the granulometry. Finally, these results also
 244

245 question the morphodynamic modelling of nearshore areas (including the shoreface and the surf
246 zone) when using phase-averaged wave model where IG waves are not represented and phase lag
247 effects have to be parameterized.

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256 and suggestions.

257 **A OBS calibration**

258 The calibration of the OBS followed the protocol detailed in Coulombier et al. (2012). Five series
259 of measurements were here carried out in a 70 L tank in which the sediment collected at the study
260 site was progressively added and maintained in suspension using a stirrer and a pump. Prior to each
261 addition of sediment, a 500 mL sample was collected to measure the concentration corresponding
262 to the $n - th$ series of measurements after filtration and drying of the sediment contained in the
263 sample. Several measurements were performed after addition of sediment so as to control that the
264 system reached a steady state. The final OBS measurements of each series were then fitted to the
265 concentration measured using a simple linear regression with a good agreement ($r^2 = 0.95$, see
266 Figure 6), such that the fitted model could be applied to the OBS measurements collected during
267 the field campaign.

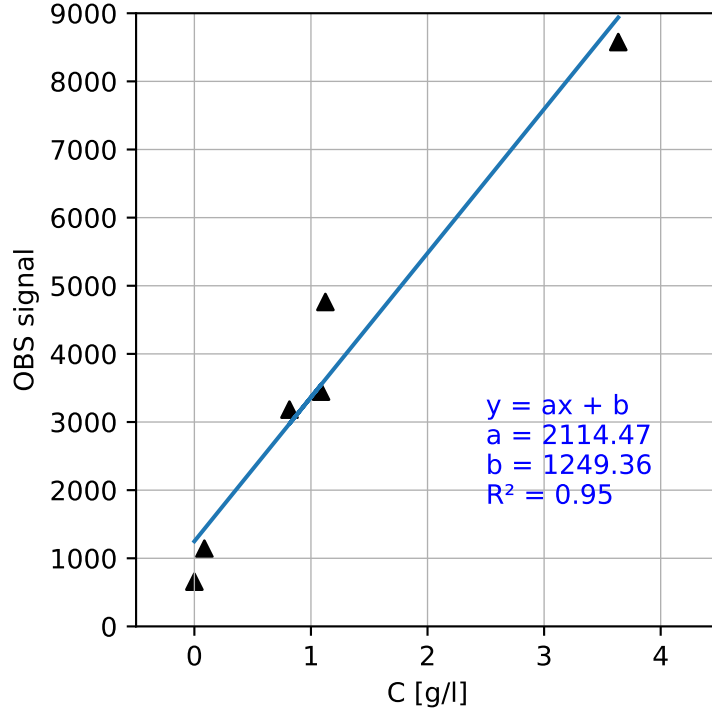


Figure 6: Linear regression resulting from the calibration procedure of the OBS sensor.

268 B Vertical decrease of short wave orbital velocities

269 The vertical decrease of short wave orbital velocities can be estimated from the depth-varying
 270 expression of orbital velocities (u_{orb} , in m/s) based on the linear wave theory:

$$u_{orb}(z) = \sigma a \frac{\cosh(k(z+d))}{\sinh(kh)} \quad (\text{B.1})$$

271 where σ is the relative pulsation (in rad/s), a is the wave amplitude (in m), k is the wavenumber
 272 (in m^{-1}) given by the dispersion relation reading $\sigma^2 = gk \tanh(kh)$, with $g = 9.81 \text{ m/s}^2$, the
 273 gravitational acceleration, and z is the vertical coordinates ranging from $-d$ at the bottom to $\bar{\eta}_h$
 274 at the free surface, such that the local mean water-depth reads $h = d + \bar{\eta}_h$. Considering a relative
 275 pulsation based on the continuous peak period representative of the sea-state (i.e. $\sigma = 2\pi/T_{pc}$), the
 276 relative decrease of short wave orbital velocities from the wave cell to the OBS vertical locations
 277 thus reads:

$$\chi_{u_{orb}} = \frac{u_{orb}(z = z_{wc}) - u_{orb}(z = z_{OBS})}{u_{orb}(z = z_{wc})} = 1 - \frac{\cosh(k(z_{OBS} + d))}{\cosh(k(z_{wc} + d))} \quad (\text{B.2})$$

278 The result of the computation carried out over the entire studied period is presented Figure 7.

279 C Prevailing transport regime for fine sands

280 The dominance of suspended load over bedload transport usually occurs when the skin friction
 281 velocity exceeds the sediment settling velocity (Soulsby, 1997). As such, the maximum skin friction
 282 velocity under combined waves and currents (u_{max}^* , in m/s) is compared with the settling velocity

283 (W_s , in m/s) considering a grain size $D = 200 \mu\text{m}$ representative of the fine sand fraction of the
 284 sediment (see Figure 7). The mean (τ_m) and maximum (τ_{max}) values of the skin friction bed shear
 285 stress over a wave cycle (in Pa) were first computed by means of the following formula (Soulsby,
 286 1997, Eq. 69 and 70):

$$\begin{aligned}\tau_m &= \tau_c \left(1 + 1.2 \left(\frac{\tau_w}{\tau_c + \tau_w} \right)^{3.2} \right) \\ \tau_{max} &= \sqrt{(\tau_m + \tau_w \cos \phi)^2 + (\tau_w \sin \phi)^2}\end{aligned}\tag{C.1}$$

287 where τ_w is the wave-only peak bed shear-stress, τ_c is the current-only bed shear stress and ϕ is
 288 the angle between current direction and mean wave direction, which was readily estimated through
 289 the cross-spectral analysis of orbital velocities, while classical expressions were used for τ_w and τ_c
 290 with a bed roughness length (z_0) set to $D/12$ (see Example 5.1 of Soulsby, 1997):

$$\tau_c = \rho C_d \bar{u}^2\tag{C.2}$$

291

$$\tau_w = \frac{1}{2} \rho f_w U_{orb}\tag{C.3}$$

292 with C_d , the drag coefficient and f_w , the wave friction factor, that respectively read:

$$C_d = \left(\frac{\kappa}{1 + \log \left(\frac{z_0}{h} \right)} \right)^2\tag{C.4}$$

293

$$f_w = 1.39 \left(\frac{A_{orb}}{z_0} \right)^{-0.52}\tag{C.5}$$

294 where $\kappa \simeq 0.41$ is the von kármán constant and U_{orb} , A_{orb} are the local estimates of the orbital
 295 velocity and semi-orbital excursion at the bottom level at peak frequency:

$$U_{orb} = \frac{\pi}{\sqrt{2}} \frac{H_{m0}}{T_{pc} \sinh(kh)}\tag{C.6}$$

296

$$A_{orb} = \frac{U_{orb} T_{pc}}{2\pi}\tag{C.7}$$

297 The maximum skin friction velocity thus reads:

$$u_{max}^* = \sqrt{\frac{\tau_{max}}{\rho}}\tag{C.8}$$

298 Second, the settling velocity for the single class sediment was given by (Soulsby, 1997, Eq. 102):

$$W_s = \frac{\nu}{D} \left((10.36^2 + 1.049 D_*^3)^{1/2} - 10.36 \right)\tag{C.9}$$

299 where $\nu \simeq 1.36 \cdot 10^{-6} \text{m}^2/\text{s}$ is the sea-water kinematic viscosity and D_* is the dimensionless grain
 300 size (see Soulsby, 1997, Eq. 98).

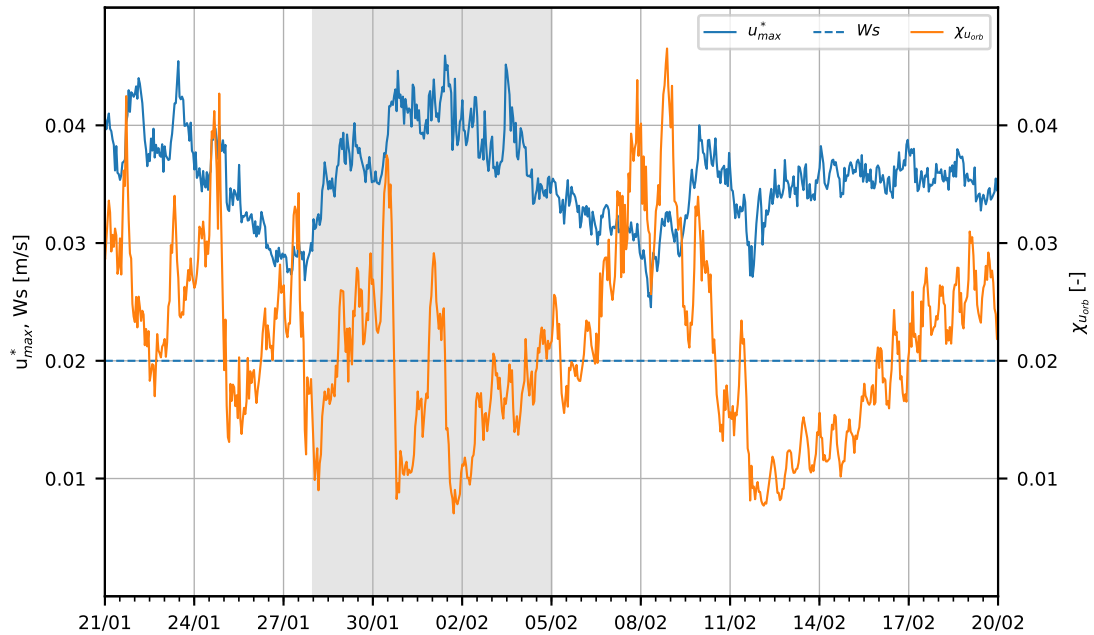


Figure 7: Skin friction velocity versus settling velocity of fine sand and relative short wave orbital velocities decrease between the wave cell and the OBS vertical locations.

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