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► **To cite this version:**

Alphonse Nahon, Déborah Idier, X. Bertin, Thomas Guérin, Vincent Marieu, et al.. Modelling the contribution of wind waves to Cap Ferret's updrift erosion. *Coastal Engineering*, 2022, 172, pp.104063. 10.1016/j.coastaleng.2021.104063 . hal-03461439

**HAL Id: hal-03461439**

**<https://univ-rochelle.hal.science/hal-03461439>**

Submitted on 1 Dec 2021

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## **Modelling the contribution of wind waves to Cap Ferret's updrift erosion**

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### **Highlights**

- Residual sediment transport is modelled at a mixed energy tidal inlet.
- Waves with tides create transport gradients near the inlet's updrift spit extremity.
- Larger or more oblique waves deepen this gradient and reinforce the sand deficit.
- Physically sounded explanation to enhanced spit erosion during NAO+ winters.
- Bed roughness models need to be improved for 2DH inlet-spit morphodynamic modelling.

### **Keywords**

Sandspit; Tidal inlet; Sediment transport; Wave power; NAO; SCHISM.

# 1 **Modelling the contribution of wind waves to Cap Ferret's updrift erosion**

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## 13 **Abstract**

14 Wind waves breaking at an angle with the shoreline force the drifting of littoral sediments, which is  
15 known for contributing to the formation and growth of barrier spits. Intriguingly, increased rates of  
16 longshore wave power have also been associated with the erosion of some barrier spits on the  
17 updrift margin of tidal inlets. Therefore, a numerical experiment was designed and is presented here,  
18 which investigates the possible links between the longshore wave power and the shortening of these  
19 elongated coastal barriers. Based on a process-based model, this experiment provides new insights  
20 into the forces at play in the redistribution of sediments between a sandspit and its adjacent inlet,  
21 respectively the Cap Ferret and the Bay of Arcachon's tidal inlet, in SW France. More particularly,  
22 model scenarios were defined that show how combined waves and tide create gradients of residual  
23 sediment transport responsible for a sediment deficit at the spit – inlet boundary. The deficit was  
24 also found to deepen with increasing longshore wave energy, as if the transfer of sediment from the  
25 spit to inlet shoals was accelerated. This physically explains the previously observed retreat of the

26 spit's distal end during periods dominated by the positive phase of North Atlantic Oscillation (NAO) in  
27 winter. Indeed, according to model results, higher and/or more oblique waves associated with the  
28 positive phase of the NAO are expected to increase the transfer and storage of the drifting sediments  
29 to and by the inlet shoals, and this at the expense of the spit. While these conclusions remain valid,  
30 we noticed that the sensitivity of model results to the parametrized bottom friction enhanced the  
31 importance of accurately representing the spatio-temporal distribution of bed roughness when  
32 investigating the morphodynamic interactions between real-world tidal inlets and their margins.

### 33 **Highlights**

- 34 • Residual sediment transport is modelled at a mixed energy tidal inlet.
- 35 • Waves with tides create transport gradients near the inlet's updrift spit extremity.
- 36 • Larger or more oblique waves deepen this gradient and reinforce the sand deficit.
- 37 • Physically sounded explanation to enhanced spit erosion during NAO+ winters.
- 38 • Bed roughness models need to be improved for 2DH inlet-spit morphodynamic modelling.

### 39 **Keywords**

40 Sandspit; Tidal inlet; Sediment transport; Wave power; NAO; SCHISM.

## 41 **1. Introduction**

42 Interactions of sandy and gravelly barriers with tidal inlets play a key role in the behaviour and  
43 resilience of coastal barrier systems (Nienhuis and Lorenzo-Trueba, 2019; Oertel, 1985). Empirical  
44 theories have long been supporting that the waves work in favour of the barriers, while the tides  
45 maintain the inlets open (Davis and FitzGerald, 2004; Hayes, 1979; Nichols and Allen, 1981). More  
46 quantitatively, the wave-driven longshore sediment transport (LST) is the main contribution to spit  
47 growth in tidal inlet reservoir models (Hoan et al., 2011; Kraus, 2000; Larson et al., 2007). This is also  
48 supported by process-based morphodynamic modelling of idealized and migrating tidal inlets  
49 (Nienhuis and Ashton, 2016). At the same time, increased longshore wave energy is thought to have  
50 contributed to barrier erosion and shortening at the entrance of two mixed-energy barrier systems in  
51 Europe. The first example is Skallingen barrier spit, along the Danish North Sea. The distal end of the  
52 spit is bounded by a tidal inlet and Aagaard et al. (2004) have reported a shift in the wind regime,  
53 between 1970 and 1999, which has increased the rate of wave-driven LST. Aagaard and Sørensen  
54 (2013) further quantified this increase as the main contributor to the updrift erosion of the barrier  
55 spit's distal end. The second example is Cap Ferret barrier spit. Along the SW Atlantic coast of France,  
56 this sandspit dips into the tidal inlet of the Bay of Arcachon. According to a 250-year long  
57 geomorphological record, higher and more oblique waves associated with the predominance of the  
58 positive phase of the North Atlantic Oscillation (NAO) in winter can trigger the updrift erosion of this  
59 sandspit (Nahon et al., 2019). These observations recall that the processes underlying the  
60 relationship between LST and the edification of barrier spits could be more subtle than generally  
61 assumed. Thus, process-based sediment transport models can help clarifying the role of wave driven  
62 LST at the inlet – spit boundary.

63 Applied to tidal inlets, process-based models have brought a wealth of physically sounded insights  
64 into the interaction between waves, tides, and morphology. These models may be fully three  
65 dimensional (Bertin et al., 2020) or integrated in the vertical (i.e., 2DH). Nowadays, 2DH models have

66 reached a level of maturity allowing their use to, for instance, investigate the role of non-linear  
67 interactions between wave and tidal forces in the redistribution of sediments between the updrift  
68 and downdrift margins of tidal inlets (Herrling and Winter, 2018). Also of particular interest for the  
69 present study, Bertin et al. (2009) and Hansen et al. (2013) have detailed how barotropic pressure  
70 gradients and wave refraction over ebb-tidal shoals affect longshore circulation near tidal inlets, and,  
71 in the case of Bertin et al. (2009), how sediment transport is impacted. Before that, Cayocca's  
72 pioneer works (2001, 1996) had shown how waves speed up the formation of the (tidal) channel that  
73 bounds the Cap Ferret. This importance of waves in the development of tidal channels was recently  
74 confirmed by Lenstra et al. (2019). To build on these results, the present study involved a new  
75 implementation of a fully coupled circulation, waves, and sediment transport modelling system to  
76 the Bay of Arcachon's tidal inlet. To investigate the role that increased longshore wave energy may  
77 play in Cap Ferret's updrift erosion, the model was forced using a regional wave hindcast to derive  
78 scenarios representative of the local variability of the wave climate at this mixed-energy location.

79 The next section presents in more detail the study area and its wave climate since the second half of  
80 the 20th century. Then, section 3 presents the implementation of the modelling system as well as the  
81 simulation scenarios, and further describes how the modelled sediment transport was processed and  
82 analysed. In section 4, model results are presented in terms of residual longshore sediment transport  
83 and sedimentation patterns. This leads to the subsequent discussion of the role of waves and wave  
84 climate variability in the erosion of barrier spit bounded by tidal inlets. The question of the  
85 parameterization of bottom friction is also discussed as it appeared to largely influence the rates of  
86 sediment exchange between the spit and the inlet shoals.

## 87 **2. Study area and local wave climate**

### 88 *2.1. Study area*

89 The barrier system of the Bay of Arcachon (**Figure 1**) lies in the highly infilled incised-valley segment  
90 of the Leyre river's estuary (Allard et al., 2009; Féniès and Lericolais, 2005). On the updrift margin of

91 the estuary, NW dominant waves generate an estimated  $661 \times 10^3 \text{ m}^3$  net annual littoral drift (Idier et  
92 al., 2013). Southward drifting littoral sediments have progressively edified the Cap Ferret sandy spit  
93 which now semi-encloses a mesotidal lagoon and dips into a tidal inlet exposed to a mixed-energy  
94 environment (annual mean significant wave height of 1.68 m and mean spring tidal range of 3.80 m;  
95 Nahon, 2018). According to navigation charts, the inlet has stopped its southward migration around  
96 1900 (Nahon, 2018). Nonetheless, channels and bars continue to swept the inlet ebb-tidal delta in  
97 the downdrift –north to south— direction (Capo et al., 2014; Cayocca, 2001). This sustained  
98 southward displacement contrasts with the large scale north-south oscillations the distal end of the  
99 spit has displayed over the last 250 years (Nahon, 2018). So, the spit has eroded updrift at some  
100 point, and the last updrift erosion (**Figure 1**, lower panels) began as the longshore wave power  
101 sharply increased in the early 1970s (**Figure 2**, upper panel). In the same time, the enlargement of  
102 the adjacent inlet only began after 1980 (Nahon, 2018). This enlargement of the inlet could be the  
103 results of an increasing tidal prism, driven by higher rates of SLR, which also leads to larger  
104 equilibrium volumes of the inlet’s flood- and ebb-tidal deltas (Walton and Adams, 1976). Therefore,  
105 it was hypothesized that the spit oscillations were a combined response to the variations of the  
106 longshore wave power, associated with the phase (negative or positive) of the North Atlantic  
107 Oscillation (NAO; Hurrell & Deser, 2009), and to the variations in the rate of SLR (Nahon et al., 2019).

## 108 *2.2. Wave hindcast*

109 Across the last spit-end oscillation (**Figure 1**, lower panels), the local wave climate was hindcasted  
110 (from 1949 to 2014) using the storm surge modelling system of Bertin et al. (2015). The model was  
111 extended to the whole North Atlantic Ocean as described in Arnoux et al. (2018) and was forced with  
112 wind fields originating from the NCEP/NCAR reanalysis (Kalnay et al., 1996). Modelled wave  
113 parameters were previously validated against directional wave buoy measurements. **Figure 3** shows  
114 the model – data comparison in terms of longshore wave power, and the position of the wave buoy,  
115 15 km offshore of the Cap Ferret and in 54 m water depth, is indicated in **Figure 4** (left panel, W).

116 Using a total of 5.3 years of record between 2007 and 2014, Nahon et al. (2019) indicated that the  
117 averaged wave power (WP) and the longshore wave power (WPy) were underestimated by 2.4% and  
118 1.7% respectively. Also, at the observation sampling frequency (i.e., one observation every 30 min),  
119 linear correlation coefficients between modelled and measured values were of 0.93 and 0.81 for WP  
120 and WPy. When considering 90-day running mean of WP and WPy, these coefficients grow to 0.99  
121 and 0.95, respectively.

122 Following Charles et al. (2012), this hindcast was used to further highlight the apparent relationship  
123 between the North Atlantic Oscillation (NAO) and the local wave climate in winter (December to  
124 March; DJFM). During the positive phase of the NAO, the significant wave height ( $H_s$ ) increases, and  
125 the mean direction of wave incidence deviates clockwise from shore normal. As a result, winter  
126 averaged WPy is positively correlated with the NAO index. To illustrate this correlation, upper panel  
127 on **Figure 2** shows the decadal average of the winter NAO index (station-based index; Hurrell, 2015)  
128 and of the winter WPy since 1950. From winter 1950 to winter 1972, the decreasing trend in the  
129 decadal average of the winter NAO index indicates the negative phase of the NAO has prevailed.  
130 By opposition, the positive phase has dominated from 1972 to mid-1990s, as revealed by the  
131 ascending trend of the averaged index. Also, over the period 1950-1972, the average of winter WPy  
132 was equal to 89.27 % of its 1950-2014 average value, when over 1973-1995 it reaches 112.49 % of it.  
133 Those two 23-year periods respectively cover the last phase of Cap Ferret elongation (1950-1972)  
134 and the onset of the rapid spit retreat around 1972-1973. Therefore, average WPy over these periods  
135 were used to derive different scenarios to force the sediment transport model, which are presented  
136 in section 3.2.

### 137 **3. Modelling methodology**

#### 138 *3.1. Model implementation and validation*

139 The modelling system SCHISM (Zhang et al., 2016) was used to simulate hydrodynamics and  
140 sediment transport around the spit distal end. **Figure 4** shows the computational domain and the



141 model bathymetry. SCHISM's circulation model was used in 2DH mode, fully coupled to the spectral  
142 wave model WWM-II (Roland et al., 2012). Both models' equations are solved over the same  
143 unstructured grid; they were run in parallel and used the same domain decomposition. The grid  
144 resolution ranges from 2 km at the open boundary to 60 m within the inlet and lagoon; along the last  
145 6 km of spit's ocean side, the resolution reaches 20 m which is expected to be fine enough to  
146 generate wave induced circulation at this type of beaches (e.g., Bruneau et al., 2014, used a 15 m  
147 resolution grid at the nearby Biscarosse Beach for waves less than half the size of those in the  
148 present study).

149 In WWM-II, the wave energy spectra was discretized over a 24 x 24 grid, spanning directions from 0°  
150 to 360° and frequencies from 0.04 Hz to 0.4 Hz; wave breaking was parameterized according to  
151 Battjes and Janssen (1978), with a breaking criteria equal to 0.78. The hydrodynamic timestep was  
152 set to 1 min and the sediment transport fluxes were computed with Camenen & Larson (2007; Larson  
153 et al., 2011) formula developed for tidal inlet applications. A relevant characteristic of this formula is  
154 to account for the wave and current colinear interactions. To do so, bedload and suspended load  
155 transport rates in the wave propagation direction are respectively computed with the "net sediment  
156 transporting velocity" and the "net mean current" (Camenen and Larson, 2007; equations 226, 227  
157 and 231 for bedload and equations 232 and 241 for suspended load). Both quantities are deduced  
158 from the root-mean-square total of the velocity over a wave period. This total velocity is defined as  
159 the wave orbital velocity plus the component of the ambient current aligned with the wave  
160 propagation direction. As a result, if waves and current are not perpendicular, waves add to the net  
161 transport (either onshore or offshore) in the direction of wave propagation. This, independently of  
162 the asymmetric character of the waves which was not considered here.

163 In the circulation model, the bottom friction was parametrized with a Manning formulation. To avoid  
164 the modulation of modelled sediment transport patterns by a spatially varying Manning coefficient,  
165 the choice was made to use a spatially uniform coefficient, equal to  $0.032 \text{ s/m}^{1/3}$ . This value falls

166 within the range of values found in the literature for tidal inlets (Bruneau et al., 2011; Orescanin et  
167 al., 2016) and was set after calibration tests performed with tidal forcing only. For the tests, the  
168 model's open boundary was forced with 16 tidal components from the regional tide model of Bertin  
169 et al. (2012). The model's performances were evaluated in terms of modelled elevation at three  
170 locations across the tidal inlet (**Figure 4**, A, B and C), compared with observations from June 2014.  
171 From the west to the east, these observations include: a 28-day long record from a bottom moored  
172 pressure sensor located on the terminal lobe of the ebb-tidal delta (A; Senechal et al., 2013), a 14.8-  
173 day record from a bottom moored pressure sensor located within the inlet Southern channel (B;  
174 Doré, 2015) and a 28-day extraction from *Eyrac's* tide gauge record (C; SHOM). Root mean square  
175 errors of modelled elevation ranged from 7.35 cm at the inlet entrance to 11.84 cm in the lagoon  
176 with biases on the order of a few centimetres. These values are summarized in **Table 1** with  
177 additional details on the evolution M2 and M4 tidal components across the inlet. The evolution of  
178 these components suggests the model qualitatively reproduces the propagation of the tidal wave  
179 across the inlet, although tidal asymmetry and distortion may be overestimated locally as the  
180 comparison at location B reveals.

### 181 *3.2. Output processing*

182 The model was then used to assess the impacts of waves and of winter wave climate variability on  
183 the residual sediment transport near the distal end of the spit. This was carried on with tidal forcing  
184 reduced to a monochromatic tide, represented only with local M2 component imposed along the  
185 model's open boundary. The tidal amplitude was set following Cayocca (2001, 1996) who found that  
186 residual sediment transport patterns are better approached with a single M2 constituent of  
187 amplitude 1.8 m (i.e., 3.6 m tidal range). After a one-day simulation spin up, sediment fluxes sampled  
188 at 10-min intervals were averaged over 2 tidal cycles and converted into annual rates of residual  
189 sediment transport. Residual fluxes were then interpolated onto a regular grid with a thinner  
190 resolution than the unstructured grid. This regular grid served to compute the divergence of the

191 residual fluxes and the divergence was spatially integrated to estimate the annual sedimentation  
192 rates over an array of 58 alongshore-overlapping cells (**Figure 5**, centre panel). Each cell is 300-meter  
193 wide in the north-south direction. In this same direction, centres of two neighbouring cells are  
194 spaced by 100 m which creates a 67% overlap between cells. To discuss the integrated sedimentation  
195 rate, the average southward sediment transport was computed for each cell. This quantity was  
196 calculated by integrating in the west-east direction the north-to-south component of the residual  
197 transport and then averaging of the integrated values over the north-south extension of each cell.

198 Three groups of cells were defined and are delimited on **Figure 5** centre panel: a first group away  
199 from the inlet (dashed black frame, delimiting cells number 5 to 10), a second group on the western  
200 face of the spit's last kilometre (dashed red frame, delimiting cells number 45 to 53, hereafter  
201 referred to as the western group) and a last group at the aerial spit – inlet boundary (dashed blue  
202 frame, delimiting cells number 55 to 57, hereafter referred to as southern group). Western and  
203 southern groups overlap by 100 m in the north-south direction and were used to present and discuss  
204 the modelled patterns of residual sediment transport and sedimentation rates along the last  
205 kilometre of the subaerial portion of the spit.

206 To give a broader perspective on the sediment budget, the sedimentation rate over the spit's  
207 subtidal platform was also estimated for simulations with both waves and tide (SIM3-8, see section  
208 2.5). As for the array of cells, the annual sedimentation rate was computed by integrating the  
209 divergence of the sediment fluxes. The right panel in **Figure 4** shows the area (delimited in blue) over  
210 which the spit-platform sedimentation rate was computed. This area overlaps with the array of cells  
211 after cell number 39. Furthermore, the sedimentation rate is integrated only in zones shallower than  
212 -7 m NGF which distinguishes inlet shoals and channels (black contour on **Figure 4**, right panel).

### 213 *3.3. Simulation scenarios*

214 Residual transport rates and sedimentation estimations were computed for a first set of 3  
215 simulations that uncovered the respective and combined roles of waves and tide; the model was run

216 with M2 tide forcing only (SIM1), with waves and no tides (SIM2) and then with waves and M2 tide  
 217 (SIM3). A second set of 5 simulations were run to assess the sensitivity of model results to changes in  
 218 the winter wave climate described in the section 2.2. **Table 2** summarizes wave climate parameters  
 219 for the 8 simulations.

220 In SIM1 and SIM3 parameters were representative of the annual average wave power hindcasted  
 221 between 1949 and 2014. The wave direction (Dir) was the one of the average incoming wave power;  
 222 the wave peak period ( $T_p$ ) and the significant wave height ( $H_s$ ) were the average values (over 1949-  
 223 2014) multiplied by  $\alpha = 1.095$  and  $\beta = 1.095^2$ , respectively. These coefficients were defined as in Eq. 1  
 224 to Eq. 5, where representative significant wave height ( $H_{s,rep}$ ) and peak period ( $T_{p,rep}$ ) are defined so  
 225 that, along the model's open boundary, the representative wave energy ( $E_{rep}$ ) and the  
 226 representative group velocity ( $c_{g,rep}$ ) match the average wave power  $\overline{WP}$  over a given period of time  
 227 T:

$$\overline{WP} = \frac{1}{T} \int_0^T WP(t) dt \quad \text{Eq. 1}$$

$$\overline{WP} = E_{rep} c_{g,rep} \quad \text{Eq. 2}$$

$$c_{g,rep}, \text{ computed with } T_{p,rep} = \alpha \overline{T_p} \quad \text{Eq. 3}$$

$$E_{rep} = \frac{1}{8} g \rho H_{s,rep}^2 \quad \text{Eq. 4}$$

$$H_{s,rep} = \beta \overline{H_s}, \text{ with } \beta = \alpha^2 \quad \text{Eq. 5}$$

228

229 Winter wave climates were defined in similar fashion, with December to March averages instead of  
 230 annual averages. SIM4 was representative of all 65 complete winters between December 1949 and  
 231 March 2014, SIM5 was representative of the low energy winters between 1950 and 1972, and finally  
 232 SIM6 is representative of the high energy winters between 1973 and 1995. The last two simulations  
 233 (SIM7 and SIM8) were set to further infer on the role of the wave angle of incidence. As depicted by

234 **Figure 2**'s lower panel, higher values of winter NAO indices shift clockwise the incoming wave mean  
235 direction, producing higher wave angle of incidence. SIM7 and SIM8 were set up with  $\pm 4^\circ$  shifted Dir  
236 to mimic this influence of negative and positive phases of the NAO.

237 Additionally, tests were made to evaluate the impact of tidal range (**Table 3**). To this end, simulations  
238 SIM4 to SIM8 were reproduced with tidal ranges of 1.8 m and 4.5 m (**Table 3**); 1.8 m corresponding  
239 to the mean neap tidal range at the entrance of the Bay and 4.5 m being greater than the mean  
240 spring tidal range (of 3.8 m) but smaller than the maximum astronomical tidal range of 4.9 m (SHOM,  
241 2014). Later in the discussion, results from four ultimate simulations (SIM3a,b,c and d; **Table 3**) will  
242 be presented, that highlights the limits of the current models of spatio-temporally variable bed  
243 roughness, at least for the present inlet – spit morphodynamic interactions.

## 244 **4. Results**

### 245 *4.1. Waves, tides, and residual sediment transport*

246 **Figure 6** shows the modelled sediment transport and sedimentation patterns along the last 6 km of  
247 the subaerial sandspit. In terms of residual sand transport (upper panel), the southward component  
248 of the transport denotes that the LST was insignificant until along the spit's last kilometre in the tide-  
249 only case (SIM1, grey curve). Then, away from the spit-end (cells > 55) the southward component  
250 became almost equal to that of the wave induced transport (SIM2, red curve). When waves were the  
251 only forcing, the updrift, or incoming, rate LST was in the order of  $150 \times 10^3 \text{ m}^3/\text{year}$  and the  
252 southward transport steadily increased with the curvature of the spit. The transport peaked within  
253 the western group of cells (in cell number 51) before it was halved in the southern group. Finally,  
254 when both average waves and M2 tide forced the circulation (SIM3, blue curve), the residual  
255 southward transport mimicked that of the wave-only simulation until cell number 50 and then  
256 increased to be multiplied by a factor of two in cell number 58.

257 The sedimentation rates associated with those transport patterns were then estimated by  
258 integrating the divergence of the residual sediment transport (middle panel, **Figure 6**). With tidal  
259 forcing only, little sediment accretes or erodes before the very end of the spit, and the average  
260 accretion was of  $6 \times 10^3 \text{ m}^3/\text{year}$  per cell in the western group (**Table 2**). In the wave-only case, this  
261 value grew to  $186 \times 10^3 \text{ m}^3/\text{year}$  per cell. Contrastingly, combined average waves with the  
262 representative M2 tide produced an erosion of  $51 \times 10^3 \text{ m}^3/\text{year}$  per cell on average. Sedimentation  
263 rates were less contrasted in the southern group. There, in all three cases, erosion was predicted  
264 immediately after the spit extremity. Still, the estimated erosion was much smaller and would  
265 happen in a narrower area in SM1 and SIM2 compared to SIM3. Indeed, with combined waves and  
266 tide the estimated erosion in the southern group was of  $312 \times 10^3 \text{ m}^3/\text{year}$  per cell on average,  
267 compared to  $95 \times 10^3 \text{ m}^3/\text{year}$ , in the wave-only case.

268 When both waves and tide were considered SIM3, the lower panel in **Figure 6** further displays  
269 alongside the contribution of the southward (NS) component and of the eastward (WE) component  
270 of the sediment transport to the overall sedimentation rates. It indicates the southward transport  
271 would generate mostly erosion (filled-in blue bars), as for instance in the southern group. By  
272 opposition, the eastward transport would promote accretion, which mostly fails to compensate the  
273 erosion produced by the gradients in the southward transport (red bars).

274 Lastly, the black curve on the upper panel of **Figure 6** also shows the southward residual sediment  
275 transport in the case of average winter waves, combined with the representative M2 tide (SIM4).  
276 Compared to SIM3, the incoming rate of LST was multiplied by a factor 1.83 (**Table 2**). Then, the  
277 southward transport's curve presents a similar shape which denotes a similar acceleration of the  
278 transport. Still, the overall acceleration was less pronounced as the increase at the spit – inlet  
279 boundary was only about a factor of 1.25 compared to SIM3 (in cell number 58). In the same time,  
280 the estimated erosion was reduced in the western group and reinforced in the southern group  
281 (**Figure 7**, middle panel, black bars).

282 *4.2. Winter wave climate and tidal range modulations*

283 **Figure 7** shows how the estimated sedimentation rates varied within the simulations for the different  
284 winter wave climates, and Table 2 summarizes the values of the rate of incoming LST in those  
285 simulations. As, the wave direction and the significant wave height varied, according to the  
286 hindcasted local wave climate (see section 2.2), the rate LST ranged from  $205.0 \times 10^3 \text{ m}^3/\text{year}$  in SIM7  
287 to  $411.1 \times 10^3 \text{ m}^3/\text{year}$  in SIM8 (**Table 2**). This corresponds respectively to 1.23 times and 2.47 times  
288 the rate in SIM3. As the tidal range was also changed according to its local variations, those rates  
289 slightly decreased and increased for smaller and greater M2 tidal amplitudes, respectively (**Table 3**).

290 In the five simulations with the representative tidal range (middle panel, SIM4 to SIM8), the erosion  
291 estimated from the divergence of the residual transport in the southern group increased when the  
292 rate of incoming LST also increased. This was the case either with higher or more oblique waves. For  
293 instance, from SIM7 to SIM8 the average erosion rate per cell grows by  $36 \times 10^3 \text{ m}^3/\text{year}$ . This  
294 corresponds to about 10.1 % of the average erosion rate for this area in simulation SIM4. At the same  
295 time, from SIM5 to SIM6 the increase was limited to 2.1 %. By opposition, in the simulation with  
296 greater values of LST, the erosion in the western group was contained, even with some accretion  
297 estimated in SIM6 and SIM8. At the greater scale of the spit platform, positive sedimentation rates  
298 prevailed independently of the wave climate. In the case of average winter wave climate (SIM4), the  
299 sedimentation rate over the platform was estimated to  $+704.6 \times 10^3 \text{ m}^3/\text{year}$ , which was modulated  
300 by -8% and +16% as the wave direction was shifted by  $-4^\circ$  (SIM7, less oblique) and  $+4^\circ$  (SIM8, more  
301 oblique) respectively (**Table 2**).

302 Simulations SIM4, 7 and 8, were further reproduced with mean neap and high spring tidal ranges  
303 (upper and lower panels on **Figure 7** respectively). Similar to the wave only simulation (SIM2),  
304 accretion was predicted in the western group during neap tides. This accretion increased when the  
305 wave climate was rotated clockwise and the rate of incoming LST more than multiplied by a factor of  
306 two (**Table 3**, SIM7a to SIM8a). Compared to simulations with a representative tidal range, lower

307 erosion rates were estimated in the southern group. Still, this erosion grew from SIM7a to SIM8a as it  
308 did from SIM7 to SIM8. This behaviour was also observed with a greater tidal range (lower panel). So,  
309 the erosion driven by the divergence of the residual sediment transport at the spit – inlet boundary  
310 (i.e., in the southern group) has increased with incoming rate of LST, and this throughout the entire  
311 neap tide – spring tide cycle.

#### 312 4.3. Contribution of the southward transport

313 As shown first for SIM3 (**Figure 6**, lower panel), the contribution of the gradient of the southward  
314 (NS) to the divergence of total residual transport was calculated for all simulations. **Figure 8** shows  
315 the comparison of both integrated quantities in cells number 45 to number 58. For the twelve  
316 simulations, two patterns emerged from the Pearson's linear correlation coefficients (upper panel).  
317 First, little to no correlation existed between the two quantities in the cases with only tide or only  
318 waves. The scatter plot of the sedimentation rates by the NS sedimentation rates (lower left panel on  
319 **Figure 8**) further confirm this as the point clouds for SIM1 and SIM2 are dispersed. By opposition, the  
320 point cloud for SIM3 confirms the dominant contribution of NS sedimentation rates to the total  
321 sedimentation rate, those quantities also displaying a correlation coefficient greater than 0.9.  
322 Secondly, the relationship between those two quantities grew either when the tidal range or the  
323 longshore wave power increases. This is well visible from SIM4a to SIM4b, where the correlation  
324 coefficient grew from 0.56 to 0.93 when the tidal range went from 1.8 m to 4.5 m, and from SIM7a to  
325 SIM8a, where the correlation coefficient grew from 0.44 to 0.67 when the wave angle of incidence  
326 was shifted 8° clockwise and the incoming LST multiplied by a factor of 2.14. Then, the scatter plots  
327 for simulations with winter-representative wave climates (lower right panel on **Figure 8**) confirm that  
328 the southward component of the residual sediment transport had a dominant contribution in the  
329 cells with maximum erosion.

## 330 5. Discussion

### 331 5.1. Updrift erosion by waves and tide



332 The twelve morphostatic simulations (i.e., without morphological evolution) spanned a substantial  
333 range of forcing and modelling choices. Those with forcing parameters representative of yearly  
334 average waves or tide conditions first revealed that it takes both waves and tide to cause erosion  
335 near the subaerial spit terminus (i.e., in SIM3). The normalized sedimentation rates shown on **Figure**  
336 **9** illustrate this, with waves or tide alone resulting in accretion along the ocean flank of the spit (i.e.  
337 in the western group,  $\Delta V_w$ ) and causing little to hardly any erosion at the spit edge bounded by the  
338 inlet (i.e., in the southern group,  $\Delta V_s$ ). By opposition, erosion was predicted in both regions with  
339 combined forcings. Furthermore, the simulations with wave parameters representative of average  
340 winter conditions indicated these patterns were preserved in winter (SIM4), although with less  
341 erosion along the spit flank (**Figure 9**, left table). Also, simulations SIM4a and SIM4b indicated the  
342 erosion at the spit edge was a permanent feature throughout the entire neap-spring tidal cycle  
343 (**Figure 7**).

344 These first results are coherent with morphodynamic simulations of Lenstra et al. (2019) and of  
345 Cayocca (2001, 1996). Indeed, in the case of Lenstra et al. (2019) it was the combination of waves  
346 and tides that caused the breaching and the deepening of new channels near the updrift spit of their  
347 idealized tidal inlet. Then, before this, simulations of Cayocca (2001, 1996) for the present real-world  
348 inlet had suggested the breaching of new secondary channels across Cap Ferret's spit-platform  
349 occurred faster with both forcings, as shown in **Figure 10**. The figure shows the initial bathymetry  
350 and the simulated inlet morphology after 50 morphological timesteps, in the case with tidal forcing  
351 only (center panel) and in the case with both tidal and wave forcing (right panel). In the latter case,  
352 the newly breached secondary channel is deeper in the alignment of the spit and is slightly more  
353 bended clockwise. The addition of waves also leads to greater accretion on the beach along the spit's  
354 ocean side and on the intertidal shoals of the spit platform. In the present simulations, the sole  
355 computation of sediment transport fluxes further revealed which components of the sediment  
356 transport could explain this behaviour.

357 The southward component of the sediment transport produced negative sedimentation rates and  
358 appeared to promote the erosion of the spit (**Figure 6**, lower panel). Perpendicular to this, the  
359 transport was directed eastward (not shown) and produced positive sedimentation rates. This could  
360 be mostly explained by the bulldozer effect of waves at the vicinity of the inlet as described by Bertin  
361 et al. (2009), although the colinear wave and current interactions (Camenen & Larson, 2007; as  
362 described in sub-section 3.1) may also contribute. Going back to Cayocca's modelled morphology  
363 under waves and tide (**Figure 10**, right panel), the acceleration of the southward sediment transport  
364 would explain the erosion at the spit edge and the deepening of the channel, and the eastward  
365 sediment transport would explain the bending of the channel and the accretion along the flank of the  
366 spit.

367 Therefore, when the acceleration of the sediment transport along the coast (i.e., southward) tears  
368 off more sediments than the transport towards the coast (i.e., eastward) brings in, the Cap Ferret  
369 erodes and retreats. The sediment budget of the spit then appears to be linked to the balance  
370 between these two contributions, and the simulated scenarios indicate this balance was affected on  
371 the rate of sediment transport along the coast (**Figure 8**). The sensitivity tests of this balance to the  
372 wave climate and to the modelling choices further came to discuss the apparent relationship of the  
373 spit behaviour with the winter phase of the NAO and to discuss the apparent limitations of the  
374 present numerical experiment.

## 375 5.2. Influence of shifting atmospheric circulations

376 The sensitivity tests to the wave climate aimed to investigate the possible relationships between the  
377 updrift erosion of barrier spits and the increase of longshore wave power and/or sediment transport.  
378 In agreement with the observations described by Nahon et al. (2019), every time the longshore wave  
379 power and the incoming (or updrift) longshore sediment transport either increased or decreased, the  
380 erosion at the spit edge (i.e., in the southern group) also increased or decreased respectively (LST  
381 and  $\Delta V_s$  values in **Table 2**). These variations remained within a modest +/- 5% range centred on the

382 average winter estimates (**Figure 9**). However, they are fully coherent with the observed relationship  
383 between Cap Ferret's behaviour and the variations of the longshore wave power associated with  
384 decadal trends of the winter phase of the NAO. Indeed, higher and/or more oblique waves  
385 associated with NAO positive winters (**Figure 2**) would cause more erosion at the edge of the spit and  
386 this pattern prevailed from neap to spring tidal range (**Figure 7**). Alongside the geomorphological  
387 record at Cap Ferret, this brings a physically sounded explanation to the links between the updrift  
388 erosion of this barrier spit and the changes of nearshore wave climate caused by a shifting  
389 atmospheric circulation. Such relationship was also reported for the Skallingen spit by Aagaard and  
390 Sørensen (2013) and it is expected that similar processes may be at play at other places. For instance,  
391 on the Pacific coast of Northern America, the updrift (southern) margin of the entrance to Willapa  
392 Bay was remarkably eroded during the 2009-2010 El Niño, which had caused the increase of the  
393 southerly longshore wave power (Long Beach, in Barnard et al., 2011). Also, along the Dutch North  
394 Sea shores, the western margin of Ameland inlet is eroding updrift since 1974 (Elias et al., 2019). At  
395 both these locations, the causes to the observed erosion remain to be explicitly stated and the  
396 present results may well provide new insights into the forces at play.

### 397 5.3. From barriers to shoals

398 The sensitivity tests also revealed the accretion rates above the spit platform increased with higher  
399 longshore wave power and sediment transport (**Table 2**). Then, the capacity of the spit platform to  
400 retain sediment increases, likely due to the role that waves play in the formation of inlet shoals  
401 (Ridderinkhof et al., 2016). Also, the predicted positive sedimentation rates are coherent with  
402 satellite observations of Capo et al. (2014), who documented the accretion of the spit platform  
403 between 1986 and 2012. Therefore, both model results and observations suggest the Cap Ferret  
404 subaerial spit erodes at the benefit of its subtidal platform, like in Meistrell's (1972) spit-platform  
405 concept in which the subaerial spit release sediments to its platform until this latter one has reached  
406 a vertical equilibrium. In this process, the energy brought in by the waves to create this transfer also

407 increases the capacity of the adjacent platform and shoals to fix the material eroded from the barrier  
408 spit. This leads to the hypothesis that increased wave energy may accelerate the release of  
409 sediments stored in coastal barriers to the subtidal shoals of tidal inlets, for instance when those are  
410 expending due to sea level rise. This would become even more relevant for establishing  
411 morphodynamic prediction of barrier systems along shores where the longshore sediment transport  
412 presents a climatically-driven variability (Almar et al., 2015; Anderson et al., 2018; Marchesiello et al.,  
413 2020; Poirier et al., 2017; Silva et al., 2020; Splinter et al., 2012; Wiggins et al., 2020). Another  
414 fundamental aspect for such morphodynamic predictions was found to be the reliance of the results  
415 on the bottom friction parameterization.

#### 416 5.4. Friction parameterization as a major limitation

417 The adopted morphostatic approach aimed to identify the dominant sediment transport mechanism  
418 at the distal end of Cap Ferret. An unexpected behaviour in simulations with both tide and waves  
419 was that the model never predicted the retention of sand near the spit-end. This, even when the  
420 model was forced with waves representative of the period 1950 –1972, during which aerial  
421 photographs documented its growth (**Figure 1**). In fact, according to earlier and recent  
422 morphodynamic simulations referred above, this may not be completely surprising. Indeed, like it  
423 was reported by van Ormondt et al. (2020), 2DH models may struggle to reproduce the growth of  
424 real-world spits near tidal inlets. Some aspects certainly lie in the simplified wave-current  
425 interactions in these type of model and would only be overcome with fully coupled 3D models  
426 (Bertin et al., 2020). Other aspects may instead be related to the modelling choices and have  
427 motivated the further testing of three parameterizations of the spatio-temporal variability of the bed  
428 roughness in the circulation model. Indeed, near the spit-end, large-scale bedforms (Vaucher et al.,  
429 2018) attest of the spatial variability and the increase of the bottom roughness towards the inlet.  
430 Nearly absent along the northern beaches, these bedforms results from the erratic interactions  
431 between the waves and the strong ebb- and (dominant) flood-tidal currents. On the intertidal beach,

432 these bedforms have wavelength of more than a meter, for a height in the order of 30 cm (Vaucher  
433 et al. 2018). As a result, they are not resolved by the unstructured grid and need to be  
434 parameterized. Brakenhoff et al. (2020) recently reviewed the different approaches to represent the  
435 bed friction in these environments, and how modelled sediment transport rates were sensitive to its  
436 parameterization. So, simulation SIM3 was reproduce with four different parameterization of the  
437 bottom friction (**Table 3**, SIM3a-d).

438 In the case of a reduced Manning coefficient ( $0.02 \text{ s/m}^{1/3}$ , SIM3a), more representative of open  
439 beaches (Smith et al., 1993), although the incoming rate of LST was increased by a factor of 2.3  
440 compared to SIM3, the erosion patterns in the southern group were multiplied by a factor in the  
441 order of two (**Figure 11**). Then, two bedform predictors were tested as they are a way to estimate  
442 the spatio-temporal variability of the bed roughness length. In SIM3b and SIM3c (**Table 3**) the  
443 bedform predictors from Soulsby (1997) and van Rijn (2007) were respectively tested, using a log law  
444 formula to compute the drag coefficient. In the former case, the predictor returns the maximum  
445 value between the skin friction and the sand wave roughness plus the roughness of the current- or  
446 wave-generated ripple, in the latter case the predictor returns the skin friction plus the roughness of  
447 wave and current ripple roughness, mega-ripple and dunes. **Figure 11** presents the results of those  
448 simulations. In the case of Soulsby's (1997) predictor, in black, the results mimicked those obtained  
449 with Manning coefficient scaled to reproduce the tidal distortion through the inlet (SIM3, in blue),  
450 but which is also expected to overestimate the friction on the updrift open beaches. By opposition,  
451 results with van Rijn's (2007) predictor (SIM3c) were very similar those with a Manning coefficient  
452 representative of an open beach (SIM3a). Therefore, these predictors produced similar results to  
453 spatially constant Manning coefficients and, in the present case, *a priori* failed to represent the  
454 expected larger spatial variability. As a result, the model could only be expected to overestimate the  
455 erosion of the spit, in one case because the littoral drift would not bring in enough sand, in the other  
456 because the transiting sand would not be slowed down sufficiently.

457 Therefore, an alternative was to use a spatially varying Manning coefficient (like for instance in Bertin  
458 et al., 2009; Bruneau et al., 2011; Elias and Hansen, 2013; Teske, 2013), with lower values of the  
459 Manning coefficient for the open beaches than for the inlet and lagoon areas. The inset on **Figure 11**  
460 presents the spatial repartition of the Manning coefficient, which was derived after Mugica et al.  
461 (2016) and empirically based on sediment characteristics and vegetation cover. This repartition led to  
462 increased values of the updrift LST while the acceleration of this transport remained relatively limited  
463 towards the inlet (SIM3d). As a result, the predicted deficit at the spit-end was reduced, although  
464 most significantly within the western group.

465 These tests highlight how crucial it is to finely tune the bed friction to model spit – inlet  
466 morphodynamic interactions, in the present case and highly likely for other real (by opposition to  
467 synthetic) tidal inlet morphologies. Although some improvements of the bed roughness predictors  
468 (Elias et al., 2015) or the use of anisotropic Manning coefficient should be further tested (Demissie  
469 and Bacopoulos, 2017), such tuning appeared possible using a spatially varying Manning coefficient.  
470 However, such an approach should be supported/validated by measured elevations and/or velocities  
471 data much closer from the spit-end. Here, the spatial distribution was shown to give similar  
472 performances, than the calibrated ( $0.032 \text{ m}^{1/3} \cdot \text{s}$ ) Manning coefficient, although this was down with  
473 elevation data collected too far away from the area of interest. So, to build a realistic 2DH  
474 morphodynamic model of the Cap Ferret barrier spit, the next step would be to acquire the missing  
475 hydrodynamic data for validating the chosen parameterization of the bed friction.

## 476 **6. Conclusions**

477 A numerical experiment was performed to investigate the physical processes behind the apparent  
478 relationship between the Cap Ferret's updrift erosion and increasing rates of the longshore wave  
479 power. Residual sediment transport patterns near the spit-end were simulated with a 2DH process-  
480 based model, for a series of twenty morphostatic scenarios of forcing and parameterized bottom  
481 friction. Simulations confirmed that waves were contributing to the over-deepening of the secondary

482 tidal channel bounding the spit, they also suggested it was the acceleration of the longshore  
483 component of the sediment transport that was responsible for this over-deepening. This acceleration  
484 appeared to create a sediment deficit at the spit – inlet boundary which was found to increase with  
485 the wave power. So, the leading hypothesis was confirmed as higher and/or more oblique waves  
486 during NOA positive winters would favour the northward (updrift) retreat of the spit-end. By  
487 opposition, the cross-shore component was found to promote accretion, although the erosion  
488 induced by the longshore component dominated along the western southern flanks of the spit-end.  
489 Beyond the spit's bounding channel, accretion became dominant. Indeed, the sedimentation rates  
490 above the spit platform were always positive and increased with the wave power. In agreement with  
491 previous observations, the erosion of the spit therefore appeared to benefit to the spit-platform.  
492 Model results further support this transfer of sediment from the barrier to the shoals was  
493 accelerated under increased longshore wave power. However, the modelled transfer was found to  
494 be extremely sensitive to the parameterized bottom friction. Possibly because of this, the model  
495 failed to predict any positive sediment budget near the spit-end. Indeed, it is suspected the bottom  
496 was either too rough to generate a realistic amount of littoral drift, or too smooth to allow the  
497 fixation of the drifting sand near the spit end. A finely tuned spatially varying Manning coefficient  
498 could be a solution and would require an appropriate hydrodynamic dataset. Finally, the processes  
499 and limitations highlighted here are expected to be valid near other barrier spit with similar  
500 behaviour. In those places, greater attention should be given to the parameterization of the bottom  
501 friction in the morphodynamic modelling of spit – inlet interaction. Especially when inlets are  
502 expending at the expense of the spits.

503 **CRedit authorship contribution statement**

504 **Alphonse Nahon:** Conceptualization, Methodology, Formal analysis, Writing - Original Draft Writing;  
505 **Déborah Idier:** Methodology, Resources, Writing - Review & Editing; **Xavier Bertin:** Resources,  
506 Writing - Review & Editing; **Thomas Guérin:** Resources, Writing - Review & Editing; **Vincent Marieu:**  
507 Resources; **Nadia Sénéchal:** Supervision, Writing - Review & Editing; **Julie Mugica:** Supervision,  
508 Project administration, Funding acquisition.

509 **Declaration of competing interest**

510 The authors declare that they have no known competing financial interests or personal relationships  
511 that could have appeared to influence the work reported in this paper.

512 **Funding**

513 The work was part of a PhD grant co-funded by the BRGM, the Conseil Régional de Nouvelle-  
514 Aquitaine and the Syndicat Intercommunal du Bassin d'Arcachon, and part of the project MOSAIC.pt  
515 funded by the Portuguese Fundação para a Ciência e Tecnologia (FCT; PTDC/CTA-AMB/28909/2017).  
516 Numerical simulations relied on the Avakas computational facility, made available by the Région  
517 Nouvelle-Aquitaine, and used resources of the Portuguese National Distributed Computing  
518 Infrastructure (INCD) funded by Lisboa2020 Operational Program through the INCD project (LISBOA-  
519 01-0145-FEDER-022153).

520 **Acknowledgement**

521 The authors would like to thank Joseph Zhang for making the numerical model SCHISM available.

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- 524 Aagaard, T., Nielsen, J., Jensen, S.G., Friderichsen, J., 2004. Longshore sediment transport and coastal  
525 erosion at Skallingen, Denmark. *Geografisk Tidsskrift-Danish Journal of Geography* 104, 5–14.  
526 <https://doi.org/10.1080/00167223.2004.10649499>
- 527 Aagaard, T., Sørensen, P., 2013. Sea level rise and the sediment budget of an eroding barrier on the  
528 Danish North Sea coast. *Journal of Coastal Research* 65, 434–439.  
529 <https://doi.org/10.2112/SI65-074.1>
- 530 Allard, J., Chaumillon, E., Féliès, H., 2009. A synthesis of morphological evolutions and Holocene  
531 stratigraphy of a wave-dominated estuary: The Arcachon lagoon, SW France. *Continental  
532 Shelf Research* 29, 957–969. <https://doi.org/10.1016/j.csr.2008.11.017>
- 533 Almar, R., Kestenare, E., Reyns, J., Jouanno, J., Anthony, E.J., Laibi, R., Hemer, M., Du Penhoat, Y.,  
534 Ranasinghe, R., 2015. Response of the Bight of Benin (Gulf of Guinea, West Africa) coastline  
535 to anthropogenic and natural forcing, Part1: Wave climate variability and impacts on the  
536 longshore sediment transport. *Continental Shelf Research* 110, 48–59.  
537 <https://doi.org/10.1016/j.csr.2015.09.020>
- 538 Anderson, D., Ruggiero, P., Antolínez, J.A.A., Méndez, F.J., Allan, J., 2018. A Climate Index Optimized  
539 for Longshore Sediment Transport Reveals Interannual and Multidecadal Littoral Cell  
540 Rotations. *Journal of Geophysical Research: Earth Surface*.  
541 <https://doi.org/10.1029/2018JF004689>
- 542 Arnoux, F., Abadie, S., Bertin, X., Kojadinovic, I., 2018. A Database to Study Storm Impact Statistics  
543 along the Basque Coast. *Journal of Coastal Research* 85, 806–810.  
544 <https://doi.org/10.2112/SI85-162.1>
- 545 Barnard, P.L., Allan, J., Hansen, J.E., Kaminsky, G.M., Ruggiero, P., Doria, A., 2011. The impact of the  
546 2009–10 El Niño Modoki on U.S. West Coast beaches. *Geophysical Research Letters* 38, n/a-  
547 n/a. <https://doi.org/10.1029/2011GL047707>
- 548 Battjes, J.A., Janssen, J.P.F.M., 1978. Energy Loss and Set-Up Due to Breaking of Random Waves, in:  
549 *Coastal Engineering 1978. Presented at the 16th International Conference on Coastal  
550 Engineering, American Society of Civil Engineers, Hamburg, Germany, pp. 569–587.*  
551 <https://doi.org/10.1061/9780872621909.034>
- 552 Bertin, X., Bruneau, N., Breilh, J.-F., Fortunato, A.B., Karpytchev, M., 2012. Importance of wave age  
553 and resonance in storm surges: The case Xynthia, Bay of Biscay. *Ocean Modelling* 42, 16–30.  
554 <https://doi.org/10.1016/j.ocemod.2011.11.001>
- 555 Bertin, X., Fortunato, A.B., Oliveira, A., 2009. A modeling-based analysis of processes driving wave-  
556 dominated inlets. *Continental Shelf Research* 29, 819–834.  
557 <https://doi.org/10.1016/j.csr.2008.12.019>
- 558 Bertin, X., Li, K., Roland, A., Bidlot, J.-R., 2015. The contribution of short-waves in storm surges: Two  
559 case studies in the Bay of Biscay. *Continental Shelf Research* 96, 1–15.  
560 <https://doi.org/10.1016/j.csr.2015.01.005>
- 561 Bertin, X., Mengual, B., de Bakker, A., Guérin, T., Martins, K., Pezert, M., Lavaud, L., 2020. Recent  
562 Advances in Tidal Inlet Morphodynamic Modelling. *Journal of Coastal Research* 95, 1016.  
563 <https://doi.org/10.2112/SI95-198.1>
- 564 Brakenhoff, L., Schrijvershof, R., van der Werf, J., Grasmeyer, B., Ruessink, G., van der Vegt, M., 2020.  
565 From Ripples to Large-Scale Sand Transport: The Effects of Bedform-Related Roughness on  
566 Hydrodynamics and Sediment Transport Patterns in Delft3D. *JMSE* 8, 892.  
567 <https://doi.org/10.3390/jmse8110892>
- 568 Bruneau, N., Bertin, X., Castelle, B., Bonneton, P., 2014. Tide-induced flow signature in rip currents on  
569 a meso-macrotidal beach. *Ocean Modelling* 74, 53–59.  
570 <https://doi.org/10.1016/j.ocemod.2013.12.002>
- 571 Bruneau, N., Fortunato, A.B., Dodet, G., Freire, P., Oliveira, A., Bertin, X., 2011. Future evolution of a  
572 tidal inlet due to changes in wave climate, Sea level and lagoon morphology (Óbidos lagoon,

573 Portugal). *Continental Shelf Research* 31, 1915–1930.  
574 <https://doi.org/10.1016/j.csr.2011.09.001>

575 Camenen, B., Larson, M., 2007. A unified sediment transport formulation for inlet application (No.  
576 ERDC/CHL CR-07-1). U.S. Army Engineer Research and Development Center, Vicksburg, MS.

577 Capo, S., Lubac, B., Marieu, V., Robinet, A., Bru, D., Bonneton, P., 2014. Assessment of the decadal  
578 morphodynamic evolution of a mixed energy inlet using ocean color remote sensing. *Ocean*  
579 *Dynamics* 64, 1517–1530. <https://doi.org/10.1007/s10236-014-0762-1>

580 Cayocca, F., 2001. Long-term morphological modeling of a tidal inlet: the Arcachon Basin, France.  
581 *Coastal Engineering* 42, 115–142. [https://doi.org/10.1016/S0378-3839\(00\)00053-3](https://doi.org/10.1016/S0378-3839(00)00053-3)

582 Cayocca, F., 1996. Modélisation morphodynamique d'une embouchure tidale: application aux passes  
583 d'entrée du Bassin d'Arcachon (Thèse de Doctorat). Université Bordeaux I.

584 Charles, E., Idier, D., Thiébot, J., Le Cozannet, G., Pedreros, R., Ardhuin, F., Planton, S., 2012. Present  
585 Wave Climate in the Bay of Biscay: Spatiotemporal Variability and Trends from 1958 to 2001.  
586 *Journal of Climate* 25, 2020–2039. <https://doi.org/10.1175/JCLI-D-11-00086.1>

587 Davis, R.A., FitzGerald, D.M., 2004. Tidal Inlets, in: *Beaches and Coasts*. Blackwell Science Ltd, pp.  
588 209–244.

589 Demissie, H.K., Bacopoulos, P., 2017. Parameter estimation of anisotropic Manning's n coefficient for  
590 advanced circulation (ADCIRC) modeling of estuarine river currents (lower St. Johns River).  
591 *Journal of Marine Systems* 169, 1–10. <https://doi.org/10.1016/j.jmarsys.2017.01.008>

592 Doré, A., 2015. Modelling of the morphodynamic evolution of submarine sand dunes (Thèse de  
593 Doctorat). Université de Bordeaux, Bordeaux, France.

594 Elias, E., Teske, R., Van Der Spek, A., Lazar, M., 2015. Modelling tidal-inlet morphodynamics on  
595 medium time scales, in: *Coastal Sediments 2015*. WORLD SCIENTIFIC.  
596 [https://doi.org/10.1142/9789814689977\\_0230](https://doi.org/10.1142/9789814689977_0230)

597 Elias, E.P.L., Hansen, J.E., 2013. Understanding processes controlling sediment transports at the  
598 mouth of a highly energetic inlet system (San Francisco Bay, CA). *Marine Geology* 345, 207–  
599 220. <https://doi.org/10.1016/j.margeo.2012.07.003>

600 Elias, E.P.L., Van der Spek, A.J.F., Pearson, S.G., Cleveringa, J., 2019. Understanding sediment  
601 bypassing processes through analysis of high-frequency observations of Ameland Inlet, the  
602 Netherlands. *Marine Geology* 415, 105956. <https://doi.org/10.1016/j.margeo.2019.06.001>

603 Féliès, H., Lericolais, G., 2005. Internal architecture of an incised valley-fill on a wave- and tide-  
604 dominated coast (the Leyre incised valley, Bay of Biscaye, France). *Comptes Rendus*  
605 *Geoscience* 337, 1257–1266. <https://doi.org/10.1016/j.crte.2005.06.005>

606 Hansen, J.E., Elias, E., List, J.H., Erikson, L.H., Barnard, P.L., 2013. Tidally influenced alongshore  
607 circulation at an inlet-adjacent shoreline. *Continental Shelf Research* 56, 26–38.  
608 <https://doi.org/10.1016/j.csr.2013.01.017>

609 Hayes, M.O., 1979. Barrier island morphology as a function of tidal and wave regime, in: *Barrier*  
610 *Islands, from the Gulf of St. Lawrence to the Gulf of Mexico*. Leatherman, S., pp. 1–27.

611 Herrling, G., Winter, C., 2018. Tidal inlet sediment bypassing at mixed-energy barrier islands. *Coastal*  
612 *Engineering* 140, 342–354. <https://doi.org/10.1016/j.coastaleng.2018.08.008>

613 Hoan, L.X., Hanson, H., Larson, M., Kato, S., 2011. A mathematical model of spit growth and barrier  
614 elongation: Application to Fire Island Inlet (USA) and Badreveln Spit (Sweden). *Estuarine,*  
615 *Coastal and Shelf Science* 93, 468–477. <https://doi.org/10.1016/j.ecss.2011.05.033>

616 Hurrell, J.W., 2015. *The Climate Data Guide: Hurrell North Atlantic Oscillation (NAO) Index (Station-*  
617 *based)*.

618 Hurrell, J.W., Deser, C., 2009. North Atlantic climate variability: The role of the North Atlantic  
619 Oscillation. *Journal of Marine Systems* 78, 28–41.  
620 <https://doi.org/10.1016/j.jmarsys.2008.11.026>

621 Idier, D., Castelle, B., Charles, E., Mallet, C., 2013. Longshore sediment flux hindcast: spatio-temporal  
622 variability along the SW Atlantic coast of France. *Journal of Coastal Research* 165, 1785–  
623 1790. <https://doi.org/10.2112/SI65-302.1>

624 Kalnay, E., Kanamitsu, M., Kistler, R., Collins, W., Deaven, D., Gandin, L., Iredell, M., Saha, S., White,  
625 G., Woollen, J., Zhu, Y., Leetmaa, A., Reynolds, R., Chelliah, M., Ebisuzaki, W., Higgins, W.,  
626 Janowiak, J., Mo, K.C., Ropelewski, C., Wang, J., Jenne, R., Joseph, D., 1996. The NCEP/NCAR  
627 40-Year Reanalysis Project. *Bulletin of the American Meteorological Society* 77, 437–471.  
628 [https://doi.org/10.1175/1520-0477\(1996\)077<0437:TNYRP>2.0.CO;2](https://doi.org/10.1175/1520-0477(1996)077<0437:TNYRP>2.0.CO;2)

629 Kraus, N.C., 2000. Reservoir Model of Ebb-Tidal Shoal Evolution and Sand Bypassing. *Journal of*  
630 *Waterway, Port, Coastal, and Ocean Engineering* 126, 305–313.  
631 [https://doi.org/10.1061/\(ASCE\)0733-950X\(2000\)126:6\(305\)](https://doi.org/10.1061/(ASCE)0733-950X(2000)126:6(305))

632 Larson, M., Camenen, B., Nam, P.T., 2011. A Unified Sediment Transport Model for Inlet Application.  
633 *Journal of Coastal Research* 59, 27–38. <https://doi.org/10.2112/SI59-004.1>

634 Larson, M., Kraus, N.C., Connell, K.J., 2007. MODELING SEDIMENT STORAGE AND TRANSFER FOR  
635 SIMULATING REGIONAL COASTAL EVOLUTION, in: *Coastal Engineering 2006*. Presented at the  
636 Proceedings of the 30th International Conference, World Scientific Publishing Company, San  
637 Diego, California, USA, pp. 3924–3936. [https://doi.org/10.1142/9789812709554\\_0330](https://doi.org/10.1142/9789812709554_0330)

638 Lenstra, K.J.H., Ridderinkhof, W., Vegt, M., 2019. Unraveling the Mechanisms That Cause Cyclic  
639 Channel-Shoal Dynamics of Ebb-Tidal Deltas: A Numerical Modeling Study. *J. Geophys. Res.*  
640 *Earth Surf.* 124, 2778–2797. <https://doi.org/10.1029/2019JF005090>

641 Marchesiello, P., Kestenare, E., Almar, R., Boucharel, J., Nguyen, N.M., 2020. Longshore drift  
642 produced by climate-modulated monsoons and typhoons in the South China Sea. *Journal of*  
643 *Marine Systems* 211, 103399. <https://doi.org/10.1016/j.jmarsys.2020.103399>

644 Meistrell, F.J., 1972. The spit-platform concept: laboratory observation of spit development, in:  
645 Schwartz, M.L. (Ed.), *Spits and Bars*. Dowden, Hutchinson & Ross, pp. 225–283.

646 Mugica, J., Paris, F., Nicolae Lerma, A., Pedreros, R., Ayache, B., Garcin, M., Bulteau, T., Hoareau, A.,  
647 2016. Caractérisation de l'aléa submersion marine dans le cadre des PPRM du Bassin  
648 d'Arcachon - Approche dynamique avec prise en compte des ouvrages de protection côtière.  
649 Rapport final. (No. BRGM/RP-64807-FR). BRGM, Orléans, France.

650 Nahon, A., 2018. Ongoing morphological evolution of a Holocene coastal barrier spit - The Cap Ferret,  
651 at the entrance of the Bay of Arcachon (PhD Thesis). University of Bordeaux, Bordeaux,  
652 France.

653 Nahon, A., Idier, D., Sénéchal, N., Féniès, H., Mallet, C., Mugica, J., 2019. Imprints of wave climate  
654 and mean sea level variations in the dynamics of a coastal spit over the last 250 years: Cap  
655 Ferret, SW France. *Earth Surface Processes and Landforms* esp.4634.  
656 <https://doi.org/10.1002/esp.4634>

657 Nichols, M., Allen, G., 1981. Sedimentary processes in coastal lagoons. *Unesco Technical Papers In*  
658 *Marine Science* 27–80.

659 Nienhuis, J.H., Ashton, A.D., 2016. Mechanics and rates of tidal inlet migration: modeling and  
660 application to natural examples: InletMigration. *Journal of Geophysical Research: Earth*  
661 *Surface*. <https://doi.org/10.1002/2016JF004035>

662 Nienhuis, J.H., Lorenzo-Trueba, J., 2019. Can Barrier Islands Survive Sea-Level Rise? Quantifying the  
663 Relative Role of Tidal Inlets and Overwash Deposition. *Geophys. Res. Lett.* 46, 14613–14621.  
664 <https://doi.org/10.1029/2019GL085524>

665 Oertel, G.F., 1985. The barrier island system. *Marine Geology* 63, 1–18.  
666 [https://doi.org/10.1016/0025-3227\(85\)90077-5](https://doi.org/10.1016/0025-3227(85)90077-5)

667 Orescanin, M.M., Elgar, S., Raubenheimer, B., 2016. Changes in bay circulation in an evolving multiple  
668 inlet system. *Continental Shelf Research* 124, 13–22.  
669 <https://doi.org/10.1016/j.csr.2016.05.005>

670 Poirier, C., Tessier, B., Chaumillon, É., Bertin, X., Fruergaard, M., Mouazé, D., Noël, S., Weill, P.,  
671 Wöppelmann, G., 2017. Decadal changes in North Atlantic atmospheric circulation patterns  
672 recorded by sand spits since 1800CE. *Geomorphology* 281, 1–12.  
673 <https://doi.org/10.1016/j.geomorph.2016.12.028>

674 Ridderinkhof, W., de Swart, H.E., van der Vegt, M., Hoekstra, P., 2016. Modeling the growth and  
675 migration of sandy shoals on ebb-tidal deltas. *Journal of Geophysical Research: Earth*  
676 *Surface*. <https://doi.org/10.1002/2016JF003823>  
677 Roland, A., Zhang, Y.J., Wang, H.V., Meng, Y., Teng, Y.-C., Maderich, V., Brovchenko, I., Dutour-Sikiric,  
678 M., Zanke, U., 2012. A fully coupled 3D wave-current interaction model on unstructured  
679 grids. *Journal of Geophysical Research: Oceans* 117, n/a-n/a.  
680 <https://doi.org/10.1029/2012JC007952>  
681 Senechal, N., Sottolichio, A., Bertrand, F., Goeldner-Gianella, L., Garlan, T., 2013. Observations of  
682 waves' impact on currents in a mixed-energy tidal inlet: Arcachon on the southern French  
683 Atlantic coast. *Journal of Coastal Research* 165, 2053–2058. [https://doi.org/10.2112/SI65-](https://doi.org/10.2112/SI65-347.1)  
684 [347.1](https://doi.org/10.2112/SI65-347.1)  
685 SHOM, 2014. Références altimétriques maritimes - édition 2014.  
686 SHOM, n.d. Tide gauge Arcachon Eyrac [WWW Document]. URL  
687 <https://data.shom.fr/donnees/refmar/190> (accessed 14/05/2020).  
688 Silva, A.P., Klein, A.H.F., Fetter-Filho, A.F.H., Hein, C.J., Méndez, F.J., Broggio, M.F., Dalinghaus, C.,  
689 2020. Climate-induced variability in South Atlantic wave direction over the past three  
690 millennia. *Sci Rep* 10, 18553. <https://doi.org/10.1038/s41598-020-75265-5>  
691 Smith, J.M., Larson, M., Kraus, N.C., 1993. Longshore current on a barred beach: Field measurements  
692 and calculation. *Journal of Geophysical Research* 98, 22717.  
693 <https://doi.org/10.1029/93JC02116>  
694 Soulsby, R., 1997. *Dynamics of Marine Sands*. Thomas Telford Ltd.  
695 <https://doi.org/10.1680/doms.25844>  
696 Splinter, K.D., Davidson, M.A., Golshani, A., Tomlinson, R., 2012. Climate controls on longshore  
697 sediment transport. *Continental Shelf Research* 48, 146–156.  
698 <https://doi.org/10.1016/j.csr.2012.07.018>  
699 Teske, R., 2013. Tidal inlet channel stability in long term process based modelling (MSc Thesis).  
700 Utrecht University.  
701 van Ormondt, M., Nelson, T.R., Hapke, C.J., Roelvink, D., 2020. Morphodynamic modelling of the  
702 wilderness breach, Fire Island, New York. Part I: Model set-up and validation. *Coastal*  
703 *Engineering* 157, 103621. <https://doi.org/10.1016/j.coastaleng.2019.103621>  
704 van Rijn, L.C., 2007. Unified View of Sediment Transport by Currents and Waves. I: Initiation of  
705 Motion, Bed Roughness, and Bed-Load Transport. *Journal of Hydraulic Engineering* 133, 649–  
706 667. [https://doi.org/10.1061/\(ASCE\)0733-9429\(2007\)133:6\(649\)](https://doi.org/10.1061/(ASCE)0733-9429(2007)133:6(649))  
707 Vaucher, R., Pittet, B., Humbert, T., Ferry, S., 2018. Large-scale bedforms induced by supercritical  
708 flows and wave-wave interference in the intertidal zone (Cap Ferret, France). *Geo-Mar Lett*  
709 38, 287–305. <https://doi.org/10.1007/s00367-017-0526-2>  
710 Walton, Jr., T.L., Adams, W.D., 1976. Capacity of inlet outer bars to store sand, in: *Proceedings of the*  
711 *15th Coastal Engineering Conference, Honolulu, HI*. ASCE, pp. 1919–1937.  
712 Wiggins, M., Scott, T., Masselink, G., McCarroll, R.J., Russell, P., 2020. Predicting beach rotation using  
713 multiple atmospheric indices. *Marine Geology* 426, 106207.  
714 <https://doi.org/10.1016/j.margeo.2020.106207>  
715 Zhang, Y.J., Ye, F., Stanev, E.V., Grashorn, S., 2016. Seamless cross-scale modeling with SCHISM.  
716 *Ocean Modelling* 102, 64–81. <https://doi.org/10.1016/j.ocemod.2016.05.002>  
717

719 **Table 1** – Comparison of modelled and observed elevations at locations A, B and C shown on Figure  
 720 1;  $a_{M2}$  and  $a_{M4}$  are the amplitude of tidal component M2 and M4 respectively and  $\theta_{M2}$  and  $\theta_{M4}$  their  
 721 respective phase (adapted from Nahon, 2018).

	Bias (cm)	Erms (cm)	$a_{M2}$ (m)	$a_{M4}$ (m)	$\theta_{M2}$ (°)	$\theta_{M4}$ (°)	$a_{M4}/a_{M2}$	$2\theta_{M2} - \theta_{M4}$
Obs. A	--	--	1.36	0.05	93.1	322.0	0.035	224.4
Mod. A	-0.21	7.35	1.36	0.04	91.5	324.0	0.029	219.0
Obs. B	--	--	1.45	0.03	105.0	30.3	0.019	179.7
Mod. B	2.25	10.39	1.42	0.07	105.0	57.1	0.049	152.9
Obs. C	--	--	1.36	0.08	121.0	35.6	0.061	206.4
Mod. C	-1.93	11.84	1.35	0.11	116.0	26.6	0.078	205.0

722

723 **Table 2** – Tidal (M2) amplitude, mean wave parameters and estimated transport and sedimentation  
 724 rates near the spit's distal end for 8 selected simulations; LST values correspond to the average  
 725 southward transport in cells 5 to 10 (dashed black frames on **Figure 5** and **6**);  $\Delta V_w$  and  $\Delta V_s$  are the  
 726 average estimated volumetric changes (per 300-m wide cell) in the western and southern groups,  
 727 respectively.

<u>Simulations</u>	<u>M2</u> (m)	<u>Hs</u> (m)	<u>Tp</u> (s)	<u>Dir</u> (°)	<u>LST</u> ( $10^3 \text{ m}^3 \cdot \text{y}^{-1}$ )	<u><math>\Delta V_w</math></u> ( $10^3 \text{ m}^3 \cdot \text{y}^{-1}$ )	<u><math>\Delta V_s</math></u> ( $10^3 \text{ m}^3 \cdot \text{y}^{-1}$ )	<u>Platform</u> ( $10^3 \text{ m}^3 \cdot \text{y}^{-1}$ )
SIM1	3.6	0.00	0.00	-	1.1	6	16	-
SIM2	0.0	2.02	11.79	291.83	150.2	186	-95	-
SIM3	3.6	2.02	11.79	291.83	166.5	-51	-312	644
SIM4	3.6	2.56	13.00	290.00	304.2	-6	-329	705
SIM5	3.6	2.41	13.00	290.00	258.5	-17	-328	676
SIM6	3.6	2.70	13.00	290.00	357.2	8	-335	763
SIM7	3.6	2.56	13.00	286.00	205.0	-15	-311	648
SIM8	3.6	2.56	13.00	294.00	411.1	3	-347	819

728

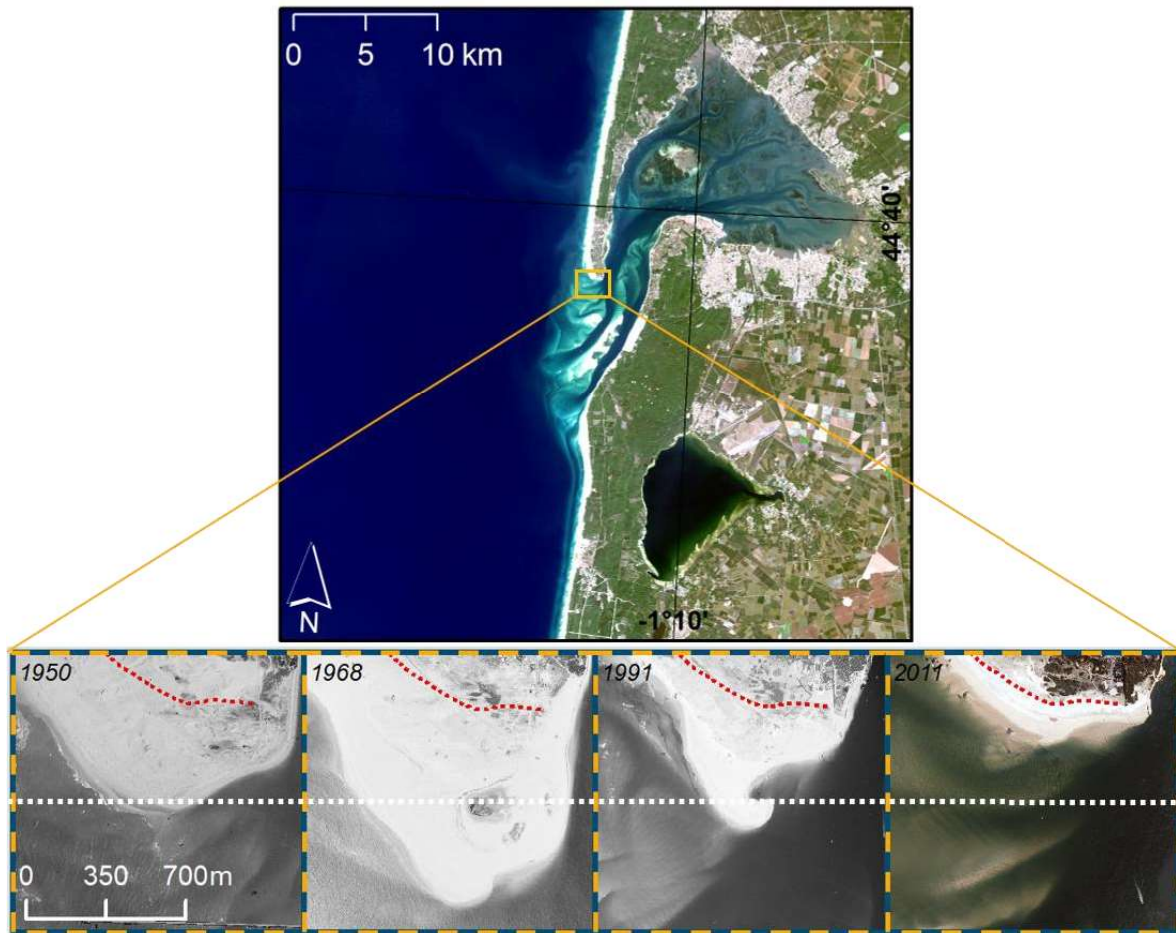
729

730 **Table 3** – Tidal (M2) amplitude and mean wave parameters for 9 sensitivity analysis simulations, in  
 731 bold are indicated changes from simulations given in **Table 2**; LST values are the same as in **Table 2**.

<u>Simulations</u>	<u>M2</u> (m)	<u>Hs</u> (m)	<u>Tp</u> (s)	<u>Dir</u> (°)	<u>LST</u> (10 <sup>3</sup> m <sup>3</sup> .y <sup>-1</sup> )	other
SIM3a	3.6	2.02	11.79	291.83	379.1	<b>Manning's n = 0.02 s/m<sup>1/3</sup></b>
SIM3b	3.6	2.02	11.79	291.83	145.5	<b>Bedforms (RS97)</b>
SIM3c	3.6	2.02	11.79	291.83	461.8	<b>Bedforms (VR07)</b>
SIM3d	3.6	2.02	11.79	291.83	411.8	<b>0.019 ≤ n ≤ 0.040 s/m<sup>1/3</sup></b>
SIM4a	<b>1.8</b>	2.56	13.00	290.00	291.4	
SIM7a	<b>1.8</b>	2.56	13.00	286.00	185.9	
SIM8a	<b>1.8</b>	2.56	13.00	294.00	397.9	
SIM4b	<b>4.5</b>	2.56	13.00	290.00	316.2	
SIM7b	<b>4.5</b>	2.56	13.00	286.00	219.8	
SIM8b	<b>4.5</b>	2.56	13.00	294.00	420.9	

732

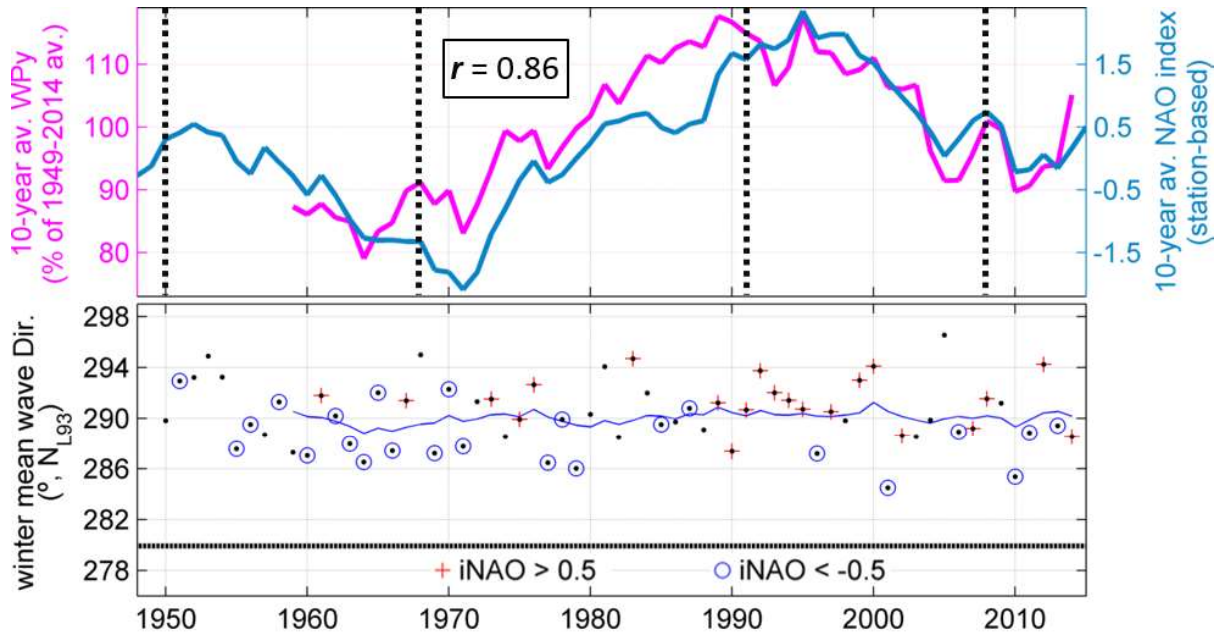
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734

735 **Figure 1** – Satellite view of the Bay of Arcachon (Landsat 8, October 2014), with aerial photos of Cap  
 736 Ferret’s distal end from 1950 to 2011 (red dotted line indicates the position of the dune toe in  
 737 October 2014).

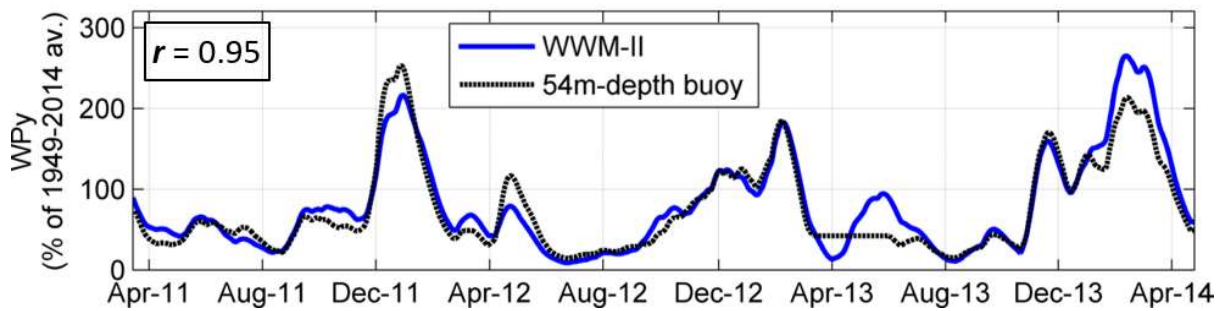
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740 **Figure 2** – Winter -DJFM- wave climate. **Upper panel:** decadal average of the normalized longshore  
 741 wave power (WPy) and of the NAO station-based index (running mean of the winter averaged values,  
 742 decentred over the ten preceding years), vertical dotted lines stand for the date of the photos on  
 743 **Figure 1**,  $r$  is Pearson's linear correlation coefficient between the two curves; **lower panel:** direction  
 744 of winter mean incident wave power, expressed in nautical convention into Lambert-93 projection  
 745 (subtract  $3.12^{\circ}$  for true north), the thin blue line is the decadal average and the dotted black line is  
 746 the direction normal to the 20-km spit orientation.

747



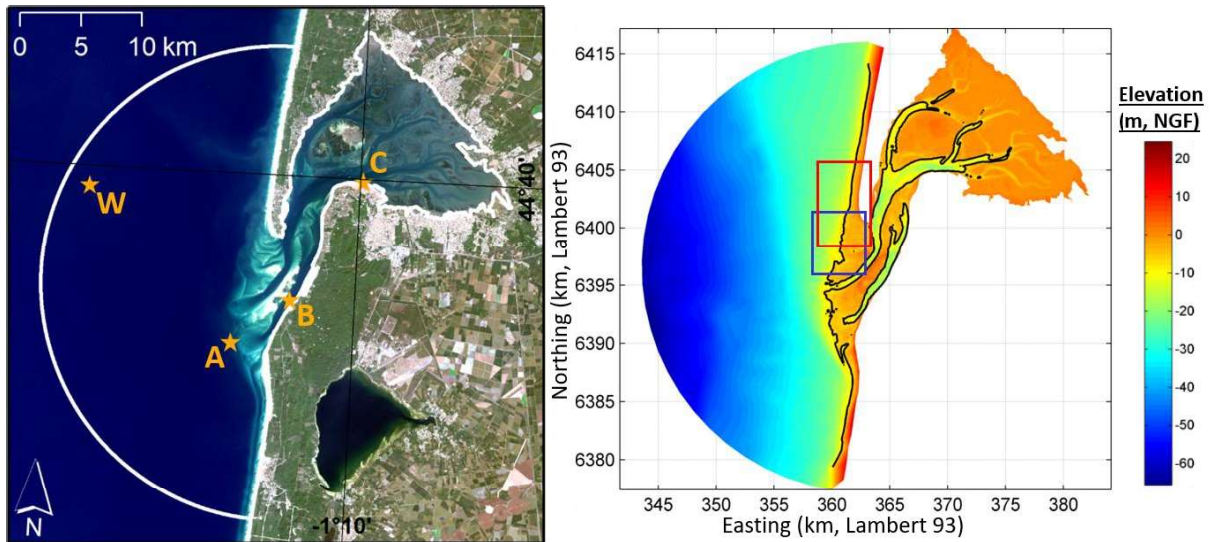
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749 **Figure 3** – Comparison of hindcasted (solid blue curve) and observed (dotted black curve) normalized  
 750 longshore wave power (WPy), 90-day centred average,  $r$  is Pearson's linear correlation coefficient  
 751 between the two curves.

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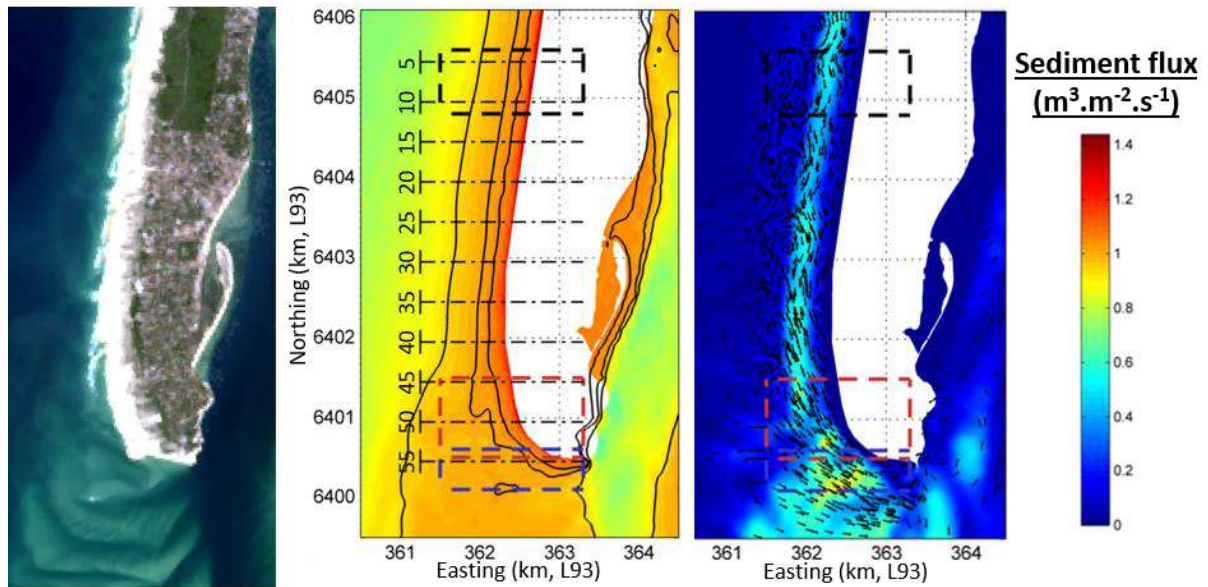




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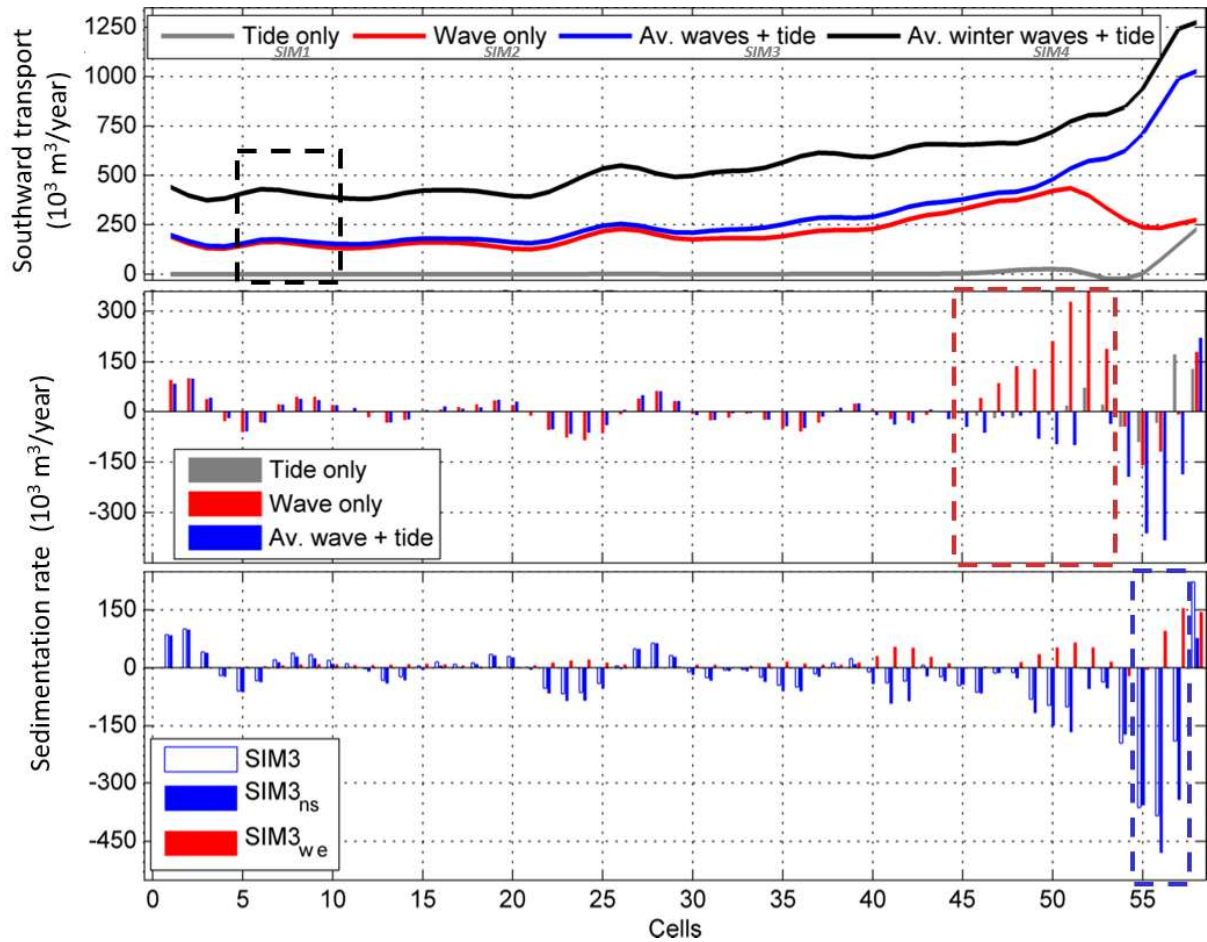
755 **Figure 4** – Model domain. **Left panel:** satellite Landsat 8 view of the Bay of Arcachon and its tidal  
 756 inlet on October 1st, 2014, the white contour represents the boundary of SCHISM’s computational  
 757 grid and the four yellow stars stand for the Cap Ferret wave buoy (W), the ebb-tidal delta pressure  
 758 sensor (A), the southern channel pressure sensor (B) and *Eyrac*’s tide gauge (C); **right panel:** model  
 759 computational domain and bathymetry (local mean sea level is at +0.36 m NGF), black contours  
 760 correspond to -7 m NGF, the red frame shows the area of interest at the distal end of Cap Ferret and  
 761 corresponds to **Figure 5** extent, the blue frame is the area over which the spit platform  
 762 sedimentation rate was computed.

763



764

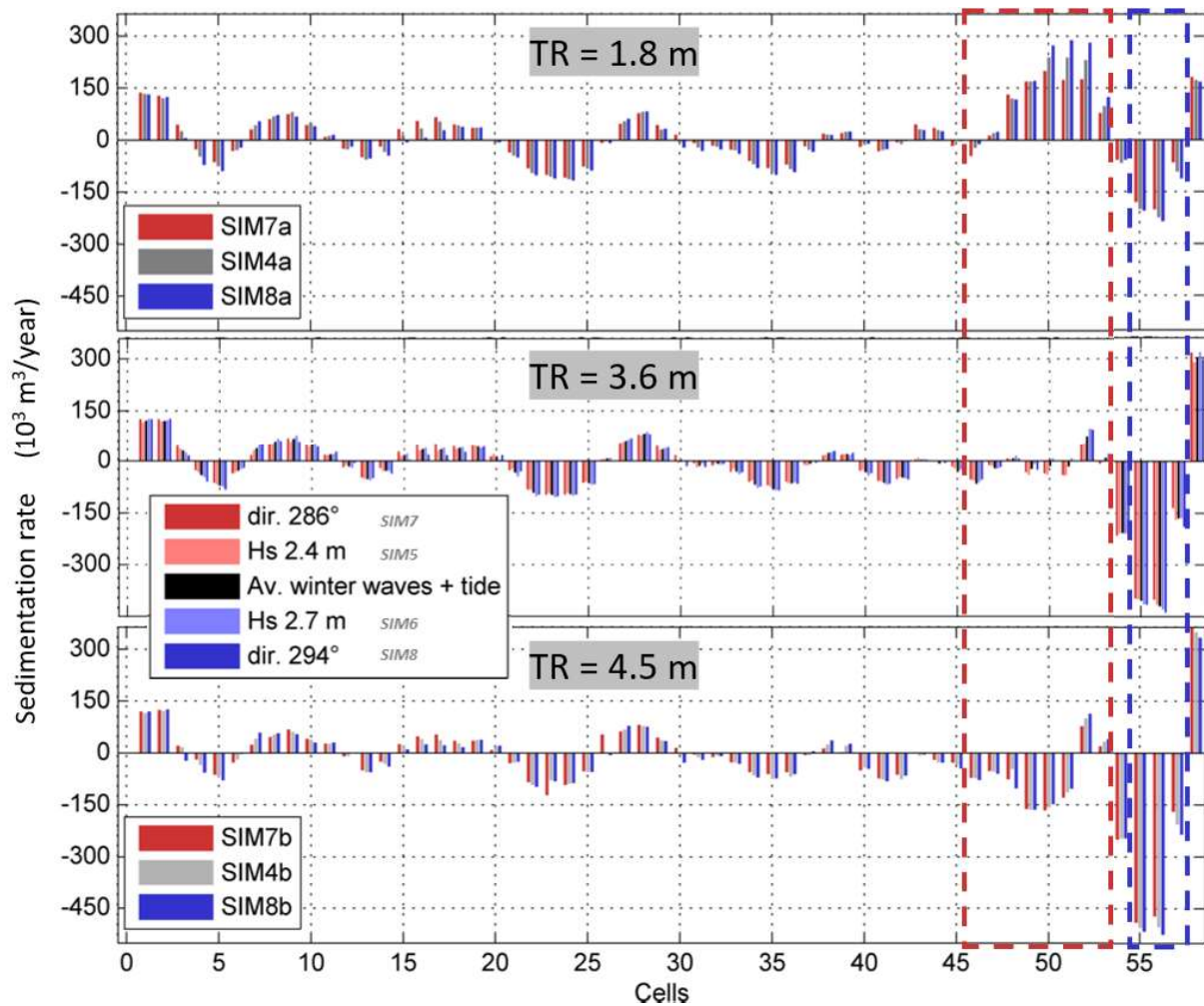
765 **Figure 5** – Modelling sediment transport patterns at the distal end of Cap Ferret. **Left panel:** Satellite  
 766 Landsat 8 view of the subaerial spit's last 5 km on October 1st, 2014; **centre panel:** model  
 767 bathymetry, solid black lines stand for the -7 m NGF, -2 m NGF and 0 m NGF contours (-2 m is close  
 768 to the offshore limit of the intertidal area), dashed frames delimit sediment transport integration  
 769 cells highlighted in **Figures 6-9 & 14**, with in red the western group and in blue the southern group,  
 770 the dot-dashed lines stand for the centres of cell number 5 to 55 with the vertical standing for the  
 771 cells' width; **right panel:** residual sediment transport patterns in the average winter case (SIM4).



772

773 **Figure 6** – Integrated sediment transport patterns. **Upper panel:** cell-averaged southward sediment  
 774 transport, integrated in the west-east direction (SIM1 to SIM4); **centre panel:** cell-integrated  
 775 sedimentation rates for SIM1 to SIM3; **lower panel:** cell-integrated sedimentation rates for SIM3,  
 776 with the respective contributions of the southward (ns) and eastward (we) transport components.  
 777 Simulation names refer to simulation parameters given in **Table 2** and the dashed framed cells  
 778 correspond to delimited areas on **Figure 5** centre panel.

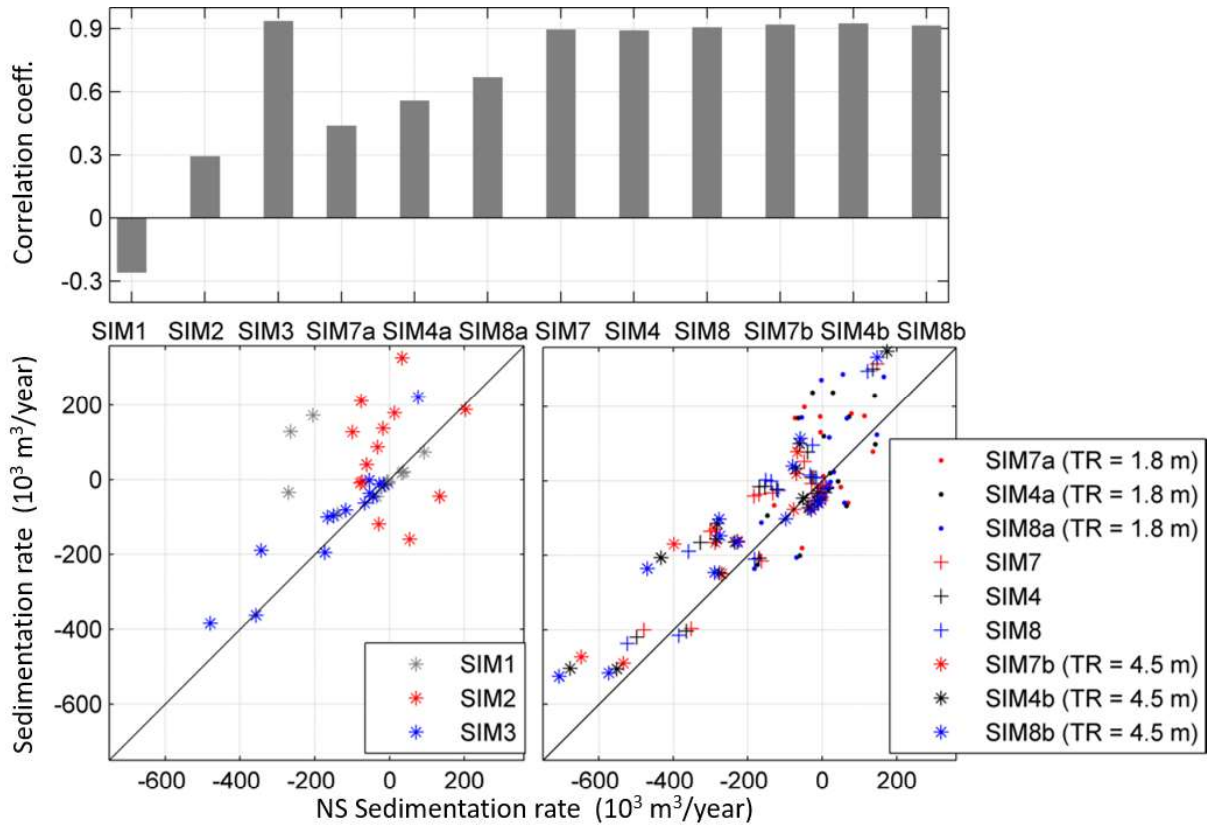
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780

781 **Figure 7** – Cell-integrated sedimentation rates for winter wave climate scenarios and for three  
 782 different values of tidal range (TR). Simulation names refer to simulation parameters given in **Table 2**  
 783 and **Table 3**, and the dashed framed cells correspond to delimited areas on **Figure 5** centre panel.

784



785

786 **Figure 8** – Total sedimentation rate vs. sedimentation associated with south-north sediment fluxes  
 787 from cell number 45 to cell number 58. **Upper panel:** Pearson's linear correlation coefficients  
 788 between both cell-averaged sedimentation rates; **lower panels:** scatter plot of this quantities, with  
 789 the black straight curve standing for equation  $y = x$ . Simulation names refer to simulation parameters  
 790 given in **Table 2** and **Table 3**.

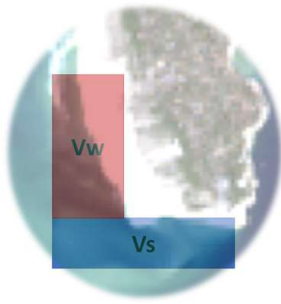
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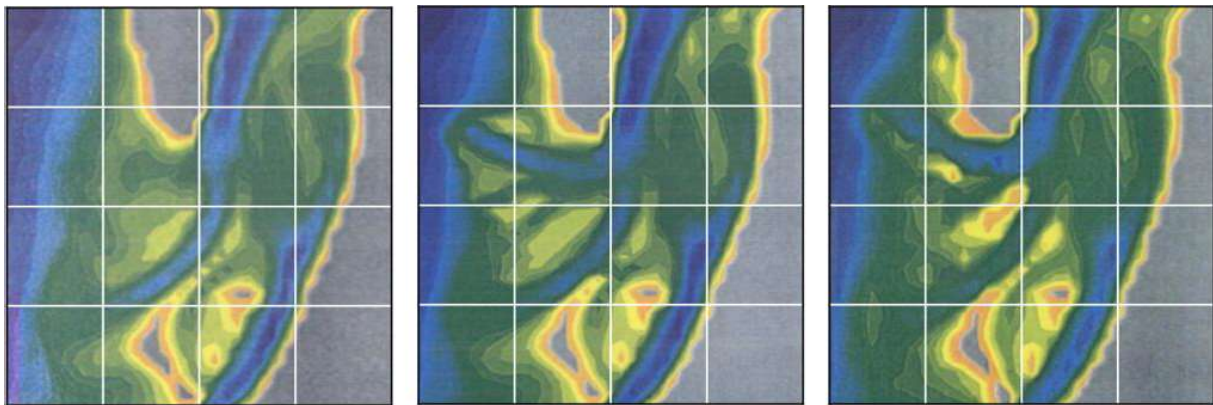


Representative conditions			Winter representative conditions		
	$\Delta V_w$	$\Delta V_s$		$\Delta V_w$	$\Delta V_s$
Tide only	0.10	0.05	NAO - ; -4°	-0.33 ; -0.31	-0.99 ; -0.94
Wave only	3.66	-0.29	Average	-0.11	-1.00
Wave + tide	-1.00	-0.95	NAO + ; +4°	0.16 ; 0.06	-1.02 ; -1.05

795

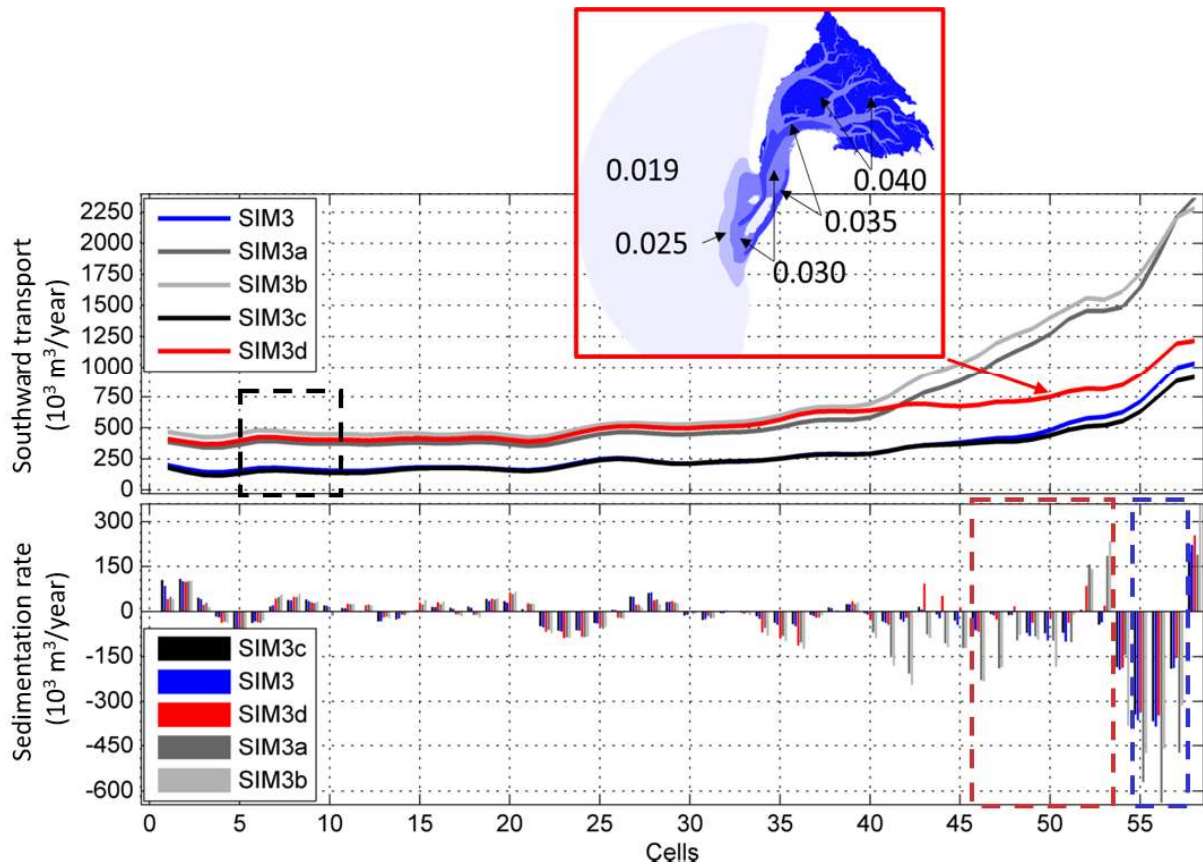
796 **Figure 9** – Normalized averaged volumetric variations per cell in the western ( $\Delta V_w$ ) and southern  
 797 ( $\Delta V_s$ ) group of cells indicated in **Figure 5**, for annual representative wave climate and winter  
 798 representative wave climates associated with the dominant phase of the NAO. Positive values in red  
 799 stand for accretion while negative value in blue stand for erosion,  $\Delta V_w$  values were normalized by  
 800 SIM3 (waves and tide) absolute value while  $\Delta V_s$  were normalized by the one of SIM4 (winter waves  
 801 and tide)

802



803

804 **Figure 10** – Morphodynamic results of Cayocca (2001, 1996), adapted from Cayocca (1996). **Left**  
 805 **panel:** initial model bathymetry contours, from -20.8 m NGF (dark blue shade) to +2 m NGF (grey  
 806 shade) at 1.2 m interval; **centre panel:** simulated morphology after 50 morphological timesteps with  
 807 tidal forcing only; **right panel:** simulated morphology after 50 morphological timesteps with tidal and  
 808 wave forcing.



809

810 **Figure 11** – Sensitivity of integrated sediment transport patterns to spatio-temporally varying bottom  
 811 friction coefficient (SIM3a-d). **Upper panel:** cell-averaged southward sediment transport, integrated  
 812 in the west-east direction; **lower panel:** cell-integrated sedimentation rates; **red inset:** spatial  
 813 repartition of the Manning coefficient ( $s/m^{1/3}$ ) used in SIM3d, adapted from Mugica et al. (2016).  
 814 Simulation names refer to simulation parameters given in **Table 2** and **Table 3**, and dashed framed  
 815 cells correspond to delimited areas on **Figure 5** centre panel.

**Declaration of interests**

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

The authors declare the following financial interests/personal relationships which may be considered as potential competing interests: