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Julie Mugica

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**Imprints of wave climate and mean sea level variations in the dynamics of a coastal spit over the last 250 years: Cap Ferret, SW France**

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3 1 **Imprints of wave climate and mean sea level variations in the**  
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6 2 **dynamics of a coastal spit over the last 250 years: Cap Ferret, SW**  
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8 **France**  
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39 16 **Key words:**

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41 17 Barrier systems; Tidal inlet; Sea level rise; Climate; NAO; Dalton Minimum.  
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3 18 **Abstract**  
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5  
6 19 In coastal areas, sea level rise (SLR) and changing wave climates are expected to be  
7  
8 20 the main oceanic drivers of shoreline adjustments. These drivers have been shown to  
9  
10 21 vary on a wide spectrum of spatial and temporal scales. Nonetheless, a general rule  
11  
12 22 about how this variability impacts global shorelines remains to be articulated. Here, we  
13  
14 23 discussed the impacts of wave climate changes and SLR on the evolution of a barrier-  
15  
16 24 spit – inlet system over the last 250 years. The distal end of the Cap Ferret barrier-spit,  
17  
18 25 SW France, has undergone large-scale oscillations that were well correlated with  
19  
20 26 variations of the decadal average of the winter North Atlantic Oscillation (NAO) index.  
21  
22 27 The local wave climate hindcast supports that increased alongshore wave energy fluxes  
23  
24 28 associated with the positive phase of the NAO were responsible for the updrift retreat of  
25  
26 29 the spit. In another case, the spit has elongated downdrift when waves were less  
27  
28 30 energetic and more shore normal, as during the negative phase of the NAO. In addition,  
29  
30 31 lower rates of SLR appeared necessary for the spit to develop, as higher rates of SLR  
31  
32 32 very likely forced the adjacent inlet to enlarge, at the expense of the spit. These results  
33  
34 33 should help to predict and detect coastal adjustments driven by climate change and  
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36 34 variability.  
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## 36 Introduction

37 Slower eustatic sea level rise (SLR) over the last 6 ka has allowed the formation of  
38 barrier spits at the entrance of many of the world's estuaries and bays (Davis, 2013).  
39 Growing barriers would then create bar-built estuaries or coastal lagoons where rich  
40 ecosystems have developed. Provided that such sheltered areas remained open to the  
41 sea, they have become attractive for human activities. This has added an economic  
42 dimension to their ecological value and role (Newton et al., 2014). Such factors are  
43 among the many reasons that continue to motivate multidisciplinary studies to improve  
44 our understanding of barrier spit dynamics and their interactions with estuary and  
45 lagoon entrances.

46 Where sediments are abundant, it is recognised that sort of a competition between  
47 wave-driven alongshore sediment transport (LST) and estuarine (fresh and saltwater)  
48 flows, controls the existence of spits as it does for inlets (FitzGerald et al., 2015; Hayes,  
49 1979; Nichols and Allen, 1981). From the spit perspective, LST generally plays a  
50 constructive role while flows favour inlets (Kraus, 1998; Kraus and Seabergh, 2002).

51 Inlet (cross-section) stability can be rated as good, fair or poor based on the ratio of the  
52 spring tidal prism by the total littoral drift (Bruun, 1978). Larger freshwater inputs  
53 ultimately favour inlet stability as confirmed by numerical modelling (Zhou et al., 2014).

54 These forces vary as waves, sea level and rainfalls change with seasons and climate  
55 shifts. In mixed-energy environments (Hayes, 1979), such variability can lead to  
56 noticeable changes of local morphology. Such changes may occur fortnightly, as it does  
57 for small systems (Fortunato et al., 2014), and at any larger timescale at which driving  
58 forces may change. For instance, if most barrier spits were able to build during periods

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3 59 with relatively low rates of sea-level rise (SLR), questions remain about how they would  
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6 60 respond to predicted higher rates of SLR and changing wave climate for the coming  
7  
8 61 decades (Cazenave and Le Cozannet, 2014; Semedo et al., 2013).

9  
10 62 There is indeed a persistent and immediate need to improve the prediction of climate  
11  
12 63 change impacts on coastal systems worldwide (Wong et al., 2014). Main drivers include  
13  
14 64 SLR, the alteration of rainfall and runoff and changing wave climates which are crucial  
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16  
17 65 for spits and inlets' evolution (Ranasinghe et al., 2012). In most coastal embayments,  
18  
19 66 SLR leads to larger values of tidal prisms that force inlets to increase their cross-section  
20  
21 67 (FitzGerald et al., 2008). In places with low continental sediment inputs, SLR would also  
22  
23 68 create accommodation space in embayments that may become sediment sinks for the  
24  
25  
26 69 adjacent coasts (Ranasinghe et al., 2012). Where freshwater runoff contributes to  
27  
28 70 maintain inlets, it may also cause adjacent spit retreat or progradation as rainfall  
29  
30  
31 71 respectively increases or diminishes (Ranasinghe et al., 2012). Finally, changing storm  
32  
33 72 tracks, storm intensity and frequency are likely to modify coastal resilience (Masselink et  
34  
35 73 al., 2016) and have a direct impact on the wave driven LST (Chowdhury and Behera,  
36  
37 74 2017; Splinter et al., 2012).

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40 75 Along with natural climate variability, these drivers have already varied over a wide  
41  
42 76 range of temporal and spatial scales. At inter-annual to decadal timescales,  
43  
44 77 atmospheric teleconnection patterns, such as the North Atlantic Oscillation (NAO), and  
45  
46 78 climatic cycles such as the El Niño Southern Oscillation (ENSO) over oceanic basins  
47  
48 79 have remote effects on coastal environments (Barnard et al., 2011; Masselink et al.,  
49  
50 80 2014; Wiggins et al., 2017). On longer timescales, the stratigraphic records of coastal  
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52 81 barrier spits have also revealed the control exerted by sea level changes (Clemmensen  
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3 82 et al., 2001; Fruergaard et al., 2015) and major storm events (Fruergaard et al., 2013).  
4  
5 83 This climate-driven variability can overlay the system internal cyclicity and play a part in  
6  
7 84 rhythmic behaviours that are common along barrier coasts (Allard et al., 2008;  
8  
9 85 Ridderinkhof et al., 2016). These cycles have substantial effects on shoreline dynamics.  
10  
11 86 Nonetheless, due to limited number of detailed geomorphological datasets, on the  
12  
13 87 longer term it is often unclear whether cyclicity is driven by the system intrinsic nature  
14  
15 88 (i.e., autocyclic) or by external (climatic) factors (i.e., allocyclic).  
16  
17 89 On the other hand, global knowledge on climate variability and its effects is constantly  
18  
19 90 growing. Coupled ocean-atmosphere numerical models help to describe processes  
20  
21 91 underlying relationships between parameters of different nature as, for instance, sea  
22  
23 92 level pressure, anomalous coldness and storm tracks (van der Schrier and Barkmeijer,  
24  
25 93 2005), solar irradiance and teleconnection patterns (Ineson et al., 2011) or  
26  
27 94 teleconnection patterns and sea level (Calafat et al., 2012). Going back to interactions  
28  
29 95 between barrier spits and inlets, recent studies address the impact of climate change  
30  
31 96 through the application of process-based morphodynamic models to tidal embayments  
32  
33 97 (Bruneau et al., 2011; Dissanayake et al., 2012; van Maanen et al., 2013; van der  
34  
35 98 Wegen, 2013). Those studies primarily discuss the effects of SLR and describe the  
36  
37 99 evolution of the sediment source-to-sink nature of inlets over time. Such growing  
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39 100 understanding of both the dynamic nature of climate and of the processes changing the  
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41 101 coastal landscape widens the spectrum of possibilities for interpreting documented  
42  
43 102 coastal evolutions.  
44  
45 103 This study thrives on these possibilities to discuss the coherence between large-scale  
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47 104 coastal changes and variable environmental conditions. The next section presents a  
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3 105 dataset with outstanding temporal coverage and resolution. It covers the morphological  
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5 106 evolution of the Cap Ferret sand spit (SW France, Atlantic coast) over the last 250 years  
6  
7 107 and contemporaneous environmental conditions. Then section 3 investigates the  
8  
9 108 synchronisation of the apparent cyclical dynamic of the spit distal end with climatic shifts  
10  
11 109 and sea level variations. Potential mechanisms underlying this synchronisation are  
12  
13 110 subsequently discussed based on current understanding of barrier spits and tidal inlets  
14  
15 111 morphodynamics, allowing to identify the dominant climatic drivers (section 4).  
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## 113 **Materials and methods**

### 114 ***Study area***

115 The Cap Ferret is a baymouth spit bordered by the Bay of Biscay (Figure 1). It lies at  
116 the southern end of a 110 km long uninterrupted sandy beach bounded by the Gironde  
117 estuary to the north (Aubié and Tastet, 2000; Castelle et al., 2018; Figure 1b). The  
118 subaerial fraction of the spit accounts for the beach last twenty kilometres and is built  
119 upon a subtidal platform that dips into the Bay of Arcachon tidal inlet (Figure 1c). This  
120 inlet connects a 160 km<sup>2</sup> coastal lagoon to the Atlantic Ocean, the lagoon – inlet system  
121 being the vestige of the Leyre River estuary (Féniès and Lericolais, 2005). Around 3 *ka*  
122 ago the Cap Ferret started to build up on the estuary northern margin (Féniès et al.,  
123 2010), pushing the river mouth southward until it semi-enclosed the Bay of Arcachon.  
124 At present, water circulation between the lagoon and the ocean constricts the spit  
125 progradation. Twice a day between 260 and 490 Mm<sup>3</sup> of water flow in and out through  
126 the inlet, confirming that at any time tidal exchanges largely take over on freshwater  
127 inputs (Plus et al., 2009). Such high values of tidal prism (*P*) largely overcome the total



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3 128 LST (M), estimated at 0.661 Mm<sup>3</sup> (Idier et al., 2013), so that this opening falls into the  
4  
5 129 “good stability” category ( $P/M \gg 150$ ) of (Bruun, 1978) classification. Nonetheless,  
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8 130 northwesterly dominant waves have deviated the estuary mouth some thirty kilometres  
9  
10 131 downdrift in the past 3 *ka* (Féniès et al., 2010). Annual mean significant wave height is  
11  
12 132 of 1.77 m (in 50 m water depth, according to Charles et al., 2012) and mean tidal range  
13  
14 133 is of 2.7 m (twice the semidiurnal tidal component amplitude, measured on the ebb-  
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16  
17 134 delta’s shield, by Senechal et al., 2013). This mixed-energy environment, as defined by  
18  
19 135 Hayes (1979), has moulded an inlet with a transitional morphology (Hubbard et al.,  
20  
21 136 1979). The inlet ebb-tidal delta is more developed than the flood-tidal shoals (Figure  
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23  
24 137 1c). This disequilibrium reflects the ebb dominance confirmed by morphodynamic  
25  
26 138 modelling (Cayocca, 2001). Model results further showed the role of the ebb-dominant  
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28 139 tide in the breaching of new channels across the spit platform. Indeed, historical charts  
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31 140 report various occurrences of the formation of a new channel across the platform. Over  
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33 141 the last 300 years, newly opened channels have migrated southward and the detached  
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35 142 shoals ultimately reached and merged to the inlet southern margin, in a movement that  
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38 143 resembles the spit-platform breaching model of sediment by-passing of FitzGerald et al.  
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40 144 (2001). The apparent cyclical nature of this process has led to the hypothesis of an 80-  
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42 145 year autocyclic behaviour. According to charts, the inlet has been alternatively  
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45 146 composed of one or two channels and Michel and Howa (1997) further interpreted the  
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47 147 retreat of the subaerial fraction of the spit as a feedback from the breaching of new  
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49 148 (secondary) channels. However, a thorough discussion of the role of variable  
50  
51 149 environmental conditions is still lacking. The premises of such a discussion were given  
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54 150 by Nahon et al. (2015). In this first analysis, it was noted that the autocyclic model fails

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3 151 to predict a major spit progradation event. Instead, spit elongation was found  
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5 152 synchronous with periods dominated by the negative phase of the NAO.  
6  
7 153 The present study builds upon Nahon et al. (2015) data to further infer on the role that  
8  
9 154 climate and SLR may have played in inlet – spit interactions. Cap Ferret is well suited  
10  
11 155 for this study because of 3 reasons: 1) updrift sandy beaches prevent from any  
12  
13 156 sediment deficit; 2) it is backed up by a mature estuary where an elevated level of  
14  
15 157 infilling makes tidal prism sensitive to mean sea level variations; 3) the first  
16  
17 158 quantitatively valuable chart dates from 1768 and the evolution of the spit-inlet system  
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19 159 can be reconstructed over the last 2.5 centuries.  
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### 161 ***Shoreline data***

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28 162 The geomorphological record encompasses the entire 19<sup>th</sup> and 20<sup>th</sup> centuries and  
29  
30 163 consists of nautical charts and aerial photographs of the Bay of Arcachon's tidal inlet  
31  
32 164 (Figures 2 to 4).

33 165 Charts were retrieved from Caspari (1872), Lapeyre (1925) and Bordeaux harbour  
34  
35 166 authorities (PAB, 1985). All made appear 3 reference positions on charts (Figure 2): the  
36  
37 167 still existing Cap Ferret's lighthouse and the church *Notre-Dame d'Arcachon*, both  
38  
39 168 located north of the spit distal end, and a former semaphore on the southern margin.  
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41 169 This later position was retrieved from the map of the Gironde department in 1888  
42  
43 170 (Figure 2a). These positions were used for georeferencing.

44  
45 171 Clear and stable features along the spit were used to estimate the error on the charts  
46  
47 172 before 1900. Preserved former spit-end positions and small bights on the lagoon side of  
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49 173 the spit, indicate a distance error well below 200 m. Therefore, a +/- 200 m error bars  
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3 174 marks off the results to avoid any misinterpretation. Early 20<sup>th</sup> century charts also  
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5 175 indicate a (still existing) semaphore located 1 km north of the 2014 shoreline. The root  
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7 176 mean square error for the semaphore positions is 55 m and results are also shown with  
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9  
10 177 +/- 100 m error bars to account for possible shoreline misinterpretation at the time of the  
11  
12 178 survey. The spit terminus was then defined as the southernmost position of the  
13  
14 179 coastline represented on charts (Figure 3). These positions were orthogonally projected  
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16  
17 180 onto the axis defined by 2 of the reference positions used for georeferencing nautical  
18  
19 181 charts (Figure 3).  
20  
21 182 Positions measured on charts from 1932 and 1936 presented a very good match with  
22  
23 183 those on photographs from 1934 and 1946 respectively (Figure 4). Between 1934 and  
24  
25 184 2000, aerial photographs were used. They were georeferenced using the current road  
26  
27 185 and pathway network and, at some point, World War II bunkers on top of the dunes.  
28  
29 186 After 2000, 7 high resolution orthophotomaps were used. Georeferencing errors are  
30  
31 187 lower than the shoreline photo-interpretation error that was estimated well below +/- 50  
32  
33 188 m. On photographs, the southernmost position of the interpreted shoreline (i.e. berm  
34  
35 189 crest when perceptible, high-tide wrack line otherwise) was taken to measure the spit  
36  
37 190 extension. Overall, 50 positions of the Cap Ferret distal end were retrieved since 1768  
38  
39 191 (24 on charts from 1768 to 1936, Figure 3; 26 on photographs from 1934 to 2014,  
40  
41 192 Figure 4). These positions were used to trace the path of the spit distal-end. In the  
42  
43 193 following, periods over which the spit advanced across the inlet were identified from this  
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45 194 path. Still, it is a limited indicator of erosion or accretion patterns as it does not consider  
46  
47 195 the behaviour of the spit in the direction perpendicular to the reference axis.  
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197 ***Bathymetric data***

198 Shoreline data document north-south oscillations of the spit distal end, the last large-  
199 scale oscillation occurring between 1950 and 2014 (Figure 4). Over this period,  
200 bathymetric evolution of the adjacent inlet throat was quantified using soundings from  
201 years 1949, 1969, 1979, 2001 and 2014.

202 Over the years, along transect resolution increased from 100 m to 1 m and transect  
203 positioning and spacing (~200m) remained constant (Figure 5a). For each survey,  
204 transects have been interpolated into digital elevation models (DEMs); 50 m x 50 m  
205 surfaces were created using a nearest-neighbour interpolation (Figure 5b). Bottom  
206 elevation and volumetric changes were computed over a 12.74 km<sup>2</sup> overlapping surface  
207 between all 5 DEMs. Figure 5d presents the erosion-deposition patterns over the 65  
208 years.

209 A distinction was made between shoals and channels using the -7 m NGF contour  
210 (NGF is the French vertical datum; locally Mean Sea Level is currently around +0.40 m  
211 NGF). This contour match well with shoals visible on contemporary Landsat 8 image  
212 (Figure 5c). For each date, 4 quantities were computed. The first 3 are the mean  
213 elevation and the shoal and channel volumes which are respectively the sand volume  
214 above -7 m NGF and water volume below -7 m NGF. As little to none information exists  
215 to estimate errors due the instrumental bias or vertical datum changes over time (others  
216 than the known and corrected ones), a fourth quantity, defined as the inlet throat  
217 *morphological amplitude* was calculated. It corresponds to the equal volumes of sand  
218 and water, respectively above and below the mean elevation of each DEM. In this way,  
219 this later quantity is free of any artificial variations. Results are summarised in Table 1.

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3 220 Figure 5 (e-h) further presents the variation of the computed quantities with the  
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5 221 identified periods of elongation and retreat of the spit-end.  
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9  
10 223 **Sea level data**

11  
12 224 To infer on the role of sea level variations in the evolutions depicted by cartographic and  
13  
14 225 bathymetric data, the annual mean sea level (MSL) record from Brest tidal gauge  
15  
16 226 (Figure 1) was retrieved from the permanent mean sea level observatory (Holgate et al.,  
17  
18 227 2013); it is the nearest tidal gauge with an appropriate temporal coverage (Wöppelmann  
19  
20 228 et al., 2008). Over the entire period the data is fitted to a 2<sup>nd</sup> order polynomial function  
21  
22 229 which reveals the lowering of MSL during the first third of the 19<sup>th</sup> century (Figure 6,  
23  
24 230 upper panel). Then over the 20<sup>th</sup> century, the data is both averaged and fitted to a  
25  
26 231 polynomial function. The running mean over an 11-year-centred window is used to  
27  
28 232 emphasise the significant decadal variability of the regional MSL. Then, fitting the record  
29  
30 233 with 5<sup>th</sup> order polynomial function allows identifying a period of slower SLR around the  
31  
32 234 second third of the 20<sup>th</sup> century (Figure 6, lower panel).

33  
34 235 Table 2 further indicates average SLR rates over identified periods of spit growth and  
35  
36 236 decay. Rates are computed as the linear trend in the overlapping sea level data. The  
37  
38 237 overlap is defined by adding sea level data until preceding and following spit-end  
39  
40 238 positions to the actual identified period (excepted for 1909 because of the important 20-  
41  
42 239 year interval between positions). This lengthens the periods over which the trend is  
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44 240 computed. When computed over the exact (shorter) intervals, differences between  
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46 241 successive periods of growth and decay are more pronounced.  
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243 **Wave climate data**

244 To assess the impact from variations of wave-driven forces, the local wave climate  
245 during the last oscillation of the spit-end was characterized by means of a hindcast  
246 simulation, performed with the storm surge modelling system of Bertin et al. (2015). The  
247 model was extended to the whole North Atlantic Ocean and forced with wind fields  
248 originating from the NCEP/NCAR reanalysis (Kalnay et al., 1996), over the period 1949-  
249 2014. Hindcasted wave parameters were validated against measurements from a  
250 directional wave buoy moored in 54 m water depth, and located 15 km offshore the  
251 study site (CETMEF, n.d.). Between 2007 and 2014, 92602 measurements,  
252 representative of 5.3 years in record length, were compared to interpolated modelled  
253 parameters. For both modelled and measured parameters, the wave power (WP) is  
254 approximated with the linear wave theory (Svendsen, 2006; Eq. 1-2), as the wave  
255 energy (E) times the wave group velocity ( $c_g$ ), after computing the phase velocity (c)  
256 using an iterative method to solve the dispersion relationship for calculating (in  
257 intermediate water depth) the wave number (k) associated with the peak angular  
258 frequency ( $\omega$ ).

260 Eq. 1:  $WP = Ec_g$

261 Eq. 2:  $E = \frac{1}{8}\rho g Hs^2; c_g = c * \frac{1}{2} \left(1 + \frac{2kh}{\sinh 2kh}\right); c = \frac{\omega}{k}$

262  
263 The WP is then decomposed into cross-shore (WPx) and alongshore (WPy) energy flux,  
264 positive when landward and northward respectively. Table 3 summarizes the  
265 comparison outcomes for the significant wave height (Hs), the WP and mean wave

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2  
3 266 energy direction ( $wDir$ ), computed with mean  $WP_x$  and  $WP_y$ . Modelled and observed  
4  
5 267 parameters are in good agreements. Pearson linear correlation coefficient are equal to  
6  
7 268 0.93 and 0.80 for respectively  $WP$  and  $WP_y$  and the modelled wave field only slightly  
8  
9 269 underestimating the total  $WP$  by 2.4%. In terms of direction,  $wDir$  has an initial bias of  
10  
11 270 3.8°. After removing the bias, correlation between modelled and observed  $WP_y$  is  
12  
13 271 improved to 0.81 and the error on averaged  $WP_y$  is minimal (1.7%). Low-pass filtered  
14  
15 272  $WP$  and  $WP_y$  on Figure 7 further reveal the good performance of the model, as  
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17 273 correlation coefficients grow to 0.99 and 0.95 for  $WP$  and  $WP_y$  respectively.  
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### 23 24 275 ***North Atlantic Oscillation (NAO)***

25  
26 276 Over the entire period covered by geomorphological data, winter indices of the North  
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28 277 Atlantic Oscillation (NAO) provide additional information about the environmental  
29  
30 278 forcing. The NAO is the main atmospheric mode of variability over the North Atlantic  
31  
32 279 basin in winter (Hurrell and Deser, 2009), and from December to March, its negative  
33  
34 280 and positive phases have a demonstrated influence on ocean waves and sea level  
35  
36 281 anomalies on Western Europe coastlines (Calafat et al., 2012; Dodet et al., 2010).  
37  
38 282 Therefore, 3 different NAO indices were used as proxy for these environmental drivers.  
39  
40 283 Indices of the winter NAO in the literature are either reconstructions based on a  
41  
42 284 combination of instrumental and proxy data (Cook et al., 2002; Glueck and Stockton,  
43  
44 285 2001; Luterbacher et al., 1999; Ortega et al., 2015), indices computed from the  
45  
46 286 difference of normalized sea level pressure between the Azores High and the Icelandic  
47  
48 287 Low (Hurrell, 1995; Jones et al., 1997), or indices based on a principal component  
49  
50 288 analysis of reconstructed sea level pressure fields over the North Atlantic Ocean  
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3 289 (Barnston and Livezey, 1987; Hurrell, 2019). This study used the station-based index of  
4  
5 290 Hurrell (2015) which provides empirical evidence until 1864. This record is extended  
6  
7 291 back to 1768, with two proxy-based reconstructions selected for their robustness  
8  
9  
10 292 against this instrumental data (Luterbacher et al., 2001; DJFM index, accessed from  
11  
12 293 <https://www.ncdc.noaa.gov/paleo-search/study/6275>) and their novelty (Ortega et al.,  
13  
14 294 2015; DFJ index, accessed from <https://www.ncdc.noaa.gov/paleo-search/study/18935>).  
15  
16  
17 295 In the following, these indices are either showed in their cumulative form or as decadal  
18  
19 296 averages. To highlight the dominance of either the negative or positive phase of the  
20  
21 297 NAO in winter, cumulative indices were computed. After Mazzarella and Scafetta  
22  
23 298 (2012), the cumulative value of each index for a given year is taken as the inclusive sum  
24  
25  
26 299 of values for all previous years. Then over the 20<sup>th</sup> century onward, more frequent spit-  
27  
28 300 end positions are compared to the station-based index, averaged over the 10 years  
29  
30  
31 301 preceding each observation. This averaging procedure was needed because of the  
32  
33 302 irregularly sampled observations; the 10-year window was found to fit well with this  
34  
35 303 irregular resolution and the inertia of the system compared to the high interannual  
36  
37 304 variability of the NAO (Hurrell, 1995).  
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## 306 **Results and basis for discussion**

### 307 ***Three periods of spit lengthening***

308 Upper panel in Figure 8 represents the path of the Cap Ferret spit-end from 1768 to  
309 2014; lower panels show cumulated NAO indices. The spit-end path reveals 3 periods  
310 over which the spit has dominantly grown across the inlet. These periods, over 40 years  
311 each, are delimited by local extrema of the spit-end position, on maps and photos of



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3 312 1768-1826, 1865-1909 and 1932-1972. The 1<sup>st</sup> and 3<sup>rd</sup> periods are found when the  
4  
5 313 negative phase of the NAO has dominated (Figure 8, lower panels). Likewise, according  
6  
7 314 to the station-based index (Hurrell, 2015), the negative phase of the NAO has  
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9  
10 315 dominated during most of the 2<sup>nd</sup> period, until 1902, or 7 years before the map of 1909.  
11  
12  
13 316

### 14 317 ***The Dalton Minimum***

16  
17 318 The 1<sup>st</sup> period of spit lengthening, characterized by a 3.5 km lengthening of the spit  
18  
19 319 between 1805 and 1826, was also contemporaneous with the Dalton Minimum (DM,  
20  
21 320 1780-1830) climate anomaly. Figure 9 details this record lengthening and the  
22  
23 321 subsequent retreat. The spit has grown while the tidal inlet had 2 well defined channels.  
24  
25  
26 322 Both northern and southern channels were preserved as the spit grew, but the inlet  
27  
28 323 minimal width was divided by 3 in 1826 compared to that of 1768. Then between 1826  
29  
30 324 and 1865, the spit has retreated while the inlet width regained a width comparable to its  
31  
32 325 1768 level. In 1865 the southern channel was buried by the main inlet shoal that have  
33  
34 326 steadily migrated southward since 1768 attached to the inlet's southern margin.  
35  
36  
37 327 Therefore, during a century, the spit has advanced against a double channelled inlet  
38  
39 328 and has retreated when the inlet's southern channel closed. In terms of climate, this  
40  
41 329 century encompassed the Dalton Minimum which was a period of anomalous coldness  
42  
43 330 (van der Schrier and Barkmeijer (2005) and references therein).  
44  
45  
46 331 DM is explained as a conjunction of low solar activity and major volcanic eruptions and  
47  
48 332 has culminated in Europe with "the year without a summer" in 1816 (van der Schrier and  
49  
50 333 Barkmeijer, 2005). Using a global circulation model, van der Schrier and Barkmeijer  
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52 334 (2005) have pointed out the "higher occurrence probability at the European mid-  
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3 335 latitudes of the strongest cyclone under DM atmospheric conditions”. Their results recall  
4  
5 336 those from Shindell (2001) and Ineson et al. (2011), who showed that periods, or years,  
6  
7 337 with low solar activity produce NAO negative like atmospheric configurations. Indeed,  
8  
9 338 during the negative phase of the NAO low pressure systems are deviated southward  
10  
11 339 (Hurrell and Deser, 2009). Furthermore, there is converging evidence of enhanced mid-  
12  
13 340 latitude storminess during such cold periods, as recorded by the dune fields north and  
14  
15 341 south of Cap Ferret (Clarke et al., 2002).

16  
17 342 The dominance of the negative phase of the NAO during DM is also suggested by the  
18  
19 343 cumulative NAO winter-index of Luterbacher et al. (2001), that decreased until 1830  
20  
21 344 (Figure 8, mid panel). However, as recalled by Poirier et al. (2017), NAO reconstruction  
22  
23 345 tend to diverge in this period. For instance, the data from Ortega et al. (2015) only  
24  
25 346 confirm Luterbacher’s data until 1810 when, according to the more recent  
26  
27 347 reconstruction, the positive phase begins to dominate again. On the other hand, when  
28  
29 348 compared to the station-based index computed by (Hurrell, 2015) from the instrumental  
30  
31 349 record, Luterbacher’s reconstruction performs better. Indeed, Luterbacher’s (DJFM)  
32  
33 350 reconstruction is well correlated with the station-based index from 1864 onward (Figure  
34  
35 351 8 mid panel), while the correlation between Ortega’s (DJF) reconstruction and  
36  
37 352 instrumental data between 1864 and 1969 is weaker (Figure 8 lower panel); Pearson  
38  
39 353 linear correlation coefficient ( $r$ ) equal to 0.89 ( $p$ -value  $< 10^{-47}$ ) in the first case, against a  
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41 354 correlation coefficient of 0.47 ( $p$ -value  $< 10^{-6}$ ) in the second case.  
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51 356 ***Correlation with NAO winter-indices***

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3 357 After the DM, the other 2 periods of spit growth happened while the negative phase of  
4  
5 358 the NAO have dominated in winter, as supported by Hurrell's (2015) station-based index  
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7 359 (Figure 8, middle panel). By opposition, periods of spit retreat systematically took place  
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9  
10 360 in periods dominated by the positive phase of the winter NAO; on Figure 8, periods  
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12 361 delimited by the local extrema of the spit-end position, on maps and photos of 1826-  
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14 362 1865, 1909-1932 and 1972-2014.

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16  
17 363 Over the 20<sup>th</sup> century, sharp movements of the spit-end like those around 1915 and  
18  
19 364 1972 were also synchronous with remarkable shifts of the NAO (Figure 10, upper  
20  
21 365 panel): between 1909 and 2000, the spit-end path is significantly correlated with the  
22  
23 366 decadal average of the NAO winter-index ( $r = -0.67$ ,  $p\text{-value} < 10^{-4}$ ).

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25  
26 367 Locally, the positive phase of NAO is known to produce higher and more oblique  
27  
28 368 (clockwise shift) winter waves (Charles et al., 2012). The hindcast used in this study  
29  
30 369 agrees with this and further shows how decadal averages of the winter NAO index and  
31  
32 370 the alongshore wave power are well correlated (Figure 10, lower panel;  $r = 0.86$ ,  $p\text{-value}$   
33  
34 371  $< 10^{-16}$ ). Correlation is greater than for the total wave power (WP), which has a  
35  
36 372 correlation coefficient of 0.76 ( $p\text{-value} < 10^{-11}$ ). This highlights the impact of the NAO on  
37  
38 373 the local wave direction as well. Consequently, according to wave climate between  
39  
40 374 1949 and 2014, decades dominated by the positive phase of the NAO may produce an  
41  
42 375 average alongshore wave power up to 30% greater than when the negative phase  
43  
44 376 dominates (figure 10, lower panel).

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47 377 Nonetheless, although winter wave power and NAO index remain correlated until 2014,  
48  
49 378 their apparent relationship with the spit-end track is less clear at the beginning of the  
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3 379 21<sup>st</sup> century. Since 2000, averaged wave power and NAO index have regained a more  
4  
5 380 neutral value while the spit have continued to retreat.  
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10 382 ***Sea level and inlet width variations***

11  
12 383 The ongoing erosion and retreat of the spit-end comes after the last significant  
13  
14 384 lengthening which has culminated in 1972 (Figures 4 & 10). In addition to a dominant  
15  
16 385 negative phase of the NAO in winter (Figures 8 & 10), the spit has then advanced  
17  
18 386 across the inlet while, since the 1930s, sea level was rising at a relative slow pace (1.26  
19  
20 387 mm.y<sup>-1</sup>) compared to that of the beginning and end of the 20<sup>th</sup> century (above 2.4 mm.y<sup>-1</sup>  
21  
22 388 1; see Table 2 and Figure 6 lower panel). During this period, the inlet channel volume  
23  
24 389 and the morphological amplitude of the inlet throat were also relatively stable or slightly  
25  
26 390 decreased (Figure 5).  
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31 391 Around 1972, the spit-end growth was sharply interrupted. In the same time the  
32  
33 392 morphological amplitude of the inlet throat remained stable until 1979. Instead, as the  
34  
35 393 winter NAO abruptly shifts towards the dominance of the positive phase, the decadal  
36  
37 394 averaged of the alongshore wave power increased and reached its highest level  
38  
39 395 between 1970 and 1990. It is only after 1979 that the inlet channel volume has begun to  
40  
41 396 increase, synchronous with an acceleration of sea level rise (Figure 6). Since then  
42  
43 397 channel volume has increased over 25%, while the wave energy remains at an average  
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45 398 level.  
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48  
49 399 The increase in channel volume recalls the dramatic expansion of the inlet observed  
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51 400 after 1826 (Figure 9). This first *breathing* then occurred as sea level has successively  
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53 401 fallen and risen (Table 2). In the absence of detailed analysis of the evolution of the  
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3 402 inlet's dimensions over the second period of spit lengthening, linear trends presented in  
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5 403 Table 2 reveal that for all 3 periods of spit lengthening, sea level has either fallen or has  
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8 404 risen slower than during preceding and following periods.  
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For Peer Review

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3 405 **Discussion**

4  
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6 406 ***The auto-cyclic hypothesis***

7  
8 407 From the 1960s to until recently (2014), the present set of maps and photos has been  
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10 408 gathered and analysed numerous times by several authors, mainly for engineering  
11  
12 409 purposes (see Nahon, 2018, for references). The main idea which raised is that Cap  
13  
14 410 Ferret's spit-end north-south oscillations could be an auto-cyclic response to the  
15  
16 411 apparent cyclicity of the inlet channel configuration (with one or two channels). Michel  
17  
18 412 and Howa (1997) detailed the cyclical nature of the inlet channels and conceptualized  
19  
20 413 its impacts on the inlet's northern and southern margins. Among other impacts, their  
21  
22 414 conceptual model predicts that the Cap Ferret spit extends southward when a single  
23  
24 415 channel exists and by opposition retreats when two channels split the inlet. This  
25  
26 416 feedback interaction was derived from approximately the same cartographic data  
27  
28 417 presented in here. However, such hypothesis is far less than evident during over a third  
29  
30 418 of the data coverage. Indeed, as shown in Figure 9, let's recall here that between 1769  
31  
32 419 and 1825, the spit has elongated while the inlet had two channels and that from 1825 to  
33  
34 420 1865, the spit has retreated when the southern channel was progressively buried. On  
35  
36 421 the other hand, there is an apparent relationship between the spit-end behaviour and  
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38 422 the North Atlantic climate over the entire study period.  
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45 423  
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47 424 ***Wave climate influence***

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49 425 Links between the NAO and the local wave climate could be part of the explanation to  
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51 426 the spit – climate apparent relationship. Indeed, it appears that the spit grows across the  
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53 427 inlet when NAO-negative ocean waves are more shore normal and/or the alongshore  
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3 428 wave power is below average. By opposition, the spit-end retreats when the alongshore  
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5 429 wave power is stronger than average, and as so when the wave-driven alongshore  
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7 430 sediment transport (LST) is more intense. Aagaard et al. (2004) have observed a similar  
8  
9 431 relationship along the western Danish coast in northern Europe. Based on wind data  
10  
11 432 from the first and last 30 years of the 20<sup>th</sup> century, they have pointed out the contribution  
12  
13 433 of increased wind-wave driven LST, to the updrift erosion of the spit-end towards the  
14  
15 434 end of the century. More recently, Aagaard and Sørensen (2013) have suggested the  
16  
17 435 increase of the LST surpasses all terms in the sediment *imbalanced* equation, including  
18  
19 436 sea-level rise. Underlying physical processes are yet to be identified, however they  
20  
21 437 could either lie within LST – inlet flow interactions (Bertin et al., 2009), or within wave  
22  
23 438 orientation – spit-growth relationship (Ashton et al., 2016).  
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25  
26 439 On the one hand, Bertin et al. (2009) have shown with a process-based morphodynamic  
27  
28 440 model, that sediment retention within tidal inlets could increase when waves approach  
29  
30 441 the coast at a lower angle of incidence; therefore when, to all other parameters equal,  
31  
32 442 LST is reduced. Locally, this is quite meaningful because the nonlinear superimposition  
33  
34 443 of the southward LST induced waves and tidal residuals increase alongshore gradients  
35  
36 444 in sediment transport (Cayocca, 2001).  
37  
38  
39 445 On the other hand, simulations with coastal evolution models (CEM) suggest that the  
40  
41 446 sole relationships between spit growth and wave orientation may also be at play  
42  
43 447 (Ashton et al., 2016): less oblique waves (and a decreased LST) leads to the erosion of  
44  
45 448 the spit on the updrift side of the fulcrum point and to the lengthening of spit distal end  
46  
47 449 (Figure 11a), whereas increasing the wave incidence and the LST induces sediment  
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49 450 retention updrift of the fulcrum point while the spit-end loses sediment and retreats  
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3 451 (Figure 11b). This apparent oscillation of the sediment budget around the fulcrum point  
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5 452 recalls the behaviour of the Cap Ferret between 1826 and 1865 (Figures 9 & 11c).  
6  
7 453 These examples provide some insights on the physics underlying the apparent  
8  
9 454 relationship between local geomorphology and local waves. Increased LST and more  
10  
11 455 oblique waves during the positive phase of the NAO may well have forced the spit-end  
12  
13 456 to retreat, while the spit can advance across the inlet when the negative phase of the  
14  
15 457 NAO produce waves with a reduced angle of incidence and lower rates of LST.  
16  
17 458 From a regional perspective, this behaviour differs from that of the Arçay sandspit (less  
18  
19 459 than 250 km north of Cap Ferret). Allard et al. (2008) first suggested the rhythmic  
20  
21 460 growth of Arçay sandspit was boosted by LST, and Poirier et al. (2017) further put into  
22  
23 461 evidence the hierarchical control of the NAO and the East Atlantic–West Russia pattern.  
24  
25 462 In the case of Arçay, higher rates of LST during the positive phase of the NAO are  
26  
27 463 found to be responsible for the enhanced spit growth. Here we found it is the opposite at  
28  
29 464 Cap Ferret. Therefore, the attempt by Poirier et al. (2017) to, in a second time,  
30  
31 465 associate both spits behaviour is questionable. The reason may be the important  
32  
33 466 difference of their back-barrier lagoon dimensions, which in the case of Arçay is a lot  
34  
35 467 smaller. In the case of Cap Ferret, the large lagoon of the Bay of Arcachon engenders a  
36  
37 468 large tidal prism that has shaped large inlet shoals. Then, the repartition of wave-driven  
38  
39 469 sediment inputs, between the spit and the shoals, become more complex (Hoan et al.,  
40  
41 470 2011; Kraus, 2000; Larson et al., 2007). This is particularly true when relative sea levels  
42  
43 471 are subject to rise, ultimately turning shoals into sediment sinks because their  
44  
45 472 equilibrium volume increases with the larger value of tidal prism engendered by higher  
46  
47 473 sea level (Walton, Jr. and Adams, 1976). Also, wave climate variations alone may not  
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3 474 fully explain inlet constrictions and expansions such as those observed during and after  
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5 475 the Dalton Minimum (Figure 2c-d), or the ongoing inlet throat expansion since 1979  
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7  
8 476 (Figure 5).

9  
10 477 **Sea level influence**

11  
12 478 A tidal inlet's cross-section increases with tidal prism (O'Brien, 1931) and, to a lower  
13  
14 479 degree, decreases with wave energy (Jarrett, 1976; Nahon et al., 2012). Tidal prism  
15  
16  
17 480 variations could then have contributed to observed changes in inlet dimensions. For  
18  
19 481 instance, in 1990s when the wave energy is maximal (Figure 10, lower panel), only an  
20  
21 482 increase of the tidal prism could have caused the inlet to enlarge (Figure 5e). In  
22  
23 483 addition, the magnitude of the inlet's narrowing and widening in the first half of the 19<sup>th</sup>  
24  
25 484 century (Figure 9) suggests that changes in the tidal prism have certainly added to the  
26  
27 485 effects of variations of the wave climate associated with the Dalton Minimum. Over the  
28  
29 486 study period, the Bay of Arcachon's contours remained stable. Instead, variations of the  
30  
31 487 sea level must have modulated tidal prism.

32  
33 488 Tidal flats and salt marshes occupy about 75 % of the Bay of Arcachon (Féniès and  
34  
35 489 Faugères, 1998; Nahon, 2018), so that the tidal prism is mostly controlled by the relative  
36  
37 490 elevation of those. Due to low sediment input from freshwater streams, the  
38  
39 491 sedimentation rate above these tidal flats is expected to respond, at most, with a  
40  
41 492 temporal lag to changes in the rate of sea level variations (Kirwan and Murray, 2008).  
42  
43 493 So that when sea level starts to increase or when sea level rise (SLR) accelerates, tidal  
44  
45 494 flats' relative elevation decreases temporarily, leading to a greater tidal prism that  
46  
47 495 ultimately forces the inlet to enlarge. On the contrary, inlets may narrow in response to  
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49 496 falling sea level or stable to decelerating SLR.  
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3 497 Part of the explanation would also be related to the ongoing morphological evolution  
4  
5 498 whole barrier system. Indeed, the large tidal flats of the lagoon and the relatively high  
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7 499 tidal range promote the ebb dominance of the system (Fortunato and Oliveira, 2005;  
8  
9 500 Friedrichs and Aubrey, 1988). The overall negative sediment budget of the lagoon  
10  
11 501 between charts of 1865 and 2001 (Allard et al., 2009) confirmed this characteristic.  
12  
13 502 Charts further revealed that sediment loss was due to headward erosion of its tidal  
14  
15 503 channels which, according to process-based models, can be a response of an inlet –  
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17 504 lagoon system to SLR (van Maanen et al., 2013). This similitude between simulated  
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19 505 morphologies and the observed evolution of the lagoon, suggests the Bay of Arcachon  
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21 506 is currently adapting to secular regional SLR (Jevrejeva et al., 2014). Van Maanen et al.  
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23 507 (2013) numerical experiments further explain how, under progressive SLR, such  
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25 508 systems can remain ebb-dominated and how the amount of exported sediment  
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27 509 increases with the rate of SLR.  
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33 510 Therefore, multiple processes could explain how decadal to pluri-decadal variations in  
34  
35 511 the rate of SLR (Jevrejeva et al., 2014), like those observed at Brest (Figure 6 and  
36  
37 512 Table 2), must have impacted the inlet sediment budget, at temporal scales matching  
38  
39 513 those of documented spit-end oscillations.  
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#### 514 515 ***Other influences***

516 In winter, the NAO also modifies effective rainfalls and wind regimes as well as sea  
517 level anomalies (Calafat et al., 2012). Locally, NAO-positive winters tends to be drier  
518 (Hurrell and Deser, 2009), so that freshwater inputs to the lagoon are expected to be  
519 further reduced. Therefore, it is very unlikely that the impact of the NAO on precipitation

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3 520 could contribute to the observed inlet breathing and spit oscillation. However, the  
4  
5 521 morphodynamic impact of positive sea level anomalies during NAO negative winters  
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7 522 (Calafat et al., 2012) and how it may interact with longer term trends of SLR is an open  
8  
9  
10 523 question. As is the impact of the complex wind-induced circulation (Salles et al., 2015)  
11  
12 524 on inlet morphodynamics. Nonetheless, these questions fall beyond the scope of the  
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14  
15 525 present work.  
16  
17 526

### 19 527 **Synthesis**

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21 528 Despite these latter questions, a combination of processes exists that may well have  
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23 529 contributed to the north-south oscillations of the spit-end observed on charts and aerial  
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26 530 photographs, between 1768 and 2014.

27  
28 531 Variations in the wave climate associated with North Atlantic atmospheric state, very  
29  
30 532 likely modified sediment transport patterns near the spit-end; and trends in sea level  
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32 533 variations have modulated inlet dimensions, ultimately forcing the spit-end to retreat  
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34  
35 534 when the inlet enlarged faster than it was migrating and allowing the spit-end to advance  
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37  
38 535 otherwise. These two drivers make it possible to explain the spit decays and growths  
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40 536 over the entire study period and at the temporal scale resolved by that cartographic  
41  
42 537 data. Identifying a dominant mechanism is tricky though.

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44 538 On the one hand, since the turn of the 20<sup>th</sup> century, the spit has then been able to  
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46  
47 539 durably grow when the rate of SLR was moderate, as between 1930 and 1970.

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49 540 Otherwise, when sea level has risen faster, like before 1930 or since 1970, the spit-end  
50  
51 541 has been unstable or has retreated. This indicates the existence of an upper limit to the  
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54 542 rate of SLR (around 2 mm.y<sup>-1</sup> according data in Table 2), above which spit growth is

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3 543 annihilated. On the other hand, even though regional SLR remained moderate until the  
4  
5 544 late 1970s, the intensification of the alongshore wave power alone have trigger the spit  
6  
7 545 retreat in the early 1970s. The next step would then be to implement a numerical  
8  
9 546 morphodynamic model to rigorously evaluate the respective and relative contribution of  
10  
11 547 both forcing.  
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14 548 Finally, independently from the existence of a dominant mechanism, the exceptional  
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16 549 evolution during the first half of the 19<sup>th</sup> century was certainly produced by in-phase  
17  
18 550 destructive and constructive forces: from the spit point of view, it must have resulted  
19  
20 551 from synchronous constructive forces, exacerbated by the Dalton Minimum, followed by  
21  
22 552 synchronous destructive forces. This certainly agrees with the duality of wave climate  
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24 553 and sea level relations in terms of geomorphological impacts. However, it does not  
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26 554 answer the question of the interplay between these two drivers.  
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## 32 33 556 **Conclusions**

34  
35 557 Detailed observations of the Cap Ferret spit-end over the last 250 years reveal  
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37 558 geomorphological changes of a remarkable magnitude. Owing to the temporal  
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39 559 coverage, it seems possible to explain the apparent cyclical nature of this dynamic by a  
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41 560 combination of climatic shifts and sea-level variations, although without considering an  
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43 561 eventual relationship between these two drivers.  
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46 562 First, all documented phases of spit growth were found during periods dominated by the  
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48 563 negative phase of the North Atlantic Oscillation in winter. Inversely, the spit has  
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50 564 retreated when the NAO shifted towards a positive-phase dominance. The relationship  
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52 565 between NAO and local wave climate make the wave climate a good candidate to  
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3 566 explain this behaviour. During the positive phase of the NAO, higher ocean waves  
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5 567 and/or clockwise shifts of the waves' mean direction accelerate alongshore sediment  
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7 568 transport (LST), which, counter intuitively and like it has been observed and modelled  
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9  
10 569 elsewhere, contribute the updrift erosion of the spit-end.  
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12 570 In the second place, spit-end retreats and instabilities were also in phase with periods of  
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14 571 rapid or accelerating sea-level rise SLR. Synchronous expansions of the adjacent inlet  
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16 572 pointed at the impact of SLR on tidal prism. Indeed, enhanced tidal prism under  
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18 573 accelerating SLR may well exacerbate the ongoing adaptation of the whole Bay of  
19  
20 574 Arcachon to secular SLR, ultimately forcing the spit-end to retreat when the inlet  
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22 575 enlarges faster than it migrates downdrift.  
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26 576 Locally, these findings advocate a dominance of allocyclic mechanisms over the  
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28 577 autocyclic behaviour proposed by Michel and Howa (1997). Somehow this goes along  
29  
30 578 with the results of Allard et al. (2008) and Poirier et al. (2017), although the opposed  
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32 579 behaviour of Cap Ferret and Arçay spits regarding wave climate suggest that coastal  
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34 580 barrier spits response to increased LST may differ regarding if SLR is or not turning  
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36 581 spits' adjacent shoals into sediment sinks.  
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38  
39 582 In a broader perspective, the present results further highlight the vulnerability to SLR of  
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41 583 barrier-spit backed by large estuaries. Given this, they may also serve to detect coastal  
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43 584 adjustments driven by climate change.  
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19 597 [aquitaine.fr/](http://www.observatoire-cote-aquitaine.fr/)).  
20  
21  
22  
23  
24  
25

## 26 599 **References**

- 27  
28  
29 600 Aagaard T, Nielsen J, Jensen SG, Friderichsen J. 2004. Longshore sediment transport and coastal erosion at  
30 601 Skallingen, Denmark. *Geografisk Tidsskrift-Danish Journal of Geography* **104** : 5–14. DOI:  
31 602 10.1080/00167223.2004.10649499
- 32  
33 603 Aagaard T, Sørensen P. 2013. Sea level rise and the sediment budget of an eroding barrier on the Danish North Sea  
34 604 coast. *Journal of Coastal Research* **65** : 434–439. DOI: 10.2112/SI65-074.1
- 35  
36 605 Allard J, Bertin X, Chaumillon E, Pouget F. 2008. Sand spit rhythmic development: A potential record of wave  
37 606 climate variations? Arçay Spit, western coast of France. *Marine Geology* **253** : 107–131. DOI:  
38 607 10.1016/j.margeo.2008.05.009
- 39  
40 608 Allard J, Chaumillon E, Féliès H. 2009. A synthesis of morphological evolutions and Holocene stratigraphy of a  
41 609 wave-dominated estuary: The Arcachon lagoon, SW France. *Continental Shelf Research* **29** : 957–969. DOI:  
42 610 10.1016/j.csr.2008.11.017
- 43  
44 611 Ashton AD, Nienhuis J, Ells K. 2016. On a neck, on a spit: controls on the shape of free spits. *Earth Surface*  
45 612 *Dynamics* **4** : 193–210. DOI: 10.5194/esurf-4-193-2016
- 46  
47 613 Aubié S, Tastet J-P. 2000. Coastal Erosion, Processes and Rates: An Historical Study of the Gironde Coastline,  
48 614 Southwestern France. *Journal of Coastal Research* **16** : 756–767.
- 49  
50 615 Barnard PL, Allan J, Hansen JE, Kaminsky GM, Ruggiero P, Doria A. 2011. The impact of the 2009-10 El Niño  
51 616 Modoki on U.S. West Coast beaches. *Geophysical Research Letters* **38** : n/a-n/a. DOI: 10.1029/2011GL047707
- 52  
53 617 Barnston AG, Livezey RE. 1987. Classification, Seasonality and Persistence of Low-Frequency Atmospheric  
54 618 Circulation Patterns. *Monthly Weather Review* **115** : 1083–1126. DOI: 10.1175/1520-  
55 619 0493(1987)115<1083:CSAPOL>2.0.CO;2

- 1  
2  
3 620 Bertin X, Fortunato AB, Oliveira A. 2009. A modeling-based analysis of processes driving wave-dominated inlets.  
4 621 Continental Shelf Research **29** : 819–834. DOI: 10.1016/j.csr.2008.12.019  
5  
6 622 Bertin X, Li K, Roland A, Bidlot J-R. 2015. The contribution of short-waves in storm surges: Two case studies in  
7 623 the Bay of Biscay. Continental Shelf Research **96** : 1–15. DOI: 10.1016/j.csr.2015.01.005  
8  
9 624 Bruneau N, Fortunato AB, Dodet G, Freire P, Oliveira A, Bertin X. 2011. Future evolution of a tidal inlet due to  
10 625 changes in wave climate, Sea level and lagoon morphology (Óbidos lagoon, Portugal). Continental Shelf Research  
11 626 **31** : 1915–1930. DOI: 10.1016/j.csr.2011.09.001  
12  
13 627 Bruun P. 1978. Stability of Coastal Inlets: Theory and Engineering . Elsevier. [online] Available from:  
14 628 <http://www.sciencedirect.com/science/bookseries/01651250/23>  
15  
16 629 Calafat FM, Chambers DP, Tsimplis MN. 2012. Mechanisms of decadal sea level variability in the eastern North  
17 630 Atlantic and the Mediterranean Sea. Journal of Geophysical Research: Oceans **117** : n/a-n/a. DOI:  
18 631 10.1029/2012JC008285  
19  
20 632 Caspari E. 1872. Rapport sur l'exploration des Passes du Bassin d'Arcachon  
21  
22 633 Castelle B, Guillot B, Marieu V, Chaumillon E, Hanquiez V, Bujan S, Poppeschi C. 2018. Spatial and temporal  
23 634 patterns of shoreline change of a 280-km high-energy disrupted sandy coast from 1950 to 2014: SW France.  
24 635 Estuarine, Coastal and Shelf Science **200** : 212–223. DOI: 10.1016/j.ecss.2017.11.005  
25  
26 636 Cayocca F. 2001. Long-term morphological modeling of a tidal inlet: the Arcachon Basin, France. Coastal  
27 637 Engineering **42** : 115–142. DOI: 10.1016/S0378-3839(00)00053-3  
28  
29 638 Cazenave A, Le Cozannet G. 2014. Sea level rise and its coastal impacts. Earth's Future **2** : 15–34. DOI:  
30 639 10.1002/2013EF000188  
31  
32 640 CETMEF. n.d. CANDHIS (03302) - Bouée Cap Ferret [online] Available from:  
33 641 [http://candhis.cetmef.developpement-](http://candhis.cetmef.developpement-durable.gouv.fr/campagne/?idcampagne=b6d767d2f8ed5d21a44b0e5886680cb9)  
34 642 [durable.gouv.fr/campagne/?idcampagne=b6d767d2f8ed5d21a44b0e5886680cb9](http://candhis.cetmef.developpement-durable.gouv.fr/campagne/?idcampagne=b6d767d2f8ed5d21a44b0e5886680cb9) (Accessed 6 March 2017)  
35  
36 643 Charles E, Idier D, Thiébot J, Le Cozannet G, Pedreros R, Arduin F, Planton S. 2012. Present Wave Climate in the  
37 644 Bay of Biscay: Spatiotemporal Variability and Trends from 1958 to 2001. Journal of Climate **25** : 2020–2039. DOI:  
38 645 10.1175/JCLI-D-11-00086.1  
39  
40 646 Chowdhury P, Behera MR. 2017. Effect of long-term wave climate variability on longshore sediment transport  
41 647 along regional coastlines. Progress in Oceanography **156** : 145–153. DOI: 10.1016/j.pocean.2017.06.001  
42  
43 648 Clarke M, Rendell H, Tastet J-P, Clavé B, Massé L. 2002. Late-Holocene sand invasion and North Atlantic  
44 649 storminess along the Aquitaine Coast, southwest France. The Holocene **12** : 231–238. DOI:  
45 650 10.1191/0959683602hl539rr  
46  
47 651 Clemmensen LB, Richardt N, Anderson C. 2001. Holocene sea-level variation and spit development: data from  
48 652 Skagen Odde, Denmark. The Holocene **11** : 323–331. DOI: 10.1191/095968301667877044  
49  
50 653 Cook ER, D'Arrigo RD, Mann ME. 2002. A Well-Verified, Multiproxy Reconstruction of the Winter North Atlantic  
51 654 Oscillation Index since a.d. 1400. Journal of Climate **15** : 1754–1764. DOI: 10.1175/1520-  
52 655 0442(2002)015<1754:AWVMRO>2.0.CO;2  
53  
54 656 Davis RA. 2013. Evolution of Coastal Landforms. In Treatise on Geomorphology , . Elsevier; 417–448. [online]  
55 657 Available from: <http://linkinghub.elsevier.com/retrieve/pii/B9780123747396002931> (Accessed 20 April 2016)  
56  
57  
58  
59  
60



- 1  
2  
3 658 Dissanayake DMPK, Ranasinghe R, Roelvink JA. 2012. The morphological response of large tidal inlet/basin  
4 659 systems to relative sea level rise. *Climatic Change* **113** : 253–276. DOI: 10.1007/s10584-012-0402-z  
5
- 6 660 Dodet G, Bertin X, Taborda R. 2010. Wave climate variability in the North-East Atlantic Ocean over the last six  
7 661 decades. *Ocean Modelling* **31** : 120–131. DOI: 10.1016/j.ocemod.2009.10.010  
8
- 9 662 Féliens H, Faugères J-C. 1998. Facies and geometry of tidal channel-fill deposits (Arcachon Lagoon, SW France).  
10 663 *Marine Geology* **150** : 131–148. DOI: 10.1016/S0025-3227(98)00049-8  
11
- 12 664 Féliens H, Lericolais G. 2005. Internal architecture of an incised valley-fill on a wave- and tide-dominated coast (the  
13 665 Leyre incised valley, Bay of Biscaye, France). *Comptes Rendus Geoscience* **337** : 1257–1266. DOI:  
14 666 10.1016/j.crte.2005.06.005  
15
- 16 667 Féliens H, Lericolais G, Posamentier HW. 2010. Comparison of wave- and tide-dominated incised valleys: specific  
17 668 processes controlling systems tract architecture and reservoir geometry. *Bulletin de la Societe Geologique de France*  
18 669 **181** : 171–181. DOI: 10.2113/gssgfbull.181.2.171  
19
- 20 670 FitzGerald D, Georgiou I, Miner M. 2015. Estuaries and Tidal Inlets. In *Coastal Environments and Global Change* ,  
21 671 Masselink G and Gehrels R (eds). John Wiley & Sons, Ltd: Chichester, UK; 268–298. [online] Available from:  
22 672 <http://doi.wiley.com/10.1002/9781119117261.ch12> (Accessed 19 May 2016)  
23
- 24 673 FitzGerald DM, Fenster MS, Argow BA, Buynevich IV. 2008. Coastal Impacts Due to Sea-Level Rise. *Annual*  
25 674 *Review of Earth and Planetary Sciences* **36** : 601–647. DOI: 10.1146/annurev.earth.35.031306.140139  
26
- 27 675 FitzGerald DM, Kraus NC, Hands EB. 2001. Natural mechanisms of sediment bypassing at tidal inlets . U.S. Army  
28 676 Engineer Research and Development Center: Vicksburg, MS  
29
- 30 677 Fortunato AB et al. 2014. Morphological evolution of an ephemeral tidal inlet from opening to closure: The  
31 678 Albufeira inlet, Portugal. *Continental Shelf Research* **73** : 49–63. DOI: 10.1016/j.csr.2013.11.005  
32
- 33 679 Fortunato AB, Oliveira A. 2005. Influence of Intertidal Flats on Tidal Asymmetry. *Journal of Coastal Research* **215** :  
34 680 1062–1067. DOI: 10.2112/03-0089.1  
35
- 36 681 Friedrichs CT, Aubrey DG. 1988. Non-linear tidal distortion in shallow well-mixed estuaries: a synthesis. *Estuarine,*  
37 682 *Coastal and Shelf Science* **27** : 521–545. DOI: 10.1016/0272-7714(88)90082-0  
38
- 39 683 Fruergaard M, Andersen TJ, Johannessen PN, Nielsen LH, Pejrup M. 2013. Major coastal impact induced by a  
40 684 1000-year storm event. *Scientific Reports* **3** DOI: 10.1038/srep01051 [online] Available from:  
41 685 <http://www.nature.com/articles/srep01051> (Accessed 11 April 2016)  
42
- 43 686 Fruergaard M, Andersen TJ, Nielsen LH, Johannessen PN, Aagaard T, Pejrup M. 2015. High-resolution  
44 687 reconstruction of a coastal barrier system: impact of Holocene sea-level change. *Sedimentology* **62** : 928–969. DOI:  
45 688 10.1111/sed.12167  
46
- 47 689 Glueck MF, Stockton CW. 2001. Reconstruction of the North Atlantic Oscillation, 1429-1983. *International Journal*  
48 690 *of Climatology* **21** : 1453–1465. DOI: 10.1002/joc.684  
49
- 50 691 Hayes MO. 1979. Barrier island morphology as a function of tidal and wave regime. In *Barrier Islands, from the*  
51 692 *Gulf of St. Lawrence to the Gulf of Mexico* , . Leatherman, S.; 1–27.  
52
- 53 693 Hoan LX, Hanson H, Larson M, Kato S. 2011. A mathematical model of spit growth and barrier elongation:  
54 694 Application to Fire Island Inlet (USA) and Badreveln Spit (Sweden). *Estuarine, Coastal and Shelf Science* **93** : 468–  
55 695 477. DOI: 10.1016/j.ecss.2011.05.033  
56  
57  
58  
59  
60



- 1  
2  
3 696 Holgate SJ, Matthews A, Woodworth PL, Rickards LJ, Tamisiea ME, Bradshaw E, Foden PR, Gordon KM,  
4 697 Jevrejeva S, Pugh J. 2013. New Data Systems and Products at the Permanent Service for Mean Sea Level. *Journal of*  
5 698 *Coastal Research* **288** : 493–504. DOI: 10.2112/JCOASTRES-D-12-00175.1  
6  
7 699 Hubbard DK, Oertel GF, Nummedal D. 1979. The Role of Waves and Tidal Currents in the Development of Tidal-  
8 700 inlet Sedimentary Structures and Sand Body Geometry: Examples from North Carolina, South Carolina, and  
9 701 Georgia. *SEPM Journal of Sedimentary Research* **Vol. 49** DOI: 10.1306/212F78B5-2B24-11D7-  
10 702 8648000102C1865D [online] Available from: [http://jsedres.sepmonline.org/cgi/doi/10.1306/212F78B5-2B24-11D7-](http://jsedres.sepmonline.org/cgi/doi/10.1306/212F78B5-2B24-11D7-8648000102C1865D)  
11 703 [8648000102C1865D](http://jsedres.sepmonline.org/cgi/doi/10.1306/212F78B5-2B24-11D7-8648000102C1865D) (Accessed 26 September 2016)  
12  
13 704 Hurrell JW. 1995. Decadal Trends in the North Atlantic Oscillation: Regional Temperatures and Precipitation.  
14 705 *Science* **269** : 676–679. DOI: 10.1126/science.269.5224.676  
15  
16 706 Hurrell JW. 2015. The Climate Data Guide: Hurrell North Atlantic Oscillation (NAO) Index (Station-based) [online]  
17 707 Available from: <https://climatedataguide.ucar.edu/climate-data/hurrell-north-atlantic-oscillation-nao-index-pc-based>  
18 708 (Accessed 14 August 2015)  
19  
20 709 Hurrell JW. 2019. The Climate Data Guide: Hurrell North Atlantic Oscillation (NAO) Index (PC-based) [online]  
21 710 Available from: <https://climatedataguide.ucar.edu/climate-data/hurrell-north-atlantic-oscillation-nao-index-pc-based>  
22 711 (Accessed 26 January 2019)  
23  
24 712 Hurrell JW, Deser C. 2009. North Atlantic climate variability: The role of the North Atlantic Oscillation. *Journal of*  
25 713 *Marine Systems* **78** : 28–41. DOI: 10.1016/j.jmarsys.2008.11.026  
26  
27 714 Idier D, Castelle B, Charles E, Mallet C. 2013. Longshore sediment flux hindcast: spatio-temporal variability along  
28 715 the SW Atlantic coast of France. *Journal of Coastal Research* **165** : 1785–1790. DOI: 10.2112/SI65-302.1  
29  
30 716 Ineson S, Scaife AA, Knight JR, Manners JC, Dunstone NJ, Gray LJ, Haigh JD. 2011. Solar forcing of winter  
31 717 climate variability in the Northern Hemisphere. *Nature Geoscience* **4** : 753–757. DOI: 10.1038/ngeo1282  
32  
33 718 Jarrett JT. 1976. Tidal Prism - Inlet Area Relationships . Army Engineer Waterways Experiment Station: Vicksburg  
34 719 MS [online] Available from: [http://cirp.usace.army.mil/pubs/archive/GITI-Report\\_Number\\_3.pdf](http://cirp.usace.army.mil/pubs/archive/GITI-Report_Number_3.pdf)  
35  
36 720 Jevrejeva S, Moore JC, Grinsted A, Matthews AP, Spada G. 2014. Trends and acceleration in global and regional  
37 721 sea levels since 1807. *Global and Planetary Change* **113** : 11–22. DOI: 10.1016/j.gloplacha.2013.12.004  
38  
39 722 Jones PD, Jonsson T, Wheeler D. 1997. Extension to the North Atlantic oscillation using early instrumental pressure  
40 723 observations from Gibraltar and south-west Iceland. *International Journal of Climatology* **17** : 1433–1450. DOI:  
41 724 10.1002/(SICI)1097-0088(19971115)17:13<1433::AID-JOC203>3.0.CO;2-P  
42  
43 725 Kalnay E et al. 1996. The NCEP/NCAR 40-Year Reanalysis Project. *Bulletin of the American Meteorological*  
44 726 *Society* **77** : 437–471. DOI: 10.1175/1520-0477(1996)077<0437:TNYRP>2.0.CO;2  
45  
46 727 Kirwan ML, Murray AB. 2008. Tidal marshes as disequilibrium landscapes? Lags between morphology and  
47 728 Holocene sea level change. *Geophysical Research Letters* **35** DOI: 10.1029/2008GL036050 [online] Available from:  
48 729 <http://doi.wiley.com/10.1029/2008GL036050> (Accessed 19 April 2016)  
49  
50 730 Kraus NC. 1998. Inlet Cross-Sectional Area Calculated by Process-Based Model. presented at the Conference on  
51 731 Coastal Engineering. Copenhagen, Denmark. 3265–3278 pp. [online] Available from: [https://icce-ojs-](https://icce-ojs-tamu.tdl.org/icce/index.php/icce/article/view/5839)  
52 732 [tamu.tdl.org/icce/index.php/icce/article/view/5839](https://icce-ojs-tamu.tdl.org/icce/index.php/icce/article/view/5839) (Accessed 28 September 2016)  
53  
54 733 Kraus NC. 2000. Reservoir Model of Ebb-Tidal Shoal Evolution and Sand Bypassing. *Journal of Waterway, Port,*  
55 734 *Coastal, and Ocean Engineering* **126** : 305–313. DOI: 10.1061/(ASCE)0733-950X(2000)126:6(305)  
56  
57  
58  
59  
60

- 1  
2  
3 735 Kraus NC, Seabergh WC. 2002. Inlet spits and maintenance of navigation channels . U.S. Army Engineer Research  
4 736 and Development Center: Vicksburg, MS [online] Available from:  
5 737 <http://acwc.sdp.sirsi.net/client/search/asset/1000403> (Accessed 13 June 2016)  
6  
7 738 Lapeyre. 1925. Cartes du Pays de Buch, 1318 à 1913  
8  
9 739 Larson M, Kraus NC, Connell KJ. 2007. MODELING SEDIMENT STORAGE AND TRANSFER FOR  
10 740 SIMULATING REGIONAL COASTAL EVOLUTION. presented at the Proceedings of the 30th International  
11 741 Conference. San Diego, California, USA. 3924–3936 pp. April [online] Available from:  
12 742 [http://www.worldscientific.com/doi/abs/10.1142/9789812709554\\_0330](http://www.worldscientific.com/doi/abs/10.1142/9789812709554_0330) (Accessed 22 March 2019)  
13  
14 743 Luterbacher J et al. 2001. Extending North Atlantic oscillation reconstructions back to 1500. Atmospheric Science  
15 744 Letters **2** : 114–124. DOI: 10.1006/asle.2002.0047  
16  
17 745 Luterbacher J, Schmutz C, Gyalistras D, Xoplaki E, Wanner H. 1999. Reconstruction of monthly NAO and EU  
18 746 indices back to AD 1675. Geophysical Research Letters **26** : 2745–2748. DOI: 10.1029/1999GL900576  
19  
20 747 van Maanen B, Coco G, Bryan KR, Friedrichs CT. 2013. Modeling the morphodynamic response of tidal  
21 748 embayments to sea-level rise. Ocean Dynamics **63** : 1249–1262. DOI: 10.1007/s10236-013-0649-6  
22  
23 749 Masselink G, Austin M, Scott T, Poate T, Russell P. 2014. Role of wave forcing, storms and NAO in outer bar  
24 750 dynamics on a high-energy, macro-tidal beach. Geomorphology **226** : 76–93. DOI: 10.1016/j.geomorph.2014.07.025  
25  
26 751 Masselink G, Castelle B, Scott T, Dodet G, Suarez S, Jackson D, Floc'h F. 2016. Extreme wave activity during  
27 752 2013/2014 winter and morphological impacts along the Atlantic coast of Europe. Geophysical Research Letters **43** :  
28 753 2135–2143. DOI: 10.1002/2015GL067492  
29  
30 754 Mazzarella A, Scafetta N. 2012. Evidences for a quasi 60-year North Atlantic Oscillation since 1700 and its  
31 755 meaning for global climate change. Theoretical and Applied Climatology **107** : 599–609. DOI: 10.1007/s00704-011-  
32 756 0499-4  
33  
34 757 Michel D, Howa HL. 1997. Morphodynamic behaviour of a tidal inlet system in a mixed-energy environment.  
35 758 Physics and Chemistry of the Earth **22** : 339–343. DOI: 10.1016/S0079-1946(97)00155-9  
36  
37 759 Nahon A. 2018. Ongoing morphological evolution of a Holocene coastal barrier-spit - The Cap Ferret, bounded by  
38 760 the Bay of Arcachon tidal inlet, PhD Thesis, University of Bordeaux: Bordeaux, France, October  
39  
40 761 Nahon A, Bertin X, Fortunato AB, Oliveira A. 2012. Process-based 2DH morphodynamic modeling of tidal inlets:  
41 762 A comparison with empirical classifications and theories. Marine Geology **291–294** : 1–11. DOI:  
42 763 10.1016/j.margeo.2011.10.001  
43  
44 764 Nahon A, Idier D, Fenies H, Mugica J, Senechal N, Mallet C. 2015. Role of north atlantic climate variability on  
45 765 barrier-spit oscillations: the Cap Ferret. presented at the Coastal Sediments 2015. San Diego. July [online] Available  
46 766 from: [http://www.worldscientific.com/doi/abs/10.1142/9789814689977\\_0240](http://www.worldscientific.com/doi/abs/10.1142/9789814689977_0240) (Accessed 21 April 2016)  
47  
48 767 Newton A et al. 2014. An overview of ecological status, vulnerability and future perspectives of European large  
49 768 shallow, semi-enclosed coastal systems, lagoons and transitional waters. Estuarine, Coastal and Shelf Science **140** :  
50 769 95–122. DOI: 10.1016/j.ecss.2013.05.023  
51  
52 770 Nichols M, Allen G. 1981. Sedimentary processes in coastal lagoons. Unesco Technical Papers In Marine Science :  
53 771 27–80.  
54  
55 772 O'Brien MP. 1931. Estuary tidal prisms related to entrance areas. Civil Engineering **1** : 738–739.  
56  
57  
58  
59  
60

- 1  
2  
3 773 Ortega P, Lehner F, Swingedouw D, Masson-Delmotte V, Raible CC, Casado M, Yiou P. 2015. A model-tested  
4 774 North Atlantic Oscillation reconstruction for the past millennium. *Nature* **523** : 71–74.  
5  
6 775 PAB. 1985. Etude de l'évolution du littoral de la presqu'île du Cap Ferret . Port Autonome de Bordeaux, Direction  
7 776 de l'aménagement et de l'environnement maritimes: Bordeaux, France  
8  
9 777 Plus M, Dumas F, Stanisière J-Y, Maurer D. 2009. Hydrodynamic characterization of the Arcachon Bay, using  
10 778 model-derived descriptors. *Continental Shelf Research* **29** : 1008–1013. DOI: 10.1016/j.csr.2008.12.016  
11  
12 779 Poirier C, Tessier B, Chaumillon É, Bertin X, Fruergaard M, Mouazé D, Noël S, Weill P, Wöppelmann G. 2017.  
13 780 Decadal changes in North Atlantic atmospheric circulation patterns recorded by sand spits since 1800CE.  
14 781 *Geomorphology* **281** : 1–12. DOI: 10.1016/j.geomorph.2016.12.028  
15  
16 782 Ranasinghe R, Duong TM, Uhlenbrook S, Roelvink D, Stive M. 2012. Climate-change impact assessment for inlet-  
17 783 interrupted coastlines. *Nature Climate Change* **3** : 83–87. DOI: 10.1038/nclimate1664  
18  
19 784 Ridderinkhof W, Hoekstra P, van der Vegt M, de Swart HE. 2016. Cyclic behavior of sandy shoals on the ebb-tidal  
20 785 deltas of the Wadden Sea. *Continental Shelf Research* **115** : 14–26. DOI: 10.1016/j.csr.2015.12.014  
21  
22 786 Roland A, Zhang YJ, Wang HV, Meng Y, Teng Y-C, Maderich V, Brovchenko I, Dutour-Sikiric M, Zanke U. 2012.  
23 787 A fully coupled 3D wave-current interaction model on unstructured grids. *Journal of Geophysical Research: Oceans*  
24 788 **117** : n/a-n/a. DOI: 10.1029/2012JC007952  
25  
26 789 Salles P, Valle-Levinson A, Sottolichio A, Senechal N. 2015. Wind-driven modifications to the residual circulation  
27 790 in an ebb-tidal delta: Arcachon Lagoon, Southwestern France. *Journal of Geophysical Research: Oceans* **120** : 728–  
28 791 740. DOI: 10.1002/2014JC010311  
29  
30 792 van der Schrier G, Barkmeijer J. 2005. Bjerknes' hypothesis on the coldness during AD 1790–1820 revisited.  
31 793 *Climate Dynamics* **25** : 537–553. DOI: 10.1007/s00382-005-0053-0  
32  
33 794 Semedo A, Weisse R, Behrens A, Sterl A, Bengtsson L, Günther H. 2013. Projection of Global Wave Climate  
34 795 Change toward the End of the Twenty-First Century. *Journal of Climate* **26** : 8269–8288. DOI: 10.1175/JCLI-D-12-  
35 796 00658.1  
36  
37 797 Senechal N, Sottolichio A, Bertrand F, Goeldner-Gianella L, Garlan T. 2013. Observations of waves' impact on  
38 798 currents in a mixed-energy tidal inlet: Arcachon on the southern French Atlantic coast. *Journal of Coastal Research*  
39 799 **165** : 2053–2058. DOI: 10.2112/S165-347.1  
40  
41 800 Shindell DT. 2001. Solar Forcing of Regional Climate Change During the Maunder Minimum. *Science* **294** : 2149–  
42 801 2152. DOI: 10.1126/science.1064363  
43  
44 802 Splinter KD, Davidson MA, Golshani A, Tomlinson R. 2012. Climate controls on longshore sediment transport.  
45 803 *Continental Shelf Research* **48** : 146–156. DOI: 10.1016/j.csr.2012.07.018  
46  
47 804 Svendsen IA. 2006. Linear Waves. In *Introduction to Nearshore Hydrodynamics* , . World Scientific; 49–205.  
48 805 [online] Available from: <https://books.google.fr/books?id=g7-PHmrvQcC>  
49  
50 806 Walton, Jr. TL, Adams WD. 1976. Capacity of inlet outer bars to store sand. In *Proceedings of the 15th Coastal*  
51 807 *Engineering Conference, Honolulu, HI* , . ASCE; 1919–1937. [online] Available from: [https://icce-ojs-](https://icce-ojs-tamu.tdl.org/icce/index.php/icce/article/viewFile/3161/2825)  
52 808 [tamu.tdl.org/icce/index.php/icce/article/viewFile/3161/2825](https://icce-ojs-tamu.tdl.org/icce/index.php/icce/article/viewFile/3161/2825)  
53  
54 809 van der Wegen M. 2013. Numerical modeling of the impact of sea level rise on tidal basin morphodynamics. *Journal*  
55 810 *of Geophysical Research: Earth Surface* **118** : 447–460. DOI: 10.1002/jgrf.20034  
56  
57  
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59  
60

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2  
3 811 Wiggins M, Scott T, Masselink G, Russell P, Castelle B, Dodet G. 2017. THE ROLE OF MULTI-DECADAL  
4 812 CLIMATE VARIABILITY IN CONTROLLING COASTAL DYNAMICS: RE-INTERPRETATION OF THE  
5 813 ‘LOST VILLAGE OF HALLSANDS’’.’ Helsingor, Denmark. June [online] Available from: [https://hal.archives-](https://hal.archives-ouvertes.fr/hal-01578356)  
6 814 [ouvertes.fr/hal-01578356](https://hal.archives-ouvertes.fr/hal-01578356)  
7  
8 815 Wong PP, Losada IJ, Gattuso J-P, Hinkel J, Khattabi A, McInnes KL, Saito Y, Sallenger A. 2014. Coastal systems  
9 816 and low-lying areas. In *Climate Change 2014: Impacts, Adaptation, and Vulnerability. Part A: Global and Sectoral*  
10 817 *Aspects. Contribution of Working Group II to the Fifth Assessment Report of the Intergovernmental Panel of*  
11 818 *Climate Change*, Field, C.B. et al. (eds). Cambridge, United Kingdom and New York, NY, USA; 361–409.  
12  
13 819 Wöppelmann G, Pouvreau N, Coulomb A, Simon B, Woodworth PL. 2008. Tide gauge datum continuity at Brest  
14 820 since 1711: France’s longest sea-level record. *Geophysical Research Letters* **35** DOI: 10.1029/2008GL035783  
15 821 [online] Available from: <http://doi.wiley.com/10.1029/2008GL035783> (Accessed 2 June 2016)  
16  
17 822 Zhou Z, Coco G, Jiménez M, Olabarrieta M, van der Wegen M, Townend I. 2014. Morphodynamics of river-  
18 823 influenced back-barrier tidal basins: The role of landscape and hydrodynamic settings. *Water Resources Research* **50**  
19 824 : 9514–9535. DOI: 10.1002/2014WR015891  
20  
21 825  
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827 **Tables**828 **Table 1.** Morphological parameters of the inlet throat between 1949 and 2014.

year	Channel volume (Mm <sup>3</sup> )	Shoal volume (Mm <sup>3</sup> )	Mean elevation (m NGF)	Morphological amplitude (Mm <sup>3</sup> )
1949	32.14	59.48	-5.50	42.69
1969	30.02	54.20	-5.61	38.80
1979	30.02	43.21	-6.17	34.74
2001	33.93	61.41	-5.53	44.02
2014	38.62	60.58	-5.82	46.65
mean	32.53	55.78	-5.77	41.38

830 **Table 2.** Trends in Brest tidal gauge annual sea level data.

Periods	Linear trend, mm.y <sup>-1</sup>	Spit behaviour
1807-1835	-1.08	extending
1826-1872	0.14	retreating
1865-1909	-0.33	extending
1900-1934	2.5	retreating
1932-1972	1.26	extending
1968-2015	2.42	retreating

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832 **Table 3.** Comparison of hindcasted wave parameters and wave buoy observations.

Wave Field	BIAS (% of mean)	ERMS (% of mean)	Pearson Corr. coeff.
<i>Hs</i>	-1.16	43 cm	0.93
<i>WP</i>	-2.4	-	0.93
<i>wDir</i>	3.75° (0°)	-	-
<i>WPx</i>	-3.4 (1.7)	-	0.93 (0.93)
<i>WPy</i>	27.1 (1.7)	-	0.80 (0.81)

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3 834 **Figures**  
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8 836 **Figure 1** – Location maps. **a**) Bathymetry of the Bay of Biscay, bordering the  
9  
10 837 SW coast of France; **b**) Gironde sandy coast, between the Gironde Estuary and  
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12 838 the Bay of Arcachon, former estuary of the Leyre river and now a lagoon, semi-enclosed  
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14 839 by the Cap Ferret sand spit (coordinates given in Lambert 93, in meters); **c**)  
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16 840 Satellite Landsat 8 image of the Bay's tidal inlet in October 2014, showing the main  
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18 841 geomorphological units and the red dotted line indicating the 2014 coastline of the Cap  
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20 842 Ferret spit-end, on the inlet's updrift margin.  
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26 844 **Figure 2** – Nautical charts. **a**) Ordnance survey map from 1888, black circled  
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28 845 dots indicate control points used for georeferencing the map, red dashed line is the  
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30 846 coastline measured by GPS on 2014 spring, red stars indicate the 3 reference locations  
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32 847 used for georeferencing nautical charts; **b-e**) examples of nautical charts used  
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34 848 to measure the spit-end positions, indicated by asterisk symbols for every given date;  
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36 849 **f-g**) Satellite Landsat image of the Bay of Arcachon and its tidal inlet in October  
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38 850 2014.  
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45 852 **Figure 3** – Measuring the spit's extension relative to the position of Cap Ferret's  
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47 853 lighthouse. **Upper panels**: charts and coastlines from 1768 and 1826 and  
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49 854 Satellite Landsat 8 image from October 2014, red dotted lines represent the 2014  
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51 855 coastline and the red stars indicate the lighthouse's position, white dashed lines  
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53 856 materialize the axe onto which spit-end positions were orthogonally projected and  
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3 857 measured, the black sector in the right panel represents the spit-end migration range;  
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5 858 **lower panel**: the grey curve shows the spit-end path between 1768 and 1936,  
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7 859 grey error bars stand for historical charts and the black ones for modern charts, asterisk  
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10 860 symbol stand for spit-end positions in Figure 2**b-e**.

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12 861  
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14 862 **Figure 4** – Positions of the Cap Ferret spit-end measured on aerial photographs.

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17 863 **Upper panels**: series of aerial photographs of the spit-end, red dotted lines  
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19 864 represent the 2014 coastline, black asterisk symbols materialize the spit-end positions  
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21 865 for the given years, superimposed on photographs from 1934 and 1946 are the  
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23 866 shorelines reproduced from charts of 1932 and 1936 respectively; **lower panel**:  
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25 867 distance of the spit-end to lighthouse's position, black asterisk symbols correspond to  
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27 868 those on the upper panel.

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33 870 **Figure 5** – Inlet throat bathymetric data. **a**) summer 2014 along transect  
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35 871 soundings; **b**) interpolated bathymetry with the -7 m NGF contours (black lines);  
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37 872 **c**) -7 m NGF contours superimposed onto October 2014 Landsat image;  
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39 873 **d**) erosion (blue) – deposition (red) patterns over the 12.74 km<sup>2</sup> area covered  
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41 874 by all 5 surveys between 1949 and 2014; **e-h**) channel volume (below -7 m  
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43 875 NGF) , shoal volume (above -7 m NGF), overlapping area mean depth (m NGF) and  
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45 876 inlet morphological amplitude over panel **d**'s area (**e**, **f** and  
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47 877 **h** are in percent of averaged values for all 5 surveys).

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3 879 **Figure 6** – Brest tidal gauge Mean Sea Level (MSL) variation. **Upper panel**:  
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5 880 annual MSL data from 1807 to 2015 recovered from the PMSL observatory (Holgate et  
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7 881 al., 2013), black lines are the 11-year centred running mean of annual measurements  
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9 882 and the thick dotted red line is the quadratic fit of all the data; **lower panel**:  
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11 883 same as above with data from 20<sup>th</sup> century onward fitted with a 5<sup>th</sup> order polynomial  
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13 884 function.  
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19 886 **Figure 7** – Local, 54 m water depth wave hindcast. **Upper panel**: 90-day  
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21 887 running mean of the wave power (WP; linear wave theory approximation) 15 km  
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23 888 offshore Cap Ferret, solid blue line are wind wave model (WWM-II, Roland et al., 2012)  
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25 889 results with Bertin et al. (2015) setup, dotted black line are the wave buoy observations  
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27 890 (CETMEF, n.d.) between April 2011 and May 2014, data is normalized using mean WP  
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29 891 value over the hindcast period (1948-2014); **lower panel**: same as above with  
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31 892 the alongshore component of the wave power (WP<sub>y</sub>); *r* values are  
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33 893 Pearson's linear correlation coefficients between plotted curves.  
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40 895 **Figure 8** – Spit-end path and cumulated winter NAO indices between 1768 and 2014.  
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42 896 **Upper panel**: the grey curve shows the distance from the spit-end to the Cap  
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44 897 Ferret lighthouse as on Figures 3&4 (red dotted frame is for next Figure 8's zoom);  
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46 898 **mid panel**: darker blue curve is the cumulated winter (DJFM) index of the NAO  
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48 899 reconstruction of Luterbacher et al. (1999), lighter blue curve is the same but for  
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50 900 (Hurrell's (2015) station-based (DJFM) index; **lower panel**: darker blue curve is  
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52 901 the cumulated winter (DJF) index of the NAO reconstruction of Ortega et al. (2015),  
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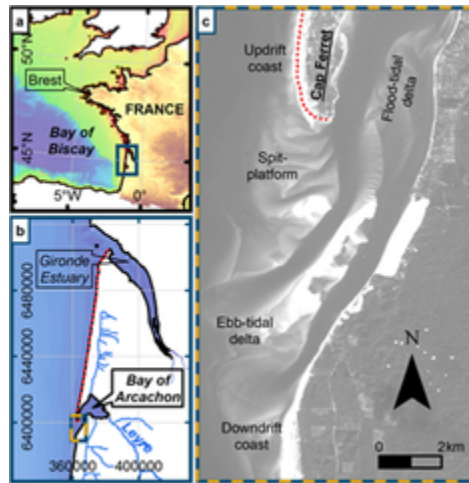
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7 904 period in darker orange,  $r$  values are Pearson's linear correlation  
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10 905 coefficients between plotted curves.

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13 907 **Figure 9** – Evolution of the spit-inlet system across the Dalton Minimum (1780 – 1830).  
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16 908 Navigation charts from years 1768, 1813, 1826 and 1865; blue asterisks indicate the  
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18 909 positions of the spit terminus at the respective dates; black arrows materialize the inlet  
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20 910 minimal width; red and blue arrows materialize the position of respectively the northern  
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22 911 and southern channels; the dotted black line depicts the migration of the main inlet  
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24 912 shoal barycentre; common red stars, dotted white axis and dotted red line are the same  
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26 913 as in Figures 2 and 3.

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32 915 **Figure 10** – NAO relationships with Cap Ferret's spit-end and local wave climate.  
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34 916 **Upper panel**: in grey is the path of the spit-end, as the distance from the spit-  
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36 917 end to the Cap Ferret lighthouse, as on Figures 3&4 and over since 1909,  
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38 918 superimposed blue curve is Hurrell's (2015) station-based NAO winter index, averaged  
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40 919 of 10 years preceding each observation; **lower panel**: same winter NAO index  
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42 920 curve as above, along with hindcasted alongshore winter wave power (WPy), averaged  
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44 921 in same fashion as the NAO index and normalized using mean WPy value over the  
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46 922 hindcast period (1948-2014); on both panels, orange shades indicate periods of spit  
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48 923 elongation,  $r$  values are Pearson's linear correlation coefficients between  
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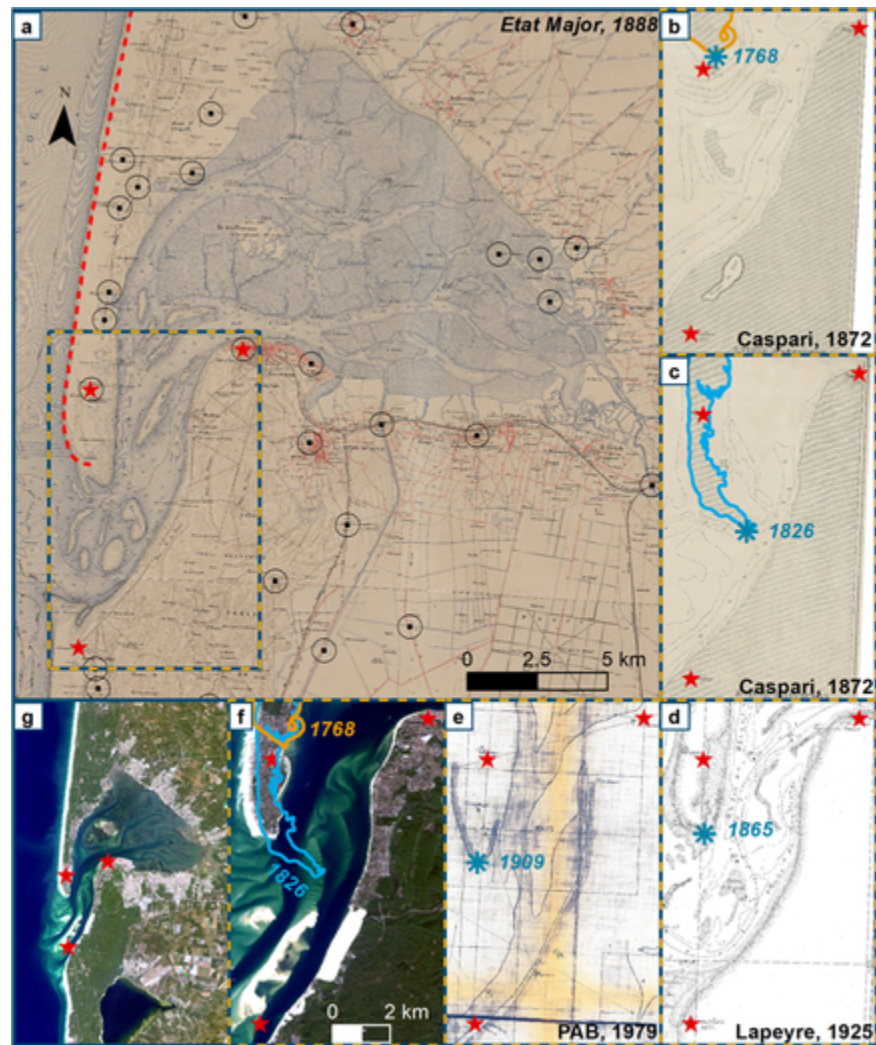
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3 926 **Figure 11** – Sandspit’s morphological response to changes in mean wave direction.  
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6 927 **a**) spit-end response to a decrease of the averaged wave angle of incidence,  
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8 928 modelled by (and reproduced from) Ashton et al. (2016), dashed line represents initial  
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10 929 shoreline and shaded area represents the spit-end contours after waves became more  
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12 930 shore normal; **b**) same as panel **a**), but for an increase of the averaged  
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14 931 wave angle of incidence, or after waves became more oblique; **c**) dashed line  
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16 932 represents Cap Ferret spit-end in 1826, at the end of the Dalton Minimum, and shaded  
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19 933 area represents the spit-end in 1865.  
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Location maps. **a)** Bathymetry of the Bay of Biscay, bordering the SW coast of France; **b)** Gironde sandy coast, between the Gironde Estuary and the Bay of Arcachon, former estuary of the Leyre river and now a lagoon, semi-enclosed by the Cap Ferret sand spit (coordinates given in Lambert 93, in meters); **c)** Satellite Landsat 8 image of the Bay's tidal inlet in October 2014, showing the main geomorphological units and the red dotted line indicating the 2014 coastline of the Cap Ferret spit-end, on the inlet's updrift margin.

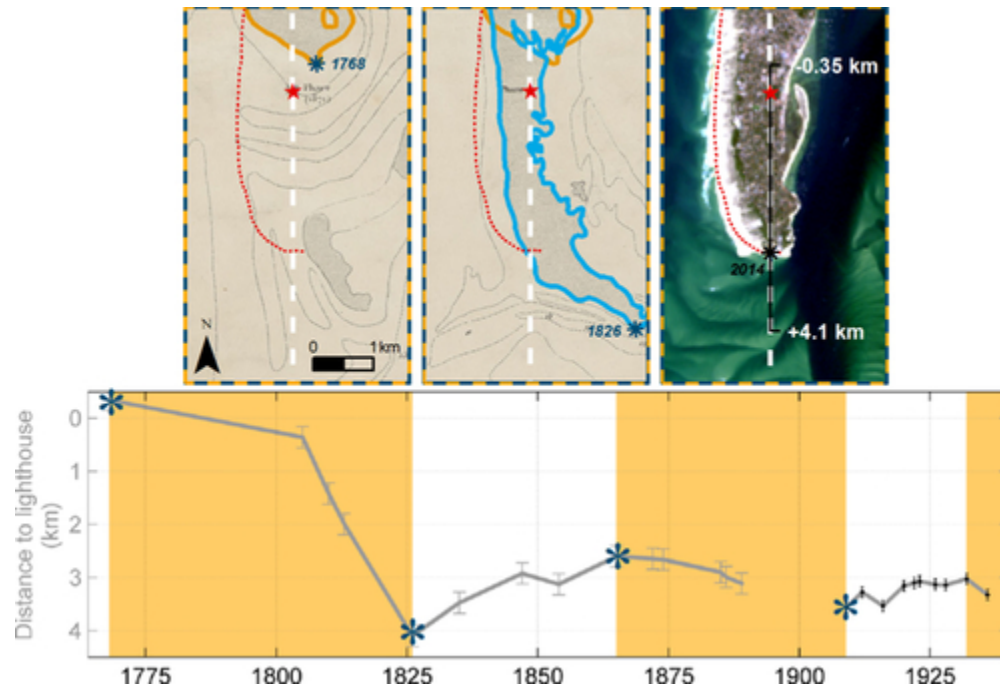
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Nautical charts. **a)** Ordnance survey map from 1888, black circled dots indicate control points used for georeferencing the map, red dashed line is the coastline measured by GPS on 2014 spring, red stars indicate the 3 reference locations used for georeferencing nautical charts; **b-e)** examples of nautical charts used to measure the spit-end positions, indicated by asterisk symbols for every given date; **f-g)** Satellite Landsat image of the Bay of Arcachon and its tidal inlet in October 2014.

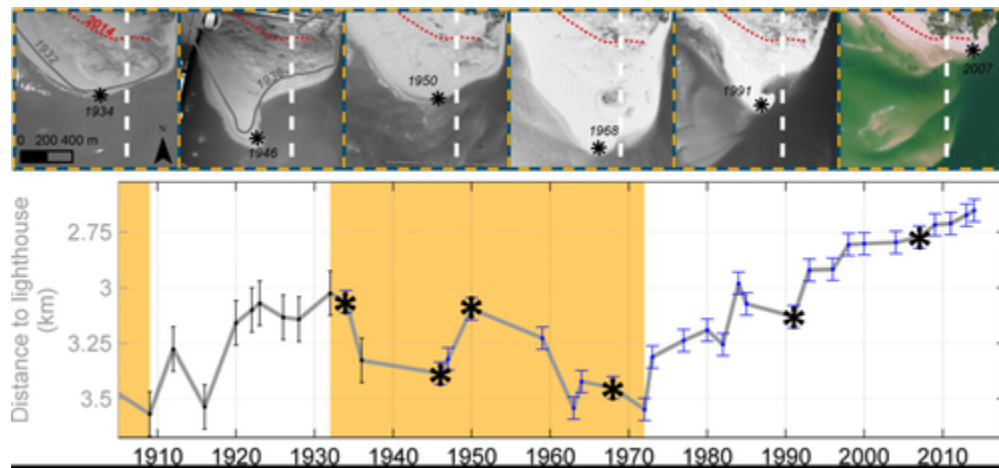
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Measuring the spit's extension relative to the position of Cap Ferret's lighthouse. **Upper panels:** charts and coastlines from 1768 and 1826 and Satellite Landsat 8 image from October 2014, red dotted lines represent the 2014 coastline and the red stars indicate the lighthouse's position, white dashed lines materialize the axe onto which spit-end positions were orthogonally projected and measured, the black sector in the right panel represents the spit-end migration range; **lower panel:** the grey curve shows the spit-end path between 1768 and 1936, grey error bars stand for historical charts and the black ones for modern charts, asterisk symbol stand for spit-end positions in Figure 2**b-e**.

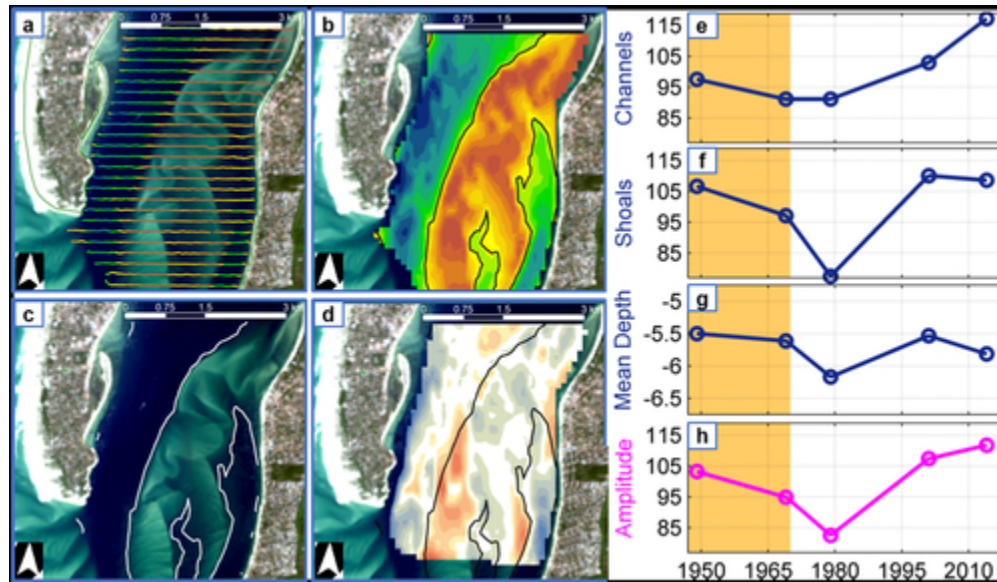
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Positions of the Cap Ferret spit-end measured on aerial photographs. **Upper panels:** series of aerial photographs of the spit-end, red dotted lines represent the 2014 coastline, black asterisk symbols materialize the spit-end positions for the given years, superimposed on photographs from 1934 and 1946 are the shorelines reproduced from charts of 1932 and 1936 respectively; **lower panel:** distance of the spit-end to lighthouse's position, black asterisk symbols correspond to those on the upper panel.

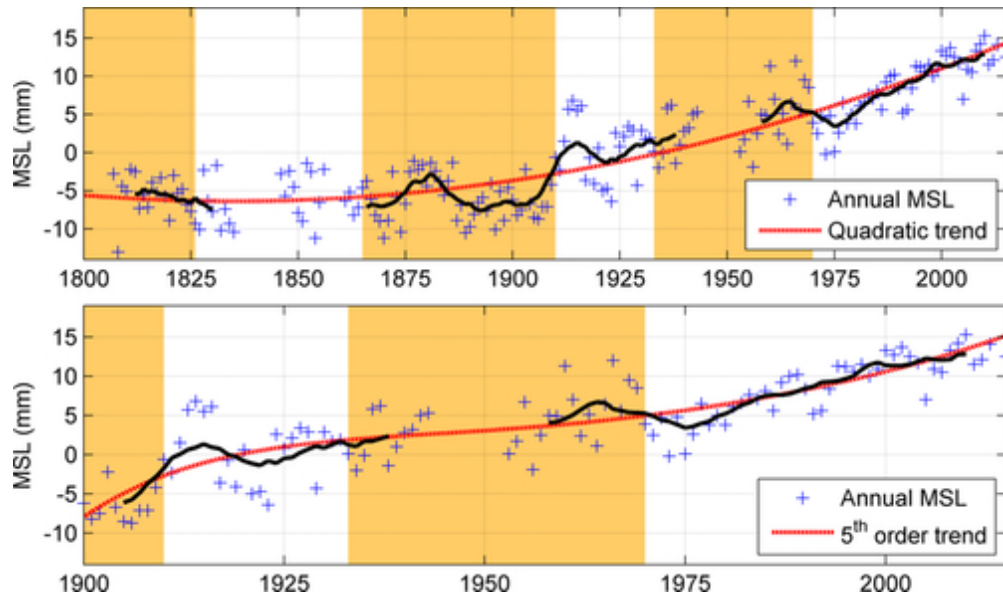
42x19mm (300 x 300 DPI)





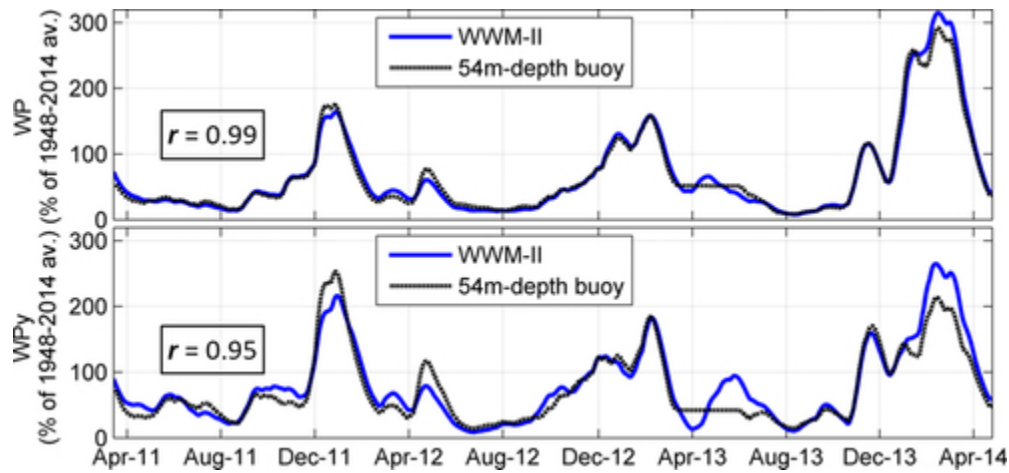
Inlet throat bathymetric data. **a)** summer 2014 along transect soundings; **b)** interpolated bathymetry with the -7 m NGF contours (black lines); **c)** -7 m NGF contours superimposed onto October 2014 Landsat image; **d)** erosion (blue) – deposition (red) patterns over the 12.74 km<sup>2</sup> area covered by all 5 surveys between 1949 and 2014; **e-h)** channel volume (below -7 m NGF), shoal volume (above -7 m NGF), overlapping area mean depth (m NGF) and inlet morphological amplitude over panel **d)**'s area (**e,f** and **h** are in percent of averaged values for all 5 surveys).

42x24mm (300 x 300 DPI)



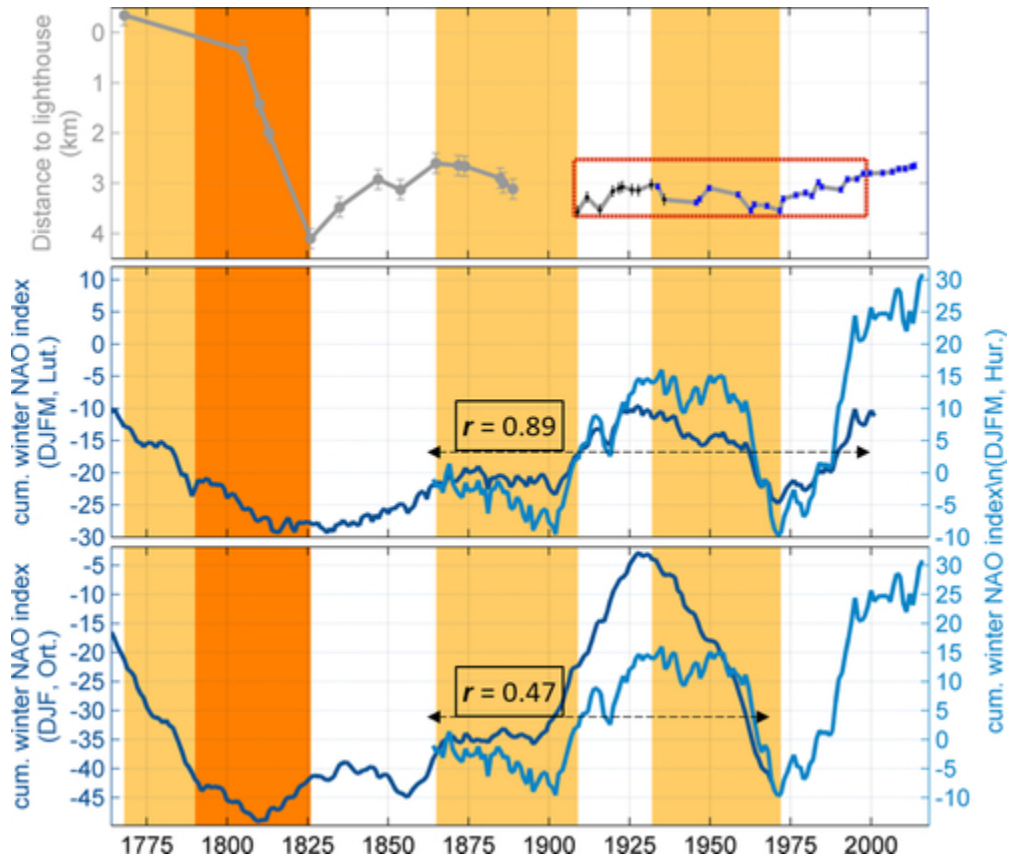
Brest tidal gauge Mean Sea Level (MSL) variation. **Upper panel:** annual MSL data from 1807 to 2015 recovered from the PMSL observatory (Holgate et al., 2013), black lines are the 11-year centred running mean of annual measurements and the thick dotted red line is the quadratic fit of all the data; **lower panel:** same as above with data from 20<sup>th</sup> century onward fitted with a 5<sup>th</sup> order polynomial function.

42x24mm (300 x 300 DPI)



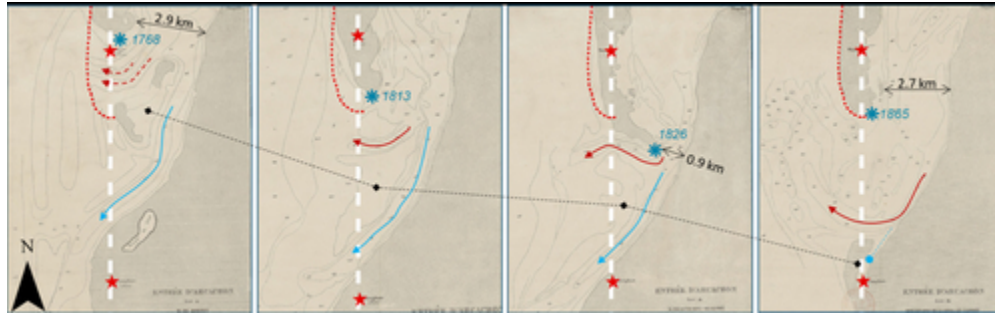
Local, 54 m water depth wave hindcast. **Upper panel:** 90-day running mean of the wave power (WP; linear wave theory approximation) 15 km offshore Cap Ferret, solid blue line are wind wave model (WWM-II, Roland et al., 2012) results with Bertin et al. (2015) setup, dotted black line are the wave buoy observations (CETMEF, n.d.) between April 2011 and May 2014, data is normalized using mean WP value over the hindcast period (1948-2014); **lower panel:** same as above with the alongshore component of the wave power (WPy);  $r$  values are Pearson's linear correlation coefficients between plotted curves.

42x19mm (300 x 300 DPI)



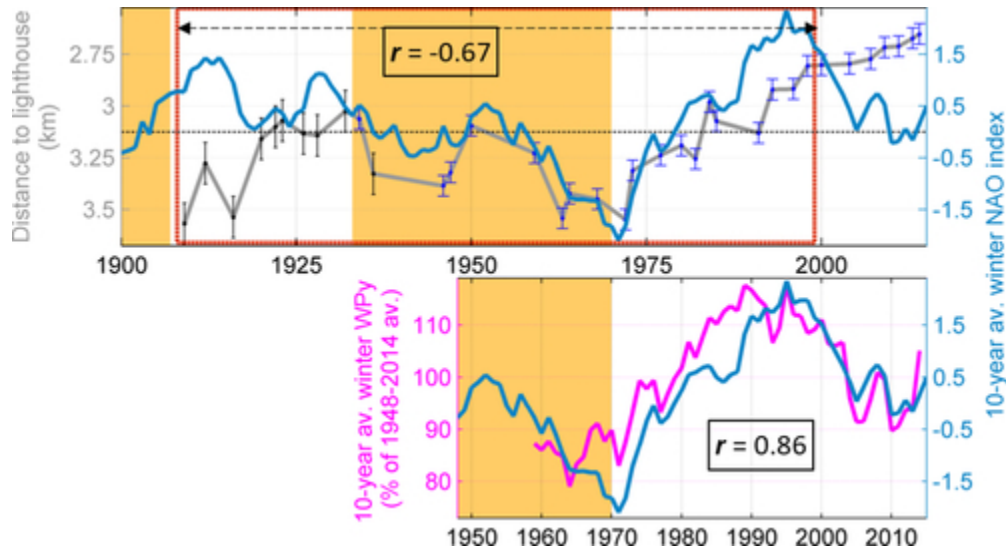
Spit-end path and cumulated winter NAO indices between 1768 and 2014. **Upper panel:** the grey curve shows the distance from the spit-end to the Cap Ferret lighthouse as on Figures 3&4 (red dotted frame is for next Figure 8's zoom); **mid panel:** darker blue curve is the cumulated winter (DJFM) index of the NAO reconstruction of Luterbacher et al. (1999), lighter blue curve is the same but for (Hurrell's (2015) station-based (DJFM) index; **lower panel:** darker blue curve is the cumulated winter (DJF) index of the NAO reconstruction of Ortega et al. (2015), lighter blue curve is the same but for Hurrell's (2015) station-based (DJFM) index; on all panels, orange shades indicate periods of spit elongation with the Dalton Minimum period in darker orange,  $r$  values are Pearson's linear correlation coefficients between plotted curves.

42x35mm (300 x 300 DPI)



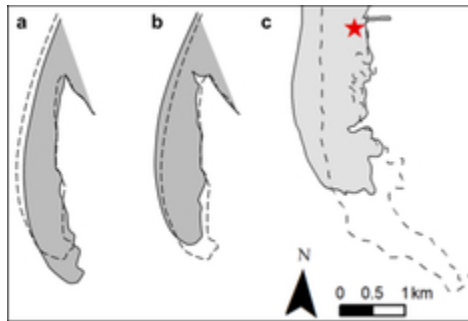
Evolution of the spit-inlet system across the Dalton Minimum (1780 – 1830). Navigation charts from years 1768, 1813, 1826 and 1865; blue asterisks indicate the positions of the spit terminus at the respective dates; black arrows materialize the inlet minimal width; red and blue arrows materialize the position of respectively the northern and southern channels; the dotted black line depicts the migration of the main inlet shoal barycentre; common red stars, dotted white axis and dotted red line are the same as in Figures 2 and 3.

42x13mm (300 x 300 DPI)



NAO relationships with Cap Ferret’s spit-end and local wave climate. **Upper panel:** in grey is the path of the spit-end, as the distance from the spit-end to the Cap Ferret lighthouse, as on Figures 3&4 and over since 1909, superimposed blue curve is Hurrell’s (2015) station-based NAO winter index, averaged of 10 years preceding each observation; **lower panel:** same winter NAO index curve as above, along with hindcasted alongshore winter wave power (WPy), averaged in same fashion as the NAO index and normalized using mean WPy value over the hindcast period (1948-2014); on both panels, orange shades indicate periods of spit elongation, *r* values are Pearson’s linear correlation coefficients between plotted curves.

42x22mm (300 x 300 DPI)



Sandspit's morphological response to changes in mean wave direction. **a**) spit-end response to a decrease of the averaged wave angle of incidence, modelled by (and reproduced from) Ashton et al. (2016), dashed line represents initial shoreline and shaded area represents the spit-end contours after waves became more shore normal; **b**) same as panel **a**, but for an increase of the averaged wave angle of incidence, or after waves became more oblique; **c**) dashed line represents Cap Ferret spit-end in 1826, at the end of the Dalton Minimum, and shaded area represents the spit-end in 1865.

19x13mm (300 x 300 DPI)