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The contribution of short-wave breaking to storm surges: the case Klaus in the Southern Bay of Biscay

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Abstract

This study investigates the contribution of short-wave breaking to storm surges through a high-resolution hindcast of the sea state and storm surge associated with the extra-tropical storm Klaus. This storm made landfall in January 2009 in the Southern Bay of Biscay and produced the largest storm surges observed in this region over the last 20 years, with 1.70 m in the Arcachon Lagoon and 1.10 m in the Adour Estuary. A fully-coupled 3D modelling system, which uses a vortex force formalism to represent wave-current interactions, is applied with a spatial resolution down to 35 m in the surf zones in order to properly compute the wave-induced setup. Modelling results reveal that the wave setup contributes by up to 40 % and 23 % to the storm surge peak in the Adour Estuary and the Arcachon Lagoon respectively. Accounting for wave forces in the circulation model improves storm surge predictions by 50 to 60 %. This is explained by the dominant role played by wave forces in the momentum balance at the inlets under storm waves. Numerical experiments further reveal that the wave-induced setup can be tidally-modulated, although this phenomenon seems to be site-specific. Finally, a sensitivity analysis highlights the importance of the model grid resolution in the surf zones to correctly resolve the wave setup along open-ocean coasts. Inside the lagoon, the storm surge and wave setup are

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less sensitive to the grid resolution while tidal propagation cannot be accurately represented with a resolution of 1000 m, which is typically used in operational storm surge forecast.

Keywords: Storm surge, Wave setup, SCHISM, Klaus

1. Introduction

Coastal flooding can be one of the most destructive natural catastrophes. In recent years, the combined effects of demographic growth and economic development of coastal zones with the ongoing sea level rise increased coastal flooding risk (Muis et al., 2016). This risk can be locally aggravated by land subsidence in some regions worldwide such as the Ganges-Brahmaputra Delta in Bangladesh (Karpytchev et al., 2018; Krien et al., 2019) or along the Mississippi Delta in the Gulf of Mexico (Letetrel et al., 2015). To assess future population changes in low-lying coastal zones, Neumann et al. (2015) conducted a global analysis combining socio-economic and sea level rise scenarios. These authors suggested that the number of people living in low-lying coastal zones in 2000 (~ 625 million) will increase by 50 % by 2030 and will double by 2060, which stresses the need to improve coastal communities resilience in the near future. On a more fundamental perspective, a better knowledge on the physical processes controlling storm-induced flooding is crucial to mitigate the consequences of these phenomena. Hurricane Katrina in the Gulf of Mexico (2005), storm Xynthia in the central part of the Bay of Biscay (2010), hurricane Sandy in the region of New York (2012) or typhoon Haiyan in the Philippines (2013) are major disasters which occurred over the past 20 years and illustrate this necessity.

Storm-induced coastal flooding results from extreme sea levels, which mostly occur when a high spring tide coincides with a large storm surge, although the importance of this combination depends on the ratio between the storm surge and the local tidal range. Storm surges correspond to variations of the ocean free-surface mainly caused by wind-induced surface stress and atmospheric pressure gradients associated with extra-tropical storms, tropical hurricanes and typhoons (Flather, 2001). Since the wind effect is inversely proportional to the water depth, low-lying coastal zones bordered by a large continental shelf, and located on storm tracks, are particularly vulnerable to storm surges and coastal flooding hazards.

While wind-induced surface stress and atmospheric pressure gradients have been identified as the main storm surge drivers since the early twentieth century (Doodson, 1924), the contribution of wind-generated surface waves

34 (hereafter short waves) to storm surges has received much less attention
35 and remains only partly understood. Charnock (1955) and Stewart (1974)
36 revealed that a young sea state can result in a higher surface stress and
37 thus a higher storm surge, which was corroborated by Donelan et al. (1993),
38 Mastenbroek et al. (1993), Brown and Wolf (2009), Nicolle et al. (2009)
39 and Bertin et al. (2015a) among others. Besides this effect, the breaking
40 of short waves in coastal zones drives an increase in the mean water levels
41 along the shoreline, referred to as wave setup. This phenomenon was first
42 explained physically by the radiation stress formalism of Longuet-Higgins
43 and Stewart (1962, 1964), which corresponds to the momentum flux associ-
44 ated with propagation of short waves. The dissipation of short-wave energy
45 in the nearshore induces spatial gradients of radiation stresses, which act as
46 a horizontal pressure force driving currents and a setup along the shoreline.
47 The absence of consensus on the representation of wave-current interactions
48 in 3D has long restricted the computation of wave setup to 2DH radiation
49 stress formalism. Over the last 15 years, new theories have emerged to
50 represent wave-current interactions in 3D (Mellor, 2003; McWilliams et al.,
51 2004; Ardhuin et al., 2008). Also, a few studies have shown that the depth-
52 varying circulation in surf zones can increase the maximum wave setup along
53 the shoreline of sandy beaches (Apotsos et al., 2007; Guérin et al., 2018).
54 However, the relevance of 3D fully-coupled models to compute storm surges
55 at regional scale with a resolution sufficiently fine to represent surf zones
56 has yet to be evaluated.

57 While wave setup on beaches is well documented and has been stud-
58 ied for several decades (e.g., see Holman and Sallenger Jr, 1985; Nielsen,
59 1988; King et al., 1990; Raubenheimer et al., 2001; Apotsos et al., 2007),
60 its correct representation in storm surge numerical models requires a good
61 description of the surf zones through refined meshes, which poses a serious
62 challenge in terms of computational time for regional applications. How-
63 ever, thanks to the recent development of parallel computing techniques
64 and the access to more computational resources, it is nowadays possible
65 to represent the wave setup in storm surge modelling systems at regional
66 scale (Dietrich et al., 2010; Bertin et al., 2015a; Krien et al., 2017) and bet-
67 ter understand its impact. Several authors revealed that the wave-induced
68 setup can substantially contribute to the storm surge under energetic wave
69 conditions, and even dominate the other drivers along coasts characterised
70 by narrow to moderately-wide shelves (Lerma et al., 2017) or at volcanic
71 islands (Kim et al., 2010; Kennedy et al., 2012; Pedreros et al., 2018). The
72 wave setup can range from several tens of centimetres to values of about 1 m
73 near the shoreline (Pedreros et al., 2018; Guérin et al., 2018) while regional

74 wave setup can reach tens of centimetres (Bertin et al., 2015a; Fortunato
75 et al., 2017). However, the contribution of wave setup in harbours where
76 tide gauges are usually located is not fully clear in the scientific community
77 (e.g., Thompson and Hamon, 1980). Melet et al. (2018) suggested that the
78 wave setup is negligible in most of the sheltered areas, while Aucan et al.
79 (2012) reported that the Midway tide gauge, located in the interior lagoon
80 of Midway Atoll in the Northern Hawaiian Islands, recorded high sea level
81 anomaly (SLA) events corresponding to the wave setup driven by breaking
82 waves during storms. The authors even suggested that the seasonal number
83 of SLA events recorded at this tide gauge can be used as an index of the
84 storminess in the Central North Pacific over climatic time-scales, as they
85 found a good correlation between the two.

86 Recently, several studies combining numerical modelling with field ob-
87 servations suggested that the breaking of short waves over the ebb deltas of
88 shallow inlets (Malhadas et al., 2009; Olabarrieta et al., 2011; Dodet et al.,
89 2013; Wargula et al., 2018) or large estuaries (Bertin et al., 2015a; Fortunato
90 et al., 2017) can induce a wave setup that extends at the scale of the whole
91 lagoon or estuary. Thus, modelling the wave setup appears to be fundamen-
92 tal for the prediction of flood inundation levels and floodplain management
93 of embayments, estuaries and river entrances (Hanslow and Nielsen, 1992).

94 This study presents a high-resolution hindcast of the sea state and storm
95 surge induced by the violent extra-tropical storm Klaus, which made landfall
96 in the Bay of Biscay on the 24th of January 2009. As Klaus produced the
97 most energetic waves ever recorded in the southern part of the bay, this storm
98 represents a unique opportunity to investigate the contribution of short-wave
99 breaking to storm surges. This process is examined in two sheltered areas
100 of the French Aquitanian Coast where Klaus drove the largest storm surges
101 observed over the last 20 years: the Arcachon Lagoon and the Adour Estu-
102 ary. A fully-coupled 3D modelling system with the vortex force formalism
103 of Bennis et al. (2011) is applied at the scale of the Bay of Biscay and the
104 English Channel. The relevance of the 3D model in terms of storm surge and
105 wave setup is compared against a conventional 2DH approach. Additional
106 numerical experiments are conducted in order to analyse the impact of the
107 wave forces on the momentum balance at the inlet of the Arcachon Lagoon
108 and their tidal modulation at both studied locations. Lastly, a sensitivity
109 analysis is carried out to analyse the impact of the grid resolution on storm
110 surge and wave setup predictions.

111 2. The studied area and storm

112 2.1. Study area

113 The Bay of Biscay is located in the North-East Atlantic Ocean, bordered
114 by France to the east and Spain to the south. The study area is the Aquitaine
115 coast in the south-eastern part of the Bay of Biscay, which comprises two
116 major geomorphologic settings: a first unit from the northern Spanish coast
117 to the Adour Estuary, characterised by rocky cliffs and small creeks, and
118 bordered by a continental shelf only 20 km-wide, and a second one from the
119 Adour Estuary to the Gironde Estuary, with a sandy coast bordered by a
120 continental shelf which width increases up to 150 km in front of the Gironde
121 Estuary. This study focuses on two specific locations, the Arcachon Lagoon
122 and the Adour Estuary further south (Fig. 1), which allows to investigate
123 the influence of short-wave breaking in areas sheltered from this process.

124 The Arcachon Lagoon (Fig. 1-B) is a semi-enclosed bay, which extends
125 at high-tide over an area of 160 km². The head of the embayment is occupied
126 by intertidal muddy and sandy flats that account for 75 % of the lagoon, and
127 divided by a large and complex network of secondary channels. The lagoon
128 is connected to the ocean by a 5 km-wide tidal inlet, bounded to the north
129 by the 18 km-long Cap Ferret sand-spit. The inlet is characterised by a well-
130 developed ebb-tidal delta covering 12 km², two deep channels, called North
131 Pass and South Pass, and a poorly-developed flood-tidal delta of 2.3 km²
132 (Michel and Howa, 1997). The tidal regime is semi-diurnal and mesotidal,
133 with a tidal range from 0.94 m to 4.93 m and a mean value of 2.94 m (Do-
134 det et al., 2019). The channels are tide-dominated, with currents 20-30 %
135 stronger in the North Pass than in the South Pass (Salles et al., 2015).
136 Because of the well developed ebb delta and the sandbar (continuation of
137 Cap-Ferret), the swells do not propagate inside the Arcachon Lagoon (Na-
138 hon, 2018) and the outer inlet can be often saturated with wave breaking
139 (Senechal et al., 2013). According to the hydrodynamic classification pro-
140 posed by Hayes (1980), the Arcachon Lagoon corresponds to a "transitional
141 inlet" under a "mixed-energy regime".

142 The Adour Estuary (Fig. 1-C), located approximately 40 km north of
143 the Spanish border, is defined by a narrower channel with a width varying
144 between 150 (inlet mouth) and 500 m over the last 6 km of the river. Two
145 breakwaters protect the entrance of the harbour of Bayonne from longshore
146 currents and swell waves and help stabilizing the navigation channel. The
147 influence of the breakwaters on the storm surge in the Adour Estuary will
148 be discussed later in this study. The tidal regime of the area is semi-diurnal
149 and mesotidal, with a tidal range varying from 0.78 to 4.32 m and a mean

150 value of 2.53 m (Dodet et al., 2019). Tidal currents are weak in the outer
 151 part of the estuary with values lower than $0.20 \text{ m}\cdot\text{s}^{-1}$ while in the river
 152 mouth, velocities reach values between 1 and $2 \text{ m}\cdot\text{s}^{-1}$ during spring tides
 153 (Brière, 2005). The river flow discharge ranges from 30 to $2000 \text{ m}^3\cdot\text{s}^{-1}$ with
 154 an annual mean of about $300 \text{ m}^3\cdot\text{s}^{-1}$ (Bellafont et al., 2018).

155 Dodet et al. (2019) analysed wave regimes along the metropolitan coasts
 156 of France and provided yearly means of wave parameters along the 30 m iso-
 157 bath line. According to their study, yearly-averaged significant wave height
 158 in front of Arcachon and Bayonne is about 1.65 m. Yearly averages of mean
 159 wave period and mean wave direction at Arcachon (respectively Bayonne)
 160 are about 6.3 s (resp. 7.15 s) and about 290° (resp. 310°). The wave cli-
 161 mate is however characterised by important seasonal variations: at Arcachon
 162 (resp. Bayonne), the significant wave height has a winter average of 2.08 m
 163 (resp. 2.06 m) and a summer average of 1.24 m (resp. 1.20 m) and the mean
 164 period decreases by 2.5 s (resp. 1.5 s) between winter and summer. Seaward
 165 of the Arcachon Lagoon, storm waves can exceed 9 m in water depths of
 166 26 m (Butel et al., 2002).

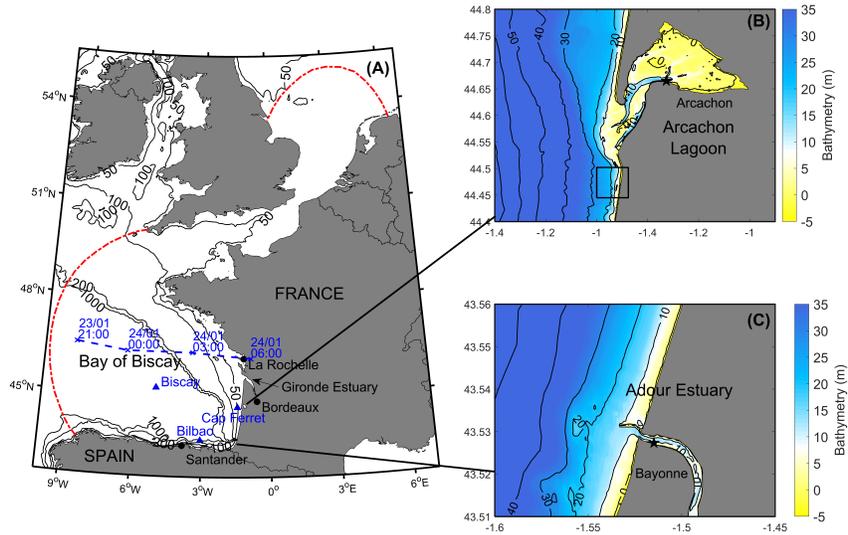


Figure 1: (A) Bathymetric map and extension of the computational domain (red dash-dotted line), the storm track (blue dashed line and crosses) and the wave buoys (blue triangles) used in this study. (B) and (C) Detailed bathymetry of the studied areas with location of the tide gauges (black stars). The black box in (B) corresponds to the adjacent beach where the sensitivity of storm surge and wave setup predictions to the grid resolution is analysed.

167 *2.2. The storm Klaus*

168 The extra-tropical storm Klaus hit the French coasts in the night of
169 the 23rd to the 24th of January 2009. It induced the largest storm surge
170 observed over the last 20 years in this region of the Bay of Biscay, with
171 1.10 m at Bayonne Boucau station and 1.70 m at Arcachon Eyrac station
172 (Mugica et al., 2010; Arnaud and Bertin, 2014). Previous long-term records
173 of wind speeds were exceeded in some French stations like Bordeaux and
174 Bayonne with wind gusts of 44-50 m.s⁻¹ and 33-39 m.s⁻¹ respectively. It
175 was considered as the most damaging wind storm to affect Northern Iberia
176 and Southern France since the destructive storm Martin in late December
177 1999 (Liberato et al., 2011). In 2009, Klaus was the most costly weather
178 events worldwide with over US\$ 6.0 billion in losses reported, mainly in
179 France and Spain (Aon-Benfield, 2010). Liberato et al. (2011) described
180 the storm from its genesis to its impact on the French and Spanish coasts
181 and the main features of its evolution are summarized here. Klaus was first
182 detected on 21 January 2009 as a small atmospheric wave perturbation. Due
183 to the southward displacement of the polar jet stream, the winter cyclone
184 moved eastward at an unusually low latitude (between 35°N and 45°N), on
185 the southern edge of the typical North-Atlantic storm track climatological
186 envelope. It underwent an explosive development on 23 January around
187 21°W, with a deepening rate of 37 hPa in 24 hours, probably supported by
188 an important tropical moisture export.

189 The storm rapidly reached the Bay of Biscay and followed a WNW to
190 ESE track toward the coasts (Fig. 1-A). The Spanish Oceanographic In-
191 stitute (IEO) registered two individual wave heights over 24 m from a buoy
192 35 km north of Santander between 06:00 and 07:00 in the morning of Jan-
193 uary 24th. Bilbao and Cap-Ferret buoys recorded significant wave heights
194 reaching 13 m with a peak period of 15 s during the storm. The centre of
195 the low-pressure system passed at 5:00 am on January 24th over La Rochelle
196 with a minimum of 965.8 hPa recorded at the nearby station of Chassiron
197 (Fig. 1-A). The highest sustained wind speeds were measured further south,
198 with a maximum of 36 m.s⁻¹ at Cap-Ferret station (Arcachon Lagoon) for
199 a lowest pressure of 976 hPa. At Bayonne, sustained wind speed reached
200 21 m.s⁻¹ with a minimum pressure of 983.6 hPa.

201 **3. Methods and data**

202 *3.1. The modelling system*

203 *3.1.1. Overview of the modelling system*

204 This study uses the modelling system SCHISM (Semi-implicit Cross-
 205 scale Hydroscience Integrated System Model) of Zhang et al. (2016) which
 206 is a 3D unstructured-grid model. The model uses a combination of a semi-
 207 implicit scheme and an Eulerian-Lagrangian Method to treat the momen-
 208 tum advection, which allows to relax the associated numerical stability con-
 209 straints. Compared to the original model SELFE from which it is derived
 210 (Zhang and Baptista, 2008), SCHISM now integrates many enhancements
 211 and upgrades including new extension to large-scale eddying regime and a
 212 seamless cross-scale capability from creek to ocean (Zhang et al., 2016; Ye
 213 et al., 2020). A detailed description of SCHISM, the governing equations
 214 and its numerical implementation can be found in Zhang et al. (2015, 2016).
 215 The hydrostatic solver of SCHISM can be coupled with other modules incor-
 216 porated in the modelling system such as short waves, sediment transport,
 217 water quality, oil spills and biology. The generation and propagation of
 218 short waves are simulated with the Wind Wave Model WWMII of Roland
 219 et al. (2012). In this study, the contribution of short-wave breaking to storm
 220 surges is analysed from 3D fully-coupled (wave-current) simulations. The
 221 hydrodynamic and spectral wave models share the same unstructured grid
 222 and domain decomposition, which reduces the exchange of information be-
 223 tween the models and eliminates errors associated with interpolation.

224 *3.1.2. Vortex force formalism*

225 In the modelling system, the 3D wave-current interactions are repre-
 226 sented with the vortex force formalism proposed by Ardhuin et al. (2008),
 227 as described in Bennis et al. (2011). Its detailed implementation in SCHISM
 228 can be found in Gu erin et al. (2018). In the vortex force framework, the
 229 mass conservation and momentum equations of the hydrodynamic model
 230 read:

$$\nabla \cdot \hat{\mathbf{u}} = 0 \tag{1}$$

$$\frac{D\hat{u}}{Dt} = f\hat{v} - \frac{1}{\rho} \frac{\partial P_A}{\partial x} - g \frac{\partial \zeta}{\partial x} + \frac{\partial}{\partial z} \left(\nu \frac{\partial \hat{u}}{\partial z} \right) + F_{wave,x} \tag{2}$$

$$\frac{D\hat{v}}{Dt} = -f\hat{u} - \frac{1}{\rho} \frac{\partial P_A}{\partial y} - g \frac{\partial \zeta}{\partial y} + \frac{\partial}{\partial z} \left(\nu \frac{\partial \hat{v}}{\partial z} \right) + F_{wave,y} \tag{3}$$

233 In Eq. (1), $\nabla = (\frac{\partial}{\partial x}, \frac{\partial}{\partial y}, \frac{\partial}{\partial z})$ and $\hat{\mathbf{u}} = (\hat{u}, \hat{v}, \hat{w})$ is the quasi-Eulerian veloc-
 234 ity, equal to the mean Lagrangian velocity $\mathbf{u} = (u, v, w)$ minus the Stokes

235 velocity $\mathbf{u}_s = (u_s, v_s, w_s)$. In Eqs. (2) and (3), f is the Coriolis parameter,
 236 ρ is the water density, P_A is the sea-level atmospheric pressure, g is the
 237 acceleration caused by gravity, ζ is the free surface elevation and ν is the
 238 vertical eddy viscosity. $F_{wave,x}$ and $F_{wave,y}$ are the two components of the
 239 wave forces, given by:

$$F_{wave,x} = v_s \left[f + \left(\frac{\partial \hat{v}}{\partial x} - \frac{\partial \hat{u}}{\partial y} \right) \right] - w_s \frac{\partial \hat{u}}{\partial z} - \frac{\partial J}{\partial x} + \hat{F}_{d,x} \quad (4)$$

240

$$F_{wave,y} = -u_s \left[f + \left(\frac{\partial \hat{v}}{\partial x} - \frac{\partial \hat{u}}{\partial y} \right) \right] - w_s \frac{\partial \hat{v}}{\partial z} - \frac{\partial J}{\partial y} + \hat{F}_{d,y} \quad (5)$$

241 where J is the wave-induced mean pressure, and $\hat{\mathbf{F}}_d$ is the wave-induced
 242 non conservative forces due to depth-induced wave breaking. A detailed
 243 description of the wave-induced non conservative forces can be found in
 244 Gu erin et al. (2018).

245 3.1.3. Model parametrisations for 2DH and 3D models

246 There are noticeable differences between 2DH and 3D configurations.
 247 In 2DH, the model uses a Manning coefficient and the depth-integrated
 248 current velocity to evaluate the bottom stress, while in 3D, the model uses
 249 the bottom roughness and the velocity computed at the top of the bottom
 250 cell. In the 3D model, several parametrisations are available to compute
 251 the wave-enhanced bottom stress but a sensitivity analysis did not result
 252 in significant improvements, which corroborates the findings of Bertin et al.
 253 (2015a) in the central part of the Bay of Biscay. Therefore, wave effects
 254 on the bottom stress are not considered in the study. In the 3D model,
 255 the wave effects on vertical mixing are integrated in the turbulence closure
 256 scheme (Umlauf and Burchard, 2003) following the approach of Moghimi
 257 et al. (2013), as described in Gu erin et al. (2018). For both 2DH and 3D
 258 models, the surface stress can be computed with a bulk formula of the form
 259 $\rho_a C_d U_{10}^2$, where U_{10} is the 10 m wind speed and C_d is the drag coefficient
 260 calculated with the formulation of Hwang et al. (2019). The surface stress
 261 can also be computed using a wave-dependent parametrisation using the
 262 friction velocity U_* calculated in WWMII. Donelan et al. (1993) reported
 263 that a young sea state enhances the sea surface roughness. In order to
 264 correctly represent this process and predict the subsequent storm surge,
 265 Mastenbroek et al. (1993) and Bertin et al. (2015a) showed that a wave-
 266 dependent surface stress is required. The influence of the surface stress
 267 formulation will be discussed later in this study.

268 *3.1.4. The spectral wave model*

269 WWMII solves the equation for the conservation of the wave action
270 (e.g., see Komen et al., 1994) to simulate the generation and propagation
271 of wind-generated waves. The model accounts for wind input and energy
272 dissipation by whitecapping, computed according to Ardhuin et al. (2010),
273 energy dissipation due to bottom friction, which is modelled based on the
274 results obtained during the JONSWAP project (Hasselmann et al., 1973),
275 and depth-induced breaking computed according to the model of Battjes
276 and Janssen (1978), which is parametrized with the breaker index γ and the
277 dissipation coefficient B . As wave measurements in the surf zone during the
278 storm were not available, γ and B are set to the default values of 0.73 and 1
279 respectively, which will be discussed later. Finally, the non-linear wave-wave
280 interactions are calculated following the Discrete Interaction Approximation
281 of Hasselmann et al. (1985) and the Lumped Triad Approximation of El-
282 deberky (1996) in deep water and shallow water respectively. A detailed
283 description of the coupling between SCHISM and WWMII can be found in
284 Roland et al. (2012) and Schloen et al. (2017). At the coupling time step,
285 SCHISM provides WWMII with fields of 2DH currents and water levels
286 while SCHISM receives wave forces from WWMII.

287 *3.1.5. Model implementation*

288 The unstructured computational grid used to perform the hindcast of
289 the storm covers the whole Bay of Biscay from 10°W to the French coasts,
290 the English Channel and a part of the North Sea (up to 55°N) (Fig. 1-A).
291 The grid has ~ 281000 nodes in the horizontal, with a spatial resolution
292 ranging from 5000 m along the open boundary to 35 m along the shoreline
293 of the studied areas (*i.e.* the Arcachon Lagoon and the Adour Estuary). In
294 the vertical, the grid is discretized in 35 S levels for the 3D simulations.

295 The circulation model is forced at its open boundaries by the 16 main
296 astronomical constituents linearly interpolated from the regional model of
297 Bertin et al. (2012). The tidal potential is switched off since a sensitivity
298 analysis revealed a negligible effect on tidal predictions. After calibration of
299 the tidal model, the bed roughness in the 3D model is set to 0.0001 m in the
300 open ocean and 0.002 m in the Arcachon Lagoon and the Adour Estuary. In
301 the 2DH model, a Manning coefficient of 0.02 is employed for the open ocean
302 while a value of 0.029 is considered for the Adour Estuary and the Arcachon
303 Lagoon. The Manning coefficient used for the Arcachon Lagoon is between
304 the values used by Cayocca (1996) (~ 0.028) and Nahon (2018) (0.032).
305 The simulations are started on the 22th of January 2009, two days before
306 the peak of the storm and last 4 days. The time step is set to 60 s for both

307 the hydrodynamic and the wave models, in the 2DH and 3D simulations.

308 Over the whole domain, the circulation model is forced by hourly 10 m
309 wind speed and sea-level pressure fields from the Climate Forecast System
310 Reanalysis CFSR (Saha et al., 2010). The datasets are provided on a reg-
311 ular grid with a spatial resolution of 0.312° and 0.5° for the wind and the
312 atmospheric pressure respectively. WWMII is forced with the CFSR wind
313 fields over the whole domain. WWMII is also forced along the open bound-
314 aries by time series of directional wave spectra, previously computed from
315 a regional application of the WaveWatchIII (WWIII) spectral wave model
316 described in Bertin et al. (2015a). Wind fields from CFSR are also used to
317 run the WWIII model over the North Atlantic Ocean.

318 *3.2. Wave and water level observations*

319 The accuracy of the wave predictions is evaluated with the measurements
320 recorded by three buoys in the Bay of Biscay (see Fig. 1-A, for their loca-
321 tion). The Biscay buoy is a non-directional buoy located by 4500 m depth,
322 operated by Météo-France and UK Met Office. The Cap Ferret and Bilbao
323 buoys are located in more intermediate water depths of the southern part
324 of the Bay of Biscay (depths of 50 m and 600 m respectively) and are op-
325 erated by CEREMA and Puerto del Estado respectively. The three buoys
326 provide time series of significant wave height (H_s) while the mean wave pe-
327 riod (T_{m02}) is available at Cap Ferret and Biscay buoys and the peak wave
328 period (T_p) at Bilbao buoy. Wave bulk parameters are estimated every 60
329 minutes at Biscay and Bilbao buoys and every 30 minutes at Cap Ferret
330 buoy. Since the atmospheric data used to force the model has a hourly time
331 resolution, the wave predictions cannot represent the sub-hourly variability
332 and the measurements at Cap Ferret buoy are therefore averaged over one
333 hour to yield a consistent comparison with the model predictions.

334 Simulated water levels are validated through a comparison against obser-
335 vations recorded with a 10-min sampling interval during the storm period
336 at the two tide gauges of Arcachon Eyrac and Bayonne Boucau (see Fig.
337 1-B and 1-C, for their respective location). A tidal prediction is obtained
338 based on a 5 year-long time series (2008-2012) with a harmonic analysis
339 using the UTide code (Codiga, 2011). Tides are reconstructed with the
340 67 main astronomical constituents previously computed. Note that in the
341 North-East Atlantic Ocean, the constituent Sa results from a combination
342 of thermo-steric and atmospheric effects (Bertin et al., 2015b; Payo-Payo
343 and Bertin, 2020). Therefore, it is not included in the tidal prediction since
344 storm surges are computed as the difference between the observed water
345 level and the astronomic tidal prediction.

346 **4. Modelling results**

347 *4.1. Atmospheric forcing*

348 In order to validate the atmospheric forcing originating from CFSR,
 349 a comparison is performed against field observations available during the
 350 storm and collected at the meteorological stations of Cap Ferret and Bayonne
 351 (see Fig. 1-A, for their location). The comparison (Fig. 2) of modelled
 352 against observed 10 m wind speeds (hereafter U_{10}) and sea-level pressure
 353 (hereafter SLP) reveals that SLP is well reproduced with a root mean square
 354 error (hereafter RMSE) lower than 1.5 hPa at both locations. At Cap Ferret,
 355 U_{10} is accurately predicted with a RMSE of 2.3 m.s^{-1} , although with a slight
 356 underestimation of approximately 4 m.s^{-1} two hours before the peak of the
 357 storm. Since the meteorological station at Bayonne is located at 75 m above
 358 sea level and 3 km inland, the model, providing 10 m wind speed with a 0.3°
 359 resolution, does not accurately reproduce the observations, which probably
 360 explains the positive bias of 1.6 m.s^{-1} . Overall, it should be noted that
 361 the intensity of the storm is correctly represented: peak values of U_{10} are
 362 reasonably predicted with stronger values at Cap Ferret (34 m.s^{-1}) than at
 363 Bayonne (22 m.s^{-1}).

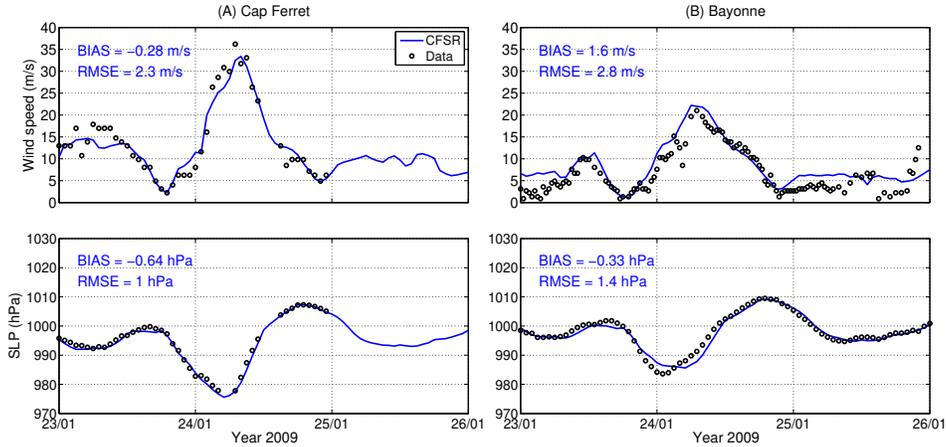


Figure 2: Modelled (blue solid line) against observed wind speeds and sea level pressure (black dots) at Cap Ferret (A) and Bayonne (B) stations.

364 *4.2. Wave predictions*

365 Modelled wave bulk parameters are compared against the measurements
 366 available during Klaus at Cap Ferret, Bilbao and Biscay buoys (Fig. 3). The
 367 comparison reveals a good agreement between modelled and measured data:

368 H_s is well reproduced with a RMSE ranging from 0.51 to 1 m which corre-
 369 sponds to a 10-17 % error once normalized by the mean of the observations
 370 (hereafter NRMSE). However, for the three stations, the model displays a
 371 positive bias of 0.35-0.50 m at Cap Ferret and Biscay buoys and 0.70 m at
 372 Bilbao buoy. It should be noted that the larger error at Bilbao buoy is
 373 partly due to a one-hour time lag, representing 35 % of the bias and the
 374 NRMSE, which we are unable to explain. The model correctly captures the
 375 peak storm wave height with less than 10 % error at the three buoys. T_{m02} ,
 376 available at Cap Ferret and Biscay buoys, is well predicted with a NRMSE
 377 less than 6 % while at Bilbao buoy, T_p is adequately reproduced with a 10
 378 % NRMSE.

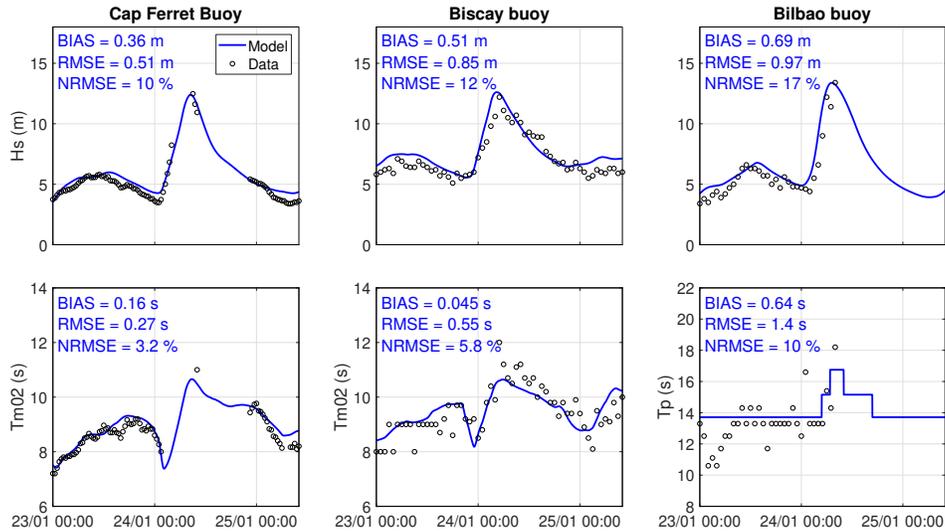


Figure 3: Observed (black dots) against modelled wave parameters (blue solid line) at Cap Ferret, Biscay and Bilbao buoys during Klaus.

379 4.3. Storm surge and water level predictions

380 A tide-only simulation is first performed and the modelled water levels
 381 are compared against the tidal predictions based on the observations at each
 382 station. The tidal forcing together with the distribution of the Manning
 383 coefficient yields good results with a RMSE on tides of 0.11 m at Bayonne
 384 and 0.08 m at Arcachon (not shown).

385 The effect of the parametrisation of the surface stress on the storm
 386 surge is investigated by comparing simulations using the bulk formula of
 387 Hwang et al. (2019) and the wave-dependent approach (see Section 3.1.3).

388 This comparison reveals moderate differences between both parametrisations
 389 (lower than 0.05 m), with the predictions of the model using the bulk
 390 formula slightly better matching the observations. To explain the negligible
 391 effects of the wave-dependent approach on the storm surge predictions, the
 392 sea state is characterised by the wave age, defined as C_p/U_{10} where C_p is the
 393 peak wave phase speed. Considering a 20-hour window centred on the storm
 394 peak, the wave age varies from 0.7 to 2.3, with an average value of 1.32 (with
 395 a standard deviation of 0.5), which is characteristic of a mature sea state
 396 and explains the very slight impact of the wave-dependent approach on the
 397 results. This behaviour corroborates the study of Bertin et al. (2015a), who
 398 showed that the surface stress was little dependent on the sea state for the
 399 storm Joachim, characterised by comparable wave height and peak period
 400 as during Klaus. According to these results, the bulk formula of the surface
 401 stress is adopted in the rest of the study.

402 The contribution of short-wave breaking to the storm surge is analysed
 403 by comparing a first simulation without wave forces and a fully-coupled
 404 simulation, *i.e.* including wave forces, hereafter referred to as the baseline
 405 model. The modelled storm surges are obtained by subtracting the tide-only
 406 simulation to each case of simulation.

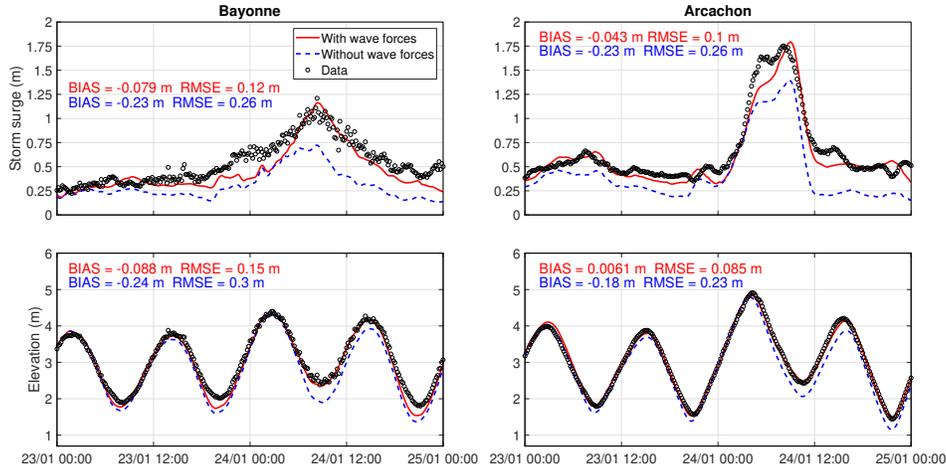


Figure 4: Observed (black dots) against modelled storm surges with the baseline model (red solid line) and the model without wave forces (blue dashed line) at Bayonne (left) and Arcachon (right).

407 The results are presented in Figure 4, where each simulation is compared
 408 against storm surges and water levels observed at Bayonne and Arcachon
 409 during the storm. The baseline model accurately reproduces the water levels

410 with a RMSE of 0.09 and 0.15 m at Arcachon and Bayonne respectively.
411 The storm surges are well predicted by the model, with a RMSE of 0.12 m
412 at Bayonne and 0.10 m at Arcachon, although a 0.25 m underestimation
413 is noticed at this station approximately two hours before the storm peak.
414 Without wave forces, storm surge and water level predictions considerably
415 deteriorate compared to the baseline model with a RMSE two to three times
416 larger at both locations. The modelled water levels display a negative bias
417 ranging from 0.18 to 0.24 m. The surge peak is underestimated by 0.40-
418 0.45 m at Arcachon and Bayonne, which results in a negative bias of 0.23 m
419 over the duration of the storm.

420 The comparison of both simulations reveals that the wave setup driven
421 by the wave forces in the baseline model accounts for 40 % and 23 % of
422 the surge peak in the Adour Estuary and the Arcachon Lagoon respectively,
423 which explains that the baseline model much better matches the observed
424 peak values.

425 In order to get a spatial overview of this process, modelled storm surges
426 with and without wave forces, as well as their difference, are computed at
427 the scale of the Arcachon Lagoon and the Adour Estuary (Fig. 5). In the
428 Adour Estuary, the storm surge in the fully-coupled simulation increases by
429 more than 0.5 m at adjacent beaches while being almost constant inside the
430 estuary (Fig. 5-a). The comparison between Figure 5-c (atmospheric surge
431 only) and Figure 5-e (wave setup only) reveals that this behaviour is due to
432 the development of a wave setup along adjacent shorelines, reaching up to
433 0.75 m and extending at the scale of the whole estuary where it raises the
434 water level by 0.45 m. A different pattern can be observed in the Arcachon
435 Lagoon, where the storm surge in the fully-coupled simulation increases from
436 the inlet to the lagoon head (Fig. 5-b). The comparison between Figure 5-
437 d and Figure 5-f suggests that this behaviour results from the increase in
438 atmospheric surge towards the lagoon head combined with the development
439 of a wave setup reaching 0.40 m at the scale of the lagoon. As in the Adour
440 Estuary, the wave setup develops at the inlet and then exhibits a plateau
441 inside the lagoon. Along the adjacent shorelines of the lagoon, the maximum
442 wave setup reaches 0.80 m (The maximum wave setup along the adjacent
443 shorelines are not shown in Figures 5-e and 5-f as computational nodes dry
444 in the tide-only simulation are not represented).

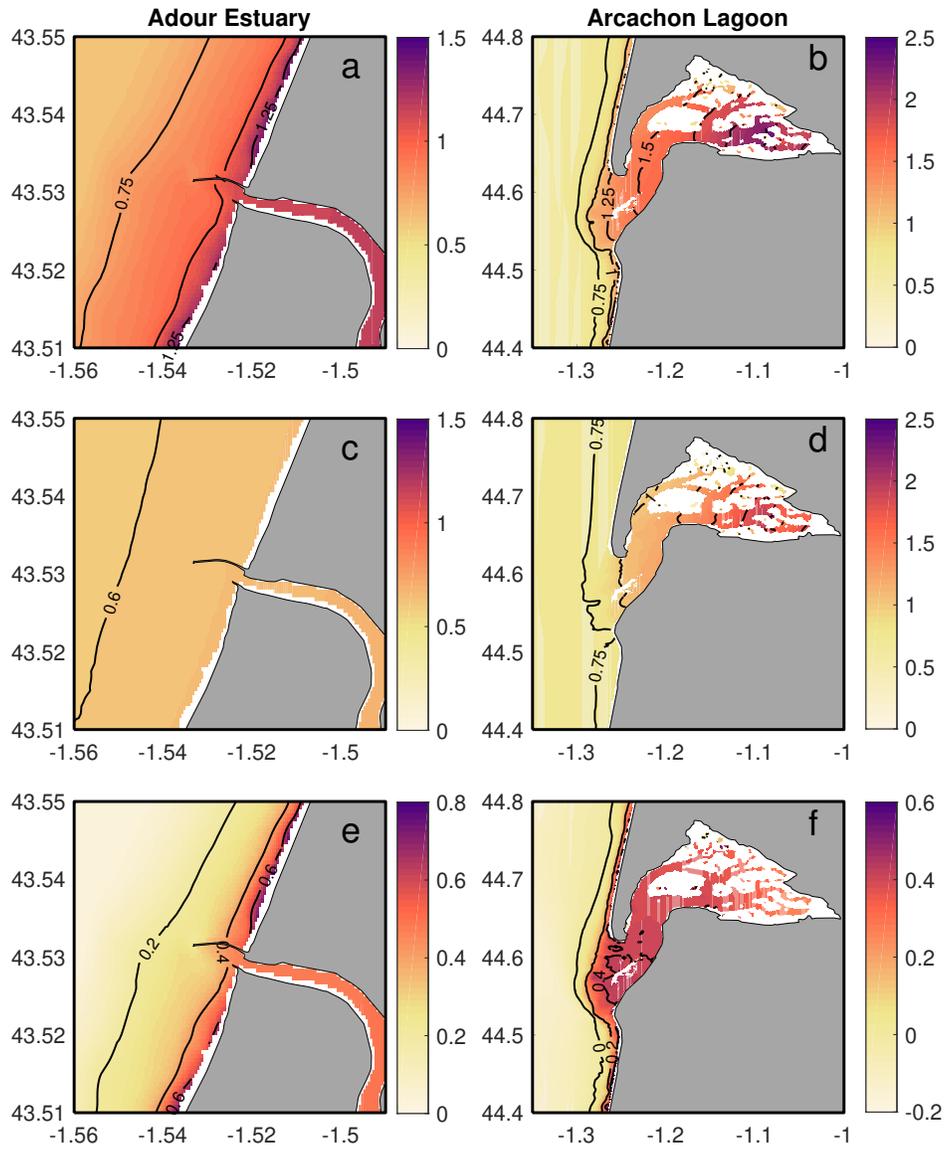


Figure 5: Storm surge (in m) simulated with wave forces (a,b), without wave forces (c,d) and their difference (e,f), at the Adour Estuary (left) and the Arcachon Lagoon (right). The white color corresponds to dry nodes in the tide-only simulation.

445 **5. Contribution of wave breaking to storm surges**

446 *5.1. Model predictive skills*

447 Wave parameters are accurately reproduced by the model and corre-
448 spond to the state-of-the-art considering previously published studies led
449 under storm wave conditions (e.g. Kerr et al., 2013; Bertin et al., 2012;
450 Staneva et al., 2016). Storm surges are also well predicted, with errors simi-
451 lar or even lower compared to previously published studies (e.g. Kerr et al.,
452 2013; Brown et al., 2013; Bertin et al., 2015a). In details, the storm surge
453 is underestimated by up to 0.25 m during the first part of the storm peak
454 at Arcachon. This can be explained by an underestimation of the CFSR
455 sustained wind speeds by up to 4 m.s^{-1} during this period (Fig. 2), which
456 leads to a wind-induced surge lower than expected. This hypothesis was
457 tested by correcting wind speeds empirically on the time steps correspond-
458 ing to this period (cf. Appendix A). The results reveal that this correction
459 almost cancels out the local underestimation in the surge, thus supporting
460 this hypothesis. In the Adour Estuary, model results at an earlier stage of
461 this study showed a 0.05 to 0.1 m negative bias in the storm surge before the
462 storm peak when the breakwaters bounding the estuary mouth were con-
463 sidered as impermeable wall. In fact, these breakwaters are made of large
464 blocks (4 to 40 tons) that allow large amounts of water to flow through the
465 structures when a gradient in water levels exists on both sides of the struc-
466 ture (Prof. Abadie, pers. com.). Such flows can take place when a wave
467 setup develops at adjacent beaches, a process already reported at other engi-
468 neered estuaries (Hanslow and Nielsen, 1992; Hanslow et al., 1996). In order
469 to account for these possible flows, we took advantage of hydraulic structure
470 options implemented in the code. Although this parametrisation improves
471 storm surge prediction by 0.04 m, verifying the adequate representation of
472 these flows is outside the scope of the study and would deserve a specific
473 analysis.

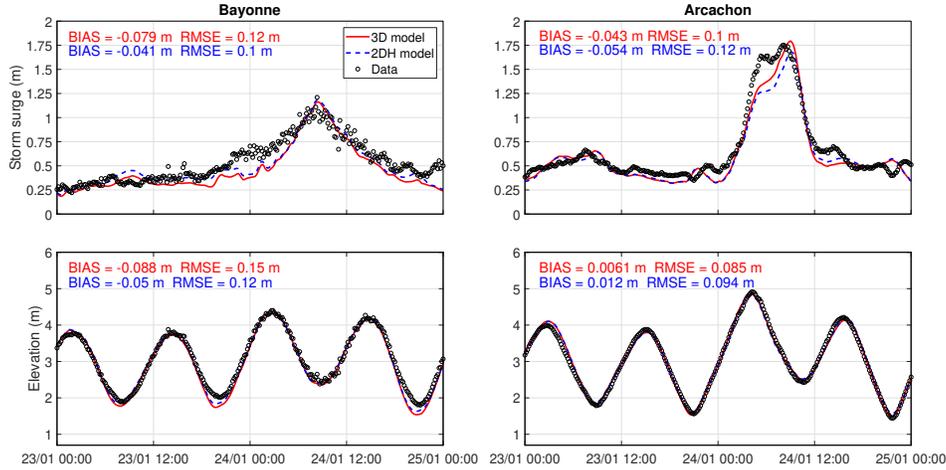


Figure 6: Observed (black dots) against modelled storm surges with the 3D baseline model (red solid line) and the 2DH model (blue dashed line) at Bayonne (left) and Arcachon (right).

474 The comparison of the results between the baseline model and the model
475 without wave forces (Fig. 4) reveals that including wave forces in the cir-
476 culation model substantially improves its predictive skills. The analysis of
477 the different terms included in wave forces (Eq. 4 and 5) shows that the
478 wave dissipation term by depth-induced breaking is clearly dominant over
479 the vortex force and the wave mean pressure terms. In accordance with pre-
480 vious studies (Staneva et al., 2016), this analysis highlights the importance
481 of accounting for short waves in storm surge modelling systems, provided
482 that wave energy dissipation due to wave breaking is correctly represented.
483 Gu erin et al. (2018) investigated wave-induced circulation in a surf zone with
484 varying bed slope. The authors computed the wave breaking process accord-
485 ing to the model of Thornton and Guza (1983) in which they calculated the
486 breaking index γ and the dissipation coefficient B as a linear function of the
487 beach slope. The authors showed that this adaptive approach improved the
488 predictions of short-waves bulk parameters and wave setup by 30 %. Follow-
489 ing this study, Pezerat et al. (2020) showed that a dissipation coefficient B
490 taken at 40 times the local bed slope strongly improves wave predictions at
491 gently sloping shorefaces ($\sim 1:1000$). At both study sites, bottom slopes are
492 much steeper (1:50 to 1:100), so that this adaptive parametrisation results
493 in values for B close to the default value of 1. Indeed, a sensitivity analysis
494 shows that the adaptive parametrisation of Pezerat et al. (2020) yields very
495 similar short wave and setup predictions compared to the default values for

496 γ and B in the model of Battjes and Janssen (1978). New field experiments
497 are required to investigate further wave dissipation mechanisms in coastal
498 zones and validate the numerical model under very high energetic conditions,
499 although such field deployments remain very challenging.

500 Finally, the comparison between 2DH and 3D simulations reveals only
501 modest differences, with water level and surge predictions slightly improved
502 in 3D in Arcachon and slightly deteriorated in Bayonne (Fig. 6). In the
503 Arcachon Lagoon, improved storm surge predictions are obtained before
504 and during the storm peak, when winds blow from SW to W and drive an
505 Ekman transport towards the coast, a process better represented with a 3D
506 model (Roland et al., 2012). In the case of the Adour Estuary, maximum
507 wave setup at adjacent beaches is slightly lower in 2DH compared to 3D but
508 extends further offshore, thereby more impacting water levels in the estuary.
509 Gu erin et al. (2018) showed that the depth-varying circulation driven by
510 short waves in surf zones can increase the wave setup along the coast but
511 this process is only substantial at steep beaches (i.e. mean slope of 1:30 and
512 over). Also, these authors reported that 3D runs yield larger wave setup
513 compared to 2DH runs very close to the shoreline, so that reproducing these
514 differences requires a spatial resolution of a few meters, that is one order a
515 magnitude finer than in this study.

516 5.2. Momentum balance

517 Previous studies already reported the development of a wave setup in
518 inlets, river entrances and shallow lagoons (Hanslow and Nielsen, 1992;
519 Hanslow et al., 1996; Dunn et al., 2001; Oshiyama et al., 2001; Tanaka et al.,
520 2001, 2003; Nguyen et al., 2007; Bertin et al., 2009; Malhadas et al., 2009;
521 Olabarrieta et al., 2011; Dodet et al., 2013; Wargula et al., 2018), which
522 can be further investigated by analysing the momentum balance at the in-
523 let. Hench and Luettich Jr (2003) analysed the momentum balance without
524 waves in the Beaufort Inlet in North Carolina and in an idealized inlet and
525 reported that near maximum flood and ebb, the along-stream momentum
526 balance in both cases is dominated by advection, barotropic pressure gradi-
527 ent and bottom friction. Olabarrieta et al. (2011) corroborated these results
528 in a study conducted in Willapa Bay (USA) during a storm event. By ac-
529 tivating the wave forces in their fully-coupled modelling system, they also
530 revealed that they can substantially change the barotropic pressure gradi-
531 ent and the bottom friction while being one of the dominant terms in the
532 momentum balance in the inlet area. These findings were then corroborated
533 by Dodet et al. (2013) and Wargula et al. (2014). In particular, Dodet et al.

534 (2013) combined both modelling and observations to study wave-current in-
535 teractions on the Albufeira Lagoon, a shallow wave-dominated tidal inlet in
536 Portugal, during energetic oceanic swells conditions. The authors showed
537 that the wave forces term oriented toward the lagoon was of the same or-
538 der of magnitude as the other terms in the momentum balance in the inlet,
539 which therefore had a significant impact on the hydrodynamics, including a
540 setup that developed within the lagoon. Recently, Fortunato et al. (2017)
541 conducted a high-resolution hindcast of the storm surge associated with the
542 1941 storm that made landfall in the North of Portugal and has driven the
543 development of a large surge in the Tagus Estuary. Their model results
544 suggested that the breaking of storm waves generated a wave setup up to
545 0.50 m in the Tagus Estuary, showing that a substantial wave setup can also
546 impact water levels at the scale of a large estuary. This phenomenon is
547 explained by the authors as the result of large onshore-directed wave forces
548 owing to storm waves breaking over the ebb delta, generating a wave setup
549 that extended beyond the surf zone and in the inlet. The previous analysis
550 of Fig. 4 and Fig. 5 suggests that such a phenomenon occurred at the Arca-
551 chon Lagoon during Klaus: large wave breaking on the ebb delta generated
552 a wave setup that affected the whole lagoon.

553 To understand the underlying mechanisms, the magnitude of the leading
554 terms of the momentum equations, *i.e.* the barotropic pressure gradient term
555 (third term of the right hand side of Eqs. (2) - (3)), the wave forces (last term
556 of the right hand side of Eqs. (2) - (3)), the bottom stress and surface stress
557 terms are computed at the inlet of the Arcachon Lagoon (Fig. 7). In order
558 to analyse the momentum balance at mid-flood and mid-ebb under similar
559 forcing corresponding to the peak of the storm, two additional simulations
560 are performed where tides are shifted.

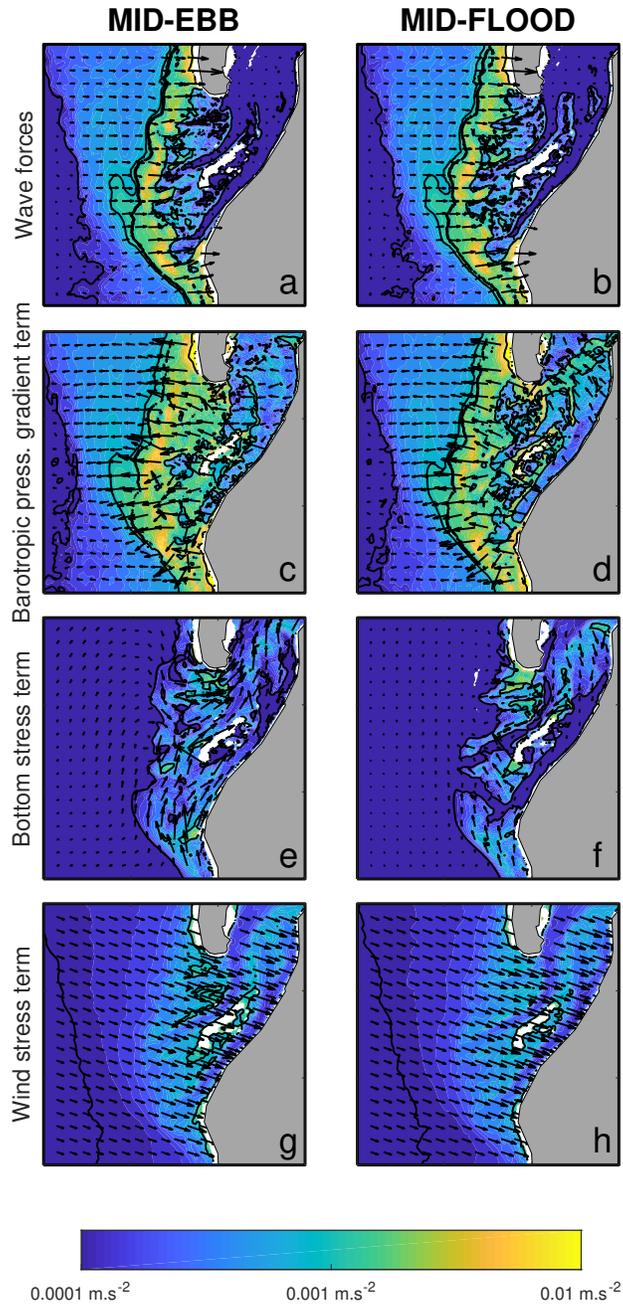


Figure 7: Leading terms of the momentum balance at mid-ebb and mid-flood: wave forces (a-b), barotropic pressure gradient term (c-d), bottom stress term (e-f) and wind stress term (g-h), at mid-ebb (left) and mid-flood (right). The magnitude and direction of each term are represented by the map colors and the vectors respectively.

561 The analysis of Fig. 7 shows that the outer part of the inlet behaves
562 like a sandy beach, with a balance between the wave forces (hereafter WF)
563 and the barotropic pressure gradient (hereafter BPG) term (Battjes and
564 Stive, 1985; Lentz and Raubenheimer, 1999). In this area, the WF reach
565 values one order of magnitude larger than the bottom stress (hereafter BS)
566 and the surface stress (hereafter SS) terms. The dominant role of WF in the
567 momentum balance at the inlet corroborates the findings of Olabarrieta et al.
568 (2011) and Dodet et al. (2013). In the inlet channel, the WF become much
569 weaker and the alongstream dynamics is controlled by a balance between
570 the BPGR and the BS terms, which is typical of tidal channels (Hench and
571 Luetlich Jr, 2003). Between the flood and the ebb, the signs of the BPGR
572 and the BS terms are inverted, except in the outer part of the inlet where
573 the BPGR term compensates the WF during all tidal phases.

574 The major contribution of the wave forces to the momentum balance
575 in the inlet directly explains the strong effect of short-wave breaking on
576 the hydrodynamics, the main impact being a wave setup that reaches several
577 tens of centimetres within the lagoon (Fig. 4). In more details, the
578 rapid decrease in WF inside the lagoon explains that wave setup displays a
579 plateau inside the lagoon (Fig. 5-f). Over the ebb delta, the wind-driven
580 surge reaches approximately 0.4 m (Fig. 5-d and assuming 0.35 m of inverse
581 barometer effect), which is comparable to the wave setup while SS are one
582 order of magnitude lower than WF. This behaviour is explained by the fact
583 that strong WF only extend over the 3 km-wide ebb delta while the wind
584 effect is integrated across the 60 km-wide shelf. Inside the lagoon, the at-
585 mospheric surge further grows as the water depth decreases (Fig. 5-d). In
586 the Adour Estuary, the weaker atmospheric surge (Fig. 5-c) is explained not
587 only by weaker winds (Fig. 2) but also by the narrower continental shelf.
588 Indeed, many studies already demonstrated that, for a given wind speed,
589 the wind-driven surge is also controlled by the shelf width, such as in the
590 Bay of Biscay (Bertin et al., 2012), in North Sea (Wolf and Flather, 2005)
591 or in the Gulf of Mexico (Kennedy et al., 2012).

592 *5.3. Tidal modulation of the wave setup*

593 Some of the studies that highlighted the development of a wave setup
594 in tidal inlets also suggested that the wave setup can be tidally-modulated
595 (Olabarrieta et al., 2011; Dodet et al., 2013; Fortunato et al., 2017). Fortu-
596 nato et al. (2017) showed that the wave setup that developed in the Tagus
597 Estuary mouth during the 1941 storm was strongly tidally-modulated with
598 values of 0.10-0.15 m at high tide while being three times larger at low tide
599 with values of 0.30-0.35 m. The authors attributed this phenomenon to

600 more intense wave breaking on the ebb delta at low tide. When waves do
601 not break over the ebb delta, they propagate into the inlet or to the coast in
602 the vicinity of the estuary mouth, and thus, their contribution to the setup
603 inside the estuary is lower. In this section, the tidal modulation of the wave
604 setup is investigated at the Arcachon Lagoon and the Adour Estuary with
605 additional numerical experiments.

606 The Arcachon lagoon exhibits large intertidal flats, which makes the
607 tidal propagation and asymmetry very sensitive to the mean water depth.
608 Therefore, tidal propagation is different when the wave setup raises the mean
609 water level of the lagoon. Computing the wave setup as the difference be-
610 tween a simulation including tides and waves and a simulation with tides
611 only results in difference not only including the wave setup but also the
612 differences in tidal levels due to the higher mean water level in the coupled
613 simulation, a process also referred to as tide surge interactions. To overcome
614 this problem, a series of stationary runs is performed with constant water
615 levels and wave forcing (Fig. 8-A). Two sub-grids of smaller extent covering
616 each studied area are forced at the ocean boundary by constant water eleva-
617 tions ranging from -1.5 m to 1.5 m, and a JONSWAP spectrum to simulate
618 short waves. The spectrum is characterized by a significant wave height of
619 14 m and a peak period of 15 s, which corresponds to the peak values reached
620 during Klaus in the region.

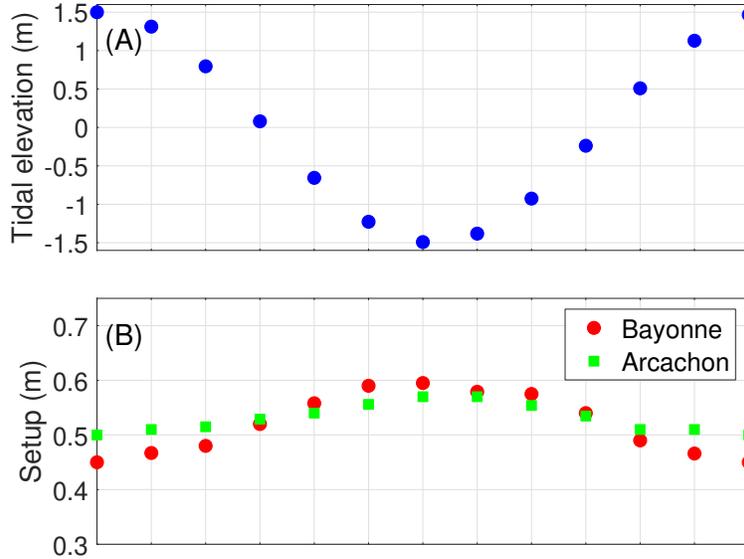


Figure 8: (A) Constant water elevations prescribed at the open boundary (blue circles). (B) Tidal modulation of the wave setup at Arcachon (green squares) and Bayonne (red circles) during a tidal cycle.

621 In the case of the Arcachon Lagoon, the results reveal a small tidal
 622 modulation of the order of 0.07 m (Fig. 8-B), the wave setup being larger
 623 at low tide. At the Adour Estuary, the tidal modulation is stronger with a
 624 wave setup reaching 0.60 m when the mean sea level is lowered by 1.5 m
 625 and decreasing to 0.45 m when the mean sea level is increased by 1.5 m.

626 Contrary to the Tagus Estuary where the ebb delta is submerged, with
 627 depths of the order of 5 m relative to the mean sea level, the ebb delta
 628 of the Arcachon Lagoon extends 3 km offshore and includes an elongated
 629 supratidal bank, the Arguin Bank. This setting causes the wave breaking
 630 to be almost full in front of the inlet, even at high tide. At lower tidal ele-
 631 vations, wave energy mostly dissipates on the terminal lobe while at higher
 632 tidal stages, waves also break over the supratidal sand bank. The wave
 633 setup exhibits therefore a slight tidal modulation, unlike the Tagus Estuary
 634 (Fortunato et al., 2017). At the Adour Estuary, the bathymetry is subtidal,
 635 which implies that the lower the water level, the larger is the wave energy
 636 dissipation and the wave setup.

637 These results indicate that tide-induced water level variations change
 638 the spatial gradients of short-wave energy dissipation rates, which in turn
 639 controls the wave setup. Depending on the morphology of the inlet, the
 640 wave setup along the shoreline and in the lagoons or estuaries can experi-

641 ence significant tidal modulations as well. Tidal currents, which are strong
642 in estuaries or tidal inlets, can also affect the propagation of short waves
643 (Ardhuin et al., 2012; Rusu et al., 2011; Dodet et al., 2013; Bertin et al.,
644 2019) and subsequently the wave setup. During flood, waves following cur-
645 rents decrease while during ebb, waves propagating against currents increase,
646 shifting the position of the breaking point seaward (Dodet et al., 2013). The
647 impact of tidal currents on short-waves propagation is verified by comparing
648 water elevations from runs including tides and waves, and activating or not
649 the feedback of currents on waves. Model results at the Arcachon Lagoon
650 and at the Adour Estuary show that switching off the effects of tidal cur-
651 rents on short-wave propagation has a small impact on wave setup (lower
652 than 0.01 m). This finding corroborates the study of Fortunato et al. (2017)
653 which reported that the tide-induced water level variations at the mouth of
654 the Tagus Estuary are the main driver for the tidal modulation of the wave
655 setup compared to tidal currents effects.

656 The comparison of the effect of tides on wave setup between both studied
657 locations emphasises that tidal modulation is site-specific. In areas such as
658 the Adour Estuary, the higher wave setup is produced close to low tide,
659 and the tidal modulation amplitude increases with increasing tidal range.
660 Such tidal modulation can therefore limit the contribution of short-wave
661 breaking to coastal flooding, which mostly occurs during high tide in macro
662 tidal regions.

663 *5.4. Sensitivity of storm surge and wave setup calculation to spatial resolu-* 664 *tion*

665 Recently, several storm surge numerical models using unstructured grids
666 have been developed (e.g. Kerr et al., 2013). Such models allow to correctly
667 represent complex shorelines and coastal embayments, using a variable grid
668 resolution, usually coarse in the deep ocean (several kilometers to tens of
669 kilometers) and down to few hundreds of meters in the nearshore. However,
670 such resolution in coastal areas may not be sufficient to adequately represent
671 small coastal morphological features, such as lagoons, and thus, enable the
672 model to provide accurate storm surge predictions (Shen et al., 2006). Also,
673 this study reveals that the wave setup generated by wave breaking during
674 extreme events can greatly contribute to the storm surge, even in areas
675 sheltered from wave breaking such as lagoons and estuaries. Accounting for
676 short waves in storm surge operational modelling is thus of key importance
677 to correctly predict water levels in coastal areas during storm events and
678 thereby, improve emergency responses. However, a good evaluation of the
679 wave setup requires a resolution fine-enough in the surf zones, which is

680 not always possible in operational modelling systems (Kohno et al., 2018).
 681 Therefore, an important question rises here: how well do surf zones need to
 682 be spatially resolved in order to correctly estimate the contribution of wave
 683 setup to the storm surge?

684 The sensitivity of the storm surge/wave setup to the model resolution
 685 is analysed at the Arcachon Lagoon region by simulating the sea state and
 686 storm surge associated to Klaus with different grid resolutions. The grid
 687 resolution used for the baseline model (hereafter BM), which goes down to
 688 35 m in the nearshore, is modified to get two additional computational grids
 689 with spatial resolution from the inner shoreface to the nearshore degraded to
 690 200 and 1000 m. The surge is evaluated at two locations along the coastline:
 691 in the inner part of the lagoon at the Eyrac tide gauge and at the shoreline
 692 exposed to the ocean, computed as the average value of the storm surge
 693 in an area defined to the south of the inlet (see Fig. 1-B). This sensitivity
 694 analysis is not carried out at the Adour Estuary since the inlet mouth, with
 695 a maximum width of 150 m, cannot be represented with such resolutions.

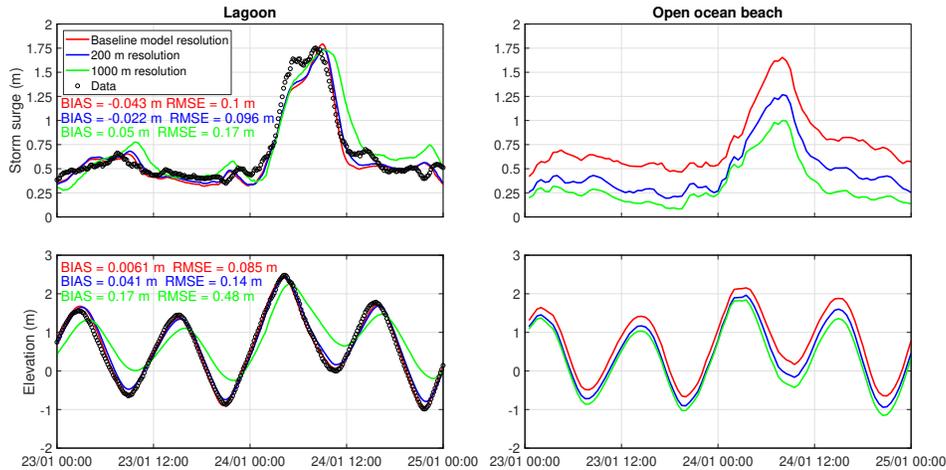


Figure 9: Simulated storm surge in the inner part of the lagoon and at the adjacent beach to the south, with the baseline model resolution, 200 m resolution and 1000 m resolution.

696 The results show that the modelled water levels and storm surge on the
 697 open ocean beach are lower when the grid resolution coarsens (Fig. 9).
 698 Indeed, while tidal predictions show little sensitivity to the grid resolution,
 699 the peak of the surge simulated with the BL resolution reaches 1.65 m while
 700 being 30 % and 65 % higher than the surges obtained with the 200 m
 701 and 1000 m resolutions respectively. A detailed analysis reveals that these

702 differences are mostly explained by wave setup, which is poorly represented
703 with a coarse grid.

704 In the lagoon, the results reveal a different behaviour of the model (Fig.
705 9). Surprisingly, the predicted storm surge is less sensitive to the grid res-
706 olution compared to the open ocean beach. The three grid resolutions well
707 reproduce the peak of the surge, with the 200 m and 1000 m grid resolu-
708 tions resulting in a slightly lower surge than the BM resolution (0.05 m).
709 However, the storm surge modelled over the total duration of the storm
710 is deteriorated with the 1000 m resolution (RMSE of 0.17 m) compared to
711 the BM and 200 m resolutions (RMSE of ~ 0.1 m). Also, water elevation is
712 poorly predicted with the 1000 m resolution, which yields a RMSE of 0.48 m,
713 against 0.085 m and 0.14 m with the BM and 200 m grid resolutions respec-
714 tively. As soon as the channels of the lagoon are not correctly represented,
715 the tidal propagation in the lagoon is poorly reproduced, which impacted
716 the predictions of water level and storm surge.

717 This sensitivity analysis of model results to grid resolution reveals a
718 contrasting situation between the inner lagoon, where wave setup is rea-
719 sonably represented even with a coarse resolution and adjacent sandy
720 beaches, where modelled wave setup is almost nil when using a coarse reso-
721 lution. This behaviour is directly explained by the cross shore extension of
722 the surf zone, which is of the order of 1000 m at adjacent beaches but range
723 from 3000 to 5000 m in front of the lagoon. As a rough guideline, we esti-
724 mate that accounting for wave setup in storm surge models requires at least
725 5 grid elements across the surf zone, which implies the use of a finer spatial
726 resolution when the beach slope increases and the wave height decreases.
727 This corroborates the findings of Nayak et al. (2012), who investigated the
728 sensitivity of wave setup predictions to grid resolution considering idealized
729 beaches of slope ranging from 1:80 to 1:10.

730 6. Conclusion

731 The fully-coupled modelling system SCHISM using a vortex force for-
732 malism was used to hindcast the sea state and storm surge associated to the
733 strongest storm that occurred in the southern part of the Bay of Biscay for
734 the last 20 years. After the verification of the model with wave and water
735 level observations available during the storm, the analyses of the simulations
736 revealed that the predictions of the storm surges at the Arcachon Lagoon
737 and the Adour Estuary were improved by 50 to 60 % when the wave forces
738 were accounted for. The wave setup induced by the storm waves break-
739 ing in the vicinity of these two inlets extended outside the surf zones and

740 significantly increased the water level at the scale of the whole lagoon and
741 estuary.

742 To understand the impact of storm wave breaking on the hydrodynamics
743 of the tidal inlets, the local momentum balance was analysed at the inlet
744 of the Arcachon Lagoon. By reaching values one order of magnitude larger
745 than the bottom stress and the surface stress terms, the wave forces were one
746 of the leading terms of the momentum balance and thereby greatly affected
747 hydrodynamics in the inlet, the main impact being the development of a
748 wave setup at the scale of the whole lagoon.

749 Further analysis showed that the wave setup in tidal inlets can be tidally-
750 modulated while this phenomenon is site-specific and depends on the mor-
751 phology of the inlet. At Arcachon, as the ebb delta is characterised by
752 supra-tidal sand banks, wave breaking is total at all tidal phases, the wave
753 setup exhibits therefore a slight tidal modulation. At Bayonne, waves are
754 subjected to more intense breaking at low tide than at high tide, the tidal
755 modulation of the wave setup is thus more pronounced.

756 Finally, a sensitivity analysis of the storm surge and wave setup to the
757 spatial resolution of the computational grid was carried out. This work
758 revealed that the calculated wave setup at the shoreline is highly sensitive to
759 the grid resolution. In the lagoon, the modelled storm surge and wave setup
760 were found to be comparable between different grid resolutions, while tidal
761 propagation cannot be accurately represented with a resolution of 1000 m.
762 This study highlighted the need to account for wave breaking in operational
763 storm surge models, although resolving the wave setup requires a spatial
764 resolution that depends on the width of the surf zone, itself controlled by
765 the bottom slope and the wave height.

766 In a context of upcoming altimetry satellite missions with spatial foot-
767 prints below 1000 m (SWOT, Durand et al., 2010), the results presented in
768 this study are of key importance as they show that the wave setup can im-
769 pact the water level in sheltered areas such as harbours, large lagoons and
770 estuaries. As these coastal areas are usually instrumented with tide gauges
771 that are used to calibrate altimeter measurement systems, it is crucial to
772 determine the physical drivers of the water level variations recorded at these
773 stations.

774 **Appendix A.**

775 The underestimation of the storm surge before the peak of the surge can
776 be attributed to a negative bias in the 10 m wind speed of CFSR. In order

777 to verify this hypothesis, the modelled wind speed from CFSR is increased
778 by 12-15 % over three time steps before the storm peak (Fig. A.10-A).

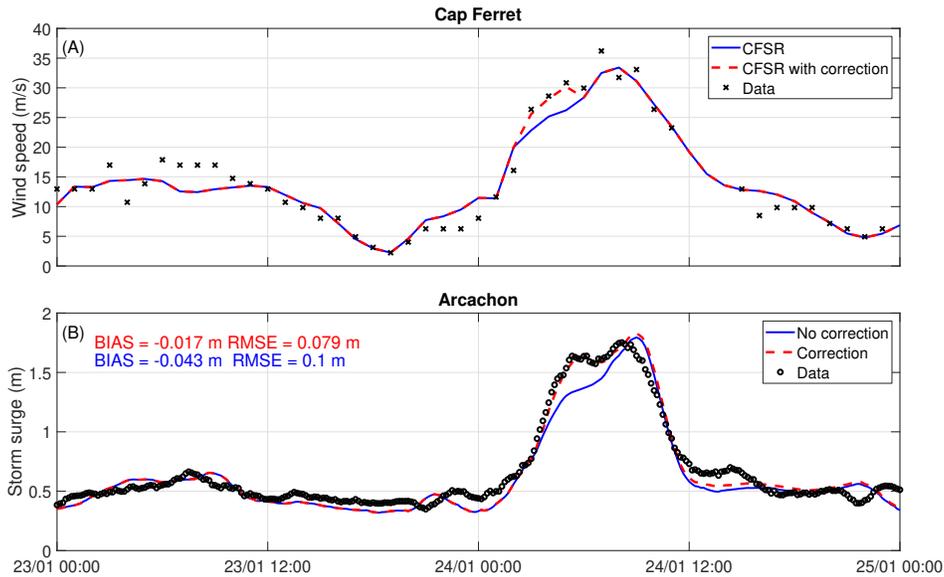


Figure A.10: (A) Measured (black crosses) against CFSR wind speed with correction (red dashed line) and original data (blue line). (B) Observed (black dots) against modelled storm surges with the corrected wind speed (red dashed line) and the original wind speed (blue line).

778

779 The comparison between the original modelled storm surge and the storm
780 surge computed with the tuned wind speed (Fig. A.10-B) shows a signifi-
781 cant difference at the considered period. The RMSE is improved by 20 %
782 and the localised error is cancelled out. These results confirm that the un-
783 derestimation of the storm surge at this stage of the storm is due to a local
784 negative bias in the modelled wind speed of the order of 4 m.s^{-1} .

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