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- ¹ Interactions Between Mean Sea Level, Tide, Surge,
- ² Waves and Flooding: Mechanisms and Contributions
- ³ to Sea Level Variations at the Coast
- 4 Déborah Idier · Xavier Bertin · Philip
- 5 Thompson · Mark D. Pickering

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Abstract Coastal areas epitomize the notion of 'at risk' territory in the context 8 of climate change and sea-level rise (SLR). Knowledge of the water level changes at the coast resulting from the mean sea level variability, tide, atmospheric surge 10 and wave setup is critical for coastal flooding assessment. This study investigates 11 how coastal water level can be altered by interactions between SLR, tides, storm 12 surges, waves and flooding. The main mechanisms of interaction are identified, 13 mainly by analyzing the shallow-water equations. Based on a literature review, 14 the orders of magnitude of these interactions are estimated in different environ-15 ments. The investigated interactions exhibit a strong spatio-temporal variability. 16 Depending on the type of environments (e.g. morphology, hydro-meteorological 17 context), they can reach several tens of centimeters (positive or negative). As a 18 19 consequence, probabilistic projections of future coastal water levels and flooding should identify whether interaction processes are of leading order, and, where ap-20 propriate, projections should account for these interactions through modeling or 21

²² statistical methods.

Keywords Water level · Hydrodynamics · Interaction processes · Implications ·
 Flood · Quantification · Modeling

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25 1 Introduction

²⁶ Coastal areas are considered 'at-risk' territories in the context of climate change
²⁷ and sea-level rise (SLR). Knowledge of water level variability at the coast, espe²⁸ cially the still water level and storm tide, is critical for coastal flooding assessment,
²⁹ both for present and future climate. Still water level includes mean sea level, tide
³⁰ and atmospheric surge (Figure 1a). Several definitions can be found for storm
³¹ tides; here, we consider that they include mean sea level, tide, atmospheric storm
³² surge and wave setup.

According to the IPCC 5th Assessment Report, the likely range of future global mean sea level (GMSL) for the high emissions scenarios is +0.5 to +1 m by 2100 (Church et al. 2013b), which does not preclude more extreme scenarios (Church et al. 2013a). Importantly, sea level will continue to rise beyond 2100 (Church et al. 2013b) and is likely to reach several meters by 2200 (Kopp et al. 2014). Such changes in mean sea level will produce significant societal impact.

A widely used approach to account for mean sea level changes in flood hazard 39 assessment is to linearly add the sea-level rise to selected still water level scenarios 40 to estimate the flood hazard. Such an approach relies on the underlying assumption 41 that there is no significant nonlinear interaction between contributions to water 42 level variability, such as SLR, tide and surge. However, some studies show evidence 43 that these interactions can represent a significant part of water level changes. For 44 instance, using a numerical modeling approach, focusing on the German Bight 45 area (SE of North Sea) and assuming a sea-level rise of 0.54 m, Arns et al. (2015) 46 show that taking into account the interactions between mean sea level, tide and 47 atmospheric surge leads to a 50-year return still water level 12 cm larger than when 48 these interactions are neglected, corresponding to a doubling of the frequency of 49 50-year return still water level obtained neglecting the interactions. 50

The present review focuses on water level resulting from mean sea level, tide 51 and surge (atmospheric surge and wave setup) and investigates how this water level 52 can be altered by interaction processes occurring between SLR, tides, storm surges, 53 waves and flooding. Indeed, many mechanisms can affect the still water level (e.g. 54 changes in sea-bed morphology, oceanographic circulations, tide-surge interactions, 55 ...). Figure 1b schematizes some of these interactions. The main interactions to be 56 investigated in this review are: (1) the SLR effect on tides, atmospheric surges and 57 waves, (2) the tide effect on atmospheric storm surges, waves and wave setup, (3)58 the flooding effect on tide and still water level (including human adaptation on 59 tides), (4) the effect of short waves on atmospheric surges. 60

The present paper aims at highlighting these interaction mechanisms and at 61 providing the orders of magnitude of the interactions contributions to the water 62 level at the coast. One of the difficulties is to isolate the influence of each inter-63 action. Observations (e.g. from tide gauges) can provide insights into changes in 64 mean sea level, tides, surge, or even sometimes the wave setup, but, as highlighted 65 by Woodworth (2010) and Haigh et al. (submitted), many processes can affect 66 tide changes (SLR, harbor infrastructures, long-term changes in the tidal poten-67 tial, changes in internal tide, morphological changes, ...), such that it is difficult 68 to properly isolate the influence of each parameter based on these observations. 69 Thus, the present review is mainly based on modeling studies. 70

The paper is organized as follows. First, the main mechanisms leading to changes in mean sea level, tidal amplitude, atmospheric surge and wave setup ⁷³ are reviewed together with their orders of magnitude (Section 2). Then, each of

⁷⁴ the interaction mechanisms is described and orders of magnitude are provided in

⁷⁵ different environments (Section 3). Section 4 provides a synthesis and examples

⁷⁶ of combined interactions, before discussing the limits of the review and the im-⁷⁷ plications for projections of water level at the coast and flood hazard estimation.

⁷⁸ Remaining questions are also highlighted. Section 5 draws the main conclusions.

⁷⁹ 2 Mean sea level, tide, atmospheric surge and wave setup: mechanisms ⁸⁰ and orders of magnitude

What follows is a summary of the main mechanisms or factors leading to changes in mean sea level and leading to tide, atmospheric storm surge and wave setup. Here, mean sea level can be considered as a zeroth order component (low frequency), while tide, atmospheric surge and wave setup (higher frequency) are considered as first order components.

As highlighted by Woodworth et al. (2019), mean sea level can be affected by 86 many factors. In addition to long-term trends related to climate change, mean 87 sea level is subject to seasonal variability due to changes in thermal expansion 88 and salinity variations (steric effect), air pressure and winds, land and sea ice 89 melt, oceanographic circulation, river runoff, etc. Some interannual and decadal 90 variability can also be observed as a result of the effect of climate modes like El 91 Niño Southern Oscillation (ENSO), North Atlantic Oscillation (NAO) or Pacific 92 Decadal Oscillation (POD) resulting in large-scale sea level change. Fifteen thou-93 sand years ago, GMSL was more than 100 m below present (Clark et al. 2016). 94 According to Church et al. (2013a), GMSL will increase by several tens centimeters 95 and could exceed 1 m in 2100. 96

Tides are generated by gravitational forces acting over the whole water col-97 umn in the deep ocean, with a gravitational feedback in tidal dynamics known as 98 self-attraction and loading (SAL, Hendershott (1972)). In deep water, tides have 99 wavelengths of several hundreds of kilometers, i.e. much longer than the water 100 depth. They propagate as shallow-water waves, influenced by the Earth rotation 101 (Coriolis force), and are dissipated by bottom friction in shallow water on conti-102 nental shelves and by energy loss to internal tides (Ray 2001). Local or basin-scale 103 enhancements can occur due to resonance producing very large tides (Godin 1993). 104 At the global scale, depending on the location, spring tidal ranges vary from a few 105 tens of centimeters to several meters, and can locally exceed 10 meters as in the 106 Bay of Fundy (Pugh 1987). The atmospheric storm surges can also be regarded 107 as long waves. They are generated by changes in atmospheric pressure and wind 108 stress acting on the sea surface. Similar to tides, storm surges propagate to the 109 coast as shallow-water waves, and are subject to sea-bed friction, Earth's rotation 110 (Coriolis force), and local enhancements due to resonance producing large surges. 111 Atmospheric surge of tens of centimeters and up to 1 m are frequently observed 112 on the Northwest European shelf (e.g. Brown et al. 2010; Idier et al. 2012; Breilh 113 et al. 2014; Pedreros et al. 2018) and can exceed 2–3 m under specific stormy 114 conditions, such as in the North Sea in 1953 (Wolf and Flather 2005). It should 115 be noted that these values correspond to the so-called practical storm surge, i.e. 116 the difference between the still water level and the tide. In cyclonic environments, 117

¹¹⁸ pure atmospheric surge (i.e. without accounting for tide-surge interaction) can ¹¹⁹ reach almost 10 m (e.g. Nott et al. 2014).

In a first approximation, tides and surge can be modeled by the shallow-water equations, which can be written as follows, omitting the horizontal viscosity term $(A\nabla^2 u)$ for the sake of clarity:

$$\frac{\partial\xi}{\partial t} + \nabla.\left(D.\mathbf{u}\right) = 0\tag{1}$$

 $\frac{\partial \mathbf{u}}{\partial t} + \mathbf{u} \cdot \nabla \mathbf{u} - f \mathbf{k} \cdot \mathbf{u} = -g \nabla \xi + \frac{1}{\rho} \nabla p_a + \frac{\tau_s}{\rho D} - \frac{\tau_b}{\rho D} + \mathbf{F} + \mathbf{\Pi}$ (2)

with **u** the depth-integrated current velocity, ξ the free surface, D the total water depth (equal to the sum of undisturbed water depth H and the free surface elevation ξ), ρ the density of sea water, g the gravitational acceleration, p_a the sealevel atmospheric pressure, f the Coriolis parameter ($2\omega \sin \phi$, with ω the angular speed of Earth rotation and ϕ the latitude) and **k** a unit vector in the vertical. τ_b and τ_s are respectively the bed and wind shear stress, which can be written as follow (assuming a quadratic law):

τ

$$\tau_{\mathbf{s}} = \rho_a C_{D_s} \mathbf{U}_{\mathbf{10}} |\mathbf{U}_{\mathbf{10}}| \tag{3}$$

$$\mathbf{b} = C_{D_b} \mathbf{u} |\mathbf{u}| \tag{4}$$

where C_{D_s} and C_{D_b} are the free surface and bottom drag coefficients, respec-132 tively, ρ_a is the air density and U₁₀ is the wind velocity at z = 10 m. Finally, F 133 includes other forces such as the wave-induced forces leading to wave setup, and Π 134 includes tide-related forcing terms (e.g. self attracting load, tidal potential forces). 135 In the case of pure tides, the second, third and fifth terms of the right hand side 136 of equation (2) are equal to zero. In the case of pure atmospheric storm surge, the 137 last term is equal to zero. From the equations, it can be readily seen that the effect 138 of wind depends on water depth and increases as the depth decreases, whereas the 139 atmospheric pressure effect is depth independent. In deep water, surge elevations 140 are therefore approximately hydrostatic, while surge production by the wind stress 141 can be large on shallow continental shelves. 142

In the nearshore, the dissipation of short-waves through depth-limited breaking 143 (with a small contribution from bottom friction) results in a force that drives cur-144 rents and wave setup along the coast, which can substantially contribute to storm 145 surges. Under energetic wave conditions, wave setup can even dominate storm 146 surges along coasts bordered by narrow to moderately wide shelves or at volcanic 147 islands (Kennedy et al. 2012; Pedreros et al. 2018). Several studies combining field 148 observations with numerical modeling also demonstrated that wave breaking over 149 the ebb shoals of shallow inlets (Malhadas et al. 2009; Dodet et al. 2013) as well 150 as large estuaries (Bertin et al. 2015; Fortunato et al. 2017) results in a setup that 151 can propagate at the scale of the whole backbarrier lagoon or estuary. Local wave 152 setup of several tens of centimeters up to about 1 m have been observed (Pedreros 153 et al. 2018; Guérin et al. 2018), while regional wave setup can reach values of tens 154 of centimeters (Bertin et al. 2015). This does not imply that larger values could 155 not exist. The first theoretical explanation for the development of wave setup is 156 due to Longuet-Higgins and Stewart (1964), who proposed that the divergence of 157 the short-wave momentum flux associated with wave breaking acts as a horizontal 158 pressure force that tilts the water level until an equilibrium is reached with the 159

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160 subsequent barotropic pressure gradient. However, several studies reported that 161 this model could result in a severe underestimation of wave setup along the coast

(Raubenheimer et al. 2001; Apotsos et al. 2007), suggesting that other processes
may be involved. Apotsos et al. (2007) and Bennis et al. (2014) proposed that

bed-shear stress associated with the undertow (bed return flow that develops in

¹⁶⁵ surfzones) could contribute to wave setup. Recently, Guérin et al. (2018) showed

that the depth-varying currents that take place in the surfzone (horizontal and

¹⁶⁷ vertical advection and shear of the currents) could contribute to wave setup sub-¹⁶⁸ stantially, particularly when the bottom is steep.

A broader overview of all forcing factors causing sea level changes at the coast and of their orders of magnitude, from sub-daily (seiche, infra-gravity waves) to long-term scales (centuries), is provided by Woodworth et al. (2019), while Dodet

¹⁷² et al. (2019) provide a review on wind-generated waves (processes, methods) and

¹⁷³ their contributions to coastal sea level changes.

¹⁷⁴ 3 Interaction mechanisms and orders of magnitude

¹⁷⁵ 3.1 Sea-level rise effect on tide and atmospheric surge

176 3.1.1 Mechanisms

Tides behave as shallow-water waves, and thus are strongly affected by water 177 depth. There are several mechanisms by which mean sea level changes can alter 178 tidal dynamics (Wilmes 2016; Haigh et al. submitted). Here, we focus on the direct 179 effect of SLR on tides. First, large tidal amplitudes and dissipation occur when 180 the tidal forcing frequency lies close to the natural period of an ocean basin or sea 181 (Hendershott 1973). Therefore, increases in water depth due to MSL rise, could 182 push a shelf sea or embayment closer to resonance, increasing tidal range (e.g. in 183 the Tagus estuary after the study of Guerreiro et al. (2015)), or away from reso-184 nance, reducing tidal range (e.g. in the Western English Channel according to Idier 185 et al. (2017)). The sensitivity to such water depth change increases as the basin 186 approaches resonance. Second, the greater water depth implies a reduction in the 187 energy dissipation at the bottom (see Equations 2 and 4) and thus contributes to 188 an increase in tidal range (see e.g. the study of Green (2010) for numerical experi-189 ments in real cases). Third, increase in water depth alters the propagation speed of 190 the tidal wave (and thus causes a spatial re-organization of the amphidrome). For 191 instance, in a semi-enclosed basin and neglecting the dissipation terms, sea-level 192 rise causes the amphidromic point to shift towards the open boundary after the 193 analytical solution of Taylor (1922). 194

In case of SLR, as for the tides, storm surges are affected by the bottom 195 friction reduction which tends to increase storm surges. However, as illustrated 196 by Arns et al. (2017) in the North Sea, the decreased bottom friction appears to 197 be counteracted by the lessened effectiveness of surface wind stress. Indeed, the 198 same wind forcing (surface stress) is less effective at dragging water and produces 199 a smaller surge when water depth is larger (see the wind forcing term in Equation 200 2). Finally, as a counterbalancing effect, SLR will also result, in some locations, 201 in new flooded areas, which will act as additional dissipative areas for tide and 202 surge. In the present section, for sake of clarity, we do not consider the effect of 203

these additional wet areas. The effect of flooding on tides and on still water level is discussed in sections 3.3 and 4.2, respectively.

206 3.1.2 Orders of magnitude

The SLR effect on tides has been investigated at different scales (global, regional, 207 local), and in different environments. The locations of the main studies presented 208 below are indicated in Figure 2 (in black). The values given in the next paragraphs 209 correspond to changes in the tidal component alone (i.e. relative to the mean sea 210 level) and not to the absolute tide level (which includes the mean sea level and thus 211 212 SLR itself). In addition, the orders of magnitude provided below are extracted from model results obtained assuming a fixed shoreline (i.e. impermeable walls along 213 the present day shoreline). 214

First, at the global scale, tide response (M2 amplitude or mean high water level 215 for instance) to SLR is widespread globally with spatially coherent non-uniform 216 amplitude changes of both signs in many shelf seas, even considering uniform SLR 217 (Green 2010: Pickering et al. 2017: Schindelegger et al. 2018). Response in the open 218 ocean, where the relative depth change with SLR is small, is generally of a smaller 219 magnitude but with a much greater horizontal length scale. Pickering et al. (2017), 220 focusing on 136 cities of population larger than 1 million (in 2005), assuming fixed 221 shorelines and a uniform SLR of 2 m, found changes in the mean high waters 222 (MHW) varying from -0.25 (Surabaya, Indonesia) to +0.33 m (Rangoon, Myan-223 mar) (Figure 3b). The comparison of bathymetry (here, GEBCO, Figure 2) with 224 the SLR-induced tide changes map of Pickering et al. (2017) (Figure 2a therein, 225 or Figure 3b in the present paper) illustrates that along open coasts with narrow 226 continental shelf, the SLR effect on tide appears negligible. In that study, they also 227 investigated the effect of a spatially varying SLR, focusing on fingerprints of the 228 initial elastic response to ice mass loss. These SLR perturbations weakly alter the 229 tidal response with the largest differences being found at high latitudes (see the 230 cumulative distribution of MHW changes for the 136 cities and for the assumption 231 of a fixed shoreline, Figure 3a). 232

Focusing on the NW European Shelf (spring tidal range varying from a few 233 centimeters to more than 10 m mainly in the Bay of Mont Saint Michel), Idier 234 et al. (2017), Pickering et al. (2012) and Pelling et al. (2013a) show that for 235 SLR=+2 m and under a 'no flood' assumption (also called fixed shoreline), the 236 M2 amplitude changes up to $\pm 10-15\%$ of SLR. In terms of high tide changes, 237 Idier et al. (2017) show that depending on the location, changes in the highest 238 tide of the year (in that study: 2009) range from -15% to +15% of SLR, i.e. 239 several centimeters to about 15 cm for SLR=1 m. They also show that, when it is 240 assumed that land areas are protected from flooding, the tide components and the 241 maximum tidal water levels vary proportionally to SLR over most of the domain, 242 up to at least SLR=+2 m. Some areas show non-proportional behavior (e.g. the 243 Celtic Sea and the German Bight). Consistently with the studies of Pelling et al. 244 (2013a) and Pickering et al. (2017), the high tide level decreases in the western 245 English Channel and increases in the Irish Sea, the southern part of the North 246 Sea and the German Bight. The overall agreement between the different modeling 247 experiments is fairly remarkable. Even using different tidal boundary conditions, 248 different spatial resolutions, different models (even if all based on the shallow-water 249 equations), Idier et al. (2017), Pickering et al. (2012) and Pelling et al. (2013a) 250

²⁵¹ provide similar results in terms of M2 changes (see Figures 3, 4 and 6d in these ²⁵² studies, respectively). These studies agree on a decrease in the western English ²⁵³ Channel and in the SW North Sea, and an increase in the eastern English Channel, ²⁵⁴ the central part of the North Sea, the German Bight and the Irish Sea. The main ²⁵⁵ discrepancy is the positive trend along the Danish coast given in (Pickering et al. ²⁵⁶ 2012), probably due to the closed model boundary to the Baltic Sea in that study.

As an additional comparison, we can also refer to the study of Palmer et al. (2018),
which shows strikingly similar spatial patterns of increase and decrease to the one

of Pickering et al. (2012), except for the region that spreads out from the Bristol
Channel. For a more detailed comparison on the modeling studies of the SLR effect
on tides over the European shelf, see Idier et al. (2017).

Other regions have been investigated, such as the Taiwan Strait, which is a 262 long (> 300 km) and wide (\sim 140 km) shelf channel characterized by a mean tidal 263 range varying from less than 1 m in the southeast end to more than 4 m in its 264 northwest end. After Kuang et al. (2017), a SLR of 2 m induces an increase of 1 to 265 4 cm for M2 tidal component amplitude, and 1 cm for K1 amplitude. For this area, 266 the global study of Pickering et al. (2017) indicates even larger changes, with an 267 increase of 17 cm of M2 amplitude at Xiamen (city located on the Chinese coast) 268 and an increase of the maximal tidal range of 36 cm (for the same SLR of 2 m, 269 assuming fixed present-day coastline). 270

The SLR effect on tides has also been investigated in the Bohai Sea (China, north of the Taiwan Strait, Figure 2). For a 'no flood' case and SLR=2 m, Pelling et al. (2013b) estimated M2 amplitude changes ranging from about -0.1 to +0.1 m. In addition, the effect of the sea-level rise on M2 amplitude in this area is not proportional to the SLR amount.

In the Southern Hemisphere, along the coast of Australia, Harker et al. (2019) 276 found M2 amplitude changes ranging from -0.1 to +0.1 m for a SLR of 1 m with 277 amplitude changes that are not proportional to the SLR in this area, even if the 278 patterns appear similar (in terms of areas of increase or decrease of M2 amplitude). 279 The effect of SLR on tides in the San Francisco Bay has also been investigated. 280 This bay is characterized by an inlet (width of about 2.5 km) and two main bays 281 (one to South, one to the North) of tens kilometers length and widths varying 282 from few kilometers up to almost 20 km. The mean tidal range varies from 1.8 to 283 about 2.7 m, after the tide data provided by NOAA. With scenarios of hardened 284 shoreline and a SLR of 1 m, the high tide exhibits an increase of 6 cm and 5 cm 285 in the southern and northern bay, respectively, i.e. more than 10% of the natural 286 tide amplification in these bays (Holleman and Stacey 2014). 287

Regarding the U.S. East coast, the M2 amplitudes range from about 0.4 to 1.5 m (values extracted from the FES2014 tidal components database ; Carrere et al. 2015). Ross et al. (2017) show that for SLR=1 m, more than half of the Delaware Bay is projected to experience an M2 amplitude increase of at least 15 cm, while the Chesapeake Bay exhibits a small decrease at the mouth (-2 cm), and an increase over most of the bay up to 10 cm at the head. It is also highlighted that changes are proportional to the SLR (especially for SLR=[-1;+1]m).

Along the Patagonian Shelf (spring tidal range between 0 and 3 m), Carless et al. (2016) show that for SLR=1 m, M2 amplitude changes range between -0.1 and +0.1 m ('no flood' scenario).

All these studies converge to highlight that the effect of metric SLR can lead to tide changes (M2 component or spring high tide) up to $\pm 10-15\%$ of SLR. We could not identify any studies focusing on the sole effect of SLR on atmospheric
 surge. However, several studies investigated the effect of SLR on still water level or

practical storm surge. These studies implicitly take into account not only the effect

³⁰³ of SLR on atmospheric surge but also on tides and the tide-surge interactions. The

³⁰⁴ changes induced by these combined interactions are discussed in Section 4.2.

305 3.2 Tide effect on surge

 $_{\tt 306}$ $\,$ The present subsection focuses on the effect of tide on surge. However, it should be

 $_{\rm 307}$ $\,$ kept in mind that this is a matter of point of view, and that strictly speaking the

mechanisms can be referred to as tide-surge interactions. In addition, we focus on

the practical storm surge (also called residual), which is the difference between the

still water level and the tide level, such that it includes both the pure atmospheric

 $_{\rm 311}$ $\,$ storm surge and the effect of the tide on the surge.

312 3.2.1 Mechanisms

From Equations 1, 2, 3, 4, it can be readily seen that there are several nonlinear terms of tide-surge interaction which can be classified in three nonlinear effects:

- the advective effect arising from the advective terms of the momentum equation (2).

- the shallow-water effect, which arises from nonlinearity related to $D = (H + \xi)$

in equations (1), (2) and (3) in the following terms: advective term of the continuity equation, division by the depth D for the bed-friction and wind forcing.

- the nonlinear effect of the bottom friction term with the quadratic parametrisation in equation (4).

Thus, tidal current and tidal water level interact directly with the hydrodynam-323 ics induced by wind and pressure through the advection term, the shallow-water 324 effect and the nonlinear friction term related to velocity interactions (Flather 2001; 325 Zhang et al. 2010). The advective term implies that everything else being equal, the 326 tide-surge interaction is larger for larger tidal currents, such that areas of strong 327 tidal currents are potential areas of strong tide-surge interactions. Regarding the 328 shallow-water effect, it contributes to the modulated surge production. Indeed, 329 under some assumptions (mainly 1-D flow and constant wind field) Pugh (1987) 330 shows that the sea surface slope $(\partial \xi / \partial x)$ is in equilibrium with $C_{D_c} U_{10}^2 / (qD)$, 331 such that the wind stress produces more surge in shallow water, leading to more 332 surge at low tide than at high tide with other fators being equal. As explained 333 by Horsburgh and Wilson (2007), such phenomena can lead to an increase in the 334 phase lag of the practical storm surge compared to the tide, such that the storm 335 surge can precede high water by more than four hours. Regarding the nonlin-336 ear friction term, it appears to be the dominant term in tide-surge interaction in 337 shallow-water areas of strong tidal currents (Zhang et al. 2010; Idier et al. 2012). 338 Wolf (1978) investigated the contribution of this nonlinear friction term to the 339 tide-surge interaction. In that study, the definition was slightly different with the 340 quadratic friction term including the water depth variations, i.e. one part of the 341 shallow-water effect. They solved analytically the motion equation of two plane 342

progressive waves traveling together in a semi-infinite uniform channel and show 343 that the increase of the interaction on rising tide is due to shallow water and ad-344 vection effect, whereas the quadratic friction effect tends to reduce it at high tide. 345 Besides, from equation (2), the increased force of bed stress due to the alignment 346 of tidal and storm-induced current is offset by the pressure gradient force (surface 347 slope) and then the surge residual, such that tide-surge interaction is intensified 348 in cases of strong alignment of tide and storm surges. From the tide perspective, 349 it should be noted that a positive storm surge (leading to a larger water depth) 350 induces a faster propagating tide (see e.g. Flather 2001). 351

352 3.2.2 Orders of magnitude

The locations of the main studies presented below are indicated in Figure 2 (in 353 red). At the scale of the NW European shelf, Horsburgh and Wilson (2007) show 354 that for the storm of 29-30/01/2000, tide modified the instantaneous surge of sev-355 eral tens of centimeters (exceeding 50 cm of changes at high tide for instance in 356 the German Bight). Their analysis of results from selected locations along the East 357 coast of the United Kingdom show changes in the surge peak of tens of centime-358 ters. Focusing on the English Channel, Idier et al. (2012) made surge computations 359 with and without tide using a shallow-water model (Figure 4). For the two selected 360 events (the November 2007 North Sea and March 2008 Atlantic storms), the in-361 stantaneous tide-surge interaction is seen to be non-negligible in the eastern half 362 of the English Channel, reaching values of 74 cm (i.e. 50% of the same event's 363 maximal storm surge) in the Dover Strait. Using the same hydrodynamic model, 364 simple computations are performed with the same meteorological forcing while 365 varying the tidal amplitude. Skew surges (defined as the difference between the 366 maximum still water level and the maximum predicted tidal level regardless of 367 their timing during the tidal cycle) appear to be tide-dependent, with negligible 368 values (<0.05 m) over a large portion of the English Channel, but reaching several 369 tens of centimeters in some locations (e.g. the Isle of Wight and Dover Strait). 370

The Bay of Bengal is another region where tide-surge interactions are known 371 to be very relevant (Johns et al. 1985). Krien et al. (2017b) performed numerical 372 experiments during cyclone Sidr (2007), focusing on the head of the bay. They 373 investigated the interaction between the tide and the total storm surge (including 374 wave setup). As the computed wave setup ranged from 0.2 to 0.3 m and varies 375 little over a tidal cycle, the tide-surge interactions analyzed by these authors mostly 376 corresponds to the effect of the tide on atmospheric surge and vice versa. They 377 showed that tide-surge interactions in the range ± 0.6 m develop in shallow areas of 378 this large deltaic zone. In addition, such interactions occurred at a maximum of 1-2379 hours after low tide due to the combination of a stronger wind contribution during 380 periods of shallow depth and a faster propagating tide compared to a situation 381 without surge. These findings corroborate those of Johns et al. (1985), Antony and 382 Unnikrishnan (2013) and Hussain and Tajima (2017). 383

The Taiwan Strait is one example where the pattern of strong tidal currents and storm-induced currents along the channel direction enhances tide-surge interaction via nonlinear bottom friction (Zhang et al. 2010). This strait is subject to large storm surges frequently occurring during the typhoon season; from 1949 to 1990 there were 69 typhoons inducing storm surges over 1 m along at the western bank of the Taiwan Strait (Fujian coast), including four with storm surge larger than 2 m. Oscillations of about 0.4 m have been observed at tide gauges along the northern

³⁹¹ Fujian coast, the west bank of the Taiwan Strait, during Typhoon Dan (1999) (for

 $_{392}$ this event, surge ranged from 0.6 m to more than 1 m at the tide gauges). The

³⁹³ numerical experiments of Zhang et al. (2010) show that these oscillations are due ³⁹⁴ to tide-surge interaction.

At the East of the Taiwan strait, tide-surge interactions have been investigated along the coast of the Leizhou Peninsula (LP) by Zhang et al. (2017) (Figure 2). This area is characterized by extensive mudflats, large tidal ranges and a complex coastline. The largest amplitudes of tide-surge interaction are found in the shallow-water region of the Leizhou Bay, with values up to 1 m during typhoon events. Numerical experiments reveal that nonlinear bottom friction is the main contributor to tide-surge interaction in this area.

Further east, the effect of tide on surge has been investigated in the Bohai Sea and the East China Sea. Xu et al. (2016) selected one typhoon, assumed that this typhoon arrives at 12 different times (the other conditions remaining constant) and analysed the results in four tidal stations. The modeled storm surge elevations exhibit wide variations across the twelve cases, reaching differences up to 58 cm (at Yingkou tidal station).

Tide-surge interactions have also been investigated on the Patagonian Shelf. Etala (2009) found differences of tens of centimeters at the head of the bay and mouth of Rio de la Plata (Brazil).

The above studies converge to highlight that tide-surge interactions can produce tens of centimeters of water level at the coast with up to 1 m contributions in some cases.

414 3.3 Flooding effect on tide

415 3.3.1 Mechanisms

In sections 3.1 and 3.2, we considered that the shoreline was fixed. In other words 416 we assume that the coastal defenses (natural or man-made) are high enough to 417 protect the lands from the flood. Removing this assumption can lead locally to 418 more space for water, especially adding very shallow areas, i.e. areas of additional 419 energy dissipation. Such effect can balance for instance the pure effect of SLR 420 on tides. At the global scale, Pickering et al. (2017) investigated the impact of 421 flood defenses and SLR on the tidal regime: SLR scenarios allowing for coastal 422 recession (i.e. allowing for flood) tend increasingly to result in a reduction in tidal 423 range. At this global scale, according to Pickering et al. (2017), the fact that the 424 fixed and recession shoreline scenarios result mainly in changes of opposing sign 425 is explained by the effect of the perturbations on the natural period of oscillation 426 in the basin. At a regional scale, the effect of allowing dry land to flood is more 427 complex. For instance, Pelling et al. (2013a) show that the North Sea is dominated 428 by the flooding of the Dutch coast which shifts the areas of tidal energy dissipation 429 from the present coastline to the new cells and thus moves the amphidromic points 430

431 towards the coast.

432 3.3.2 Orders of magnitude

The locations of the main studies presented below are indicated in green in Figure 2. As in section 3.1, the values given in the next paragraphs correspond to changes in the sole tidal component (i.e. relative to the mean sea level).

At the global scale, according to Pickering et al. (2017), assuming a receding 436 shoreline except around Antarctica and a SLR of 2 m tends to result in a reduc-437 tion of the tidal range, with more cities exhibiting mean high water reduction. 438 For instance, Rotterdam (Netherlands) experiences a change of -0.69 m in MHW 439 compared to the present day MHW of 1.31 m. Changes occur at the coast but 440 also in the open ocean, highlighting that the flood effect is not only local. In that 441 study, tide changes appear more sensitive to the shoreline evolution (recession or 442 fixed), than to the non-uniformity of SLR (Figure 3a). Comparing the M2 ampli-443 tude changes of Schindelegger et al. (2018) with the ones of Pickering et al. (2017), 444 M2 appears less sensitive to flooding in the former study. However, it should be 445 kept in mind that these global modeling studies are run at a scale of $1/12^{\circ}$ and 446 $1/8^{\circ}$ respectively. 447

On the NW European shelf, comparing 'flood' and 'no flood' scenarios, Pelling 448 et al. (2013a) found M2 amplitude changes up to more than 10 cm (for SLR=2 m), 449 especially along the German and Danish coasts. From a quantitative point of view, 450 the numerical experiment of Idier et al. (2017) shows that the sign of high-tide level 451 changes obtained for the 'flood' and 'no flood' scenarios are the same across most 452 of the NW European shelf area (57% of the computational domain). Significant 453 local changes are observed especially along the German and Danish coasts (see e.g. 454 point C of Figure 5). As highlighted by Pelling et al. (2013a), local flooding can 455 have an effect at the basin scale: the flooding of the low-lying Dutch coast is the 456 main forcing for the response of the M2 amplitude to SLR seen in the North Sea. It 457 should be noted that there is a strong consistency between these modeling studies. 458 Idier et al. (2017) and Pelling and Green (2014) provide very similar M2 amplitude 459 changes in terms of order of magnitude and patterns. The main discrepancy is in 460 the Bristol Channel, which may be due to the differences in spatial resolution and 461 quality of the topographic data used in each case. 462

In San Francisco Bay, Holleman and Stacey (2014) and Wang et al. (2017) also 463 investigated the effect of coastal defense scenarios. We discussed in section 3.1 464 that a SLR of 1 m induces an increase of 6 and 5 cm in the southern and northern 465 bay in the case of hardened shoreline scenario based on the study of Holleman 466 and Stacey (2014). These authors made the same experiment with present coastal 467 defenses and topography, finding a decrease of the high tide level (relative to the 468 mean sea level) ranging from a few centimeters to 13 cm, i.e. an opposite change 469 compared to results obtained with the hardened shoreline scenarios. Wang et al. 470 (2017) also investigated the effect of coastal defenses on tides, considering two 471 scenarios: existing topography and full-bay containment that follows the existing 472 land boundary with an impermeable wall. Comparing the model results obtained 473 for both scenarios, they found that the semidiurnal mode exhibits local changes 474 at the shoreline up to 2 mm, changes in the diurnal mode extend into the bay 475 (reaching values of about 1 mm), and overtide changes exhibit a significant spatial 476 variability (with changes exceeding locally 2 mm). But the most important im-477 pact of the full-bay containment appears to be in the long-term process, with the 478

changes in the long-term tidal mode being almost uniform in space, and exceeding
5 mm.

Along the coast of Australia, Harker et al. (2019) found that the effect of coastal defense on M2 and K1 amplitude is very small for a SLR of 1 m, with changes smaller than 1 cm. As stated by the authors, this is likely due to the fact that allowing land to flood only increases the wetted area by few cells for this SLR scenario. For a larger SLR (7 m), the effect of allowing dry land to flood is much larger with changes ranging between -20 cm and +20 cm.

On the Patagonian Shelf, Carless et al. (2016) show that allowing model cells to flood leads to M2 amplitude changes larger than in the case of a fixed shoreline, with more negative changes in the 'flood' scenario, but also that the sign of changes can be locally reversed. The absolute difference of M2 amplitude changes between the 'flood' and 'no flood' scenarios can locally exceed 15 cm for SLR=1 m (Figure 6).

The above studies show that when previously dry land is allowed to flood, it can decrease the amplitude of high tide by a few or even tens of centimeters relative to the case where flooding is prevented. A key point is that initially flood defense schemes have localized benefits (they defend the coast line they protect). However, as shown for instance by Pelling and Green (2014), the process of flooding or not can have an impact on tides at the entire basin scale (e.g. North Sea).

⁴⁹⁹ 3.4 Wave effect on atmospheric storm surges

500 3.4.1 Shear stress at the sea surface

For a long time, the drag coefficient C_{Ds} used to compute the surface stress 501 due to wind (see Eq. 3) was assumed to increase linearly with wind speed. Al-502 though such a simple approach appears attractive for implementation in storm 503 surge models, it has several major shortcomings. First, several studies relying on 504 field and laboratory measurements suggested that under extreme winds the sea-505 surface roughness—and therefore the drag coefficient—could plateau after 30 m/s 506 and even decrease for higher winds (e.g. Powell et al. 2003). This behavior was 507 attributed to the development of wave-induced streaks of foam and sprays, which 508 tend to smooth the sea surface. Second, for a given wind speed, significant scatter 509 exists and C_{Ds} could vary by 30% or more. This scatter was partly explained by 510 the fact that the sea-surface roughness does not only depend on the wind speed 511 but also on the sea state. Following the pioneering work of Van Dorn (1953) and 512 Charnock (1955), Stewart (1974) proposed that, for a given wind speed, the sea-513 surface roughness should also depend on the wave age, which is defined as the 514 ratio between the short-wave velocity and the friction velocity. Using a coupled 515 wave and storm surge model, Mastenbroek et al. (1993) showed that using a wave-516 dependent surface stress could increase the surface stress by 20% while better 517 matching the observations. Performing a numerical hindcast of the February 1989 518 storm in the North Sea, they showed that using a wave-dependent drag parame-519 terization rather than the ones of Smith and Banke (1975) (quadratic wind-shear 520 stress with $C_{Ds} = f(U_{10})$ leads to an increase of 20 cm for the modeled highest 521 water level reached during the storm. In the Taiwan Strait, Zhang and Li (1996) 522 show that the wave effect on atmospheric storm surges is significant, reaching 523

about 20 cm for the typhoon Ellen (September 1983, 297 km/h gusts) and for 524 an observed surge peak of almost 1 m. The dependence of the surface stress to 525 the sea state was then corroborated in many studies (Moon et al. 2004; 2009; 526 Bertin et al. 2012; 2015). In particular, Bertin et al. (2015) performed a high-527 resolution hindcast of the storm surge associated with the Xynthia (2010) event 528 in the Bay of Biscay and showed that the young sea state associated with this 529 storm increased the surface stress by a factor of two. More precisely, they com-530 pared the surges obtained with quadratic formulation to the ones obtained with 531 wave-dependent parameterization to compute wind stress. They found that both 532 approaches perform similarly except during the storm peak, where the surge with 533 the wave-dependent parameterization for wind stress is 30% larger, i.e. several 534 tens of centimeters larger, bringing the model results closer to the observations. 535 All these studies show that the wave effect on the sea-surface roughness can lead 536 to increases in storm surge of a few centimeters up to tens of centimeters. 537

538 3.4.2 Shear stress at the sea bed

In coastal zones, the near-bottom orbital velocities associated with the propagation 539 of short waves become large and enhance bottom stress (Grant and Madsen 1979). 540 Several studies investigated the impact of wave-enhanced bottom stress on storm 541 surges (Xie et al. 2003; Nicolle et al. 2009) although it is not really clear whether 542 accounting for this process improves storm surge predictions or not (Jones and 543 Davies 1998). Bertin et al. (2015) argued that, most of the time, the tide gauges 544 used to validate storm surge models are located in harbors connected to deep 545 navigation channels, where bottom friction is rather a second order process and 546 affect storm surges by less than 0.1 m. Further research is needed, including field 547 measurements in shallow water. 548

549 3.5 Tide effect on waves and wave setup

550 3.5.1 Mechanisms

In the nearshore, tides can have a significant effect on short waves. First, tide-551 induced water level variations shifts the cross-shore position of the surfzone, so 552 that wave heights are modulated along a tidal cycle (Brown et al. 2013; Dodet 553 et al. 2013; Guérin et al. 2018). Second, in coastal zones subjected to strong tidal 554 currents such as estuaries and tidal inlets, tidal currents can substantially affect 555 the wave field (Ardhuin et al. 2012; Rusu et al. 2011; Dodet et al. 2013). Ne-556 glecting dissipation, the conservation of the short-wave energy flux implies that, 557 during the flood phase, waves following currents decrease while, during the ebb 558 phase, waves propagating against currents increase (Dodet et al. 2013; Bertin and 550 Olabarrieta 2016). However, during this ebb phase, as and when currents increase, 560 the increase in wave height together with the decrease in wavelength increases the 561 wave steepness. This increase in steepness can induce dissipation by whitecapping, 562 although this process affects mostly higher frequencies (Chawla and Kirby 2002; 563 Dodet et al. 2013; Bertin and Olabarrieta 2016; Zippel and Thomson 2017). In very 564 shallow inlets and estuaries, tidal currents can also reach the group speed of the 565 short waves so that full blocking can eventually occur (Dodet et al. 2013; Bertin 566

and Olabarrieta 2016). As a consequence of wave setup being controlled mainly
 by the spatial rate of dissipation of short waves, wave setup along the shoreline or
 inside estuaries can exhibit large tidal modulations. Such modulation is illustrated

inside estuaries can exhibit large tidal modulations. Such modulation is illustrated
by Dodet et al. (2013) and Fortunato et al. (2017) for the case of inlets; they show
that wave breaking is more intense on ebb shoals at low tide so that the associated

⁵⁷² setup in the lagoon/estuary is higher at low tide.

573 3.5.2 Orders of magnitude

In the Taiwan Strait, the numerical experiment of Yu et al. (2017) for Typhoon 574 Morakot (2009) shows that an increase of significant wave height (H_s) of about 575 0.5 m occurred at Sansha (China coast) during high water levels induced by tidal 576 change and atmospheric storm surge. The H_s difference at other estuary regions 577 is also significant with a range of -0.4 m at low water levels and 0.4 m at high 578 water levels. In locations with larger currents or depths, H_s appear less correlated 579 with water level variations. From the results of the numerical experiments, a weak 580 modulation of the wave setup is observed (up to about 2 cm, based on Figure 8 581 therein) for H_s of about 2 to 3 m. 582

Along the French Atlantic Coast (Truc Vert beach) during storm Johanna (10th March 2008), Pedreros et al. (2018) estimated wave setup modulations of tens of centimeters, sometimes exceeding 40 cm. These modulations were found at a given nearshore location, i.e. a fixed point located in the surf zone. At the waterline, the modulation is found to be much weaker (few centimeters).

Fortunato et al. (2017) performed a high resolution hindcast of the storm surge 588 associated with the 1941 storm in the Tagus Estuary (Portugal, Figure 2), which 589 corresponds to the most damaging event to strike this region over the last century. 590 Their numerical results suggest that the significant wave height of incident waves 591 exceeded 10 m and dissipation of the waves on the ebb shoal resulted in the 592 development of a wave setup reaching up to 0.35 m at the scale of the whole 593 estuary. This setup was tidally modulated and ranged from 0.10 to 0.15 m at high 594 tide (when dissipation was lowest) and 0.30 to 0.37 m at low tide. In addition, 595 Fortunato et al. (2017) revealed a more subtle mechanism, where wave setup is 596 also amplified along the estuary up to 25% through a resonant process. 597

⁵⁹⁸ 3.6 Sea-level rise effect on waves and wave setup

The impact of SLR on waves and wave setup is difficult to evaluate at sandy 599 beaches. On the one hand, assuming an unchanged bathymetry (i.e. no morpho-600 logical adaptation to SLR) and considering that the bottom slope increases along 601 the beach profile (i.e. Dean 1991), SLR would imply that waves would break over 602 a steeper bottom, which would result in an increase in wave setup. On the other 603 hand, it is more likely that the beach profile will translate onshore due to SLR 604 (Bruun 1962), so that wave setup would be globally unchanged. Due to large un-605 certainties concerning the response of sandy coastlines to SLR, we focus the review 606 on studies investigating the effect of SLR on reef environment, assuming that the 607 reef will weakly change for metric SLR. 608

Quataert et al. (2015) investigated this effect on Roi-Namur Island (Marshall islands, Figure 2), showing that an offshore water level increase (e.g. SLR) leads to a decrease of wave setup on the reef. According to their computations, a SLR of 1 m leads to a wave setup decrease of tens up to 50 cm (for $H_s=3.9$ m and $T_p=14$ s).

⁶¹³ They also show that these changes are increased by narrow reef and steep fore reef

⁶¹⁴ slope. On the Molokai coral reefs (Hawaii), Storlazzi et al. (2011) show that the ⁶¹⁵ SLR effect on wave height and wave setup leads to changes in total water level

(relative to SLR) of a few centimeters for SLR=1m. From these studies, assuming

⁶¹⁷ unchanged bathymetry in these coral reef environments, SLR has the potential

to induce wave setup decreases of a few centimeters up to tens of centimeters or

619 more.

620 4 Discussion

621 4.1 Synthesis

The literature review shows that site-specific knowledge of the importance of each 622 interaction mechanism is very heterogeneous (Figure 2) with some sites or regions 623 receiving little attention (see e.g. the West Africa coast) and others subject to many 624 studies (e.g. NW European shelf). In the literature, there are few environments or 625 sites where all the interactions investigated here have been quantified separately. 626 However, we can still extract some ranges (in order of magnitude) for some of the 627 interactions (Figure 6), to compare to the orders of magnitude of the main water 628 level components, keeping in mind that the importance of interactions strongly 629 depends on the location, the amount of SLR and the considered meteorological 630 event. 631

High tide amplitudes (relative to mean sea level) range from decimetric to 632 metric magnitudes (locally exceeding 8 m in the Bay of Fundy), but sea-level rise 633 can cause significant modifications. Changes in M2 amplitude or MHW changes 634 of several centimeters to more than 10 cm for SLR scenarios of \sim 1-2 m have been 635 obtained from modeling studies. Flooding of previously dry land can also induce 636 tide changes of several centimeters to more than 10 cm nearshore with potential 637 effects at the basin scale (as in the North Sea for instance). The flood effect is 638 mainly negative (i.e. it reduces the high tide level). 639

Wave setup ranges between a few to tens of centimeters and up to about 1 m. 640 We cannot, however, exclude that larger wave setup could occur. Tides can mod-641 ulate wave setup from several centimeters on open beaches to tens of centimeters 642 at inlet or estuary mouths (shoals). Nearshore areas (e.g. open sandy coasts and 643 inlets) are known to have dynamic morphologies on event, seasonal or pluriannual 644 time scales. On longer time scales, under the effect of SLR, the morphology is 645 expected to change significantly such that providing specific estimates of the SLR 646 effect on wave setup is perhaps too ambitious. However, assuming an unchanged 647 bathymetry such as in reef environments, wave setup (at the lagoon scale) is ex-648 pected to decrease O(10 cm), while on fixed open beaches, setup is expected to 649 increase. 650

Atmospheric surges range from a few centimeters to tens of centimeters, reaching a few meters in exceptional cases and up to 10 m in cyclonic areas. Again, we cannot exclude larger atmospheric storm surges. Tide-surge interactions can lead to surge changes of a few to tens of centimeters and up to almost 1 m. In addition, the wave effect on the sea-surface roughness can increase atmospheric surges up to a few tens of centimeters, especially under young sea states.

In summary, there exist locations where each interaction discussed in the 657 present review can reach values of a few tens of centimeters while some interaction 658 processes can contribute almost 1 m to coastal water level. These contributions to 659 total water level are far from negligible and must be considered in planning and 660 mitigation efforts. As highlighted above, site-specific knowledge of the importance 661 of each interaction mechanism is heterogeneous. However, the sites discussed here 662 give some indication where the significant interactions are likely to occur. Accord-663 ing to available literature, along the US East coast, the Patagonian coast, the 664 NW European coasts, and in San Francisco Bay and Taiwan Strait, there exists 665 a significant effect of SLR on tide and tide-surge interaction. All these sites are 666 characterized by substantial shallow-water areas. This suggests that there should 667 be other shallow areas subject to non-negligible SLR-tide-surge interactions, such 668 as Indonesia, for instance. Along open coasts with narrow continental shelves, the 669 SLR effect on tide, surge and tide-surge interaction is expected to be negligible. 670 Similarly, based on the shallow-water equations and the literature review, signifi-671 cant effects from waves on atmospheric storm surges are expected only in shallow-672 water areas. Regarding the effect on tides of letting previously dry areas flood or 673 not, it is not straightforward to identify the potential areas where this will be a 674 significant effect. However, we can still expect significant changes in shallow-water 675 areas characterized by low-lying coasts, especially in estuaries or tidal inlets. As a 676 result, the following areas (not exhaustive) appear sensitive to at least one of the 677 interactions investigated in the present paper: Hudson bay, European continental 678 shelf, straits with significant tidal currents (e.g. Taiwan), the northern coast of 679 Australia, the Patagonian shelf, Gulf of Mexico and San Francisco bay. 680

⁶⁸¹ 4.2 Combined interactions

The present review focused on a few of the possible interactions and on studies providing results concerning each individual interactions. However, multiple studies address the combined effect of multiple interactions, such as taking into account coincident interactions between tides, atmospheric surges, waves and wave setup (e.g. Brown et al. 2013), mainly at local scales.

At the scale of the U.S. East and Gulf coasts Marsooli and Lin (2018) investi-687 gated historical storm tides resulting from tide, atmospheric surge and wave setup 688 and the interactions between these components using a numerical modeling ap-689 proach, focusing on past tropical cyclones (1988-2015). Model results show that the 690 maximum water level rise due to nonlinear tide-surge interaction in most regions 691 along the U.S. East and Gulf Coasts (especially in Long Island Sound, Delaware 692 Bay, Long Bay and along the western coast of Florida) was relatively large but did 693 not occur at the timing of the peak storm tide. Model results at the location of 694 selected tide gauges showed that the contribution of tide-surge interaction to the 695 peak storm tide was between -25% and +20% (from -0.35 to 0.31 m). During the 696 most extreme storm events (i.e. TCs that caused a storm tide larger than 2 m), 697 the tide-surge interaction contribution ranged between -12% and 5%. Brown et al. 698 (2013) found a similar behavior on the Liverpool Bay, showing that the maximum 699

wave setup occurs at low water, stressing the effect of the banks (or shoals) at the mouth of the estuary.

Several studies quantified the effect of large coastal flooding on the still water 702 level (i.e. on tide and surge, implicitly including the tide effect on surge), mainly 703 at regional or local scales. For instance, Townend and Pethick (2002) showed that 704 the removal of coastal defenses along several English estuaries allowed the flooding 705 of extensive areas and result in a water level reduction locally exceeding 1 m. 706 Bertin et al. (2014) and Huguet et al. (2018) showed that the massive flooding 707 associated with Xynthia (February 2010, central part of the Bay of Biscay) induced 708 a still water level decrease ranging from 0.1 m at the entrance of the estuaries 709 to more than 1.0 m inside estuaries compared to a situation where the flooding 710 would have been prevented. Note that these two studies revealed that the impact 711 of freshwater discharges was negligible in the case of Xynthia, mostly because 712 713 freshwater discharges were close to yearly-mean values. Different conclusions might 714 be drawn for tropical hurricanes, which are usually associated with heavy rainfall. 715 These findings and the corresponding orders of magnitude have been corroborated in other French estuarine environments (Chaumillon et al. 2017; Waeles et al. 716 2016) 717

While there are few studies focusing on the sole effect of SLR on pure atmo-718 spheric surges, there are several studies investigating the effect on the still water 719 level or practical storm surge. For instance Arns et al. (2017) show that SLR sce-720 narios of 54 cm, 71 cm, and 174 cm lead to decreases of the practical atmospheric 721 storm surge (still water level minus tide) of 1.8%, 2.3% and 5.1%, on average in 722 the German Bight area. They show that, with SLR and a fixed bathymetry, the 723 modulation of waves by the storm tide should decrease. The observed correlation 724 between waves and storm tides decreases with SLR, such that waves and storm 725 tides become more independent with SLR. 726

As a last example of studies providing quantitative values on combined in-727 teractions, Krien et al. (2017a) investigated the effect of SLR on 100-year surge 728 levels (including atmospheric surge and wave setup) along the coasts of Martinica 729 (Caribbean islands). They show that a 1 m SLR leads to changes in the 100-year 730 surge levels ranging between -0.3 and +0.5 m in areas of coral reef and shelf. Most 731 of the domain (especially between the eastern shoreline and coral reefs) is sub-732 ject to a decrease (larger water depths induce a decrease in wave setup and wind 733 induced storm surge), while increase is observed in very local low-lying regions 734 where the inundation extent is strongly enhanced by the sea level rise. 735

⁷³⁶ 4.3 Implications for water-level and flood projections

Interactions between the contributions to coastal water level have significant im-737 plications for projected changes in the frequency and amplitude of future extreme 738 events, yet most projections neglect the interactions discussed in this paper and 739 consider only linear additions of the relevant processes (Muis et al. 2016; Vitousek 740 et al. 2017; Melet et al. 2018; Vousdoukas et al. 2018). Often, projected SLR is 741 added to current tidal datums with the assumption that a given increment of SLR 742 corresponds to an equivalent increment in the average high tide (i.e. MHW) or 743 some other relevant datum. As discussed above and demonstrated for instance by 744 Pickering et al. (2017), this assumption is not valid in many shelf seas across the 745

global ocean. The obvious implication of the nonlinear relationship between SLR 746 and tidal amplitude is an enhancement or reduction of future water level at high 747 tide. From a planning perspective, however, it is also important to consider how 748 the nonlinear response of tidal amplitude to SLR affects the timing of impacts. 749 Rates of SLR will be on the order of 10 cm per decade under the plausible scenario 750 that GMSL rises by 1 m during the 21st century. The nonlinear response of tidal 751 amplitude to 1 m of SLR is also on the order of ± 10 cm in some locations (Picker-752 ing et al. 2017). Thus, the effect of tidal amplification on the MHW datum can be 753 roughly equivalent to the effect of ± 1 decade of SLR. Depending on the sign and 754 magnitude of the tidal amplification at a given location, optimal planning horizons 755 may need to be adjusted earlier or later to account for the nonlinear response of 756 tidal amplitude and its effect on the frequency of high-tide flooding. 757

In addition, changes in mean sea level will cascade through the different in-758 teraction mechanisms with the net effect potentially resulting in several tens of 759 centimeters of additional (or reduced) water level at the coast during extreme 760 events (e.g. Arns et al. 2017). SLR will directly affect water levels associated with 761 tides, atmospheric storm surges and wave setup with additional indirect effects 762 on surges through tide-surge interactions. Even excluding the interaction between 763 these processes, a 10 cm increase in mean sea level corresponds to a doubling of 764 the frequency of former 50-year return total water levels over much of the global 765 ocean, with approximately 25 cm required for the doubling in areas most affected 766 by storm surge (Vitousek et al. 2017). These mean sea level changes are of the 767 same order of magnitude as the interaction effects (Figure 6), suggesting that the 768 frequency of the most extreme water level events could increase much faster than 769 currently projected in locations where constructive interactions between processes 770 are strongest. 771

Explicitly accounting for the interactions between processes in projections of 772 future total water level is not trivial. In practice, a detailed accounting of the in-773 teractions requires high-resolution tide, wave, and surge models (including wave 774 setup) with high-quality bathymetry and reasonable estimates of local bottom 775 friction. Indeed, even if in many cases a resolution of few hundred meters is suffi-776 cient to capture tide, atmospheric surge and their interaction (see e.g. Idier et al. 777 2012; Muller et al. 2014, in which 2 km resolution models are used), it is not 778 sufficient to capture regional and local wave setups, which, depending on the en-779 vironment, require resolutions of tens and few meters, respectively, but also high-780 quality bathymetry (see e.g. Bertin et al. 2015; Pedreros et al. 2018). The quality 781 of the topography (especially on the coastal defenses) is also crucial to account for 782 the flooding effect on coastal water level. As highlighted here and in section 4.4 783 (penultimate paragraph), there are downscaling challenges, but there are also up-784 scaling issues (indeed, local processes as for instance dissipation by floods or on 785 intertidal areas or tidal bedforms may also affect the regional dynamic). In ad-786 dition, to account for the uncertainties in the future climate projections, many 787 simulations should be done. Thus, global modeling assessments currently exceed 788 what is computationally feasible, while accurate bathymetric and bottom friction 789 information is not available in many locations. 790

Given these challenges, a statistical approach should be employed when possible to include process interactions in projections of flood frequency and return
periods of extreme events. Probabilistic projections of extreme water levels (e.g.
Vousdoukas et al. 2018) offer an opportunity to include statistical representations

of process interactions, but this approach has yet to be implemented. Fortunato 795 et al. (2016) proposed a solution to account for tide-surge interactions in extreme 796 value statistics for the coasts of the Iberic Peninsula, although the interactions are 797 not the strongest in this region. To include process interactions in probabilistic 798 projections of extreme water level events, statistical covariance relationships must 799 be established between the processes contributing to total water level. For exam-800 ple, one can envision a spatially variable covariance relationship between mean 801 sea level and tidal amplitude derived from scenario-based tide modeling studies 802 such as Pickering et al. (2017) or Schindelegger et al. (2018). General covariance 803 relationships between other processes could be estimated by leveraging regional 804 and local modeling frameworks designed for specific local case studies. Explor-805 ing physically-reasonable parameter spaces in a small number of locations that 806 span general classes of coastal environments (e.g. reef-lined islands, shelf seas, 807 etc.) could provide reasonable estimates of covariance relationships that could be 808 applied more generally to coastal environment classes in a global probabilistic 809 810 framework. Regardless of approach, a high priority must be placed on develop-811 ing covariance relationships between contributions to total water level in order to provide accurate projections of water level extremes. 812

⁸¹³ 4.4 Limits, gaps and remaining questions

The values and orders of magnitude provided in the present review should be used 814 with some caution. Indeed, this study is based on a literature review, keeping 815 in mind that some areas or environments have been subject to fewer investiga-816 tions than others, and that the values are dependent on the considered scenarios 817 (e.g. SLR values) and/or meteorological events. In addition, the review relies on 818 modeling studies (as this is the only way to distinguish every component and con-819 tribution), such that even if most of the modeling experiments have been validated 820 with observations, there are still some epistemic uncertainties related either to the 821 water level measurements or the modeling (errors in input data, e.g. bathymetry, 822 atmospheric forcing, etc.; representation or omission of the underlying dominant 823 process). For instance, Apotsos et al. (2007) showed that wave setup estimates 824 based on the depth-integrated approach of Longuet-Higgins and Stewart (1964) 825 can be biased low by a factor of two along the shoreline. While most storm surge 826 modeling systems rely on a similar approach, one can expect that wave setup is not 827 always well represented in 2D models. Also, field measurements in surf zones under 828 storm waves are limited to a very few studies (Guérin et al. 2018; Pedreros et al. 829 2018). These difficulties highlight an urgent need for detailed field observations of 830 wave setup under storm wave conditions. 831

Regarding validation of modeled tide changes, Schindelegger et al. (2018) in-832 vestigated the SLR effect on tides and made a thorough comparison of model 833 results with 45 tide gauge records. Their model reproduces the sign of observed 834 amplitude trends in 80% of the cases and captures considerable fractions of the 835 absolute M2 variability, specifically for stations in the Gulf of Mexico and the 836 Chesapeake-Delaware Bay system. This result gives strong support to the fact 837 that, in many locations, a significant part of observed tide changes could be at-838 tributed to the effect of SLR. Still, discrepancies in models/data remain in several 839 key locations, such as the European Shelf. Indeed, as highlighted by Haigh et al. 840

(submitted), many processes can cause changes in tides (sea-ice coverage, sea-bed 841 roughness, ocean stratification and internal tides, etc.), and it remains difficult to 842 associate observed changes in tide over the instrumental period with particular 843 forcing factors. Furthermore, it remains unclear how some of these forcing fac-844 tors would change with SLR and more 'generally' with the future climate. Finally, 845 whilst the agreement in local to regional scale tidal changes from differing models 846 (e.g. on the NW European Shelf) provides some confidence, it would be of value to 847 conduct further investigations of macro-tidal regions (such as the Bristol Channel) 848 with consideration of interim 'partial recession' or 'flood' scenarios depending on 849 coastal management priorities and sensitivities to the resulting change. 850

In the present review, we focus on SLR, tide, atmospheric surge, wave and 851 wave setup interactions. However, in estuaries and inlets, when not negligible, the 852 interaction between water level and river discharge should also be considered. Wa-853 ter depth changes are dependent on river discharge (Q_r) in estuarine locations. 854 An increase in Q_r will increase mean sea level locally, but the increased friction 855 of the incoming tide interacting with the outgoing river discharge will lead to a 856 decrease in tidal amplitude (Devlin et al. 2017). Krien et al. (2017b) performed a 857 high-resolution hindcast of the storm surge and flooding associated with cyclone 858 Sidr (2007) in the head Bay of Bengal. They showed that while tide-surge interac-859 tions can impact the storm surge by -0.5 to +0.5 m, accounting for river discharge 860 can also impact storm-induced flooding substantially. 861

The interactions investigated in the present review exhibit different spatial 862 scales; the wave setup changes induced by tidal water level and currents or SLR 863 occur mainly at a local scale, while tide-surge interactions and SLR effects have 864 local, regional and global scales. Indeed, tide and atmospheric surge dynamics re-865 sult from global, regional and local scale mechanisms. A few studies have provided 866 maps of the SLR effect on tide at the global scale. First, as much as possible, at 867 least in regions exhibiting a significant effect of SLR on tides, it is recommended 868 to take into account the global scale changes in the open boundary conditions of 869 regional or local modeling studies (as in Harker et al. 2019, for instance). Sec-870 ond, to the authors' knowledge there is no study at a global scale investigating 871 tide-surge interaction or SLR effect on atmospheric surge (except tide gauge based 872 studies, which provide local information, non-uniformly spread around the world; 873 see e.g. Arns et al. 2019). Such studies would be very helpful for hindcasts and 874 projections of still water level or for coastal flood hazard assessment by allowing 875 areas sensitive to SLR, tide and atmospheric surge interactions to be identified. 876

As mentioned above, one of the main issues when focusing on nearshore water 877 level is the temporal evolution of the sea-bed topography/morphology. Indeed, 878 such changes, especially for sandy beaches exposed to waves, can have a significant 879 effect on the wave setup (at different time scales: event, seasonal, inter-annual to 880 longer term). For instance, Thiébot et al. (2012) and Brivois et al. (2012) show 881 how, depending on the wave and water level characteristics, different morphologies 882 can emerge on doubled sandbar systems, and thus can alter water level at the 883 coast, and especially the wave setup. Morphological changes can also alter the 884 tidal dynamics and related water levels. Ferrarin et al. (2015) investigated the 885 effect of morphology changes and MSL on tides in Venice lagoon over the last 70 886 years. While tidal amplitudes in the North Adriatic Sea did not change significantly 887 (even if they exhibit some fluctuations), morphological changes that occurred in 888 the lagoon in the last century produced an increase in the amplitude of major 889

tidal constituents (e.g. M2 increase up to about 20% with respect to the imposed tidal wave). This raises the question of how the sole effect of a changing seabed morphology (related to SLR, wave climate change, or human intervention)

compares with the interactions investigated in this review.

⁸⁹⁴ 5 Conclusions and perspectives

The present paper focused on water level at the coast resulting from the interac-895 tion between SLR, tides, atmospheric surge and wave setup. While the discussed 896 mechanisms of interaction were previously known, we provide an overview of quan-897 tifications of the interactions based on modeling studies. The largest identified 898 interaction is the tide-surge interaction which can lead to changes in the practical 899 atmospheric surge up to 1 m or more. SLR with a metric value can induce high 900 tide changes (positive or negative) exceeding 10 cm, while waves can increase the 901 atmospheric storm surge a few centimeters up to tens of centimeters. The flood 902 effect on tide and still water level can induce changes (mainly negative) ranging 903 from a few centimeters up to more than 1 m in estuaries. The modulation of wave 904 setup by the tide can represent a few centimeters at the shoreline, but can reach 905 tens of centimeters on ebb shoals. The effect of SLR on wave setup is more debat-906 able, but could be positive or negative depending on the nearshore bathymetry 907 and beach slope. 908

The review also suggests that these interactions have smaller magnitudes in deeper areas, and can thus be considered as negligible in some locations. On the contrary, many studies show significant interactions in shallow-water. The sensitive areas we identified include Hudson bay, European continental shelves, straits with significant tidal currents (e.g. Taiwan), the northern coast of Australia, the Patagonian shelf and the Gulf of Mexico. However, the spatial distribution of studies focusing on these interactions is heterogeneous.

In areas where non-negligible interactions are expected, we suggest these interactions to be taken into account (either by numerical modeling or by statistical
methods) in assessments of nearshore water levels and induced coastal flooding.
Including interaction mechanisms is especially important when projecting future
coastal flooding including SLR.

The present review focused on a subset of the possible interactions, but within 921 the estimation of future water levels, additional complexity arises. Regarding fu-922 ture tide changes, further investigations are needed to better identify if and which 923 phenomena other than the sea-level rise could have a significant effect on tides. 924 Finally, the nearshore area is a key zone for tide and surge dissipation, but also 925 wave setup. However, especially under a rising sea level, significant morphological 926 changes are expected in these nearshore areas. The effect of changes in nearshore 927 bathymetry deserves more attention, at least considering scenarios to better char-928 acterize the sensitivity of water level to these changes in comparison with changes 929 induced by the interactions discussed in the present paper. 930

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Fig. 1 (a) Components of storm tide, terminology and sketch of interactions. (b) main interactions between mean sea level, waves, atmospheric storm surges, tide, wave setup and flooding. In bold and black: the focus of the present paper.



Fig. 2 Global bathymetry (seabed level (m), from GEBCO) and areas investigated in the papers selected to provide orders of magnitude of each of the interactions described in the present review. Stars correspond to very local studies.



Fig. 3 Mean High water changes in the 136 cities of population larger than 1 millions (in 2005) investigated by Pickering et al. (2017). (a) Empirical Cumulative Distribution for various SLR scenarios characterized by a global mean SLR of +2 m: case of a fixed shoreline ('no flood') with a uniform SLR (2UF) or non-uniform SLR corresponding to the initial elastic response of ice sheet melt in Greenland (2NUGF), Western Antarctica (2NUWAF), or both (2NUBF); case with recession ('flood') with a uniform SLR (2UR) or non-uniform SLR corresponding to the initial elastic response of the initial elastic response of ice sheet melt in Greenland and Western Antarctica (2NUBR). (b) MHW change for the SLR scenario 2UF. Figures produced here based on the data provided in the supplementary material of (Pickering et al. 2017).



Fig. 4 Tide-surge interactions (in m) for the storms of the 10-11 March 2008 and 9-10 November 2007 in the English Channel, after the study of Idier et al. (2012).



Fig. 5 Changes in maximum annual high tide (2009) versus SLR for points A (Mont Saint Michel Bay) and C (German Bight), for the 'no flood' and 'flood' scenarios, after the study of Idier et al. (2017).



Fig. 6 Orders of magnitude of main water level components (in color) and some of the interactions (in black) investigated in the present paper. The orders of magnitude are based on the publications reviewed in the present paper and on (Woodworth et al. 2019). It should be noted that these orders of magnitude are provided 'at the coast' (i.e. at the waterline). This does not exclude larger effect in the nearshore (for instance for tide modulation of wave setup in the surf zone).