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Interactions Between Mean Sea Level, Tide, Surge, Waves and Flooding: Mechanisms and Contributions to Sea Level Variations at the Coast

Déborah Idier · Xavier Bertin · Philip
Thompson · Mark D. Pickering

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Abstract Coastal areas epitomize the notion of ‘at risk’ territory in the context 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 and wave setup is critical for coastal flooding assessment. This study investigates how coastal water level can be altered by interactions between SLR, tides, storm surges, waves and flooding. The main mechanisms of interaction are identified, mainly by analyzing the shallow-water equations. Based on a literature review, the orders of magnitude of these interactions are estimated in different environments. The investigated interactions exhibit a strong spatio-temporal variability. Depending on the type of environments (e.g. morphology, hydro-meteorological context), they can reach several tens of centimeters (positive or negative). As a consequence, probabilistic projections of future coastal water levels and flooding should identify whether interaction processes are of leading order, and, where appropriate, projections should account for these interactions through modeling or statistical methods.

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

D. Idier
BRGM, 3 av. C. Guillemin, 45060 Orléans Cédex, France
E-mail: d.idier@brgm.fr

X. Bertin
UMR 7266 LIENSs, CNRS - La Rochelle University, 2 rue Olympe de Gouges, 17000 La Rochelle, France
E-mail: xbertin@univ-lr.fr

P. Thompson
Department of Oceanography, University of Hawai‘i at Manoa, 1000 Pope Road Honolulu, HI 96822, USA.
E-mail: philiprt@hawaii.edu

M. D. Pickering
University of Southampton, University Road, Southampton, SO17 1BJ, United Kingdom.
E-mail: mdp053@gmail.com

1 Introduction

Coastal areas are considered ‘at-risk’ territories in the context of climate change and sea-level rise (SLR). Knowledge of water level variability at the coast, especially the still water level and storm tide, is critical for coastal flooding assessment, both for present and future climate. Still water level includes mean sea level, tide and atmospheric surge (Figure 1a). Several definitions can be found for storm tides; here, we consider that they include mean sea level, tide, atmospheric storm 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 assessment is to linearly add the sea-level rise to selected still water level scenarios to estimate the flood hazard. Such an approach relies on the underlying assumption that there is no significant nonlinear interaction between contributions to water level variability, such as SLR, tide and surge. However, some studies show evidence that these interactions can represent a significant part of water level changes. For instance, using a numerical modeling approach, focusing on the German Bight area (SE of North Sea) and assuming a sea-level rise of 0.54 m, Arns et al. (2015) show that taking into account the interactions between mean sea level, tide and atmospheric surge leads to a 50-year return still water level 12 cm larger than when these interactions are neglected, corresponding to a doubling of the frequency of 50-year return still water level obtained neglecting the interactions.

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

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

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**

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

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

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

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

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

$$\frac{\partial \xi}{\partial t} + \nabla \cdot (D \cdot \mathbf{u}) = 0 \quad (1)$$

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

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

$$\tau_s = \rho_a C_{D_s} \mathbf{U}_{10} |\mathbf{U}_{10}| \quad (3)$$

$$\tau_b = C_{D_b} \mathbf{u} |\mathbf{u}| \quad (4)$$

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

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

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
162 (Raubenheimer et al. 2001; Apotsos et al. 2007), suggesting that other processes
163 may be involved. Apotsos et al. (2007) and Bennis et al. (2014) proposed that
164 bed-shear stress associated with the undertow (bed return flow that develops in
165 surfzones) could contribute to wave setup. Recently, Gu erin et al. (2018) showed
166 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.

169 A broader overview of all forcing factors causing sea level changes at the coast
170 and of their orders of magnitude, from sub-daily (seiche, infra-gravity waves) to
171 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*

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

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

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

206 *3.1.2 Orders of magnitude*

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

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

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

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.
257 As an additional comparison, we can also refer to the study of Palmer et al. (2018),
258 which shows strikingly similar spatial patterns of increase and decrease to the one
259 of Pickering et al. (2012), except for the region that spreads out from the Bristol
260 Channel. For a more detailed comparison on the modeling studies of the SLR effect
261 on tides over the European shelf, see Idier et al. (2017).

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

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

276 In the Southern Hemisphere, along the coast of Australia, Harker et al. (2019)
277 found M2 amplitude changes ranging from -0.1 to +0.1 m for a SLR of 1 m with
278 amplitude changes that are not proportional to the SLR in this area, even if the
279 patterns appear similar (in terms of areas of increase or decrease of M2 amplitude).

280 The effect of SLR on tides in the San Francisco Bay has also been investigated.
281 This bay is characterized by an inlet (width of about 2.5 km) and two main bays
282 (one to South, one to the North) of tens kilometers length and widths varying
283 from few kilometers up to almost 20 km. The mean tidal range varies from 1.8 to
284 about 2.7 m, after the tide data provided by NOAA. With scenarios of hardened
285 shoreline and a SLR of 1 m, the high tide exhibits an increase of 6 cm and 5 cm
286 in the southern and northern bay, respectively, i.e. more than 10% of the natural
287 tide amplification in these bays (Holleman and Stacey 2014).

288 Regarding the U.S. East coast, the M2 amplitudes range from about 0.4 to
289 1.5 m (values extracted from the FES2014 tidal components database ; Carrere
290 et al. 2015). Ross et al. (2017) show that for SLR=1 m, more than half of the
291 Delaware Bay is projected to experience an M2 amplitude increase of at least 15
292 cm, while the Chesapeake Bay exhibits a small decrease at the mouth (-2 cm), and
293 an increase over most of the bay up to 10 cm at the head. It is also highlighted
294 that changes are proportional to the SLR (especially for SLR=[-1;+1]m).

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

298 All these studies converge to highlight that the effect of metric SLR can lead
299 to tide changes (M2 component or spring high tide) up to ± 10 –15% of SLR. We

300 could not identify any studies focusing on the sole effect of SLR on atmospheric
 301 surge. However, several studies investigated the effect of SLR on still water level or
 302 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

306 The present subsection focuses on the effect of tide on surge. However, it should be
 307 kept in mind that this is a matter of point of view, and that strictly speaking the
 308 mechanisms can be referred to as tide-surge interactions. In addition, we focus on
 309 the practical storm surge (also called residual), which is the difference between the
 310 still water level and the tide level, such that it includes both the pure atmospheric
 311 storm surge and the effect of the tide on the surge.

312 3.2.1 Mechanisms

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

- 315 – the advective effect arising from the advective terms of the momentum equation
 316 (2).
- 317 – the shallow-water effect, which arises from nonlinearity related to $D = (H + \xi)$
 318 in equations (1), (2) and (3) in the following terms: advective term of the
 319 continuity equation, division by the depth D for the bed-friction and wind
 320 forcing.
- 321 – the nonlinear effect of the bottom friction term with the quadratic parametrisation
 322 in equation (4).

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

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

352 *3.2.2 Orders of magnitude*

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

371 The Bay of Bengal is another region where tide-surge interactions are known
372 to be very relevant (Johns et al. 1985). Krien et al. (2017b) performed numerical
373 experiments during cyclone Sidr (2007), focusing on the head of the bay. They
374 investigated the interaction between the tide and the total storm surge (including
375 wave setup). As the computed wave setup ranged from 0.2 to 0.3 m and varies
376 little over a tidal cycle, the tide-surge interactions analyzed by these authors mostly
377 corresponds to the effect of the tide on atmospheric surge and vice versa. They
378 showed that tide-surge interactions in the range ± 0.6 m develop in shallow areas of
379 this large deltaic zone. In addition, such interactions occurred at a maximum of 1–2
380 hours after low tide due to the combination of a stronger wind contribution during
381 periods of shallow depth and a faster propagating tide compared to a situation
382 without surge. These findings corroborate those of Johns et al. (1985), Antony and
383 Unnikrishnan (2013) and Hussain and Tajima (2017).

384 The Taiwan Strait is one example where the pattern of strong tidal currents and
385 storm-induced currents along the channel direction enhances tide-surge interaction
386 via nonlinear bottom friction (Zhang et al. 2010). This strait is subject to large
387 storm surges frequently occurring during the typhoon season; from 1949 to 1990
388 there were 69 typhoons inducing storm surges over 1 m along at the western bank of
389 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 Fujian coast, the west bank of the Taiwan Strait, during Typhoon Dan (1999) (for this event, surge ranged from 0.6 m to more than 1 m at the tide gauges). The numerical experiments of Zhang et al. (2010) show that these oscillations are due 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.

3.3 Flooding effect on tide

3.3.1 Mechanisms

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

432 3.3.2 Orders of magnitude

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

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

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

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

479 changes in the long-term tidal mode being almost uniform in space, and exceeding
480 5 mm.

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

487 On the Patagonian Shelf, Carless et al. (2016) show that allowing model cells
488 to flood leads to M2 amplitude changes larger than in the case of a fixed shoreline,
489 with more negative changes in the ‘flood’ scenario, but also that the sign of changes
490 can be locally reversed. The absolute difference of M2 amplitude changes between
491 the ‘flood’ and ‘no flood’ scenarios can locally exceed 15 cm for SLR=1 m (Figure
492 6).

493 The above studies show that when previously dry land is allowed to flood, it
494 can decrease the amplitude of high tide by a few or even tens of centimeters relative
495 to the case where flooding is prevented. A key point is that initially flood defense
496 schemes have localized benefits (they defend the coast line they protect). However,
497 as shown for instance by Pelling and Green (2014), the process of flooding or not
498 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

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

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

538 *3.4.2 Shear stress at the sea bed*

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

549 3.5 Tide effect on waves and wave setup

550 *3.5.1 Mechanisms*

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

567 and Olabarrieta 2016). As a consequence of wave setup being controlled mainly
568 by the spatial rate of dissipation of short waves, wave setup along the shoreline or
569 inside estuaries can exhibit large tidal modulations. Such modulation is illustrated
570 by Dodet et al. (2013) and Fortunato et al. (2017) for the case of inlets; they show
571 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

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

583 Along the French Atlantic Coast (Truc Vert beach) during storm Johanna
584 (10th March 2008), Pedreros et al. (2018) estimated wave setup modulations of
585 tens of centimeters, sometimes exceeding 40 cm. These modulations were found
586 at a given nearshore location, i.e. a fixed point located in the surf zone. At the
587 waterline, the modulation is found to be much weaker (few centimeters).

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

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

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

609 Quataert et al. (2015) investigated this effect on Roi-Namur Island (Marshall
610 islands, Figure 2), showing that an offshore water level increase (e.g. SLR) leads

611 to a decrease of wave setup on the reef. According to their computations, a SLR of
612 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
616 (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
618 to induce wave setup decreases of a few centimeters up to tens of centimeters or
619 more.

620 4 Discussion

621 4.1 Synthesis

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

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

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

651 Atmospheric surges range from a few centimeters to tens of centimeters, reach-
652 ing a few meters in exceptional cases and up to 10 m in cyclonic areas. Again, we
653 cannot exclude larger atmospheric storm surges. Tide-surge interactions can lead
654 to surge changes of a few to tens of centimeters and up to almost 1 m. In addition,

655 the wave effect on the sea-surface roughness can increase atmospheric surges up
656 to a few tens of centimeters, especially under young sea states.

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

681 4.2 Combined interactions

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

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

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

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

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

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

736 4.3 Implications for water-level and flood projections

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

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

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

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

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

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

813 4.4 Limits, gaps and remaining questions

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

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

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

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

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

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

890 tidal constituents (e.g. M2 increase up to about 20% with respect to the imposed
891 tidal wave). This raises the question of how the sole effect of a changing sea-
892 bed morphology (related to SLR, wave climate change, or human intervention)
893 compares with the interactions investigated in this review.

894 5 Conclusions and perspectives

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

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

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

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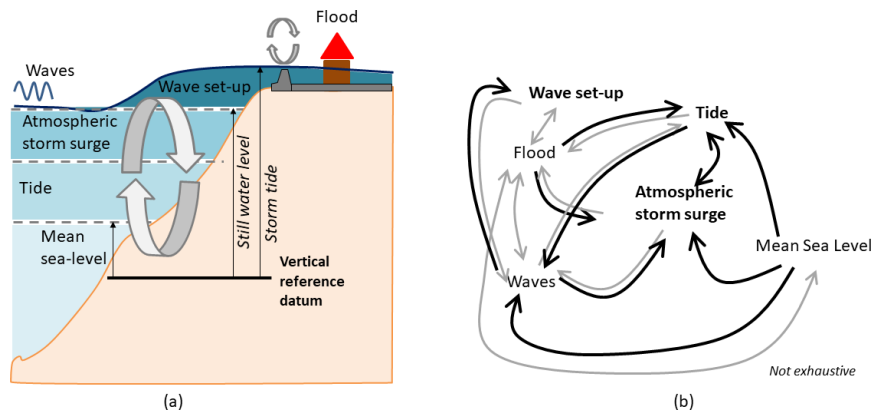


Fig. 1 (a) Components of storm tide, terminology and sketch of interactions. (b) main interactions between mean sea level, waves, atmospheric storm surges, tide, wave setup and flooding. In bold and black: the focus of the present paper.

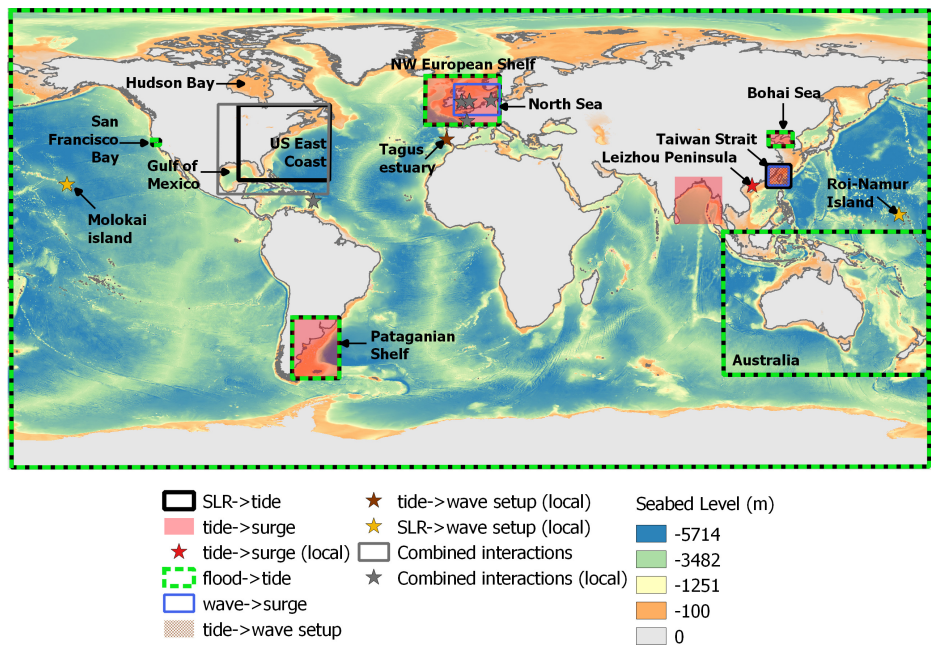


Fig. 2 Global bathymetry (seabed level (m), from GEBCO) and areas investigated in the papers selected to provide orders of magnitude of each of the interactions described in the present review. Stars correspond to very local studies.

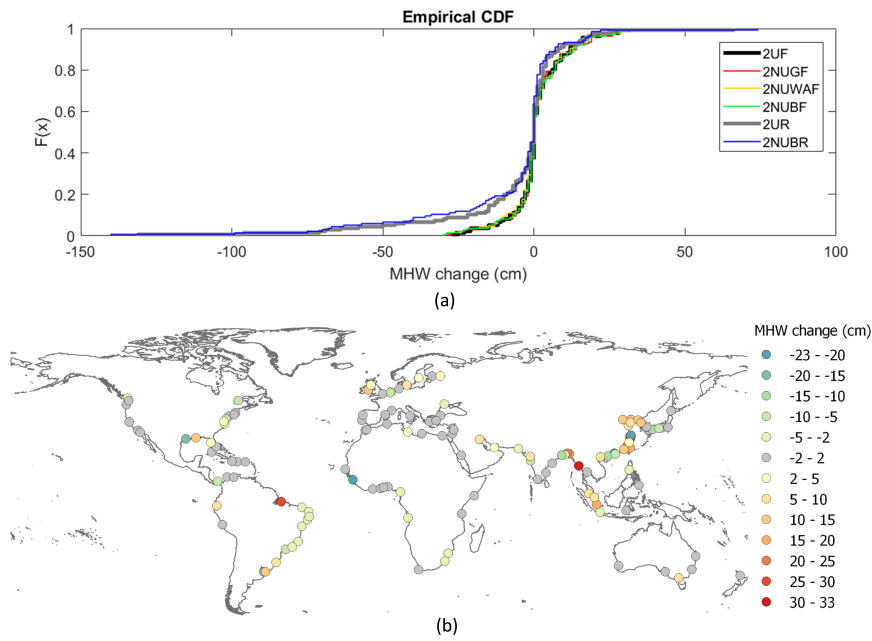


Fig. 3 Mean High water changes in the 136 cities of population larger than 1 millions (in 2005) investigated by Pickering et al. (2017). (a) Empirical Cumulative Distribution for various SLR scenarios characterized by a global mean SLR of +2 m: case of a fixed shoreline ('no flood') with a uniform SLR (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 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).

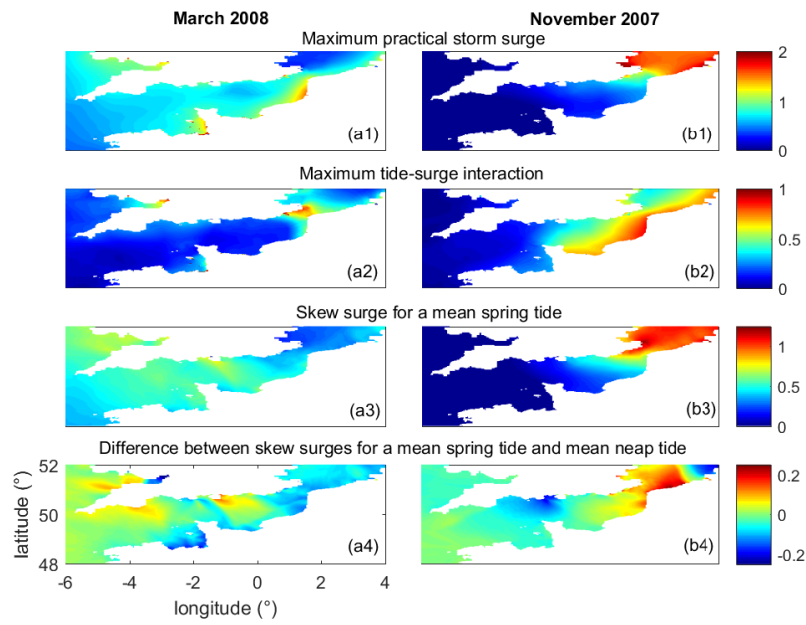


Fig. 4 Tide-surge interactions (in m) for the storms of the 10-11 March 2008 and 9-10 November 2007 in the English Channel, after the study of Idier et al. (2012).

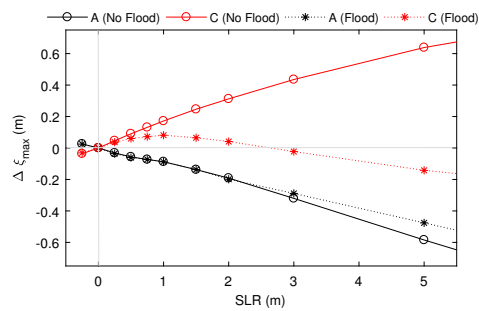


Fig. 5 Changes in maximum annual high tide (2009) versus SLR for points A (Mont Saint Michel Bay) and C (German Bight), for the 'no flood' and 'flood' scenarios, after the study of Idier et al. (2017).

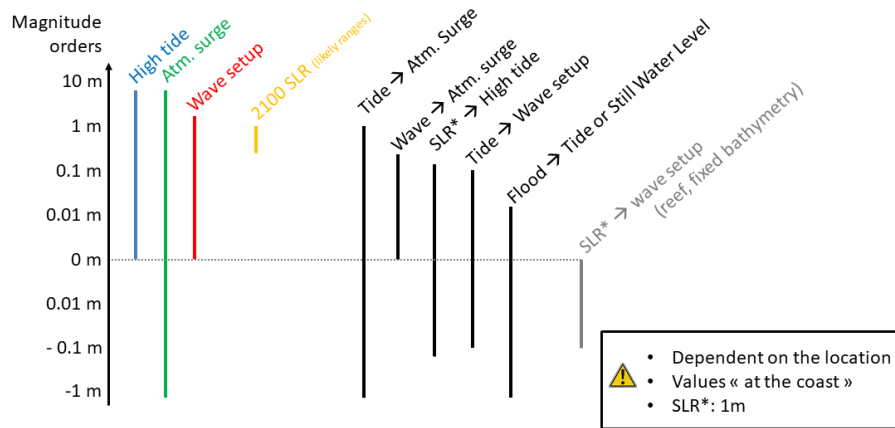


Fig. 6 Orders of magnitude of main water level components (in color) and some of the interactions (in black) investigated in the present paper. The orders of magnitude are based on the publications reviewed in the present paper and on (Woodworth et al. 2019). It should be noted that these orders of magnitude 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).