



UNIVERSITÀ POLITECNICA DELLE MARCHE
Repository ISTITUZIONALE

Wave-forced dynamics in the nearshore river mouths, and swash zones

This is the peer reviewed version of the following article:

Original

Wave-forced dynamics in the nearshore river mouths, and swash zones / Brocchini, M.. - In: EARTH SURFACE PROCESSES AND LANDFORMS. - ISSN 0197-9337. - STAMPA. - 45:1(2020), pp. 75-95. [10.1002/esp.4699]

Availability:

This version is available at: 11566/270464 since: 2024-05-30T13:16:30Z

Publisher:

Published

DOI:10.1002/esp.4699

Terms of use:

The terms and conditions for the reuse of this version of the manuscript are specified in the publishing policy. The use of copyrighted works requires the consent of the rights' holder (author or publisher). Works made available under a Creative Commons license or a Publisher's custom-made license can be used according to the terms and conditions contained therein. See editor's website for further information and terms and conditions.

This item was downloaded from IRIS Università Politecnica delle Marche (<https://iris.univpm.it>). When citing, please refer to the published version.

(Article begins on next page)

1 Wave-forced dynamics in the nearshore river mouths, and swash 2 zones

3 Maurizio Brocchini

4 DICEA, Università Politecnica delle Marche, Ancona, Italy

5

6 Abstract

7 The role of wave forcing on the main hydro-morphological dynamics evolving in the shallow
8 waters of the nearshore and at river mouths is analyzed. Focus is mainly on the cross-shore
9 dynamics that evolve over mildly sloping barred, dissipative sandy beaches from the storm
10 up to the yearly time scale, at most. Local mechanisms, nonlocal mechanisms and
11 connections across three main inter-related subsystems of the nearshore - the region of
12 generation and evolution of nearshore bars, river mouths and the swash zone - are
13 analyzed. The beach slope is a major controlling parameter for all nearshore dynamics. A
14 local mechanism that must be properly described for a suitable representation of wave-
15 forced dynamics of all such three subsystems is the proper correlation between orbital
16 velocity and sediment concentration in the bottom boundary layer; while specific
17 mechanisms are the wave-current interaction and bar generation at river mouths and the
18 sediment presuspension at the swash zone. Fundamental nonlocal mechanisms are both
19 Infragravity (IG) waves and large-scale horizontal vortices (i.e. with vertical axes), both
20 influencing the hydrodynamics, the sediment transport and the seabed morphology across
21 the whole nearshore. Major connections across the three subsystems are the upriver
22 propagation of IG waves generated by breaking sea waves and swash-swash interactions,
23 the interplay between the swash zone and along-river-flank sediment transport and the
24 evolution of nearshore sand bars.

25

26 Introduction

27 Focus, motivation and method

28 This work focuses on wave-forced nearshore and river mouth dynamics, as opposite to
29 tidally-forced dynamics - i.e. such that $RTR=TR/H_b < 3$, where TR is the tidal range and H_b is
30 the height of breaking waves (Masselink and Short, 1993).

31 In more detail, cross-shore (mostly) dynamics evolving from the yearly down to the storm
32 time scales are investigated. With reference to coastal and river mouth hydro-morphological
33 regimes the present analysis gives attention to the following two regimes.

34 First, open-coast, mildly sloping, barred ($\Omega=H_b/w_s T_i > 2$, where Ω is the Dean parameter),
35 w_s is the sediment fall velocity and T_i is the period of incoming waves - Wright and Short,
36 1984) and dissipative ($\epsilon=A_b \omega^2 / g \tan^2 \beta > 20$, where the surf-scaling parameter ϵ depends on
37 the amplitude of the breaking waves A_b , on the radian wave frequency $\omega=2\pi/T_i$ and on the
38 beach slope β -Wright and Short, 1984) sandy beaches. Second, wave-dominated or river-
39 dominated and wave-modified (which feed sediments to sea) river mouths (Cooper, 2001).

40 Scopes of the present contribution are the following. First, to provide a fairly complete
41 (obviously not exhaustive) and clear overall view of what we know and what we do not
42 know of wave-forced nearshore and river mouth dynamics. Second, to give the needed
43 focus to recent progresses made available on specific dynamics. But the main aim is to
44 **highlight links and relations among such dynamics**, this by properly highlighting
45 nonlocal agents and relations.

46 Regarding this problem as a puzzle to be solved (see figure 1), this entailing local and
47 nonlocal relations, this paper tries to: put the analysis in the proper context, with no
48 pretention for a systematic description; move from consolidated knowledge to new results;
49 move from observations to modeling; move from the large to the small scales; inspect
50 different types of models, both in terms of their structure and use.

52
53
54
55
56
57

Figure 1. Wave-forced dynamics in the nearshore and estuaries: a puzzle to be solved. This photo, taken at the microtidal Misa River estuary (Senigallia, Italy), illustrates through one single view the three subsystems analyzed in the paper.

58 The latter issue is analyzed in some detail because of its intrinsic importance and of the
59 guidance that models provide in exploring the mechanisms of the phenomena at hand.

60

61 Generalities on Models

62 The hydro-morphodynamics of the nearshore region, including that evolving at river mouths,
63 is so complex that a number of different methods and models has been proposed and
64 implemented over the years. However, such abundance of models has also brought to some
65 major ambiguities in the definition and use of such models.

66 Hence, some clarification seems useful, which is based on a detailed classification of the
67 models in terms of their: intrinsic structure and use.

68

69 **Model structure**

70 Although this classification, as many others, is somewhat arbitrary (e.g. as function of the
71 level of closure implemented), its use may help understand what model is actually applied
72 for a specific analysis. The structure of the models at hand can be classed as either “process
73 based” or “empirically based”.

74 “Process-based models” rely on equations derived from fundamental physical principles
75 and/or conservation laws. In the following three examples of such models (at different levels
76 of time/space scales and resolution) are described.

77 First, wave-averaged models: Nonlinear Shallow Water Equation (NSWE)-type (e.g.
78 Delft3D, <https://oss.deltares.nl/web/delft3d/home>; Lesser et al., 2004; NHWAVE, Ma et al.,
79 2012); Oceanographic (e.g. ROMS, <https://www.myroms.org/index.php>, Haas and Warner,

80 2009), etc. Second, wave-resolving models: NSWE-type (e.g. Postacchini et al., 2012);
81 Boussinesq-type (e.g. Brocchini, 2013; Kirby, 2016); etc. Third, models that account for
82 turbulence through proper closures: Reynolds Averaged Navier-Stokes; Large Eddy
83 Simulations (Chang and Scotti, 2004; Torres-Freyermuth, Lara and Losada, 2010; Lubin et
84 al. 2011) etc.

85 “Empirically-based models” rely on equations constructed from a mixture of conservation
86 laws and observations, with a significant weight of empirical inputs. In fact, though also
87 turbulence models like RANS and LES do require some empirism for the turbulence closure
88 relations (hence the arbitrariness of classing models), their structure heavily relies on
89 conservation laws and less on closures that are based on solid theoretical foundations.
90 Examples of empirically-based models are: Regression-type models, Machine Learning
91 Approaches (e.g. Pourzangbar et al., 2017), etc. and energetics-type, which solve a
92 sediment mass conservation equation (Exner equation) where the mass fluxes come from
93 empirically-based closure laws (e.g. Bailard, 1981).

94

95 **Model use**

96 The above models can be, each in its own way, used in terms of either a template/forced
97 approach or a stability approach (at times referred to as “self-organization”, see, for
98 example, Coco and Murray, 2007). To clarify the above an example is made with reference
99 to the classical Exner 1D sediment transport equation, which can be formulated as (e.g.
100 Plant et al., 2001):

$$101 \quad \frac{\partial z_b}{\partial t}(x, t) = I + II + III \quad (1)$$

102 where z_b is the seabed position and

103 I = is a forcing term due to initial bathymetry and wave field,

104 II = gives the modification of forcing due to externally-forced changes in wave height,

105 III = is a feedback term due to the rate of change of the seabed position z_b , which introduces
106 a feedback behavior in (1).

107 Template/forced models and stability-type models can be distinguished by the presence
108 of I, II and III. Template/forced morphological models are such that $I \neq 0$, $II \neq 0$ and $III = 0$, while
109 stability approaches must have $III \neq 0$. With reference to the specific problem of seabed
110 features generation, while template models may identify compelling mechanisms, i.e. a
111 template, for the initial stages of formation (e.g. at breakpoint, at nodes/antinodes of
112 standing waves, etc.) stability models are such that perturbations evolve simultaneously in
113 the hydrodynamic and bathymetric fields. This is illustrated by the examples given in figures
114 2 and 3. The initial bathymetry used for the simulations of figure 3 consisted of a longshore-
115 uniform sand bar.

116
117
118 Figure 2. Forced/template mechanism: “synchronous standing edge waves” (left) and flow
119 imprinted on sand giving beach cusps (right). Adapted from Coco and Murray, (2007).

120
121
122 Figure 3. Stability/Self-organization mechanism: crescentic bars emerge from the feedback
123 between flow and bathymetry (initial longshore-uniform bar bathymetry). Top: waves
124 breaking on shoals force onshore flow, which returns to sea in gaps. Bottom: for suspended
125 sediment decreasing from breaker line to shore, the onshore flow favors deposition (shoals
126 grow) and the offshore flow favors erosion (scour holes grow). Hence, the development of
127 crescentic bars. Adapted from Coco and Murray, (2007).

128
129
130 The rest of the paper is structured as follows. The next section provides detailed
131 information specific to 1) the generation and evolution of nearshore bars, 2) river mouth and
132 3) swash zone dynamics. A short Discussion and then Conclusions close the paper.

133

134 Wave forced-dynamics for three fundamental subsystems: sand
135 bars, river mouths and swash zone

136 Sand bars

137 The generation and evolution of sand bars in the nearshore is analyzed as a first applicative
138 topic because of its many implications, which help set the scene for all the rest of the paper.
139 Consolidated observations and models are first illustrated, then followed by more recent
140 findings.

141 Starting from the observations, sand bars that evolve in the nearshore are usually classed
142 in terms of their planview shape, i.e. longshore uniform (2D) and rhythmic (3D). Different
143 evolution mechanisms lead to different evolution time scales (e.g. Wijnberg and Kroon,
144 2002). We can distinguish between: 2D sand bars (left panel of figure 4), which emerge as
145 the outcome of wave-induced, cross-shore sediment transport and evolve over scales of
146 days to months (e.g. Ruessink and Kroon, 1994); 3D sand bars (right panel of figure 4),
147 which emerge as the outcome of horizontal circulation patterns and, because of this stronger
148 forcing, evolve on shorter time scales, i.e. hours to days (e.g. Caballeria et al., 2002).

149
150
151 Figure 4. Left panel: planview image of two shore-parallel sand bars at Noordwijk beach
152 (The Netherlands, 19th November 2000). Right panel: planview image of surf zone complex
153 topography at Duck (U.S.A. Atlantic coast, 10th January 1994). Adapted from Ribas Prats
154 (2003).

155
156

157 Three main interpretative models exist for the overall generation and evolution
158 mechanisms of nearshore sandbars.

159 The most known is the Net Offshore Migration (NOffM) model, which suggests that
160 multiple nearshore 2D bars have multi-annual lifetimes and pass through 3 stages (figure
161 5). Generation occurs close to shore because of the interaction of undertow, wave orbital
162 velocity asymmetry and group-forced IG orbital motion (Roelvink and Stive, 1989). This is
163 followed by a net seaward migration, which is the outcome of an alternation of gradual
164 onshore movements during calms and episodic strong offshore movements during storms
165 (Van Enckevort and Ruessink, 2003, see also the right panel of figure 5). Finally,

166 degeneration occurs at the outer surf zone when the bar moves offshore of the breaker line
167 where weakly or non-breaking waves induce a net onshore sediment transport. This step
168 governs the entire cycle and only when the offshore bar degenerates at the breaker line the
169 nearshore bar can start its offshore migration (Wijnberg and Terwindt, 1995).

170
171

172 Figure 5. Left panel: seabed cross-shore profiles at Noordwijk, The Netherlands, from 1979
173 to 1987, showing cyclic bar behavior with a recurrence interval of about 4-5 years. Adapted
174 from van Enckvort, (2001). Right panel: seabed cross-shore profiles at Duck (U.S.A. Atlantic
175 coast, September-October 1994). Thick curves with red dates are the beginning of storms,
176 resulting in offshore bar migration (red arrows). Thin curves are the subsequent daily
177 profiles. The sandbar migrated onshore (blue arrows) during calm periods between the
178 storms. Courtesy of Dr. S. Elgar.

179
180

181 The second model is the Net Onshore Migration (NO_{on}M) model, which suggests that sand
182 bars are generated inshore of the breakpoint and migrate towards the swash zone (Aagard
183 et al., 2004). The observations made on the Danish North Sea coast suggest that a net
184 onshore bar migration occurs and the onshore bar movement in the mid-outer surf zone is
185 due to a dominance of the wave-induced sediment transport during storm condition. The
186 onshore-directed sediment transport is caused by incident wave skewnesses and enhanced
187 phase couplings between orbital velocity and sediment concentration at times of high energy
188 and water levels.

189 Alternative to both NO_{off}M and NO_{on}M, the Oscillation around a Position of Equilibrium
190 (OPE, Certain and Barousseau, 2005) proposes a cyclic oscillation of the bar position due to
191 seaward (storms) and shoreward (calms) migration. The mechanisms of this model are
192 essentially similar to those of the NO_{off}M, i.e. seaward migration during storms because of
193 the dominance of the undertow and recovery during calms due to the wave-induced onshore
194 sediment transport, but equilibrium is achieved when the local climate cannot not force storm
195 strong or long enough (e.g. Mediterranean vs. North Sea, see also Armaroli and Ciavola,
196 2011). Very recent observations (Parlagreco et al., 2019) show that Mediterranean storms

197 can indeed induce a NOffM rather than an OPE behavior and that the longshore flow can
198 sensibly contribute to increase the potential of sediment suspension.

199 Both NOffM and NOnM are preferential for longshore-uniform 2D bars, which evolve for
200 $\Omega \approx 5$ (see in the following), and seem to require large waves (Ω is directly proportional to H_b)
201 and longshore flows (likely contributing to longshore uniformity, Shand et al., 1999).
202 Moreover, since both NOffM and NOnM are observed to evolve over beaches with similar
203 slopes (1:200-1:150) and sediment sizes ($d_{50} \approx 150 \mu\text{m}$), the different behaviours are probably
204 to be searched into the different incident wave field (steepness and storm sequencing) and
205 possible physiographic constraints.

206 The above observations motivate a detailed analysis, using the mentioned classification,
207 of the models currently available to predict the generation and evolution of 2D sand bars.
208 These are typically distinguished between models for single but non-interacting bars and
209 interacting sand bars.

210 Among the models for single and non-interacting multiple bars we find the “empirical
211 models”, the “template/forced” models and the “stability models”.

212 The “empirical models” are based on use of the Dean parameter Ω (Wright and Short,
213 1984) and of the the $B^* = x_s / g \tan \beta T_r^2$ parameter (Short and Aagard, 1993), where x_s is the
214 nearshore width (offshore distance to the point where $\beta \approx 0$). Single bar formation is possible
215 for $20 < B^* < 50$ and $2 < \Omega < 6$, while $\Omega < 1$ characterizes reflective beaches and $\Omega > 6$ dissipative
216 beaches. For Ω increasing from 2 to 6 the following features emerge in sequence (Wijnberg
217 and Kroon, 2002): shore-attached bars, 3D bars and 2D bars (the latter ones for $\Omega \approx 5$).
218 Formation of multiple bars requires larger values of B^* (e.g. 2 bars for $50 < B^* < 100$, 3 bars
219 for $100 < B^* < 300$, etc.);

220 The “template/forced models” not only give synthetic information on conditions, but also
221 clarify some of the mechanisms for bar formation. Most of them propose bar formation via
222 sediment accumulation at specific spatial locations. Such locations can be (see also figure

223 6): the anti/nodes of standing waves - be they induced by incoming/outgoing free IG waves
224 (Short, 1975) or edge waves from opposite directions (Holman and Bowen, 1982), this is
225 known as the “standing IG wave mechanism”; the breakpoint location, where flow/sediment
226 convergence is due to the seaward-flowing undertow and the shoreward bottom boundary
227 layer (BBL) streaming, this is known as the “breakpoint-bar mechanism”.

228

229

230 Figure 6. Bar formation because of sediment convergence at: anti/nodes of standing IG
231 waves (left panel), breakpoint (right panel).

232

233

234 Another forced model, by O’Hare and Huntley (1994), suggests bar formation is induced by
235 sediment stirring from sea/short waves and transport by free IG, hence we call this “sea-IG
236 waves correlation mechanism”. Erosion/deposition is function of a fixed phase shift between
237 the stirring given by short-wave envelope and IG waves. If IG are Bound Long Waves (BLW,
238 more efficient on dissipative/mild beaches, Longuet-Higgins and Stewart, 1964), bar
239 formation is inhibited, while if IG are generated by BreakPoint (BP, more efficient on
240 reflective/steep beaches, Symonds et al., 1982) oscillation, bar formation is fostered.

241 Forced models represent a major improvement over empirical models but are affected by
242 some significant weaknesses. The most important is their intrinsic inability to give account of
243 the morphology-on-flow- feedback (term III of equation 1 is automatically set to zero).
244 Moreover, standing IG waves and breakpoint-bar mechanisms: prescribe the simultaneous
245 emergence of all bars across the surf zone (this is only observed for rhythmic 3D bars, but
246 not for 2D bars); require very specific wave conditions (e.g. narrow-banded spectra, mode
247 selection, etc.), which are not frequent in nature.

248 Finally, “stability models” have been successfully used to describe the generation of
249 crescentic or lunate 3D bars (Vittori et al., 1999) and the evolution of a double system of
250 crescentic 3D bars (Klein and Schuttelaars, 2006). Crescentic bars emerge because of
251 steady currents caused by the interaction of the incoming sea/short waves with synchronous

252 edge waves excited by the incoming waves moving over the wavy seabed. Klein and
253 Schuttelaars (2006) studied a similar problem using a numerical, wave-averaged model
254 (NSWE+Exner). Starting from an equilibrium state and numerically added perturbations the
255 fastest-growing perturbation was computed (see also the left panel of figure 7).

256
257

258 Figure 7. Left panel: nonlinear stability solution for a double sand bar system giving
259 asymmetric crescentic patterns on inner and outer bars (adapted from Klein and
260 Schuttelaars, 2006). Right panel: Evolution of longshore uniform sand bar after 30 days of
261 model run with normal wave incidence. Longshore instability in the form of rip channels
262 (adapted from Dronen and Deigaard, 2007).

263
264
265

266 However, no study seems available on the emergence of 2D sandbars by means of
267 stability models. Using a wave-averaged model Dronen and Deigaard, (2007) have shown
268 that longshore instabilities of an initially 2D bar emerge for both normal and oblique wave
269 incidence (see also the right panel of figure 7). This inadequacy of stability models in
270 reproducing the emergence and evolution of 2D sandbars calls for intensive investigation,
271 which should also take into proper account the role of longshore currents, which are thought
272 to be fundamental to give longshore uniformity.

273 With reference to models for interacting multiple bars, Castelle et al. (2010) used a
274 numerical, wave-averaged circulation model (with wave action equation for wave driver) plus
275 Exner equation to investigate the evolution of double crescentic sandbar systems.

275
276

277 Figure 8. Computed (top-right and bottom panels) 180° out-of-phase coupled patterns with
278 an outer-bar horn facing an inner-bar bay, similar to observations (top-left panel) on a meso-
279 macro-tidal, high-energy, double-barred beach (adapted from Castelle et al., 2010).

280
281
282

283 Simulations were run by providing a template in the morphology (inner bar longshore
284 uniform and outer-bar with crescentic patterns). The computed feedback between flow,
285 sediment transport and the evolving morphology allowed to reproduce morphologies similar
286 to those of observed double crescentic sandbar systems (see figure 8). The Authors

286 advanced that a novel mechanism (“morphological coupling”) was introduced that “blurs the
287 distinction between self-organization and template mechanisms”. However, inspection of
288 equation (1) suggests that such a novel mechanism is a complete model that retains all
289 terms on the right hand side. In other words, a template of crescentic outer bar is forced
290 (term I of equation 1) in a self-interating model (term III of equation 1).

291 The above consolidated knowledge has opened the way to some recent findings on:
292 sediment transport due to IG waves, sandbar migration and seabed moulding due to large-
293 scale vortices.

294 Recent results on IG wave-induced sediment transport come from De Bakker et al. (2016)
295 who have used field data collected in the Netherlands at the gently sloping ($\beta \approx 1:80$) Ballum
296 beach and at the moderately sloping ($\beta \approx 1:35$) SandMotor beach to investigate the role of
297 the beach slope on IG wave-induced cross-shore sediment transport (q_{IG}). Based on the
298 fundamental idea that sediment is put into suspension by sea/short waves and transported
299 by IG waves (see also the “sea-IG waves correlation mechanism” by O’Hare and Huntley,
300 1994), the study compares the performances of two proxies for sediment transport in the
301 nearshore. The first proxy is the correlation between the group bound IG wave (BLW) and
302 the sea-wave envelope (see also Baldock, 2006):

303
$$R_{Uu_{IG}}(\tau) = \frac{\langle U(t)u_{IG}(t+\tau) \rangle}{\sigma_U \sigma_{u_{IG}}} \quad (2)$$

304 where u_{IG} is the IG wave cross-shore velocity, U is the sea-wave envelope cross-shore
305 velocity, σ_s give the related variances and $\langle \rangle$ denotes time averaging. The second proxy,
306 proposed by De Bakker et al. (2016), is the IG wave height to sea wave height ratio H_{IG}/H_{SW} .

307 De Bakker et al. (2016) suggest that $R_{Uu_{IG}}$ is a good proxy for the nearshore sediment
308 transport only on moderately-sloping beaches while H_{IG}/H_{SW} well represents the sediment
309 transport on all beaches. In fact, while over moderate slopes $R_{Uu_{IG}} \neq 0$, it is $R_{Uu_{IG}} = 0$ on

310 mild slopes. On the other hand, H_{IG}/H_{SW} is always positive and increasing for mild slopes,
311 because IG waves receive more energy from sea waves breaking on wide surf zones.

312 In the shoaling region it is $R_{Uu_{IG}} < 0$ because the setdown (negative BLW) is associated
313 with the peak of the sea wave envelope and the sediment flux is negative, i.e. seaward (see
314 also figure 9). In the inner surf zone of moderately-sloping beaches the correlation becomes
315 positive because breaking-induced wave setup (positive BLW and shoreward sediment flux)
316 gives deeper waters and allows the faster sea waves to ride the crest of the IG wave (see
317 figure 9). However, in the inner-surf zone of mildly-sloping beaches it is $R_{Uu_{IG}} = 0$ because
318 the sea waves have been almost completely dissipated and $q_{IG} < 0$ because the peaks of
319 suspended sediment transport (SST) correlate with $u_{IG} < 0$ and the undertow \bar{u} .

320

321

322 Figure 9. Sketch of IG-sea wave interaction and cross-shore sediment transport in the
323 nearshore. From left to right: shoaling zone, inner-surf zone of moderately-sloping beaches,
324 inner-surf zone of mildly-sloping beaches.

325

326

327 Other recent results are on the mechanisms for onshore bar migration, fundamental for
328 all bar migration models of above. Early energetics-type models could only reproduce
329 offshore migration due to strong offshore currents (undertow) that are maximum at the bar
330 crest (Gallagher et al., 1998). Observations suggest that onshore migration occurs for weak
331 currents and significant waves. Hence, the role of waves is fundamental for the onshore bar
332 migration, and, as a consequence, it is fundamental to understand what are the model
333 representations of the main wave mechanisms for onshore bar migration.

334 With reference to the models analyzed in the Introduction we first analyze the energetics-
335 type models. They base their ability to reproduce onshore bar migration on use of the wave
336 acceleration skewness (Sk_a , proportional to the third order acceleration moment) in the
337 sediment transport closure, so that the sediment transport rate is written as (Hoefel and
338 Elgar, 2003):

339 $Q_A = K_a(a_{spike} - \text{sgn}(a_{spike})a_{crit}), \text{ for } |a_{spike}| \geq a_{crit} \text{ with } a_{spike} = \frac{\langle a^3 \rangle}{\langle a^2 \rangle} = \langle a^2 \rangle^{1/2} Sk_a \quad (3)$

340 This description is such to properly give account of the rapid onshore flow acceleration under
 341 steep wave fronts, which, in turn, generates strong horizontal pressure gradients on the
 342 seabed sediment.

343 On the other hand, process based, wave-resolving models (e.g. Henderson et al., 2004;
 344 Hsu et al., 2006) can properly predict onshore bar migration if they accurately describe the
 345 correlation between the orbital velocity (u) and the sediment concentration (c) in the BBL
 346 ($q = \langle uc \rangle$).

347
 348
 349 Figure 10. From left to right: vertical profile of mean cross-shore sediment transport due to
 350 waves (q_w) and currents (q_c), total cross-shore sediment transport as function of near-bed
 351 (middle) and freestream (right) velocity moments, for the Duck94 migration event. Adapted
 352 from Henderson et al. (2004).

353
 354
 355 Figure 10 shows that onshore migration occurs when the total sediment transport (q_t) is
 356 essentially due to waves (q_w) and correlates well with the third-order moment of the nearbed
 357 velocity ($\langle u_b^3 \rangle$).

358 The apparent distance between the two models can be reconciled, showing that energetics-
 359 type models are successful if they reproduce, at the integral level, the proper sediment flux-
 360 nearbed velocity moments correlation given by advective BBL models. The proof can be
 361 summarized as follows. First, it is shown that onshore bar migration occurs when q
 362 correlates well with $\langle u_b^3 \rangle$ in BBL, i.e. with the nearbed wave velocity skewness $Sk_{u_b} =$
 363 $\langle u_b^3 \rangle / \langle u_b^2 \rangle^{3/2}$. Then, using a Fourier decomposition of the velocity field, it is shown that $Sk_{u_b} \propto$
 364 $\sin(\phi_b) A_{u_f}$, where A_{u_f} is the wave freestream velocity asymmetry and ϕ_b is the nearbed-
 365 over-freestream velocity phase lead (Henderson et al., 2004; Berni et al., 2013). Finally, for
 366 sawtooth bores, i.e. setting $\phi = \pi/2$ in the modal solution $u_f \sim \sum_n \frac{1}{2^n} \sin[(n+1)\omega t + n\phi]$

367 gives $A_{u_f} \sim Sk_{a_f}$, with Sk_{a_f} wave freestream acceleration skewness (Drake and Calantoni,
368 2001) of energetics-type models.

369 Very recently Fernandez-Mora et al., (2015) used an energetics-type model to show the
370 ability of such models to predict onshore/offshore bar migration and assess the role of both
371 Sk_{u_b} and Sk_{a_f} . A complex model was built such that:

$$372 \quad (1 - n) \frac{\partial z_b(x,t)}{\partial t} = - \frac{\partial Q(x,t)}{\partial x} \quad (4)$$

$$373 \quad \text{with } Q = \alpha_c Q_c (Sk_{u_c}) + \alpha_u Q_u (Sk_{u_f}) + \alpha_a Q_a (Sk_{a_f}) + \alpha_D Q_D \left(\frac{dz_b}{dx} \right)$$

374 and where Sk_{u_c} is the skewness of the currents velocity vector and dz_b/dx is the, stabilizing,
375 seabed slope term. Hence, if α_u is set to zero the SkA model of Hoefel and Elgar (2002) is
376 obtained, if α_a is set to zero the SkV model of Hsu et al. (2006), while the complete model
377 is the MiX model. The model performances are illustrated in figure 11, always referring to
378 the Duck94 onshore migration event.

379
380
381 Figure 11. Modeling (red line) of the morphological evolution of the Duck94 onshore
382 migration event for: (A) the SkV model, (B) the SkA model, and (C) the MiX model. The initial
383 and final measured profiles are shown by dashed and solid lines, respectively. Adapted from
384 Fernandez-Mora et al. (2015).

385
386
387 All models reproduce well the chosen event, the SkV model better in the shoaling zone
388 and the SkA model better in inner surf zone. However, they all perform poorly in the swash
389 zone, this suggesting further investigation is needed for such region.

390 Among the most recent results on seabed evolution and bar generation are those
391 dedicated to the roles of macrovortices. Breakers of finite longshore width are known to
392 introduce vertical vorticity ζ that rolls up to form large-scale vortices with vertical axes, also
393 called macrovortices (Peregrine, 1999). Important dynamics characterized by finite
394 longshore breakers are those giving rise to rip currents (e.g. Kennedy et al., 2006) and cross-
395 sea induced breakers (e.g. Postacchini et al., 2014).

396 The latter condition is seen to be a potentially important cause of seabed morphology
397 alteration with the generation of complex 3D sandbars. The example of figure 12 has been
398 obtained by computing, using the NSW morphodynamic model of Postacchini et al. (2012),
399 the morphological evolution of a planar beach under the combined action of waves and
400 macrovortices generated by finite longshore breakers due to regular waves ($H_i=1\text{m}$, $T_i=10\text{s}$,
401 $\beta=1/30$) crossing at an angle of 15° to the coastline.

402
403

404 Figure 12. Seabed evolution forced by sea waves and macrovortices over a planar beach.
405 Red thin lines give contours of the free surface elevation, while black and blue solid lines
406 are contours of clockwise and anticlockwise vorticity, respectively. Yellow gives sand
407 deposition and green seabed erosion.

408
409

410 Sandbars of 3D shape (yellow colours in figure 12) arise due to the interaction of cross-sea
411 waves and the breaking-induced macrovortices.

412

413 River mouths

414 The hydro-morphodynamics evolving at river mouths, significantly related with the evolution
415 of sandbars, is analyzed, as a second applicative topic, with the same approach used for
416 sandbars: consolidated observations and models first, followed by more recent findings.

417 In view of the main aim of the paper, i.e. to highlight interactions amongst the subsystems
418 of interest, fundamentals of river mouth dynamics are kept at the minimum, with a preference
419 for those of wave-dominated rivers debouching into sandy coasts.

420 Analysis of the literature on river mouths dynamics is fairly complex because of the
421 different perspectives taken by the various communities that study river mouths, i.e.
422 engineers, geologists, ecologists, etc. Also, the vocabulary is influenced by such different
423 views and much effort is being spent to provide a systematic description of river mouth
424 dynamics.

425 The most important efforts have been made in the classification of river deltas (Galloway,
 426 1975), where delta is here generically used as “sedimentary deposit due to river-sea
 427 interaction and protruding into the sea”. The triangular classification of deltaic depositional
 428 systems by Galloway (1975) illustrates how the relative importance of sediment input, wave
 429 energy flux and tidal energy determines both morphology and stratigraphy of the delta.
 430 Indices have also been introduced to classify deltas. Among them we find the discharge
 431 index I_q of very consolidated use (Wright et al., 1974) and the more recent fluvial dominance
 432 ratio R (Nienhuis et al., 2015):

$$433 \quad I_q = \text{discharge index} = \frac{\text{river liquid discharge per unit width of mouth}}{\text{wave power per unit width of wave crest}} \quad (5a)$$

$$434 \quad R = \text{fluvial dominance ratio} = \frac{Q_r}{Q_{s,max}} = \frac{\text{river sand flux}}{\text{max longshore sediment flux at river mouth}}$$

435 (5b)

436 The former is a dimensional index (units of $\text{m}^3/\text{Watts/s}$) and such to increase of orders of
 437 magnitude from wave-dominated ($I_q \approx 10^{-5} \text{ m}^3/\text{Watts/s}$) to river-dominated ($I_q \approx 1 \text{ m}^3/\text{Watts/s}$)
 438 systems, while the latter is a dimensionless index, also increasing from $R < 1$ for wave-
 439 dominated to $R > 1$ for river dominated systems (see also figure 13).

440
 441
 442 Figure 13. Classification of river delta morphologies by means of the fluvial dominance ratio.
 443 Adapted from Nienhuis (2016).
 444
 445

446 Reference classifications of river deltas (Galloway, 1975) also clarify use of the word
 447 “estuary”, this being attributed to tide-dominated deltas, though at times more broadly
 448 (improperly?) used for semi-enclosed regions where river fresh water meets the sea. A
 449 number of schemes to classify estuaries is available in the literature, a recent one being that
 450 of Valle-Levinson (2011), which provides different classifications based on geomorphology,
 451 salinity structure and hydrodynamics.

452 Because of the ambiguities in using “estuary” and of the present focus on river systems
453 weakly influenced by tides, “river mouth” is here used as a clear and unambiguous definition
454 of the region where the river meets the sea. Analyses of wave-dominated river mouths are
455 also available (e.g. Cooper, 2001), and the very comprehensive, recent, work by Anthony
456 (2015) provides a nice overview of the role of waves at river mouths: waves remove fine
457 sediments at the river mouth and flatten the delta by moving the sediment alongshore.
458 Hence, deltas, placed along smooth coasts with shore-parallel ridges (see figure 14), are
459 shaped from arcuate to cusped. Waves that approach the coast obliquely generate: littoral
460 (i.e. longshore) sediment transport, maximum for 45° wave attack; an important difference
461 between updrift/downdrift flank transport and sediments (updrift flank made by beach
462 sediments, downdrift flank made by river sediments); an asymmetrical delta. A recent
463 quantitative analysis on delta progradation/migration is that of Nienhuis et al. (2016), which
464 can be summarized as follows. The delta undergoes: a symmetric progradation, for small
465 river sand flux and all longshore transported sediment bypassing the mouth; a downdrift
466 migration, for small river sand flux and vanishing sediment bypassing at mouth; an updrift
467 migration, for large river sand flux, independently of sediment bypassing at mouth.

468
469
470 Figure 14. River delta geomorphic units and schematic shoreline morphology (wave-
471 dominated morphology circled in red). Adapted from Anthony (2015).
472

473
474 Being the main aim of the present work that of clarifying the interactions - through nonlocal
475 agents - between the sandbar, river mouth and swash zone subsystems, most of the
476 attention is here given to local and nonlocal wave-forced dynamics.

477 Fundamental wave-forced, river mouth dynamics are the interaction of river current with
478 the sea waves and the related sediment transport.

479 Seaward flowing currents interact with incoming waves (Chawla and Kirby, 2002) to: slow
480 them down and even block them if the depth-uniform current velocity is equal and opposite

481 to the wave group velocity, this is known as “wave blocking”; steepen and even lead to
482 breaking; decrease the high-frequency content by wave breaking and transfer part of the
483 current energy to lower wave frequencies, this leading to frequency downshift of the
484 incoming waves;

485 The bedload accumulation related to the above dynamics typically leads to the formation
486 of a mouth sandbar. The intensity of the wave field significantly influences the bar formation
487 process. With no waves, a triangular mouth bar is generated just seaward of the mouth, at
488 a distance of about twice the mouth width. The generation process is initiated by the river
489 jet expansion in the sea, which, by continuity arguments, leads to a flow reduction and a
490 consequent sediment deposition (e.g. Fagherazzi et al., 2015). On the other hand, small
491 waves promote mouth bar formation by a process similar to the above: they promote the
492 river jet expansion and flow reduction, with a consequent more intense deposition close to
493 the river mouth. Finally, large waves suppress mouth bar formation because of their intense
494 seabed erosion and longshore transport (Fagherazzi et al., 2015).

495 Analysis of the above-mentioned wave-current interactions reveals that a river mouth acts
496 as a low-pass filter by removing (by breaking) sea and swell waves and letting long waves
497 pass. Recent studies have been dedicated to the propagation and evolution of IG waves at
498 inlets. Observations at the Ría de Santiuste (Spain, see Uncles et al., 2014) reveal how IG
499 waves (with periods of 4-5minutes), believed to be the outcome of estuary-amplified edge
500 waves, propagate upriver. This is well illustrated by the top panel of figure 15, which shows
501 a positive correlation between the water level (WL) and the flow velocity (U' in the figure).

502
503
504 Figure 15. IG wave propagation at river mouths. Top: positive water level-flow velocity
505 correlation at the Ría de Santiuste (Spain). Adapted from Uncles et al. (2014). Bottom: tide-
506 modulated IG wave propagation in the Albufeira Inlet (Portugal). No IG waves propagate
507 landward of the mouth at ebbs. Adapted from Bertin and Olabarrieta (2016).

508
509

510 The bottom panel of the same figure nicely illustrates the role of tide on the propagation
 511 of IG waves at inlets. Observations and wave-averaged modeling of the Albufeira Inlet
 512 (Portugal, see Bertin and Olabarrieta, 2016) reveal that upriver-propagation (from transect
 513 0 to transect 18 of images C-Run 3 and D-Run 4) of IG-waves only occurs during tidal flood.
 514 Inspection of the same images reveals that: no propagation is possible at ebb (white areas
 515 in correspondence of the ebbs), IG waves are equally generated by BLW (image C-Run 3)
 516 and BP (image D-Run 4) mechanisms.

517 Other recent field observation support the role of river mouths and the upriver propagation
 518 of IG waves. Among these, notable are the EsCoSed and MORSE project studies (Brocchini
 519 et al., 2017; Melito et al., 2018a, 2019) performed at the mouth of the Misa River (Italy, see
 520 figures 1 and 17), located at a microtidal site ($RTR < 3$) with a dissipative-to-intermediate
 521 beach ($5 < \Omega < 20$).

522
 523
 524 Figure 16. The Misa River estuary in Senigallia (Italy) depicting locations of measuring
 525 stations within the final river reach (TGdown, QR1, QR2, QR3, TGup) and in the sea (QS1,
 526 QS2, QS3). Adapted from Melito et al. (2018a).

527
 528
 529 On the basis of water elevation measurements collected at various locations both in the
 530 sea (QS in figure 17) and in the river (TG and QR in figure 17), the energy flux density dE_f
 531 and the energy flux E_f :

532
$$dE_f = S(f)C_g \quad dE_f = \int S(f)C_g df \quad (6)$$

533 have been computed - where $S(f)$ is the power spectral density, C_g the wave group velocity
 534 and f the frequency. The fundamental result is that while in the sea (QS1, QS2) most of the
 535 energy content is due to swell and sea waves, the IG band becomes dominant within the
 536 river (e.g. QR2 of the left panel of figure 17) with a weak upriver decay (TGup of the right
 537 panel of figure 17).

538

539
540
541
542
543
544

Figure 17. Energy density flux (left) and energy flux (right) at the Misa River during storm conditions. Adapted from Melito et al. (2018a).

545 The EsCoSed field campaign has also provided useful information on the seabed
546 morphology at the river mouth, highlighting the formation of large deposits/bars (blue in the
547 right panel of figure 18) and scours holes (red in the right panel of figure 18). The sand
548 deposits are found to be at a typical location, i.e. about twice the mouth width seaward of
549 the mouth itself (Fagherazzi et al., 2015), but characterized by a very complex planview
550 pattern.

551
552

553 Figure 18. Seabed variation between May-September 2013 (left) and between September
554 2013-February 2014 (right) at the Misa River. Adapted from Brocchini et al. (2017).

555
556
557

The above observations have motivated important efforts in the modeling of the dynamics
558 evolving at a river mouth characterized by a complex morphology, with a specific focus on
559 the role of waves.

560
561

562 Figure 19. Velocity maps for a strong river outflow opposed to three different wave regimes
563 (from top to bottom: $H_s = 0.0\text{m}$, 0.5 m , 1.5 m). Left column: results from Olabarrieta et al.
564 (2014). Right column: results from the wave-resolving solver of Brocchini et al. (2001),
565 averaged over 10 wave periods. Adapted from Melito et al. (2018b).

566
567

568 Olabarrieta et al. (2014) used the wave-averaged 3D Regional Ocean Modeling System
569 (ROMS) and an idealized river mouth shoal bathymetry to explore the wave-river current
570 interactions. The increased, with respect to a planar beach, steepening and breaking of
571 waves over the shoal leads to stronger seaward-flowing currents (undertow). Moreover,
572 waves are seen to increase both the lateral spreading of the river current and the longshore
573 transport.

574 In the attempt to better describe the dynamics forced by the sea waves, Melito et al.
575 (2018b) run the same cases studied by Olabarrieta et al. (2014) using the intra-wave NSWE
576 model of Brocchini et al. (2001), this neglecting vertical flow gradients, but improving the
577 time resolution of the wave forcing. The comparison between the two different numerical
578 solutions, illustrated in figure 19 for the case of a strong river outflow (1.1 m/s river current),
579 suggests that the main features of the wave-current interactions are captured, i.e. the river
580 current intensity and instability (e.g. Canestrelli et al., 2014). However, some significant
581 differences are also evident, among which: a weaker longshore spreading of the river jet,
582 the appearance of a number of large-scale eddies. Both differences seem to be caused by
583 the stronger wave breaking predicted by the NSWE, which removes significant amounts of
584 momentum and energy from the wave field, thus the reduced capacity of pushing shoreward
585 the river jet, in favour of the generation by differential wave breaking of large-scale eddies
586 or macrovortices (see previous subsection and Kennedy et al., 2006).

587 Such macrovortices not only contribute to the near-mouth flow mixing (right panels of
588 figure 19), but, as seen at the end of the previous subsection, also induce significant seabed
589 morphological changes. Falcini et al. (2014) recently sought for a mechanistic relationship
590 between potential vorticity Π ($\Pi = \zeta/d$ in a shallow-water framework) and sediment transport
591 at a river mouth. Lateral advection and diffusion of suspended sediment were found to be
592 directly proportional to the river jet vorticity, so that levees building is observed in
593 correspondence of regions of large vorticity and mouth bar deposition in regions of low
594 vorticity.

595 In view of the large similarities between the analyses of vorticity-induced dynamics within
596 the nearshore by Kennedy et al. (2006) and Brocchini (2013) and at river mouths by Falcini
597 et al. (2014), an integration of the two analytical approaches seems useful to clarify the role
598 of breaking-wave-induced vorticity on the river mouth morphodynamics.

599 In the meantime, some numerical experiments are being carried out, by means of the
600 hydro-morphodynamic NSW solver of Postacchini et al. (2012), to describe the main
601 morphological features that evolve at a simplified (e.g. shoal of Olabarrieta et al., 2014) and
602 natural (e.g. Misa River) river mouth bathymetry. Preliminary numerical results are illustrated
603 in figure 20, describing both a significant sediment deposition at the seaward edge of the
604 idealized shoal and a weaker deposition at the typical location of a mouth bar (left panel)
605 and a complex erosion deposition pattern at the mouth of the Misa River (right panel).

606
607
608
609
610
611

Figure 20. Estimated morphological evolution at the simplified river mouth bathymetry of Olabarrieta et al. (2014), left panel and at the Misa River, right panel.

612 Comparison of the right panels of figures 18 and 20 reveals that these preliminary
613 numerical solution can qualitatively capture the main characteristics of the complex
614 erosion/deposition pattern that evolves just seaward of the river mouth.

615

616 Swash zone

617 The swash zone is that region of the beach that is alternatively covered and uncovered by
618 water because of the action of sea waves (Brocchini and Baldock, 2008). Because of such
619 process, fundamental characteristics of the swash zone are: the flow intermittency; the
620 infiltration/exfiltration of water at the seabed; the very large sediment transport within the
621 swash, which can be quantified, for example, in terms of the cross-shore sediment mass per
622 beach longshore length transported by each swash event (Blenkissopp et al., 2011)

$$623 \quad -40 \text{ Kg/m} < \hat{q}_x < 40 \text{ Kg/m}, \quad \hat{q}_x \equiv \int_{T_{swash}}^{\square} q dt \quad (7)$$

624 or in terms of the longshore mass flux (Ribeiro et al., 2012).

$$625 \quad Q_y \sim (1 - 10) \text{ Kg/s} \quad (8)$$

626 The swash zone is also a locus for the generation of IG waves by two main mechanisms.

627 The first is the seaward reflection of BLW released by breaking (e.g., Schäffer, 1993).
 628 This is well illustrated in the left panel of figure 21, where a group of sea waves releases
 629 through breaking its BLW that propagates to the shoreline and it is there reflected to
 630 propagate out to sea as a Free Long Wave (FLW). The process of faster sea waves climbing
 631 the crest of the BLW is evident, as also illustrated in figure 9.

632 The second mechanism is the frequency downshifting of the sea waves that interact in
 633 the swash. This interaction occurs because waves in shallow water propagate with a velocity
 634 that is proportional to their height. Therefore, large waves travel faster than small waves and
 635 catch them up near or inside the swash zone (Mase, 1995). The assimilation of small waves
 636 by large waves leads to a reduced number of waves emerging from the swash zone, i.e. to
 637 a frequency downshifting, as illustrated in the right panel of figure 21.

638
 639 Figure 21. Left: example of BLW release on a beach because of sea wave breaking. The
 640 shoreward-propagating (from bottom to top) wave group releases its BLW that is reflected
 641 at the shoreline and propagated (from top to bottom) out to sea as a FLW (adapted from
 642 Watson, Barnes and Peregrine, 1995). Right: example of frequency downshifting within the
 643 swash zone. Bichromatic waves propagate from seaward to the shoreline (from top to
 644 bottom) over a steep beach and interact in the swash zone giving wave of doubled period
 645 (adapted from Mase, 1995).
 646
 647
 648

649 Most of the modelling of swash zone hydrodynamics, essentially forced by sea waves,
 650 has been done by using the NSW. For an horizontally-1D flow evolving in the crossshore
 651 direction x , these read (e.g. Brocchini et al., 2001):

$$652 \quad \begin{cases} d_{,t} + (ud)_{,x} = 0 \\ u_{,t} + uu_{,x} + gd_{,x} = gh_{,x} + c_f \frac{u|u|}{d} \end{cases} \quad (9)$$

653 where commas gives partial differentiation, d and h are the total and still-water depths,
 654 respectively, u is the depth-averaged flow velocity, g is gravity acceleration and c_f a
 655 dimensionless coefficient for seabed friction.

656 Solution of such equations not only provides a quantitative means for studying the
657 nearshore hydrodynamics, but can also be of help to illustrate some of the phenomena
658 above described, like, for example, the interaction of sea waves within the swash zone. An
659 example is provided by figure 22 for the case of a group of 5 sea waves.

660
661

662 Figure 22. Waves interacting at a wall (left) and at a swash zone (right). Top: incoming and
663 outgoing characteristic curves at the shoreline. Bottom: a group of 5 sea waves (thin line)
664 incoming to shore and outgoing IG waves (S-, thick line) resulting from the interaction of sea
665 waves (adapted from Brocchini, 2006).

666
667

668 On the basis of the NSWE a number of analytical solutions has been made available over
669 the years for both non-breaking and breaking waves. The pioneering work of Carrier and
670 Greenspan (1958), who used the method of hodograph transformation, opened the way to
671 some solutions that are currently used as reference for both analytical and numerical studies
672 for the evolution of non-breaking sea waves on a beach. In particular, Brocchini and
673 Peregrine (1996) extended the solution by Carrier and Greenspan (1958) to compute for the
674 longshore flow due to waves propagating almost orthogonally to the beach (see figure 23).
675 This solution can be used to provide estimates of the longshore mass flux to be compared
676 with field measurements, like that of equation (8). More recent studies, e.g. Antuono and
677 Brocchini (2010), suggest that analytical solutions for both non-breaking and breaking waves
678 can be achieved also in the physical space, i.e. avoiding use of the hodograph
679 transformation.

680
681

682 Figure 23. Left: maps, in the (x,t) plane of hodograph coordinates, free surface elevation η
683 crossshore (u) and longshore (v) velocities for two sea waves interacting onto a beach. Right:
684 mean longshore mass flux near and inside the swash zone for waves of different amplitude
685 (adapted from Brocchini and Peregrine, 1996).

686
687

688 Cornerstone of analytical solutions for breaking sea waves, i.e. bores, on a beach is that of
689 Shen and Meyer (1963), in the following SM63. A bore of velocity u_0 collapsing onto a beach
690 of slope β forces a swash that is shoreward bounded by a parabolic shoreline x_s :

691
$$x_s = u_0 t - \frac{1}{2} g \tan \beta t^2 \quad (10)$$

692 i.e. similar to the trajectory of a massive point under the action of gravity. This very neat
693 solution, which has been used for long times, has, however, the fundamental weakness of
694 predicting runup lenses much thinner than those observed on a natural beach.

695 Very recently Guard and Baldock (2007), in the following GB07, analyzed this issue and
696 suggested that the problem is caused by the assumption of regarding the collapsing bore
697 as a dam-break flow, where part of the fluid accelerates seaward. This is illustrated in the
698 sketch of the top panel of figure 24, whose left part illustrates the evolution of the SM63
699 solution in similarity with a dam-break flow (dashed line) that leads to both shoreward- and
700 seaward-accelerating fluid. The right image of the same panel illustrates how the GB07
701 solution, achieved by means of a linearly-time-varying (rather than constant) flow forcing at
702 collapse, predicts only shoreward-accelerating fluid after bore collapse. The bottom panel
703 of the same figure shows the fundamental differences in the water depths predicted by the
704 two solutions over a swash period. The SM63 solutions (thick lines) predicts more
705 asymmetric (downrush much longer than uprush) and much thinner (about a half) swash
706 lenses than those predicted by the GB07 solution (thin lines) that well compares with the
707 experimental data given by the symbols.

708 Recent remote sensing data, collected at 6 different beaches in USA and Australia, have
709 been used to assess the value and validity of the assumptions at the basis of the GB07
710 solutions (Power et al., 2011). Such observations confirm that: natural swash lenses are
711 deeper and less asymmetric than those predicted by SM63; a linearly-time-varying forcing
712 of the swash flow is better suited than a time-independent forcing to describe the evolution
713 of natural swashes.

714
715
716
717
718
719
720
721

Figure 24. Top: sketches of the flow by bore collapse. The left sketch shows the evolution of the SM63 solution in comparison to a dam-break flow (dashed lines), the right sketch suggests that only shoreward-accelerating fluid motion is predicted by the GB07 solution. Bottom: swash water thickness predicted by the SM63 solution (thick lines) and by the GB07 solution (thin lines) also compared with experimental data (symbols). Bottom panel adapted from GB07.

722
723
724

The fact that natural swash are fairly symmetric (still asymmetric in favour of the downrush) in time and characterized by fairly deep water lenses has important consequences on the understanding and modeling of the swash zone sediment transport and related morphodynamics. Thicker swash lenses lead to increased sediment pickup areas, because the sediment transport rate is directly proportional to the flow depth for a uniform sediment concentration (Pritchard and Hogg, 2005), and increased seaward sediment transport, because of the still existing swash asymmetry in favour of the rundown phase. However, such seaward sediment transport is counter-balanced by two main mechanisms: the presuspension of sediments due to the bore collapse onto the beach (which allows for a significant transport from the surf zone into the swash zone) and the settling lag effects due to the inertia of suspended sediment particles advected into the swash zone (Pritchard and Hogg, 2005). The above competing mechanisms, i.e. reduced swash asymmetry with increased thickness (promoting seaward transport) and sediment presuspension with settling lag effects (promoting shoreward transport) force a very complex swash zone morphology. This has been well described by Kelly and Dodd (2010), who implemented one of the most advanced and physically complete models for the swash zone hydro-morphodynamics. Some of their results reveal the complexities in beach profile changes due to a swash event (see figure 25).

742
743
744
745
746

Figure 25. Left: space-time map of the changes in beach profile due a swash event. The thick dashed lines give the zero contour. Right: snapshots of the swash lens cross-shore sections (adapted from Kelly and Dodd, 2010).

747

748

749 While beach degradation (light tones of grey) occurs virtually over all of the seaward
750 boundary of the swash zone, the most inshore portion of the swash zone undergoes
751 significant, though non-uniform, aggradation (dark tones of grey). Bed discontinuities, typical
752 of the evolution of an erosive bore, are seen to characterize both the uprush and the
753 downrush stages.

754 The increasing awareness of the fundamental role of the swash zone for the entire
755 nearshore hydro-morphodynamics stimulates always new research focused on links
756 between the surf zone and swash zone dynamics.

757 From the hydrodynamic viewpoint, one of the most interesting recent studies is that by
758 Moura and Baldock (2017), who focused on the relation between the shoreline (x_s) motion
759 and the IG waves, both BP forced (horizontal BP excursion x_{br}) and BLW (vertical BLW
760 excursion η_{BLW}), generated within the surf zone.

761

762

763 Figure 26. Shoreline relation with BLW and BP forcing. (a) Wave group. The vertical blue
764 line gives the BP between its outer (red), x_o , and inner (green), x_i location. Grey and red
765 lines give the released BLW and incident BP-forced IG wave, respectively. (b) BP excursion
766 (black) and shoreline response to BP-forced (red) and BLW (grey) IG waves. Dashed line
767 gives path for a shallow water wave from the BP to the shoreline. Horizontal colored lines
768 are BP positions as in Figure 2a. (c) Cross-correlation between BP and shoreline excursion,
769 BLW released (grey) and BP-foced IG wave (red). Adapted from Moura and Baldock, (2017).

770

771

772 Such relation can be illustrated with reference to the schematic of figure 26. When a group
773 arrives at breaking (panel a) the first, smaller waves break at x_i (green line) and the
774 subsequent, larger waves, more and more seaward, moving the BP to x_o (red line), the
775 subsequent smaller waves force the BP to move back to x_i . Hence, the BP position $x_{br}(t)$ is
776 – large (x_i), small(x_o), large (x_i) – i.e. in phase with the BLW vertical oscillation $\eta_{BLW}(t)$ (which
777 is large, small, large, i.e. concave upwards). The amplitude of the BP-forced IG wave $\eta_{BP}(t)$
778 is proportional to the width of the breaking region. Hence, it increases during the passage

779 from small to large waves ($x_{br}(t)$ from zero to max) and decreases from large to small waves
780 ($x_{br}(t)$ from max to zero). Hence, $\eta_{BP}(t)$ is concave downward or in phase opposition to $x_{br}(t)$
781 and $\eta_{BLW}(t)$. The released BLW and the incident BP-forced wave propagate with the same
782 speed (\sqrt{gh}) and reach the shoreline at the same time (after traveling through the surf zone),
783 but with opposite phase (panel b). This leads to a positive correlation between $x_{s,BLW}$ and x_{br}
784 and a negative correlation between $x_{s,BP}$ and x_{br} (panel c).

785 On the basis of the above and using data coming from pressure sensors (water depth,
786 significant wave height and period, etc.) and video observations (shoreline and BP
787 excursion) collected at 3 Australian beaches, correlations between IG waves and shoreline
788 excursion were made. The main findings suggest that: the shoreline excursion is a good
789 proxy for infragravity waves in the inner surf and swash zone, IG wave generation by BLW
790 release is stronger for conditions with relatively narrow surf zones and plunging waves while
791 BP forcing is dominant for wider surf zones and spilling breaker conditions. This seems in
792 contradiction with the established understanding (Battjes et. al, 2004) that the BP forcing is
793 most efficient on steep beaches and the BLW release is most efficient on mildly-sloping
794 beaches, see also the discussion provided in "Template models" of above. Such apparent
795 contradiction can, perhaps, be solved by recalling that the surf zone width is not only function
796 of the beach slope but also of the characteristics of the incoming wave. However, some
797 further research seems needed here.

798 From the morphological point of view, one fundamental link between the surf zone and
799 swash zone dynamics is represented by the feeding of the most inshore sand bar (see also
800 figures 4, 5, 11) by swash zone sediments. This has been recently investigated by Alsina et
801 al. (2012), on the basis of dedicated laboratory experiments. Such experiments have been
802 carried out by using flow conditions (surf zone characterized by plunging breakers, $\epsilon \approx 12-13$)
803 that, during an initial stage of run of about 3500 waves, led to the generation, from an almost-
804 planar beach, of a nearshore bar that slowly migrated off to sea (see blue-solid to green-

805 dashed lines, and vertical red lines in top panel of figure 27). The run was then stopped and
806 the beach profile manually reshaped to a milder slope only inside the swash zone (pink-solid
807 line). The same flow conditions were subsequently run, but bar migration stopped.

808
809
810 Figure 27. Cross-shore distribution of beach profile (top panel) and of cross-shore sediment
811 transport rate (bottom panel) at various times before and after beach reshaping (adapted
812 from Alsina et al., 2012).

813
814
815 This behavior was attributed to a reduced sediment transport from the swash to the surf
816 zone, because of the beach slope reduction forced in the swash zone. Such a conjecture
817 was also verified by analyzing the sediment transport rate illustrated in the bottom panel of
818 figure 27. This shows that, while before beach reshaping it was $Q_x < 0$, i.e. sediment transport
819 rate was seaward, in the surf zone and inner swash zone, the sediment transport rate
820 became vanishing to positive after beach reshaping. This is a clear indication of a major
821 reduction of bar sediment feeding from the swash zone.

822 The above observations can be interpreted on the basis of a simple mechanistic model:
823 the reduction of the swash zone beach slope (β_{swash} decreasing) forces an increase in the
824 natural swash period T_{swash} defined as (Baldock and Holmes, 1999):

825
$$T_{swash} = \frac{2C\sqrt{gH_b}}{g \sin \beta}, \quad C = O(1) \quad (12)$$

826 and, as a consequence, an increase in the swash-swash interactions, this being measured
827 by the T_{swash}/T_i ratio (Baldock and Holmes, 1999); such increased swash-swash interaction
828 makes the size of intense backwash events to decrease and, therefore, also the SST to sea,
829 i.e. to a reduced sediment input to the nearshore bar (Alsina et al., 2012).

830 This section is closed with some promising ongoing research aimed at including the
831 swash zone dynamics into wave-averaged solvers for the nearshore circulation (see also
832 “Generalities on Models” in the Introduction). By definition, wave-averaged models cannot
833 describe the swash zone dynamics associated to the sea waves, this leading to problems in

834 correctly predicting IG waves radiating out to sea. This is well illustrated by figure 22, whose
835 left panels describe the interaction of groups of 5 sea waves at a rigid wall used in wave-
836 averaged solvers to represent the shoreline, while the right panels describe the same
837 interaction at a swash zone reproduced by means of a wave-resolving model (in the specific
838 that of Brocchini et al., 2001). Two main features are very clear: the interaction at a wall,
839 being a mere reflection of the incoming wave groups, leads to outgoing groups with the
840 same structure of the incoming ones; the interaction at a swash zone allows for a significant
841 swash-swash interaction, for example because of large waves catching up and engulfing
842 small waves (see Mase, 1995 and figure 21), which contributes to increasing the intensity
843 (S^-) of the IG waves radiated out to sea (in the specific 20% larger for the interaction within
844 the swash zone). In view of the fundamental importance of wave-wave interaction within the
845 swash zone Brocchini and co-workers have designed some Shoreline Boundary Conditions
846 (SBCs) for the motion of the mean shoreline x_l and of the flow at at such boundary ($d(x_l)$,
847 $u(x_l)$) on the basis of an integral swash zone model (Brocchini and Peregrine, 1996;
848 Brocchini and Bellotti, 2002). Such SBCs have been successfully validated by means of
849 large-scale laboratory experiments (Bellotti et al., 2003) and have recently been
850 implemented in the ROMS wave-averaged solver (Haas and Warner, 2009). The main aim
851 of such preliminary implementation was to gauge the capabilities of reproducing swash zone
852 dynamics at a reduced computational cost, the final goal being that of predicting such
853 dynamics by simply running a wave-averaged solution. Figure 28 summarizes the main
854 findings of the mentioned implementation (Memmola, 2017; Memmola et al., 2019). The top
855 six panels give the shoreline motion computed by running ROMS alone at the unreasonably-
856 high resolution of 120 grid nodes (Δx) per offshore wavelength L_0 (black line), i.e. using it as
857 a wave-resolving model (benchmark solution), and ROMS coupled with the proposed SBCs
858 at lower and lower resolutions (red line), down to 2 grid nodes per wavelength. It is clear
859 that coupling ROMS with the SBCs allows one to run very cheap calculations that predict

860 reasonably well the benchmark solution by only using 3 grid nodes per wavelength (with $\Delta x/$
861 $L_0 \approx 2$ the solution is too deteriorated). The lower two panels of the same figure show that not
862 only the shoreline motion, but the entire flow field is well predicted by using the SBCs. The
863 left and right panels give, respectively, the swash zone free surface elevation and onshore
864 velocity for the benchmark solution (top panels) and ROMS+SBCs solutions for two very
865 coarse meshes (middle and lower panels).

866
867
868 Figure 28. Top 6 panels: comparison of the ROMS ultra-highly resolved shoreline (black line) and ROMS+SBCs shorelines for lower and lower grid resolution. Bottom 2 panels: free
869 surface elevation (left) and onshore velocity (right) for the benchmark ROMS solution (top
870 panels) and ROMS+SBCs solutions for two very coarse meshes (middle and lower panels).
871 Adapted from Memmola (2017).
872

873
874
875 These preliminary results demonstrate that the SBCs of Brocchini and co-worker can, at
876 least, allow for cheap solutions that also provide the entire structure of the swash zone flow.
877 Work is currently underway to achieve the more ambitious goal of running the ROMS-SBCs
878 system in a predicting mode (two-way coupled system), i.e. so that the SBCs are actually
879 used to time-step the swash solution to be returned to ROMS.

880

881 Discussion

882 An important contribution of the present work is the identification of nonlocal
883 forcing/dynamics, i.e. those that influence the entire nearshore as a whole. These are, in a
884 cascading order of mutual control the: beach slope (obvious nonlocal forcing), IG waves and
885 large-scale vortices with vertical axes (macrovortices).

886 The **beach slope** controls the role of IG waves, generation of macrovortices and the
887 related sediment transport. In particular, BLW are regarded to be more efficient on mildly-
888 sloping beaches, while IG waves are more likely generated by the oscillation of the BP on
889 steep beaches (Battjes et al., 2004); although some recent studies by Moura and Baldock

890 (2017) seem to provide different information. The generation of macrovortices is controlled
891 by the beach slope through the longshore differential breaking of sea waves that leads to
892 generation and reorganization of potential vorticity at the lateral edges of breaking wave
893 fronts (e.g. Kennedy et al. 2006). Although the role of SST at the lower boundary of the
894 swash zone can be regarded as a local effect (see above), changes in the swash zone
895 beach slope can significantly alter the swash-swash interactions and intense backwash
896 events, this either promoting or stopping the sediment transport needed for the formation
897 and migration of nearshore sandbars (Alsina et al., 2012).

898 **IG waves** provide a nonlocal forcing because the lengths involved in their generation and
899 evolution are comparable to the surf zone width. IG are thought to have a fundamental role
900 in the generation of sandbars, though various different models have been proposed like the
901 “standing IG wave mechanism” (e.g. Short, 1975) and the “sea-IG waves correlation
902 mechanism” of O’Hare and Huntley (1994) in which IG waves are responsible for the
903 transport of sediments put into suspension by sea waves. Hence, the importance of the
904 correlation between BLW and sea-wave envelope or the IG wave height to sea wave height
905 ratio H_{IG}/H_{SW} , which is a good proxy for the sediment transport on beaches of all slopes (De
906 Bakker et al., 2016). IG waves are also important for the riverine sediment transport because
907 they can propagate upriver (e.g. Uncles et al, 2014), while sea and swell waves are
908 dissipated by breaking at river mouths, which act as “low-pass filters”. IG waves are not only
909 generated at the seaward boundary of the surf zone, but also in the very shallow waters of
910 the swash zone. This occurs by the wave-wave interaction due to large waves traveling
911 faster than small waves and catching them up in the swash zone (Mase, 1995).

912 **Macrovortices** are a nonlocal forcing of the nearshore dynamics because they can
913 maintain their coherent over large distances after their generation. Macrovortices contribute
914 to sediment transport both by mobilizing the nearbed sands and, like IG waves, by carrying
915 the sediments put into suspension by sea waves. The differential breaking that generates

916 macrovortices is particularly intense at river mouths characterized by seabed shoals. Hence,
917 macrovortices, beyond the river jet expansion, provide a significant contribution to both flow
918 mixing and sediment transport at river mouths (Melito et al., 2018b).

919 The analysis of the three subsystems of interest and of the related wave-induced
920 dynamics suggests some lines for future research, proposed with reference to each of such
921 subsystems.

922 Inspection of the state-of-the-art sandbar modeling reveals that we are still in need of
923 suitable stability models for the generation of 2D sandbars. As recalled in the analysis of
924 sandbars, available numerical solutions for the generation of 2D sandbars are characterized
925 by longshore instabilities in the form of rip channels (Dronen and Deigaard, 2007) and no
926 stable 2D sandbar systems can actually be reproduced for either normal or oblique wave
927 incidence. It seems, therefore, essential to investigate the mechanisms for the emergence
928 of 3D instabilities, also inspecting the role of the longshore currents, which are believed to
929 contribute to the maintenance of the stability of a 2D pattern (Shand et al., 1999).

930 The role of longshore flows is becoming of growing interest, as believed to significantly
931 control not only the generation, but also the migration of sandbars. In particular, longshore
932 currents due to waves obliquely incident to the shore may have the potential to increase
933 the shear stress acting on the seabed and responsible for the sediment mobility (Parlagreco
934 et al., 2019). In this perspective longshore currents are not directly responsible for the
935 direction of cross-shore migration of sandbars (onshore vs offshore) but provide an indirect
936 important control through increased sediment mobility.

937 Observed opposite migration behaviours on beaches of similar steepness and sediment
938 size (Van Enckevort and Ruessink, 2003; Aagard et al., 2004) suggest that other important
939 mechanisms that may characterize conditions for offshore (NOffM model) vs onshore
940 (NOnM model) migration are the steepness and storm sequencing of the incident wave field.

941 In brief, future studies on sandbar generation and evolution should pay great attention to
942 **wave obliquity, steepness and storm sequencing.**

943 A more detailed inspection of the steepness of the incident wave field and related
944 modalities of breaking is also advisable to try to explain the apparent contradiction between
945 the recent findings of Moura and Baldock, (2017), suggesting that IG wave generation by
946 BLW release is stronger for narrow surf zones and plungers while BP forcing is dominant
947 for wider surf zones and spillers, and the established understanding (Battjes et. al, 2004)
948 that the BP forcing is most efficient on steep beaches and the BLW release is most efficient
949 on mildly-sloping beaches. This investigation should also try to encompass an analysis of
950 the different modes of transport of turbulence associated with the two mentioned breaking
951 types (seaward under spilling breakers and landward under plunging breakers – Ting and
952 Kirby, 1994).

953 The implementation of **novel Shoreline Boundary Conditions** into wave averaging
954 models for the nearshore circulation is certainly an important way forward for the inclusion
955 of swash zone dynamics in long-term numerical simulations of the nearshore dynamics. The
956 recent works of Memmola (2017) and Memmola et al. (2019) have shown the potentials of
957 this approach, which will enable accurate and relatively cheap long-term numerical
958 solutions. Work is currently underway for a predictive implementation of such approach.

959 In relation to the interactions occurring at a river mouth, studies are needed to clarify the
960 roles of the different wave modes acting there. Recent field observations by Melito et al.
961 (2019) confirm that **the tide has an important role of modulating the IG waves** that
962 propagate upstream the river, result already obtained through numerical means by
963 Olabarrieta (2016). This modulating effect, considerable even for microtidal environments
964 (e.g. Misa River), should be better analyzed through field studies carried out in macrotidal
965 environments.

966 The interactions of a river jet and the incoming sea waves should be better investigated
967 also in relation to the generation of macrovortices. Preliminary evidence shows that they can
968 remove significant amounts of energy from the incident wave field, reducing the opposing
969 action of waves to the river stream (Melito et al., 2018b), and also force 3D seabed
970 morphological features, e.g. mouth bars. Hence, both theoretical studies aimed at
971 integrating the analytical approach of Brocchini (2013) for vorticity and potential vorticity
972 generation with that of Falcini et al. (2014) as well as accurate morphological wave-resolving
973 numerical simulations will likely clarify the role of ***breaking-wave-induced vorticity on the***
974 ***river mouth morphodynamics***.

975

976 Conclusions

977 The role of waves on the nearshore hydro-morphodynamics has been analyzed with specific
978 focus to three subsystems of the nearshore i.e. sandbar region, river mouths and swash
979 zone.

980 Each of these is characterized by very specific local dynamics. For sandbar generation
981 and evolution fundamental is the role the nearbed sediment transport, with the correlation
982 between the orbital velocity and sediment concentration in the BBL being of paramount
983 importance for suitably predicting onshore bar migration (e.g. Henderson et al., 2004). It has
984 also been shown that energetics-type models can achieve similarly good results if they
985 reproduce such correlation through an integral approach, i.e. by including the effects of the
986 wave freestream acceleration skewness (e.g. Hoefel and Elgar, 2002).

987 Local dynamics specific to river mouths are: the wave-current interaction by which the
988 river current steepens the incoming waves, if intense enough to the point of breaking and/or
989 blocking them, and forces a frequency downshifting (Chawla and Kirby, 2002); the related
990 formation of a mouth bar (Fagherazzi et al., 2015) or the formation of a shoal that largely

991 influences the local dynamics by forcing stronger wave-breaking-induced return currents
992 (Olabarrieta et al., 2014) and wave breaking/sediment deposit farther from the mouth, i.e. at
993 the seaward edge of the shoal.

994 BBL dynamics largely influence the very shallow flows of the swash zone and, like for
995 sandbars, a proper modeling of the swash zone sediment transport can only be achieved if
996 correlation between the orbital velocity and sediment concentration in the BBL is well
997 described. Fundamental is also the description of the SST transport at the lower boundary
998 of the swash zone, which affects the sediment fluxes both from the surf zone, through
999 sediment presuspension at bore collapse (Pritchard and Hogg, 2005), and to the surf zone,
1000 in association to intense backwash events.

1001 A closure can be reached by recalling the main connections among the three subsystems
1002 of interest. IG waves provide a fundamental connection because generated both at the
1003 seaward edge of the surf zone (by BLW release, BP oscillation and edge wave interaction)
1004 and at the swash zone (because of wave-wave interaction) can propagate almost unaltered
1005 upstream at river mouths. Another important connection is provided by the evolution of
1006 macrovortices, which generated by differential breaking move over long distances in the surf
1007 zone and may reach the shoreline where they can rebound and alter the near-shoreline
1008 morphology. From the morphological viewpoint, obvious links are those between: the swash
1009 zone morphology and the along-river-flank sediment transport that can feed the swash zone
1010 and the swash zone and the nearshore sandbars, which can drive sediments for their
1011 evolution from the swash zone.

1012

1013 Acknowledgements

1014 This paper is essentially a written transposition of the same-titled invited talk given at
1015 RCEM2017. The organizers are thanked for the opportunity given to reflect on and present
1016 this material. The following colleagues and friends are thanked for providing both useful

1017 material and insightful discussions: Tom Baldock (University of Queensland, Australia),
1018 Riccardo Briganti (University of Nottingham, England), Joe Calantoni (Naval Research
1019 Laboratory, USA), Dan Conley (University of Plymouth, England), Steve Elgar (Woods Hole
1020 Oceanographic Institution, USA), Tom Hsu (University of Delaware, USA), Matteo
1021 Postacchini (Università Politecnica delle Marche, Italy), Gerben Ruessink (Utrecht
1022 University, The Netherlands), Alex Sheremet (University of Florida, USA). Financial support
1023 from the ONR Global (UK), through the MORSE Research Grant (N62909-17-1-2148) is
1024 gratefully acknowledged.

1025

1026 References

- 1027 Aagard, T., Davidson-Arnott, R., Greenwood, B., Nielsen, J. (2004). Sediment supply from
1028 shoreface to dunes: linking sediment transport measurements and long-term
1029 morphological evolution. *Geomorphology* **60**: 205-224.
- 1030 Alsina, J.M., Caceres, I., Brocchini, M., Baldock, T.E. (2012). An experimental study on
1031 sediment transport and bed evolution under different swash zone morphological
1032 conditions. *Coastal Engineering* **68**: 31-43.
- 1033 Anthony, E.J. (2015). Wave influence in the construction, shaping and destruction of river
1034 deltas: A review. *Marine Geology* **361**: 53-78.
- 1035 Antuono, M., Brocchini, M. (2010). Solving the Nonlinear Shallow Water Equations in
1036 physical space. *Journal of Fluid Mechanics* **643**: 207-232.
- 1037 Armaroli, C., Ciavola, P. (2011). Dynamics of a nearshore bar system in the northern
1038 Adriatic: A video-based morphological classification. *Geomorphology* **126**: 201-216.
- 1039 Ashton, A.D., Giosan, L. (2011). Wave-angle control of delta evolution. *Geophysical*
1040 *Research Letters* **38**: L13405.
- 1041 Bailard, J.A. (1981). An energetics total load sediment transport model for a plane sloping
1042 beach. *Journal of Geophysical Research – Oceans* **86(C11)**: 10938-10954.

1043 Baldock, T.E. (2006). Long wave generation by the shoaling and breaking of transient wave
1044 groups on a beach. *Proceedings of the Royal Society of London A* **462**: 1853-1876.

1045 Baldock, T.E., Holmes, P. (1999). Simulation and prediction of swash oscillations on a steep
1046 beach. *Coastal Engineering* **36 (3)**: 219–242.

1047 Battjes, J.A., Bakkenes, H.J., Janssen, T.T., van Dongeren, A.R. (2004). Shoaling of
1048 subharmonic gravity waves. *Journal of Geophysical Research - Oceans* **109**: C02009.

1049 Bellotti, G., Archetti, R., Brocchini, M. (2003). Experimental validation of mean swash zone
1050 boundary conditions, *Journal of Geophysical Research - Oceans* **108(C8)**: 3250,
1051 doi:10.1029/2002JC001510.

1052 Berni, C., Barthelemy, E., Michallet, H. (2013). Surf zone cross-shore boundary layer
1053 velocity asymmetry and skewness: an experimental study on a mobile bed. *Journal of*
1054 *Geophysical Research - Oceans* **118**: 2188–2200.

1055 Bertin, X., Olabarrieta, M. (2015). Relevance of infragravity waves in a wave-dominated
1056 inlet. *Journal of Geophysical Research - Oceans* **121**: 5418–5435.

1057 Blenkinsopp, C.E., Turner, I.L., Masselink, G., Russell, P.E. (2011). Swash zone sediment
1058 fluxes — field observations. *Coastal Engineering* **58**: 28–44.

1059 Brocchini, M. (2006). Integral swash zone models. *Continental Shelf Research* **26**: 653-660.

1060 Brocchini, M. (2013). A reasoned overview on Boussinesq-type models: the interplay
1061 between physics, mathematics and numerics. *Proceedings of the Royal Society of*
1062 *London A* **469**: 20130496, doi:10.1098/rspa.2013.0496.

1063 Brocchini, M., Peregrine, D.H. (1996). Integral flow properties of the swash zone and
1064 averaging. *Journal of Fluid Mechanics* **317**: 241-273.

1065 Brocchini, B., Bellotti, G. (2002). Integral flow properties of the swash zone and averaging.
1066 Part 2. The shoreline boundary conditions for wave-averaged models. *Journal of Fluid*
1067 *Mechanics* **458**: 269-281.

1068 Brocchini, M., Baldock, T.E. (2008). Recent advances in modeling swash zone dynamics:
1069 influence of surf-swash interaction on nearshore hydrodynamics and morphodynamics.
1070 *Reviews in Geophysics* **46(3)**: RG3003, doi:10.1029/2006RG000215.

1071 Brocchini, M., Calantoni, J., Postacchini, M., Sheremet, A., Staples, T., Smith, J., Reed,
1072 A.H., Braithwaite III, E.F., Lorenzoni, C., Russo, A., Corvaro, S., Mancinelli, A. and
1073 Soldini, L. (2017). Comparison between the wintertime and summertime dynamics of the
1074 Misa River estuary. *Marine Geology* **385**: 27-40.

1075 Caballeria, M., Coco, G., Falqués, A., Huntley, D.A., 2002. Self-organization mechanisms
1076 for the formation of nearshore crescentic sand bars. *Journal of Fluid Mechanics* **465**: 379–
1077 410.

1078 Canestrelli, A., Nardin, W., Edmonds, D., Fagherazzi, S., Slingerland, R. (2014), Importance
1079 of frictional effects and jet instability on the morphodynamics of river mouth bars and
1080 levees, *Journal of Geophysical Research - Oceans* **119**: 509–522.

1081 Certain, R., Barousseau, J.P. (2005). Conceptual modelling of sand bars morphodynamics
1082 for a microtidal beach (Sète, France). *Bulletin de la Société Géologique de France* **176**:
1083 343–354.

1084 Chang, Y.S., Scotti, A. (2004). Modeling unsteady turbulent flows over ripples: Reynolds-
1085 averaged Navier-Stokes equations (RANS) versus large-eddy simulation (LES). *Journal*
1086 *of Geophysical Research - Oceans* **109**: 1-16.

1087 Chawla, A., Kirby, J.T. (2002). Monochromatic and random wave breaking at blocking
1088 points. *Journal of Geophysical Research - Oceans* **107**: 3067.

1089 Coco G., Murray, A.B. (2007). Patterns in the sand: from forcing templates to self-
1090 organization. *Geomorphology* **91**: 271-290.

1091 Cooper, J.A.G. (2001). Geomorphological variability among microtidal estuaries from the
1092 wave-dominated South African coast. *Geomorphology* **40**: 99-122.

1093 de Bakker, A.T.M., Brinkkemper, J.A., van der Steen, F., Tissier, M.F.S., Ruessink B.G.
1094 (2016). Cross-shore sand transport by infragravity waves as a function of beach
1095 steepness. *Journal of Geophysical Research – Earth Science* **121**: 1786-1799.

1096 Drake, T.G., J. Calantoni (2001). Discrete particle model for sheet flow sediment transport
1097 in the nearshore. *Journal of Geophysical Research - Oceans* **106**: 19,859–19,868.

1098 Dronen, N., Deigaard, R. (2007). Quasi-three-dimensional modelling of the morphology of
1099 longshore bars. *Coastal Engineering* **54**: 197-215.

1100 van Enckevort, I.M.J., Ruessink, B.G. (2003). Video observations of nearshore bar
1101 behaviour. Part 1: alongshore uniform variability. *Continental Shelf Research* **23**: 501-
1102 512.

1103 Fagherazzi, S., Edmonds, D.A., Nardin, W., Leonardi, N., Canestrelli, A., Falcini, F.,
1104 Jerolmack, D.J., Mariotti, G., Rowland, J.C., Slingerland, R.L. (2015). Dynamics of river
1105 mouth deposits. *Reviews in Geophysics* **53**, 31pp., doi:10.1002/2014RG000451.

1106 Falcini, F., Piliouras, A., Garra, R., Guerin, A., Jerolmack, D.J., Rowland, J., Paola, C. (2014)
1107 Hydrodynamic and suspended sediment transport controls on river mouth morphology
1108 *Journal of Geophysical Research – Earth Surface* **119**: 1-11, doi:10.1002/2013JF002831.

1109 Gallagher, E., Elgar, S., Guza, R.T. (1998). Observations of sand bar evolution on a natural
1110 beach. *Journal of Geophysical Research - Oceans* **103(C2)**: 3203-3215.

1111 Galloway, W.E. (1975). Process framework for describing themorphologic and stratigraphic
1112 evolution of delta depositional systems. In: Broussard, M.L (Ed.), *Deltas: Models for*
1113 *Exploration*. Houston Geological Society, Houston, pp. 87–98.

1114 Guard, P., Baldock, T. (2007). The influence of seaward boundary conditions on swash zone
1115 hydrodynamics. *Coastal Engineering* **54**: 321–331.

1116 Haas, K.A. and Warner, J.C. (2009). Comparing a quasi-3D to a full 3D nearshore circulation
1117 model: SHORECIRC and ROMS. *Ocean Modelling* **26**: 91-103.

1118 Henderson, S.M., Allen, J.S., Newberger P.A. (2004). Nearshore sandbar migration
1119 predicted by an eddy-diffusive boundary layer model. *Journal of Geophysical Research -*
1120 *Oceans* **109**: C06024, doi:10.1029/2003JC002137.

1121 Hoefel, F., Elgar, S. (2003). Wave-induced sediment transport and sandbar migration.
1122 *Science* **299**: 1885-1887.

1123 Holman, R.A., Bowen, J. (1982). Bars, bumps, and holes: models for the generation of
1124 complex beach topography. *Journal of Geophysical Research - Oceans* **87**: 457-468.

1125 Hsu, T., Elgar, S., Guza, R.T. (2006). Wave-induced sediment transport and onshore
1126 sandbar migration. *Coastal Engineering* **53**: 817–824.

1127 Kennedy, A.B., Brocchini, M., Soldini, L., Gutierrez, E. (2006). Topographically controlled,
1128 breaking-wave-induced macrovortices. Part 2. Rip current topographies. *Journal of Fluid*
1129 *Mechanics* **559**: 57–80.

1130 Kirby, J.T. (2016). Boussinesq models and their application to coastal processes across a
1131 wide range of scales. *Journal of Waterways, Port, Coastal and Ocean Engineering* **142**:
1132 03116005.

1133 Lesser, G.R., Roelvink, J.A., van Kester, J.A.T.M., Stelling, G.S. (2004). Development and
1134 validation of a three-dimensional morphological model. *Coastal Engineering* **51**: 883-915.

1135 Longuet-Higgins, M. S., Stewart, R. W. (1964). Radiation stress in water waves: a physical
1136 discussion with applications. *Deep Sea Research I* **11**: 529-562.

1137 Lubin, P., Glockner, S., Kimmoun, O., Branger, H. (2011). Numerical study of the
1138 hydrodynamics of regular waves breaking over a sloping beach. *European Journal of*
1139 *Mechanics B/Fluids* **30**: 552-564.

1140 Ma, G., Shi, F., Kirby, J.T. (2012). Shock-capturing non-hydrostatic model for fully dispersive
1141 surface wave processes. *Ocean Modelling* **43-44**: 22-35.

1142 Mase, H. (1995). Frequency down-shift of swash oscillations compared to incident waves.
1143 *Journal of Hydraulic Research* **33**: 397-411.

1144 Masselink, G., Short, A.D. (1993). The Effect of Tide Range on Beach Morphodynamics and
1145 Morphology: A Conceptual Beach Model. *Journal of Coastal Research* **9**: 785-800.

1146 Melito, L., Postacchini, M., Sheremet, A., Calantoni, J., Zitti, G., Darvini, G., Brocchini, M.
1147 (2018a). Wave-current interactions and infragravity wave propagation at a microtidal inlet.
1148 *E-proceedings of the 3rd EWAS International Conference*: 477-485.

1149 Melito, L., Postacchini, G, Darvini, G., Brocchini, M. (2018b). Waves and currents at a river
1150 mouth: the role of macrovortices, sub-grid turbulence and seabed friction. *Water* **10**: 550,
1151 doi:10.3390/w10050550 .

1152 Melito, L., Postacchini, M., Sheremet, A., Calantoni, J., Zitti, G., Darvini G., Brocchini M.
1153 (2019l). Hydrodynamics at a microtidal inlet: analysis of propagation of the main wave
1154 components. *Estuarine Coastal and Shelf Science* (under peer review).

1155 Memmola, F. (2017). Analysis and development of oceanographic models: reaching the
1156 swash zone. Università Politecnica delle Marche, PhD Thesis, 92pp.

1157 Memmola, F., Coluccelli, A., Russo, A., Warner J.C., Brocchini M. (2019). Wave-resolving
1158 Shoreline Boundary Conditions for wave-averaged coastal models. *Ocean Modeling*
1159 (under peer review).

1160 Moura, T., Baldock, T.E. (2017). Remote sensing of the correlation between breakpoint
1161 oscillations and infragravity waves in the surf and swash zone. *Journal of Geophysical*
1162 *Research - Oceans* **122**: 3106-3122.

1163 Nienhuis, J.H. (2016). Plan-view evolution of wave-dominated deltas. Massachusetts
1164 Institute of Technology and the Woods Hole Oceanographic Institution, PhD Thesis,
1165 182pp.

1166 Nienhuis, J.H., Ashton, A.D., Giosan, L. (2015). What makes a delta wave-dominated?
1167 *Geology* **43(6)**: 511-514.

1168 Nienhuis, J.H., Ashton, A.D., Giosan, L. (2016). Littoral steering of deltaic channels. *Earth*
1169 *and Planetary Science Letters* **453(1)**: 204-214.

1170 O'Hare, T.J., Huntley, D.A. (1994). Bar formation due to wave groups and associated long
1171 waves. *Marine Geology* **116**: 313-325.

1172 Olabarrieta, M., Geyer, W.R., Kumar, N. (2014). The role of morphology and wave-current
1173 interaction at tidal inlets: An idealized modeling analysis. *Journal of Geophysical*
1174 *Research – Oceans* **119**: 8818-8837.

1175 Parlagreco, L. Melito, L. Devoti, S. Perugini, E. Soldini, L. Zitti G., Brocchini M. (2019).
1176 Monitoring for coastal resilience: preliminary data from five Italian sandy beaches. **19**.
1177 1854; doi:10.3390/s19081854

1178 Peregrine, D.H. (1999). Large-scale vorticity generation by breakers in shallow and deep
1179 water. *European Journal of Mechanics B/Fluids* **18**: 403–408.

1180 Plant, N.G., Freilich, M.H., Holman, R.A. (2001). Role of morphologic feedback in surf zones
1181 and bar response. *Journal of Geophysical Research - Oceans* **106**: 973-989.

1182 Postacchini, M., Brocchini, M., Landon, M., Mancinelli, A. (2012). A multi-purpose, intra-
1183 wave, shallow water hydro-morphodynamic solver. *Advances in Water Resources* **38**: 13-
1184 26.

1185 Postacchini, M., Brocchini, M., Soldini, L. (2014). Vorticity generation due to cross-sea.
1186 *Journal of Fluid Mechanics* **744**: 286-309.

1187 Pourzangbar, A., Losada, M.A., Saber, A., Rasoul Ahari, L., Larroudé, P., Vaezi, M. and
1188 Brocchini, M. (2017). Prediction of non-breaking wave induced scour depth at the trunk
1189 section of breakwaters using Genetic Programming and Artificial Neural Networks.
1190 *Coastal Engineering* **121**: 107-118.

1191 Power, H.E., Holman, R.A., Baldock, T.E. (2011). Swash zone boundary conditions derived
1192 from optical remote sensing of swash zone flow patterns. *Journal of Geophysical*
1193 *Research - Oceans* **116**: C06007, doi:10.1029/2010JC006724.

1194 Pritchard, D., Hogg, A.J. (2005). On the transport of suspended sediment by a swash event
1195 on a plane beach, *Coastal Engineering* **52(1)**: 1–23.

1196 Ribas, F. (2003). On the growth of nearshore sand bars as instability processes of
1197 equilibrium beach states. Universitat Politecnica de Catalunya, PhD Thesis, 242pp.

1198 Ribeiro, M., Barrio, F., Taborda, R., Cascalho, J., Bosnic, I. Rodríguez, I., Sanchez, M.
1199 (2012). Longshore drift: experimental and empirical predictors. The example of the Ebro
1200 Delta. *Proceedings of the 2as Jornadas de Engenharia Hidrográfica*, Lisboa, Portugal,
1201 pp. 271–274.

1202 Roelvink, J. A., Stive, M. J. F. (1989) Bar-generating cross-shore flow mechanisms on a
1203 beach. *Journal of Geophysical Research - Oceans* **94**: 4785-4800.

1204 Ruessink, B.G., Kroon, A. (1994). The behaviour of a multiple bar system in the nearshore
1205 zone of Terschelling, the Netherlands: 1965–1993. *Marine Geology* **121**: 187–197.

1206 Schäffer, H.A. (1993). Infragravity waves induced by short wave groups. *Journal of Fluid*
1207 *Mechanics* **247**: 551–588.

1208 Shand, R.D., Bailey, D.G., Shephard, M.J. (1999). An inter-site comparison of net offshore
1209 bar migration characteristics and environmental conditions. *Journal of Coastal Research*
1210 **15**: 750–765.

1211 Shen, M.C., Meyer, E.R. (1963). Climb of a bore on a beach. Part 3. Run-up. *Journal of*
1212 *Fluid Mechanics* **16**: 113–125.

1213 Short, A.D. (1975). Multiple offshore bars and standing waves. *Journal of Geophysical*
1214 *Research - Oceans* **80**: 3838-3840.

1215 Short, A.D., Aagard, T. (1993). Single and multi-bar beach change models. *Journal of*
1216 *Coastal Research* **15**: 141-157.

1217 Symonds, G., Huntley, D.A., Bowen, A.J. (1982). Two-dimensional surf beat: long wave
1218 generation by a time-varying breakpoint. *Journal of Geophysical Research - Oceans* **87**:
1219 492-498.

1220 Ting, F.C.K., Kirby, J.T. (1994). Observation of undertow and turbulence in a laboratory surf
1221 zone. *Coastal Engineering* **24**: 51-80.

- 1222 Torres-Freyermuth, A., Lara, J.L. and Losada, I.J. (2010). Numerical modelling of short- and
1223 long-wave transformation on a barred beach. *Coastal Engineering* **57**: 317-330.
- 1224 Uncles, R.J. Stephens, J.A., Harris C. (2014). Infragravity currents in a small ría: Estuary-
1225 amplified coastal edge waves? *Estuarine Coastal and Shelf Science* **150**: 242-251.
- 1226 Valle-Levinson A. (2011). Classification of Estuarine Circulation. In: Wolanski E and
1227 McLusky DS (eds.) *Treatise on Estuarine and Coastal Science* **1**, 75–86. Waltham:
1228 Academic Press.
- 1229 Van Enckevort, I.M.J. (2001). Daily to yearly nearshore bar behaviour. Utrecht University,
1230 PhD Thesis, 174pp.
- 1231 Van Enckevort, I.M.J., Ruessink, B.G. (2003). Video observations of nearshore bar
1232 behaviour. Part 1: alongshore uniform variability. *Continental Shelf Research* **23**: 501–
1233 512.
- 1234 Vittori, G., DeSwart, H.E., Blondeaux, P. (1999). Crescentic bedforms in the nearshore
1235 region. *Journal of Fluid Mechanics* **381**: 271-303.
- 1236 Watson, G., Barnes, T. C. D., Peregrine, D. H. (1995). The generation of low-frequency
1237 waves by a single wave group incident on a beach. *24th International Conference on*
1238 *Coastal Engineering*, 776-790.
- 1239 Wijnberg, K.M., Terwindt, J.H.J. (1995). Extracting decadal morphological behavior from
1240 high-resolution, long-term bathymetric surveys along the Holland coast using
1241 eigenfunction analysis. *Marine Geology* **126**: 301-330.
- 1242 Wijnberg, K.M., Kroon, A. (2002). Barred beaches. *Geomorphology* **48**: 103-120.
- 1243 Wright, L.D., Coleman, J.M., Erickson, M.W., (1974). Analysis of major river systems and
1244 their deltas: morphologic and process comparisons. Technical Report 156, Coastal
1245 Studies Institute. Louisiana State University (125 pp.).
- 1246 Wright, L.D., Short, A.D. 1984. Morphodynamic variability of surf zones and beaches: A
1247 synthesis. *Marine Geology* **56**: 93-118.

1249 **Wave-forced dynamics in the nearshore river mouths, and swash**
1250 **zones**

1251 Maurizio Brocchini

1252 DICEA, Università Politecnica delle Marche, Ancona, Italy

1253

1254 **Abstract**

1255 The role of wave forcing on the main hydro-morphological dynamics evolving in the shallow
1256 waters of the nearshore and at river mouths is analyzed. Focus is mainly on the cross-shore
1257 dynamics that evolve over mildly sloping barred, dissipative sandy beaches from the storm
1258 up to the yearly time scale, at most. Local mechanisms, nonlocal mechanisms and
1259 connections across three main inter-related subsystems of the nearshore - the region of
1260 generation and evolution of nearshore bars, river mouths and the swash zone - are
1261 analyzed. The beach slope is a major controlling parameter for all nearshore dynamics. A
1262 local mechanism that must be properly described for a suitable representation of wave-
1263 forced dynamics of all such three subsystems is the proper correlation between orbital
1264 velocity and sediment concentration in the bottom boundary layer; while specific
1265 mechanisms are the wave-current interaction and bar generation at river mouths and the
1266 sediment presuspension at the swash zone. Fundamental nonlocal mechanisms are both
1267 Infragravity (IG) waves and large-scale horizontal vortices (i.e. with vertical axes), both
1268 influencing the hydrodynamics, the sediment transport and the seabed morphology across
1269 the whole nearshore. Major connections across the three subsystems are the upriver
1270 propagation of IG waves generated by breaking sea waves and swash-swash interactions,
1271 the interplay between the swash zone and along-river-flank sediment transport and the
1272 evolution of nearshore sand bars.

1273

1274 Introduction

1275 Focus, motivation and method

1276 This work focuses on wave-forced nearshore and river mouth dynamics, as opposite to
1277 tidally-forced dynamics - i.e. such that $RTR=TR/H_b < 3$, where TR is the tidal range and H_b is
1278 the height of breaking waves (Masselink and Short, 1993).

1279 In more detail, cross-shore (mostly) dynamics evolving from the yearly down to the storm
1280 time scales are investigated. With reference to coastal and river mouth hydro-morphological
1281 regimes the present analysis gives attention to the following two regimes.

1282 First, open-coast, mildly sloping, barred ($\Omega=H_b/w_s T_i > 2$, where Ω is the Dean parameter),
1283 w_s is the sediment fall velocity and T_i is the period of incoming waves - Wright and Short,
1284 1984) and dissipative ($\epsilon=A_b \omega^2 / g \tan^2 \beta > 20$, where the surf-scaling parameter ϵ depends on
1285 the amplitude of the breaking waves A_b , on the radian wave frequency $\omega=2\pi/T_i$ and on the
1286 beach slope β -Wright and Short, 1984) sandy beaches. Second, wave-dominated or river-
1287 dominated and wave-modified (which feed sediments to sea) river mouths (Cooper, 2001).

1288 Scopes of the present contribution are the following. First, to provide a fairly complete
1289 (obviously not exhaustive) and clear overall view of what we know and what we do not not
1290 know of wave-forced nearshore and river mouth dynamics. Second, to give the needed
1291 focus to recent progresses made available on specific dynamics. But the main aim is to
1292 **highlight links and relations among such dynamics**, this by properly highlighting
1293 nonlocal agents and relations.

1294 Regarding this problem as a puzzle to be solved (see figure 1), this entailing local and
1295 nonlocal relations, this paper tries to: put the analysis in the proper context, with no
1296 pretention for a systematic description; move from consolidated knowledge to new results;
1297 move from observations to modeling; move from the large to the small scales; inspect
1298 different types of models, both in terms of their structure and use.

1299

1300
1301
1302
1303
1304
1305

Figure 1. Wave-forced dynamics in the nearshore and estuaries: a puzzle to be solved. This photo, taken at the microtidal Misa River estuary (Senigallia, Italy), illustrates through one single view the three subsystems analyzed in the paper.

1306 The latter issue is analyzed in some detail because of its intrinsic importance and of the
1307 guidance that models provide in exploring the mechanisms of the phenomena at hand.

1308

1309 Generalities on Models

1310 The hydro-morphodynamics of the nearshore region, including that evolving at river mouths,
1311 is so complex that a number of different methods and models has been proposed and
1312 implemented over the years. However, such abundance of models has also brought to some
1313 major ambiguities in the definition and use of such models.

1314 Hence, some clarification seems useful, which is based on a detailed classification of the
1315 models in terms of their: intrinsic structure and use.

1316

1317 ***Model structure***

1318 Although this classification, as many others, is somewhat arbitrary (e.g. as function of the
1319 level of closure implemented), its use may help understand what model is actually applied
1320 for a specific analysis. The structure of the models at hand can be classed as either “process
1321 based” or “empirically based”.

1322 “Process-based models” rely on equations derived from fundamental physical principles
1323 and/or conservation laws. In the following three examples of such models (at different levels
1324 of time/space scales and resolution) are described.

1325 First, wave-averaged models: Nonlinear Shallow Water Equation (NSWE)-type (e.g.
1326 Delft3D, <https://oss.deltares.nl/web/delft3d/home>; Lesser et al., 2004; NHWAVE, Ma et al.,
1327 2012); Oceanographic (e.g. ROMS, <https://www.myroms.org/index.php>, Haas and Warner,

1328 2009), etc. Second, wave-resolving models: NSWE-type (e.g. Postacchini et al., 2012);
1329 Boussinesq-type (e.g. Brocchini, 2013; Kirby, 2016); etc. Third, models that account for
1330 turbulence through proper closures: Reynolds Averaged Navier-Stokes; Large Eddy
1331 Simulations (Chang and Scotti, 2004; Torres-Freyermuth, Lara and Losada, 2010; Lubin et
1332 al. 2011) etc.

1333 “Empirically-based models” rely on equations constructed from a mixture of conservation
1334 laws and observations, with a significant weight of empirical inputs. In fact, though also
1335 turbulence models like RANS and LES do require some empirism for the turbulence closure
1336 relations (hence the arbitrariness of classing models), their structure heavily relies on
1337 conservation laws and less on closures that are based on solid theoretical foundations.
1338 Examples of empirically-based models are: Regression-type models, Machine Learning
1339 Approaches (e.g. Pourzangbar et al., 2017), etc. and energetics-type, which solve a
1340 sediment mass conservation equation (Exner equation) where the mass fluxes come from
1341 empirically-based closure laws (e.g. Bailard, 1981).

1342

1343 **Model use**

1344 The above models can be, each in its own way, used in terms of either a template/forced
1345 approach or a stability approach (at times referred to as “self-organization”, see, for
1346 example, Coco and Murray, 2007). To clarify the above an example is made with reference
1347 to the classical Exner 1D sediment transport equation, which can be formulated as (e.g.
1348 Plant et al., 2001):

$$1349 \quad \frac{\partial z_b}{\partial t}(x, t) = I + II + III \quad (1)$$

1350 where z_b is the seabed position and

1351 I = is a forcing term due to initial bathymetry and wave field,

1352 II = gives the modification of forcing due to externally-forced changes in wave height,

1353 III = is a feedback term due to the rate of change of the seabed position z_b , which introduces
1354 a feedback behavior in (1).

1355 Template/forced models and stability-type models can be distinguished by the presence
1356 of I, II and III. Template/forced morphological models are such that $I \neq 0$, $II \neq 0$ and $III = 0$, while
1357 stability approaches must have $III \neq 0$. With reference to the specific problem of seabed
1358 features generation, while template models may identify compelling mechanisms, i.e. a
1359 template, for the initial stages of formation (e.g. at breakpoint, at nodes/antinodes of
1360 standing waves, etc.) stability models are such that perturbations evolve simultaneously in
1361 the hydrodynamic and bathymetric fields. This is illustrated by the examples given in figures
1362 2 and 3. The initial bathymetry used for the simulations of figure 3 consisted of a longshore-
1363 uniform sand bar.

1364
1365
1366 Figure 2. Forced/template mechanism: “synchronous standing edge waves” (left) and flow
1367 imprinted on sand giving beach cusps (right). Adapted from Coco and Murray, (2007).
1368
1369

1370 Figure 3. Stability/Self-organization mechanism: crescentic bars emerge from the feedback
1371 between flow and bathymetry (initial longshore-uniform bar bathymetry). Top: waves
1372 breaking on shoals force onshore flow, which returns to sea in gaps. Bottom: for suspended
1373 sediment decreasing from breaker line to shore, the onshore flow favors deposition (shoals
1374 grow) and the offshore flow favors erosion (scour holes grow). Hence, the development of
1375 crescentic bars. Adapted from Coco and Murray, (2007).
1376
1377

1378 The rest of the paper is structured as follows. The next section provides detailed
1379 information specific to 1) the generation and evolution of nearshore bars, 2) river mouth and
1380 3) swash zone dynamics. A short Discussion and then Conclusions close the paper.

1381

1382 Wave forced-dynamics for three fundamental subsystems: sand
1383 bars, river mouths and swash zone

1384 Sand bars

1385 The generation and evolution of sand bars in the nearshore is analyzed as a first applicative
1386 topic because of its many implications, which help set the scene for all the rest of the paper.
1387 Consolidated observations and models are first illustrated, then followed by more recent
1388 findings.

1389 Starting from the observations, sand bars that evolve in the nearshore are usually classed
1390 in terms of their planview shape, i.e. longshore uniform (2D) and rhythmic (3D). Different
1391 evolution mechanisms lead to different evolution time scales (e.g. Wijnberg and Kroon,
1392 2002). We can distinguish between: 2D sand bars (left panel of figure 4), which emerge as
1393 the outcome of wave-induced, cross-shore sediment transport and evolve over scales of
1394 days to months (e.g. Ruessink and Kroon, 1994); 3D sand bars (right panel of figure 4),
1395 which emerge as the outcome of horizontal circulation patterns and, because of this stronger
1396 forcing, evolve on shorter time scales, i.e. hours to days (e.g. Caballeria et al., 2002).

1397
1398
1399 Figure 4. Left panel: planview image of two shore-parallel sand bars at Noordwijk beach
1400 (The Netherlands, 19th November 2000). Right panel: planview image of surf zone complex
1401 topography at Duck (U.S.A. Atlantic coast, 10th January 1994). Adapted from Ribas Prats
1402 (2003).

1403
1404

1405 Three main interpretative models exist for the overall generation and evolution
1406 mechanisms of nearshore sandbars.

1407 The most known is the Net Offshore Migration (NOffM) model, which suggests that
1408 multiple nearshore 2D bars have multi-annual lifetimes and pass through 3 stages (figure
1409 5). Generation occurs close to shore because of the interaction of undertow, wave orbital
1410 velocity asymmetry and group-forced IG orbital motion (Roelvink and Stive, 1989). This is
1411 followed by a net seaward migration, which is the outcome of an alternation of gradual
1412 onshore movements during calms and episodic strong offshore movements during storms
1413 (Van Enckevort and Ruessink, 2003, see also the right panel of figure 5). Finally,

1414 degeneration occurs at the outer surf zone when the bar moves offshore of the breaker line
1415 where weakly or non-breaking waves induce a net onshore sediment transport. This step
1416 governs the entire cycle and only when the offshore bar degenerates at the breaker line the
1417 nearshore bar can start its offshore migration (Wijnberg and Terwindt, 1995).

1418

1419

1420 Figure 5. Left panel: seabed cross-shore profiles at Noordwijk, The Netherlands, from 1979
1421 to 1987, showing cyclic bar behavior with a recurrence interval of about 4-5 years. Adapted
1422 from van Enckvort, (2001). Right panel: seabed cross-shore profiles at Duck (U.S.A. Atlantic
1423 coast, September-October 1994). Thick curves with red dates are the beginning of storms,
1424 resulting in offshore bar migration (red arrows). Thin curves are the subsequent daily
1425 profiles. The sandbar migrated onshore (blue arrows) during calm periods between the
1426 storms. Courtesy of Dr. S. Elgar.

1427

1428

1429 The second model is the Net Onshore Migration (NO_{nm}) model, which suggests that sand
1430 bars are generated inshore of the breakpoint and migrate towards the swash zone (Aagard
1431 et al., 2004). The observations made on the Danish North Sea coast suggest that a net
1432 onshore bar migration occurs and the onshore bar movement in the mid-outer surf zone is
1433 due to a dominance of the wave-induced sediment transport during storm condition. The
1434 onshore-directed sediment transport is caused by incident wave skewnesses and enhanced
1435 phase couplings between orbital velocity and sediment concentration at times of high energy
1436 and water levels.

1437 Alternative to both NO_{ffm} and NO_{nm}, the Oscillation around a Position of Equilibrium
1438 (OPE, Certain and Barousseau, 2005) proposes a cyclic oscillation of the bar position due to
1439 seaward (storms) and shoreward (calms) migration. The mechanisms of this model are
1440 essentially similar to those of the NO_{ffm}, i.e. seaward migration during storms because of
1441 the dominance of the undertow and recovery during calms due to the wave-induced onshore
1442 sediment transport, but equilibrium is achieved when the local climate cannot not force storm
1443 strong or long enough (e.g. Mediterranean vs. North Sea, see also Armaroli and Ciavola,
1444 2011). Very recent observations (Parlagreco et al., 2019) show that Mediterranean storms

1445 can indeed induce a NOffM rather than an OPE behavior and that the longshore flow can
1446 sensibly contribute to increase the potential of sediment suspension.

1447 Both NOffM and NOnM are preferential for longshore-uniform 2D bars, which evolve for
1448 $\Omega \approx 5$ (see in the following), and seem to require large waves (Ω is directly proportional to H_b)
1449 and longshore flows (likely contributing to longshore uniformity, Shand et al., 1999).
1450 Moreover, since both NOffM and NOnM are observed to evolve over beaches with similar
1451 slopes (1:200-1:150) and sediment sizes ($d_{50} \approx 150 \mu\text{m}$), the different behaviours are probably
1452 to be searched into the different incident wave field (steepness and storm sequencing) and
1453 possible physiographic constraints.

1454 The above observations motivate a detailed analysis, using the mentioned classification,
1455 of the models currently available to predict the generation and evolution of 2D sand bars.
1456 These are typically distinguished between models for single but non-interacting bars and
1457 interacting sand bars.

1458 Among the models for single and non-interacting multiple bars we find the “empirical
1459 models”, the “template/forced” models and the “stability models”.

1460 The “empirical models” are based on use of the Dean parameter Ω (Wright and Short,
1461 1984) and of the the $B^* = x_s / g \tan \beta T_r^2$ parameter (Short and Aagard, 1993), where x_s is the
1462 nearshore width (offshore distance to the point where $\beta \approx 0$). Single bar formation is possible
1463 for $20 < B^* < 50$ and $2 < \Omega < 6$, while $\Omega < 1$ characterizes reflective beaches and $\Omega > 6$ dissipative
1464 beaches. For Ω increasing from 2 to 6 the following features emerge in sequence (Wijnberg
1465 and Kroon, 2002): shore-attached bars, 3D bars and 2D bars (the latter ones for $\Omega \approx 5$).
1466 Formation of multiple bars requires larger values of B^* (e.g. 2 bars for $50 < B^* < 100$, 3 bars
1467 for $100 < B^* < 300$, etc.);

1468 The “template/forced models” not only give synthetic information on conditions, but also
1469 clarify some of the mechanisms for bar formation. Most of them propose bar formation via
1470 sediment accumulation at specific spatial locations. Such locations can be (see also figure

1471 6): the anti/nodes of standing waves - be they induced by incoming/outgoing free IG waves
1472 (Short, 1975) or edge waves from opposite directions (Holman and Bowen, 1982), this is
1473 known as the “standing IG wave mechanism”; the breakpoint location, where flow/sediment
1474 convergence is due to the seaward-flowing undertow and the shoreward bottom boundary
1475 layer (BBL) streaming, this is known as the “breakpoint-bar mechanism”.

1476

1477

1478 Figure 6. Bar formation because of sediment convergence at: anti/nodes of standing IG
1479 waves (left panel), breakpoint (right panel).

1480

1481

1482 Another forced model, by O’Hare and Huntley (1994), suggests bar formation is induced by
1483 sediment stirring from sea/short waves and transport by free IG, hence we call this “sea-IG
1484 waves correlation mechanism”. Erosion/deposition is function of a fixed phase shift between
1485 the stirring given by short-wave envelope and IG waves. If IG are Bound Long Waves (BLW,
1486 more efficient on dissipative/mild beaches, Longuet-Higgins and Stewart, 1964), bar
1487 formation is inhibited, while if IG are generated by BreakPoint (BP, more efficient on
1488 reflective/steep beaches, Symonds et al., 1982) oscillation, bar formation is fostered.

1489 Forced models represent a major improvement over empirical models but are affected by
1490 some significant weaknesses. The most important is their intrinsic inability to give account of
1491 the morphology-on-flow- feedback (term III of equation 1 is automatically set to zero).
1492 Moreover, standing IG waves and breakpoint-bar mechanisms: prescribe the simultaneous
1493 emergence of all bars across the surf zone (this is only observed for rhythmic 3D bars, but
1494 not for 2D bars); require very specific wave conditions (e.g. narrow-banded spectra, mode
1495 selection, etc.), which are not frequent in nature.

1496 Finally, “stability models” have been successfully used to describe the generation of
1497 crescentic or lunate 3D bars (Vittori et al., 1999) and the evolution of a double system of
1498 crescentic 3D bars (Klein and Schuttelaars, 2006). Crescentic bars emerge because of
1499 steady currents caused by the interaction of the incoming sea/short waves with synchronous

1500 edge waves excited by the incoming waves moving over the wavy seabed. Klein and
1501 Schuttelaars (2006) studied a similar problem using a numerical, wave-averaged model
1502 (NSWE+Exner). Starting from an equilibrium state and numerically added perturbations the
1503 fastest-growing perturbation was computed (see also the left panel of figure 7).

1504
1505

1506 Figure 7. Left panel: nonlinear stability solution for a double sand bar system giving
1507 asymmetric crescentic patterns on inner and outer bars (adapted from Klein and
1508 Schuttelaars, 2006). Right panel: Evolution of longshore uniform sand bar after 30 days of
1509 model run with normal wave incidence. Longshore instability in the form of rip channels
1510 (adapted from Dronen and Deigaard, 2007).

1511
1512

1513 However, no study seems available on the emergence of 2D sandbars by means of
1514 stability models. Using a wave-averaged model Dronen and Deigaard, (2007) have shown
1515 that longshore instabilities of an initially 2D bar emerge for both normal and oblique wave
1516 incidence (see also the right panel of figure 7). This inadequacy of stability models in
1517 reproducing the emergence and evolution of 2D sandbars calls for intensive investigation,
1518 which should also take into proper account the role of longshore currents, which are thought
1519 to be fundamental to give longshore uniformity.

1520 With reference to models for interacting multiple bars, Castelle et al. (2010) used a
1521 numerical, wave-averaged circulation model (with wave action equation for wave driver) plus
1522 Exner equation to investigate the evolution of double crescentic sandbar systems.

1523
1524

1525 Figure 8. Computed (top-right and bottom panels) 180° out-of-phase coupled patterns with
1526 an outer-bar horn facing an inner-bar bay, similar to observations (top-left panel) on a meso-
1527 macro-tidal, high-energy, double-barred beach (adapted from Castelle et al., 2010).

1528
1529

1530 Simulations were run by providing a template in the morphology (inner bar longshore
1531 uniform and outer-bar with crescentic patterns). The computed feedback between flow,
1532 sediment transport and the evolving morphology allowed to reproduce morphologies similar
1533 to those of observed double crescentic sandbar systems (see figure 8). The Authors

1534 advanced that a novel mechanism (“morphological coupling”) was introduced that “blurs the
1535 distinction between self-organization and template mechanisms”. However, inspection of
1536 equation (1) suggests that such a novel mechanism is a complete model that retains all
1537 terms on the right hand side. In other words, a template of crescentic outer bar is forced
1538 (term I of equation 1) in a self-interating model (term III of equation 1).

1539 The above consolidated knowledge has opened the way to some recent findings on:
1540 sediment transport due to IG waves, sandbar migration and seabed moulding due to large-
1541 scale vortices.

1542 Recent results on IG wave-induced sediment transport come from De Bakker et al. (2016)
1543 who have used field data collected in the Netherlands at the gently sloping ($\beta \approx 1:80$) Ballum
1544 beach and at the moderately sloping ($\beta \approx 1:35$) SandMotor beach to investigate the role of
1545 the beach slope on IG wave-induced cross-shore sediment transport (q_{IG}). Based on the
1546 fundamental idea that sediment is put into suspension by sea/short waves and transported
1547 by IG waves (see also the “sea-IG waves correlation mechanism” by O’Hare and Huntley,
1548 1994), the study compares the performances of two proxies for sediment transport in the
1549 nearshore. The first proxy is the correlation between the group bound IG wave (BLW) and
1550 the sea-wave envelope (see also Baldock, 2006):

$$1551 \quad R_{Uu_{IG}}(\tau) = \frac{\langle U(t)u_{IG}(t+\tau) \rangle}{\sigma_U \sigma_{u_{IG}}} \quad (2)$$

1552 where u_{IG} is the IG wave cross-shore velocity, U is the sea-wave envelope cross-shore
1553 velocity, σ_s give the related variances and $\langle \rangle$ denotes time averaging. The second proxy,
1554 proposed by De Bakker et al. (2016), is the IG wave height to sea wave height ratio H_{IG}/H_{SW} .

1555 De Bakker et al. (2016) suggest that $R_{Uu_{IG}}$ is a good proxy for the nearshore sediment
1556 transport only on moderately-sloping beaches while H_{IG}/H_{SW} well represents the sediment
1557 transport on all beaches. In fact while over moderate slopes $R_{Uu_{IG}} \neq 0$, it is $R_{Uu_{IG}} = 0$ on

1558 mild slopes. On the other hand, H_{IG}/H_{SW} is always positive and increasing for mild slopes,
1559 because IG waves receive more energy from sea waves breaking on wide surf zones.

1560 In the shoaling region it is $R_{Uu_{IG}} < 0$ because the setdown (negative BLW) is associated
1561 with the peak of the sea wave envelope and the sediment flux is negative, i.e. seaward (see
1562 also figure 9). In the inner surf zone of moderately-sloping beaches the correlation becomes
1563 positive because breaking-induced wave setup (positive BLW and shoreward sediment flux)
1564 gives deeper waters and allows the faster sea waves to ride the crest of the IG wave (see
1565 figure 9). However, in the inner-surf zone of mildly-sloping beaches it is $R_{Uu_{IG}} = 0$ because
1566 the sea waves have been almost completely dissipated and $q_{IG} < 0$ because the peaks of
1567 suspended sediment transport (SST) correlate with $u_{IG} < 0$ and the undertow \bar{u} .

1568

1569

1570 Figure 9. Sketch of IG-sea wave interaction and cross-shore sediment transport in the
1571 nearshore. From left to right: shoaling zone, inner-surf zone of moderately-sloping beaches,
1572 inner-surf zone of mildly-sloping beaches.

1573

1574

1575 Other recent results are on the mechanisms for onshore bar migration, fundamental for
1576 all bar migration models of above. Early energetics-type models could only reproduce
1577 offshore migration due to strong offshore currents (undertow) that are maximum at the bar
1578 crest (Gallagher et al., 1998). Observations suggest that onshore migration occurs for weak
1579 currents and significant waves. Hence, the role of waves is fundamental for the onshore bar
1580 migration, and, as a consequence, it is fundamental to understand what are the model
1581 representations of the main wave mechanisms for onshore bar migration.

1582 With reference to the models analyzed in the Introduction we first analyze the energetics-
1583 type models. They base their ability to reproduce onshore bar migration on use of the wave
1584 acceleration skewness (Sk_a , proportional to the third order acceleration moment) in the
1585 sediment transport closure, so that the sediment transport rate is written as (Hoefel and
1586 Elgar, 2003):

1587 $Q_A = K_a(a_{spike} - \text{sgn}(a_{spike})a_{crit}), \text{ for } |a_{spike}| \geq a_{crit} \text{ with } a_{spike} = \frac{\langle a^3 \rangle}{\langle a^2 \rangle} = \langle a^2 \rangle^{1/2} Sk_a \quad (3)$

1588 This description is such to properly give account of the rapid onshore flow acceleration under
 1589 steep wave fronts, which, in turn, generates strong horizontal pressure gradients on the
 1590 seabed sediment.

1591 On the other hand, process based, wave-resolving models (e.g. Henderson et al., 2004;
 1592 Hsu et al., 2006) can properly predict onshore bar migration if they accurately describe the
 1593 correlation between the orbital velocity (u) and the sediment concentration (c) in the BBL
 1594 ($q = \langle uc \rangle$).

1595
 1596
 1597 Figure 10. From left to right: vertical profile of mean cross-shore sediment transport due to
 1598 waves (q_w) and currents (q_c), total cross-shore sediment transport as function of near-bed
 1599 (middle) and freestream (right) velocity moments, for the Duck94 migration event. Adapted
 1600 from Henderson et al. (2004).

1601
 1602
 1603 Figure 10 shows that onshore migration occurs when the total sediment transport (q_t) is
 1604 essentially due to waves (q_w) and correlates well with the third-order moment of the nearbed
 1605 velocity ($\langle u_b^3 \rangle$).

1606 The apparent distance between the two models can be reconciled, showing that energetics-
 1607 type models are successful if they reproduce, at the integral level, the proper sediment flux-
 1608 nearbed velocity moments correlation given by advective BBL models. The proof can be
 1609 summarized as follows. First, it is shown that onshore bar migration occurs when q
 1610 correlates well with $\langle u_b^3 \rangle$ in BBL, i.e. with the nearbed wave velocity skewness $Sk_{u_b} =$
 1611 $\langle u_b^3 \rangle / \langle u_b^2 \rangle^{3/2}$. Then, using a Fourier decomposition of the velocity field, it is shown that $Sk_{u_b} \propto$
 1612 $\sin(\phi_b) A_{u_f}$, where A_{u_f} is the wave freestream velocity asymmetry and ϕ_b is the nearbed-
 1613 over-freestream velocity phase lead (Henderson et al., 2004; Berni et al., 2013). Finally, for
 1614 sawtooth bores, i.e. setting $\phi = \pi/2$ in the modal solution $u_f \sim \sum_n \frac{1}{2^n} \sin[(n+1)\omega t + n\phi]$

1615 gives $A_{u_f} \sim Sk_{a_f}$, with Sk_{a_f} wave freestream acceleration skewness (Drake and Calantoni,
1616 2001) of energetics-type models.

1617 Very recently Fernandez-Mora et al., (2015) used an energetics-type model to show the
1618 ability of such models to predict onshore/offshore bar migration and assess the role of both
1619 Sk_{u_b} and Sk_{a_f} . A complex model was built such that:

$$1620 \quad (1 - n) \frac{\partial z_b(x,t)}{\partial t} = - \frac{\partial Q(x,t)}{\partial x} \quad (4)$$

$$1621 \quad \text{with } Q = \alpha_c Q_c (Sk_{u_c}) + \alpha_u Q_u (Sk_{u_f}) + \alpha_a Q_a (Sk_{a_f}) + \alpha_D Q_D \left(\frac{dz_b}{dx} \right)$$

1622 and where Sk_{u_c} is the skewness of the currents velocity vector and dz_b/dx is the, stabilizing,
1623 seabed slope term. Hence, if α_u is set to zero the SkA model of Hoefel and Elgar (2002) is
1624 obtained, if α_a is set to zero the SkV model of Hsu et al. (2006), while the complete model
1625 is the MiX model. The model performances are illustrated in figure 11, always referring to
1626 the Duck94 onshore migration event.

1627
1628
1629 Figure 11. Modeling (red line) of the morphological evolution of the Duck94 onshore
1630 migration event for: (A) the SkV model, (B) the SkA model, and (C) the MiX model. The initial
1631 and final measured profiles are shown by dashed and solid lines, respectively. Adapted from
1632 Fernandez-Mora et al. (2015).

1633
1634
1635 All models reproduce well the chosen event, the SkV model better in the shoaling zone
1636 and the SkA model better in inner surf zone. However, they all perform poorly in the swash
1637 zone, this suggesting further investigation is needed for such region.

1638 Among the most recent results on seabed evolution and bar generation are those
1639 dedicated to the roles of macrovortices. Breakers of finite longshore width are known to
1640 introduce vertical vorticity ζ that rolls up to form large-scale vortices with vertical axes, also
1641 called macrovortices (Peregrine, 1999). Important dynamics characterized by finite
1642 longshore breakers are those giving rise to rip currents (e.g. Kennedy et al., 2006) and cross-
1643 sea induced breakers (e.g. Postacchini et al., 2014).

1644 The latter condition is seen to be a potentially important cause of seabed morphology
1645 alteration with the generation of complex 3D sandbars. The example of figure 12 has been
1646 obtained by computing, using the NSW morphodynamic model of Postacchini et al. (2012),
1647 the morphological evolution of a planar beach under the combined action of waves and
1648 macrovortices generated by finite longshore breakers due to regular waves ($H_i=1\text{m}$, $T_i=10\text{s}$,
1649 $\beta=1/30$) crossing at an angle of 15° to the coastline.

1650
1651
1652 Figure 12. Seabed evolution forced by sea waves and macrovortices over a planar beach.
1653 Red thin lines give contours of the free surface elevation, while black and blue solid lines
1654 are contours of clockwise and anticlockwise vorticity, respectively. Yellow gives sand
1655 deposition and green seabed erosion.

1656
1657
1658 Sandbars of 3D shape (yellow colours in figure 12) arise due to the interaction of cross-sea
1659 waves and the breaking-induced macrovortices.

1660

1661 River mouths

1662 The hydro-morphodynamics evolving at river mouths, significantly related with the evolution
1663 of sandbars, is analyzed, as a second applicative topic, with the same approach used for
1664 sandbars: consolidated observations and models first, followed by more recent findings.

1665 In view of the main aim of the paper, i.e. to highlight interactions amongst the subsystems
1666 of interest, fundamentals of river mouth dynamics are kept at the minimum, with a preference
1667 for those of wave-dominated rivers debouching into sandy coasts.

1668 Analysis of the literature on river mouths dynamics is fairly complex because of the
1669 different perspectives taken by the various communities that study river mouths, i.e.
1670 engineers, geologists, ecologists, etc. Also, the vocabulary is influenced by such different
1671 views and much effort is being spent to provide a systematic description of river mouth
1672 dynamics.

1673 The most important efforts have been made in the classification of river deltas (Galloway,
1674 1975), where delta is here generically used as “sedimentary deposit due to river-sea
1675 interaction and protruding into the sea”. The triangular classification of deltaic depositional
1676 systems by Galloway (1975) illustrates how the relative importance of sediment input, wave
1677 energy flux and tidal energy determines both morphology and stratigraphy of the delta.
1678 Indices have also been introduced to classify deltas. Among them we find the discharge
1679 index I_q of very consolidated use (Wright et al., 1974) and the more recent fluvial dominance
1680 ratio R (Nienhuis et al., 2015):

$$1681 \quad I_q = \text{discharge index} = \frac{\text{river liquid discharge per unit width of mouth}}{\text{wave power per unit width of wave crest}} \quad (5a)$$

$$1682 \quad R = \text{fluvial dominance ratio} = \frac{Q_r}{Q_{s,max}} = \frac{\text{river sand flux}}{\text{max longshore sediment flux at river mouth}}$$

1683 (5b)

1684 The former is a dimensional index (units of $\text{m}^3/\text{Watts/s}$) and such to increase of orders of
1685 magnitude from wave-dominated ($I_q \approx 10^{-5} \text{ m}^3/\text{Watts/s}$) to river-dominated ($I_q \approx 1 \text{ m}^3/\text{Watts/s}$)
1686 systems, while the latter is a dimensionless index, also increasing from $R < 1$ for wave-
1687 dominated to $R > 1$ for river dominated systems (see also figure 13).

1688
1689
1690 Figure 13. Classification of river delta morphologies by means of the fluvial dominance ratio.
1691 Adapted from Nienhuis (2016).
1692
1693

1694 Reference classifications of river deltas (Galloway, 1975) also clarify use of the word
1695 “estuary”, this being attributed to tide-dominated deltas, though at times more broadly
1696 (improperly?) used for semi-enclosed regions where river fresh water meets the sea. A
1697 number of schemes to classify estuaries is available in the literature, a recent one being that
1698 of Valle-Levinson (2011), which provides different classifications based on geomorphology,
1699 salinity structure and hydrodynamics.

1700 Because of the ambiguities in using “estuary” and of the present focus on river systems
1701 weakly influenced by tides, “river mouth” is here used as a clear and unambiguous definition
1702 of the region where the river meets the sea. Analyses of wave-dominated river mouths are
1703 also available (e.g. Cooper, 2001), and the very comprehensive, recent, work by Anthony
1704 (2015) provides a nice overview of the role of waves at river mouths: waves remove fine
1705 sediments at the river mouth and flatten the delta by moving the sediment alongshore.
1706 Hence, deltas, placed along smooth coasts with shore-parallel ridges (see figure 14), are
1707 shaped from arcuate to cusped. Waves that approach the coast obliquely generate: littoral
1708 (i.e. longshore) sediment transport, maximum for 45° wave attack; an important difference
1709 between updrift/downdrift flank transport and sediments (updrift flank made by beach
1710 sediments, downdrift flank made by river sediments); an asymmetrical delta. A recent
1711 quantitative analysis on delta progradation/migration is that of Nienhuis et al. (2016), which
1712 can be summarized as follows. The delta undergoes: a symmetric progradation, for small
1713 river sand flux and all longshore transported sediment bypassing the mouth; a downdrift
1714 migration, for small river sand flux and vanishing sediment bypassing at mouth; an updrift
1715 migration, for large river sand flux, independently of sediment bypassing at mouth.

1716
1717
1718 Figure 14. River delta geomorphic units and schematic shoreline morphology (wave-
1719 dominated morphology circled in red). Adapted from Anthony (2015).

1720
1721
1722 Being the main aim of the present work that of clarifying the interactions - through nonlocal
1723 agents - between the sandbar, river mouth and swash zone subsystems, most of the
1724 attention is here given to local and nonlocal wave-forced dynamics.

1725 Fundamental wave-forced, river mouth dynamics are the interaction of river current with
1726 the sea waves and the related sediment transport.

1727 Seaward flowing currents interact with incoming waves (Chawla and Kirby, 2002) to: slow
1728 them down and even block them if the depth-uniform current velocity is equal and opposite

1729 to the wave group velocity, this is known as “wave blocking”; steepen and even lead to
1730 breaking; decrease the high-frequency content by wave breaking and transfer part of the
1731 current energy to lower wave frequencies, this leading to frequency downshift of the
1732 incoming waves;

1733 The bedload accumulation related to the above dynamics typically leads to the formation
1734 of a mouth sandbar. The intensity of the wave field significantly influences the bar formation
1735 process. With no waves, a triangular mouth bar is generated just seaward of the mouth, at
1736 a distance of about twice the mouth width. The generation process is initiated by the river
1737 jet expansion in the sea, which, by continuity arguments, leads to a flow reduction and a
1738 consequent sediment deposition (e.g. Fagherazzi et al., 2015). On the other hand, small
1739 waves promote mouth bar formation by a process similar to the above: they promote the
1740 river jet expansion and flow reduction, with a consequent more intense deposition close to
1741 the river mouth. Finally, large waves suppress mouth bar formation because of their intense
1742 seabed erosion and longshore transport (Fagherazzi et al., 2015).

1743 Analysis of the above-mentioned wave-current interactions reveals that a river mouth acts
1744 as a low-pass filter by removing (by breaking) sea and swell waves and letting long waves
1745 pass. Recent studies have been dedicated to the propagation and evolution of IG waves at
1746 inlets. Observations at the Ría de Santiuste (Spain, see Uncles et al., 2014) reveal how IG
1747 waves (with periods of 4-5minutes), believed to be the outcome of estuary-amplified edge
1748 waves, propagate upriver. This is well illustrated by the top panel of figure 15, which shows
1749 a positive correlation between the water level (WL) and the flow velocity (U' in the figure).

1750

1751

1752 Figure 15. IG wave propagation at river mouths. Top: positive water level-flow velocity
1753 correlation at the Ría de Santiuste (Spain). Adapted from Uncles et al. (2014). Bottom: tide-
1754 modulated IG wave propagation in the Albufeira Inlet (Portugal). No IG waves propagate
1755 landward of the mouth at ebbs. Adapted from Bertin and Olabarrieta (2016).

1756

1757

1758 The bottom panel of the same figure nicely illustrates the role of tide on the propagation
1759 of IG waves at inlets. Observations and wave-averaged modeling of the Albufeira Inlet
1760 (Portugal, see Bertin and Olabarrieta, 2016) reveal that upriver-propagation (from transect
1761 0 to transect 18 of images C-Run 3 and D-Run 4) of IG-waves only occurs during tidal flood.
1762 Inspection of the same images reveals that: no propagation is possible at ebb (white areas
1763 in correspondence of the ebbs), IG waves are equally generated by BLW (image C-Run 3)
1764 and BP (image D-Run 4) mechanisms.

1765 Other recent field observation support the role of river mouths and the upriver propagation
1766 of IG waves. Among these, notable are the EsCoSed and MORSE project studies (Brocchini
1767 et al., 2017; Melito et al., 2018a, 2019) performed at the mouth of the Misa River (Italy, see
1768 figures 1 and 17), located at a microtidal site ($RTR < 3$) with a dissipative-to-intermediate
1769 beach ($5 < \Omega < 20$).

1770
1771
1772 Figure 16. The Misa River estuary in Senigallia (Italy) depicting locations of measuring
1773 stations within the final river reach (TGdown, QR1, QR2, QR3, TGup) and in the sea (QS1,
1774 QS2, QS3). Adapted from Melito et al. (2018a).

1775
1776

1777 On the basis of water elevation measurements collected at various locations both in the
1778 sea (QS in figure 17) and in the river (TG and QR in figure 17), the energy flux density dE_f
1779 and the energy flux E_f :

1780
$$dE_f = S(f)C_g \quad dE_f = \int S(f)C_g df \quad (6)$$

1781 have been computed - where $S(f)$ is the power spectral density, C_g the wave group velocity
1782 and f the frequency. The fundamental result is that while in the sea (QS1, QS2) most of the
1783 energy content is due to swell and sea waves, the IG band becomes dominant within the
1784 river (e.g. QR2 of the left panel of figure 17) with a weak upriver decay (TGup of the right
1785 panel of figure 17).

1786

1787
1788
1789
1790
1791
1792

Figure 17. Energy density flux (left) and energy flux (right) at the Misa River during storm conditions. Adapted from Melito et al. (2018a).

1793 The EsCoSed field campaign has also provided useful information on the seabed
1794 morphology at the river mouth, highlighting the formation of large deposits/bars (blue in the
1795 right panel of figure 18) and scours holes (red in the right panel of figure 18). The sand
1796 deposits are found to be at a typical location, i.e. about twice the mouth width seaward of
1797 the mouth itself (Fagherazzi et al., 2015), but characterized by a very complex planview
1798 pattern.

1799
1800

Figure 18. Seabed variation between May-September 2013 (left) and between September 2013-February 2014 (right) at the Misa River. Adapted from Brocchini et al. (2017).

1803
1804
1805

1806 The above observations have motivated important efforts in the modeling of the dynamics
1807 evolving at a river mouth characterized by a complex morphology, with a specific focus on
1808 the role of waves.

1809
1810

Figure 19. Velocity maps for a strong river outflow opposed to three different wave regimes (from top to bottom: $H_s = 0.0\text{m}$, 0.5 m , 1.5 m). Left column: results from Olabarrieta et al. (2014). Right column: results from the wave-resolving solver of Brocchini et al. (2001), averaged over 10 wave periods. Adapted from Melito et al. (2018b).

1814
1815
1816

1817 Olabarrieta et al. (2014) used the wave-averaged 3D Regional Ocean Modeling System
1818 (ROMS) and an idealized river mouth shoal bathymetry to explore the wave-river current
1819 interactions. The increased, with respect to a planar beach, steepening and breaking of
1820 waves over the shoal leads to stronger seaward-flowing currents (undertow). Moreover,
1821 waves are seen to increase both the lateral spreading of the river current and the longshore
transport.

1822 In the attempt to better describe the dynamics forced by the sea waves, Melito et al.
1823 (2018b) run the same cases studied by Olabarrieta et al. (2014) using the intra-wave NSWE
1824 model of Brocchini et al. (2001), this neglecting vertical flow gradients, but improving the
1825 time resolution of the wave forcing. The comparison between the two different numerical
1826 solutions, illustrated in figure 19 for the case of a strong river outflow (1.1 m/s river current),
1827 suggests that the main features of the wave-current interactions are captured, i.e. the river
1828 current intensity and instability (e.g. Canestrelli et al., 2014). However, some significant
1829 differences are also evident, among which: a weaker longshore spreading of the river jet,
1830 the appearance of a number of large-scale eddies. Both differences seem to be caused by
1831 the stronger wave breaking predicted by the NSWE, which removes significant amounts of
1832 momentum and energy from the wave field, thus the reduced capacity of pushing shoreward
1833 the river jet, in favour of the generation by differential wave breaking of large-scale eddies
1834 or macrovortices (see previous subsection and Kennedy et al., 2006).

1835 Such macrovortices not only contribute to the near-mouth flow mixing (right panels of
1836 figure 19), but, as seen at the end of the previous subsection, also induce significant seabed
1837 morphological changes. Falcini et al. (2014) recently sought for a mechanistic relationship
1838 between potential vorticity Π ($\Pi = \zeta/d$ in a shallow-water framework) and sediment transport
1839 at a river mouth. Lateral advection and diffusion of suspended sediment were found to be
1840 directly proportional to the river jet vorticity, so that levees building is observed in
1841 correspondence of regions of large vorticity and mouth bar deposition in regions of low
1842 vorticity.

1843 In view of the large similarities between the analyses of vorticity-induced dynamics within
1844 the nearshore by Kennedy et al. (2006) and Brocchini (2013) and at river mouths by Falcini
1845 et al. (2014), an integration of the two analytical approaches seems useful to clarify the role
1846 of breaking-wave-induced vorticity on the river mouth morphodynamics.

1847 In the meantime, some numerical experiments are being carried out, by means of the
 1848 hydro-morphodynamic NSW solver of Postacchini et al. (2012), to describe the main
 1849 morphological features that evolve at a simplified (e.g. shoal of Olabarrieta et al., 2014) and
 1850 natural (e.g. Misa River) river mouth bathymetry. Preliminary numerical results are illustrated
 1851 in figure 20, describing both a significant sediment deposition at the seaward edge of the
 1852 idealized shoal and a weaker deposition at the typical location of a mouth bar (left panel)
 1853 and a complex erosion deposition pattern at the mouth of the Misa River (right panel).

1854
 1855
 1856
 1857
 1858
 1859

Figure 20. Estimated morphological evolution at the simplified river mouth bathymetry of Olabarrieta et al. (2014), left panel and at the Misa River, right panel.

1860 Comparison of the right panels of figures 18 and 20 reveals that these preliminary
 1861 numerical solution can qualitatively capture the main characteristics of the complex
 1862 erosion/deposition pattern that evolves just seaward of the river mouth.

1863

1864 Swash zone

1865 The swash zone is that region of the beach that is alternatively covered and uncovered by
 1866 water because of the action of sea waves (Brocchini and Baldock, 2008). Because of such
 1867 process, fundamental characteristics of the swash zone are: the flow intermittency; the
 1868 infiltration/exfiltration of water at the seabed; the very large sediment transport within the
 1869 swash, which can be quantified, for example, in terms of the cross-shore sediment mass per
 1870 beach longshore length transported by each swash event (Blenkissopp et al., 2011)

$$1871 \quad -40 \text{ Kg/m} < \hat{q}_x < 40 \text{ Kg/m}, \quad \hat{q}_x \equiv \int_{T_{swash}}^{\square} q dt \quad (7)$$

1872 or in terms of the longshore mass flux (Ribeiro et al., 2012).

$$1873 \quad Q_y \sim (1 - 10) \text{ Kg/s} \quad (8)$$

1874 The swash zone is also a locus for the generation of IG waves by two main mechanisms.

1875 The first is the seaward reflection of BLW released by breaking (e.g., Schäffer, 1993).
 1876 This is well illustrated in the left panel of figure 21, where a group of sea waves releases
 1877 through breaking its BLW that propagates to the shoreline and it is there reflected to
 1878 propagate out to sea as a Free Long Wave (FLW). The process of faster sea waves climbing
 1879 the crest of the BLW is evident, as also illustrated in figure 9.

1880 The second mechanism is the frequency downshifting of the sea waves that interact in
 1881 the swash. This interaction occurs because waves in shallow water propagate with a velocity
 1882 that is proportional to their height. Therefore, large waves travel faster than small waves and
 1883 catch them up near or inside the swash zone (Mase, 1995). The assimilation of small waves
 1884 by large waves leads to a reduced number of waves emerging from the swash zone, i.e. to
 1885 a frequency downshifting, as illustrated in the right panel of figure 21.

1886
 1887
 1888 Figure 21. Left: example of BLW release on a beach because of sea wave breaking. The
 1889 shoreward-propagating (from bottom to top) wave group releases its BLW that is reflected
 1890 at the shoreline and propagated (from top to bottom) out to sea as a FLW (adapted from
 1891 Watson, Barnes and Peregrine, 1995). Right: example of frequency downshifting within the
 1892 swash zone. Bichromatic waves propagate from seaward to the shoreline (from top to
 1893 bottom) over a steep beach and interact in the swash zone giving wave of doubled period
 1894 (adapted from Mase, 1995).

1895
 1896
 1897 Most of the modelling of swash zone hydrodynamics, essentially forced by sea waves,
 1898 has been done by using the NSW. For an horizontally-1D flow evolving in the crossshore
 1899 direction x , these read (e.g. Brocchini et al., 2001):

$$1900 \quad \begin{cases} d_{,t} + (ud)_{,x} = 0 \\ u_{,t} + uu_{,x} + gd_{,x} = gh_{,x} + c_f \frac{u|u|}{d} \end{cases} \quad (9)$$

1901 where commas gives partial differentiation, d and h are the total and still-water depths,
 1902 respectively, u is the depth-averaged flow velocity, g is gravity acceleration and c_f a
 1903 dimensionless coefficient for seabed friction.

1904 Solution of such equations not only provides a quantitative means for studying the
1905 nearshore hydrodynamics, but can also be of help to illustrate some of the phenomena
1906 above described, like, for example, the interaction of sea waves within the swash zone. An
1907 example is provided by figure 22 for the case of a group of 5 sea waves.

1908
1909

1910 Figure 22. Waves interacting at a wall (left) and at a swash zone (right). Top: incoming and
1911 outgoing characteristic curves at the shoreline. Bottom: a group of 5 sea waves (thin line)
1912 incoming to shore and outgoing IG waves (S-, thick line) resulting from the interaction of sea
1913 waves (adapted from Brocchini, 2006).

1914
1915

1916 On the basis of the NSWE a number of analytical solutions has been made available over
1917 the years for both non-breaking and breaking waves. The pioneering work of Carrier and
1918 Greenspan (1958), who used the method of hodograph transformation, opened the way to
1919 some solutions that are currently used as reference for both analytical and numerical studies
1920 for the evolution of non-breaking sea waves on a beach. In particular, Brocchini and
1921 Peregrine (1996) extended the solution by Carrier and Greenspan (1958) to compute for the
1922 longshore flow due to waves propagating almost orthogonally to the beach (see figure 23).
1923 This solution can be used to provide estimates of the longshore mass flux to be compared
1924 with field measurements, like that of equation (8). More recent studies, e.g. Antuono and
1925 Brocchini (2010), suggest that analytical solutions for both non-breaking and breaking waves
1926 can be achieved also in the physical space, i.e. avoiding use of the hodograph
1927 transformation.

1928
1929

1930 Figure 23. Left: maps, in the (x,t) plane of hodograph coordinates, free surface elevation η
1931 crossshore (u) and longshore (v) velocities for two sea waves interacting onto a beach. Right:
1932 mean longshore mass flux near and inside the swash zone for waves of different amplitude
1933 (adapted from Brocchini and Peregrine, 1996).

1934
1935

1936 Cornerstone of analytical solutions for breaking sea waves, i.e. bores, on a beach is that of
1937 Shen and Meyer (1963), in the following SM63. A bore of velocity u_0 collapsing onto a beach
1938 of slope β forces a swash that is shoreward bounded by a parabolic shoreline x_s :

1939
$$x_s = u_0 t - \frac{1}{2} g \tan \beta t^2 \quad (10)$$

1940 i.e. similar to the trajectory of a massive point under the action of gravity. This very neat
1941 solution, which has been used for long times, has, however, the fundamental weakness of
1942 predicting runup lenses much thinner than those observed on a natural beach.

1943 Very recently Guard and Baldock (2007), in the following GB07, analyzed this issue and
1944 suggested that the problem is caused by the assumption of regarding the collapsing bore
1945 as a dam-break flow, where part of the fluid accelerates seaward. This is illustrated in the
1946 sketch of the top panel of figure 24, whose left part illustrates the evolution of the SM63
1947 solution in similarity with a dam-break flow (dashed line) that leads to both shoreward- and
1948 seaward-accelerating fluid. The right image of the same panel illustrates how the GB07
1949 solution, achieved by means of a linearly-time-varying (rather than constant) flow forcing at
1950 collapse, predicts only shoreward-accelerating fluid after bore collapse. The bottom panel
1951 of the same figure shows the fundamental differences in the water depths predicted by the
1952 two solutions over a swash period. The SM63 solutions (thick lines) predicts more
1953 asymmetric (downrush much longer than uprush) and much thinner (about a half) swash
1954 lenses than those predicted by the GB07 solution (thin lines) that well compares with the
1955 experimental data given by the symbols.

1956 Recent remote sensing data, collected at 6 different beaches in USA and Australia, have
1957 been used to assess the value and validity of the assumptions at the basis of the GB07
1958 solutions (Power et al., 2011). Such observations confirm that: natural swash lenses are
1959 deeper and less asymmetric than those predicted by SM63; a linearly-time-varying forcing
1960 of the swash flow is better suited than a time-independent forcing to describe the evolution
1961 of natural swashes.

1962
1963
1964
1965
1966
1967
1968
1969
1970
1971
1972

Figure 24. Top: sketches of the flow by bore collapse. The left sketch shows the evolution of the SM63 solution in comparison to a dam-break flow (dashed lines), the right sketch suggests that only shoreward-accelerating fluid motion is predicted by the GB07 solution. Bottom: swash water thickness predicted by the SM63 solution (thick lines) and by the GB07 solution (thin lines) also compared with experimental data (symbols). Bottom panel adapted from GB07.

The fact that natural swash are fairly symmetric (still asymmetric in favour of the downrush) in time and characterized by fairly deep water lenses has important consequences on the understanding and modeling of the swash zone sediment transport and related morphodynamics. Thicker swash lenses lead to increased sediment pickup areas, because the sediment transport rate is directly proportional to the flow depth for a uniform sediment concentration (Pritchard and Hogg, 2005), and increased seaward sediment transport, because of the still existing swash asymmetry in favour of the rundown phase. However, such seaward sediment transport is counter-balanced by two main mechanisms: the presuspension of sediments due to the bore collapse onto the beach (which allows for a significant transport from the surf zone into the swash zone) and the settling lag effects due to the inertia of suspended sediment particles advected into the swash zone (Pritchard and Hogg, 2005). The above competing mechanisms, i.e. reduced swash asymmetry with increased thickness (promoting seaward transport) and sediment presuspension with settling lag effects (promoting shoreward transport) force a very complex swash zone morphology. This has been well described by Kelly and Dodd (2010), who implemented one of the most advanced and physically complete models for the swash zone hydro-morphodynamics. Some of their results reveal the complexities in beach profile changes due to a swash event (see figure 25).

1990
1991
1992
1993
1994

Figure 25. Left: space-time map of the changes in beach profile due a swash event. The thick dashed lines give the zero contour. Right: snapshots of the swash lens cross-shore sections (adapted from Kelly and Dodd, 2010).

1995

1996

1997 While beach degradation (light tones of grey) occurs virtually over all of the seaward
1998 boundary of the swash zone, the most inshore portion of the swash zone undergoes
1999 significant, though non-uniform, aggradation (dark tones of grey). Bed discontinuities, typical
2000 of the evolution of an erosive bore, are seen to characterize both the uprush and the
2001 downrush stages.

2002 The increasing awareness of the fundamental role of the swash zone for the entire
2003 nearshore hydro-morphodynamics stimulates always new research focused on links
2004 between the surf zone and swash zone dynamics.

2005 From the hydrodynamic viewpoint, one of the most interesting recent studies is that by
2006 Moura and Baldock (2017), who focused on the relation between the shoreline (x_s) motion
2007 and the IG waves, both BP forced (horizontal BP excursion x_{br}) and BLW (vertical BLW
2008 excursion η_{BLW}), generated within the surf zone.

2009

2010

2011 Figure 26. Shoreline relation with BLW and BP forcing. (a) Wave group. The vertical blue
2012 line gives the BP between its outer (red), x_o , and inner (green), x_i location. Grey and red
2013 lines give the released BLW and incident BP-forced IG wave, respectively. (b) BP excursion
2014 (black) and shoreline response to BP-forced (red) and BLW (grey) IG waves. Dashed line
2015 gives path for a shallow water wave from the BP to the shoreline. Horizontal colored lines
2016 are BP positions as in Figure 2a. (c) Cross-correlation between BP and shoreline excursion,
2017 BLW released (grey) and BP-foced IG wave (red). Adapted from Moura and Baldock, (2017).

2018

2019

2020 Such relation can be illustrated with reference to the schematic of figure 26. When a group
2021 arrives at breaking (panel a) the first, smaller waves break at x_i (green line) and the
2022 subsequent, larger waves, more and more seaward, moving the BP to x_o (red line), the
2023 subsequent smaller waves force the BP to move back to x_i . Hence, the BP position $x_{br}(t)$ is
2024 – large (x_i), small(x_o), large (x_i) – i.e. in phase with the BLW vertical oscillation $\eta_{BLW}(t)$ (which
2025 is large, small, large, i.e. concave upwards). The amplitude of the BP-forced IG wave $\eta_{BP}(t)$
2026 is proportional to the width of the breaking region. Hence, it increases during the passage

2027 from small to large waves ($x_{br}(t)$ from zero to max) and decreases from large to small waves
2028 ($x_{br}(t)$ from max to zero). Hence, $\eta_{BP}(t)$ is concave downward or in phase opposition to $x_{br}(t)$
2029 and $\eta_{BLW}(t)$. The released BLW and the incident BP-forced wave propagate with the same
2030 speed (\sqrt{gh}) and reach the shoreline at the same time (after traveling through the surf zone),
2031 but with opposite phase (panel b). This leads to a positive correlation between $x_{s,BLW}$ and x_{br}
2032 and a negative correlation between $x_{s,BP}$ and x_{br} (panel c).

2033 On the basis of the above and using data coming from pressure sensors (water depth,
2034 significant wave height and period, etc.) and video observations (shoreline and BP
2035 excursion) collected at 3 Australian beaches, correlations between IG waves and shoreline
2036 excursion were made. The main findings suggest that: the shoreline excursion is a good
2037 proxy for infragravity waves in the inner surf and swash zone, IG wave generation by BLW
2038 release is stronger for conditions with relatively narrow surf zones and plunging waves while
2039 BP forcing is dominant for wider surf zones and spilling breaker conditions. This seems in
2040 contradiction with the established understanding (Battjes et. al, 2004) that the BP forcing is
2041 most efficient on steep beaches and the BLW release is most efficient on mildly-sloping
2042 beaches, see also the discussion provided in "Template models" of above. Such apparent
2043 contradiction can, perhaps, be solved by recalling that the surf zone width is not only function
2044 of the beach slope but also of the characteristics of the incoming wave. However, some
2045 further research seems needed here.

2046 From the morphological point of view, one fundamental link between the surf zone and
2047 swash zone dynamics is represented by the feeding of the most inshore sand bar (see also
2048 figures 4, 5, 11) by swash zone sediments. This has been recently investigated by Alsina et
2049 al. (2012), on the basis of dedicated laboratory experiments. Such experiments have been
2050 carried out by using flow conditions (surf zone characterized by plunging breakers, $\epsilon \approx 12-13$)
2051 that, during an initial stage of run of about 3500 waves, led to the generation, from an almost-
2052 planar beach, of a nearshore bar that slowly migrated off to sea (see blue-solid to green-

2053 dashed lines, and vertical red lines in top panel of figure 27). The run was then stopped and
2054 the beach profile manually reshaped to a milder slope only inside the swash zone (pink-solid
2055 line). The same flow conditions were subsequently run, but bar migration stopped.

2056
2057

2058 Figure 27. Cross-shore distribution of beach profile (top panel) and of cross-shore sediment
2059 transport rate (bottom panel) at various times before and after beach reshaping (adapted
2060 from Alsina et al., 2012).

2061
2062

2063 This behavior was attributed to a reduced sediment transport from the swash to the surf
2064 zone, because of the beach slope reduction forced in the swash zone. Such a conjecture
2065 was also verified by analyzing the sediment transport rate illustrated in the bottom panel of
2066 figure 27. This shows that, while before beach reshaping it was $Q_x < 0$, i.e. sediment transport
2067 rate was seaward, in the surf zone and inner swash zone, the sediment transport rate
2068 became vanishing to positive after beach reshaping. This is a clear indication of a major
2069 reduction of bar sediment feeding from the swash zone.

2070 The above observations can be interpreted on the basis of a simple mechanistic model:
2071 the reduction of the swash zone beach slope (β_{swash} decreasing) forces an increase in the
2072 natural swash period T_{swash} defined as (Baldock and Holmes, 1999):

2073
$$T_{swash} = \frac{2C\sqrt{gH_b}}{g \sin \beta}, \quad C = O(1) \quad (12)$$

2074 and, as a consequence, an increase in the swash-swash interactions, this being measured
2075 by the T_{swash}/T_i ratio (Baldock and Holmes, 1999); such increased swash-swash interaction
2076 makes the size of intense backwash events to decrease and, therefore, also the SST to sea,
2077 i.e. to a reduced sediment input to the nearshore bar (Alsina et al., 2012).

2078 This section is closed with some promising ongoing research aimed at including the
2079 swash zone dynamics into wave-averaged solvers for the nearshore circulation (see also
2080 “Generalities on Models” in the Introduction). By definition, wave-averaged models cannot
2081 describe the swash zone dynamics associated to the sea waves, this leading to problems in

2082 correctly predicting IG waves radiating out to sea. This is well illustrated by figure 22, whose
2083 left panels describe the interaction of groups of 5 sea waves at a rigid wall used in wave-
2084 averaged solvers to represent the shoreline, while the right panels describe the same
2085 interaction at a swash zone reproduced by means of a wave-resolving model (in the specific
2086 that of Brocchini et al., 2001). Two main features are very clear: the interaction at a wall,
2087 being a mere reflection of the incoming wave groups, leads to outgoing groups with the
2088 same structure of the incoming ones; the interaction at a swash zone allows for a significant
2089 swash-swash interaction, for example because of large waves catching up and engulfing
2090 small waves (see Mase, 1995 and figure 21), which contributes to increasing the intensity
2091 (S^-) of the IG waves radiated out to sea (in the specific 20% larger for the interaction within
2092 the swash zone). In view of the fundamental importance of wave-wave interaction within the
2093 swash zone Brocchini and co-workers have designed some Shoreline Boundary Conditions
2094 (SBCs) for the motion of the mean shoreline x_l and of the flow at at such boundary ($d(x_l)$,
2095 $u(x_l)$) on the basis of an integral swash zone model (Brocchini and Peregrine, 1996;
2096 Brocchini and Bellotti, 2002). Such SBCs have been successfully validated by means of
2097 large-scale laboratory experiments (Bellotti et al., 2003) and have recently been
2098 implemented in the ROMS wave-averaged solver (Haas and Warner, 2009). The main aim
2099 of such preliminary implementation was to gauge the capabilities of reproducing swash zone
2100 dynamics at a reduced computational cost, the final goal being that of predicting such
2101 dynamics by simply running a wave-averaged solution. Figure 28 summarizes the main
2102 findings of the mentioned implementation (Memmola, 2017; Memmola et al., 2019). The top
2103 six panels give the shoreline motion computed by running ROMS alone at the unreasonably-
2104 high resolution of 120 grid nodes (Δx) per offshore wavelength L_0 (black line), i.e. using it as
2105 a wave-resolving model (benchmark solution), and ROMS coupled with the proposed SBCs
2106 at lower and lower resolutions (red line), down to 2 grid nodes per wavelength. It is clear
2107 that coupling ROMS with the SBCs allows one to run very cheap calculations that predict

2108 reasonably well the benchmark solution by only using 3 grid nodes per wavelength (with $\Delta x/$
2109 $L_0 \approx 2$ the solution is too deteriorated). The lower two panels of the same figure show that not
2110 only the shoreline motion, but the entire flow field is well predicted by using the SBCs. The
2111 left and right panels give, respectively, the swash zone free surface elevation and onshore
2112 velocity for the benchmark solution (top panels) and ROMS+SBCs solutions for two very
2113 coarse meshes (middle and lower panels).

2114

2115

2116 Figure 28. Top 6 panels: comparison of the ROMS ultra-highly resolved shoreline (black line) and ROMS+SBCs shorelines for lower and lower grid resolution. Bottom 2 panels: free
2117 surface elevation (left) and onshore velocity (right) for the benchmark ROMS solution (top
2118 panels) and ROMS+SBCs solutions for two very coarse meshes (middle and lower panels).
2119 Adapted from Memmola (2017).
2120

2121

2122

2123 These preliminary results demonstrate that the SBCs of Brocchini and co-worker can, at
2124 least, allow for cheap solutions that also provide the entire structure of the swash zone flow.
2125 Work is currently underway to achieve the more ambitious goal of running the ROMS-SBCs
2126 system in a predicting mode (two-way coupled system), i.e. so that the SBCs are actually
2127 used to time-step the swash solution to be returned to ROMS.

2128

2129 Discussion

2130 An important contribution of the present work is the identification of nonlocal
2131 forcing/dynamics, i.e. those that influence the entire nearshore as a whole. These are, in a
2132 cascading order of mutual control the: beach slope (obvious nonlocal forcing), IG waves and
2133 large-scale vortices with vertical axes (macrovortices).

2134 The **beach slope** controls the role of IG waves, generation of macrovortices and the
2135 related sediment transport. In particular, BLW are regarded to be more efficient on mildly-
2136 sloping beaches, while IG waves are more likely generated by the oscillation of the BP on
2137 steep beaches (Battjes et al., 2004); although some recent studies by Moura and Baldock

2138 (2017) seem to provide different information. The generation of macrovortices is controlled
2139 by the beach slope through the longshore differential breaking of sea waves that leads to
2140 generation and reorganization of potential vorticity at the lareral edges of breaking wave
2141 fronts (e.g. Kennedy et al. 2006). Although the role of SST at the lower boundary of the
2142 swash zone can be regarded as a local effect (see above), changes in the swash zone
2143 beach slope can significantly alter the swash-swash interactions and intense backwash
2144 events, this either promoting or stopping the sediment transport needed for the formation
2145 and migration of nearshore sandbars (Alsina et al., 2012).

2146 **IG waves** provide a nonlocal forcing because the lengths involved in their generation and
2147 evolution are comparable to the surf zone width. IG are thought to have a fundamental role
2148 in the generation of sandbars, though various different models have been proposed like the
2149 “standing IG wave mechanism” (e.g. Short, 1975) and the “sea-IG waves correlation
2150 mechanism” of O’Hare and Huntley (1994) in which IG waves are responsible for the
2151 transport of sediments put into suspension by sea waves. Hence, the importance of the
2152 correlation between BLW and sea-wave envelope or the IG wave height to sea wave height
2153 ratio H_{IG}/H_{SW} , which is a good proxy for the sediment transport on beaches of all slopes (De
2154 Bakker et al., 2016). IG waves are also important for the riverine sediment transport because
2155 they can propagate upriver (e.g. Uncles et al, 2014), while sea and swell waves are
2156 dissipated by breaking at river mouths, which act as “low-pass filters”. IG waves are not only
2157 generated at the seaward boundary of the surf zone, but also in the very shallow waters of
2158 the swash zone. This occurs by the wave-wave interaction due to large waves traveling
2159 faster than small waves and catching them up in the swash zone (Mase, 1995).

2160 **Macrovortices** are a nonlocal forcing of the nearshore dynamics because they can
2161 maintain their coherent over large distances after their generation. Macrovortices contribute
2162 to sediment transport both by mobilizing the nearbed sands and, like IG waves, by carrying
2163 the sediments put into suspension by sea waves. The differential breaking that generates

2164 macrovortices is particularly intense at river mouths characterized by seabed shoals. Hence,
2165 macrovortices, beyond the river jet expansion, provide a significant contribution to both flow
2166 mixing and sediment transport at river mouths (Melito et al., 2018b).

2167 The analysis of the three subsystems of interest and of the related wave-induced
2168 dynamics suggests some lines for future research, proposed with reference to each of such
2169 subsystems.

2170 Inspection of the state-of-the-art sandbar modeling reveals that we are still in need of
2171 suitable stability models for the generation of 2D sandbars. As recalled in the analysis of
2172 sandbars, available numerical solutions for the generation of 2D sandbars are characterized
2173 by longshore instabilities in the form of rip channels (Dronen and Deigaard, 2007) and no
2174 stable 2D sandbar systems can actually be reproduced for either normal or oblique wave
2175 incidence. It seems, therefore, essential to investigate the mechanisms for the emergence
2176 of 3D instabilities, also inspecting the role of the longshore currents, which are believed to
2177 contribute to the maintenance of the stability of a 2D pattern (Shand et al., 1999).

2178 The role of longshore flows is becoming of growing interest, as believed to significantly
2179 control not only the generation, but also the migration of sandbars. In particular, longshore
2180 currents due to waves obliquely incident to the shore may have the potential to increase
2181 the shear stress acting on the seabed and responsible for the sediment mobility (Parlagreco
2182 et al., 2019). In this perspective longshore currents are not directly responsible for the
2183 direction of cross-shore migration of sandbars (onshore vs offshore) but provide an indirect
2184 important control through increased sediment mobility.

2185 Observed opposite migration behaviours on beaches of similar steepness and sediment
2186 size (Van Enckevort and Ruessink, 2003; Aagard et al., 2004) suggest that other important
2187 mechanisms that may characterize conditions for offshore (NOffM model) vs onshore
2188 (NOnM model) migration are the steepness and storm sequencing of the incident wave field.

2189 In brief, future studies on sandbar generation and evolution should pay great attention to
2190 **wave obliquity, steepness and storm sequencing.**

2191 A more detailed inspection of the steepness of the incident wave field and related
2192 modalities of breaking is also advisable to try to explain the apparent contradiction between
2193 the recent findings of Moura and Baldock, (2017), suggesting that IG wave generation by
2194 BLW release is stronger for narrow surf zones and plungers while BP forcing is dominant
2195 for wider surf zones and spillers, and the established understanding (Battjes et. al, 2004)
2196 that the BP forcing is most efficient on steep beaches and the BLW release is most efficient
2197 on mildly-sloping beaches. This investigation should also try to encompass an analysis of
2198 the different modes of transport of turbulence associated with the two mentioned breaking
2199 types (seaward under spilling breakers and landward under plunging breakers – Ting and
2200 Kirby, 1994).

2201 The implementation of **novel Shoreline Boundary Conditions** into wave averaging
2202 models for the nearshore circulation is certainly an important way forward for the inclusion
2203 of swash zone dynamics in long-term numerical simulations of the nearshore dynamics. The
2204 recent works of Memmola (2017) and Memmola et al. (2019) have shown the potentials of
2205 this approach, which will enable accurate and relatively cheap long-term numerical
2206 solutions. Work is currently underway for a predictive implementation of such approach.

2207 In relation to the interactions occurring at a river mouth, studies are needed to clarify the
2208 roles of the different wave modes acting there. Recent field observations by Melito et al.
2209 (2019) confirm that **the tide has an important role of modulating the IG waves** that
2210 propagate upstream the river, result already obtained through numerical means by
2211 Olabarrieta (2016). This modulating effect, considerable even for microtidal environments
2212 (e.g. Misa River), should be better analyzed through field studies carried out in macrotidal
2213 environments.

2214 The interactions of a river jet and the incoming sea waves should be better investigated
2215 also in relation to the generation of macrovortices. Preliminary evidence shows that they can
2216 remove significant amounts of energy from the incident wave field, reducing the opposing
2217 action of waves to the river stream (Melito et al., 2018b), and also force 3D seabed
2218 morphological features, e.g. mouth bars. Hence, both theoretical studies aimed at
2219 integrating the analytical approach of Brocchini (2013) for vorticity and potential vorticity
2220 generation with that of Falcini et al. (2014) as well as accurate morphological wave-resolving
2221 numerical simulations will likely clarify the role of ***breaking-wave-induced vorticity on the***
2222 ***river mouth morphodynamics***.

2223

2224 Conclusions

2225 The role of waves on the nearshore hydro-morphodynamics has been analyzed with specific
2226 focus to three subsystems of the nearshore i.e. sandbar region, river mouths and swash
2227 zone.

2228 Each of these is characterized by very specific local dynamics. For sandbar generation
2229 and evolution fundamental is the role the nearbed sediment transport, with the correlation
2230 between the orbital velocity and sediment concentration in the BBL being of paramount
2231 importance for suitably predicting onshore bar migration (e.g. Henderson et al., 2004). It has
2232 also been shown that energetics-type models can achieve similarly good results if they
2233 reproduce such correlation through an integral approach, i.e. by including the effects of the
2234 wave freestream acceleration skewness (e.g. Hoefel and Elgar, 2002).

2235 Local dynamics specific to river mouths are: the wave-current interaction by which the
2236 river current steepens the incoming waves, if intense enough to the point of breaking and/or
2237 blocking them, and forces a frequency downshifting (Chawla and Kirby, 2002); the related
2238 formation of a mouth bar (Fagherazzi et al., 2015) or the formation of a shoal that largely

2239 influences the local dynamics by forcing stronger wave-breaking-induced return currents
2240 (Olabarrieta et al., 2014) and wave breaking/sediment deposit farther from the mouth, i.e. at
2241 the seaward edge of the shoal.

2242 BBL dynamics largely influence the very shallow flows of the swash zone and, like for
2243 sandbars, a proper modeling of the swash zone sediment transport can only be achieved if
2244 correlation between the orbital velocity and sediment concentration in the BBL is well
2245 described. Fundamental is also the description of the SST transport at the lower boundary
2246 of the swash zone, which affects the sediment fluxes both from the surf zone, through
2247 sediment presuspension at bore collapse (Pritchard and Hogg, 2005), and to the surf zone,
2248 in association to intense backwash events.

2249 A closure can be reached by recalling the main connections among the three subsystems
2250 of interest. IG waves provide a fundamental connection because generated both at the
2251 seaward edge of the surf zone (by BLW release, BP oscillation and edge wave interaction)
2252 and at the swash zone (because of wave-wave interaction) can propagate almost unaltered
2253 upstream at river mouths. Another important connection is provided by the evolution of
2254 macrovortices, which generated by differential breaking move over long distances in the surf
2255 zone and may reach the shoreline where they can rebound and alter the near-shoreline
2256 morphology. From the morphological viewpoint, obvious links are those between: the swash
2257 zone morphology and the along-river-flank sediment transport that can feed the swash zone
2258 and the swash zone and the nearshore sandbars, which can drive sediments for their
2259 evolution from the swash zone.

2260

2261 Acknowledgements

2262 This paper is essentially a written transposition of the same-titled invited talk given at
2263 RCEM2017. The organizers are thanked for the opportunity given to reflect on and present
2264 this material. The following colleagues and friends are thanked for providing both useful

2265 material and insightful discussions: Tom Baldock (University of Queensland, Australia),
2266 Riccardo Briganti (University of Nottingham, England), Joe Calantoni (Naval Research
2267 Laboratory, USA), Dan Conley (University of Plymouth, England), Steve Elgar (Woods Hole
2268 Oceanographic Institution, USA), Tom Hsu (University of Delaware, USA), Matteo
2269 Postacchini (Università Politecnica delle Marche, Italy), Gerben Ruessink (Utrecht
2270 University, The Netherlands), Alex Sheremet (University of Florida, USA). Financial support
2271 from the ONR Global (UK), through the MORSE Research Grant (N62909-17-1-2148) is
2272 gratefully acknowledged.

2273

2274 References

- 2275 Aagard, T., Davidson-Arnott, R., Greenwood, B., Nielsen, J. (2004). Sediment supply from
2276 shoreface to dunes: linking sediment transport measurements and long-term
2277 morphological evolution. *Geomorphology* **60**: 205-224.
- 2278 Alsina, J.M., Caceres, I., Brocchini, M., Baldock, T.E. (2012). An experimental study on
2279 sediment transport and bed evolution under different swash zone morphological
2280 conditions. *Coastal Engineering* **68**: 31-43.
- 2281 Anthony, E.J. (2015). Wave influence in the construction, shaping and destruction of river
2282 deltas: A review. *Marine Geology* **361**: 53-78.
- 2283 Antuono, M., Brocchini, M. (2010). Solving the Nonlinear Shallow Water Equations in
2284 physical space. *Journal of Fluid Mechanics* **643**: 207-232.
- 2285 Armaroli, C., Ciavola, P. (2011). Dynamics of a nearshore bar system in the northern
2286 Adriatic: A video-based morphological classification. *Geomorphology* **126**: 201-216.
- 2287 Ashton, A.D., Giosan, L. (2011). Wave-angle control of delta evolution. *Geophysical*
2288 *Research Letters* **38**: L13405.
- 2289 Bailard, J.A. (1981). An energetics total load sediment transport model for a plane sloping
2290 beach. *Journal of Geophysical Research – Oceans* **86(C11)**: 10938-10954.

2291 Baldock, T.E. (2006). Long wave generation by the shoaling and breaking of transient wave
2292 groups on a beach. *Proceedings of the Royal Society of London A* **462**: 1853-1876.

2293 Baldock, T.E., Holmes, P. (1999). Simulation and prediction of swash oscillations on a steep
2294 beach. *Coastal Engineering* **36 (3)**: 219–242.

2295 Battjes, J.A., Bakkenes, H.J., Janssen, T.T., van Dongeren, A.R. (2004). Shoaling of
2296 subharmonic gravity waves. *Journal of Geophysical Research - Oceans* **109**: C02009.

2297 Bellotti, G., Archetti, R., Brocchini, M. (2003). Experimental validation of mean swash zone
2298 boundary conditions, *Journal of Geophysical Research - Oceans* **108(C8)**: 3250,
2299 doi:10.1029/2002JC001510.

2300 Berni, C., Barthelemy, E., Michallet, H. (2013). Surf zone cross-shore boundary layer
2301 velocity asymmetry and skewness: an experimental study on a mobile bed. *Journal of*
2302 *Geophysical Research - Oceans* **118**: 2188–2200.

2303 Bertin, X., Olabarrieta, M. (2015). Relevance of infragravity waves in a wave-dominated
2304 inlet. *Journal of Geophysical Research - Oceans* **121**: 5418–5435.

2305 Blenkinsopp, C.E., Turner, I.L., Masselink, G., Russell, P.E. (2011). Swash zone sediment
2306 fluxes — field observations. *Coastal Engineering* **58**: 28–44.

2307 Brocchini, M. (2006). Integral swash zone models. *Continental Shelf Research* **26**: 653-660.

2308 Brocchini, M. (2013). A reasoned overview on Boussinesq-type models: the interplay
2309 between physics, mathematics and numerics. *Proceedings of the Royal Society of*
2310 *London A* **469**: 20130496, doi:10.1098/rspa.2013.0496.

2311 Brocchini, M., Peregrine, D.H. (1996). Integral flow properties of the swash zone and
2312 averaging. *Journal of Fluid Mechanics* **317**: 241-273.

2313 Brocchini, B., Bellotti, G. (2002). Integral flow properties of the swash zone and averaging.
2314 Part 2. The shoreline boundary conditions for wave-averaged models. *Journal of Fluid*
2315 *Mechanics* **458**: 269-281.

2316 Brocchini, M., Baldock, T.E. (2008). Recent advances in modeling swash zone dynamics:
2317 influence of surf-swash interaction on nearshore hydrodynamics and morphodynamics.
2318 *Reviews in Geophysics* **46(3)**: RG3003, doi:10.1029/2006RG000215.

2319 Brocchini, M., Calantoni, J., Postacchini, M., Sheremet, A., Staples, T., Smith, J., Reed,
2320 A.H., Braithwaite III, E.F., Lorenzoni, C., Russo, A., Corvaro, S., Mancinelli, A. and
2321 Soldini, L. (2017). Comparison between the wintertime and summertime dynamics of the
2322 Misa River estuary. *Marine Geology* **385**: 27-40.

2323 Caballeria, M., Coco, G., Falqués, A., Huntley, D.A., 2002. Self-organization mechanisms
2324 for the formation of nearshore crescentic sand bars. *Journal of Fluid Mechanics* **465**: 379–
2325 410.

2326 Canestrelli, A., Nardin, W., Edmonds, D., Fagherazzi, S., Slingerland, R. (2014), Importance
2327 of frictional effects and jet instability on the morphodynamics of river mouth bars and
2328 levees, *Journal of Geophysical Research - Oceans* **119**: 509–522.

2329 Certain, R., Barusseau, J.P. (2005). Conceptual modelling of sand bars morphodynamics
2330 for a microtidal beach (Sète, France). *Bulletin de la Société Géologique de France* **176**:
2331 343–354.

2332 Chang, Y.S., Scotti, A. (2004). Modeling unsteady turbulent flows over ripples: Reynolds-
2333 averaged Navier-Stokes equations (RANS) versus large-eddy simulation (LES). *Journal*
2334 *of Geophysical Research - Oceans* **109**: 1-16.

2335 Chawla, A., Kirby, J.T. (2002). Monochromatic and random wave breaking at blocking
2336 points. *Journal of Geophysical Research - Oceans* **107**: 3067.

2337 Coco G., Murray, A.B. (2007). Patterns in the sand: from forcing templates to self-
2338 organization. *Geomorphology* **91**: 271-290.

2339 Cooper, J.A.G. (2001). Geomorphological variability among microtidal estuaries from the
2340 wave-dominated South African coast. *Geomorphology* **40**: 99-122.

2341 de Bakker, A.T.M., Brinkkemper, J.A., van der Steen, F., Tissier, M.F.S., Ruessink B.G.
2342 (2016). Cross-shore sand transport by infragravity waves as a function of beach
2343 steepness. *Journal of Geophysical Research – Earth Science* **121**: 1786-1799.

2344 Drake, T.G., J. Calantoni (2001). Discrete particle model for sheet flow sediment transport
2345 in the nearshore. *Journal of Geophysical Research - Oceans* **106**: 19,859–19,868.

2346 Dronen, N., Deigaard, R. (2007). Quasi-three-dimensional modelling of the morphology of
2347 longshore bars. *Coastal Engineering* **54**: 197-215.

2348 van Enckevort, I.M.J., Ruessink, B.G. (2003). Video observations of nearshore bar
2349 behaviour. Part 1: alongshore uniform variability. *Continental Shelf Research* **23**: 501-
2350 512.

2351 Fagherazzi, S., Edmonds, D.A., Nardin, W., Leonardi, N., Canestrelli, A., Falcini, F.,
2352 Jerolmack, D.J., Mariotti, G., Rowland, J.C., Slingerland, R.L. (2015). Dynamics of river
2353 mouth deposits. *Reviews in Geophysics* **53**, 31pp., doi:10.1002/2014RG000451.

2354 Falcini, F., Piliouras, A., Garra, R., Guerin, A., Jerolmack, D.J., Rowland, J., Paola, C. (2014)
2355 Hydrodynamic and suspended sediment transport controls on river mouth morphology
2356 *Journal of Geophysical Research – Earth Surface* **119**: 1-11, doi:10.1002/2013JF002831.

2357 Gallagher, E., Elgar, S., Guza, R.T. (1998). Observations of sand bar evolution on a natural
2358 beach. *Journal of Geophysical Research - Oceans* **103(C2)**: 3203-3215.

2359 Galloway, W.E. (1975). Process framework for describing themorphologic and stratigraphic
2360 evolution of delta depositional systems. In: Broussard, M.L (Ed.), *Deltas: Models for*
2361 *Exploration*. Houston Geological Society, Houston, pp. 87–98.

2362 Guard, P., Baldock, T. (2007). The influence of seaward boundary conditions on swash zone
2363 hydrodynamics. *Coastal Engineering* **54**: 321–331.

2364 Haas, K.A. and Warner, J.C. (2009). Comparing a quasi-3D to a full 3D nearshore circulation
2365 model: SHORECIRC and ROMS. *Ocean Modelling* **26**: 91-103.

2366 Henderson, S.M., Allen, J.S., Newberger P.A. (2004). Nearshore sandbar migration
2367 predicted by an eddy-diffusive boundary layer model. *Journal of Geophysical Research -*
2368 *Oceans* **109**: C06024, doi:10.1029/2003JC002137.

2369 Hoefel, F., Elgar, S. (2003). Wave-induced sediment transport and sandbar migration.
2370 *Science* **299**: 1885-1887.

2371 Holman, R.A., Bowen, J. (1982). Bars, bumps, and holes: models for the generation of
2372 complex beach topography. *Journal of Geophysical Research - Oceans* **87**: 457-468.

2373 Hsu, T., Elgar, S., Guza, R.T. (2006). Wave-induced sediment transport and onshore
2374 sandbar migration. *Coastal Engineering* **53**: 817–824.

2375 Kennedy, A.B., Brocchini, M., Soldini, L., Gutierrez, E. (2006). Topographically controlled,
2376 breaking-wave-induced macrovortices. Part 2. Rip current topographies. *Journal of Fluid*
2377 *Mechanics* **559**: 57–80.

2378 Kirby, J.T. (2016). Boussinesq models and their application to coastal processes across a
2379 wide range of scales. *Journal of Waterways, Port, Coastal and Ocean Engineering* **142**:
2380 03116005.

2381 Lesser, G.R., Roelvink, J.A., van Kester, J.A.T.M., Stelling, G.S. (2004). Development and
2382 validation of a three-dimensional morphological model. *Coastal Engineering* **51**: 883-915.

2383 Longuet-Higgins, M. S., Stewart, R. W. (1964). Radiation stress in water waves: a physical
2384 discussion with applications. *Deep Sea Research I* **11**: 529-562.

2385 Lubin, P., Glockner, S., Kimmoun, O., Branger, H. (2011). Numerical study of the
2386 hydrodynamics of regular waves breaking over a sloping beach. *European Journal of*
2387 *Mechanics B/Fluids* **30**: 552-564.

2388 Ma, G., Shi, F., Kirby, J.T. (2012). Shock-capturing non-hydrostatic model for fully dispersive
2389 surface wave processes. *Ocean Modelling* **43-44**: 22-35.

2390 Mase, H. (1995). Frequency down-shift of swash oscillations compared to incident waves.
2391 *Journal of Hydraulic Research* **33**: 397-411.

2392 Masselink, G., Short, A.D. (1993). The Effect of Tide Range on Beach Morphodynamics and
2393 Morphology: A Conceptual Beach Model. *Journal of Coastal Research* **9**: 785-800.

2394 Melito, L., Postacchini, M., Sheremet, A., Calantoni, J., Zitti, G., Darvini, G., Brocchini, M.
2395 (2018a). Wave-current interactions and infragravity wave propagation at a microtidal inlet.
2396 *E-proceedings of the 3rd EWAS International Conference*: 477-485.

2397 Melito, L., Postacchini, G, Darvini, G., Brocchini, M. (2018b). Waves and currents at a river
2398 mouth: the role of macrovortices, sub-grid turbulence and seabed friction. *Water* **10**: 550,
2399 doi:10.3390/w10050550 .

2400 Melito, L., Postacchini, M., Sheremet, A., Calantoni, J., Zitti, G., Darvini G., Brocchini M.
2401 (2019l). Hydrodynamics at a microtidal inlet: analysis of propagation of the main wave
2402 components. *Estuarine Coastal and Shelf Science* (under peer review).

2403 Memmola, F. (2017). Analysis and development of oceanographic models: reaching the
2404 swash zone. Università Politecnica delle Marche, PhD Thesis, 92pp.

2405 Memmola, F., Coluccelli, A., Russo, A., Warner J.C., Brocchini M. (2019). Wave-resolving
2406 Shoreline Boundary Conditions for wave-averaged coastal models. *Ocean Modeling*
2407 (under peer review).

2408 Moura, T., Baldock, T.E. (2017). Remote sensing of the correlation between breakpoint
2409 oscillations and infragravity waves in the surf and swash zone. *Journal of Geophysical*
2410 *Research - Oceans* **122**: 3106-3122.

2411 Nienhuis, J.H. (2016). Plan-view evolution of wave-dominated deltas. Massachusetts
2412 Institute of Technology and the Woods Hole Oceanographic Institution, PhD Thesis,
2413 182pp.

2414 Nienhuis, J.H., Ashton, A.D., Giosan, L. (2015). What makes a delta wave-dominated?
2415 *Geology* **43(6)**: 511-514.

2416 Nienhuis, J.H., Ashton, A.D., Giosan, L. (2016). Littoral steering of deltaic channels. *Earth*
2417 *and Planetary Science Letters* **453(1)**: 204-214.

2418 O'Hare, T.J., Huntley, D.A. (1994). Bar formation due to wave groups and associated long
2419 waves. *Marine Geology* **116**: 313-325.

2420 Olabarrieta, M., Geyer, W.R., Kumar, N. (2014). The role of morphology and wave-current
2421 interaction at tidal inlets: An idealized modeling analysis. *Journal of Geophysical*
2422 *Research – Oceans* **119**: 8818-8837.

2423 Parlagreco, L. Melito, L. Devoti, S. Perugini, E. Soldini, L. Zitti G., Brocchini M. (2019).
2424 Monitoring for coastal resilience: preliminary data from five Italian sandy beaches. **19**.
2425 1854; doi:10.3390/s19081854

2426 Peregrine, D.H. (1999). Large-scale vorticity generation by breakers in shallow and deep
2427 water. *European Journal of Mechanics B/Fluids* **18**: 403–408.

2428 Plant, N.G., Freilich, M.H., Holman, R.A. (2001). Role of morphologic feedback in surf zones
2429 and bar response. *Journal of Geophysical Research - Oceans* **106**: 973-989.

2430 Postacchini, M., Brocchini, M., Landon, M., Mancinelli, A. (2012). A multi-purpose, intra-
2431 wave, shallow water hydro-morphodynamic solver. *Advances in Water Resources* **38**: 13-
2432 26.

2433 Postacchini, M., Brocchini, M., Soldini, L. (2014). Vorticity generation due to cross-sea.
2434 *Journal of Fluid Mechanics* **744**: 286-309.

2435 Pourzangbar, A., Losada, M.A., Saber, A., Rasoul Ahari, L., Larroudé, P., Vaezi, M. and
2436 Brocchini, M. (2017). Prediction of non-breaking wave induced scour depth at the trunk
2437 section of breakwaters using Genetic Programming and Artificial Neural Networks.
2438 *Coastal Engineering* **121**: 107-118.

2439 Power, H.E., Holman, R.A., Baldock, T.E. (2011). Swash zone boundary conditions derived
2440 from optical remote sensing of swash zone flow patterns. *Journal of Geophysical*
2441 *Research - Oceans* **116**: C06007, doi:10.1029/2010JC006724.

2442 Pritchard, D., Hogg, A.J. (2005). On the transport of suspended sediment by a swash event
2443 on a plane beach, *Coastal Engineering* **52(1)**: 1–23.

2444 Ribas, F. (2003). On the growth of nearshore sand bars as instability processes of
2445 equilibrium beach states. Universitat Politecnica de Catalunya, PhD Thesis, 242pp.

2446 Ribeiro, M., Barrio, F., Taborda, R., Cascalho, J., Bosnic, I. Rodríguez, I., Sanchez, M.
2447 (2012). Longshore drift: experimental and empirical predictors. The example of the Ebro
2448 Delta. *Proceedings of the 2as Jornadas de Engenharia Hidrográfica*, Lisboa, Portugal,
2449 pp. 271–274.

2450 Roelvink, J. A., Stive, M. J. F. (1989) Bar-generating cross-shore flow mechanisms on a
2451 beach. *Journal of Geophysical Research - Oceans* **94**: 4785-4800.

2452 Ruessink, B.G., Kroon, A. (1994). The behaviour of a multiple bar system in the nearshore
2453 zone of Terschelling, the Netherlands: 1965–1993. *Marine Geology* **121**: 187–197.

2454 Schäffer, H.A. (1993). Infragravity waves induced by short wave groups. *Journal of Fluid*
2455 *Mechanics* **247**: 551–588.

2456 Shand, R.D., Bailey, D.G., Shephard, M.J. (1999). An inter-site comparison of net offshore
2457 bar migration characteristics and environmental conditions. *Journal of Coastal Research*
2458 **15**: 750–765.

2459 Shen, M.C., Meyer, E.R. (1963). Climb of a bore on a beach. Part 3. Run-up. *Journal of*
2460 *Fluid Mechanics* **16**: 113–125.

2461 Short, A.D. (1975). Multiple offshore bars and standing waves. *Journal of Geophysical*
2462 *Research - Oceans* **80**: 3838-3840.

2463 Short, A.D., Aagard, T. (1993). Single and multi-bar beach change models. *Journal of*
2464 *Coastal Research* **15**: 141-157.

2465 Symonds, G., Huntley, D.A., Bowen, A.J. (1982). Two-dimensional surf beat: long wave
2466 generation by a time-varying breakpoint. *Journal of Geophysical Research - Oceans* **87**:
2467 492-498.

2468 Ting, F.C.K., Kirby, J.T. (1994). Observation of undertow and turbulence in a laboratory surf
2469 zone. *Coastal Engineering* **24**: 51-80.

2470 Torres-Freyermuth, A., Lara, J.L. and Losada, I.J. (2010). Numerical modelling of short- and
2471 long-wave transformation on a barred beach. *Coastal Engineering* **57**: 317-330.

2472 Uncles, R.J. Stephens, J.A., Harris C. (2014). Infragravity currents in a small ría: Estuary-
2473 amplified coastal edge waves? *Estuarine Coastal and Shelf Science* **150**: 242-251.

2474 Valle-Levinson A. (2011). Classification of Estuarine Circulation. In: Wolanski E and
2475 McLusky DS (eds.) *Treatise on Estuarine and Coastal Science* **1**, 75–86. Waltham:
2476 Academic Press.

2477 Van Enckevort, I.M.J. (2001). Daily to yearly nearshore bar behaviour. Utrecht University,
2478 PhD Thesis, 174pp.

2479 Van Enckevort, I.M.J., Ruessink, B.G. (2003). Video observations of nearshore bar
2480 behaviour. Part 1: alongshore uniform variability. *Continental Shelf Research* **23**: 501–
2481 512.

2482 Vittori, G., DeSwart, H.E., Blondeaux, P. (1999). Crescentic bedforms in the nearshore
2483 region. *Journal of Fluid Mechanics* **381**: 271-303.

2484 Watson, G., Barnes, T. C. D., Peregrine, D. H. (1995). The generation of low-frequency
2485 waves by a single wave group incident on a beach. *24th International Conference on*
2486 *Coastal Engineering*, 776-790.

2487 Wijnberg, K.M., Terwindt, J.H.J. (1995). Extracting decadal morphological behavior from
2488 high-resolution, long-term bathymetric surveys along the Holland coast using
2489 eigenfunction analysis. *Marine Geology* **126**: 301-330.

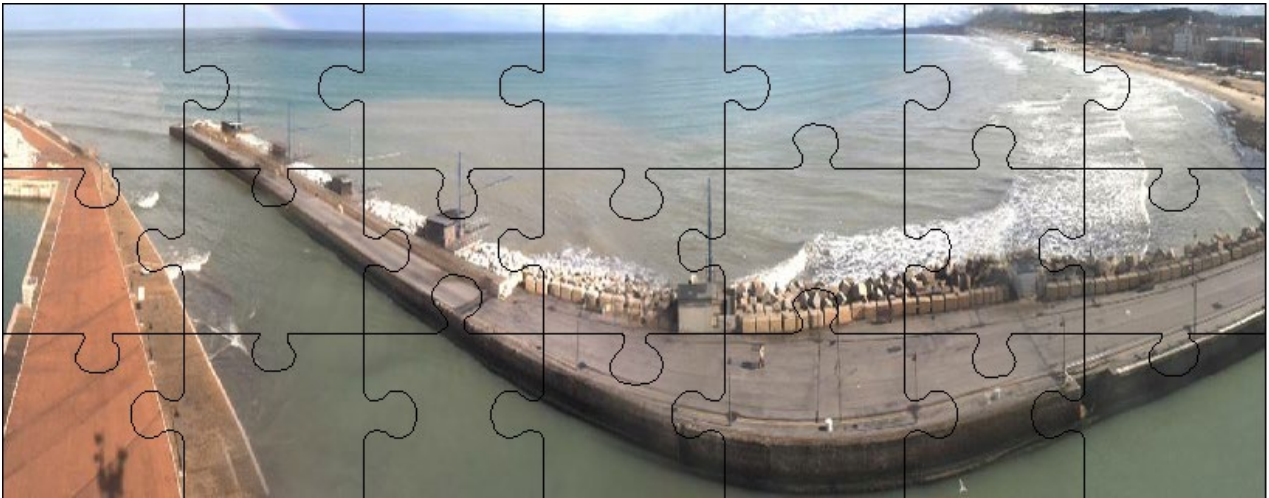
2490 Wijnberg, K.M., Kroon, A. (2002). Barred beaches. *Geomorphology* **48**: 103-120.

2491 Wright, L.D., Coleman, J.M., Erickson, M.W., (1974). Analysis of major river systems and
2492 their deltas: morphologic and process comparisons. Technical Report 156, Coastal
2493 Studies Institute. Louisiana State University (125 pp.).

2494 Wright, L.D., Short, A.D. 1984. Morphodynamic variability of surf zones and beaches: A
2495 synthesis. *Marine Geology* **56**: 93-118.

2497

LIST OF FIGURES



2498

2499 Figure 1. Wave-forced dynamics in the nearshore and estuaries: a puzzle to be solved.

2500

2501

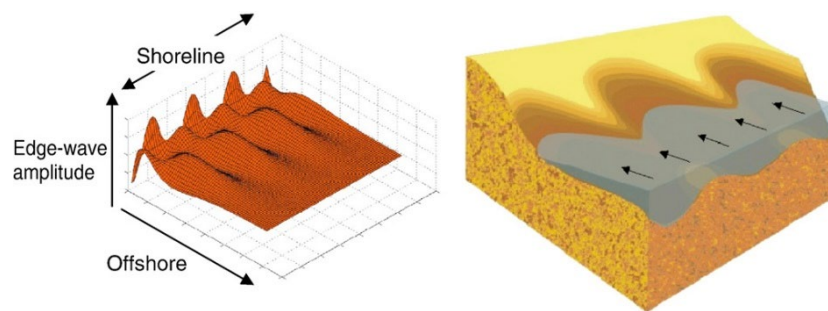
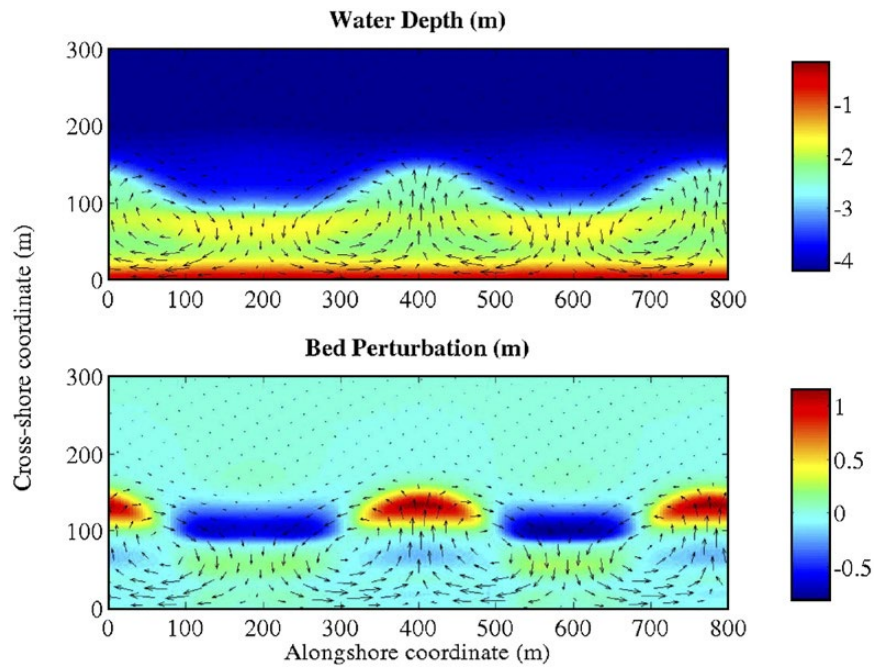


Figure 2. Forced/template mechanism: “synchronous standing edge waves” (left) and flow imprinted on sand giving beach cusps (right). Adapted from Coco and Murray, (2007).

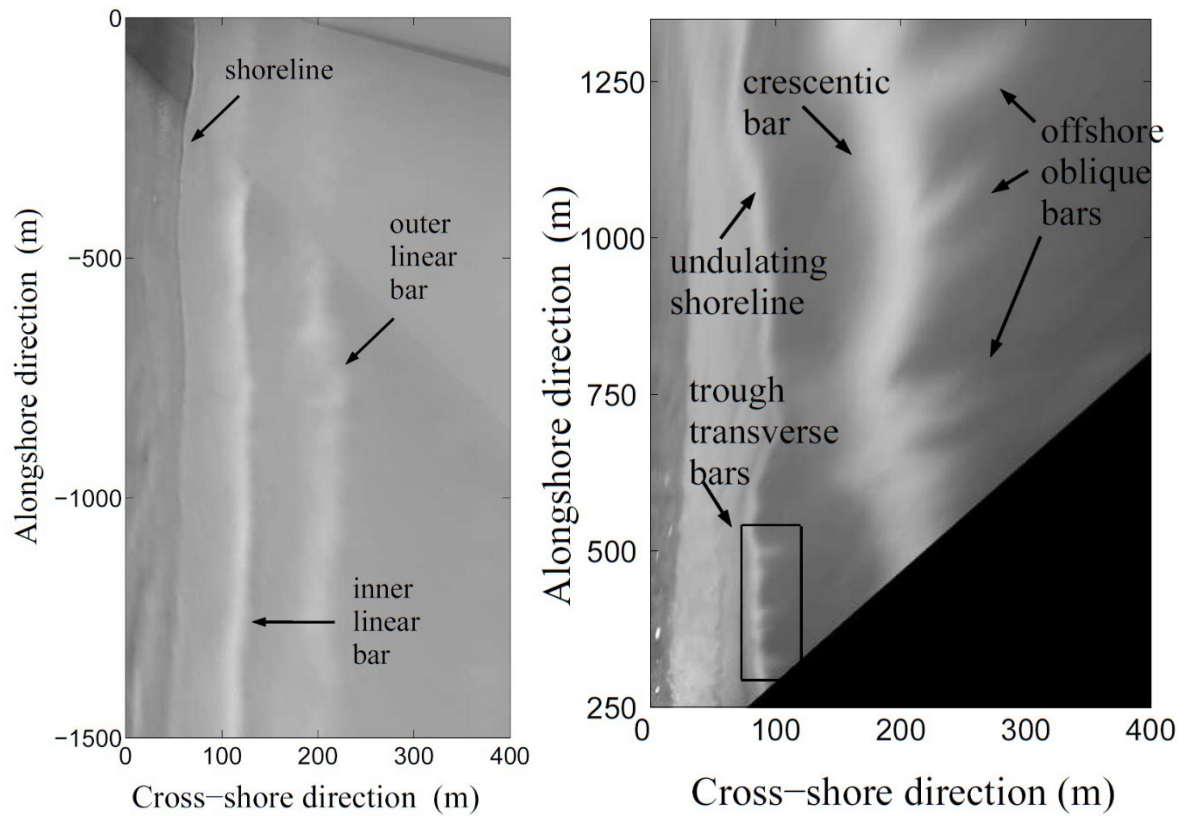
2502

2503



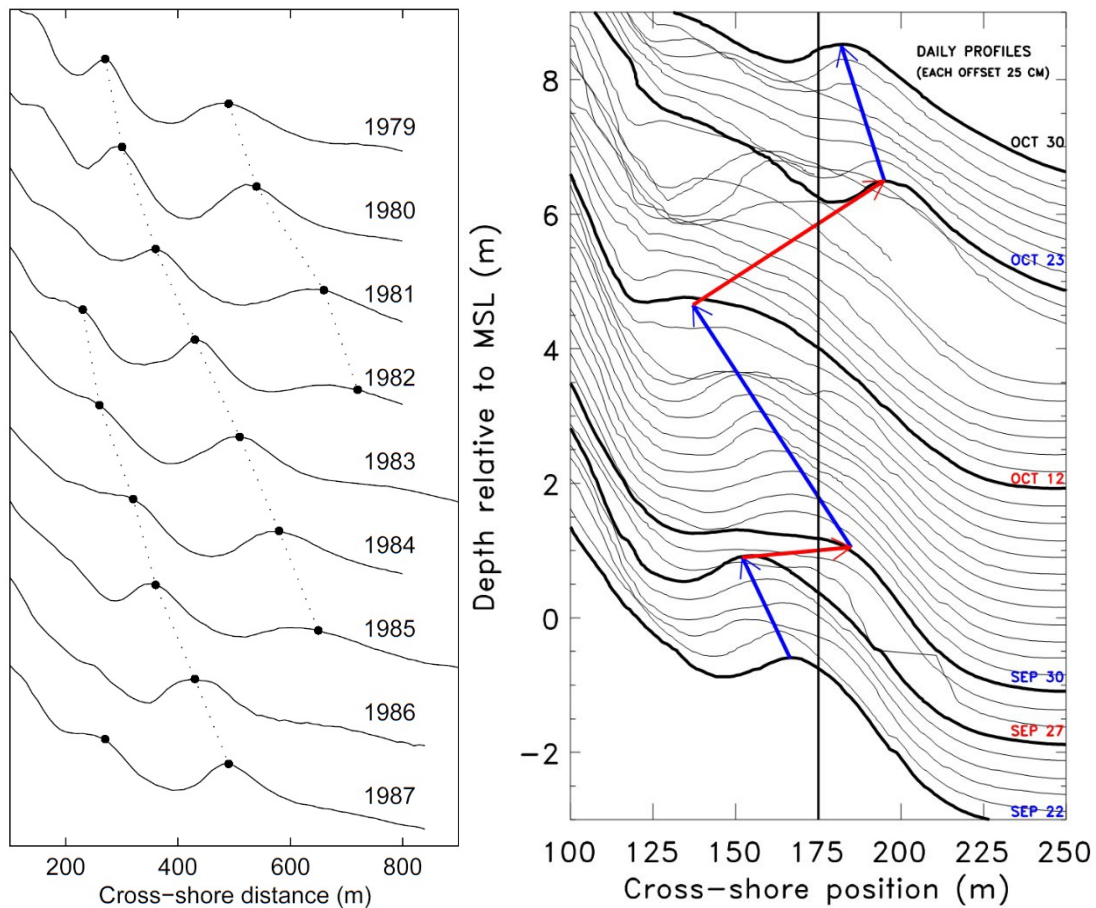
2504

2505 Figure 3. Stability/Self-organization mechanism: crescentic bars emerge from the feedback
 2506 between flow and bathymetry (initial longshore-uniform bar bathymetry). Top: waves
 2507 breaking on shoals force onshore flow, which returns to sea in gaps. Bottom: for
 2508 suspended sediment decreasing from breaker line to shore, the onshore flow favors
 2509 deposition (shoals grow) and the offshore flow favors erosion (scour holes grow). Hence,
 2510 the development of crescentic bars. Adapted from Coco and Murray, (2007).



2511

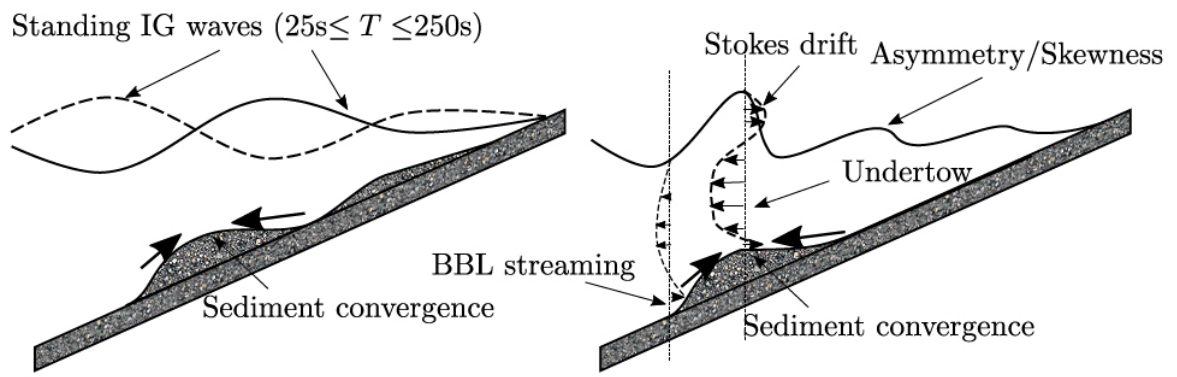
2512 Figure 4. Left panel: planview image of two shore-parallel sand bars at Noordwijk beach
 2513 (The Netherlands, 19th November 2000). Right panel: Planview image of surf zone
 2514 complex topography at Duck (U.S.A. Atlantic coast, 10th January 1994). Adapted
 2515 from Ribas Prats (2003).



2516

2517 Figure 5. Left panel: seabed cross-shore profiles at Noordwijk, The Netherlands, from 1979
 2518 to 1987, showing cyclic bar behavior with a recurrence interval of about 4-5 years.
 2519 Adapted from van Enckvort, (2001). Right panel: seabed cross-shore profiles at
 2520 Duck (U.S.A. Atlantic coast, September-October 1994). Thick curves with red
 2521 dates are the beginning of storms, resulting in offshore bar migration (red arrows).
 2522 Thin curves are the subsequent daily profiles. The sandbar migrated onshore (blue
 2523 arrows) during calm periods between the storms. Courtesy of Dr. S. Elgar.

2524



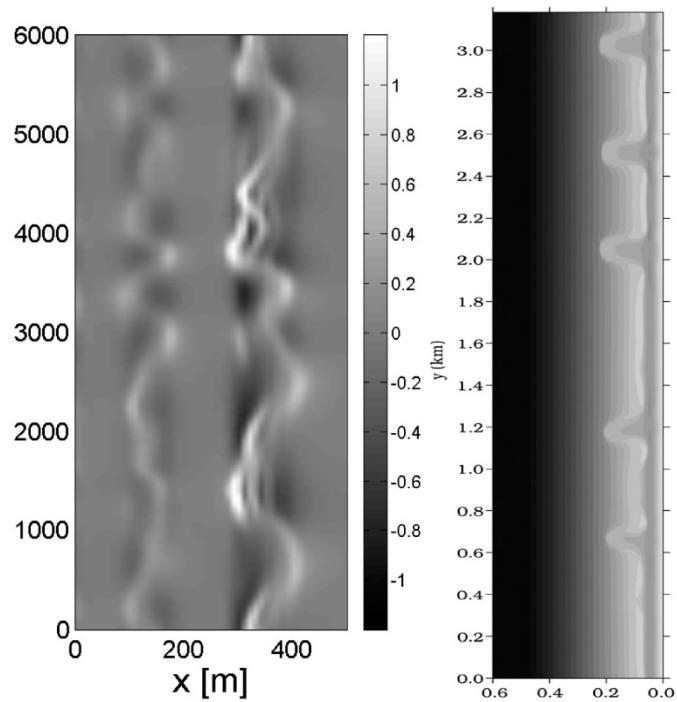
2525

2526

Figure 6. Bar formation because of sediment convergence at: anti/nodes of standing IF waves (left panel), breakpoint (right panel).

2527

2528



2529

2530

Figure 7. Left panel: non-linear stability solution for a double sand bar system giving asymmetric crescentic patterns on inner and outer bars (adapted from Klein and Schuttelaars, 2006). Right panel: Evolution of longshore uniform sand bar after 30 days of model run with normal wave incidence. Longshore instability in the form of rip channels (adapted from Dronen and Deigaard, 2007).

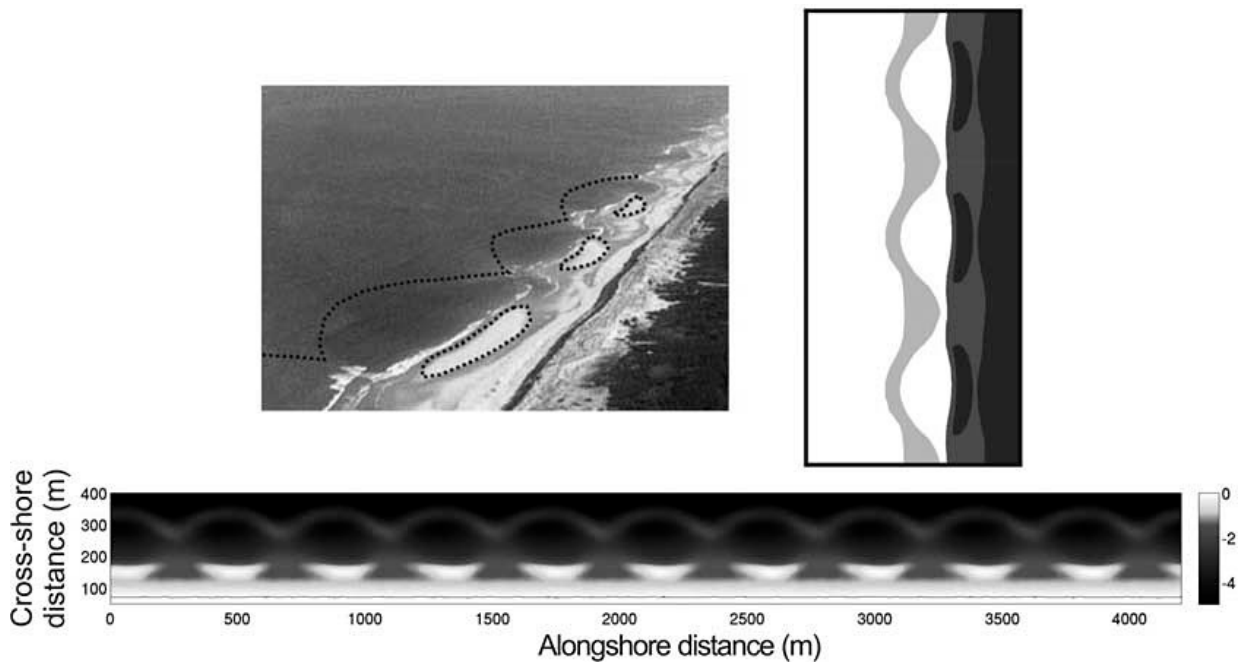
2531

2532

2533

2534

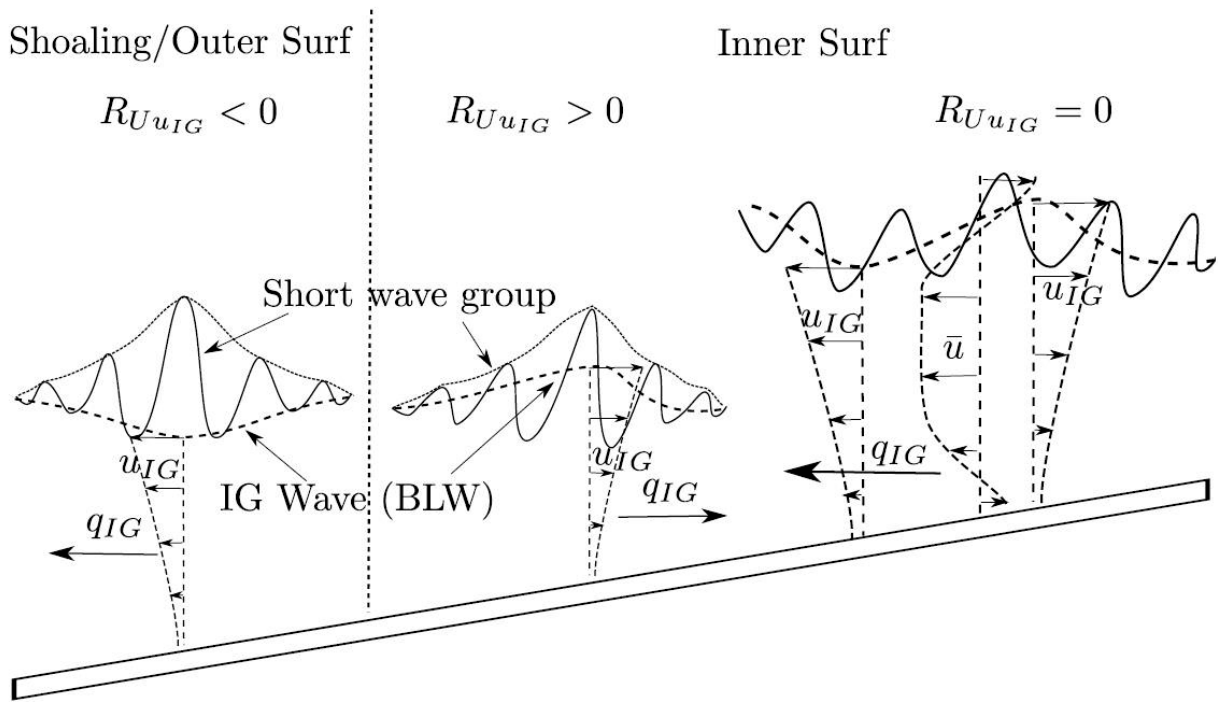
2535



2536

2537 Figure 8. Computed (top-right and bottom panels) 180° out of-phase coupled patterns with
 2538 an outer-bar horn facing an inner-bar bay, similar to observations (top-left panel)
 2539 on a meso-macro-tidal, high-energy, double-barred beach (adapted from Castelle
 2540 et al., 2010).

2541

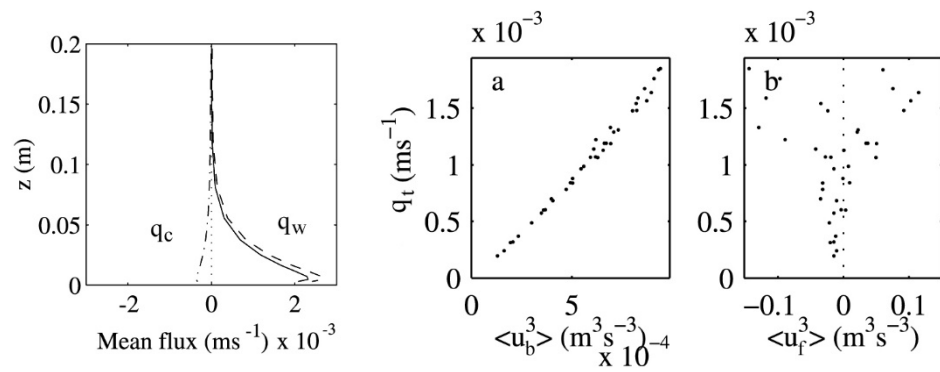


2542

2543 Figure 9. Sketch of IG-sea wave interaction and cross-shore sediment transport in the
 2544 nearshore. From left to right: shoaling zone, inner-surf zone of moderately-sloping
 2545 beaches, inner-surf zone of mildly-sloping beaches.

2546

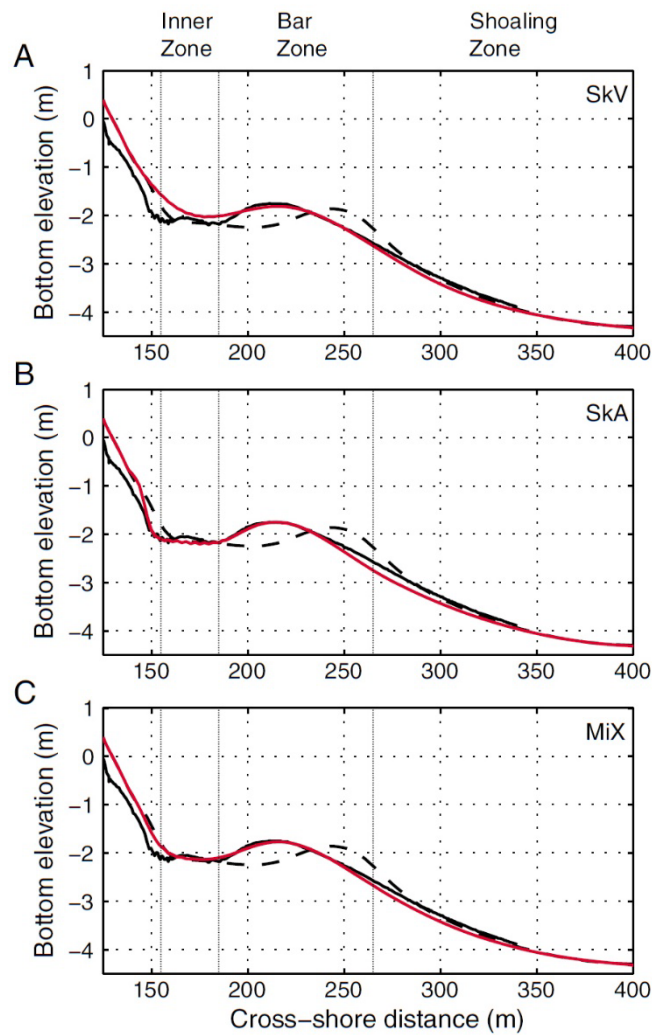
2547



2548

2549 Figure 10. Top: observed (lines) and predicted (symbols) seabed profiles at the
2550 beginning (dashed line) and end (solid line and symbols) of the Duck94
2551 onshore bar migration event. Bottom, from left to right: vertical profile of mean
2552 cross-shore sediment transport due to waves (q_w) and currents (q_c), total
2553 cross-shore sediment transport as function of near-bed (middle) and
2554 freestream (right) velocity moments. Adapted from Henderson et al. (2004).

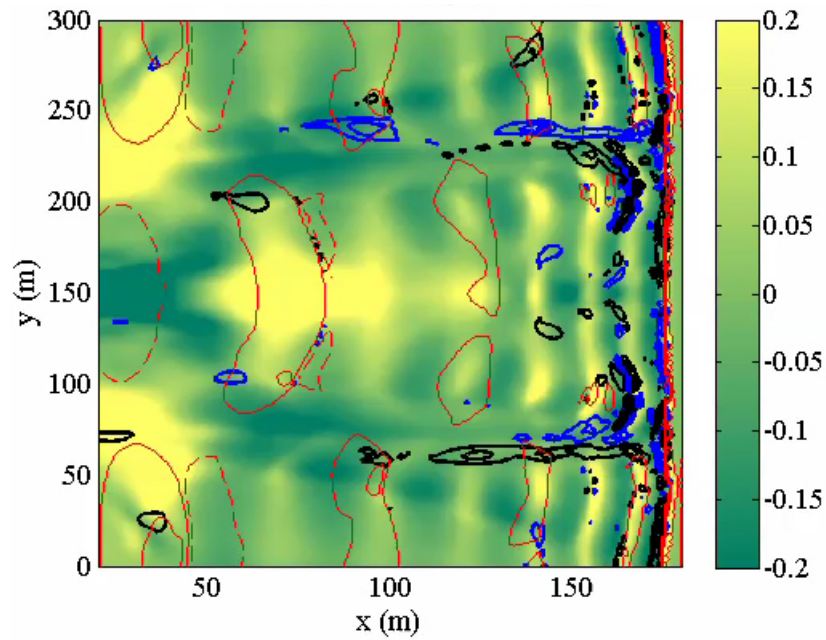
2555



2556

2557 Figure 11. Modeling (red line) of the morphological evolution of the Duck94 onshore
 2558 migration event for (A) the SkV model, (B) the SkA model, and (C) the MiX model.
 2559 The initial and final measured profiles are shown by dashed and solid lines,
 2560 respectively. Adapted from Fernandez-Mora et al. (2015).

2561



2562

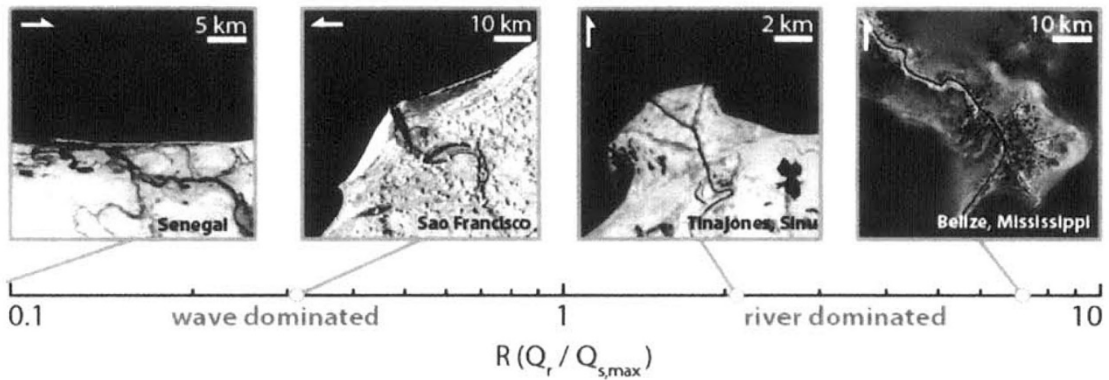
2563 Figure 12. Seabed evolution forced by sea waves and macrovortices over a planar beach.

2564 Red thin lines give contours of the free surface elevation, while black and blue

2565 solid lines are contours of clockwise and anticlockwise vorticity, respectively.

2566 Yellow gives sand deposition and green seabed erosion.

2567

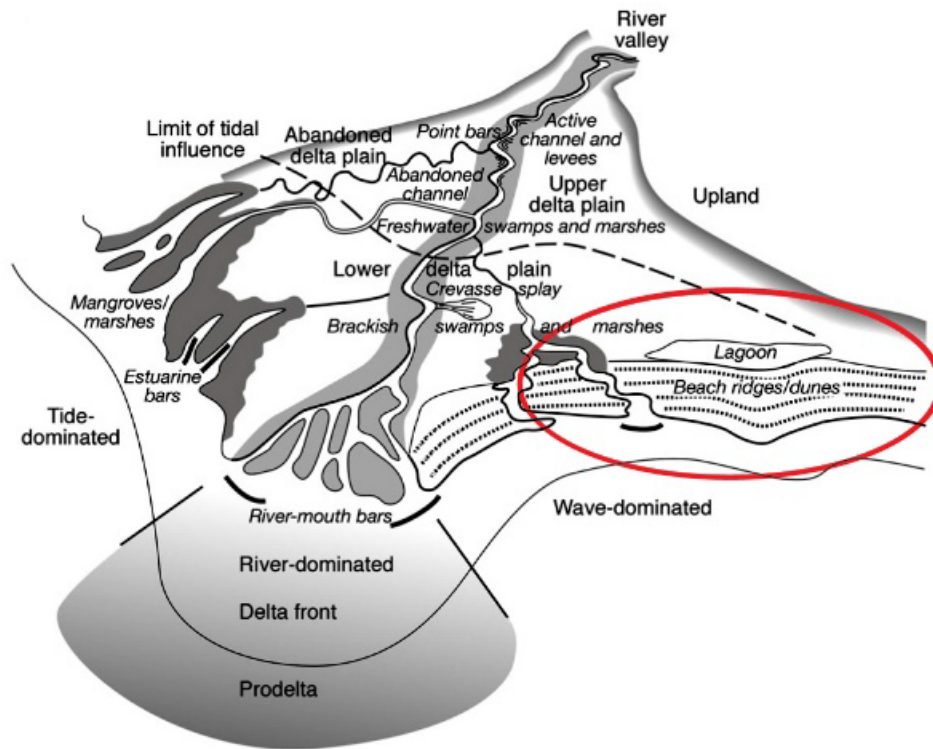


2568

2569 Figure 13. Classification of river delta morphologies by means of the fluvial dominance ratio.

2570 Adapted from Nienhuis (2016).

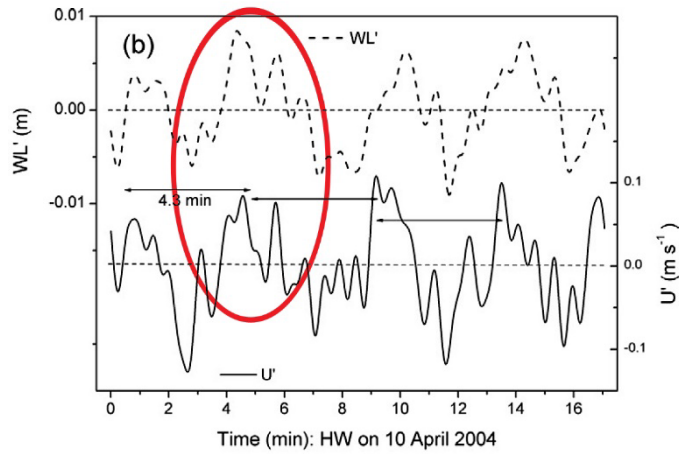
2571



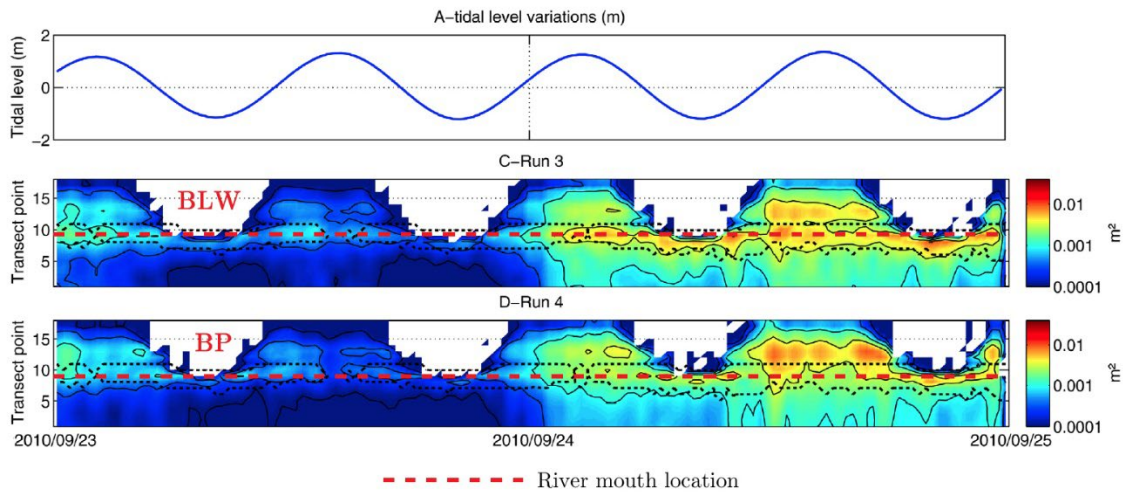
2572

2573 Figure 14. Left: river delta geomorphic units and schematic shoreline morphology (wave-

2574 dominated morphology circled in red). Adapted from Anthony (2015).



2575



2576

2577 Figure 15. IG wave propagation at river mouths. Top: positive water level-flow velocity
 2578 correlation at the Ría de Santiuste (Spain). Adapted from Uncles et al. (2014).
 2579 Bottom: tide-modulated IG wave propagation in the Albufeira Inlet (Portugal). No
 2580 IG waves propagate landward of the mouth at ebbs. Adapted from Bertin and
 2581 Olabarrieta (2016).

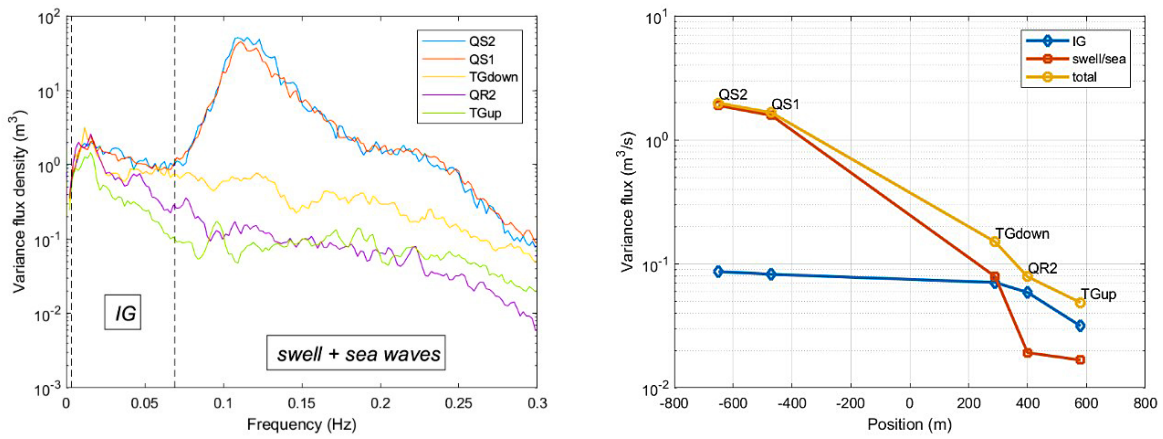
2582



2583

2584 Figure 16. The Misa River estuary in Senigallia (Italy) depicting locations of measuring
 2585 stations within the final river reach (TGdown, QR1, QR2, QR3, TGup) and in the
 2586 sea (QS1, QS2, QS3). Adapted from Melito et al. (2018a).

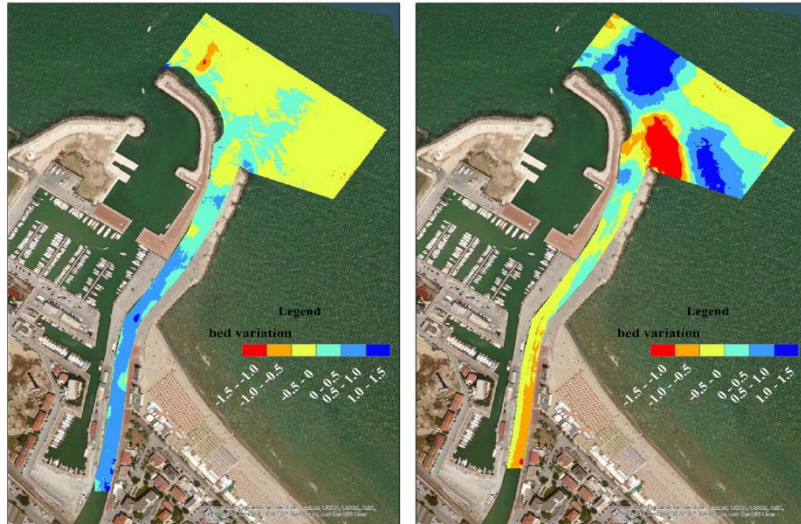
2587



2588

2589 Figure 17. Energy density flux (left) and energy flux (right) at the Misa River during storm
 2590 conditions. Adapted from Melito et al. (2018a).

2591

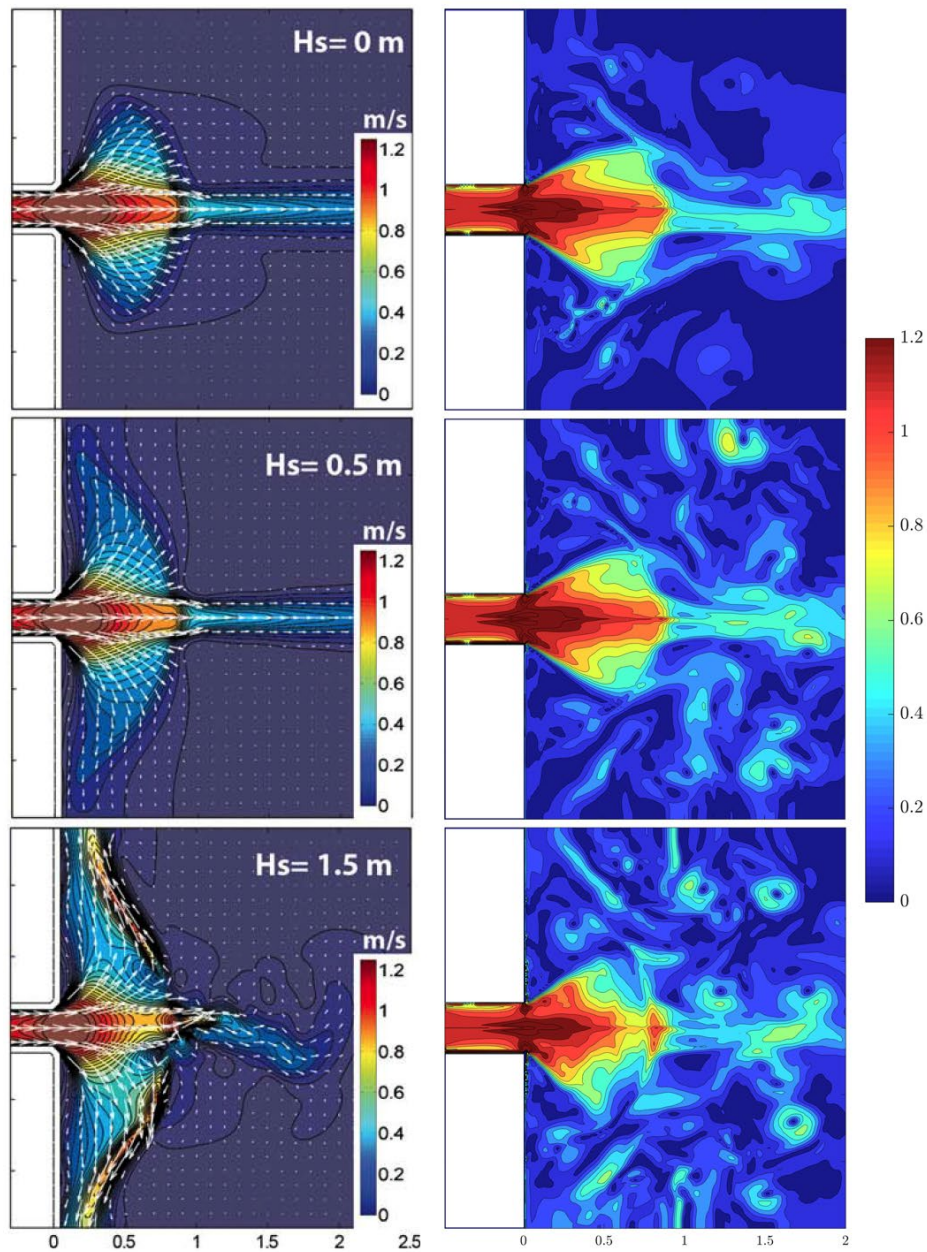


2592

2593 Figure 18. Seabed variation between May-September 2013 (left) and between September

2594 2013-February 2014 (right) at the Misa River. Adapted from Brocchini et al. (2017).

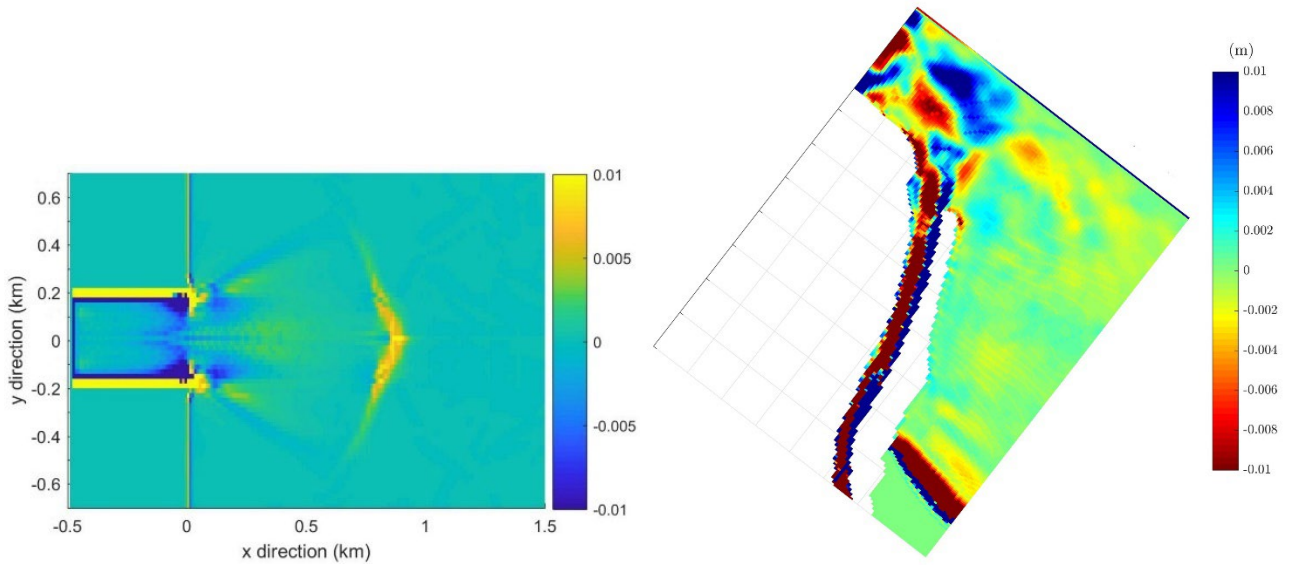
2595



2596

2597 Figure 19. Velocity maps for a strong river outflow opposed to three different wave regimes
 2598 (from top to bottom: $H_s = 0.0\text{m}$, 0.5 m , 1.5 m). Left column: results from Olabarrieta
 2599 et al. (2014). Right column: results from the wave-resolving solver of Brocchini et
 2600 al. (2001), averaged over 10 wave periods. Adapted from Melito et al. (2018b).

2601



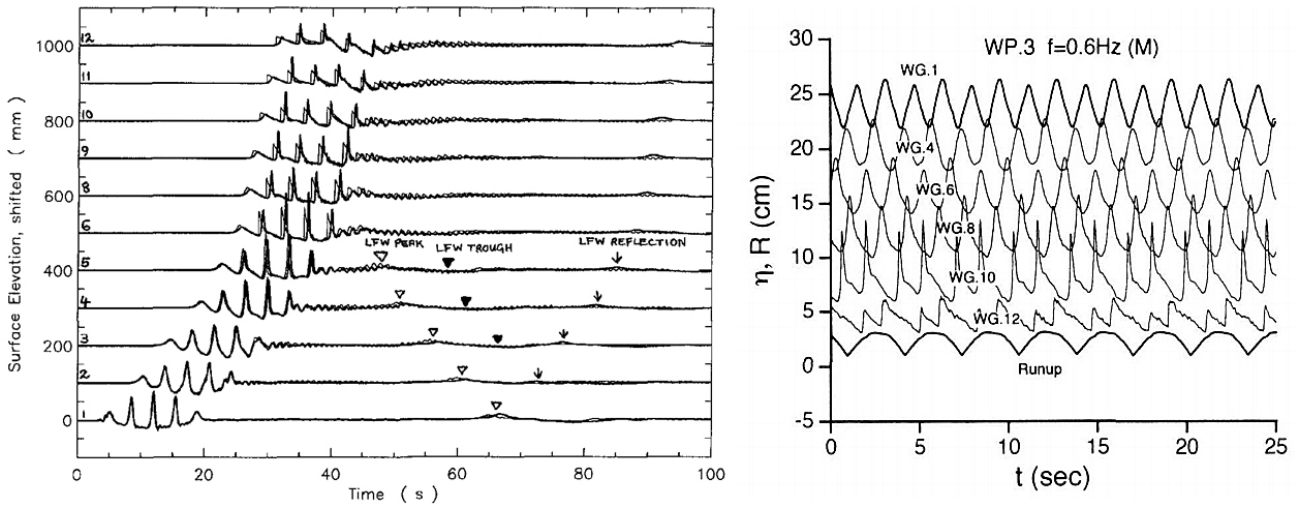
2602

2603 Figure 20. Estimated morphological evolution at the simplified river mouth bathymetry of

2604 Olabarieta et al. (2014), left panel and at the Misa River, right panel.

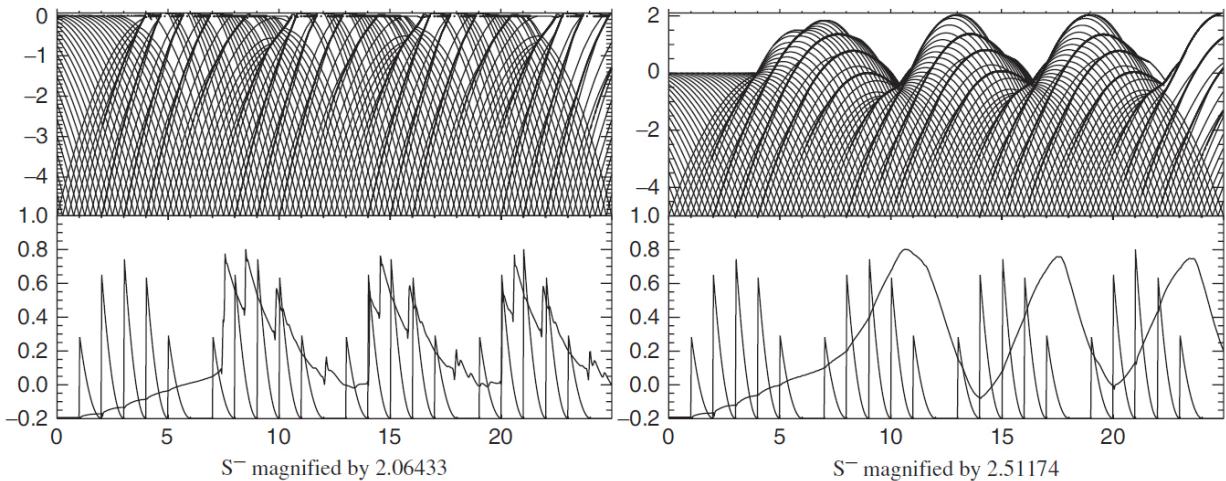
2605

2606



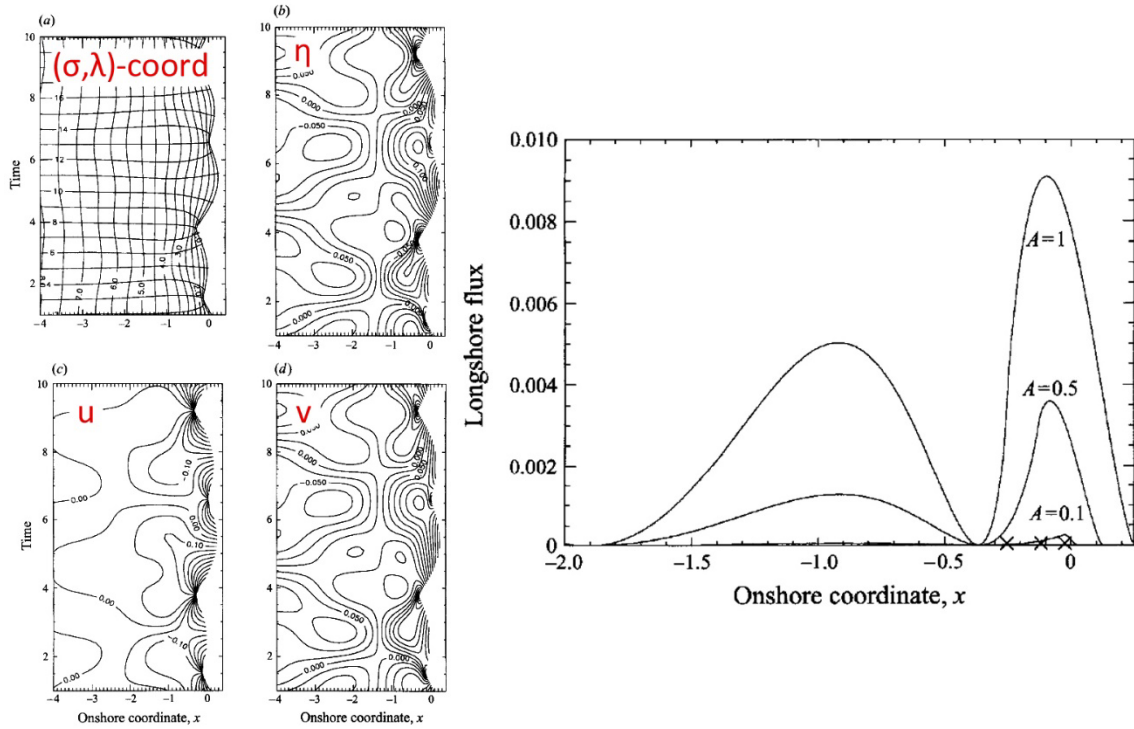
2607

2608 Figure 21. Left: example of BLW release on a beach because of sea wave breaking. The
 2609 shoreward-propagating (from bottom to top) wave group releases its BLW that is
 2610 reflected at the shoreline and propagated (from top to bottom) out to sea as a FLW
 2611 (adapted from Watson, Barnes & Peregrine 1995). Right: example of frequency
 2612 downshifting within the swash zone. Bichromatic waves propagate from seaward
 2613 to the shoreline (from top to bottom) over a steep beach and interact in the swash
 2614 zone giving wave of doubled period (adapted from Mase, 1995).



2615

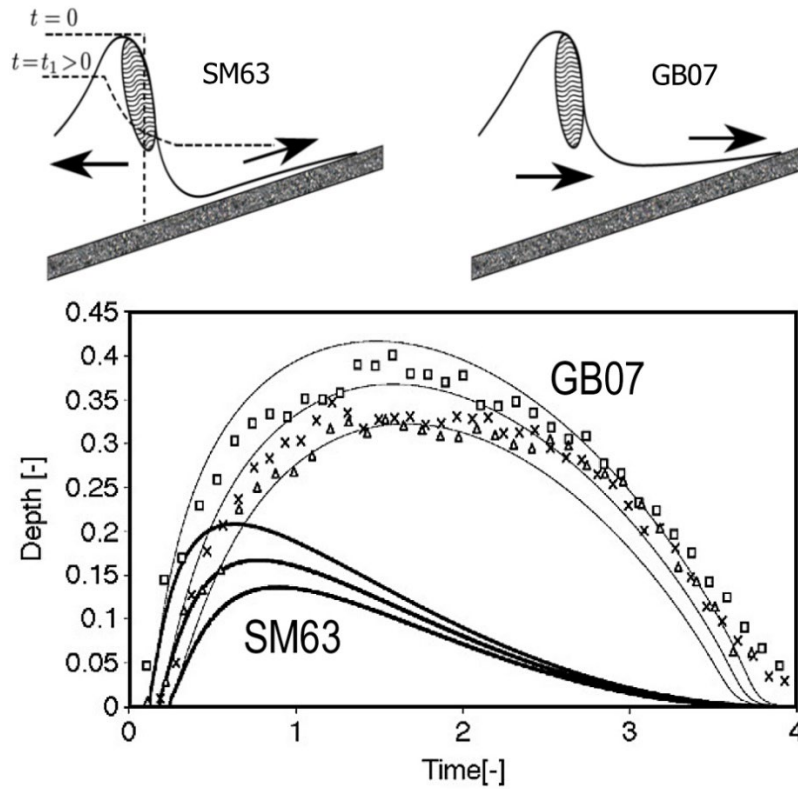
2616 Figure 22. Waves interacting at a wall (left) and at a swash zone (right). Top: incoming and
 2617 outgoing characteristic curves at the shoreline. Bottom: a group of 5 sea waves
 2618 (thin line) incoming to shore and outgoing IG waves (S^- , thick line) resulting from
 2619 the interaction of sea waves (adapted from Brocchini, 2006).



2620

2621 Figure 23. Left: maps, in the (x, t) plane of hodograph coordinates, free surface elevation η
 2622 crossshore (u) and longshore (v) velocities for two sea waves interacting onto a
 2623 beach. Right: mean longshore mass flux near and inside the swash zone for waves
 2624 of different amplitude (adapted from Brocchini & Peregrine, 1996).

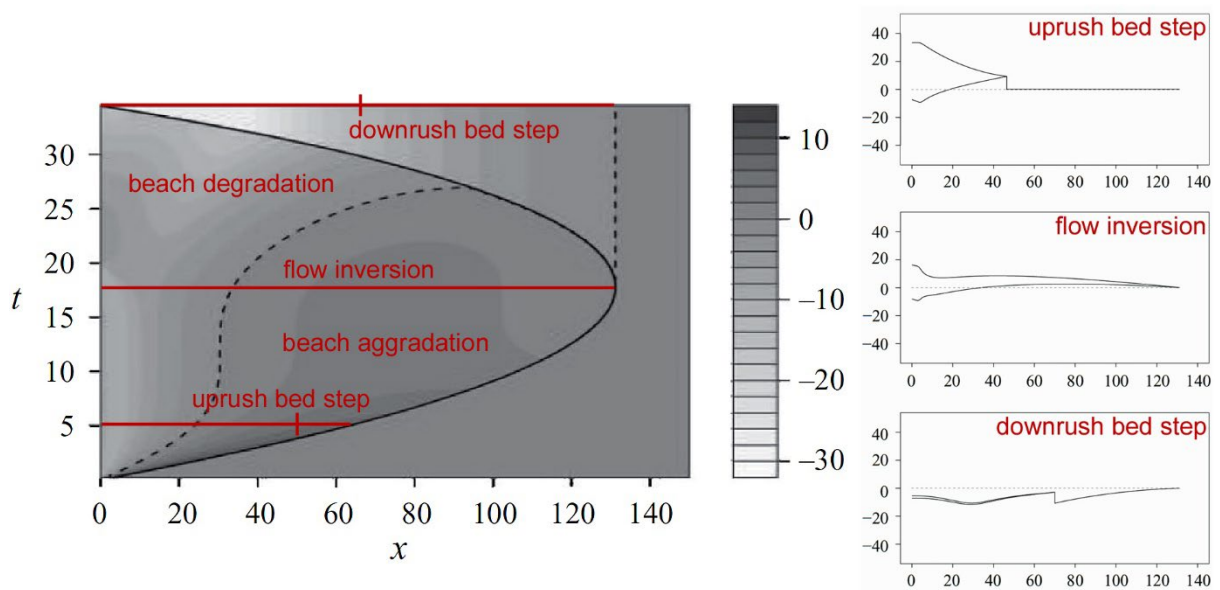
2625



2626

2627 Figure 24. Top: sketches of the flow by bore collapse. The left sketch shows the evolution
 2628 of the SM63 solution in comparison to a dam-break flow (dashed lines), the right
 2629 sketch suggests that only shoreward-accelerating fluid motion is predicted by the
 2630 GB07 solution. Bottom: swash water thickness predicted by the SM63 solution
 2631 (thick lines) and by the GB07 solution (thin lines) also compared with experimental
 2632 data (symbols). Bottom panel adapted from GB07.

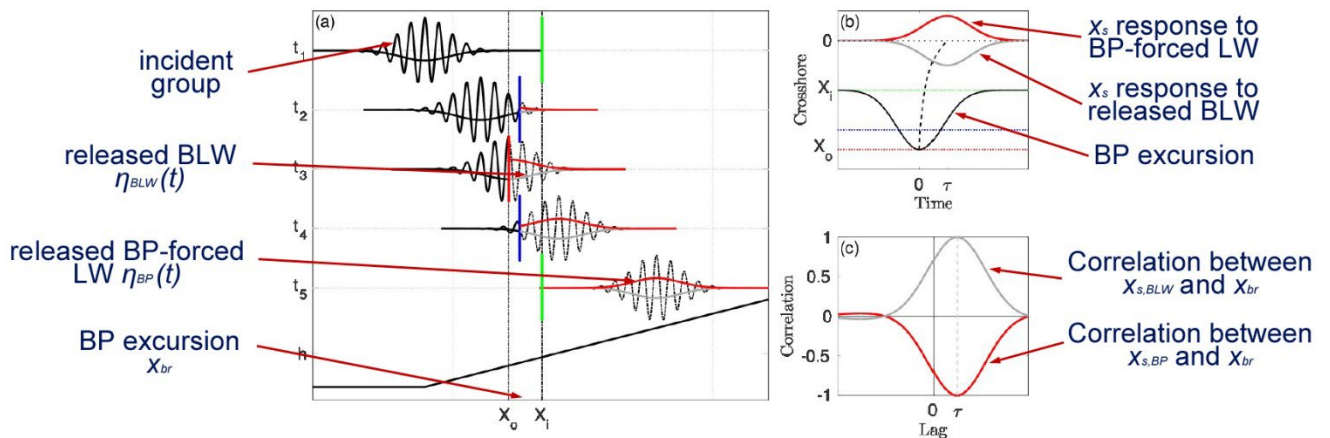
2633



2634

2635 Figure 25. Left: space-time map of the changes in beach profile due a swash event. The
 2636 thick dashed lines give the zero contour. Right: snapshots of the swash lens cross-
 2637 shore sections (adapted from Kelly & Dodd, 2010).

2638



2639

2640 Figure 26. Shoreline relation with BLW and BP forcing. (a) Wave group. The vertical blue

2641 line gives the BP between its outer (red), x_o , and inner (green), x_i location. Grey

2642 and red lines give the released BLW and incident BP-forced IG wave, respectively.

2643 (b) BP excursion (black) and shoreline response to BP-forced (red) and BLW

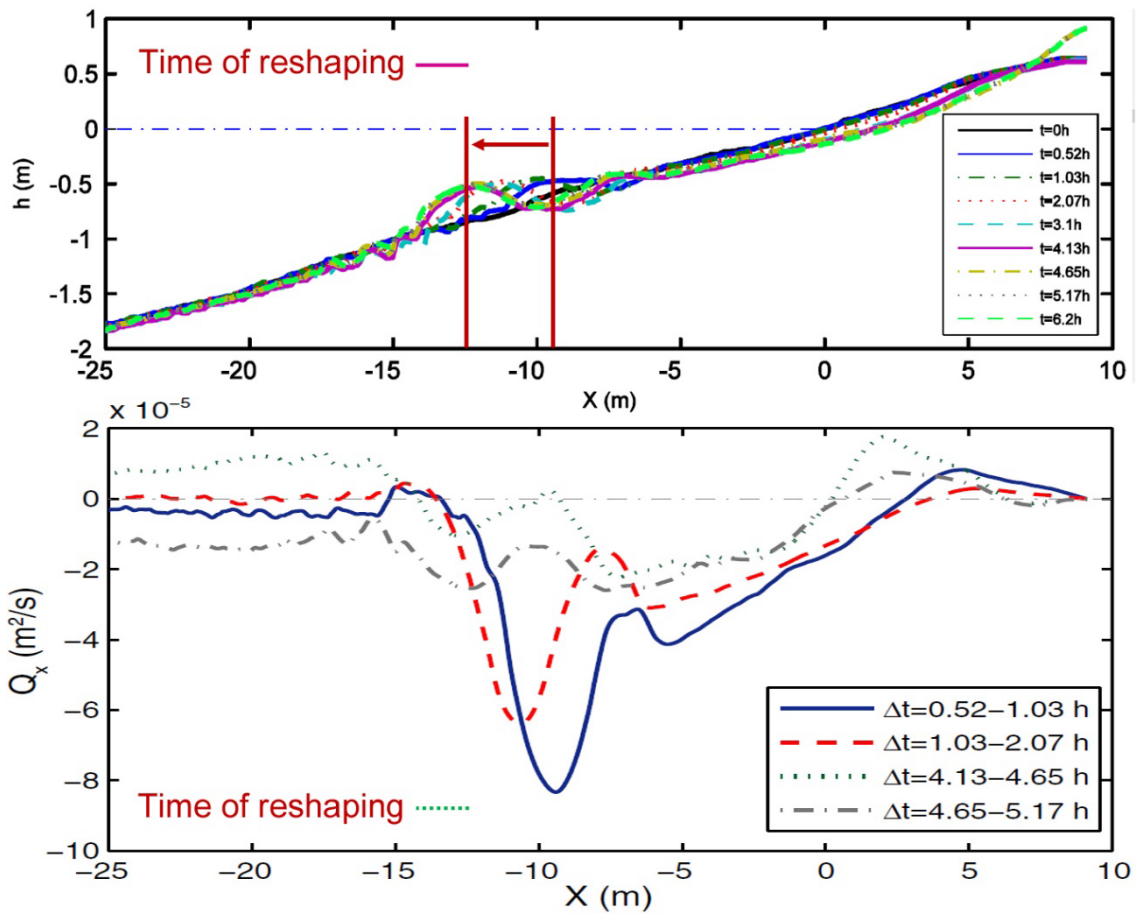
2644 (grey) IG waves. Dashed line gives path for a shallow water wave from the BP to

2645 the shoreline. Horizontal colored lines are BP positions as in Figure 2a. (c) Cross-

2646 correlation between BP and shoreline excursion, BLW released (grey) and BP-

2647 forced IG wave (red). Adapted from Moura & Baldock, (2017).

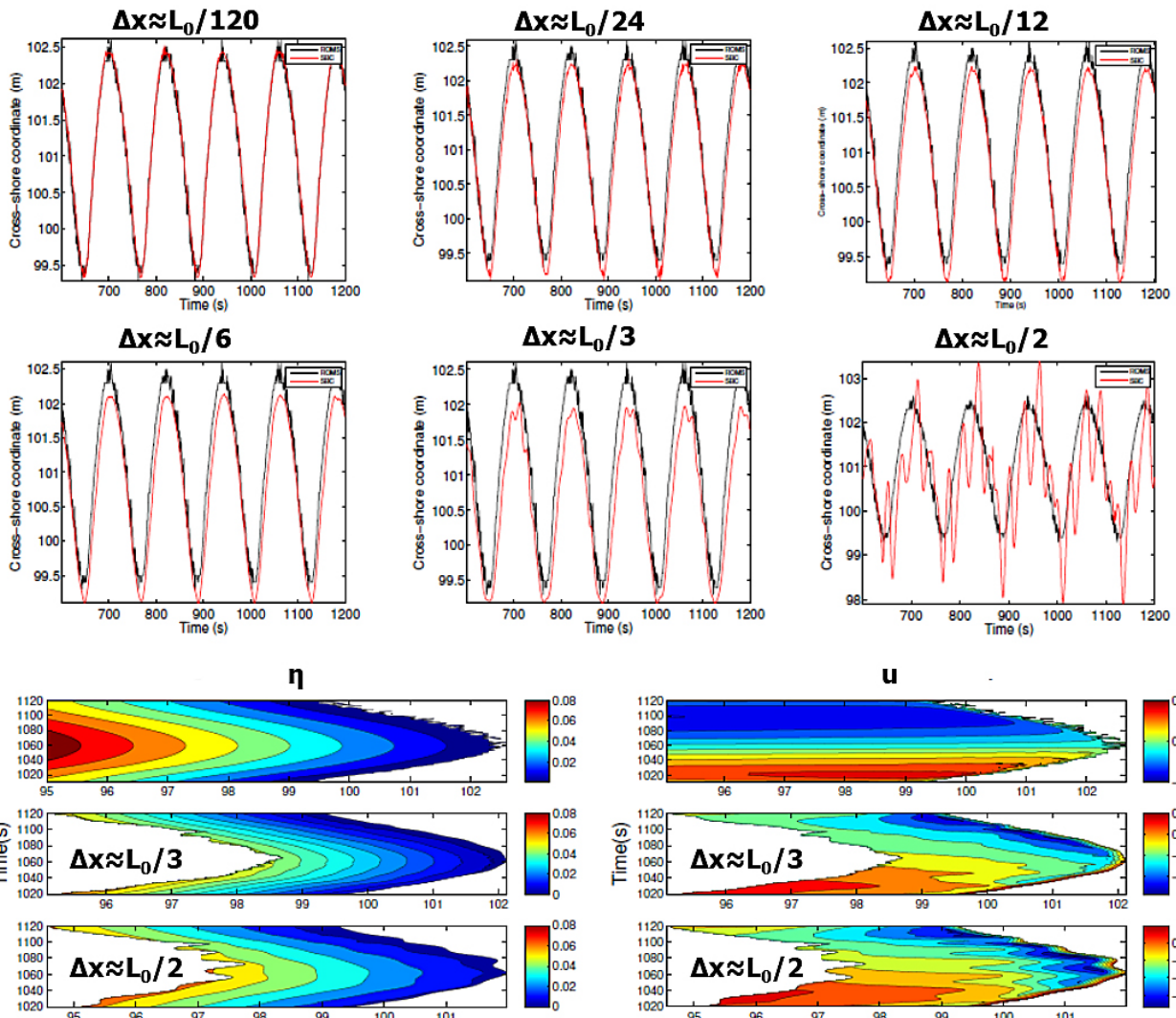
2648



2649

2650 Figure 27. Cross-shore distribution of beach profile (top panel) and of cross-shore sediment
 2651 transport rate (bottom panel) at various times before and after beach reshaping
 2652 (adapted from Alsina et al., 2012).

2653



2654

2655 Figure 28. Top 6 panels: comparison of the ROMS ultra-highly resolved shoreline (black line) and ROMS+SBCs shorelines for lower and lower grid resolution. Bottom 2
 2656 panels: free surface elevation (left) and onshore velocity (right) for the benchmark
 2657 ROMS solution (top panels) and ROMS+SBCs solutions for two very coarse
 2658 meshes (middle and lower panels). Adapted from Memmola (2017).
 2659

2660