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

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Abstract

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

Introduction

Focus, motivation and method

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

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

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

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

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

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.

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

Generalities on Models

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

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

Model structure

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

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

First, wave-averaged models: Nonlinear Shallow Water Equation (NSWE)-type (e.g. Delft3D, https://oss.deltares.nl/web/delft3d/home; Lesser et al., 2004; NHWAVE, Ma et al., 2012); Oceanographic (e.g. ROMS, https://www.myroms.org/index.php, Haas and Warner, 2009), etc. Second, wave-resolving models: NSWE-type (e.g. Postacchini et al., 2012); Boussinesq-type (e.g. Brocchini, 2013; Kirby, 2016); etc. Third, models that account for turbulence through proper closures: Reynolds Averaged Navier-Stokes; Large Eddy Simulations (Chang and Scotti, 2004; Torres-Freyermuth, Lara and Losada, 2010; Lubin et al. 2011) etc.

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

Model use

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

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

where z_b is the seabed position and

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

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

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

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

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).

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

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

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

<u>Sand bars</u>

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

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

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

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

The most known is the Net Offshore Migration (NOffM) model, which suggests that multiple nearshore 2D bars have multi-annual lifetimes and pass through 3 stages (figure 5). Generation occurs close to shore because of the interaction of undertow, wave orbital velocity asymmetry and group-forced IG orbital motion (Roelvink and Stive, 1989). This is followed by a net seaward migration, which is the outcome of an alternation of gradual onshore movements during calms and episodic strong offshore movements during storms (Van Enckevort and Ruessink, 2003, see also the right panel of figure 5). Finally, degeneration occurs at the outer surf zone when the bar moves offshore of the breaker line where weakly or non-breaking waves induce a net onshore sediment transport. This step governs the entire cycle and only when the offshore bar degenerates at the breaker line the nearshore bar can start its offshore migration (Wijnberg and Terwindt, 1995).

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

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

Alternative to both NOffM and NOnM, the Oscillation around a Position of Equilibrium (OPE, Certain and Barusseau, 2005) proposes a cyclic oscillation of the bar position due to seaward (storms) and shoreward (calms) migration. The mechanisms of this model are essentially similar to those of the NOffM, i.e. seaward migration during storms because of the dominance of the undertow and recovery during calms due to the wave-induced onshore sediment transport, but equilibrium is achieved when the local climate cannot not force storm strong or long enough (e.g. Mediterranean vs. North Sea, see also Armaroli and Ciavola, 2011). Very recent observations (Parlagreco et al., 2019) show that Mediterranean storms can indeed induce a NOffM rather than an OPE behavior and that the longshore flow can sensibly contribute to increase the potential of sediment suspension.

Both NOffM and NOnM are preferential for longshore-uniform 2D bars, which evolve for $\Omega\approx 5$ (see in the following), and seem to require large waves (Ω is directly proportional to H_b) and longshore flows (likely contributing to longshore uniformity, Shand et al., 1999). Moreover, since both NOffM and NOnM are observed to evolve over beaches with similar slopes (1:200-1:150) and sediment sizes ($d_{50}\approx150\mu m$), the different behaviours are probably to be searched into the different incident wave field (steepness and storm sequencing) and possible physiographic constraints.

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

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

The "empirical models" are based on use of the Dean parameter Ω (Wright and Short, 1984) and of the the $B^*=xs/g\ tan\beta\ T_i^2$ parameter (Short and Aagard, 1993), where x_s is the nearshore width (offhore distance to the point where $\beta\approx0$). Single bar formation is possible for $20<B^*<50$ and $2< \Omega <6$, while $\Omega<1$ characterizes reflective beaches and $\Omega >6$ dissipative beachs. For Ω increasing from 2 to 6 the following features emerge in sequence (Wijnberg and Kroon, 2002): shore-attached bars, 3D bars and 2D bars (the latter ones for $\Omega\approx 5$). Formation of multiple bars requires larger values of B* (e.g. 2 bars for $50 < B^* <100$, 3 bars for $100< B^* < 300$, etc.);

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

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

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

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

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

Finally, “stability models” have been succesfully used to describe the generation of crescentic or lunate 3D bars (Vittori et al., 1999) and the evolution of a double system of crescentic 3D bars (Klein and Schuttelaars, 2006). Crescentic bars emerge because of steady currents caused by the interaction of the incoming sea/short waves with synchronous edge waves excited by the incoming waves moving over the wavy seabed. Klein and Schuttelaars (2006) studied a similar problem using a numerical, wave-averaged model (NSWE+Exner). Starting from an equilibrium state and numerically added perturbations the fastest-growing perturbation was computed (see also the left panel of figure 7).

Figure 7. Left panel: nonlinear 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).

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

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

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

Simulations were run by providing a template in the morphology (inner bar longshore uniform and outer-bar with crescentic patterns). The computed feedback between flow, sediment transport and the evolving morphology allowed to reproduce morphologies similar to those of observed double crescentic sandbar systems (see figure 8). The Authors advanced that a novel mechanism ("morphological coupling") was introduced that "blurs the distinction between self-organization and template mechanisms". However, inspection of equation (1) suggests that such a novel mechanism is a complete model that retains all terms on the right hand side. In other words, a template of crescentic outer bar is forced (term I of equation 1) in a self-interating model (term III of equation 1).

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

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

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

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

De Bakker et al. (2016) suggest that $R_{Uu_{IG}}$ is a good proxy for the nearshore sediment transport only on moderately-sloping beaches while H_{IG}/H_{SW} well represents the sediment transport on all beaches. In fact, while over moderate slopes $R_{Uu_{IG}} \neq 0$, it is $R_{Uu_{IG}} = 0$ on mild slopes. On the other hand, H_{IG}/H_{SW} is always positive and increasing for mild slopes, because IG waves receive more energy from sea waves breaking on wide surf zones.

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

The latter condition is seen to be a potentially important cause of seabed morphology alteration with the generation of complex 3D sandbars. The example of figure 12 has been obtained by computing, using the NSWE morphodynamic model of Postacchini et al. (2012), the morphological evolution of a planar beach under the combined action of waves and macrovortices generated by finite longshore breakers due to regular waves (Hi=1m, Ti=10s, β=1/30) crossing at an angle of 15º to the coastline.

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

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

<u>River mouths</u>

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

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

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

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

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

$$R = fluvial\ dominance\ ratio = \frac{Q_r}{Q_{s,max}} = \frac{river\ sand\ flux}{max\ longshore\ sediment\ flux\ at\ river\ mouth} \quad (5b)$$

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

Figure 13. Classification of river delta morphologies by means of the fluvial dominance ratio. Adapted from Nienhuis (2016).

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

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

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

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

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

Seaward flowing currents interact with incoming waves (Chawla and Kirby, 2002) to: slow them down and even block them if the depth-uniform current velocity is equal and opposite to the wave group velocity, this is known as “wave blocking”; steepen and even lead to breaking; decrease the high-frequency content by wave breaking and transfer part of the current energy to lower wave frequencies, this leading to frequency downshift of the incoming waves;

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

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

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

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

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

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

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

$$dE_f = S(f)C_g \qquad\qquad dE_f = \int S(f)C_g df \tag{6}$$

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

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

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

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).

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

Figure 19. Velocity maps for a strong river outflow opposed to three different wave regimes (from top to bottom: H_s = 0.0m, 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).

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

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

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

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

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

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

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

<u>Swash zone</u>

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

$$-40\,Kg/m < \hat{q}_x < 40\,Kg/m, \quad \hat{q}_x \equiv \int_{Tswash} q\,dt \qquad (7)$$

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

$$Q_y \sim (1-10)\,Kg/s \qquad (8)$$

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

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

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

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

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

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

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

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

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

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

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

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

$$x_s = u_0 t - \tfrac{1}{2} g \, tan\beta \, t^2 \qquad (10)$$

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

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

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

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-moprhodynamics. Some of their results reveal the complexities in beach profile changes due to a swash event (see figure 25).

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).

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

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

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

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

Such relation can be illustrated with reference to the schematic of figure 26. When a group arrives at breaking (panel a) the first, smaller waves break at x_i (green line) and the subsequent, larger waves, more and more seaward, moving the BP to x_o (red line), the subsequent smaller waves force the BP to move back to x_i. Hence, the BP position $x_{br}(t)$ is – large (x_i), small(x_o), large (x_i) – i.e. in phase with the BLW vertical oscillation $\eta_{BLW}(t)$ (which is large, small, large, i.e. concave upwards). The amplitude of the BP-forced IG wave $\eta_{BP}(t)$ is proportional to the width of the breaking region. Hence, it increases during the passage from small to large waves ($x_{br}(t)$ from zero to max) and decreases from large to small waves ($x_{br}(t)$ from max to zero). Hence, $\eta_{BP}(t)$ is concave downward or in phase opposition to $x_{br}(t)$ and $\eta_{BLW}(t)$. The released BLW and the incident BP-forced wave propagate with the same speed ($\sqrt{gh}$) and reach the shoreline at the same time (after traveling through the surf zone), but with opposite phase (panel b). This leads to a positive correlation between $x_{s,BLW}$ *and* x_{br} and a negative correlation between $x_{s,BP}$ *and* x_{br} (panel c).

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

From the morphological point of view, one fundamental link between the surf zone and swash zone dynamics is represented by the feeding of the most inshore sand bar (see also figures 4, 5, 11) by swash zone sediments. This has been recently investigated by Alsina et al. (2012), on the basis of dedicated laboratory experiments. Such experiments have been carried out by using flow conditions (surf zone characterized by plunging breakers, $\varepsilon \approx 12$-13) that, during an initial stage of run of about 3500 waves, led to the generation, from an almost-planar beach, of a nearshore bar that slowing migrated off to sea (see blue-solid to green-dashed lines, and vertical red lines in top panel of figure 27). The run was then stopped and the beach profile manually reshaped to a milder slope only inside the swash zone (pink-solid line). The same flow conditions were subsequently run, but bar migration stopped.

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

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

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

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

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

This section is closed with some promising ongoing research aimed at including the swash zone dynamics into wave-averaged solvers for the nearshore circulation (see also "Generalities on Models" in the Introduction). By definition, wave-averaged models cannot describe the swash zone dynamics associated to the sea waves, this leading to problems in correctly predicting IG waves radiating out to sea. This is well illustrated by figure 22, whose left panels describe the interaction of groups of 5 sea waves at a rigid wall used in wave-averaged solvers to represent the shoreline, while the right panels describe the same interaction at a swash zone reproduced by means of a wave-resolving model (in the specific that of Brocchini et al., 2001). Two main features are very clear: the interaction at a wall, being a mere reflection of the incoming wave groups, leads to outgoing groups with the same structure of the incoming ones; the interaction at a swash zone allows for a significant swash-swash interaction, for example because of large waves catching up and engulfing small waves (see Mase, 1995 and figure 21), which contributes to increasing the intensity (S^-) of the IG waves radiated out to sea (in the specific 20% larger for the interaction within the swash zone). In view of the fundamental importance of wave-wave interaction within the swash zone Brocchini and co-workers have designed some Shoreline Boundary Conditions (SBCs) for the motion of the mean shoreline x_l and of the flow at at such boundary ($d(x_l)$, $u(x_l)$) on the basis of an integral swash zone model (Brocchini and Peregrine, 1996; Brocchini and Bellotti, 2002). Such SBCs have been successfully validated by means of large-scale laboratory experiments (Bellotti et al., 2003) and have recently been implemented in the ROMS wave-averaged solver (Haas and Warner, 2009). The main aim of such preliminary implementation was to gauge the capabilities of reproducing swash zone dynamics at a reduced computational cost, the final goal being that of predicting such dynamics by simply running a wave-averaged solution. Figure 28 summarizes the main findings of the mentioned implementation (Memmola, 2017; Memmola et al., 2019). The top six panels give the shoreline motion computed by running ROMS alone at the unreasonably-high resolution of 120 grid nodes (Δx) per offshore wavelength L_0 (black line), i.e. using it as a wave-resolving model (benchmark solution), and ROMS coupled with the proposed SBCs at lower and lower resolutions (red line), down to 2 grid nodes per wavelength. It is clear that coupling ROMS with the SBCs allows one to run very cheap calculations that predict reasonably well the benchmark solution by only using 3 grid nodes per wavelength (with $\Delta x/L_0 \approx 2$ the solution is too deteriorated). The lower two panels of the same figure show that not only the shoreline motion, but the entire flow field is well predicted by using the SBCs. The left and right panels give, respectively, the swash zone free surface elevation and onshore velocity for the benchmark solution (top panels) and ROMS+SBCs solutions for two very coarse meshes (middle and lower panels).

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 surface elevation (left) and onshore velocity (right) for the benchmark ROMS solution (top panels) and ROMS+SBCs solutions for two very coarse meshes (middle and lower panels). Adapted from Memmola (2017).

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

Discussion

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

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

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

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

Macrovortices are a nonlocal forcing of the nearshore dynamics because they can maintain their coherent over large distances after their generation. Macrovortices contribute to sediment transport both by mobilizing the nearbed sands and, like IG waves, by carrying the sediments put into suspension by sea waves. The differential breaking that generates macrovortices is particularly intense at river mouths characterized by seabed shoals. Hence, macrovortices, beyond the river jet expansion, provide a significant contribution to both flow mixing and sediment transport at river mouths (Melito et al., 2018b).

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

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

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

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

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

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

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

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

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

Conclusions

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

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

Local dynamics specific to river mouths are: the wave-current interaction by which the river current steepens the incoming waves, if intense enough to the point of breaking and/or blocking them, and forces a frequency downshifting (Chawla and Kirby, 2002); the related formation of a mouth bar (Fagherazzi et al., 2015) or the formation of a shoal that largely influences the local dynamics by forcing stronger wave-breaking-induced return currents (Olabarrieta et al., 2014) and wave breaking/sediment deposit farther from the mouth, i.e. at the seaward edge of the shoal.

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

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

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

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Abstract

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

Introduction

Focus, motivation and method

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

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

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

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

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

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.

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

Generalities on Models

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

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

Model structure

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

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

First, wave-averaged models: Nonlinear Shallow Water Equation (NSWE)-type (e.g. Delft3D, https://oss.deltares.nl/web/delft3d/home; Lesser et al., 2004; NHWAVE, Ma et al., 2012); Oceanographic (e.g. ROMS, https://www.myroms.org/index.php, Haas and Warner, 2009), etc. Second, wave-resolving models: NSWE-type (e.g. Postacchini et al., 2012); Boussinesq-type (e.g. Brocchini, 2013; Kirby, 2016); etc. Third, models that account for turbulence through proper closures: Reynolds Averaged Navier-Stokes; Large Eddy Simulations (Chang and Scotti, 2004; Torres-Freyermuth, Lara and Losada, 2010; Lubin et al. 2011) etc.

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

Model use

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

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

where z_b is the seabed position and

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

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

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

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

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).

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

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

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

<u>Sand bars</u>

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

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

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

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

The most known is the Net Offshore Migration (NOffM) model, which suggests that multiple nearshore 2D bars have multi-annual lifetimes and pass through 3 stages (figure 5). Generation occurs close to shore because of the interaction of undertow, wave orbital velocity asymmetry and group-forced IG orbital motion (Roelvink and Stive, 1989). This is followed by a net seaward migration, which is the outcome of an alternation of gradual onshore movements during calms and episodic strong offshore movements during storms (Van Enckevort and Ruessink, 2003, see also the right panel of figure 5). Finally, degeneration occurs at the outer surf zone when the bar moves offshore of the breaker line where weakly or non-breaking waves induce a net onshore sediment transport. This step governs the entire cycle and only when the offshore bar degenerates at the breaker line the nearshore bar can start its offshore migration (Wijnberg and Terwindt, 1995).

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

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

Alternative to both NOffM and NOnM, the Oscillation around a Position of Equilibrium (OPE, Certain and Barusseau, 2005) proposes a cyclic oscillation of the bar position due to seaward (storms) and shoreward (calms) migration. The mechanisms of this model are essentially similar to those of the NOffM, i.e. seaward migration during storms because of the dominance of the undertow and recovery during calms due to the wave-induced onshore sediment transport, but equilibrium is achieved when the local climate cannot not force storm strong or long enough (e.g. Mediterranean vs. North Sea, see also Armaroli and Ciavola, 2011). Very recent observations (Parlagreco et al., 2019) show that Mediterranean storms can indeed induce a NOffM rather than an OPE behavior and that the longshore flow can sensibly contribute to increase the potential of sediment suspension.

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

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

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

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

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

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

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

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

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

Finally, “stability models” have been succesfully used to describe the generation of crescentic or lunate 3D bars (Vittori et al., 1999) and the evolution of a double system of crescentic 3D bars (Klein and Schuttelaars, 2006). Crescentic bars emerge because of steady currents caused by the interaction of the incoming sea/short waves with synchronous edge waves excited by the incoming waves moving over the wavy seabed. Klein and Schuttelaars (2006) studied a similar problem using a numerical, wave-averaged model (NSWE+Exner). Starting from an equilibrium state and numerically added perturbations the fastest-growing perturbation was computed (see also the left panel of figure 7).

Figure 7. Left panel: nonlinear 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).

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

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

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

Simulations were run by providing a template in the morphology (inner bar longshore uniform and outer-bar with crescentic patterns). The computed feedback between flow, sediment transport and the evolving morphology allowed to reproduce morphologies similar to those of observed double crescentic sandbar systems (see figure 8). The Authors advanced that a novel mechanism (“morphological coupling”) was introduced that “blurs the distinction between self-organization and template mechanisms”. However, inspection of equation (1) suggests that such a novel mechanism is a complete model that retains all terms on the right hand side. In other words, a template of crescentic outer bar is forced (term I of equation 1) in a self-interating model (term III of equation 1).

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

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

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

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

De Bakker et al. (2016) suggest that $R_{Uu_{IG}}$ is a good proxy for the nearshore sediment transport only on moderately-sloping beaches while H_{IG}/H_{SW} well represents the sediment transport on all beaches. In fact while over moderate slopes $R_{Uu_{IG}} \neq 0$, it is $R_{Uu_{IG}} = 0$ on mild slopes. On the other hand, H_{IG}/H_{SW} is always positive and increasing for mild slopes, because IG waves receive more energy from sea waves breaking on wide surf zones.

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

The latter condition is seen to be a potentially important cause of seabed morphology alteration with the generation of complex 3D sandbars. The example of figure 12 has been obtained by computing, using the NSWE morphodynamic model of Postacchini et al. (2012), the morphological evolution of a planar beach under the combined action of waves and macrovortices generated by finite longshore breakers due to regular waves (Hi=1m, Ti=10s, β=1/30) crossing at an angle of 15º to the coastline.

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

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

River mouths

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

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

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

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

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

$$R = fluvial\ dominance\ ratio = \frac{Q_r}{Q_{s,max}} = \frac{river\ sand\ flux}{max\ longshore\ sediment\ flux\ at\ river\ mouth} \qquad (5b)$$

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

Figure 13. Classification of river delta morphologies by means of the fluvial dominance ratio. Adapted from Nienhuis (2016).

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

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

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

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

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

Seaward flowing currents interact with incoming waves (Chawla and Kirby, 2002) to: slow them down and even block them if the depth-uniform current velocity is equal and opposite to the wave group velocity, this is known as “wave blocking”; steepen and even lead to breaking; decrease the high-frequency content by wave breaking and transfer part of the current energy to lower wave frequencies, this leading to frequency downshift of the incoming waves;

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

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

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

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

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

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

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

$$dE_f = S(f)C_g \qquad\qquad dE_f = \int S(f)C_g df \tag{6}$$

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

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

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

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).

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

Figure 19. Velocity maps for a strong river outflow opposed to three different wave regimes (from top to bottom: H_s = 0.0m, 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).

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

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

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

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

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

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

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

Swash zone

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

$$-40\,Kg/m < \hat{q}_x < 40\,Kg/m, \quad \hat{q}_x \equiv \int_{Tswash} q dt \qquad (7)$$

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

$$Q_y \sim (1-10)\,Kg/s \qquad (8)$$

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

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

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

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

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

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

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

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

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

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

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

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

$$x_s = u_0 t - \tfrac{1}{2} g \tan\beta \, t^2 \tag{10}$$

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

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

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

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-moprhodynamics. Some of their results reveal the complexities in beach profile changes due to a swash event (see figure 25).

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).

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

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

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

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

Such relation can be illustrated with reference to the schematic of figure 26. When a group arrives at breaking (panel a) the first, smaller waves break at x_i (green line) and the subsequent, larger waves, more and more seaward, moving the BP to x_o (red line), the subsequent smaller waves force the BP to move back to x_i. Hence, the BP position $x_{br}(t)$ is – large (x_i), small(x_o), large (x_i) – i.e. in phase with the BLW vertical oscillation $\eta_{BLW}(t)$ (which is large, small, large, i.e. concave upwards). The amplitude of the BP-forced IG wave $\eta_{BP}(t)$ is proportional to the width of the breaking region. Hence, it increases during the passage from small to large waves ($x_{br}(t)$ from zero to max) and decreases from large to small waves ($x_{br}(t)$ from max to zero). Hence, $\eta_{BP}(t)$ is concave downward or in phase opposition to $x_{br}(t)$ and $\eta_{BLW}(t)$. The released BLW and the incident BP-forced wave propagate with the same speed ($\sqrt{gh}$) and reach the shoreline at the same time (after traveling through the surf zone), but with opposite phase (panel b). This leads to a positive correlation between $x_{s,BLW}$ *and* x_{br} and a negative correlation between $x_{s,BP}$ *and* x_{br} (panel c).

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

From the morphological point of view, one fundamental link between the surf zone and swash zone dynamics is represented by the feeding of the most inshore sand bar (see also figures 4, 5, 11) by swash zone sediments. This has been recently investigated by Alsina et al. (2012), on the basis of dedicated laboratory experiments. Such experiments have been carried out by using flow conditions (surf zone characterized by plunging breakers, $\varepsilon \approx 12$-13) that, during an initial stage of run of about 3500 waves, led to the generation, from an almost-planar beach, of a nearshore bar that slowing migrated off to sea (see blue-solid to green-dashed lines, and vertical red lines in top panel of figure 27). The run was then stopped and the beach profile manually reshaped to a milder slope only inside the swash zone (pink-solid line). The same flow conditions were subsequently run, but bar migration stopped.

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

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

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

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

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

This section is closed with some promising ongoing research aimed at including the swash zone dynamics into wave-averaged solvers for the nearshore circulation (see also "Generalities on Models" in the Introduction). By definition, wave-averaged models cannot describe the swash zone dynamics associated to the sea waves, this leading to problems in correctly predicting IG waves radiating out to sea. This is well illustrated by figure 22, whose left panels describe the interaction of groups of 5 sea waves at a rigid wall used in wave-averaged solvers to represent the shoreline, while the right panels describe the same interaction at a swash zone reproduced by means of a wave-resolving model (in the specific that of Brocchini et al., 2001). Two main features are very clear: the interaction at a wall, being a mere reflection of the incoming wave groups, leads to outgoing groups with the same structure of the incoming ones; the interaction at a swash zone allows for a significant swash-swash interaction, for example because of large waves catching up and engulfing small waves (see Mase, 1995 and figure 21), which contributes to increasing the intensity (S^-) of the IG waves radiated out to sea (in the specific 20% larger for the interaction within the swash zone). In view of the fundamental importance of wave-wave interaction within the swash zone Brocchini and co-workers have designed some Shoreline Boundary Conditions (SBCs) for the motion of the mean shoreline x_l and of the flow at at such boundary ($d(x_l)$, $u(x_l)$) on the basis of an integral swash zone model (Brocchini and Peregrine, 1996; Brocchini and Bellotti, 2002). Such SBCs have been successfully validated by means of large-scale laboratory experiments (Bellotti et al., 2003) and have recently been implemented in the ROMS wave-averaged solver (Haas and Warner, 2009). The main aim of such preliminary implementation was to gauge the capabilities of reproducing swash zone dynamics at a reduced computational cost, the final goal being that of predicting such dynamics by simply running a wave-averaged solution. Figure 28 summarizes the main findings of the mentioned implementation (Memmola, 2017; Memmola et al., 2019). The top six panels give the shoreline motion computed by running ROMS alone at the unreasonably-high resolution of 120 grid nodes (Δx) per offshore wavelength L_0 (black line), i.e. using it as a wave-resolving model (benchmark solution), and ROMS coupled with the proposed SBCs at lower and lower resolutions (red line), down to 2 grid nodes per wavelength. It is clear that coupling ROMS with the SBCs allows one to run very cheap calculations that predict reasonably well the benchmark solution by only using 3 grid nodes per wavelength (with $\Delta x/L_0 \approx 2$ the solution is too deteriorated). The lower two panels of the same figure show that not only the shoreline motion, but the entire flow field is well predicted by using the SBCs. The left and right panels give, respectively, the swash zone free surface elevation and onshore velocity for the benchmark solution (top panels) and ROMS+SBCs solutions for two very coarse meshes (middle and lower panels).

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 surface elevation (left) and onshore velocity (right) for the benchmark ROMS solution (top panels) and ROMS+SBCs solutions for two very coarse meshes (middle and lower panels). Adapted from Memmola (2017).

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

Discussion

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

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

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

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

Macrovortices are a nonlocal forcing of the nearshore dynamics because they can maintain their coherent over large distances after their generation. Macrovortices contribute to sediment transport both by mobilizing the nearbed sands and, like IG waves, by carrying the sediments put into suspension by sea waves. The differential breaking that generates macrovortices is particularly intense at river mouths characterized by seabed shoals. Hence, macrovortices, beyond the river jet expansion, provide a significant contribution to both flow mixing and sediment transport at river mouths (Melito et al., 2018b).

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

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

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

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

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

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

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

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

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

Conclusions

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

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

Local dynamics specific to river mouths are: the wave-current interaction by which the river current steepens the incoming waves, if intense enough to the point of breaking and/or blocking them, and forces a frequency downshifting (Chawla and Kirby, 2002); the related formation of a mouth bar (Fagherazzi et al., 2015) or the formation of a shoal that largely influences the local dynamics by forcing stronger wave-breaking-induced return currents (Olabarrieta et al., 2014) and wave breaking/sediment deposit farther from the mouth, i.e. at the seaward edge of the shoal.

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

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

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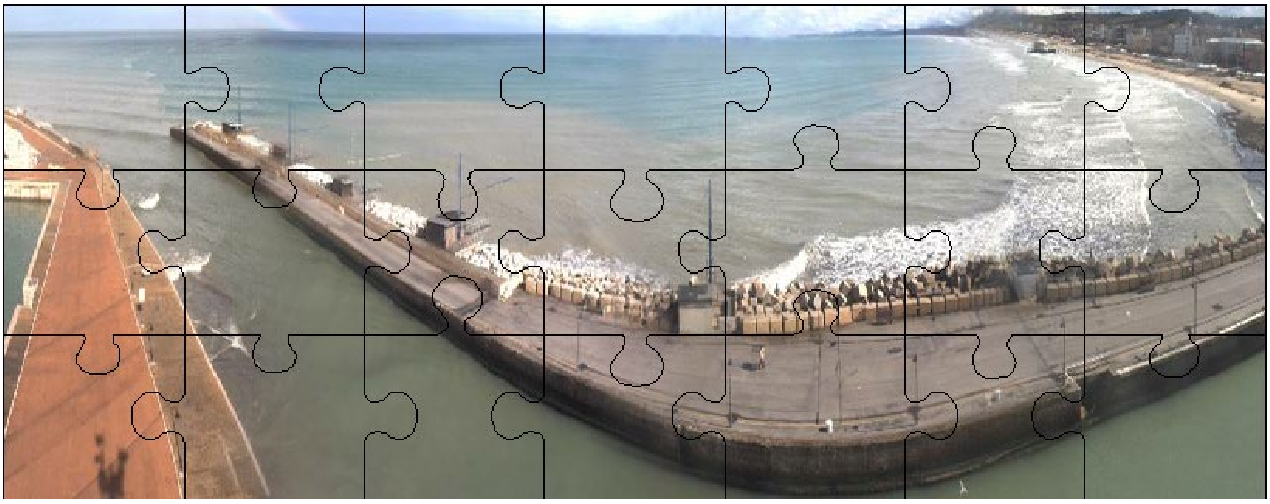

Figure 1. Wave-forced dynamics in the nearshore and estuaries: a puzzle to be solved.

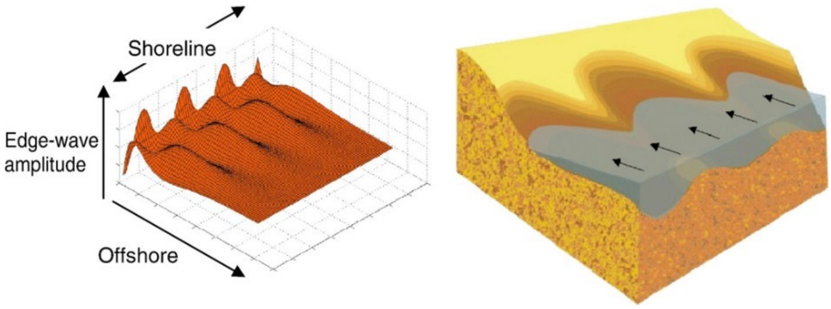

Figure 2. Forced/template mechanism: “synchronous standing edge waves” (left) and flow imprinted on sand giving beach cusps (right). Adapted from Coco and Murray, (2007).

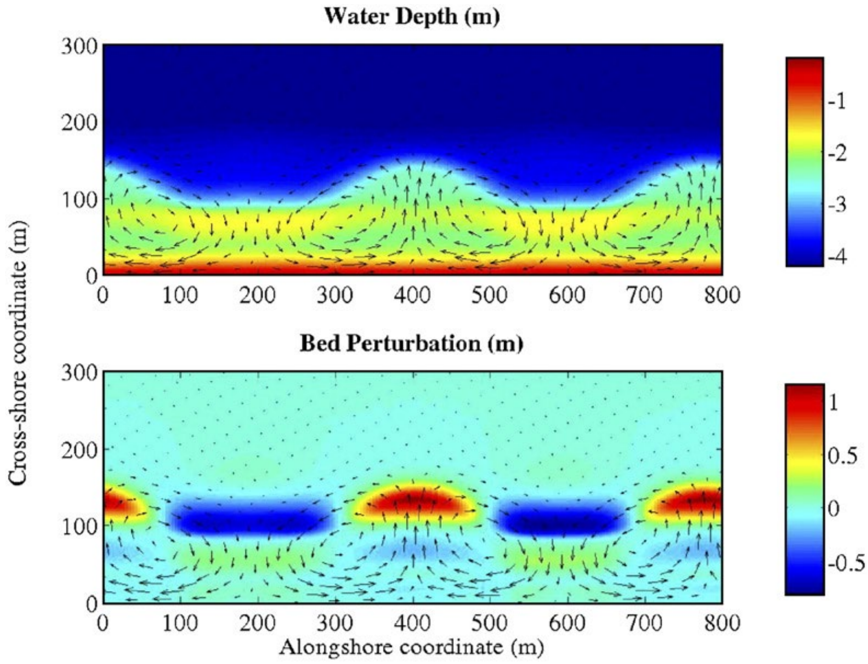

Figure 3. Stability/Self-organization mechanism: crescentic bars emerge from the feedback between flow and bathymetry (initial longshore-uniform bar bathymetry). Top: waves breaking on shoals force onshore flow, which returns to sea in gaps. Bottom: for suspended sediment decreasing from breaker line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

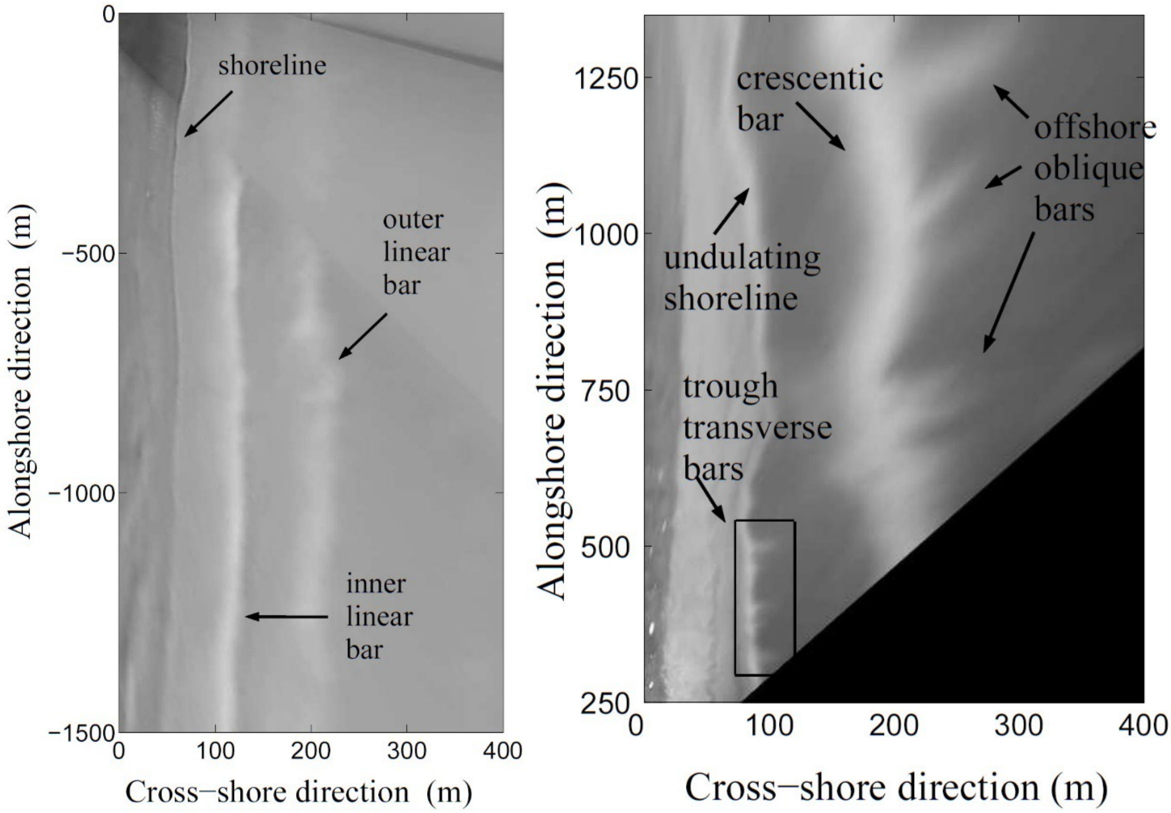

Figure 4. Left panel: planview image of two shore-parallel sand bars at Noordwijk beach (The Netherlands, 19th November 2000). Right panel: Planview image of surf zone complex topography at Duck (U.S.A. Atlantic coast, 10th January 1994). Adapted from Ribas Prats (2003).

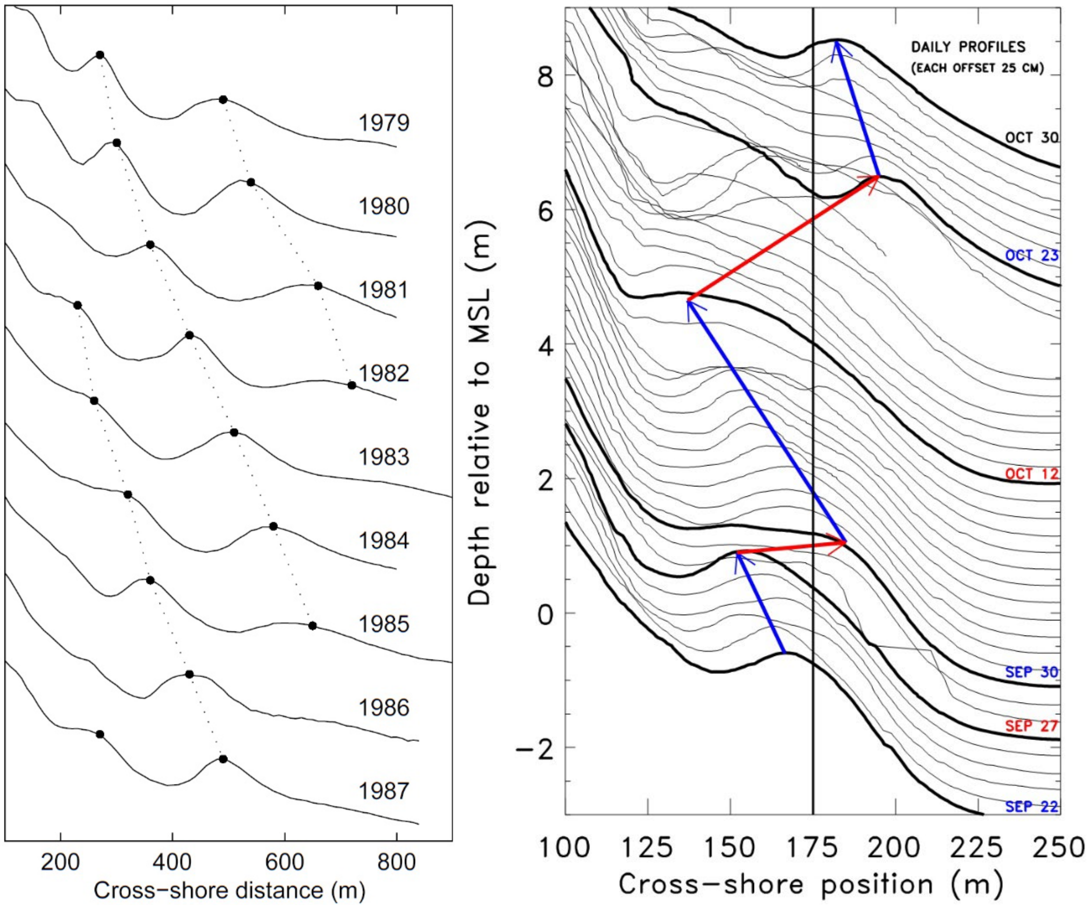

Figure 5. Left panel: seabed cross-shore profiles at Noordwijk, The Netherlands, from 1979 to 1987, showing cyclic bar behavior with a recurrence interval of about 4-5 years. Adapted from van Enckvort, (2001). Right panel: seabed cross-shore profiles at Duck (U.S.A. Atlantic coast, September-October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

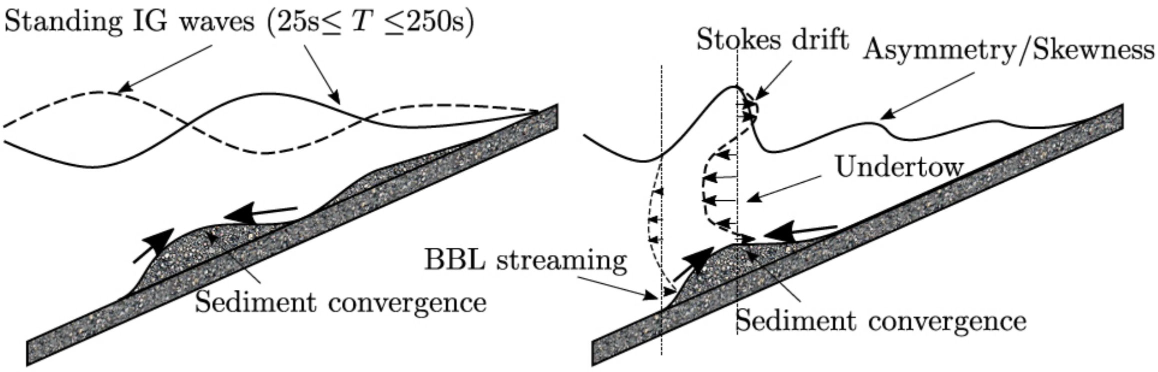

Figure 6. Bar formation because of sediment convergence at: anti/nodes of standing IF waves (left panel), breakpoint (right panel).

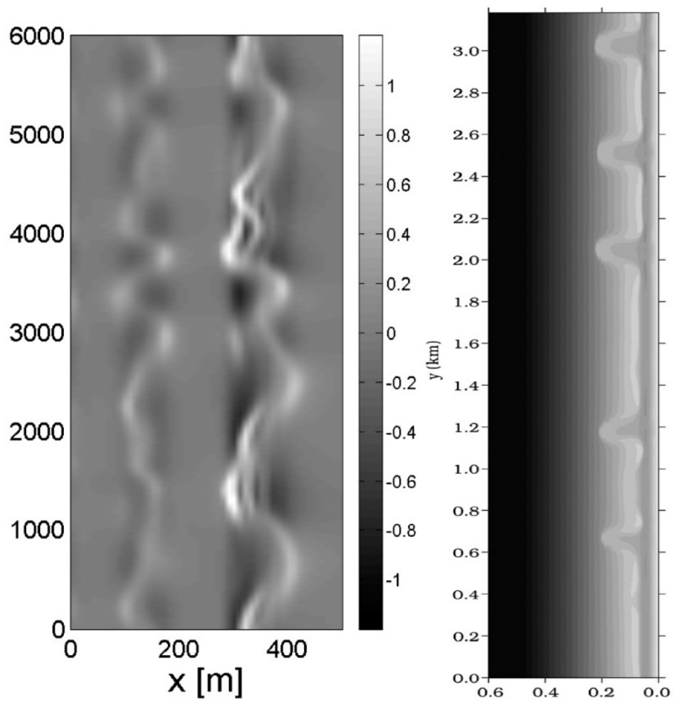

Figure 7. Left panel: non-linear stability solution for a double sand bar system giving asymmetric crescentic patterns on inner and outer bars (adapted from Klein and Schuttelaars, 2006). Right panel: Evolution of longshore uniform sand bar after 30 days of model run with normal wave incidence. Longshore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

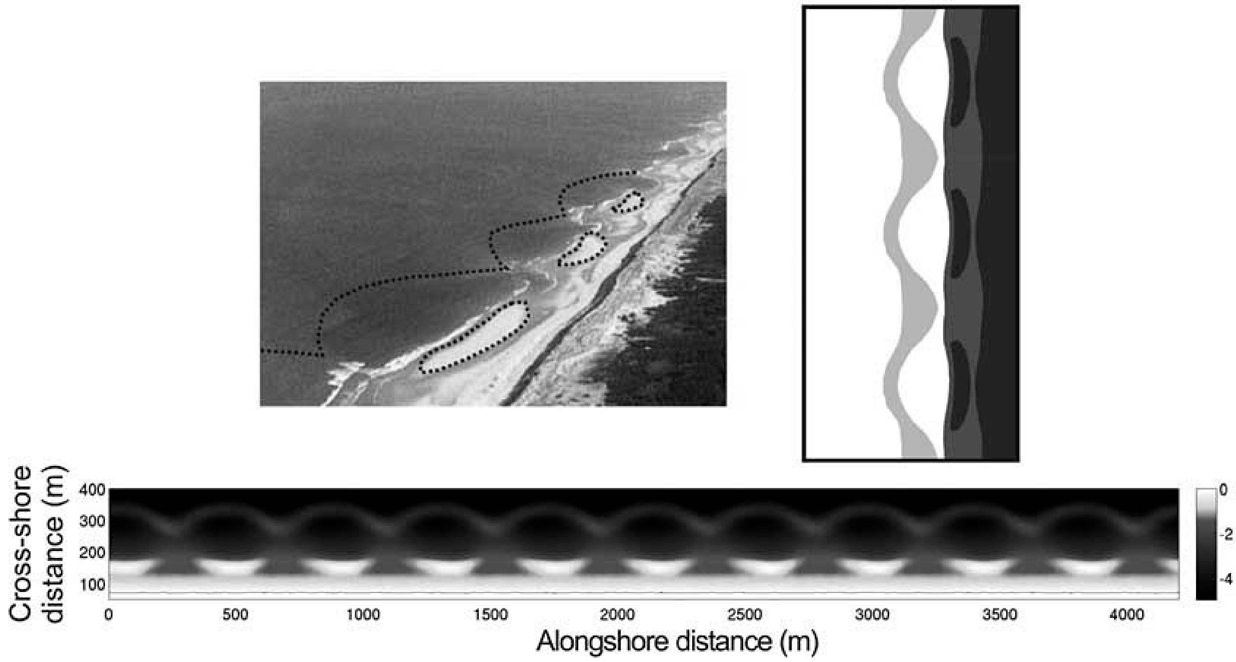

Figure 8. Computed (top-right and bottom panels) 180° out of-phase coupled patterns with an outer-bar horn facing an inner-bar bay, similar to observations (top-left panel) on a meso-macro-tidal, high-energy, double-barred beach (adapted from Castelle et al., 2010).

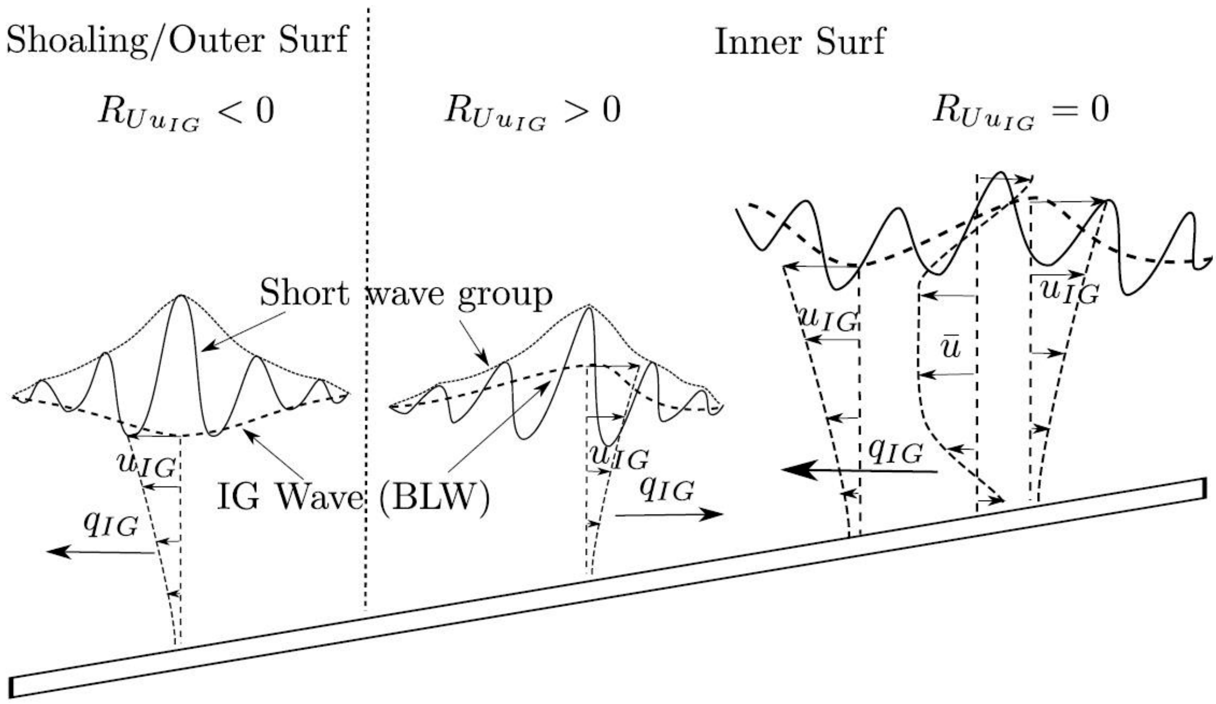

Figure 9. Sketch of IG-sea wave interaction and cross-shore sediment transport in the nearshore. From left to right: shoaling zone, inner-surf zone of moderately-sloping beaches, inner-surf zone of mildly-sloping beaches.

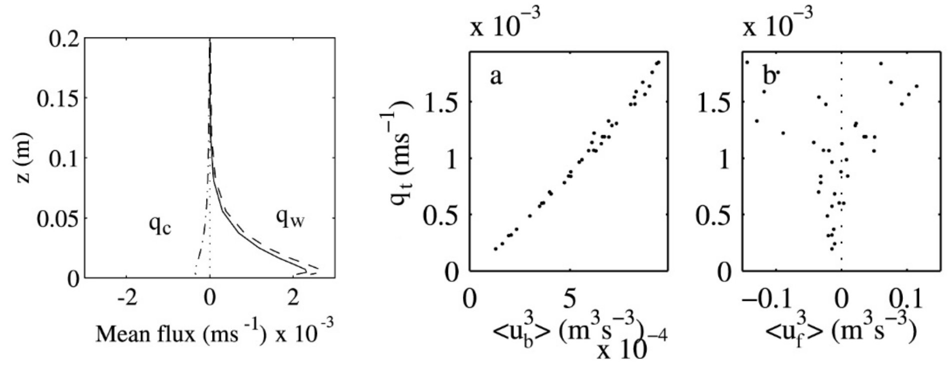

Figure 10. Top: observed (lines) and predicted (symbols) seabed profiles at the beginning (dashed line) and end (solid line and symbols) of the Duck94 onshore bar migration event. Bottom, from left to right: vertical profile of mean cross-shore sediment transport due to waves (q_w) and currents (q_c),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

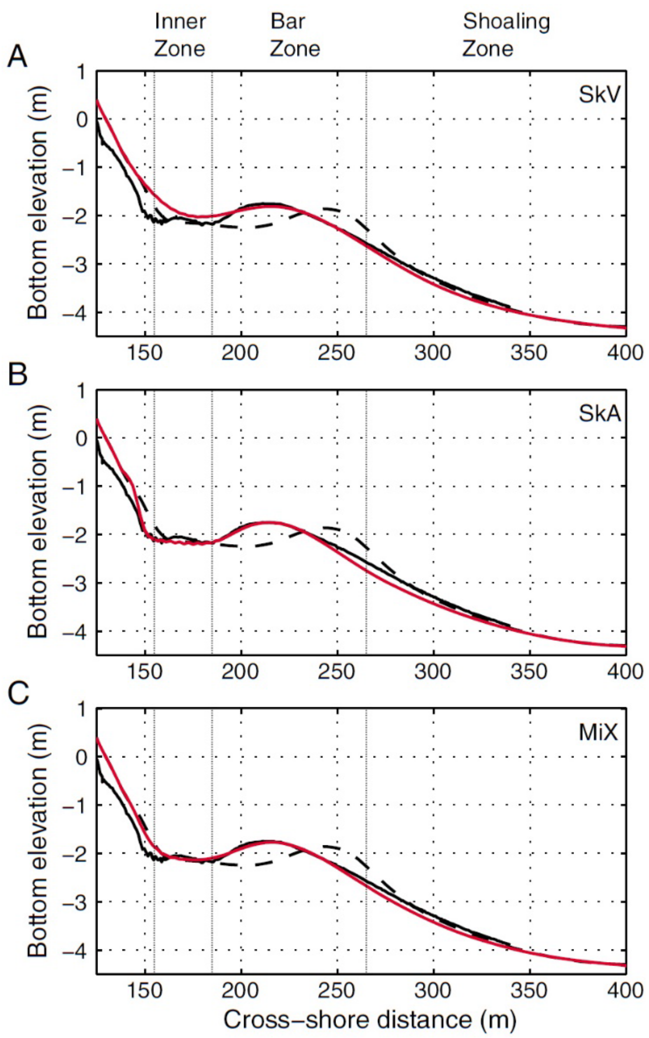

Figure 11. Modeling (red line) of the morphological evolution of the Duck94 onshore migration event for (A) the SkV model, (B) the SkA model, and (C) the MiX model. The initial and final measured profiles are shown by dashed and solid lines, respectively. Adapted from Fernandez-Mora et al. (2015).

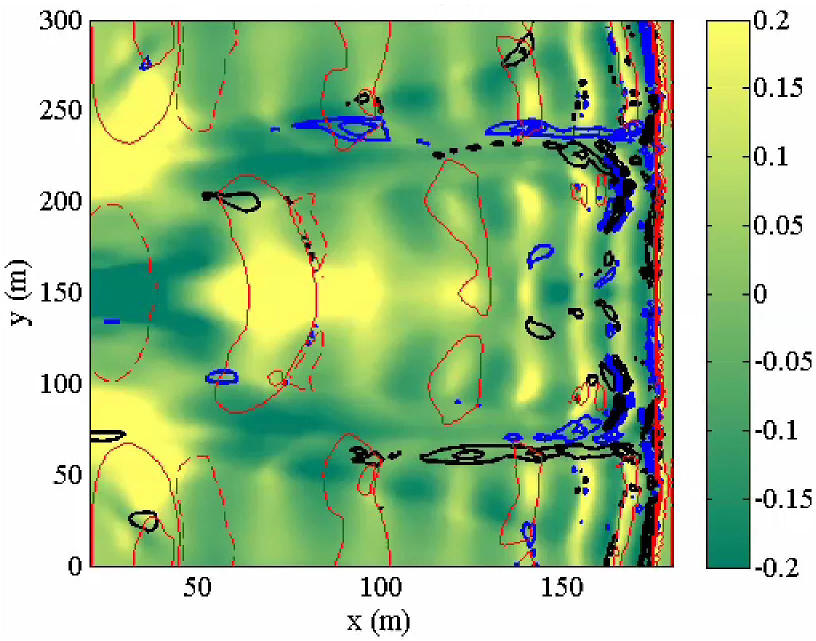

Figure 12. Seabed evolution forced by sea waves and macrovortices over a planar beach. Red thin lines give contours of the free surface elevation, while black and blue solid lines are contours of clockwise and anticlockwise vorticity, respectively. Yellow gives sand deposition and green seabed erosion.

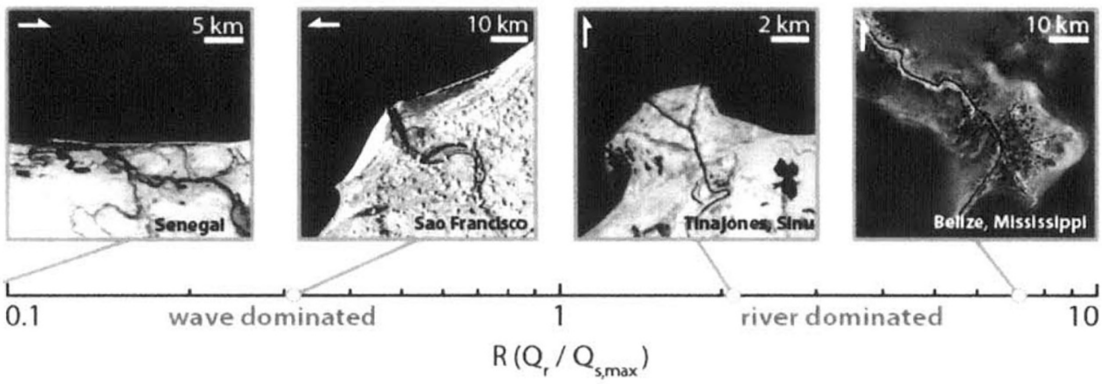

Figure 13. Classification of river delta morphologies by means of the fluvial dominance ratio. Adapted from Nienhuis (2016).

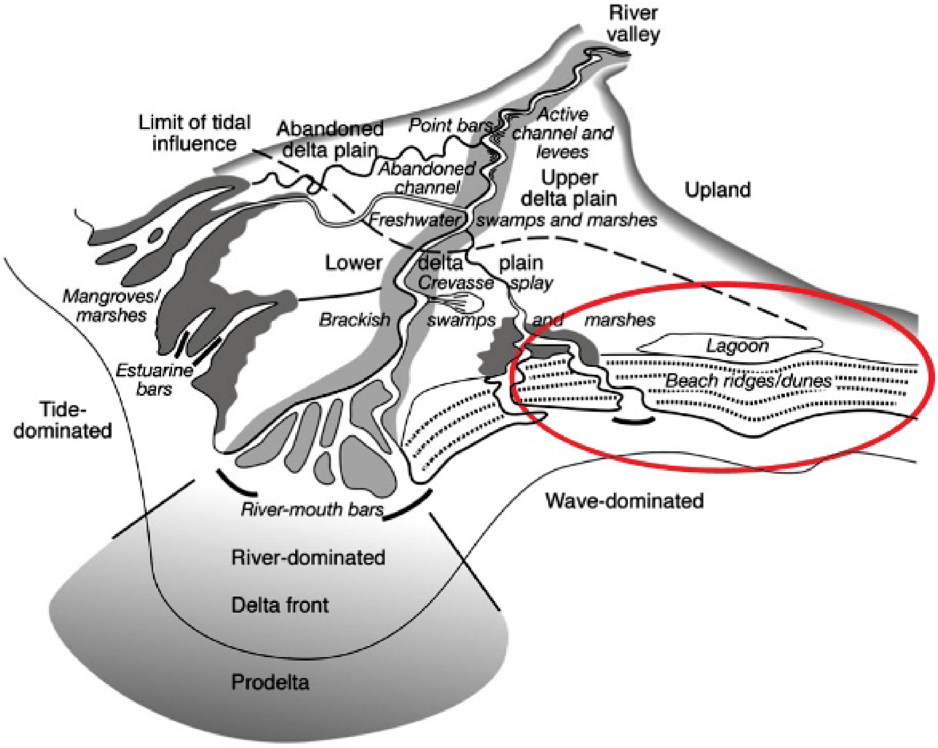

Figure 14. Left: river delta geomorphic units and schematic shoreline morphology (wave-dominated morphology circled in red). Adapted from Anthony (2015).

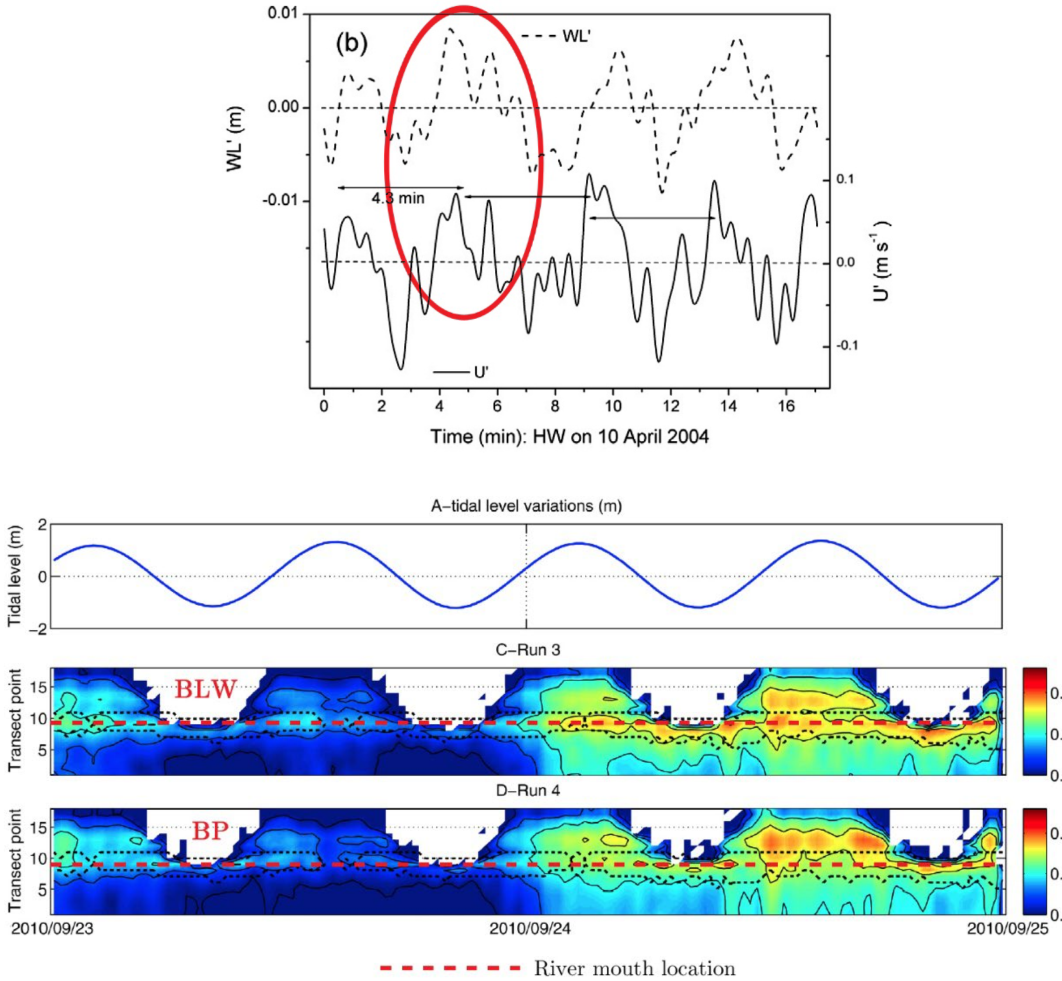

Figure 15. IG wave propagation at river mouths. Top: positive water level-flow velocity correlation at the Ría de Santiuste (Spain). Adapted from Uncles et al. (2014). Bottom: tide-modulated IG wave propagation in the Albufeira Inlet (Portugal). No IG waves propagate landward of the mouth at ebbs. Adapted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

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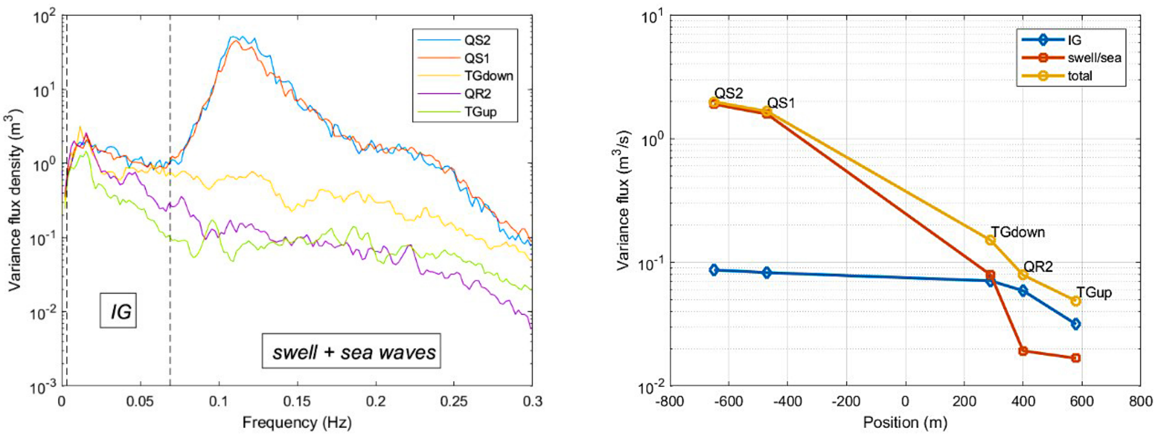

Figure 17. Energy density flux (left) and energy flux (right) at the Misa River during storm conditions. Adapted from Melito et al. (2018a).

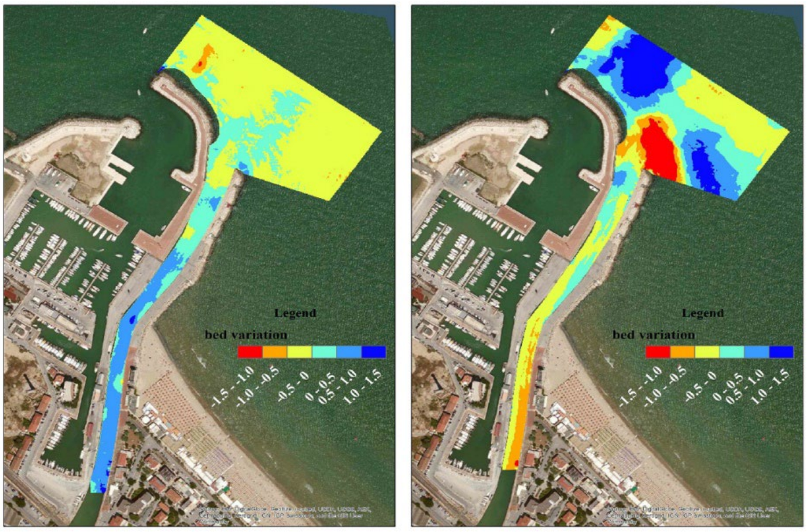

Figure 18. Seabed variation between May-September 2013 (left) and between September 2013-February 2014 (right) at the Misa River. Adapted from Brocchini et al. (2017).

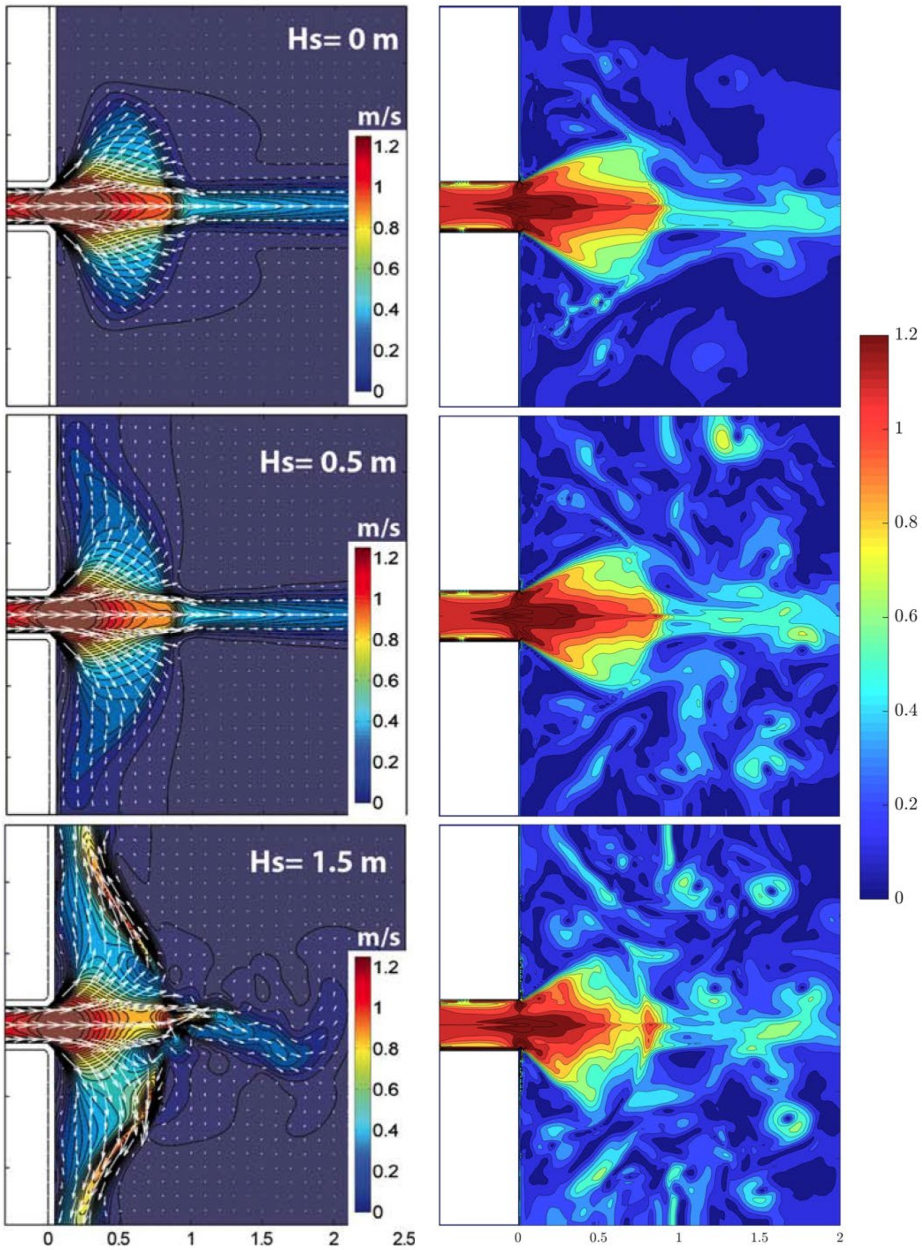

Figure 19. Velocity maps for a strong river outflow opposed to three different wave regimes (from top to bottom: H_s = 0.0m, 0.5 m, 1.5 m). Left column: results from Olabarrieta et al. (2014). Right column: results from the wave-resolving solver of Brocchini et al. (2001), averaged over 10 wave periods.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

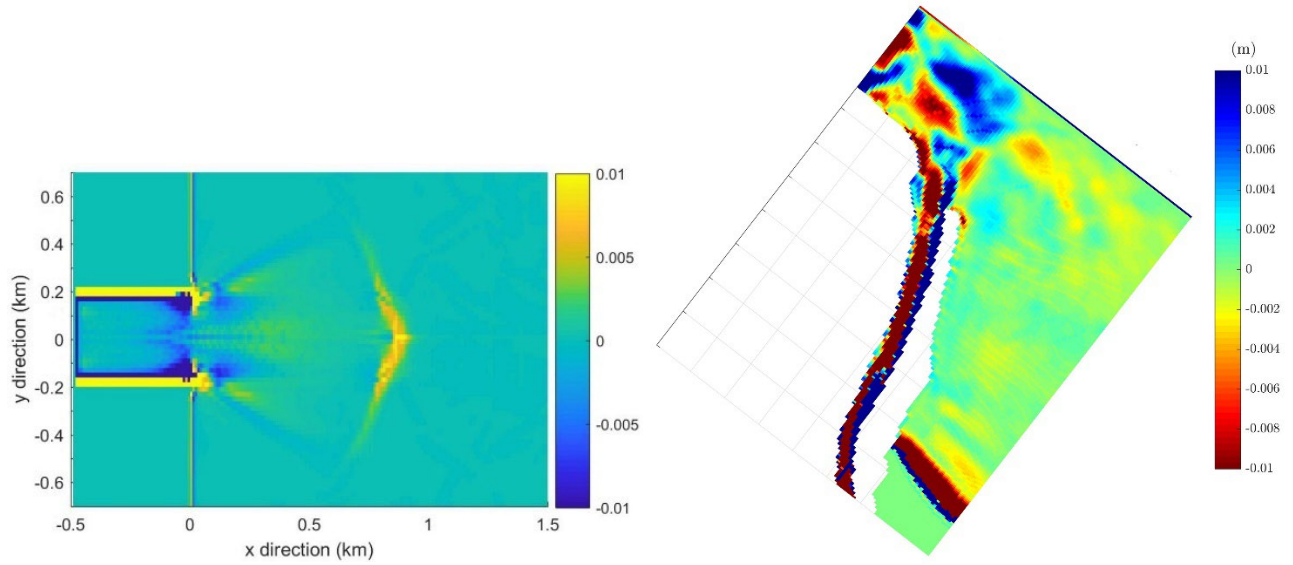

Figure 20. Estimated morphological evolution at the simplified river mouth bathymetry of Olabarieta et al. (2014), left panel and at the Misa River, right panel.

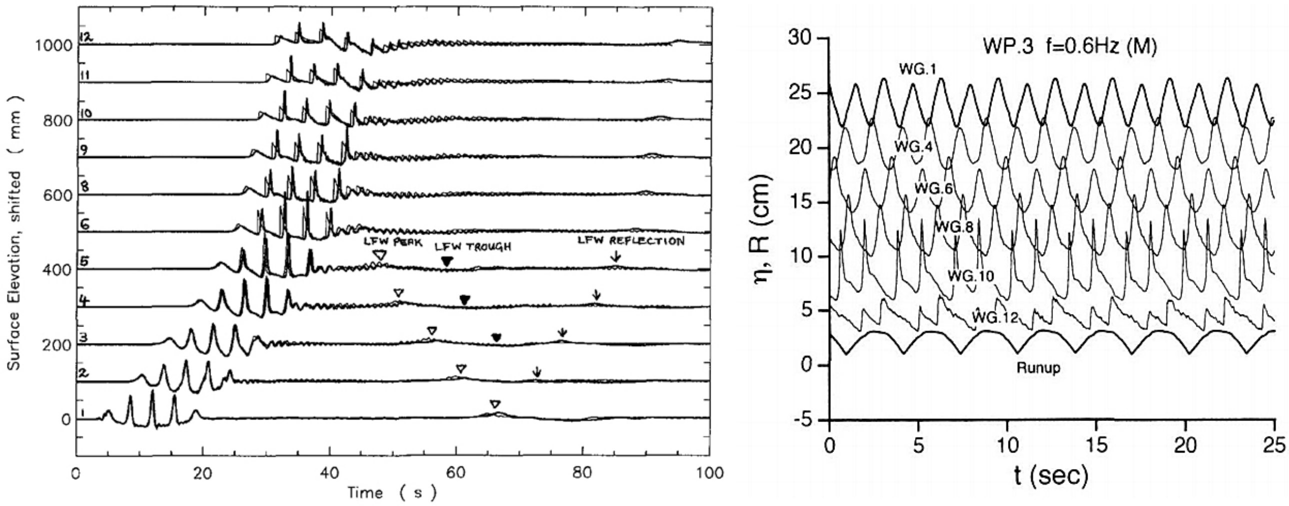

Figure 21. Left: example of BLW release on a beach because of sea wave breaking. The shoreward-propagating (from bottom to top) wave group releases its BLW that is reflected at the shoreline and propagated (from top to bottom) out to sea as a FLW (adapted from Watson, Barnes & Peregrine 1995). Right: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

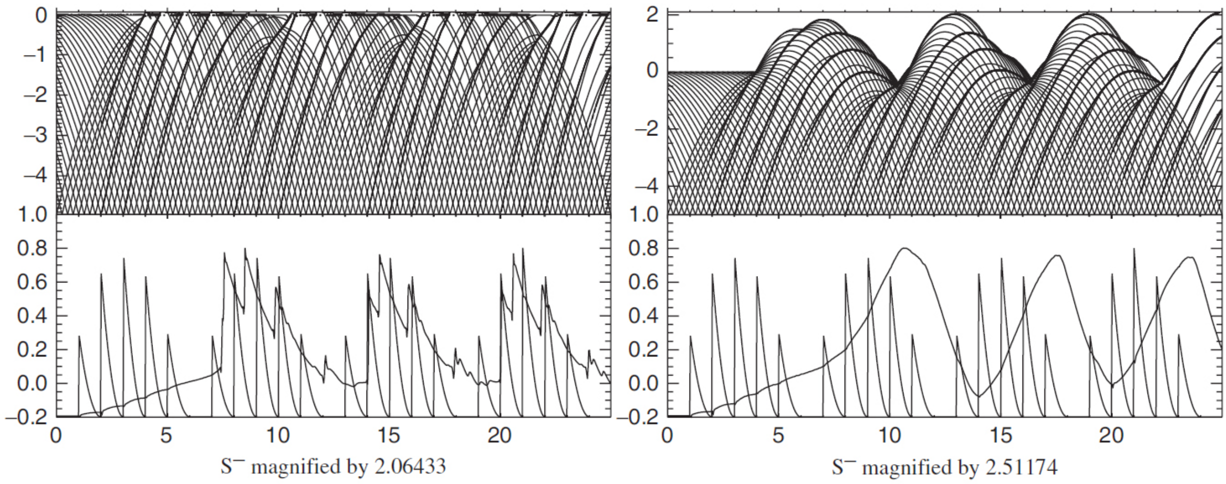

Figure 22. Waves interacting at a wall (left) and at a swash zone (right). Top: incoming and outgoing characteristic curves at the shoreline. Bottom: a group of 5 sea waves (thin line) incoming to shore and outgoing IG waves (S^-, thick line) resulting from the interaction of sea waves (adapted from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

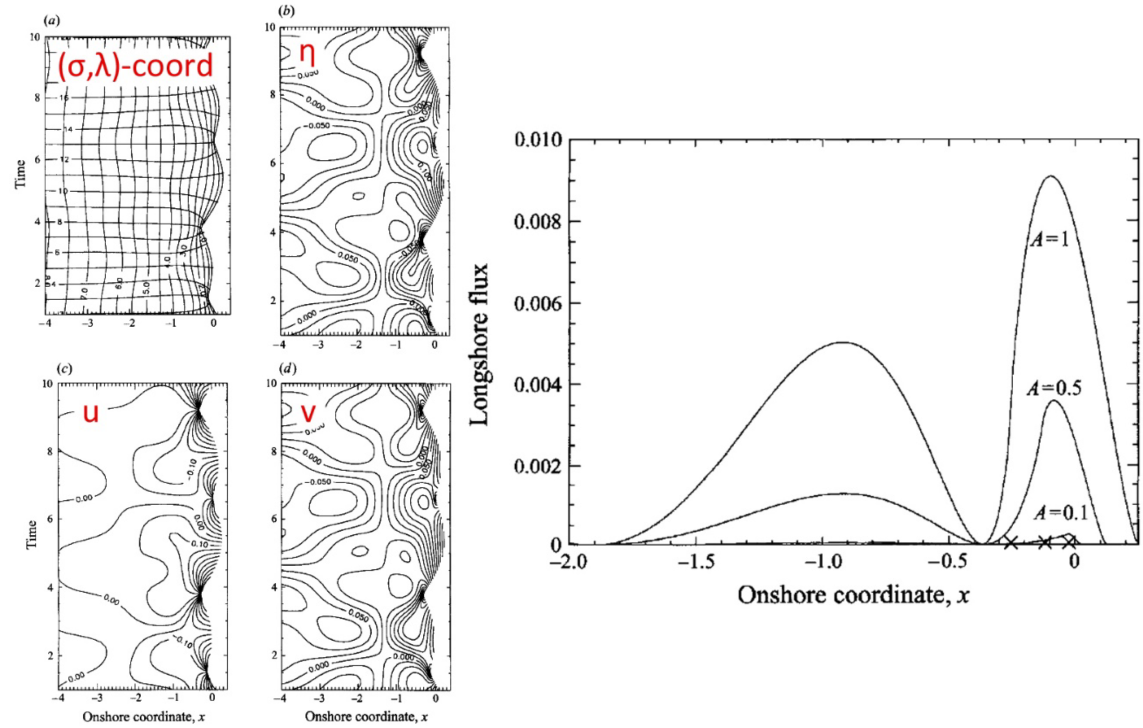

Figure 23. Left: maps, in the *(x,t)* plane of hodograph coordinates, free surface elevation *η* crosshore (*u*) and longshore (*v*) velocities for two sea waves interacting onto a beach. Right: mean longshore mass flux near and inside the swash zone for waves of different amplitude (adapted from Brocchini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

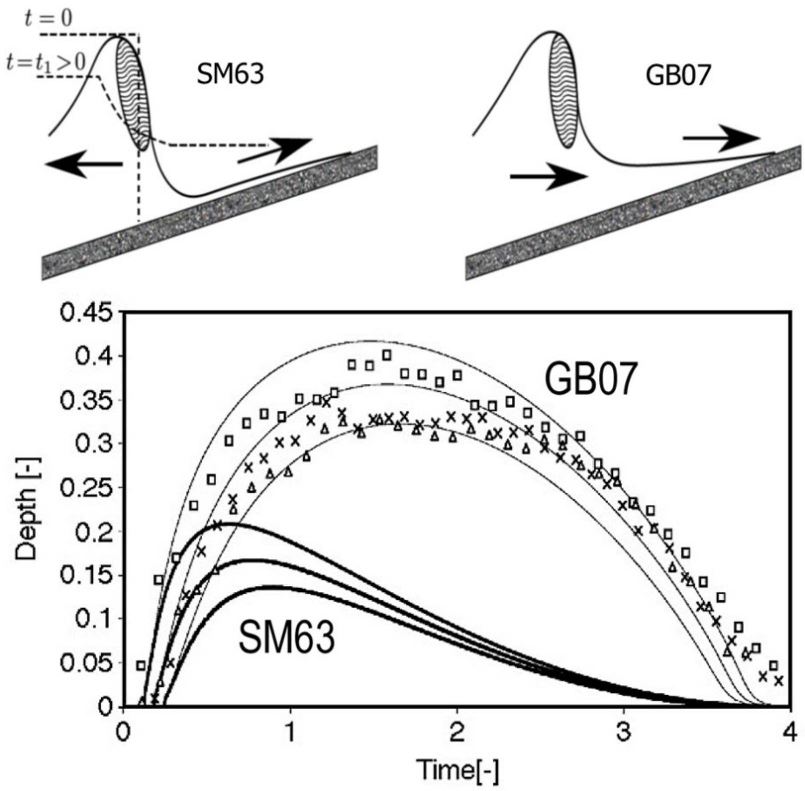

Figure 24. Top: sketches of the flow by bore collapse. The left sketch shows the evolution of the SM63 solution in comparison to a dam-break flow (dashed lines), the right sketch suggests that only shoreward-accelerating fluid motion is predicted by the GB07 solution. Bottom: swash water thickness predicted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

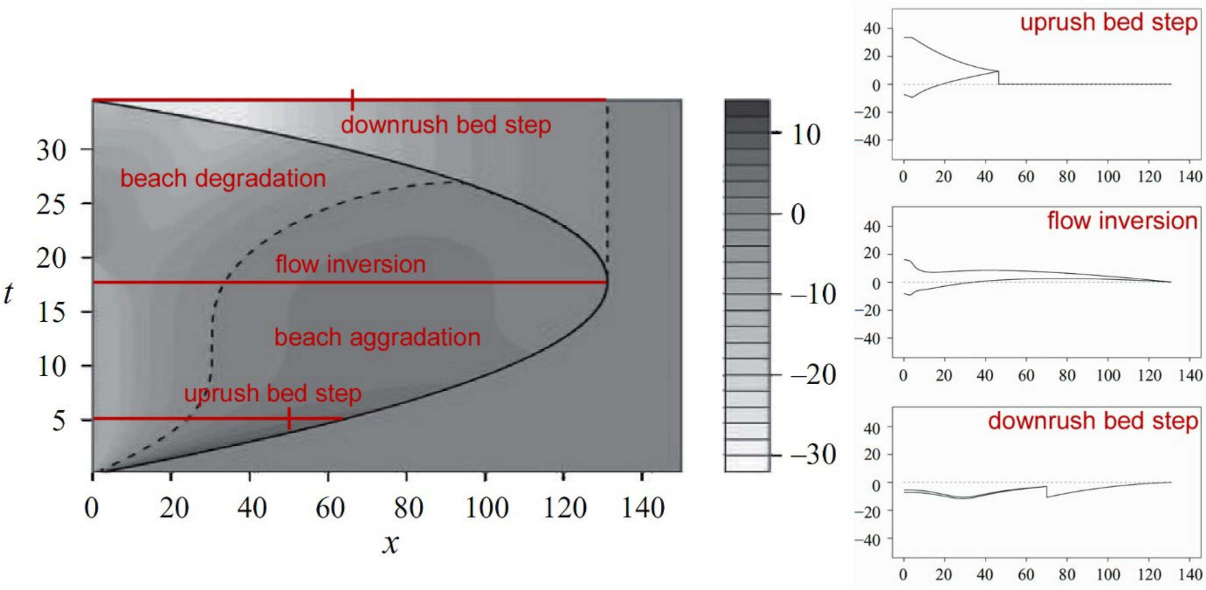

Figure 25. Left: space-time map of the changes in beach profile due a swash event. The thick dashed lines give the zero contour. Right: snapshots of the swash lens cross-shore sections (adapted from Kelly & Dodd, 2010).

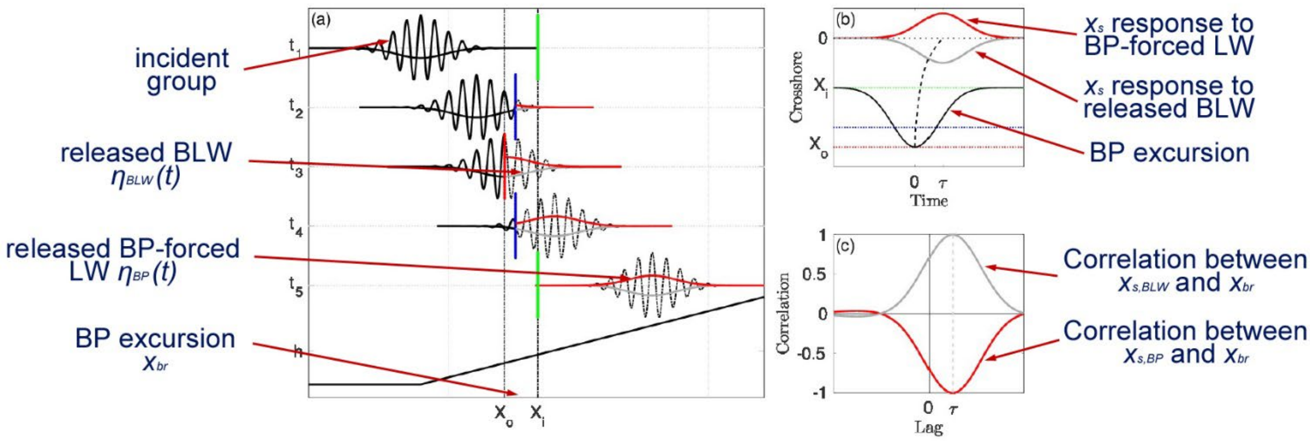

Figure 26. Shoreline relation with BLW and BP forcing. (a) Wave group. The vertical blue line gives the BP between its outer (red), x_o, and inner (green), x_i location. Grey and red lines give the released BLW and incident BP-forced IG wave, respectively. (b) BP excursion (black) and shoreline response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

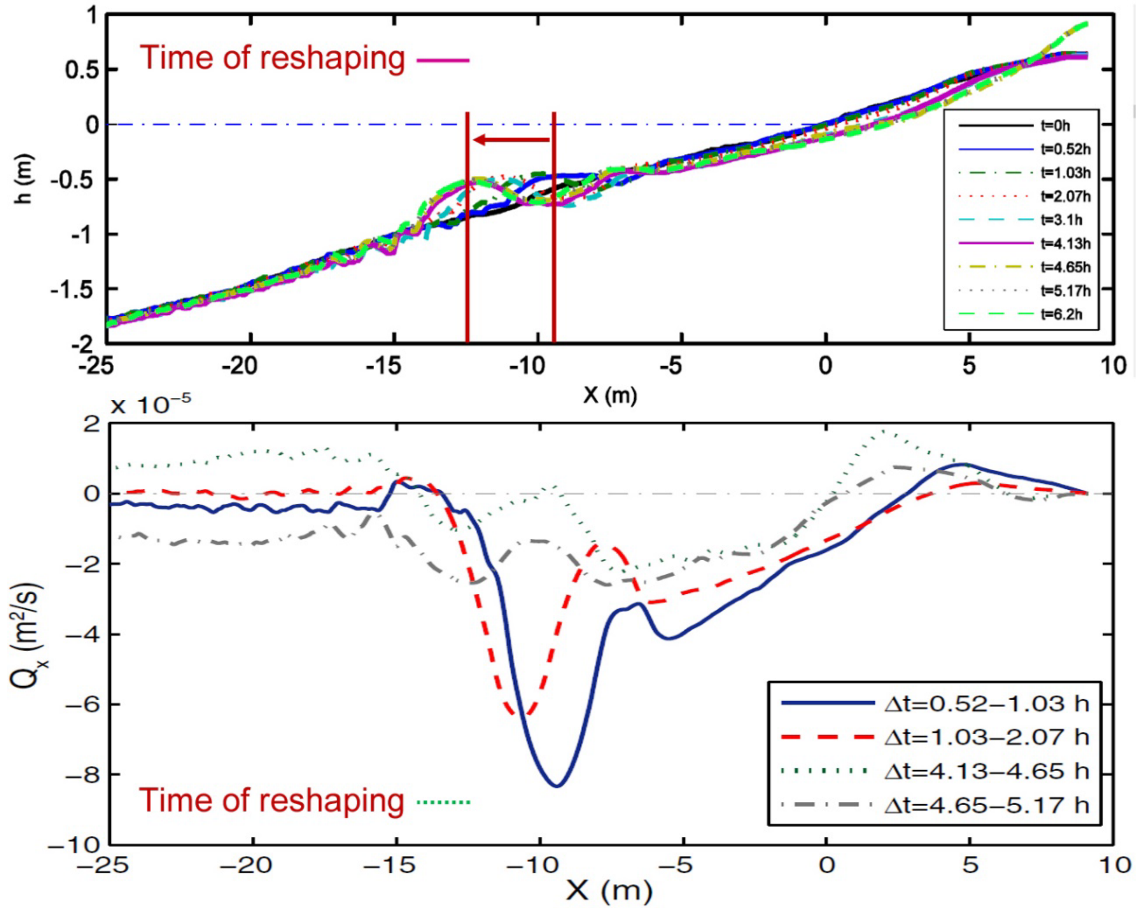

Figure 27. Cross-shore distribution of beach profile (top panel) and of cross-shore sediment transport rate (bottom panel) at various times before and after beah reshaping (adapted from Alsina et al., 2012).

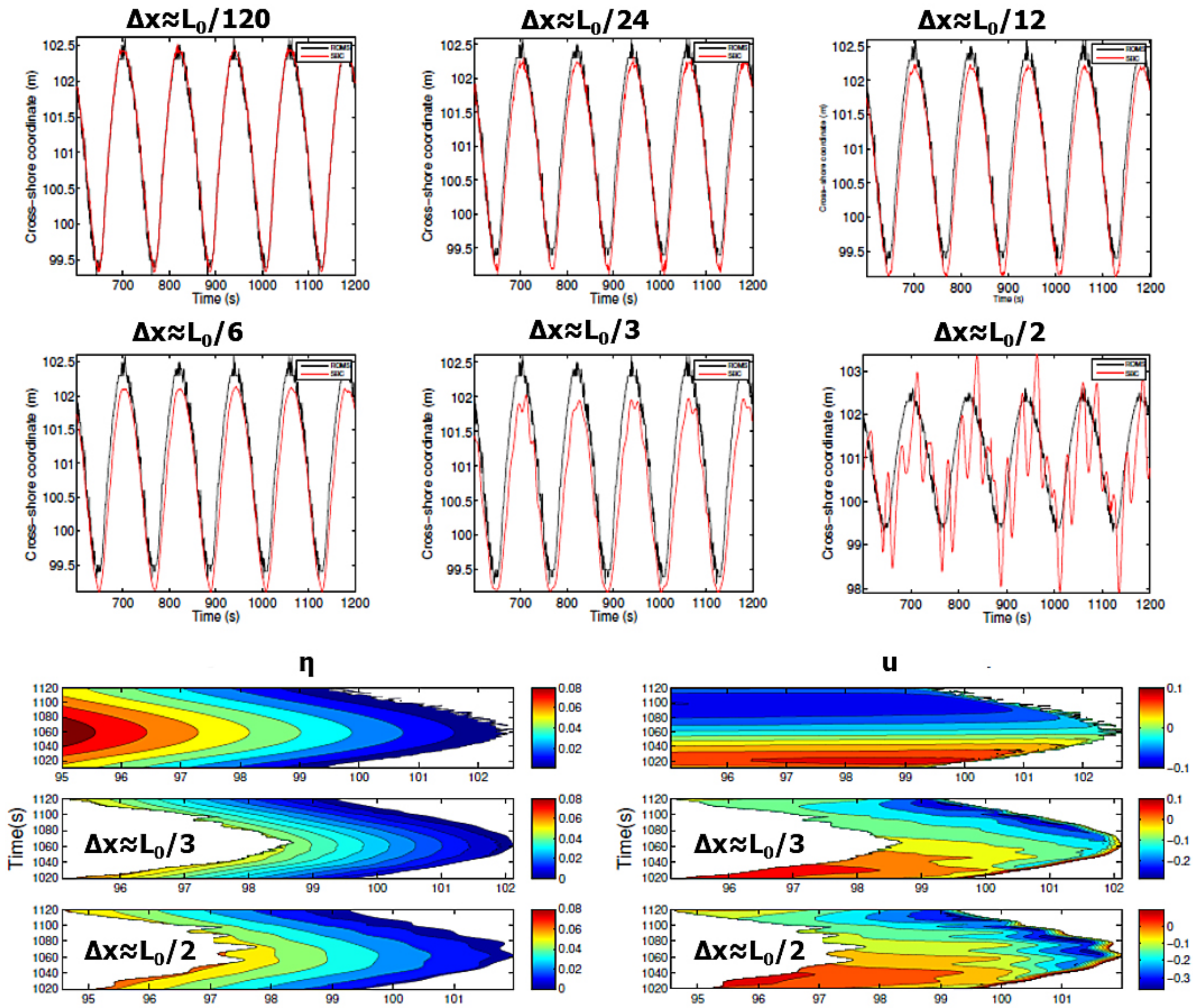

Figure 28. Top 6 panels: comparison of the ROMS ultra-highly resolved shoreline (black line) and ROMS+SBCs shorelines for lower and lower grid resolution. Bottom 2 panels: free surface elevation (left) and onshore velocity (right) for the benchmark ROMS solution (top panels) and ROMS+SBCs solutions for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