Simulating morphodynamics with unstructured grids: Description and validation of a modeling system for coastal applications

Xavier Bertin *, Anabela Oliveira, André B. Fortunato

Estuaries and Coastal Zones Division, National Laboratory of Civil Engineering, Av. do Brasil, 101, 1700-066 Lisbon, Portugal

Abstract

Morphodynamic modeling systems are being subjected to a growing development over the last decade and increasingly appear as valuable tools for understanding and predicting coastal dynamics and morphological changes. The recent improvements of a 2DH unstructured grid morphodynamic modeling system are presented in this paper and include the implementation of an adaptive morphodynamic time step, the integration and full coupling of a wave model and the forcing by large scale wave and tide models. This modeling system was first applied to a dissipative wave-dominated beach located on the French coast, where the availability of field data allowed for a fine calibration and validation of wave-induced flows and longshore transport, and an assessment of the various sediment transport formulae. The modeling system was then applied to a very dynamic Portuguese tidal inlet where numerical tests show the computational efficiency of using an adaptive time step. Morphodynamic simulations of this inlet with real wave and tidal forcings resulted in realistic morphological predictions. The two applications show that the improved modeling system is able to predict hydrodynamics, transport and morphological evolutions in complex coastal environments.

1. Introduction

Morphodynamic modeling systems consist of a set of modules to simulate shallow water flows, wave propagation, sediment transport and bottom evolution. These systems have been in extensive development over the past 15 years in Europe (de Vriend et al., 1993; Wang et al., 1995; Johnson et al., 1994; de Vriend, 1996; Nicholson et al., 1997; Lesser et al., 2004) and in the US (Holliday et al., 2002; Kubatko et al., 2006; Long et al., 2008). The increasing volume of literature on coastal morphodynamic modeling over the past five years highlights the growing interest of coastal researchers and engineers in these techniques. Among these recent publications, several were focused on the numerical methods used to solve the Exner equation (Johnson and Zyserman, 2002; Callaghan et al., 2006; Fortunato and Oliveira, 2007a; Long et al., 2008), others presented and validated new modeling systems (Lesser et al., 2004; Fortunato and Oliveira, 2004; Kubatko et al., 2006; Saied and Tsiaras, 2008), while others described applications of pre-existing modeling systems to complex coastal environments (Sutherland et al., 2004; Grunnet et al., 2004; Jones et al., 2007).

Since the developments of these modeling systems have been fuelled, to a large extent, by the need to address coastal engineering problems, one would expect successful applications to grow rapidly, as new and better tools became available. Yet, these applications remain scarce in the literature and do not seem to be increasing (Cayocca, 2001; Work et al., 2001; Grunnet et al., 2004). This observation suggests that simulating coastal morphodynamics remains a challenging task and that increasing the sophistication of the models does not necessarily improve the quality of morphological predictions. Indeed, models comparisons do not always recommend the most sophisticated approaches. For instance, Grunnet et al. (2004) suggest that the improved representation of physics brought by a fully 3D approach may not outweigh its higher computational cost, and that using a 2DH approach may perform equally well if the final goal is bathymetric evolution solely.

This study presents recent developments of the unstructured grid morphodynamic modeling system MORSYS2D (Fortunato and Oliveira, 2004, 2007a). This modeling system aims at simulating hydrodynamics, transport of non-cohesive sediments and morphological evolutions in real coastal systems driven by tides, waves, wind and river flows, such as tidal inlets. MORSYS2D was improved in the spirit of integrating less sophisticated physics than other modeling systems, which often implies excessive model tuning and large computational resources, but placing greater emphasis on the forcings and on the representation of the main processes, to perform long-term simulations with a reasonable computational cost. These recent developments are described in Section 2 and include the implementation of a time-adaptive morphodynamic time...
The morphodynamic modeling system MORSYS2D

2.1. General outline

The modeling system MORSYS2D aims at simulating the non-cohesive sediment dynamics and bottom changes in estuaries, tidal inlets and coastal regions, driven by tides, waves, wind and river flows. The system integrates the hydrodynamic models ADCIRC (Lueftich et al., 1991), www.adcirc.org and ELCIRC (Zhang et al., 2004, www.coralm.org/modeling/elcirc/index.html), the wave models SWAN (Booij et al., 1999), www.wldelft.nl/soft/swan) and REF/DIF1 (Dalrymple and Kirby (1991), chinacat.coastal.udel.edu/programs/refdif/refdif.html) and the sand transport and bottom update model SAND2D (Fortunato and Oliveira, 2004, 2007a). The system consists of C-Shell scripts that run independent models, manage the transfer of information between them, perform control checks and store result outputs. The applications presented in this study use the models SWAN, ELCIRC and SAND2D (Fig. 1).

2.2. The wave module SWAN

The spectral wave model SWAN solves the wave action density balance equation (Booij et al., 1999) and is used in MORSYS2D in stationary mode to simulate wave propagation and deformation from the open sea up to the shoreline. This model can account for bottom friction, wave breaking (Madsen et al. (1988)), wind contributions, quadruplets, triad wave-wave interaction, wave-current interactions and wave propagation within a non-uniform water level. The physical processes to be considered by SWAN are specified by the users in a parameter file and three options are available for each physical process: (1) switched off; (2) switched on with default parameters and (3) switched on with user-specified parameters. Bottom friction coefficient, wind, water level and currents can be considered: (1) constant in space and time; (2) constant in space, variable in time and (3) variable in time and space. In the case where interactions with currents or propagation within a non-uniform water level are considered, SWAN is fed by outputs of velocities and elevations from the hydrodynamic model (ELCIRC) linearly interpolated over regular grids. Depending on the purpose of the study, SWAN can be forced at its open boundary by: (1) time-independent wave parameters, (2) time series of wave parameters originating from the WAVEWATCHIII (WW3) model (Tolman et al., 2002); or (3) by wave spectra originating from the WW3 model or from an oceanic wave buoy.

Depending on the size and the geometry of the modeled domain, several computational grids can be nested, with resolutions for the corresponding spatial computational grids ranging typically from 1000 m on the external continental shelf to 10 m in the coastal area of interest. Generally, the last nested grid is oriented parallel to the coast and includes the whole area where bathymetric changes are being simulated. For computational efficiency, this grid is the only one for which the bathymetry is updated during a morphodynamic simulation. The spatial resolution of the finest computational grid is adjusted depending on the wave height and the beach gradient in order to provide a good description of the surf zone.

Significant wave height, period, direction, wavelength, and orbital velocity are outputted from SWAN and used to compute gradients of radiation stresses to force the hydrodynamic model (ELCIRC) and to compute sand fluxes into SAND2D. The radiation stress gradients are computed from the formulation of Longuet-Higgins and Stewart (1964) using centered differences in the SWAN grid. A final check is performed after linear interpolation on the hydrodynamic grid to avoid imposing radiation stress gradients on very small water depths and subsequent aberrant velocities. This check is required because of two related issues: (1) water level is taken as constant within the elements of Elcirc, which means that one element is dry if at least one of its three nodes is dry and (2) due to the different space and time resolutions of the wave and flow models, excessive gradients of radiation stresses occasionally appear in the flow grid. Finally, the updating interval of the wave and radiation stress fields is user-specified and should be adjusted for each run as a decreasing function of the ratio between tidal range and wave height. As a rough guideline, a
2.3. The hydrodynamic module ELCIRC

ELCIRC was developed as an open source community model at the Center for Coastal Margin Observation and Prediction (Zhang et al., 2004). ELCIRC solves the fully non-linear, three-dimensional, barotropic shallow water equations, coupled to transport models for salt and heat. Within MORSYS2D, a single vertical layer is used, and ELCIRC reverts to a 2D depth-averaged model. Forcings include tides, tidal potential, river flow, wind or waves-induced radiation stresses and solar radiation. A Manning friction law was integrated, with the friction coefficient either constant or variable in space. Although with important implementation differences (e.g. treatment of baroclinicity and of tangential velocities), the numerical solution is inspired from that of UnTRIM (Casulli and Zanolli, 1998).

The equations are solved with a finite volume technique for volume conservation and a natural treatment of wetting and drying. The effect of short waves on the hydrodynamics was achieved through the forcing by the gradients of wave radiation stresses, which, for depth-averaged models, can be computed through the formulation proposed by Longuet-Higgins and Stewart (1964).

Since ELCIRC is used in 2DH mode, the gradients of radiation stresses can be regarded as a surface stress \( \tau_{xy} \) (Blain and Cobb, 2003), which reads:

\[
\tau_{xx} = -\frac{\partial S_{xx}}{\partial x} + \frac{\partial S_{yx}}{\partial y} \tag{1}
\]

\[
\tau_{yy} = -\frac{\partial S_{yy}}{\partial y} + \frac{\partial S_{xy}}{\partial x} \tag{2}
\]

where \( S_{xx}, S_{yy}, S_{xy} \) and \( S_{yx} \) are the wave radiation stress terms given by Longuet-Higgins and Stewart (1964)

\[
S_{xx} = \frac{E}{2} \left( \frac{C_s^2}{C} \cos^2 \alpha + 1 \right) \tag{3}
\]

\[
S_{yy} = \frac{E}{2} \left( \frac{C_s^2}{C} \sin^2 \alpha + 1 \right) \tag{4}
\]

\[
S_{xy} = S_{yx} = \frac{E C_s}{C} \sin \alpha \cos \alpha \tag{5}
\]

In the previous equations, \( E = \frac{1}{2} \rho g H_{\text{rms}}^2 \) is the wave energy, \( \alpha \) is the wave angle to the x axis, \( C_s \) is the wave group velocity, \( C \) the wave phase velocity and \( \rho \) is the water density. Wave-induced horizontal diffusion was also taken into account in the momentum equations using an eddy viscosity approach, where the horizontal eddy viscosity coefficient \( \nu_t \) was computed at each node according to Battjes (1975):

\[
\nu_t = M \cdot H_{\text{rms}} \left( \frac{\nu_b}{\rho} \right)^\frac{1}{2} \tag{6}
\]

In the previous equation (Eq. (6)), \( M \) is a dimensionless turbulent parameter set to 1 and \( \nu_b \) is the wave energy dissipation coefficient given by Thornton and Guza (1983).

2.4. The sand transport and bottom evolution module SAND2D

The bottom update model simulates sand transport due to waves and currents using one of several semi-empirical formulae and computes the resulting bed changes through the Exner equation (Fortunato and Oliveira, 2004, 2007a). This equation is solved with a node-centered finite volume technique based on a triangular unstructured grid, and using a predictor–corrector method:

\[
\Delta h^i = \frac{1}{1 + A} \nabla Q^i \tag{7}
\]

where \( \alpha \) is the sediment porosity, \( h \) is the water depth and \( Q^i \) is the sediment flux integrated over the morphological time step which includes a diffusive term:

\[
Q = Q + \epsilon (1 - \alpha) \left( Q_x \frac{\partial h}{\partial x} + Q_y \frac{\partial h}{\partial y} \right) \tag{8}
\]

In Eq. (8), \( \epsilon \) is an user-specified constant to tune the artificial diffusion and \( Q \) is the sediment flux computed at the center of the elements and integrated in time between steps \( n \) and \( n+1 \):

\[
Q = \int_{n}^{n+1} q(u(t), \eta(t), h^n, U_{\text{orb}}, ...) dt \tag{9}
\]

In Eq. (9), \( q \) is the instantaneous sand flux, which is a function of the velocity \( u \), the tidal elevation \( \eta \) and, in the presence of waves, other parameters like the wave orbital velocity \( U_{\text{orb}} \). Instantaneous sand fluxes are computed using an empirical or a semi-empirical formula selected within a wide range implemented in MORSYS2D, including: (1) formulae for currents only (e.g., Ackers and White (1973), Van Rijn (1984), Bhattacharya et al. (2007)); and (2) formulae for waves and currents (e.g., Bijker (1971), Soulsby-Van Rijn (Soulsby, 1997), Ballard and Inman (1981), or the Ackers–White formula (Ackers and White, 1973) adapted to wave environments by Van de Graaff and Van Overeem (1979)). Details on these four latter formulations are given in Appendix A.

In earlier versions of the model (Fortunato and Oliveira, 2004), velocities and water levels were fed into SAND2D in the frequency domain (i.e., through tidal amplitudes and phases, rather than time series). While this method was efficient for slowly evolving systems, it also had severe drawbacks. First, using harmonic analysis in wave-driven flows is inadequate. Secondly, its application to very dynamic coastal systems required the use of morphological factors well below unity (Fortunato and Oliveira, 2007a) to prevent large Courant numbers and subsequent numerical oscillations. Besides being very time-consuming, this method implied using a representative tide (M2 only), which could be problematic when trying to reproduce the behavior of a coastal system.

In spite of the predictor–corrector algorithm, the stability of the solution is limited by the Courant number. Numerical experiments indicate the model is stable for Courant numbers of up to 10 (Fortunato and Oliveira, 2007a). The Courant number can be estimated as (Roelvink, 2006):

\[
Cu \approx \frac{b Q}{H_{\text{rms}}} \tag{10}
\]

where \( b \) is the velocity power in the transport formulae (typically between 3 and 5, depending on the specific formulation).

Maximum values of \( Cu \) for a fixed time step vary over more than three orders of magnitude within a tidal cycle, thereby imposing the need for very small time steps. To improve efficiency, an adaptive time step was implemented, targeting a user-specified constant value for \( Cu \) (Bertin et al., 2007). The determination of the next morphological time step is based on two assumptions: (i) the logarithm of \( \Delta h / \Delta t \) at time \( n \) divided by the time step, varies linearly in time and (ii) the morphological time step varies slowly from one time step to another. The first assumption was based on observations of \( Cu(n) \) during runs with a constant time step. Assumptions (i) and (ii) permit to make a prediction of the next Courant number \( Cu(n+1) \), using a log-linear extrapolation (Eqs. (11) and (12)):
\[
\frac{\ln(a_{n+1}) - \ln(a_n)}{\Delta t_{n+1}} = \frac{\ln(a_n) - \ln(a_{n-1})}{\Delta t_n}
\]

According to assumption (ii), \(\Delta t_{n+1} \approx \Delta t_n\), Eq. (12) simplifies to:

\[
a_{n+1} = \frac{\gamma_n^2}{\gamma_{n-1}}
\]

Then, in order to target a user-specified Courant number \(C_{u\text{target}}\), the next morphodynamic time step \(\Delta t_{n+1}\) can be computed as:

\[
\Delta t_{n+1} = C_{u\text{target}} \frac{\gamma_{n-1}}{\gamma_n}
\]

Finally, the last step of the procedure corresponds to the application of a relaxation factor and yields the new morphodynamic time step \(\Delta t_{n+1}\):

\[
\Delta t_{n+1} = \delta \cdot \Delta t_{n+1} + (1 - \delta) \cdot \Delta t_n
\]

where \(\delta\) is a relaxation factor which was set to 0.7. The application of this relaxation factor enforces approximation (ii) and also prevents the extrapolation of \(a\) to produce overshoots of the time step.

The implementation of this adaptive procedure leads to time steps that vary from about 2 min when sand fluxes are maximum (typically around mid-tide and/or for large wave conditions) to 45 min when sand fluxes are minimum (around high and low tide and/or when waves are small). Other authors mention using an adaptive time-stepping procedure (Nicholson et al., 1997; Broker et al., 2003), but the descriptions are not detailed enough to allow for a comparison with the proposed approach.

3. Application and validation of the morphodynamic modeling system

The morphodynamic modeling system was validated at two contrasting sites: a dissipative wave-dominated beach located in the middle of the French Atlantic coast and a very dynamic tidal inlet located in the middle of the Western coast of Portugal (Fig. 2). For the first application, the model was forced by tides and waves in order to test its ability to reproduce wave deformation, wave-induced currents and longshore transport. For the second application, the model was forced only by tides, to test the performance of the adaptive time step, and then by tides and waves, to evaluate the pertinence of morphological predictions.

3.1. Application to a wave-dominated dissipative beach

3.1.1. Model setting

In this section, the morphodynamic modeling system MORSYS2D was applied to a wave-dominated dissipative beach located within the SW part of Oleron Island, in the middle of the French Atlantic coast (Figs. 2 and 3). A 3-day field measurement campaign, including a bathymetric survey extending from the dune down to water depths of 10 m below MSL, tracer experiments and hydrodynamic measurements, was conducted on this beach on April 2005, and provided a large data set of wave parameters, longshore currents and sediment transport (Bertin et al., 2008a,b). This beach was also selected because it displayed an almost flat and dissipative morphology during the experiments, which induced relatively simple patterns of wave-induced currents (tidal influence is restricted to water level variation (Bertin et al., 2008b)). Another advantage of this site was the local macrotidal range combined to the deployment of measuring instruments within the intertidal zone, which allowed for the record of cross-shore profiles of wave parameters and wave-induced currents.

The model was run from the 4th to the 6th of April 2005 and this period was characterized by a moderate tidal range of 3.5 m and moderate WNW swells without local winds producing 0.9–1.3 m clean waves at the breaking point.

Fig. 2. General location map of the Iberic Peninsula and France showing the two studied sites (isobath lines 100 m and 1000 m are shown in bold). Boundaries of the regional tidal model of Fortunato et al. (2002) are shown in dashed lines and the output locations of the WW3 model are marked as crosses.
Three nested grids were used for the wave model, with spatial resolutions ranging from 1000 m for the coarsest grid which covers the continental shelf, up to 10 m for finest grid which covers the studied beach. The open boundary of the largest grid included a simulation point of the WW3 model (2.5°W, 46°N, Fig. 3) which was used as offshore forcing. Tidal level variations were provided by the hydrodynamic model to the three grids while fields of currents were only provided to the high resolution coastal grid, assuming wave–current interactions are negligible offshore.

For the hydrodynamic model, two unstructured grids using triangular elements were used. The first one has a regional extension, with mesh size ranging from 2 km at its open boundaries to 100 m at the studied beach. The flow in this grid was forced by interpolation of amplitudes and phases of the seven main tidal constituents (O1, K1, N2, M2, S2, K2 and M4) taken from the regional model of Fortunato et al. (2002), in order to simulate tidal propagation from the external continental shelf up to the studied beach and to provide amplitude and phases of water elevation at the open boundary of the second grid. These constituents represent about 95% of the tidal signal at a nearby tidal station (Idier et al., 2006). The second grid has a resolution ranging from 50 m at its open boundaries to 8 m within the surfzone. The southern limit of this grid was placed 1 km north of the Maumusson inlet (Fig. 3), thereby avoiding tidal currents above 0.1 m/s at the boundaries and limiting boundary conditions to water elevation only. The flow in this grid was forced by interpolation of the same harmonic constituents as for the tidal elevation taken from the results of the previous grid. In addition, flow in this second grid was also forced by the gradients of wave radiation stresses computed from the wave model outputs, which were also used to compute the horizontal diffusivity coefficients. While using a single grid for the hydrodynamics would be feasible, it would unnecessarily increase the computational cost. Instead, the nesting of the hydrodynamic grid was adopted here due to the complexity of the surrounding area, with multiple inlets and islands.

Finally, the finer grid was used to compute sand fluxes with SAND2D. A constant sediment grain size of $D_{50} = 0.22$ mm was used, based on the measurements of Bertin et al. (2008a) during the tracer experiments. Sand fluxes were computed using four semi-empirical formulae currently used in engineering and research studies: the Soulsby–Van Rijn formula (Soulsby, 1997, hereafter referred to as SVR), the Bijker formula (1971, hereafter referred to as BIJ), the Bailard and Inman formula (1981, hereafter referred to as BI) and the Ackers–White formula (Ackers and White, 1973) adapted to wave environments by Van de Graaff and Van Overeem (1979, hereafter referred to as AAW). These four formulae were used with standard coefficient values (i.e., as given in the literature, Appendix A) and the bottom friction coefficient $f_w$ required for the BI and the AAW formulae was computed based on the equation proposed by Swart (1974).

3.1.2. Wave simulation

Modeled wave parameters computed at the location of the S4-ADW currentmeter (Fig. 3B) were compared with data (Fig. 4). A first calibration of the bottom friction coefficient $K_n$ in SWAN was done until the modeled significant wave height fitted globally with the data obtained at high tide (i.e., the data measured outside the surf zone). A reasonable fit was achieved for $K_n = 0.05$ m, which is the default value in SWAN when the Madsen friction law (Madsen et al., 1988) is used. However, significant wave heights were slightly underestimated during the third tidal cycle. Nevertheless, this first calibration through the friction coefficient alone was not...
sufficient to reproduce the cross-shore profile of wave heights within the surfzone (Fig. 4A). The default breaking parameter in SWAN \( (c = 0.73) \), which corresponds to the ratio between the maximum possible wave height and the local water depth, appeared to be inadequate for such a low-gradient beach since the significant wave height was systematically over-predicted within the surfzone. This parameter was thus adjusted to a value of \( c = 0.55 \). This value is in borderline with the 0.6–0.83 range of Battjes and Stive (1985), but led to a fair reproduction of the wave height decay within the surfzone (Fig. 4A). This reproduction is very important for morphodynamic simulations since the rate of wave height decay determines the gradient of radiation stress, thereby controlling the magnitude and the cross-shore shape of longshore currents. Hence, the calibration of the \( c \) breaker parameter may be considered with attention for future modeling works of wave-dominated environments, particularly for low-gradient beaches.

Wave peak periods and directions were very well predicted from the first simulations, suggesting the calibrated parameters only have a negligible influence on wave angle and direction. Overall, using the WW3 model as offshore forcing resulted in very good predictions, which shows the great value of WW3 predictions when working in a coastal zone where offshore wave records are unavailable. The good quality of WW3 predictions was already reported in several areas (for instance in the Bay of Biscay by Abadie et al. (2006) and Bertin et al. (2008b) and in the Australian Gold Coast by Browne et al. (2007)).

3.1.3. Hydrodynamics simulation

Modeled longshore currents and water levels were outputted at the locations of the S4-ADW and 2D-ACM currentmeters and compared with data (Fig. 5). A good agreement was observed between measured and modeled water level (RMSE = 0.06 m, Fig. 5A), which validates the regional tidal model of Fortunato et al. (2002) for offshore tidal forcing in this area. Water elevation also integrates a setup due to the shore-normal component of the wave radiation stress gradients. To separate the tidal and the wave-induced components of the elevation, a simulation was performed without waves and the corresponding tidal curve was superimposed on Fig. 5A. This figure shows the ability of the model to reproduce this setup, which is slightly negative backward of the breaking zone (setdown) and reaches up to 0.2 m close to the shoreline. Hence, this good correlation between observed and predicted water depth also constitutes a first validation of the procedure used to couple the wave and the hydrodynamic models.

In terms of longshore currents, the calibration was performed through the adjustment of a space-constant Manning friction coefficient \( n \), until a good correlation was obtained at the location of the 2D-ACM currentmeter, which was also the location of sediment tracer experiment (Fig. 5C). The calibrated value of \( n = 0.015 \) m\(^{-1/3}\) s also leads to a fair prediction of longshore currents in the beach lower part at the location of the S4-ADW currentmeter (Fig. 5B). Although using a simple Manning law to represent bottom friction at beaches was already used successfully by Smith et al. (1993), it could appear simplistic compared to other coastal modeling systems, which integrate more sophisticated bed shear stress equations, such as those of Mei (1989), Liu and Dalrymple (1978) or Le Blond and Tang (1974). Nevertheless, these equations were developed for beaches and their validity in tidally-dominated environments remains to be demonstrated. In particular, the later two equations result in the absence of bed shear stress when the bottom orbital velocity is zero, which would lead to the over-prediction of velocities when applied to estuaries and tidal inlets.

Fig. 4. Measured and simulated wave parameters at St. Trojan Beach from the 4th to the 6th of April 2005: (A) significant wave height, (B) peak period and (C) wave direction. The dotted line in panel (A) corresponds to model results with default parameters and illustrates the necessity to calibrate the wave model.
Some differences remain between longshore current predictions and measurements, particularly at the lower part of the beach (Fig. 5B). Longshore currents were slightly underestimated during the third tidal cycle (Fig. 5B), which could be related to the underestimation of wave height as reported in the previous section (Fig. 4A). Then model/data comparison appears weaker for larger water depths, which can be due to the fact that currents were measured only at 0.4 m from the bottom and do not represent depth-integrated velocities seaward of the breaking zone, where vertical mixing is weak. Finally, the measured longshore current displayed low-frequency fluctuations (0.03–0.003 Hz) that the model was not able to reproduce. This latter variability is due to the low-frequency modulation of incident wave energy, which induces fluctuations in current intensity in the infragravity band. Such fluctuations cannot be represented in our approach since the wave forcing is constant at the time scale of these fluctuations and thus the model only computes "mean currents".

The overall very good agreement between predicted and observed longshore currents validates the procedure developed to force the hydrodynamic model by gradient of wave radiation stresses.

3.1.4. Sand transport predictions

The fine calibration of hydrodynamics at the St. Trojan Beach allowed for a comparison between the model transport predictions and tide-integrated values evaluated from tracer experiments. Sediment transport was computed over the simulated period using the SVR, the BI, the BIJ and the AAW formulae (Appendix A) and the longshore component of instantaneous fluxes was outputted at the location of the 2D-ACM currentmeter, which also coincided with the location of tracer experiments (Fig. 3B). Sand fluxes were derived from tracer experiments based on the movement of the tracer cloud centroid between two successive low tides, which provides a local estimation of the total load (i.e. bedload and suspension) integrated over a tidal cycle.

Comparison between instantaneous fluxes obtained from these four formulae revealed significant differences, with the BIJ and SVR formulae producing 5–10 times larger fluxes than the BI and AAW formulae (Fig. 6A). Instantaneous fluxes were then integrated over a tidal cycle, to be compared to the transport rates evaluated from tracer experiments. The same scatter as for instantaneous fluxes was observed, with the BIJ and SVR formulae producing 5–12 times larger sand fluxes than those deduced from tracer experiments. The BI and the AAW formulae gave values in good agreement with tracer results (Figs. 6B and 4C).

Given that the presented formulae were applied as given in the literature, it can be argued that the observed scattered results are related to the intrinsic properties of each formula. The good predictive skills of the BI and AAW formulae as well as the one order of magnitude overestimations of the BIJ formula are consistent with previous findings of Bayram et al. (2001), who tested these formulae using the field data obtained during Duck Beach experiments (North Carolina, USA). The large errors obtained with the BIJ formula could also be explained by the poor performance of this formula with fine sands (Camenen and Larroudé, 2003).

Although the results obtained for this particular beach may be site-specific and, for instance, result from the presence of fine sand, they highlight the critical problem that represents the choice of a transport formula when simulating morphological evolutions in a real case study and point out the usefulness of tracer experiments to evaluate transport predictions.
3.2. Application to a very dynamic tidal inlet

3.2.1. Model setting

The improved modeling system was tested by applying it to the very dynamic Óbidos Inlet located on the western coast of Portugal (Fig. 7). The combination of a meso-tidal range, a severe wave climate and shallow channels results in high velocities, large sediment fluxes and subsequent very fast morphological changes (Oliveira et al., 2006). Due to these characteristics, this coastal system is difficult to simulate, but provides an ideal testbed to address the numerical stability and the performance of the improved modeling system.

The modeling system was forced first by tides alone, using a constant tide represented by M2 only (producing a 2 m tidal range outside the lagoon and 1.3 m inside, Fig. 8a) and taking advantage of previous calibration work (Oliveira et al., 2006). The purpose of this test was to evaluate the performances of the adaptive morphodynamic time-stepping procedure in terms of computational time and numerical stability. The initial bathymetry for the simulation was based on two surveys. The first was conducted in June 2000 and covered the whole lagoon. The second was measured in July 2001, after the inlet was relocated and the tidal channels dredged, and covered only the lower lagoon, where the major morphological changes occur (Oliveira et al., 2006).
The same grid was used for the hydrodynamic model ELCIRC and for the sand transport model SAND2D and consists of an unstructured triangular grid where the element size ranges from more than 1000 m offshore to 5 m at the inlet channel (Fig. 7). Sand fluxes were computed with the AAW formula and from more than 1000 m offshore to 5 m at the inlet channel of an unstructured triangular grid where the element size ranges el ELCIRC and for the sand transport model SAND2D and consists of a grid bathymetry was kept constant throughout the simulations. In this scope, “morphodynamic time step” refers to the time interval between calls to the hydrodynamic model, rather than to the bathymetry update.

The morphological predictive skill of MORSYS2D was then evaluated by performing a simulation under more realistic conditions. The simulation was performed over five weeks, starting from the July 2001 bathymetry, measured shortly after the relocation of the tidal inlet and the dredging of main tidal channels inside the lagoon. Tidal forcing was obtained by interpolation of the regional tidal model of Fortunato et al. (2002) while the wave forcing was originated from a WW3 simulation point located right offshore the study area (−10°W; 40°N, Fig. 2).

3.2.2. Assessment of the adaptive time step

The first test was performed using a constant morphodynamic time step of 30 min and led to maximum Courant numbers (and thus sand fluxes) that varied over the range 0.06–20 (close to three orders of magnitude) during a single tidal cycle, with minimum values at low and high tide and maximum values about 1.5 h before and after high tide. As compared to the tests described below, the CPU time was good (1200 s to simulate one tidal cycle) but the Courant number varying over the range 0.005–3, thus preventing the appearance of numerical oscillations. The second test was performed using a constant morphodynamic time step of 3 min and led to Courant numbers achieving (18, Fig. 8b) led to the development of numerical oscillations. The second test was performed using a constant morphodynamic time step of 3 min and led to Courant number varying over the range 0.005–3, thus preventing the appearance of numerical oscillations. Unfortunately, the subsequent CPU time was three times larger (3600 s) than the CPU for the 30 min constant time step run (Fig. 8c). The last test was performed using the adaptive time step presented in Section 2.4 targeting a maximum Courant number of 5, which led to Courant numbers ranging from 0.5 to 6 over a tidal cycle (Fig. 8d), guarantying the numerical stability of the simulation. The adaptive time step requires an acceptable CPU time (1500 s to simulate one tidal cycle), only 25% larger than that obtained for the 30 min run. This approach constitutes a good compromise as compared to the use of a large, constant time step, which leads to fast but unstable runs, and the use of a small, constant time step, leading to stable but very time-consuming runs.

The advantages of using an adaptive morphodynamic time step will be even more relevant when performing long-term simulations with real forcings, including the alternation of neap and spring tides, superimposed on both small and large waves. In these conditions, numerical stability with a constant time step could only be ensured by using an even smaller morphodynamic time step, which would result in a large computational cost.

3.2.3. Preliminary validation of waves and water level predictions

The wave forcing at Óbidos was obtained using the methodology described in Section 2.2, which combines three nested SWAN runs, the first of which was forced offshore by time series of wave parameters originating from the NWW3. In order to provide a preliminary validation of this methodology, modeled significant wave height were compared to wave records intermittently recorded by the Portuguese Hydrographic Institute (IH) from November 2000 to November 2002 in front of Óbidos Inlet (Fig. 7) (Oliveira et al., 2006). Fig. 9a displays data/model comparison over 18 days from the 17th of April 2001 and shows that significant wave height $H_s$ are very well reproduced by the model, with a root mean square error of the order of 0.25 m.

The hydrodynamic model ELCIRC was forced offshore by tides interpolated from the regional model of Fortunato et al. (2002) and, on the coastal zone, by the gradients of wave radiation stresses computed based on wave model results. Model predictions of elevations were compared to elevation data available inside the lagoon and obtained from a tide gauge deployed by IH at the Bico do Corvo station (Fig. 7) as described in Oliveira et al. (2006). A model/data comparison was performed over 15 days from the 25th of July 2001 and shows that the model was able to repro-
duce very well the elevations, with root mean square errors of 0.065 m.

3.2.4. Morphodynamic simulation

The initial bathymetry used in this 5 weeks morphodynamic simulation was measured shortly after the relocation of the tidal inlet and the dredging of the North Channel inside the lagoon, which means that no significant ebb-delta was found in front of the inlet at the beginning of the simulation. The predictive skill of the model was thus tested namely by its ability to reproduce a realistic ebb-delta. Forcings during the simulated period included tides ranging from 1.2 to 3 m and short period (\( T_p = 5-10 \) s), low to moderate energy (\( H_s = 1-2 \) m) waves coming from the N to NW and inducing a southwestward longshore transport. Fig. 10 firstly shows that the model was able to reproduce the development of a well-shaped and realistic ebb-delta. The channel depth at the inlet increased from 2.5 m below mean sea-level (MSL) to 4 m below MSL, and the inlet enlarged from 115 m to 140 m and migrated about 15 m southward. The enlargement of tidal channels promoted the tidal propagation within the lagoon, which was illustrated by an increase of M2 amplitude from 0.4 m at the beginning of the simulation to 0.5 m after 5 weeks of simulations. This increase in tidal propagation within the lagoon during fair weather conditions is consistent with the observations of Oliveira et al. (2006), who reported M2 amplitudes exceeding 0.5 m within the lagoon in the end of summer 2001. The southward displacement of the inlet mouth is also consistent with observations, with indicate that the inlet tends to migrate south after relocations (Fortunato and Oliveira, 2007b). Numerical simulations will have to be performed over longer periods, to enable the morphological predictions to be compared to several bathymetric surveys. This work is in progress and will be reported on a separate paper.

4. Summary and conclusions

The MORSYS2D modeling system was firstly improved by integrating the SWAN spectral wave model (Booij et al., 1999) and forced offshore by the WW3 (Tolman et al., 2002) global wave model. The use of WW3 data as offshore forcing was shown to be very valuable, particularly in the case where no wave records are available close to the study area. The comparison with field measurements revealed that the model of wave dissipation by breaking should be calibrated, which matches the conclusions of Apostos et al. (2008). This calibration is of critical importance since the rate of wave energy decay controls the gradients of radiation stresses and thus the magnitude and cross-shore distribution of longshore currents.

Fig. 9. Measured and simulated wave height (a) and surface elevation (b) at Óbidos Inlet, showing the good predictive skills of the modeling system.

Fig. 10. Morphological simulation of Óbidos inlet from the 25th of July to the 30th of August 2001, showing the development of a realistic ebb-delta.
The effect of short waves on water circulation was taken into account by including the gradients of wave radiation stress into the momentum equations of the ELCCIRC hydrodynamic model. The horizontal momentum diffusion due to wave breaking was also implemented in Elcirk, but data/model comparison at the dissipative beach case showed that it only improves slightly the hydrodynamic predictions. A simple Manning friction law was adopted, which presents the advantage of limiting the calculation work and performs well with or without waves. Although our approach could appear simplistic compared to more sophisticated models, it was shown to perform well at a wave-dominated beach. Furthermore, model over-calibration subsequent to sophisticated physical models for friction or momentum diffusion goes against the spirit in which MORSYS2D was developed, which was to build a relatively simple and robust modeling system, but being easily applicable to realistic case studies.

The implementation of an adaptive morphodynamic time step was shown to result in a good balance between numerical stability and computational efficiency. This procedure will be of particular interest when performing long-term simulations with a real forcing (alternation of small and large waves superimposed on spring and neap tides), where the numerical stability with a constant time step could only be ensured by a very small morphodynamic time step.

Application of the model to two contrasting case studies demonstrated its ability to reproduce coastal dynamics. The fine calibration of hydrodynamics at the St. Trojan Beach allowed for a realistic comparison between model transport predictions and tracer experiments. This comparison revealed a one order of magnitude scatter from one formula to another and illustrated the problem that represents the choice of a transport formula when performing coastal morphodynamic simulations. An application to the Óbidos tidal inlet revealed that the model is able to realistically reproduce the generation of an ebb-delta.

As a general conclusion, the MORSYS2D modeling system was improved and is at present capable of simulating waves, tide and wave-induced hydrodynamics, sand transport and morphological evolutions of real and complex coastal systems. Longer term morphodynamic simulations are being performed and predictions will be compared to repetitive bathymetric surveys to validate its predictive skills in terms of morphological evolutions.

**Acknowledgements**

The first author was funded by the European Commission through a Marie Curie postdoctoral fellowship (project IMMATEL-041171). This work was also partially funded by the Fundação para a Ciência e a Tecnologia, Programa Operacional "Ciência, Tecnologia, Inovação" and FEDER, project EMERA: Estudo da Embocadura da Ria de Aveiro. The authors thank the developing teams of the models ELCCIRC and SWAN for making their source codes available, and Instituto Hidrográfico for the Óbidos lagoon data. Finally, the authors appreciated the comments of two anonymous reviewers as well as those of the associate editor Dr. Julie Pietrzak, which greatly improved this manuscript.

**Appendix A. Sediment transport formulae**

**A.1. Bijker formula (1971)**

Bijker derived the Frijlink (1952) formula for bedload by using Einstein integrals for evaluating suspended load and also modified the bottom stress to adapt the formula to coastal environments with waves. The total load $q_{sl}$ is expressed as the sum of a bedload term and a suspended load $q_{st}$ term:

$$q_{sl} = C_b \cdot d_{50} \cdot \frac{U_c}{C} \cdot \sqrt{g} \cdot \exp \left[ -\frac{0.27 \cdot (\rho_s - \rho) \cdot g \cdot d_{50}}{\mu \cdot \tau_{cw}} \right]$$

$$q_{st} = 1.83 \cdot q_{sl} \cdot \left[ I_1 \cdot \ln \left( \frac{33 \cdot h}{\delta_s} \right) + I_2 \right]$$

where $C_b$ is a wave breaking parameter (1.0 for non breaking waves and 5.0 for breaking waves with a ramp function between the two situations), $d_{50}$ is the median sediment diameter, $U_c$ is the current velocity, $C$ is the Chezy coefficient based on $d_{50}$, $g$ is the acceleration of the gravity, $\rho_s$ is the sediment density, $\rho$ is the water density, $\mu$ is a ripple factor, $\tau_{cw}$ is the combined shear stress due to waves and currents, $h$ is the water depth and $I_1$ and $I_2$ are the Einstein integrals computed following Van Rijn (1993).

The ripple factor is expressed as

$$\mu = \left( \frac{C}{C_{so}} \right)^{1.5}$$

where $C_{so}$ is the Chezy coefficient based on $d_{50}$.

The combined shear stress due to waves and currents $\tau_{cw}$ is expressed as

$$\tau_{cw} = \tau_c \left[ 1 + 0.5 \left( \frac{U_c}{Uw} \right)^2 \right]$$

where $\tau_c$ is the bed shear stress due to currents only, $Uw$ is the wave orbital velocity, $\zeta$ is a parameter for wave–current interaction and is expressed as:

$$\zeta = C \sqrt{\frac{f_{w}}{2 - g}}$$

where $f_{w}$ is the wave friction factor computed according to Jonsson (1966).

**A.2. The Bailard and Inman formula (1981)**

Bailard and Inman (1981) extended the energetic model of Bagold (1966) and proposed a formula where the total load is expressed as the sum of a bedload term $q_{sl}$ and the suspended load term $q_{st}$:

$$q_{sl} = \frac{0.5 f_w \cdot \rho \cdot v_{sb}}{(\rho_s - \rho) \cdot g \cdot \tan(\phi)} \left( U_t \right)^2 \cdot \frac{1}{U_t - \tan(\phi) \cdot U_t}$$

$$q_{st} = \frac{0.5 f_w \cdot \rho_s \cdot v_{st}}{(\rho_s - \rho) \cdot g \cdot w_s} \left( U_t \right)^3 \cdot \frac{1}{U_t - \frac{\tan(\phi)}{w_s} \cdot U_t}$$

where $\rho_s$ is the sediment density, $\rho$ is the water density, $v_{sb}$ is 0.1 and $v_{st}$ is 0.02 are efficiency factors for bedload and suspended load, respectively, $g$ is the acceleration of gravity, $\phi$ is the internal friction angle of the sediment, $U_t$ is the instantaneous velocity vector resulting from wave and currents, $\beta$ is the local bottom slope and $w_s$ is the sediment fall velocity based on $d_{50}$ and $f_w$ is a wave friction factor computed using the expression of Swart (1974).

**A.3. The Ackers and White formula (1973) adapted to waves by Van de Graaf and van Overeem (1979)**

Van de Graaf and Van Overeem (1979) adapted the formula of Ackers and White (1973) to take into account the effect of waves. The total load $q_{st}$ is expressed as:

$$q_{st} = U_c \cdot \frac{1}{1 - p} \cdot d_{50} \cdot \left[ \frac{U_{sw}}{U_{sw}} \right]^n \cdot C_{ss} \cdot A^n \cdot \left[ \frac{C_d \cdot U_{sw} \cdot (\frac{U_{sw}}{U_{sw}})^{n+1}}{C_d \cdot g^{n/2} \cdot \sqrt{(\frac{n}{n-1}) \cdot d_{50}}} - A \right]$$

Please cite this article in press as: Bertin, X. et al., Simulating morphodynamics with unstructured grids: Description and validation ..., Ocean Modell. (2008), doi:10.1016/j.ocemod.2008.11.001
where \( p \) is the sediment porosity, \( d_{50} \) the particle diameter exceeded by 65% of the weight, and \( A, n, m \) and \( C_{dgr} \) are dimensionless parameters defined as follows:

\[
\begin{align*}
n &= 1 - 0.2432 \cdot \ln(d_{50}) \\
m &= 9.66 \frac{d_{50}}{C_{dgr}} + 1.34 \\
C_{dgr} &= \exp\left[2.86 \cdot \ln(d_{50}) - 0.4343 \cdot \ln(d_{50})^2 - 8.128\right] \\
A &= 0.23 \sqrt{d_{50}} + 0.14
\end{align*}
\]

In the main equation, \( Uw_c \) and \( U_{cw} \) are the current velocity and shear velocity modified from the original formulation of Ackers and White (1973) to account for waves and currents:

\[
U_{wc} = U_c \cdot \left[1 + 0.5 \left(\frac{\zeta'}{U_c} \right)^2\right]
\]

and

\[
U_{wc} = U_{cw} \cdot \left[1 + 0.5 \left(\frac{\zeta}{U_c} \right)^2\right]
\]

where

\[
\zeta = 18 \cdot \log \left(\frac{12 \cdot h}{r}\right) \cdot \sqrt{\left(\frac{f_w}{r}\right)}
\]

and

\[
\zeta' = 18 \cdot \log \left(\frac{10 \cdot h}{d_{s0}}\right) \cdot \sqrt{\left(\frac{f_w}{2g}\right)}
\]

In these two latter equations, \( h \) is the water depth, \( r \) is the bed roughness, and \( f_w \) and \( f_c \) are the wave friction coefficient computed according to Swart (1974) and using \( r \) and \( d_{s0} \) as bed roughness, respectively.

**A.4. The Soulsby and Van Rijn formula (Soulsby, 1997)**

Soulsby and van Rijn developed a formula to compute sediment transport under the combined action of wave and currents. The total load \( q_{st} \) is expressed as the combined effects of

\[
q_{st} = A_s \cdot U_c \cdot \left[\left(U_c^2 + 0.018 \cdot U_{rms} \cdot U_{sw}\right)^{2.4} \cdot (1 - 1.6 \cdot \tan(\beta))\right]^{1.5}
\]

where \( U_c \) is the current velocity, \( U_{rms} \) is the root mean square square orbital velocity and \( \beta \) is the local bottom slope, \( Cd \) is a drag coefficient, \( U_{sw} \) is the threshold velocity, \( A_s \) is the sum of a term for bed load \( A_{ab} \) and a term for suspended load \( A_{as} \):

\[
A_s = A_{ab} + A_{as}
\]

with

\[
A_{ab} = 0.005 \cdot \frac{h \cdot \left(\frac{d_{50}}{\pi}\right)^{1.2}}{\left(\pi \cdot 0.5 \cdot g \cdot d_{s0}\right)^{1.2}}
\]

and

\[
A_{as} = 0.012 \cdot d_{s0} \cdot D^{0.6} \cdot \left(\frac{d_{50}}{\pi} \cdot g \cdot d_{s0}\right)^{1.2}
\]

In these two latter equations, \( h \) is the water depth, \( d_{s0} \) is the mean sediment grain size, \( g \) is the acceleration of gravity and \( D \) is defined as

\[
D = \left[\frac{g \cdot \left(\frac{d_{50}}{\pi} \cdot g \cdot d_{s0}\right)}{n^2}ight]^{1/3}
\]

where \( n \) is the cinematic viscosity of the water.

The drag coefficient \( Cd \) in the main equation is defined as:

\[
Cd = \left(\frac{0.40}{\ln\left(\frac{h}{d_{s0}}\right)} - 1\right)^{2}
\]

where \( z_0 \) is the bed roughness length.

In the main equation, the threshold current velocity \( U_{tr} \) reads:

\[
U_{tr} = 0.19 \cdot (d_{s0})^{0.1} \cdot \log\left(\frac{4 \cdot h \cdot d_{s0}}{d_{50}}\right)
\]

if \( 0.1 \text{ mm} \leq d_{50} < 0.5 \text{ mm} \).

And

\[
U_{tr} = 8.5 \cdot (d_{s0})^{1.6} \cdot \log\left(\frac{4 \cdot h \cdot d_{s0}}{d_{50}}\right)
\]

if \( 0.5 \text{ mm} \leq d_{50} \leq 2.0 \text{ mm} \).

**References**


