1. Introduction

In this chapter, attention is focused on the hydrodynamics and morphodynamics of salt marshes and tidal flats in shallow tidal lagoons. Shallow tidal basins are often characterized by extensive tidal flats and marshes dissected by an intricate network of channels (Fagherazzi et al., 1999; Rinaldo et al., 1999a,b; Defina, 2000a; Marani et al., 2003b). Both tidal flats and salt marshes are prevalently flat landforms located in the intertidal zone.

Salt marshes have an elevation higher than the mean sea level and are periodically flooded by high tide. They are characterized by a very irregular surface and exhibit a dendritic and meandering structure of channels of varying sizes. These channels perform a drainage function, often continuing to flow long after the tide has receded and the marshes are exposed. They usually sustain a dense vegetation canopy of halophyte plants that withstand the relative infrequent flooding periods. Besides the biological processes related to plant colonization, such as bio-stabilization or soil production, vegetation sensitively affects the local hydrodynamics, reduces the bottom erosion, and increases sediment deposition.
Tidal flats are intertidal landforms usually flat or characterized by smooth sand waves. Occasionally, wide and shallow meandering channel are present. For the sake of simplicity, the term “tidal flat” is herein used in its broadest sense to include the “shallow sub-tidal flats,” that is, the muddy platforms that do not emerge during ordinary low tide.

Salt marsh elevation is controlled by mineral and organogenic sediment accumulation (Pethick, 1981), sea-level variations, stabilizing effects of halophyte vegetation on its platform (Morris et al., 2002; Mudd et al., 2004; Silvestri et al., 2005), and the interaction between flora and fauna (Perillo et al., 2005; Minkoff et al., 2006).

Tidal flats stem from a delicate balance between sediment deposition and erosion by wind waves and tidal currents (Allen and Duffy, 1998). Indeed, biology plays a nonnegligible role on the morphology of tidal flats (for a thorough review see Uncles, 2002). In particular, microphytobenthos and other organisms which colonize shallower areas affect the bottom shear stress threshold for sediment resuspension (Amos et al., 2004), and thus the morphological equilibrium of shallow tidal flats (Marani et al., 2007). However, besides the many studies which demonstrate the important interplay between biology and morphology on salt marshes and tidal flats, reliable mathematical models predicting the biological impact on flow dynamics and sediment processes are still lacking (Dietrich and Perron, 2006).

The distinctive characters of tidal flats and salt marshes reflect on flow, wave field, transport and diffusion processes, and morphologic evolution as well. Therefore, different strategies must be followed when modeling the local hydrodynamics and morphology to maximize the accuracy and minimize the computational effort.

Although considerable progress has been made in the application of two-dimensional (2D) and three-dimensional (3D) models to simulate flow, waves and sediment transport in estuaries and coastal lagoons, a number of outstanding problems still remain in this branch of computational fluid dynamics. These problems mainly stem from the need to accurately model the key physical processes when dealing with very shallow flows, time-dependent flow domains, and complex topography.

Adequate solutions cannot be just given by adopting accurate and well-structured numerical schemes, or else extremely refined computational grids. An important effort should instead be addressed toward modeling the relevant physical phenomena, which are neglected or drastically filtered by the numerical solution. This can be accomplished through the construction of suitable subgrid models, that is, by setting up a phenomenological representation of the overall processes which ensures a statistically equivalent description of the actual physics.

Among others, wetting and drying of salt marshes and tidal flats, and the hydrodynamics of the small-scale drainage networks dissecting salt marshes are discussed in Section 2. Section 3 presents a simplified, computationally efficient wind wave model: the model generates and propagates a monochromatic wave inside tidal lagoons by solving the wave action conservation equation on an unstructured triangular mesh of arbitrary shape with a first-order
finite volume explicit scheme. The problem of evaluating the bottom shear stress distribution due to the combined action of tidal currents and wind waves is also addressed in this section. Impact of salt marsh vegetation on tidal currents and wind waves is shortly discussed in Section 4. Morphodynamics modeling of salt marshes and tidal flats is discussed in Section 5, distinguishing between long- and short-term approaches. Finally, the main conclusions are summarized in Section 6.

The examples presented in this chapter use the Venice Lagoon as typical irregular and shallow tidal basin. The Venice Lagoon is a wide tidal basin crossed by a network of deep channels departing from three inlets, namely Lido, Malamocco, and Chioggia (Figure 1). The lagoon is also characterized by the presence of wide tidal flats, small islands and salt marshes that exhibit a dendritic structure of channels of varying sizes (Rinaldo et al., 1999a, b; Defina, 2000a; Marani et al., 2003b).

2. WETTING AND DRYING, AND THE DYNAMICS OF VERY SHALLOW FLOWS

The wetting and drying problem has received considerable attention during the last two decades. Recent reviews, mainly concerned with numerical aspects of this problem, can be found in the works of Balzano (1998), Bates and Hervouet (1999), and Bates and Horritt (2005). The wetting and drying problem can be handled either by adapting the numerical grid at each time step to follow the deforming flow domain (Lynch and Gray, 1980; Kawahara and Umetsu, 1986; Akanbi and Katopodes, 1988) or by retaining a fixed computational grid and
utilizing some additional algorithms to deal with the hydrodynamics of partially wet elements. Due to the great difficulty of developing efficient deformable grid techniques, a fixed grid approach is by far preferable. In this case, a whole range of algorithms is available to identify wet elements and to control the flow over these (King and Roig, 1988; Leclerc et al., 1990; Falconer and Chen, 1991; Bates et al., 1992; Braschi et al., 1994; Defina et al., 1994; Hervouet and Janin, 1994; Defina and Zovatto, 1995; Ji et al., 2001; Oey, 2005). These algorithms are often intimately related to a particular numerical scheme and their application to a different numerical model is not straightforward (Balzano, 1998).

Moreover, when dealing with very small water depths and wetting/drying of large areas, the major source of inaccuracy comes from the fact that numerical models approximate the bottom with a piecewise homogeneous plane surface. In this way they do not properly account for the effects due to the local variations of the flow field produced by small-scale topography (Defina et al., 1994; Bates and Hervouet, 1999; Defina, 2000a), thus yielding to approximate distributions of velocity and depth. The above problems can be partially overcome by setting up a phenomenological representation of the overall processes to supply a statistically equivalent description of the physics. To deal with partially wet and very irregular domains, an effective subgrid model of ground topography was developed by the authors in the early 1990s and improved over the years (Defina et al., 1994; Defina and Zovatto, 1995; Defina, 2000a). On considering bottom irregularities from a statistical point of view, and assuming the hydrostatic approximation, the 3D Reynolds equations have been phase averaged over a representative elementary area (REA) and then integrated over the depth. The averaged equations read (Defina, 2000a):

\[ \nabla h + \frac{1}{g} \frac{d}{dt} \left( \frac{q}{Y} \right) + \mathbf{J} - \nabla \cdot \mathbf{R}_e = 0 \]  

\[ \eta(h) \frac{\partial h}{\partial t} + \nabla \cdot \mathbf{q} = 0 \]

where \( h \) is the free surface elevation, \( g \) is the gravity, \( t \) is the time, \( \mathbf{q} = (q_x, q_y) \) is the flow rate per unit width, \( Y \) the equivalent water depth, defined as the volume of water per unit area actually ponding the bottom, \( \eta \) the local fraction of wetted domain and accounts for the actual area that can be wetted or dried during the tidal cycle, \( \mathbf{R}_e \) accounts for the horizontal turbulent stresses, and \( \mathbf{J} = (J_x, J_y) \) is energy dissipation per unit length due to bottom shear stress and vegetation, and energy gain due to wind shear stress acting on the free surface.

\[ J = \frac{\tau_b + \tau_v - \tau_w}{g \rho Y} \]

where \( \tau_b \) is bottom shear stress, \( \tau_v \) is an equivalent shear stress accounting for vegetation resistance, \( \tau_w \) is wind shear stress (see Section 3), and \( \rho \) is fluid density.
The bottom topography within a REA is assumed to be irregular with bottom elevations distributed according to a Gaussian probability density function. In this case the functions $\eta$ and $Y$ are found to be (Defina, 2000a):

$$\eta = \frac{1}{2} \left\{ 1 + \text{erf} \left[ \frac{2D}{a_r} \right] \right\}$$

$$Y = a_r \left\{ \eta \left( \frac{D}{a_r} \right) + \frac{1}{4\sqrt{\pi}} \exp \left[ -4 \left( \frac{D}{a_r} \right)^2 \right] \right\}$$

where $\text{erf}(\cdot)$ is the error function, $a_r$ is the typical height of bottom irregularities (i.e., the amplitude of bottom irregularities or, approximately twice the standard deviation of bottom elevations), and $D = h - z_b$ is the average water depth, $z_b$ being the average bottom elevation within a REA (Figure 2).

For the case of turbulent flow over a rough wall, the energy dissipations due to bed shear stress can be written as (Defina, 2000a):

$$\frac{\tau_b}{g \rho Y} = \left( \frac{n^2 |q|}{H^{10/3}} \right) q$$

where $n$ is the Manning bed roughness coefficient and $H$ is an equivalent water depth which can be approximated with the following interpolation formula (Defina, 2000a):

$$\frac{H}{a_r} \approx \frac{Y}{a_r} + 0.27 \sqrt{\frac{Y}{a_r}} e^{-2Y/a_r}$$

The resulting subgrid model for ground irregularities requires the statistics of small-scale bottom topography. At present, remote sensing of topography (i.e., airborne laser altimetry, global positioning system-linked side-scan sonar and wide swath bathymetry) is proving very effective in providing high-resolution terrain data capable of parameterizing the proposed approach, even at subgrid scale (Bates and Hervouet, 1999; Bates et al., 2003, 2005).

The above model proved very effective in the simulation of tide propagation in shallow lagoons, over salt marshes and tidal flats. Examples can be found in the literature (D’Alpaos et al., 1994; Defina and Zovatto, 1995; Defina, 2000a; Lanzoni and Seminara, 2002). All these examples clearly demonstrated the efficiency of the proposed equations. However a number of open issues need to be addressed: (1) the model does not account for water which may remain trapped within the REA during the drying phase. On the contrary, experimental evidence suggests that sometimes small pools or pans remain after the tidal wave recedes; (2) the model for bed shear stress (Equation (6)) neglects momentum exchange due to convective acceleration at the subgrid scale; and (3) Equation (6) for bed shear stress was derived on the assumption of an isotropic distribution of bottom irregularities. This is not always the case, often the creeks dissecting the marshes drive tidal
flow along preferential directions (Figure 3). The two latter issues are shortly discussed in the following paragraphs.

The overall effects of momentum exchange due to convective acceleration at the subgrid scale can be accounted for by adjusting the friction coefficient. In fact, unresolved accelerations (i.e., subgrid accelerations) averaged over a sufficiently large area mostly produce extra dissipation, which is usually accounted for by suitably increasing the Manning friction coefficient. This is done during the model calibration step, since, at present, no relationships are available relating the friction coefficient to filtered velocity distributions. It would clearly be useful to have relationships allowing the a priori estimation of the appropriate roughness corrections. Finding suitable solutions to this problem is a quite demanding task.
due to the large variety of mechanisms producing a spatially heterogeneous velocity field. Among the many, bottom topography in the presence of small water depths is possibly the easiest to handle. In this case, bottom topography generates small-scale momentum mixing thus enhancing energy dissipation. The problem can be handled in a way similar to the “mixing length” approach in turbulence and a relationship relating the ratio of the equivalent ($n_{eq}$) to the actual ($n$) Manning coefficient to water depth and bottom topography can be established (Defina, 2000b; D’Alpaos and Defina, 2007)

$$\frac{n_{eq}}{n} = \frac{36(Y/a_t)^2 + 5\eta^3}{36(Y/a_t)^2 - \eta^3 + 6C_{bw}(r)\eta^3}$$  (8)

where $C_{bw}(r)$ is an autocorrelation function of bottom elevations, given as

$$C_{bw}(r) = \frac{\int z(x)z(x+r)\,dA - z_{wb}^2}{\sigma_{wb}^2}$$  (9)

where $z_{wb}$ and $\sigma_{bw}$ are the average and the root mean square of bottom elevations within the wetted part of the REA, $r$ is a horizontal “mixing length,” and $z$ is local bottom elevation.

Equation (8) for $C_{bw}(r) = 0$ and $C_{bw}(r) = 1$ is plotted in Figure 4. The behavior of $n_{eq}/n$ is not symmetric about $D/a_t = 0$. When $D/a_t \ll 0$, that is, when the REA is nearly dry (Figure 2b), the flow field is characterized by a braiding pattern with the flow in each branch being independent from the others. In this case, momentum mixing is negligibly small and equivalent Manning coefficient recovers its original value (i.e., $n_{eq}/n \approx 1$). When $D/a_t \gg 1$, that is, when the water depth is nearly uniform within the REA (Figure 2a), bottom irregularities have a minor impact on the velocity field and $n_{eq} \approx n$. Note that when the bottom is smooth then $C_{bw}(r) = 1$ and $n_{eq} = n$. 

Figure 3 Sample of creeks aligned in the E–W direction, dissecting a salt marsh of the Venice Lagoon (site 1 in Figure 1). The gray-scale aerial photograph together with the extracted creek patterns are shown. Adapted from Marani et al. (2003b).
The results of two numerical experiments (Defina, 2000b; D’Alpaos and Defina, 2007) are also plotted in Figure 4 for the sake of comparison. The numerical points lay within the region described by the two theoretical curves and the behavior of the numerical solution is in quite good agreement with the theoretical curve for $C_{bw}(r) = 0$. Importantly, the numerical solution approaches this limit curve as the Manning friction coefficient decreases. This is an expected result since increasing the Manning coefficient reduces friction and enhances large scale momentum mixing. As a consequence, the mixing length, $r$, increases and $C_{bw}(r)$ decreases toward zero.

Equation (8) must be considered as a first, promising attempt at quantifying the effects produced by small scale (i.e., subgrid) momentum mixing triggered by bottom irregularities. Indeed, further research is required to evaluate the mixing length $r$ to be used in Equation (9).

The second issue here shortly discussed focuses on the problem of modeling the creeks dissecting the marshes. The small-scale drainage network, comprising channels having a very small width cannot be resolved by a 2D model when the flow domain is comparably large as this would require a large number of very small computational elements. An example is shown in Figure 5 where a refined grid of the Venice Lagoon, comprising nearly $4 \times 10^4$ elements (Carniello et al., 2005), overlaps an aerial photograph of a marsh zone in the northern part of the lagoon. A tangle of small, highly meandering creeks with a width in range between 0.1 and 2 m covers most of the marsh surface. The mesh resolution required to describe all these channels is beyond present computational capabilities. In this case, and as a first approximation, the smaller channels can be treated as “topographic irregularities.” A clear dividing line between actual ground irregularities, which are expected to behave quasi-isotropically, and small-scale channels and creeks, cannot be traced, as it depends on the domain extension and on the required accuracy.

Here the attention is focused on channels with a size which is small enough to prevent the use of 2D elements for their description, and large enough to dissuade
one from including them as bottom irregularities. These narrow channels are usually very numerous and their importance turns out to be comparable to that of large channels (Defina, 2004). The problem of accounting for this channel network can be tackled by observing that the flow within relatively deep and narrow channels flanked by shallow intertidal areas exhibits a distinctive one-dimensional (1D) character, thus suggesting the use of 1D elements to include them in the model.

The problem of coupling 1D and 2D elements has sometimes been found in the literature, for example SOBEK, developed by WL|Delft Hydraulics (http://www.sobek.nl) and LISFLOOD-FP (Bates and De Roo, 2000; Horritt and Bates, 2001a,b; Bates et al., 2005). To keep a high accuracy and reduce the computational effort, a particular way of coupling a 2D model to describe the shallow water hydrodynamics with a 1D model to simulate the flow in the channels has been proposed by D’Alpaos and Defina (1993, 1995, 2007). In the model,
channels are superimposed on the 2D domain. The effects due to the momentum exchange between the channel and the two-dimensional flow are neglected. In the model, the 2D and 1D flow equations are solved by a finite element scheme. The domain is divided into triangular and linear elements with each channel lying along the common side of two adjacent triangular elements. In this way each channel can be added to or removed from the domain without any change in the main 2D discretization. The number of nodes in the computational grid remains unchanged and the computational effort is only slightly increased due to inclusion of 1D elements. In the example shown in Figure 5, Canale S. Felice, which is a very large channel, is described using 2D triangular elements, the very small creeks dissecting the tidal marsh are included in the model as bottom irregularities, while the large creek through the salt marsh, departing from Canale S. Felice, is described with 1D elements (dotted segments) aligned along the edges of the 2D triangular elements. The model has been tested against the numerical solution computed with a 2D model for a number of test cases (D’Alpaos and Defina, 1993, 1995, 2007; D’Alpaos et al., 1995) and proved very effective.

3. Wind and Wind Waves

Surface gravity waves are one of the most important phenomena in shallow, coastal lagoons and estuaries. Estimation of wave characteristics in estuaries, tidal basins and coastal areas is essential to analyze sediment transport and local shoreline erosion processes (Anderson, 1972; Ward et al., 1984; Shoelhamer, 1995; Möller et al., 1999; Umgiesser et al., 2004; Carniello et al., 2005). Swell waves approaching the coast from the open sea are relevant to study the shoreline morphodynamic in coastal areas. On the contrary, the locally generated wave field is of importance for lagoonal morphodynamics: on the one hand, wind wave-induced bottom shear stress is the decisive process mobilizing tidal flat sediments (Carniello et al., 2005) and influencing their equilibrium configuration (Fagherazzi et al., 2006; see also Section 5); on the other hand salt marsh, because of their elevation and the presence of halophytic vegetation, greatly affect wind wave field by reducing wave energy (see Section 4).

Two alternative methods are available to model wind wave generation and propagation, that is, a phase-resolving approach, based on mass and momentum balance equations (for a review see Dingemans, 1997); or a phase-averaged approach that solves the energy or wave action balance equation (e.g., Booij et al., 1999).

Phase-resolving models reproduce the sea surface in space and time and account for effects such as refraction and diffraction. Bottom friction and depth-induced wave breaking can be included in the model but wind wave generation is usually absent or poorly reproduced. Phase-resolving models are thus unsuitable in lagoons where storm conditions and local wave generation are key processes. Furthermore, space and time resolutions required by phase-resolving models are of the order of
a fraction of the wavelength and wave period, respectively, thus restricting their use to small domains and short duration events. For large-scale applications phase-averaged models are by far more suitable. Since the pioneering work of Gelci et al. (1956), many models that use the phase-averaged approach have been developed. Among them: the GLERL model developed by Donelan (1977) and revised by Schwab et al. (1984); the HISWA model (Hindcast Shallow water WAVes model) (see Holthuijsen et al., 1989), and its successor the SWAN model (Simulating WAVe Nearshore) (Booij et al., 1999; Ris et al., 1999); the WAVAD model (Resio, 1987; Resio and Pierre, 1989), and the ACES model (Automated Coastal Engineering System) (Leenknecht et al., 1992). Lin et al. (1998) tested all the models mentioned above against a wind and wave data set collected in the northern Chesapeake Bay, USA, during September 1992, when the tropical storm Danielle passed over the area. They found that no single model seems to be good at predicting all aspects of the surface wave field in that specific and morphologically irregular domain, but the GLERL and SWAN models were the most promising. Moreover, in shallow basins, the instantaneous local water depth is crucial to correctly predict the wave field, since water depths strongly affects wave propagation. Wave prediction can therefore be accomplished only by coupling a wave model with a hydrodynamic model. Umgiesser et al. (2004) moved a preliminary step toward this direction. They combined a 2D finite elements model with the finite difference SWAN model run in stationary mode. For consistency, all the results produced by the hydrodynamic model were interpolated to the grid of the wave model, thus introducing significant numerical approximations. Since shallow tidal basins have a very irregular morphology with large and sudden changes in bottom elevation, islands, and salt marshes which are periodically flooded and exposed, a specific framework must be adopted to model wind wave propagation in these environments.

A simplified, computationally efficient model has been recently developed (Carniello et al., 2005). The wind wave module solves for the conservation of the wave action (Hasselmann et al., 1973), defined as the ratio of the wave energy density $E$ to the wave frequency $\sigma$, using a first-order finite volume explicit scheme. The wave model is coupled with a finite element hydrodynamic model (D'Alpaos and Defina, 1995, 2007) sharing the same computational grid. At each time step, the hydrodynamic model yields nodal water levels which are used by the wind wave model to assess wave group celerity and wave energy dissipations.

The wind wave model propagates a monochromatic wave, neglects nonlinear wave–wave and wave–current interactions and assumes that the direction of wave propagation instantaneously adjusts to the wind direction (Carniello et al., 2005). The wave action conservation equation can thus be written as

$$\frac{\partial E}{\partial t} + \nabla \cdot (c_w E) = S_w - S_{bf} - S_{wc} - S_{brk}$$ (10)

The first term of Equation (10) is the local rate of change of wave energy density in time, the second term represents the energy convection, $c_w = (c_{gx}, c_{gy})$ being the
wave group velocity. The source terms on the right-hand side of Equation (10) account for the wind energy input ($S_w$), the energy dissipation by bottom friction ($S_b$) and by whitecapping ($S_{wc}$), and the energy dissipation by depth-induced breaking ($S_{bd}$). The significant wave height, $H$, is then computed using the linear theory as

$$H = \sqrt{\frac{8E}{\rho g}}$$ (11)

Examples showing the impact of the wind action on the hydrodynamics and the generated wave field inside the Venice Lagoon under different wind and tidal conditions are briefly discussed. All simulations presented here are performed using a refined mesh reproducing the present topography of the Venice Lagoon (Carniello et al., 2005). Numerical results are compared with field measurements available at two different stations inside the lagoon: station 1BF is on a shoal in the northern part of the lagoon and station 2BF is in a deeper area in the southern part of the lagoon (Fondo dei Sette Morti) (Figure 1).

Two stormy events (i.e., 3 April 2003 – see Figure 6a, and 16–17 February 2003) characterized by the Bora wind blowing from the northeast with a speed in the range between 12 and 16 m/s are simulated. The Bora wind affects near-coast sea levels resulting in lower elevations at the Lido inlet than at the Chioggia inlet (Figure 6b). Importantly, wind setup strongly affects the hydrodynamics within the lagoon. This is shown in Figure 6e,f where the measured water levels at 1BF and 2BF (see Figure 1) are compared with the water levels computed with the hydrodynamic model when wind shear stresses are included in the model or neglected. Neglecting wind shear stress produces water levels that do not fit the measured ones.

The impact of wind action on the hydrodynamics is even more evident when the flow rate through the three inlets is considered. When including wind shear stress, the computed flow rate through the Lido inlet increases during flood and a decreases during ebb (Figure 6c). The opposite occurs at the Chioggia Inlet (Figure 6d), while no substantial changes occur at the Malamocco inlet. Overall, wind setup during Bora stormy conditions produces a residual current flowing from the Lido inlet toward the Chioggia inlet.

Figure 7 compares the computed significant wave height with the measured one at stations 1BF and 2BF. The agreement for both storm events is quite good. Plotted wave heights follow a sinusoidal-like variation having the same phase as the tidal oscillation, confirming the influence of water level on wave height and the strong feedback between wind waves and hydrodynamics. The result is even clearer in the lower panel of Figure 7 where an example of the computed wave field at low tide and high tide during the stormy event of 16–17 February 2003 is given.

The complementary effect of tidal currents and wind waves on bottom shear stresses is crucial to predict the morphodynamic evolution of tidal flats in shallow
Figure 6  Stormy event of 3 April 2003. Wind velocity and direction (a); tidal level at the three inlets (b); computed flow rate through the Lido inlet (c) and the Chioggia inlet (d) when wind action is included in the model or neglected; and comparison between computed and measured water levels at stations 1BF (e) and 2BF (f). Adapted from Carniello et al. (2005).
tidal basins (see Section 5). Bottom shear stress due to waves ($\tau_{b,\text{wave}}$) is computed by the wind wave-tidal model as:

$$\tau_{b,\text{wave}} = \frac{1}{2} f_{\infty} \rho u_m^2$$

with $u_m = \frac{\pi H}{T \sinh(kY)}$ and $f_{\infty} = 1.39 \left[ \frac{u_m T}{2\pi(D_{50}/12)} \right]^{-0.52}$ (12)
where \( u_m \) is the maximum horizontal orbital velocity at the bottom according to the linear theory, \( f_r \) is the wave friction factor as given by Soulsby (1997), \( T \) is the wave period, \( k \) is the wave number, and \( D_{50} \) is the median grain diameter.

Since maximum shear stress \( \tau_{\text{max}} \), rather than average stress \( \tau_m \), is responsible for the bottom sediments mobilization, all the results presented and discussed herein after refer to the maximum total bottom shear stress, which is evaluated with the empirical formulation suggested by Soulsby (1997):

\[
\tau_{\text{max}} = \left[ (\tau_m + \tau_{b,\text{wave}} \cos \phi)^2 + (\tau_{b,\text{wave}} \sin \phi)^2 \right]^{1/2}
\]

(13)

\[
\tau_m = \tau_b \left[ 1 + 1.2 \left( \frac{\tau_{b,\text{wave}}}{\tau_b + \tau_{b,\text{wave}}} \right)^{3.2} \right]
\]

(14)

where \( \tau_b \) is given by Equation (6), and \( \phi \) is the angle between the current and the wave directions.

Figure 8 shows the time evolution of the bottom shear stress at three different sites within the lagoon during the stormy event of 16–17 February 2003. The sites are chosen in a deep channel close to the Lido inlet (site 1H in Figure 1), on a tidal flat close to the Murano island (site 3H in Figure 1), and on a tidal flat next to the Casse di Colmata (site 2H in Figure 1). Each plot compares model results obtained with three different simulations, that is, (1) the hydrodynamics is forced by the recorded tidal levels at the three inlets, wind shear stress and wind waves are neglected; (2) the wind shear stress are included whereas wind waves are not; and (3) both wind shear stress and wind waves are included in the model.

The results show that in deep channels wind waves slightly affect bottom shear stresses (Figure 8a), while no influence of wind stresses at the surface can be observed. On the contrary, bottom shear stresses on tidal flats are strongly enhanced when wind waves are included in the model (Figure 8b,c). In this case, wind shear stress gives a minor contribution, feebly enhancing the bottom shear stresses produced by tidal currents.

In shallow areas the bottom shear stresses exceed the critical value \( \tau_{\text{cr}} \approx 0.7 \) Pa, Amos et al., 2004) for sediment erosion only in the presence of waves. On the contrary, the bottom shear stresses are always smaller than the critical value when wind waves are not included in the model. This result is further supported by Figure 8d,e mapping the regions where the bottom shear stress exceeds \( \tau_{\text{cr}} \). No resuspension is possible on tidal flats and salt marshes if wind waves are not considered.

Figure 8d,e also confirms that wind wave resuspension is complementary to tidal current resuspension since waves are able to produce high bed shear stresses in shallower areas whereas bed shear stresses due to tidal currents are high only for the deep channels where the tidal flow concentrates.
Figure 8  Storm event of 16–17 February 2003: comparison of the shear stress at the bottom produced by the combined effect of wind waves and tidal currents (----), by tidal currents when the wind shear stress at the free surface is included (.....), or neglected (-----). The comparison refers, respectively, to the Lido inlet (site 1H in Figure 1) (a), a tidal flat close to Murano Island (site 3H in Figure 1) (b), and a tidal flat close to “Casse di Colmata” (site 2H in Figure 1) (c). Spatial distribution of the area experiencing a bottom shear stress greater than 0.7 Pa inside the Venice Lagoon. The simultaneous effect of tidal currents and wind waves (d) is compared to the effect of tidal currents alone (e). The dotted line (wind – no waves) in panel as is hidden by the solid line (wind – waves). Adapted from Carniello et al. (2005).
4. SALT MARSH VEGETATION

Tidal marshes are colonized by halophytic vegetation, that is, macrophytes adapted to complete their life cycle in salty environments. Vegetation has a strong impact on the hydrodynamics over salt marshes as it affects both tidal current (Burke and Stolzenbach, 1983; Kadlec, 1990; Leonard and Luther, 1995; Shi et al., 1995; Dunn et al., 1996; Nepf and Vivoni, 1999, 2000; Neumeier and Amos, 2006; Neumeier, 2007) and wind waves (Wayne, 1976; Knutson et al., 1982; Pethick, 1992; Koch and Gust, 1999; Möller et al., 1999; Möller and Spencer, 2002; Swales et al., 2004; Möller, 2006). Moreover, vegetation reduces bed shear stress, hence erosion, and strongly affects transport and diffusion processes (Lopez and Garcia, 1998; Nepf, 1999; Nepf and Koch, 1999; Leonard and Reed, 2002; Bouma et al., 2007).

Resistance to flow produced by vegetation can be included in the hydrodynamic model as an additional, equivalent shear stress \( \tau_v \) [see Equation (3)]. To compute the equivalent shear stress \( \tau_v \) any model (Shimizu and Tsujimoto, 1994; Klopstra et al., 1997; Lopez and Garcia, 2001; Righetti and Armanini, 2002; Defina and Bixio, 2005) able to predict the velocity profile in a uniform flow in the presence of vegetation can be used. Here we focus on the case of rigid vegetation and consider a uniform flow in the \( x \)-direction. In this case the velocity profile \( u_x(z) \), \( z \) being the vertical direction, can be written as

\[
    u_x(z) = f(C_D, m, A_z, h_p, Y) \sqrt{S_{0x}}
\]  

where \( h_p \) is plant height, \( C_D \) the drag coefficient, \( A_z \) the frontal area of vegetation per unit depth, \( m \) the number of stems per unit area, \( S_{0x} \) the bottom slope, and \( f \) a function of water depth and vegetation characteristics. The flow rate per unit width is then given as

\[
    q_x = \sqrt{S_{0x}} \int_{z_b}^{h} f(C_D, m, A_z, h_p, Y) \, dz = \sqrt{S_{0x}} F(C_D, m, A_z, h_p, Y)
\]  

Recalling that \( \tau_x = g \rho Y S_{0x} \), extension to 2D flow gives

\[
    \tau_v = \left( \frac{g \rho Y}{F^2} \right) |q| q
\]

Once the velocity profile \( u_x(z) \) is computed for a given slope \( S_{0x} \), function \( F \) can be easily computed from Equation (15).

\[
    F(C_D, m, A_z, h_p, Y) = \int_{z_b}^{h} \frac{u_x(z)}{\sqrt{S_{0x}}} \, dz
\]

Figure 9 shows the behavior of the function \( F \) for three vegetation species which colonize the salt marshes of the Venice Lagoon. In this case the velocity profiles
under uniform flow condition have been computed using the model proposed by Defina and Bixio (2005).

While the above results illustrate the effects of vegetation in producing additional flow resistance, the link between vegetation parameters and wave transformation remains qualitative in the absence of well experimented quantitative relationships between vegetation structure and wave energy.

Wave–vegetation interaction has been investigated to predict wave attenuation produced by vegetation. Standard approaches to model wave attenuation by vegetation are based on the time-averaged conservation equation of wave energy and assume linear wave theory or linearized momentum equations to describe the local flow field (Dalrymple et al., 1984; Kobayashi et al., 1993). For the case of small amplitude monochromatic waves impacting an array of rigid vertical cylinders of diameter $A_z$, Kobayashi et al. (1993) proposed the following dissipation term to be added to the source term of Equation (10).

$$S_{\text{veg}} = - \frac{4\sqrt{2g^3/\rho}}{3\pi} C_{D m A_z} \left( \frac{k}{\omega} \right)^3 \frac{\sinh^3(k h_p) + 3 \sinh (kh_p)}{3k \cosh^3(kY)} E^{3/2}$$

(19)

The above equation can be extended to describe conditions of emergent vegetation by substituting $h_p$ with $Y$. Wave number $k$ in Equation (19) depends not only on water depth and wave period but on vegetation characteristics as well. However, in the limit of small wave energy damping, the standard dispersion equation based on linear wave theory can be used to compute $k$ (Kobayashi et al., 1993).

The above model for wave dissipation due to vegetation suffers a number of shortcomings: (1) in the model, the flow depth is split into a lower layer containing the vegetation and a surface layer; velocity profiles are described separately for the vegetation layer and the surface layer, reflecting the different physical phenomena acting in the two layers; however, they do not match at the interface; (2) in the model turbulent shear stresses are neglected; (3) the model is not sufficiently tested (model results were compared only with experiments conducted by Asano et al.
(1988) who used a very flexible artificial vegetation); and (4) Equation (16) does not account for the variation in both $C_T$ and $A_z$ along $z$. It is thus clear that the above model is deserving of further research.

Once the resistance to flow and the wave energy dissipation due to vegetation have been modeled, the problem of assessing the spatial distribution of vegetation must be addressed. Halophytic vegetation over salt marshes is not randomly distributed nor spatially uncorrelated but is, on the contrary, organized in characteristic patches (Pignatti, 1966; Chapman, 1976; Silvestri et al., 2000, 2005; Marani et al., 2004). Therefore wide areas, which may extend over a few computational elements, are colonized by the same species, that is, a set of parameters describing a single vegetation species can be associated to each computational element. Quantitative remote sensing, integrated with field observations, proved very effective for mapping the different species (Marani et al., 2003a, 2006; Silvestri and Marani, 2004; Belluco et al., 2006) thus allowing for a very accurate parameterization of vegetation in the numerical model.

5. SALT MARSHES AND TIDAL FLATS MORPHODYNAMICS

Different approaches are usually adopted to model short- and long-term morphologic evolution of coastal and lagoonal environments. Long-term models were first introduced to investigate salt marsh formation and evolution. In the pioneering point model suggested by Krone (1987), changes in marsh elevation are calculated as a function of sediment concentration, settling velocity of the suspended sediment flocs, and hydroperiod. When the marsh platform becomes emergent the inundation period decreases, so that less sediment has time to deposit leading to a reduction of marsh accretion. The model was then improved by considering sediment supply and sea-level rise (Allen, 1990; French, 1993), sediment composition (Allen, 1995), differences in sedimentation rates between creek levees and marsh platform (Temmerman et al., 2004a), and variations in sediment concentration as a function of tidal inundation (Temmerman et al., 2004b). In recent years a major development has been the inclusion in the marsh model of the vegetation effects on sediment dynamics, accumulation rates, and organogenic production by linking all these processes to the biomass of halophyte vegetation that colonizes the marsh surface (Morris et al., 2002; Mudd et al., 2004; D’Alpaos et al., 2006).

Besides the vertical accretion of the marsh platform, the latest long-term analysis of salt marsh morphologic evolution consider the formation and the planimetric development of the tidal creek network. A model describing this complex morphodynamic process taking also into account the importance of vegetation distribution, and the consequence of marine transgression and regression, is presented and discussed in D’Alpaos et al. (2009).

However, these models disregard the incipient formation of salt marshes. As a consequence, they can only be applied to locations in which the salt marsh is already present, but are ineffective in determining under what conditions the salt
marsh has evolved from tidal flats. On the contrary, the evolution of tidal flats is less considered in the literature. An original contribution to tidal flat morphodynamics is given by Fagherazzi et al. (2006) and is based on the observation that the bathymetric data points out an abrupt transition between salt marshes and tidal flats with very few areas lying at intermediate elevations. To describe this evidence, Fagherazzi et al. (2006) developed a conceptual model which indicates that this bimodal distribution of elevations strictly relates to wind wave shear stresses. It has been demonstrated (Carniello et al., 2005), in fact, that the role of sediment resuspension by wind waves is decisive in shallow tidal basins, whereas tidal fluxes alone are unable to produce the bottom shear stresses necessary to mobilize tidal flat sediments. The conceptual model follows from the wave model described in Section 3. It mainly assumes (1) that wind waves are the main source of bottom shear stress (i.e., the model does not apply to tidal channels where bottom shear stress is mainly due to tidal current), (2) that in shallow basins waves quickly adapt to external forcing, and (3) the fetch required to attain fully developed condition is short. Therefore, as a first approximation, the conservation Equation (10) can be reduced to the local equilibrium between the source terms describing the unlimited fetch fully developed local wave field.

The conceptual model is based on the stability curve obtained by plotting the wind wave–induced bottom shear stress as a function of water depth (Figure 10a). The model assumes that the rate of sediment erosion, \( E_S \), is proportional to the difference between bottom shear stress \( (\tau_b) \) and the critical shear stress for sediment erosion \( (\tau_{eq}) \). Therefore, the curve of Figure 10a is a proxy for bed erosion rate. The model further assumes some prescribed average annual sedimentation rate, \( D_S \). Dynamic equilibrium, \( E_S = D_S \) is achieved when \( \tau_b = \tau_{eq} \) (points U and S in Figure 10a). When \( \tau_b < \tau_{eq} \) then deposition exceeds erosion and the bottom evolves toward higher elevations. On the contrary, when \( \tau_b > \tau_{eq} \), then erosion exceeds deposition and the bottom sinks toward lower elevations. Therefore, any point S

![Figure 10](image)

(a) Bed shear stress distribution as a function of bottom elevation; (b) frequency area distributions as a function of bottom elevation for the Southern Venice lagoon (1901 and 2000 bathymetries).
along the right branch of the curve is a stable point while any point U on the left branch of the curve is an unstable point. The conceptual model demonstrates that a stable morphodynamic equilibrium is possible only for salt marshes (i.e., \( Z_h > Z_c \)) and tidal flats (i.e., \( Z_{c2} < Z_h < Z_{max} \)).

Recently we tested the conceptual model through comparison with numerical results obtained with the 2D wind waves—tidal model described in Section 3 (Defina et al., 2007). We performed two numerical simulations with two different computational grids describing the present and the 1901 bathymetries of the Venice Lagoon. The analysis focuses on the central-southern part of the lagoon where the condition of fully developed wave field establishes over most of the domain.

The computed bottom shear stress plotted versus bottom elevation shows a remarkable concentration of points (corresponding to the 79.5% of the total area analyzed) around the theoretical curve (see Figure 3 in Defina et al., 2007). Whereas the few points which do not cluster along the curve pertain to tidal channels or to fetch limited areas, that is, regions that do not meet the main model hypotheses.

We further showed that all points falling along the stable branch of the curve are indeed tidal flats. The few points on the unstable branch of the curve (corresponding to less than 10% of the total area) are located on tidal flats close to salt marsh edges where the lagoon morphology is likely far from equilibrium since salt marshes are progressively reducing their extension.

After removing the few points that do not meet the model assumptions, the bottom elevation density functions of the two bathymetries of the Venice Lagoon are evaluated (Figure 10b). The curves show a minimum corresponding to elevations in the unstable range thus confirming that just a small fraction of the basin is characterized by these intermediate elevations, in agreement with the conceptual model.

More consistent with the approach assumed in the present chapter is the short-term morphological evolution of tidal basins. Such analysis requires to take into account all the processes acting at the daily timescale while processes such as organic soil production, soil compaction, eustatism, and sea-level rise, extremely important in long-term evolution, are usually neglected.

Short-term bottom evolution of shallow tidal basins can be studied by coupling an hydrodynamic model which includes wave dynamics with a sediment transport model which includes one or a set of equations for bed evolution. In the following is a short description of the sediment transport model which is being implemented to study the short- and mid-term morphodynamic evolution of the Venice Lagoon.

The bed composition of the Venice Lagoon is characterized by cohesive clayey silt with the exception of the bigger channels branching from the three inlets (see Figure 1) which are sandy and noncohesive (Amos et al., 2004). In the model we use two sedimentological classes: fine sand as a proxy of pure noncohesive sediments and mud (grain size less than 0.063 mm, i.e., silt and clay) as a proxy of pure cohesive sediments. We further assume an average grain size \( d_{50} = 150 \mu m \) for the sand class and the grain size \( d_{m50} = 20 \mu m \) for the mud class. The model considers a 10% of mud content by dry weight to discriminate between noncohesive and cohesive behavior (van Ledden, 2003; van Ledden et al., 2004). The actual bed composition is locally computed by the model as a mix of the two sedimentological classes.
The sediment transport model neglects the horizontal diffusion which is small compared to advection (Pritchard and Hogg, 2003) and solves the advection equation for different sediment classes:

\[
\frac{\partial C_i Y}{\partial t} + \nabla q C_i = E_{\text{sand}} + E_{\text{mud}} - D_{\text{sand}} - D_{\text{mud}}
\]  

(20)

where \( C_i \) (m\(^3\)/m\(^3\)) is the depth-averaged sediment concentration, \( E_{\text{sand}} \) and \( E_{\text{mud}} \) (m/s) are the entrainment of sand and mud computed according to the equations suggested by van Rijn (1993), van Ledden (2003), and van Ledden et al. (2004), which account for the different possible behaviors (i.e., noncohesive or cohesive) of the sand–mud mixture, \( D_{\text{sand}} \) (m/s) is the deposition rate for noncohesive sediments which is proportional to the local sand concentration and still water settling velocity, and \( D_{\text{mud}} \) (m/s) is the deposition rate for cohesive mud evaluated according to the Krone’s formula.

Equation (20) is solved for each of the two sedimentological classes with a first-order finite volume explicit scheme to obtain the time and spatial evolution of suspended sand and mud concentrations. Importantly, all the models share the same computational grid.

A specific bed evolution module, based on the mixing layer concept (Hirano, 1971, 1972), has been developed to predict the time variation of bed elevation and bed composition as a consequence of sand/mud deposition and erosion. Bed elevation is governed by the following equation:

\[
(1 - n) \frac{\partial z_b}{\partial t} = D_{\text{sand}} + D_{\text{mud}} - E_{\text{sand}} - E_{\text{mud}}
\]  

(21)

where \( z_b \) is the bed elevation and \( n \) is the bed porosity.

![Figure 11](image_url) Comparison of measured and suspended sediment concentration at station 1BF and 2BF during the stormy events of 3 April 2003 (left) and 16–17 February 2003 (right).
At each time step the bed evolution model calculates the net variation of bed elevation distinguishing between sand and mud contribution. Based on the deposition and erosion rates of sand and mud, the composition of the active (mixing) layer is then updated.

Results obtained in some preliminary runs are shown in Figure 11. The simulated events are the same discussed in Section 3 (i.e., 3 April 2003, and 16–17 February 2003). Figure 11 compares measured mud concentration with concentration computed when the hydrodynamic model includes or not the wind wave module.

6. CONCLUSIONS

This review has attempted to make an examination of many aspects that must be considered when modeling the hydrodynamics and the morphodynamics of salt marshes and tidal flats in shallow tidal lagoons. While it is clear that considerable progress has been made in the development and application of shallow water models to simulate flow, waves, and sediment transport in estuaries and coastal lagoons it is also clear that, in this branch of computational fluid dynamics, many outstanding problems of physical process representation still remain for the future.

Among the many, a weak area is that of short-term morphodynamics modeling. Algorithms for calculating erosion, transport, and deposition of multiple sediment classes and evolution of sediment stratigraphy caused by wind waves and currents in tidal environments need improvements.

Feedback between hydrodynamics, biology, and morphology represents a further crucial aspect to be dealt with when modeling the tidal flow over salt marshes and tidal flats. This is especially true with respect to the study of morphodynamic equilibrium. To this end, algorithms to describe each specific biological process, at tidal timescale, deserve to be developed whereas long-term hydro-biological models still need to be improved.

In this work it is shown that adequate solutions to many modeling problems can be accomplished through the construction of suitable subgrid models, that is, by setting up a phenomenological representation of the overall processes which provides a statistically equivalent description of the actual physics. This approach also simplifies the coupling of different models conceived to describe specific physical processes coming from different disciplines (e.g., hydraulics, hydrology, morphology, biology, ecology).

Finally, setup and validation procedures based on spatially distributed field data (e.g., wind and wave fields, spatial distribution of bottom sediments and sediment concentration, vegetation seasonal patterns), is a key task to future model design in environmental science. To this end remote sensing, which have become increasingly popular over the past few years owing to large advances in the technology sector, is expected to be a very helpful tool.
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