Development and Validation of a Quasi-Three-Dimensional Coastal Area Morphological Model

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**Abstract:** A quasi-three-dimensional coastal area morphological (Q3DCAM) model has been developed using the process-based approach. It is for simulating complex multiscale coastal processes, primarily morphodynamic changes of the seabed. This software package has integrated three key submodels for simulating irregular wave deformations, nearshore currents, and morphological processes. The quasi-three-dimensional capability of the depth-averaged model has been developed to consider the vertical flow structure inside the surf zone and the cross-shore movement mechanisms of nearshore currents, e.g., undertow and mass flux. To this end, the calculations of the radiation stresses inside the surf zone have been improved by introducing the nonsinusoidal wave model for surface roller effects due to the breaking wave. To predict accurately the wave field and the morphological processes near coastal structures, the wave diffraction effects were included in a multidirectional spectral wave transformation model. The morphodynamic change was modeled by considering the sediment transport due to the combinations of waves and currents. These three submodels were validated by simulating three laboratory experimental cases in regard to: (1) irregular wave deformations over a shoal; (2) longshore currents in a wave basin; and (3) moveable bed evolutions around an offshore breakwater under attack of an incident wave. The numerical results of the morphological modeling confirmed that the Q3DCAM model consisting of the diffraction effects and the surface roller effects is capable of predicting waves, currents, and morphodynamic changes more accurately than before. Therefore, this validated model can be applied to simulate more realistic morphological processes in coastal zones including structures.

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**Introduction**

Understanding morphological processes in coastal zones driven by wave and current is crucial to coastal sediment management, navigation channel maintenance, and designing erosion protection structures. Correct prediction of deposition/erosion process and estimation of sand budget in areas from rivers to coasts/estuaries are key tasks in the regional sediment management. Coastal morphological changes result mainly from transformation and deformation of surface gravity waves propagating across the continental shelf to the beach, the wave-induced currents in the surf zone, and longshore and cross-shore sediment movements. In the past decades, significant progress has been made in the studies of coastal processes by means of physical experiments and computational simulations. Especially, with the aid of the advanced numerical techniques, the simulation of the wave-breaking process has revealed the details of wave propagation through the surf zone (Liu and Losada 2002). However, due to the extreme complexities of natural morphological processes, the mechanisms of sediment transport have neither been fully understood nor described adequately by physical principles and mathematical analyses. Direct simulation of long-term (daily to yearly) morphological evolutions in a real-scale coast coupled with irregular waves and wave-induced currents has been a challenging goal. With the process-based approach having been employed to the development of the coastal area morphological (CAM) model, the simulation of morphodynamic changes and shoreline evolutions has become feasible (e.g., Shimizu et al. 1997; Zyserman and Johnson 2002). In general, this was accomplished by computing sequentially the wave field, the current field, and the seabed changes. Then a new bathymetry will be fed back to affect the computations of the wave and current fields in the next time step (Fig. 1). By this iterative procedure going through the wave-current-morphological models, it is possible to simulate the long-term morphological process by using an empirical sediment transport model for the fine time-scale morphological process (e.g., Reniers et al. 2004; Ding and Wang 2005). However, before the process-based morphological model can be applied to realistic coastal problems, it has to be verified and validated systematically to find out if: (1) the models have included the most important processes for modeling waves, currents, and sediment transport; (2) each module, which serves as a numerical solver of some physical variables, can predict the results of the corresponding variables with reasonable accuracy; and (3) it can run robustly to...
simulate long-term morphodynamic changes by selecting a reasonable long feedback period $T_{sed}$ (see Fig. 1).

According to the existence of different spatial scales, De Vriend et al. (1993) have classified the practical numerical models for simulating morphological processes into four types: (1) one-dimensional (1D) longshore coastline models; (2) two-dimensional (2D) cross-shore coastal profile models; (3) 2D horizontal morphological models; and (4) fully three-dimensional (3D) local morphological models. 1D coastline models can only describe behaviors of the longshore sediment transport and shoreline evolutions by using the sand budget approach; 2D cross-shore coastal profile models are able to predict the vertical variations of coastal profiles, but not the variations of the longshore sediment transport; 2D horizontal morphological models can simulate the morphological variations over a coastal area with a rather wide range of spatial scales (e.g., $10^3$ m$^2$–$10^7$ km$^2$) with the vertical variations of waves and currents ignored. Nevertheless, only fully 3D morphological models are expected to take into account both the vertical and horizontal variations of wave and current (e.g., Lesser et al. 2004). However due to the time-consuming nature for practical problems, 3D models are restricted generally to predict the temporal–spatial morphological changes in a relatively small near field and in a short duration. As mentioned above, the horizontal 2D models have the potential to assess the bed evolutions in large-scale areas, e.g., tidal inlets and estuaries. Therefore, a quasi-3D model with the features of the 2D depth-averaged model and vertical effects of waves and currents would be a feasible tool for the long-term morphological simulations in a large-scale coastal engineering problem. Zyserman and Johnson (2002) have presented a quasi-3D morphological process model in which an empirical 3D shear stress distribution was used to take into account a quasi-3D effect of sediment transport. But, they used a classical hydrodynamic module to compute the depth-averaged velocity without the consideration of the non-uniformities of vertical current in the momentum equations due to the surface rolling effect in surf zone. However, the quasi-3D coastal area morphological (Q3DCAM) model presented in the paper has the following basic characteristics to meet the demand of accuracy and robustness in practical applications: (1) a 2D depth averaged current model including the effects of vertical variations of currents due to the surface rolling of wave breaking; (2) the nonsinusoidal radiation stresses used inside the surf zone and the shortwave volume flux included in the current model for taking into account the undertow; and (3) an accurate and efficient wave driver (wave spectral model) for computing irregular wave deformations including diffraction, refraction, shoaling, and energy dissipation due to the wave breaking. Therefore, this Q3DCAM model enables to compute accurately cross-shore sediment transport and morphological changes around coastal structures.

By means of the process-based approach, the present Q3DCAM model has integrated systematically the three major submodels for simulating irregular wave deformations, wave-induced currents, and coastal morphodynamic changes. As far as irregular wave models are concerned, wave spectral models are more efficient than phase-resolving wave models, but omitting the diffraction effects will be a concern in a case with coastal structures. A multidirectional spectral wave transformation (MDSWT) model with the diffraction effect terms proposed by Mase (2001) was therefore used. In order to take into account the 3D features of the vertical current structures (e.g., the surface rollers or the undertow currents) in the surf zone, the improved radiation stresses formulae derived from the nonsinusoidal wave assumption (Svendsen 1984) were employed in the 2D depth-averaged momentum equations. Svendsen et al. (2003a, b) showed that the nonsinusoidal wave model could give more accurate nearshore currents in surf zone than those without the consideration of the vertical current variations. This Q3DCAM model for simulation of coastal processes has been built in a developed software package called the CCHE2D (Jia and Wang 1999; Jia et al. 2002), which is a systematically verified and validated tool to analyze 2D shallow water flows, sediment transport, and water quality, with natural flow boundary conditions. Similar to the CCHE2D model, the three submodels were discretized in a nonorthogonal grid system so that the models have high accuracy in simulating physical variables in complex coastal zones with irregular coastlines. A time-marching algorithm proposed by Jia et al. (2002) was used for computing the wave-induced currents. A validated algorithm in the CCHE2D for the treatment of wetting and drying in the computational area was directly used for predicting the shoreline movement (Jia and Wang 1999).

These three submodels were sequentially validated by simulating three laboratory experimental cases in regard to: (1) irregular wave deformations over an elliptical shoal (Vincent and Bridggs 1989); (2) longshore currents in a wave basin called the large-scale sediment transport facility (LSTF) (Hamilton and Ebersole 2001); and (3) moveable bed evolutions around an offshore breakwater under attack of an incident wave (Mimura et al. 1983). Modeling the diffraction effect was investigated rigorously by computing the irregular wave deformations in the first case. In comparisons with the measured currents in the second case, it was confirmed that the improved radiation stresses could give much more accurate currents than those by the classical radiation stresses, in which the sinusoidal wave model applied to all over a coast. The validation results about the morphodynamic changes in the third case showed that the nearshore currents predicted by the classical radiation stresses could not convey sediment toward offshore; however, the currents resulted from the improved radiation stresses could produce reasonable morphodynamic changes including sand depositions behind the structure, offshore sand bars,
sediment flux in the model is much slower than the wave and transport process represented by temporal/spatial variations of the seabed were then computed by a sediment balance model with energy density on a spectral energy balance equation. The variations of wave directions due to refraction, diffraction, and wave breaking, is based regular waves such as significant heights, periods, and mean directions in a large-scale coastal zone. Adding the diffraction effects was to extend the spectral model capability to simulate the wave fields in the lee of coastal structures where the effects may be dominant. An improved radiation stress model (Svensden 1984) was used to take into account the effects of the variations of vertical currents due to the surface rolling in the surf zone, in which the sinusoidal and non-sinusoidal wave models were applied to represent the radiation stresses inside and outside the surf zone, respectively. Because of the discontinuity in the radiation stress model, a transition model was proposed to calculate the stresses in a transition zone between the deep water zone and the surf zone. In the wave-induced depth-averaged current model, a bed friction stress due to the combined wave and current (Tanaka and Thu 1994) was used. The cross-shore and longshore sediment transport rates were calculated by means of a total sediment flux model (Watanabe et al. 1986). The morphodynamic changes of the seabed were then computed by a sediment balance model with the downslope gravitational effect. Simulations of the morphodynamic changes in one cycle of the coastal processes were implemented for a period $T_{ed}$ (Fig. 1). Because the fine-scale sediment transport process represented by temporal/spatial variations of sediment flux in the model is much slower than the wave and current processes, this feedback period $T_{ed}$ could be a relative long timescale. The temporal/spatial variations of waves and currents were represented by a series of quasi-steady wave and current fields over the whole period for morphological computation. Therefore this feedback system including the wave–current–morphology interactions becomes efficient to perform a long-term morphological simulation.

Model Descriptions

In the Q3DCAM model, the coastal-process submodels were formulated by several partial differential equations. A phase-averaged model with a term representing the diffraction effects was developed as a fast wave driver to predict statistical variables of irregular waves such as significant heights, periods, and mean directions in a large-scale coastal zone. The MDSWT model, which produces statistical variables of irregular waves such as significant heights, periods, and mean directions in a large-scale coastal zone. Therefore this feedback system including the wave–current–morphology interactions becomes efficient to perform a long-term morphological simulation.

Multidirectional Spectral Wave Transformation Model

The MDSWT model, which produces statistical variables of irregular waves such as significant heights, periods, and mean directions due to refraction, diffraction, and wave breaking, is based on a spectral energy balance equation. The variations of wave energy density $S(x, y, \theta, t)$ in a temporal–spatial–frequency domain under the attack of multidirectional incident waves is described as

$$\frac{\partial S}{\partial t} + \frac{\partial v_S}{\partial x} + \frac{\partial v_y S}{\partial y} + \frac{\partial v_\theta S}{\partial \theta} = Q$$  

where $t=$ time; $\theta =$ wave direction; $x$, $y =$ horizontal coordinates; $Q =$ source term which represents generation, wave–wave interaction, and energy dissipation due to wave breaking and bottom friction; and $v =$ energy transport velocity, of which three components are

$$v_x = C_g \cos \theta, \quad v_y = C_g \sin \theta, \quad v_\theta = \frac{C_g}{\sqrt{\cos \theta - \cos \theta \frac{\partial C}{\partial y}}}$$  

where $C =$ wave celerity; and $C_g =$ wave group celerity. The directional spread of the wave energy is frequency dependent, so the directional formulations are commonly defined as

$$S(f, \theta) = S(f)D(f, \theta)$$  

in which $S(f) =$ 1D frequency spectrum; and $D(f, \theta) =$ directional spreading function. A number of the 1D wave spectra can be found in the relevant literature. In this study, the TMA spectrum (Texel-Marsden-Arsloe, named after the three data sets used in its development) (Bouws et al. 1985) and the Bretschneider–Mitsuyasu ($B$–$M$) spectrum (Mitsuyasu 1970; Goda 1998) were used for simulating the irregular wave deformations in the validation test cases. Similar to the breaking wave dissipation term in Thornton and Guza (1983), Takayama et al. (1991) assumed that the probability density function of the irregular breaking wave height follows a Rayleigh distribution. Therefore, the dissipation term due to wave breaking can be calculated by using the loss of energy flux in a local computational grid. In the computation of the wave energy equation, the frequency domain of a wave spectrum is divided into a set of individual representative wave frequencies. An energy loss due to an individual wave breaking is calculated by taking into account the Goda’s wave breaking criterion, i.e.,

$$H_b = \frac{L_0}{h} \times 0.17 \left(1 - \exp \left(-1.5 \frac{\pi h}{L_0} (1 + 15 m^{2/3})\right)\right)$$  

where $H_b =$ breaking wave height; $L_0 =$ deep water wave length corresponding to the significant wave period; $h =$ water depth; and $m =$ beach slope at breaking. The total wave energy at each nodal point is obtained through summing up the individual wave energy, which is used to calculate the statistical variables of the irregular waves.

It has been well known that the energy balance Eq. (1) can predict correctly the refraction effects of irregular waves, but not the diffraction effects of the waves generated in the lee of coastal structures. One of the existing approaches for adding the diffraction effects is to mimic diffraction with spatial or spectral diffusion (e.g., Resio 1989; Booji et al. 1997; Mase 2001). By analogy with a parabolic wave refraction–diffraction equation, Mase (2001) proposed an improved energy balance equation including the diffusion effects as follows:

$$\frac{\partial S}{\partial t} + \frac{\partial v_S}{\partial x} + \frac{\partial v_y S}{\partial y} + \frac{\partial v_\theta S}{\partial \theta} = \frac{\kappa}{2} \left( \frac{\partial}{\partial \theta} \left( C_g \cos^2 \theta \frac{\partial S}{\partial \theta} \right) - \frac{1}{2} C_g \cos^2 \theta \frac{\partial^2 S}{\partial y^2} \right) + Q$$  

where the new term in the right hand side represents the energy dissipation due to the diffraction effects in the alongshore $y$ direction, which is implicitly perpendicular to wave direction; $\omega =$ wave angular frequency; $\kappa =$ empirical coefficient ($=2.0$–$3.0$ suggested by Mase 2001). Mase (2001) concluded that Eq. (5) also has the advantages of simplicity and robustness in predicting the wave conditions in a large-scale coastal area. However, prior to application of the spectral wave to practical wave prediction, the empirical parameter $\kappa$ has to be calibrated.
Wave-Induced Current Model

The depth- and shortwave-averaged 2D continuity and momentum equations are used for simulating nearshore currents in coastal zones, namely

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

(6)

\[
\frac{\partial \mathbf{u}}{\partial t} + \mathbf{u} \cdot \nabla \mathbf{u} = -g \left( \eta + \frac{1}{\rho_h} \nabla \cdot (h \tau) \right) - \frac{1}{\rho_h} \nabla \cdot \mathbf{R} + \tau^s - \tau^b
\]

(7)

where \( \eta \) is water elevation; \( h \) is water depth; \( \mathbf{u} \) is depth- and shortwave-averaged velocity vector in the horizontal coordinates; \( g \) is gravitational acceleration; \( \rho \) is water density; \( \tau^s \) is depth-averaged Reynolds stress; \( \tau^b \) is wind stress; \( \mathbf{R} \) is radiative stress which represents the net (shortwave-averaged) force the short wave exert on a water column, is defined as (Svendsen et al. 2003a)

\[
\mathbf{R} = \int_{Z_0}^{\eta} (\rho \mathbf{I} + \rho \mathbf{u}_w \mathbf{u}_w) dz - \frac{1}{2} \rho g h^2 \mathbf{I} - \rho \frac{Q_w Q_w}{h}
\]

(8)

where the overbar denotes the time averaging over a short wave period; \( z \) is vertical coordinate; \( Z_0 \) is elevation of the seabed; \( \rho \) is total pressure; \( \mathbf{I} \) is identity matrix (or Kronecker delta); \( \mathbf{u}_w \) is horizontal shortwave-induced velocity; and \( Q_w \) is wave volume flux induced by the short wave motion. As a result, the time-averaged contribution of the short wave forcing is represented in the mass and momentum equations. The radiation stresses can be calculated using the results of wave heights and wave angles obtained by the above-mentioned wave model Eq. (5). A generalized form of the radiation stress is

\[
\mathbf{R} = S_m \mathbf{e} + S_p \mathbf{I} - \rho \frac{Q_w Q_w}{h}
\]

(9)

where the tensor \( \mathbf{e} \) is

\[
\mathbf{e} = \begin{bmatrix} \cos^2 \theta & \sin \theta \cos \theta \\ \sin \theta \cos \theta & \sin^2 \theta \end{bmatrix}
\]

(10)

and the scalar \( S_m \) and \( S_p \) are calculated according to the following two aspects: Outside the surf zone the two terms can be calculated by the sinusoidal wave assumption, i.e.,

\[
S_m = \frac{1}{16} \rho g H^2 \left( 1 + \frac{2kh}{\sinh 2kh} \right)
\]

(11)

\[
S_p = \frac{1}{16} \rho g H^2 \frac{2kh}{\sinh 2kh}
\]

(12)

where \( H \) is wave height; \( k \) is wave number for irregular waves; and \( H \) is significant wave height. Inside the surf zone, the waves are usually nonsinusoidal long waves (wavelength is greater than water depth). In case of wave breaking, a volume of water in breakers, the so-called surface roller is carried with the local wave speed \( C \) (Svensen 1984) proposed improved formulations to calculate the radiation stresses inside the surf zone

\[
S_m = \rho g H^2 \left( B_0 + \frac{Ah}{H^2 L} \right)
\]

(13)

\[
S_p = \frac{1}{2} \rho g H B_0
\]

(14)

where \( A \) is surface roller area which may be represented as \( 0.9 H^2 \); \( B_0 \) is wave shape parameter with \( B_0 = 1/8 \) being the value for sinusoidal waves; and \( L \) is wavelength. Kaihatu et al. (2002) discussed two different approaches to calculate the roller effect. In the study, one of them, the ‘‘static roller,’’ is used to describe the roller shape because it is only dependent on local wave properties. Similarly, the wave volume flux \( Q_w \) outside the surf zone can be calculated by

\[
Q_w = B_0 \frac{g H^2}{C} \mathbf{I}_w
\]

(15)

where \( \mathbf{I}_w \) is unit vector of wave direction = (\( \cos \theta \), \( \sin \theta \)). Inside the surf zone, the wave volume flux reads

\[
Q_w = \frac{g H^2 C^2}{C g h} \left( B_0 + \frac{Ah}{H^2 L} \right) \mathbf{I}_w
\]

(16)

Most existing nearshore current models use Eqs. (11) and (12) derived from the sinusoidal wave theory to calculate the radiation stress. However, it has been already known that these classical radiation stress formulations could not generate accurately nearshore currents inside surf zone when especially wave breaking. The improved radiation stresses consisting of Eqs. (11)–(16) take into account the vertical variations of wave breaker structures inside the surf zone. The 3D features of the cross-shore movement mechanisms, e.g., undertow and mass flux, are reflected accordingly in the model. Putrevu and Svendsen (1999) further derived the comprehensive depth-averaged momentum equations including 3D dispersion terms. Because of the complexities in determining the 3D dispersion coefficients, this wave-induced current model only considers the influence of vertical flow structures due to the surface rollers in the surf zone on the radiation stresses.

The bottom friction stress \( \tau^b \) can be represented as a shortwave-averaged combination of wave and current, namely

\[
\tau^b = \rho C_f |\mathbf{u} + \mathbf{u}_s| (\mathbf{u} + \mathbf{u}_s)
\]

(17)

where \( \mathbf{u}_s \) is wave orbital velocity at the bottom; and \( C_f \) is friction coefficient due to combination of wave and current. The friction law of the combined wave and current proposed by Tanaka and Thu (1994) is used to estimate the friction coefficient in the different flow regimes, i.e., rough turbulent flow, smooth turbulent flow, and laminar flow. The friction coefficient is generally given as follows:

\[
C_f = 0.5 f_w + \sqrt{f_w \beta f_w |\cos \phi| + 0.5 f_w}
\]

(18)

where \( f_w \) and \( f_c \) are friction coefficients due to wave and current, respectively; \( \beta \) is coefficient due to nonlinear interactions of waves and currents; and \( \phi \) is angle between wave orthogonal and current vector. One may refer to Tanaka and Thu (1994) for more details of calculating the friction coefficients in different flow regimes.

In Eq. (7) the depth-averaged Reynolds stress \( \tau^s \) can be represented as a model of turbulence closure. The Boussinesq eddy-viscosity approximation is used for formulating the turbulence stresses

\[
\tau^s = \rho v_e [\nabla \mathbf{u} + (\nabla \mathbf{u})^T]
\]

(19)

where \( v_e \) is eddy viscosity coefficient; and superscript \( T \) = transpose of a tensor. The present nearshore current model provides users with two eddy viscosity turbulence models pertai-
ing to the characteristics of waves. One eddy viscosity model is the Longuet-Higgins model (Longuet-Higgins 1970) which has been proven to be effective in simulations of uniform longshore currents

\[ v_e = Nl \sqrt{gh} \]  

(20)

where \( N = \text{empirical coefficient} \ (0.001--0.01) \); \( l = \text{distance from shoreline toward offshore} \). Another is the Larson–Kraus model (Larson and Kraus 1991), which is suitable for simulations of circulations in the coasts with installation of structures from shoreline toward offshore. The variation of the seabed elevation \( \frac{\partial Z_b}{\partial t} \) is generally utilized to initiate the simulations of the wave-induced transport.

Sediment Transport and Morphodynamic Change Models

The variation of the seabed elevation \( Z_b \) is calculated by considering the local sediment balance and the downslope gravitational transport.

\[ \frac{\partial Z_b}{\partial t} = - \nabla \cdot \mathbf{q} + \frac{\partial}{\partial x} \left( \varepsilon [q_x] \frac{\partial Z_b}{\partial x} \right) + \frac{\partial}{\partial y} \left( \varepsilon [q_y] \frac{\partial Z_b}{\partial y} \right) \]  

(22)

where \( \mathbf{q} = (q_x, q_y) \) = local sediment transport rate; and \( \varepsilon = \text{empirical coefficient} \). The bed evolution is described by a divergence term at the right-hand side and the other two terms for the anisotropic downstream gravitational effect. De Vriend et al. (1993) pointed out that this slope-related transport mechanism enables a coastal profile to reach the equilibrium bed topography; otherwise the morphodynamic simulation only based on the law of mass balance will encounter an inherent instability of bed evolution. The local sediment transport rate has two contributions from wave and current (Watanabe et al. 1986)

\[ \mathbf{q} = \mathbf{q}_w + \mathbf{q}_c \]  

(23)

where \( \mathbf{q}_w \) and \( \mathbf{q}_c \) = local sediment transport rates due to wave and current, respectively

\[ \mathbf{q}_w = A_w F_D \frac{\tau_m - \tau_c}{p g} \mathbf{u}_0 \]  

(24)

where \( \tau_m = \text{maximum bottom shear stress under the action of wave and current} \), \( \tau_c = \text{critical shear stress} \), and \( A_w = \text{empirical coefficient of the wave-induced sediment transport parameter} \) (0.11 for fine sand 0.06 for rough sand); and \( F_D \) was calculated implicitly; and the term of the bed change was calculated as follows:

\[ \tau_c = (\rho_c - \rho)gd\psi_c \]  

(26)

where \( \rho_c = \text{density of sand}; d = \text{grain diameter}; \psi_c = \text{critical Shields parameter} \), and \( A_w = \text{empirical coefficient, which has a form proposed by Shimizu et al. (1997), i.e.,} \)

\[ A_w = \frac{W_0(0.5f_w)^{0.5}B_w}{(1 - \lambda)(s' - 1)(s'gd)^{0.5}} \]  

(27)

where \( s' = \rho_c / \rho - 1; \lambda = \text{porosity of sand}; W_0 = \text{settling velocity}; \) and \( B_w = \text{empirical coefficient of the wave-induced sediment transport rate}. \) Shimizu et al. (1997) suggested that the value of \( B_w \) is 7.0 for laboratory experimental cases, 3.0–5.0 for field cases. \( F_D \) represents the direction function (+1 for onshore, −1 for offshore). The sediment transport rate \( \mathbf{q}_w \) due to mean currents has a similar form to \( \mathbf{q}_w \):

\[ \mathbf{q}_c = A_c \frac{\tau_m - \tau_c}{p g} \mathbf{u} \]  

(28)

where \( A_c = \text{empirical coefficient} \). Shimizu et al. (1997) suggested that the value of \( A_c \) could be approximately ten times as much as that of \( A_w \).

Numerical Approaches in Nonorthogonal Mesh System

For simulation of the coastal morphological processes, the integrated numerical models have been developed to solve the above mentioned four partial differential equations, i.e., the energy balance Eq. (5), the depth-averaged continuity Eq. (6), the momentum Eq. (7), and the seabed level evolution Eq. (22), together with their boundary conditions. A special numerical discretization methodology called the efficient element method (Jia and Wang 1999) was used for discretizing the above four equations in a nonorthogonal grid system. This numerical model is therefore capable of simulating the morphological processes in coastal zones with complex coastlines.

As shown in Fig. 1, the morphodynamic modeling was implemented sequentially. First, the energy balance Eq. (5) describing irregular wave deformations was solved by means of the parabolic approximation, by which the waves were assumed to have a principal propagation direction from offshore toward onshore. The calculations of the waves were therefore carried out line by line from offshore to onshore. In this study, the reflection effect of the waves in the negative x direction was neglected (Ding et al. 2003). Second, a velocity correction method (Jia et al. 2002) was used to decouple and solve the continuity Eq. (6) and the momentum Eqs. (7). And a time-marching algorithm proposed by Jia et al. (2002) was employed for computing the nearshore current field with a time interval (called the current time step). The status of current field may be steady, quasi-steady, or unsteady, in accordance with the status of incident wave. Then, the term of the downslope gravitational transport in the bed level evolution Eq. (22) was calculated implicitly; and the term of the bed change was calculated by means of the Eulerian forward scheme. However, the time interval for simulating the morphodynamic change could be different from the current time step. The bed levels were calculated at each morphological time step by updating the local
sediment transport rate due to the variations of bed levels and bottom frictions. Finally, after a computational duration $T_{\text{calc}}$ for simulating the morphodynamic changes of the seabed, the waves and currents were updated according to the computed new bed levels. This integrated process-based model therefore considered the interaction of wave, current, and sediment transport by adjusting the period $T_{\text{calc}}$ for controlling the frequency of this feedback processes. This consideration is important to facilitate the long-term morphological processes calculation. In addition, to predict the shoreline changes due to morphodynamic changes in the coastal zone, a moving boundary treatment is capable of handling these complex and dynamic wetting and drying processes. The wetting and drying treatment procedure checked locally the situations of flows and the seabed levels at each time step and grid. A typical criterion of critical water depth was given to activate a grid (wet) or to freeze an element (dry) in a computational domain.

**Validation of Q3DCAM Model**

As above mentioned, this Q3DCAM model consists of the three submodels for predicting wave, current, and morphological change. These three submodels were sequentially validated by simulating three laboratory experimental cases in regard to: (1) irregular wave deformations over an elliptical shoal carried out by Vincent and Briggs (1989); (2) longshore currents in a wave basin called the LSTF conducted at the U.S. Army Engineer Research and Development Center’s Coastal and Hydraulics Laboratory (USA-ERDC’s CHL) by Hamilton and Ebersole (2001); and (3) moveable bed evolutions around an offshore breakwater under attack of an incident wave conducted by Mimura et al. (1983). The first case was to test the wave model’s capability for predicting irregular wave heights, periods, and directions; the second case was to test both the wave model and the current model, and especially to confirm the advantages of the improved radiation stresses in taking into account the surface roller effects; the third case was to validate the morphodynamic change models through the wave–current–morphological interaction on simulating the seabed changes around an offshore breakwater.

**Validation of Wave Model**

In order to validate the MDSWT model, the distributions of irregular waves over an elliptical shoal were computed. This elliptical shoal in an experimental flume (Vincent and Briggs 1989) had a major radius of 3.96 m, a minor radius of 3.05 m, and a maximum height of 30.48 cm or 0.3048 m at the center. The region outside the shoal was of constant depth (45.72 cm or 0.4572 m). In the experiments, a directional spectral wave generator produced incident irregular wave conditions which were described by the TMA spectrum with the mean wave directions identical to the axis of the minor radius. The wave heights along several sections in several cases with different spectral parameters were measured. This well-known benchmark experimental case was investigated intensively by a number of modelers by means of the phase-resolving method (e.g., Panchang et al. 1990) and the phase-averaged method (e.g., Mase 2001). In this study, two selected cases with different directional spreading spectra were computed by the MDSWT model. One was the narrow directional spreading spectrum (called Case N1); another was the broad directional spreading spectrum (Case B1). The input parameters for generating the identical TMA frequency spectrum in the two cases were specified as: the incident significant wave height $H_{\text{rms}}=7.75$ cm, the peak period of wave spectrum $T_p=1.3$ s, the alpha constant capable of adjusting variance $\alpha=0.0144$, the peak enhancement factor $\gamma=2$. Fig. 2(a) compares the predicted frequency spectrum by the MDSWT model with that measured by Vincent and Briggs (1989). The TMA frequency spectrum reproduced very well the incident wave conditions.

A Fourier series representation for the wrapped normal spreading function $D(\theta, f)$ in the TMA spectrum was used for producing two different wave spreading conditions, i.e., a narrow spreading in the Case N1 and a broad spreading in the Case B1 (Vincent and Briggs 1989)

$$D(\theta, f) = \frac{1}{2\pi} + \frac{1}{\pi} \sum_{j=1}^{J} \left\{ \exp \left[ -\frac{1}{2} (j\sigma_m)^2 \right] \cos[j(\theta - \theta_m)] \right\}$$

(29)

where $\theta_m =$ mean wave direction; $J =$ total number of terms in the series; and $\sigma_m =$ spreading parameter which determines the width of the directional spreading. For the narrow spreading function,
In both the two cases, 50 harmonics, i.e., $J=50$, were chosen to represent the directional spreading function. Fig. 2(b) compares the two measured and predicted directional spreading functions, respectively. It shows that the two functions almost replicated the directional spreading in the two different TMA spectra.

Meantime, because the B-M spectrum is also commonly used to study irregular wave deformations in engineering applications, the performance of the B-M spectrum was also investigated in the two cases. To do so, the Mitsuyasu-type directional spreading function (or “cosine squared” function) (Mitsuyasu et al. 1975) was employed in the B-M spectrum, i.e.,

$$D(\theta,f) = \frac{2^{2s-1}}{\pi} \frac{\Gamma(s+1)}{\Gamma(2s+1)} \cos^s\left(\frac{\theta - \theta_m}{2}\right)$$  \hspace{1cm} (30)

where $\Gamma$=gamma function; and $s$=spreading function assumed to vary with wave frequency $f$, peak frequency $f_p$, and a peak value of $s$ denoted as $s_{\max}$

$$s = \begin{cases} 
  s_{\max} \frac{(f/f_p)^5}{(f/f_p)^2} & \text{when } f \leq f_p \\
  s_{\max} \frac{(f/f_p)^2}{(f/f_p)^2.5} & \text{when } f > f_p 
\end{cases}$$  \hspace{1cm} (31)

In this study, two values of $s_{\max}$ were selected, in which $s_{\max}=75$ represented the narrow spreading for Case N1, and $s_{\max}=10$ the broad spreading for Case B1.

In the computations of the wave fields, the lower and upper frequency bounds were set to 0.5 and 10 Hz, respectively. The frequency interval 0.095 (i.e., 101 frequency bins) and the angle interval 5.0° (i.e., 37 directional bins between −90 and +90°) were adopted. The spatial increment 10 cm was used in the horizontal coordinates to create a uniform mesh. Due to physically no wave breaking, the terms of breaking wave effects were omitted in the simulation. At first, the wave deformations in the two cases were computed without the consideration of the diffraction effect ($\kappa=0.0$). The distributions of the normalized significant wave heights ($H_s/H_0$) in Case N1 and B1, which were computed by means of the TMA spectra without the diffraction, are shown in

$$H_s/H_0 = \frac{H_s}{H_0}$$  \hspace{1cm} (32)

where $H_s$=significant wave height; and $H_0$=significant wave height at the deep water.

In Fig. 3, the normalized wave heights ($H_s/H_0$) for narrow and broad directional spreading generated by TMA spectra without diffraction (dashed line on figures represents outline of elliptical shoal)

![Fig. 3](image)

In Fig. 4, the comparisons of normalized wave heights between computations and measurements (dashed dot lines represent measurement transects; $\kappa=0.0$)

![Fig. 4](image)
Figs. 3(a and b), respectively. The waves generated by the narrow directional spreading spectrum in Case N1 show a stronger convergent region behind the shoal than that by the broad one in Case B1. In Figs. 4(a and b), the normalized wave heights along three transects obtained by the TMA and the B-M spectra in Case N1 and Case B1 are, respectively, compared with the corresponding measured wave heights. Overall, the numerical results in Fig. 4 show that both the TMA and the B-M spectrum predicted wave heights with quite good accuracy. However, due to ignoring the diffraction effect, the computed wave heights in the lee of the shoal in the narrow directional spreading spectrum (Case N1) are much higher than the measurements. Then this influence of diffraction behind the shoal needs to be concerned.

To take the diffraction effect into account correctly, the $\kappa$ value has been calibrated under the wave conditions of the narrow and broad directional spreading spectra. According to the range of the value suggested by Mase (2001), the computations of the wave fields with different $\kappa$ values from 0.5 to 2.5 were implemented. These numerical simulations with diffraction terms were quite stable and robust because of its virtual effect of diffusion in the equation. The unstable phenomenon pointed out by Holthuijzen et al. (2003) never happened in the simulations. Figs. 5(a and b) show, respectively, comparisons of the normalized significant wave heights along three transects between the measurements by Vincent and Briggs (1989) and the computations. The inclusion of the diffraction term in the wave equation indeed improved the predictions of wave heights right behind the shoal. This term however has a little influence on the wave heights in the broad directional spreading spectrum (Case B1). The transverse profiles of the significant wave heights in the second transect ($x=9.14$ m) obtained from different $\kappa$ values are plotted in Figs. 6(a and b) for Cases N1 and B1, respectively. Similarly, the diffraction made more improvement on the predictions of the wave heights in the narrow directional spectrum than that in the broad one. From Fig. 6(a), a reasonable $\kappa$ value, i.e., $\kappa = 1.5$, was finally suggested for studying the wave deformation around the shoal, which is less than that value ($\kappa = 2.5$) obtained by Mase (2001). From the present results, that value of $\kappa = 2.5$ underestimated the wave heights in most places far from the shoal.
Validation of Current Model

The current module in the Q3DCAM model has been validated by simulating the wave-induced longshore current in a wave basin called the LSTF, which was built at USA-ERDC’s CHL (Hamilton and Ebersole 2001). The facility consists of a 30-m cross-shore, 50-m longshore, 1.4-m deep basin, and includes wave generators, a pumped recirculation system, and an instrumentation bridge. Hamilton and Ebersole (2001) reported two comprehensive test cases conducted on a concrete beach with straight and parallel contours (1:30 slope) which were to verify the facility’s capability to generate the desired longshore uniform currents under regular and irregular incident wave conditions. The currents were driven by long crested waves, which were generated by four piston-type wavemakers. The concrete beach had a longshore dimension of 31 m and a cross-shore dimension of 21 m. The constant still water depth at the offshore was 0.677 m. The computational domain in the study covered the parallelogram-like wave basin in the facility (30 m cross shore and 31 m longshore), and was uniformly divided into nonorthogonal quadrilateral grids with a mesh size of 149×156 (Fig. 7). The irregular incident wave case was chosen to validate the quasi-3D current model, which was named Test-8E in Hamilton and Ebersole (2001). In the case, a TMA spectrum was used to define the spectral shape, in which the spectral width parameter $\gamma$ was 3.3. The TMA spectrum generated an irregular wave condition by which the offshore significant wave height $H_s$ was 0.233 m, the peak period $T_p$ was 2.5 s, and the incident angle 10.0°.

The irregular wave deformation in the wave basin was computed by using the MDSWT model with the parameterized TMA spectrum. Due to physically negligible diffraction effects in the beach, the empirical parameter $\kappa$ was set to 0.0. The energy dissipation term in the wave equation was considered through the wave breaking criterion in Eq. (4). The computed normalized significant wave heights ($H_s/H_0$) and mean wave directions (wave rays) are shown in Fig. 7. The wave breaker line in the figure was determined by the wave breaking criterion of the saturated breakers, in which the breaking wave height was assumed to have a linear relation with the water depth, i.e., $H_b = \gamma h$, where $\gamma = 0.75$. The value of the empirical constant $\gamma$ was the same as that measured by Hamilton and Ebersole (2001). The comparisons of the cross-shore wave height profiles between the measurements by Hamilton and Ebersole (2001) and the computations with and without the wave breaking effects are shown in Fig. 8. The computed wave heights with the wave breaking effects are in good agreement with the measurements.

Hamilton and Ebersole (2001) and the computations with and without the wave breaking effects were shown in Fig. 8. The computed wave heights with the wave breaking effects are in good agreement with the measurements.

The computed significant wave heights and mean wave angles were used to calculate the radiation stresses, and then to finally compute the wave-induced currents. In this study, according to the wave features varying in the cross-shore direction, the radiation stresses outside the surf zone were computed by using the sinusoidal wave model; the stresses inside the surf zone by employing the nonsinusoidal wave model. However, if one would use wave breaking line to define the boundary of the surf zone (e.g., the breaker line in Fig. 7), and would directly apply the two different wave models to compute the stresses inside and outside the surf zone, the discontinuity might happen in the cross-shore profiles of the radiation stresses and eventually could result in a discontinuous longshore current profile. This problem is essentially caused by a nonphysical gap in the values of the roller area $A$ which jump from zero to a finite value at the boundary of the surf zone. Therefore, it is necessary to introduce a transition zone between the surf zone and the deep water zone. Actually, Svendsen et al. (2003b) already pointed out this discontinuity existing in the radiation stress and the currents; and they further proposed an
approach to smooth two wave parameters, i.e., roller area $A$ and phase speed $C$ over a transition zone. But, the detailed implementation and the definition of the transition region were not clear.

In this study, a transition zone illustrated in Fig. 9 was defined to connect the deep water zone with the surf zone: Assuming that the wave breaking happens at the water depth $h_b$, from the location of the wave breaker $x_b$, the transition zone is supposed to extend toward offshore to $x_0$. The width of the transition zone is assumed to be $\gamma_1 h_b$, where $\gamma_1$ is an empirical parameter needs to be calibrated in the application. Inside the transition zone, it can be assumed that the roller area $A$ varies continuously from zero at the deep water to the $0.9H^2$ inside the surf zone

$$A = 0.9 \left(1 + \frac{x - x_b}{\gamma_1 h_b} \right)H^2$$  \hspace{1cm} (32)

The wave-induced currents were computed in the computational domain covered the LSTF wave basin. The irregular wave properties, i.e., significant wave height, mean wave direction, and significant wave period, were used to calculate the radiation stresses and the wave volume fluxes inside and outside the surf zone. The Longuet-Higgins model in Eq. (20) was employed as the turbulence model in the case, in which the empirical coefficient $N$ was 0.001. The recirculation system was simulated by setting the inflow in the left opening boundary and the outflow in the right opening boundary (Fig. 10). The measured longshore-averaged velocities by Hamilton and Ebersole (2001) were specified on the inflow boundary. The so-called open boundary condition was imposed on the outflow boundary. Fig. 10 also shows the circulations illustrated by the streamlines in the upper part and the longshore current profiles in six transects which were projected from the velocity field computed in the nonorthogonal grid points. The computed flow pattern reproduced very well the uniform longshore currents and the circulations in the facility. To confirm the performance of the two models for calculating the

![Fig. 10. Computed wave-induced longshore currents and circulations in LSTF in case of $\gamma_1 = 25.0$ (cross sections named Y19, Y23, Y27, and Y31 are four of measured sections in experiments)](image)

![Fig. 11. Comparisons of longshore currents at cross sections Y19, Y23, Y27, and Y31 where locations are shown in Fig. 10)](image)
radiation stresses inside and outside the surf zone, the longshore current profiles computed by the classical sinusoidal wave model were compared with that by the wave model with the roller effect. The four pictures from Figs. 11(a–d) show comparisons of the current profiles between the measurements and simulations in the four transects, i.e., Y19, Y23, Y27, and Y31, respectively, of which the locations are illustrated in Fig. 10. The currents by the classical sinusoidal wave model without the roller effect (long dash lines) were always underestimated. Although the current profiles inside the surf zone (short dash lines) by the improved radiation stresses (without the transition zone, i.e., $\gamma_1=0.0$) were produced better than the sinusoidal wave model, the unexpected discontinuity appeared in all the four profiles. In contrast, the implementations of the transition zone improved the longshore current profiles quite well. The parameter values of the $\gamma_1$ in Eq. (32) were calibrated, and only the results of the currents at the $\gamma_1$ values 25.0, 40.0, and 50.0 are shown in Fig. 11. A calibrated value of $\gamma_1=40.0$ (bold solid lines) was finally found to predict the most accurate longshore currents both inside and outside the surf zone. This result indicates that the transition zone in the offshore was approximately 8 m wide.

**Validation of Morphological Change Model**

This integrated Q3DCAM model has been validated systematically by simulating the morphodynamic changes due to the interaction of wave and current in a movable bed laboratory experiment conducted by Mimura et al. (1983). This experiment
was carried out in a wave basin being 14 m long, 7.5 m wide, and 0.42 m deep. The beach with 1/20 slope was initially covered with 10-cm thick sand, which had a uniform diameter of 0.2 mm and the density of 2.65 g/cm³. The incident irregular wave with 5.7 cm height and 0.9 s period attacked normally the beach for approximately 12 h. Then, an offshore breakwater of an iron plate with 1.5 m long and 0.5 m height was installed approximately at the wave breaking line (1.8 m offshore from the initial shoreline). The experiment of the morphodynamic changes lasted more than 12 h after the installation of the offshore breakwater. The simulation by this numerical model was started just from the installation of the structure and terminated after 6 h. The measured beach topography at the initial time of the installation was used for creating a computational mesh with a 10-cm uniform spatial increment. This measured initial bathymetry used for the following simulations of the morphodynamic changes was not symmetrical. The wave heights and directions have been computed by using the B-M spectrum as the incident wave spectrum. To include the wave diffraction in the lee of the breakwater, the calibrated empirical parameter \( \kappa \) in the wave Eq. (5) was 2.5. Fig. 12 shows the distributions of the wave heights and the directions computed after 6 h of the structure installation. Fig. 13 compares the computed breaking wave heights with the measurements by Mimura et al. (1983), for which the breaking wave index \( \gamma \) was 0.65. The simulated breaking wave heights are in good agreement with the measured ones. The transition zone between the surf zone and the deep water was defined as the area from the breakwater to the second wave breaking line near the shoreline. Thus, the parameter of the zone width \( \gamma_1 \) was set to 20.0. In Fig. 12, a transition zone for computing the currents at \( t=6 \) h is illustrated as the area surrounded by two dashed dot lines.

By using the abovementioned two radiation stress models, two cases with and without the surface roller effects were implemented to simulate the morphological processes around the breakwater. The Larson–Kraus eddy viscosity model in Eq. (21) was chosen to calculate the turbulence stress with the \( \Lambda \) value equal to 0.3. The time step for simulating the currents was 1.0 s. Each quasi-steady current field in the domain under the action of a steady wave field was obtained after approximately 1,000 steps of time marching computations. The time step for simulations of the seabed evolutions was 0.5 s. The local sediment transport rate coefficient \( B_w \) in Eq. (27) was 7.0 suggested by Shimizu et al. (1997). The critical water depth for checking the wetting and drying process was set to 0.2 mm. The empirical coefficient \( \varepsilon \) in Eq. (22) was 5.0. Through a number of test runs, the feedback period \( T_{sed} \) was finally set to 5 min, namely, the simulations of waves and currents were repeated after every 5 min of the seabed evolution computations. Figs. 14(a) and b) show two computed currents at \( t=6 \) h by the classical radiation stresses and the improved stresses, respectively. Mimura et al. (1983) had roughly measured currents at some locations cross the surf zone by which they proved the existence of the circulations behind the structure. Unfortunately, they did not carry out an overall survey of currents in the coast. Nevertheless, some important differences between the two currents computed by the two radiation stresses can be remarked: (1) the improved stresses produced clearly four circulations in Fig. 14(b) behind the structure and offshore, but the classical stresses only detected two behind the breakwater in Fig. 14(a); (2) the improved stress generated the longshore currents near the first breaking line (almost parallel to the breakwater) more strongly than the classical stresses did; and (3) the onshore currents washing the two tips of the structure shown in Fig. 14(b) were only captured by the improved stresses. The offshore current patterns illustrated in Fig. 14(b), which are virtually connected with the appearance of undertow flow in the surf zone, play an important role in simulating the evolutions of scours around the structure and bars in the offshore.

Consequently, two types of the seabed changes over the 6 h wave attack were computed by using, respectively, the two different current fields. Two snapshots of the bed changes at \( t=6 \) h are shown in Figs. 15(a) and b), respectively. In addition, the measured bed change distribution shown in Fig. 15(c) was obtained by comparing the initial bathymetry with the bed form after 6 h measured by Mimura et al. (1983). The bed changes in the experiment have exhibited abundant sediment depositions behind the breakwater and developing sand bars in the offshore; it also showed the severe shoreline erosions and local scours at the tips of the structure. The results of bed changes shown in Fig. 15(b) by the improved radiation stresses precisely reproduced these changes of the coastal topography both behind the structure and offshore. The local sizes of the depositions and erosions onshore and offshore are quite similar to the observations, although the strength of shoreline erosion in the simulation is not as strong as the experiment. However, the bed changes shown in Fig. 15(a) by the classical stresses could not match the overall morphodynamic changes; the scours and the offshore bars were almost missed, but only the depositions behind the structure.

Furthermore, Figs. 16(a) and b) presented the seabed levels and the total sediment transport rates at \( t=6 \) h obtained by the two radiation stresses. It has been known that both of the sediment fluxes can generate the tombolo-like topography behind the breakwater. However, because the classical radiation stress model

![Fig. 16. Computed local sediment transport rates and bed levels (cm) at t=6 h by: (a) classical radiation stresses; (b) improved stresses](image-url)
did not include the mechanisms of undertow current and mass flux in the surf zone, the sediment transport in Fig. 16(a) was confined inside the lee side of the structure, and thus the sediments were barely transported offshore. In contrast, the local sediment fluxes shown in Fig. 16(b) almost followed the directions of currents obtained by the improved radiation stress shown in Fig. 14(b). The strong cross-shore sediment transport outside the surf zone oriented offshore was obtained. This effect resulted from the inclusion of the 3D features of the currents by considering the surface rolling effect in the surf zone. This morphological model therefore was also capable of producing the offshore sand bars and the local scours at the tips of the breakwater as shown in Fig. 15(b). Figs. 17(a and b) compares the contour lines of bed elevations at −2.0 and −3.0 cm behind the breakwater at t=6 h computed by the improved radiation stresses with the measured contour lines, respectively. Although the shoreline erosions are underpredicted due to lack of knowledge of the sediment transport in the swash zone, the overall agreement between the simulations and measurements is quite reasonable.

Conclusions

In this paper, a newly developed quasi-3D coastal area morphological model (Q3DCAM) is presented, which consists of three key submodels for simulating irregular wave deformations, nearshore currents, and morphodynamic changes. According to our literature review at present, this study is the first investigation to validate systematically the morphological processes model by considering the surface role effect in the surf zone using the non sinusoidal radiation stresses. The wave diffraction effects are taken into account in the wave spectral equation. To remedy the discontinuity in the radiation stresses between the surf zone and deep water region, the concept for modeling the transition zone is proposed and tested.

These three submodels were sequentially validated by simulating three laboratory experimental cases. First, the effectiveness of the diffraction term was confirmed by simulating irregular wave deformations over a shoal in an experimental flume (Vincent and Briggs 1989). Second, improvements on the radiation stresses were mainly concerned with the hydrodynamic model in order to take into account the 3D features of currents induced by the surface role effect, e.g., undertow current and mass flux in the surf zone. The non sinusoidal wave formulations derived by Svendsen (1984) were then adopted to improve the accuracy of the radiation stresses. To remedy the discontinuity problem in the computed currents, a special approach has been proposed to consider the transition zone in the calculation of the roller area. The simulation results about the longshore currents generated in the LSTF facility at the USA-ERDC-CHL showed that the inclusion of the roller effects and the transition zone could remarkably improve the predictions of currents. Finally, the modeling for the morphodynamic changes in an experimental setup of a coast with the installation of a breakwall (Mimura et al. 1983) was conducted. Because of the inclusions of the 3D flow features in the improved radiation stresses, the computed currents could correctly transport sediments onshore and offshore in a complex bathymetry with the structure installation. Therefore, this morphological process model was finally applied to the reproduction of complicated patterns of morphodynamic changes around the structure including sand depositions behind the structure, offshore sand bars, scours at the structure tips, and shoreline erosions. Numerical results have demonstrated that this Q3DCAM model is capable of simulating seabed changes including coastal structures. Thus, it can help researchers and engineers improve their understanding of morphological processes driven by waves and currents and further support coastal sediment management and planning of practical coastal structures.

In order to achieve higher accuracy in the predictions of morphological processes for engineering applications, in the near future additional research to include other mechanisms of sediment transport, e.g., suspended sediment, transport in the swash zone, non equilibrium sediment transport of nonuniform sediment size classes, etc. are to be carried out.

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Notation

The following symbols are used in this paper:

\[
\begin{align*}
A &= \text{roller area;} \\
A_w, A_c &= \text{empirical coefficients;} \\
B_0 &= \text{wave shape parameter;} \\
B_w &= \text{empirical coefficient;} \\
C &= \text{wave celerity;} \\
C_f &= \text{friction coefficient due to combination of wave and current;} \\
C_g &= \text{wave group celerity;} \\
\end{align*}
\]
\[ D = \text{directional spreading function;} \]
\[ d = \text{grain diameter;} \]
\[ e = \text{tensor;} \]
\[ F_D = \text{direction function;} \]
\[ f = \text{wave frequency;} \]
\[ f_c, f_w = \text{friction coefficients due to mean currents and waves;} \]
\[ f_p = \text{peak value of frequency;} \]
\[ g = \text{gravitational acceleration;} \]
\[ H = \text{wave height;} \]
\[ H_b = \text{breaking wave height;} \]
\[ h = \text{water depth;} \]
\[ I = \text{identity matrix;} \]
\[ i_0 = \text{unit vector of wave direction;} \]
\[ J = \text{number of terms in Fourier series;} \]
\[ L = \text{wavelength;} \]
\[ L_0 = \text{deep water wavelength;} \]
\[ i = \text{distance from shoreline toward offshore;} \]
\[ m = \text{beach slope at breaking;} \]
\[ N = \text{empirical coefficient;} \]
\[ p = \text{total pressure;} \]
\[ Q = \text{a source term in wave spectral equation;} \]
\[ Q_m = \text{wave volume flux;} \]
\[ q_s = \text{local sediment transport rates;} \]
\[ q_s, q_v = \text{local sediment transport rate due to current and wave;} \]
\[ R = \text{radiation stress;} \]
\[ S = \text{wave energy density;} \]
\[ S_m, S_p = \text{two components of radiation stress;} \]
\[ s = \text{directional spreading parameter;} \]
\[ s = \text{peak value of s;} \]
\[ T_p = \text{peak spectral wave period (s);} \]
\[ T_{	ext{sed}} = \text{feedback period;} \]
\[ t = \text{time;} \]
\[ U_w = \text{magnitude of wave orbital velocity at bottom;} \]
\[ u = \text{depth-averaged and short-wave-averaged velocity vector;} \]
\[ u_w = \text{horizontal shortwave-induced velocity;} \]
\[ W_0 = \text{particle settling velocity;} \]
\[ x, y = \text{horizontal coordinates;} \]
\[ Z_b = \text{elevation of seabed;} \]
\[ z = \text{vertical coordinate;} \]
\[ \beta = \text{coefficient due to nonlinear interactions of waves and currents;} \]
\[ \Gamma = \text{the gamma function;} \]
\[ \gamma = \text{wave breaker index;} \]
\[ \gamma_1 = \text{an empirical parameter for defining width of transition zone;} \]
\[ \varepsilon = \text{empiric coefficient;} \]
\[ \eta = \text{water elevation;} \]
\[ \theta = \text{wave direction;} \]
\[ \theta_m = \text{mean wave direction;} \]
\[ \kappa = \text{empirical coefficient;} \]
\[ \Lambda = \text{empirical coefficient;} \]
\[ \lambda = \text{porosity of sand;} \]
\[ \nu = \text{eddy viscosity coefficient;} \]
\[ \rho = \text{water density;} \]
\[ \rho_s = \text{density of sand;} \]
\[ \sigma_m = \text{spreading parameter which determines width of directional spreading;} \]
\[ \tau_c, \tau_s = \text{critical shear stress;} \]
\[ \tau_s = \text{seabed friction stress and wind stress;} \]
\[ \tau_c = \text{depth-averaged Reynolds stress;} \]
\[ \phi = \text{angle between wave orthogonal and current vector;} \]
\[ \psi_c = \text{critical Shields parameter; and} \]
\[ \omega = \text{wave angular frequency.} \]

References


