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Short communication

Combined tidal and wind driven flows and residual currents

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ABSTRACT

The effect of a residual current on the combined tidal and wind driven flow and the resulting bedload sediment transport in the ocean has been investigated, using a simple one dimensional two-equation turbulence closure model. Predictions of the combined tidal and wind driven flow with given residual currents are presented, showing that the residual current has a substantial effect on both the depth averaged mass transport and the mean bedload transport directions; in some cases the effect of the residual current is to almost reverse the mean bedload transport direction. The residual current affects the rotation of the flow due to the Coriolis effect in the lower part of the water column (the near-surface flow is wind dominated), causing a larger or smaller clockwise rotation of the depth averaged mass transport, depending on the direction of the residual current.

1. Introduction

Since residual currents are crucially important for the transport of sediments, larvae and contaminants in tidally dominated flows, a considerable effort has been made to investigate the impact of these currents on tidal flows. This has led to a vast number of papers, such as Harris and Collins (1991) looking into sediment transport paths in the Bristol Channel; Fry and Aubrey (1990) finding that tidal asymmetry in shallow embayments can cause net sediment transport; Ranasinghe and Pattiaratchi (2000) revealing that tidal asymmetry in inlets is more important for the net sediment transport than the occurences of flood or ebb dominant diurnal tides; van Maren and Gerritsen (2012) concluding that finer and coarser sediments are transported in different directions in the Singapore Straits, since the transport of finer sediments is dominated by residual currents while the transport of coarser sediments is governed by the tidal asymmetry. Further predictions of the sediment transport in tidal inlets are given by, among others, Fietcher et al. (2006) and Wu et al. (2011).

In a previous work (Holmedal and Myrhaug, 2013) the combined tidal and wind driven flow at intermediate water depths was investigated using a simple tidal model. This model was validated against field measurements tabulated by King et al. (1985) of a tidal flow in the Celtic Sea over a flat bottom at 120 meters water depth. These field measurements contained a tidal drift over the water column; the drift velocity outside the tidal boundary layer was taken as the residual current. By accounting for this residual current and taking into account the wind stress in their model, Holmedal and Myrhaug (2013) found a fair agreement between the predicted and measured tidal drift over the water column. It appears that the tidal drift is mainly caused by the residual current, although the simulations also show that the wind direction is important for the direction of the drift through the water column.

The purpose of the present work is to investigate the effect of residual currents on the combined tidal and wind driven flow and the resulting sediment transport at intermediate water depths, relevant to near-coastal waters. It will be shown that the residual currents have a large impact both on the direction of the mean (averaged over a tidal period) depth averaged velocity and on the mean bedload sediment transport, while the direction of the mean surface velocity is less affected. Visualizations of the bedload transport are provided using the near-bed tidal ellipse. The effect of residual currents on the combined tidal and wind driven flows has, to the author's knowledge, not been investigated in detail in an idealized setting. Overall, the present work yields new insight into combined tidal and wind driven flows, and represents an extension to the work by Holmedal and Myrhaug (2013).

2. Model formulation

2.1. Governing equations

The tidal turbulent flow is modeled as a horizontally uniform boundary layer where the tidal forcing is driven by the mean ocean surface slope oscillating with the tidal frequency. The

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Reynolds-averaged equations for conservation of the mean momentum and mass become

$$\frac{\partial u}{\partial t} = -\frac{1}{\rho}\frac{\partial p}{\partial x} + \frac{\partial}{\partial z}\left(\nu_T \frac{\partial u}{\partial z}\right) + f v \tag{1}$$

$$\frac{\partial v}{\partial t} = -\frac{1}{\rho}\frac{\partial p}{\partial y} + \frac{\partial}{\partial z}\left(\nu_T \frac{\partial v}{\partial z}\right) - fu \tag{2}$$

where u and v are the horizontal velocity components, p is the pressure, ρ is the density of the water, ν_T is the kinematic eddy viscosity, f is the Coriolis parameter, and g is the gravity acceleration. The turbulence closure is given by a $k-\epsilon$ model using the logarithmic wall law near the rough bottom (see e.g. Rodi, 1993)

$$\frac{\partial k}{\partial t} = \frac{\partial}{\partial z}\left(\frac{\nu_T}{\sigma_k}\frac{\partial k}{\partial z}\right) + \nu_T\left(\left(\frac{\partial u}{\partial z}\right)^2 + \left(\frac{\partial v}{\partial z}\right)\right) - \epsilon \tag{3}$$

$$\frac{\partial \epsilon}{\partial t} = \frac{\partial}{\partial z}\left(\frac{\nu_T}{\sigma_\epsilon}\frac{\partial \epsilon}{\partial z}\right) + c_{\epsilon 1}\frac{\epsilon}{k}\nu_T\left(\left(\frac{\partial u}{\partial z}\right)^2 + \left(\frac{\partial v}{\partial z}\right)^2\right) - c_{\epsilon 2}\frac{\epsilon^2}{k} \tag{4}$$

where k is the turbulent kinetic energy and ϵ is the turbulent dissipation rate. The kinematic eddy viscosity is given by

$$\nu_T = c_1 \frac{k^2}{\epsilon}. \tag{5}$$

where the standard values of the model constants have been adopted, i.e. $(c_1, c_{\epsilon 1}, c_{\epsilon 2}, \sigma_k, \sigma_\epsilon) = (0.09, 1.44, 1.92, 1.00, 1.30)$.

The bedload sediment transport is given by a formula by Nielsen (1992)

$$\Phi = 12\theta^{\frac{1}{2}}(\theta - \theta_c)\frac{\theta}{|\theta|} \tag{6}$$

where

$$\Phi = \frac{q_b}{\left(g(s-1)d_{50}^3\right)^{\frac{1}{2}}} \tag{7}$$

$$\theta = \frac{\tau_b}{\rho g(s-1)d_{50}} \tag{8}$$

Here q_b is the instantaneous dimensional bedload transport, τ_b is the dimensional instantaneous sea bed shear stress, $s = 2.65$ is the density ratio between the bottom sediments and the water, and d_{50} is the median grain size diameter. The critical Shields parameter $\theta_c = 0.05$ must be exceeded for bedload transport to take place.

2.2. Boundary conditions and tidal forcing

The sea bed is assumed to be hydraulically rough, and a logarithmic wall law is applied here in conjunction with zero velocity at the bottom A wind stress τ_s is specified at the surface and is related to the wind speed 10 meters above sea surface (U_{10}) by the empirical relation $\tau_s = \rho_a c_d U_{10}^2$, where $\rho_a = 1\ \mathrm{kg\,m^{-3}}$ is the density of air and c_d is a friction factor. The forcing is assumed to be in the East–West direction (i.e. along the x-axis), while the wind direction is varied. Here the tidal forcing is driven by the mean ocean surface slope which is oscillating with the tidal frequency $\sigma = 2\pi/T_p$ where T_p is the tidal period. The assumption of hydrostatic pressure (following from the boundary layer approximation) yields

$$-\frac{1}{\rho}\frac{\partial p}{\partial x} = -g\frac{\partial \xi}{\partial x}\cos(\sigma t) \tag{9}$$

$$-\frac{1}{\rho}\frac{\partial p}{\partial y} = 0 \tag{10}$$

Here $\partial\xi/\partial x$ is the amplitude of the mean ocean surface slope.

2.3. Inclusion of residual currents

First the tidal boundary layer without wind is calculated from Eqs. (1)–(5) using Eqs. (9) and (10). This yields the free stream velocities $U_0(t)$ and $V_0(t)$, taken from the vertical zone outside the purely tidal boundary layer. These free stream velocities are applied to incorporate residual currents. Due to the boundary layer approximation, the following relations apply:

$$-\frac{1}{\rho}\frac{\partial p}{\partial x} = \frac{\partial U_0}{\partial t} - fV_0 \tag{11}$$

$$-\frac{1}{\rho}\frac{\partial p}{\partial y} = \frac{\partial V_0}{\partial t} + fU_0 \tag{12}$$

When residual currents are considered, the tidal forcing in Eqs. (9) and (10) is replaced with the tidal forcing in Eqs. (11) and (12), using U_0 and V_0. Furthermore, U_0 and V_0 in Eqs. (11) and (12) are substituted with i.e., $U_0 \pm U_s$ or $V_0 \pm U_s$ (depending on the direction of the residual current) where U_s is the magnitude of the residual current. The hydrostatic horizontal pressure gradients are then evaluated from Eqs. (11) and (12), using $U_0 \pm U_s$ and $V_0 \pm U_s$ instead of U_0 and V_0.

2.4. Numerical method

A finite difference method was used to solve the parabolic Eqs. (1)–(5) using second order central differences in space. Geometric stretching of the mesh was applied to obtain a fine resolution near the bed and close to the free surface; here 800 gridpoints were applied in the vertical direction. The spatial discretization of the equations for the horizontal velocity components, turbulent kinetic energy and dissipation rate give a set of stiff differential equations that were integrated simultaneously in time by the integrator VODE (Brown et al., 1989). Further details about the numerical method are given in Holmedal and Myrhaug (2013).

3. Results and discussion

Residual currents may exist locally in the ocean, due to e.g. differences in temperature and salinity or by the presence of large-scale ocean currents. These residual currents may have a different direction from both the wind and the tidal forcing. Generally there is a co-existence of wind, tidal forcing and residual currents in the ocean, and for intermediate and shallow water depths, the interactions between these three components are important for both the local mass transport and the sediment transport. Here the effect of the residual current on the combined tidal and wind driven flow and sediment transport is investigated.

The choice of parameters is similar to that of Holmedal and Myrhaug (2013), except for the presence of the residual current: The tidal period is 12.5 h, the Coriolis parameter is $f = 1.112 \cdot 10^{-4}\ \mathrm{s^{-1}}$, the amplitude of the mean ocean surface slope is given by $\partial\xi/\partial x = 3 \cdot 10^{-6}$ and the water depth is 120 m. The effect of the surface waves is represented as a surface roughness z_s; in the present work this is chosen as $z_s = 0.3$ cm. A sandy flat bottom, consisting of medium sand with $d_{50} = 0.21$mm, is considered; the bottom roughness is related to d_{50} by the empirical formula $z_0 = d_{50}/12$. The residual current towards the East (+) or West (−) has been taken into account by substituting U_0 with $U_0 \pm 0.02$ m/s in Eqs. (11) and (12). Similarly, the residual current towards the North (+) or South (−) has been taken into account by substituting V_0 with $V_0 \pm 0.02$ m/s. Here $U_s = 0.02$ m/s is about the same magnitude as the residual currents tabulated in King et al. (1985). In the present setting the effect of the wind is taken into account by specifying a wind stress at the surface as $\tau_s = 0.2$ Pa. For the stable and neutral atmospheric conditions more common