A model of suspended sediment transport by internal tides

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12 January 1999

Submitted to Continental Shelf Research

ABSTRACT

The ability of internal tides to resuspend and advect sediment over continental shelfs and slope regions is investigated by applying an internal wave and sediment transport model. Numerical experiments are carried out, firstly, with the ratio of bathymetry and internal waves characteristics creating critical, subcritical, and supercritical conditions, and secondly, for an observed section of the Australian North West Shelf. In the former cases, the model is forced with an internal tide propagating through the model domain. The latter application involves forcing by a barotropic tide which in turn generates internal waves at the shelf slope. Internal wave generated bottom layer shear stresses are large enough to resuspend sediment. The application of a turbulence closure scheme results in the creation and maintenance of a thin nepheloid layer. The thickness of the suspended sediment layer is controlled by vertical diffusion which is large within the bottom boundary layer, but very small outside. The residual velocity and the asymmetry associated with the velocity field, result in both down- and upslope net suspended sediment fluxes, and deposition of resuspended material onto the shelf. These suspended sediment fluxes are largest for critical bottom slopes. The parting point between down- and upslope net sediment flux is found to be sensitive to the formulation of vertical mixing with the parting point moving downslope for increased mixing. At the Australian North West Shelf, near the shelf break and upper slope, the net flux of resuspended material is influenced by both the barotropic and internal tide. The phase relationship in bottom layer shear stresses generated from those two tides causes regions of both enhancement and reduction in the resuspension rates and net suspended sediment fluxes.

1. Introduction

Cacchione and Southard (1974) suggested that bottom shear stresses generated by internal tides may be large enough to resuspend sediment over continental shelfs and slope regions. This mechanism may even play a dominant role in controlling the distribution of sediments where the water depth is large enough to attenuate any direct impact upon sediment distributions by wind generated surface waves and currents (Cacchione and Drake, 1986). It is possible that these internal waves create and maintain a layer of high turbidity near the sea bed referred to as nepheloid zone. In a review of continental shelf transport mechanisms, Nittrouer and Wright (1994) concluded that internal tides are one of several important mechanisms which may lead to the across shelf transport of resuspended sediment.

Internal tides are energetic oscillatory flows often generated in stratified water from barotropic tidal flow over steep topography. Observations suggest that internal tides are a common phenomenon on many continental slopes and shelves (Huthnance, 1989). In addition to directly observing internal tides, numerical models are powerful tools to advance the understanding of these phenomena. Results obtained from applications of primitive equation models are consistent with observations (e.g. Sherwin and Taylor, 1990; Holloway, 1996). While the potential of internal waves in resuspending sediment has been implied in several studies, it is difficult to directly observe the contribution made by internal tides upon resuspending sediment. Bogucki et al. (1997) explained observed high sediment concentrations on the Californian shelf to resonant internal solitary waves.

Over sloping topography, bottom intensified flows associated with internal waves can occur when the slope of the topography ($\alpha = \Delta h / \Delta x$) is similar to that of the internal wave characteristics: $s = \pm\sqrt{\left(\omega^2 - f^2\right) / \left(N^2 - \omega^2\right)}$, where ω is the wave frequency, f the Coriolis parameter, and N the buoyancy frequency. Such slopes are termed critical slopes (e.g. Holloway, 1985). These intensified flows may well be the dominant process contributing to bottom shear stresses on continental slopes and hence be important in determining sediment resuspension and transport.

It is the aim of this study to investigate the interaction between the flow associated with internal tides over sloping seabeds and the suspended sediment dynamics. For this purpose, a sediment transport equation is incorporated into the Princeton Ocean Model - POM (Blumberg and Mellor 1978; Mellor, 1996), a non-linear, free-surface, sigma coordinate, hydrostatic, primitive equation model which incorporates the Mellor-Yamada level 2.5 turbulence closure scheme (Mellor and Yamada, 1982; referred to hereafter as MY2.5). In particular, the intensification of bottom boundary layer (BBL) currents and the asymmetry in the flow from up- to downslope, as shown by Holloway and Barnes (1998), is expected to impact on the sediment dynamics. The work here, extends that of Holloway and Barnes (1998) to investigate the relationship between the BBL dynamics and the resuspension and transport of fine sediment. At the seabed, sediment fluxes are formulated as dependent upon the locally computed BBL shear stress and an assumed critical shear stress for resuspension and deposition. The model developed in this study excludes any feed back between suspended sediment and the density stratification. It is applied only to fine mode, non-cohesive material characterized by very small settling velocities, and excludes any bed load sediment transport.

In a test case of an earlier version of the MY2.5 turbulence closure scheme, Weatherly and Martin (1978) found that model predicted BBL thickness was consistent with observations from the Florida Continental Shelf. A similar method was chosen by Adams and Weatherly (1981) who included the effects of resuspended sediment and investigated its interaction with the BBL circulation. It was found that resuspended sediment can reduce BBL shear stresses by about 45%. Jewell et al. (1993) were able to explain observed sedimentological features of the Amazon continental shelf. The circulation was computed by applying POM and forcing the model with climatological winds and the four major tidal constituents (M_2, S_2, K_1 and O_1). No solution to the sediment transport equation was provided, but observations of the sediment distribution at the seabed indicated that high sediment accumulation rates correlated well with locations of minima in BBL shear stresses.

Section 2 of this paper describes the sediment transport model. A number of computational experiments are carried out for varying bottom slopes and parameters. These experiments consider idealized model bathymetry (subcritical, critical, and supercritical seabed slopes) and forcing, and also a realistic across shelf section from the Australian North West Shelf (NWS). In the former three cases, the model is forced at the western boundary by specifying a first mode internal wave (IW) that propagates through the domain. A linear temperature stratification and a constant salinity is prescribed giving a constant buoyancy frequency. The latter application employs observed stratification and barotropic tidal forcing. Results are presented in Section 3 for idealized topography. Section 4 discusses results from several sensitivity experiments with critical slopes, and in Section 5 results are shown for the application of the model to the NWS. A discussion and summary of the model results is given in Section 6.

2 The Sediment Transport Model

The configuration of the POM based model for internal tides is that reported by Holloway (1996) and Holloway and Barnes (1998). A first mode semi-diurnal internal tide is specified at the offshore boundary and allowed to propagate through the model domain and over a sloping sea bed. The model predicts vertical eddy viscosity for turbulent mixing of momentum, vertical eddy diffusivity for turbulent mixing of heat and salt, and bottom drag coefficients. The Mellor-Yamada level 2.5 turbulence closure scheme is used and is modified by adding a weak background value of $2x10^{-5}$ m^2s^{-1} to the mixing coefficients, as was done by Holloway and Barnes (1998). This tends to smooth out discontinuities at the top of the BBL. BBL shear stresses are computed by a quadratic friction law. In those particular configurations, the model represents an across shelf section for three idealized topographies. Also a realistic across shelf section from the NWS is considered where the internal tide is generated through the specification of barotropic tidal surface forcing.

The transport of sediment is described through an equation solving for the advection and diffusion of sediment, similar to the equations for temperature and salinity in POM (Mellor, 1996). As the model is only forced by periodic flows, net advective velocities will only arise from any nonlinearity in the flow. Net sediment transports may also arise from the asymmetry between up- and downslope flows and the vertical profile of the sediment concentration. The sediment fluxes in and out of the water column are specified at the seabed, dependent on critical BBL shear stresses for resuspension and deposition. Within the interior of the water column, sediment is assumed to be a conservative property. It is allowed to sink toward the seabed by specifying a sinking velocity as an additional component to computed vertical advection.

Several assumptions were made in solving the transport of resuspended sediment. Firstly, suspended sediment is assumed to be composed of only one non-cohesive size class. Secondly, the class of sediment is assumed to contribute to the fine mode with very small settling velocities. For example, Hill et al. (1994) found that on the northern California continental shelf, the size distribution of resuspended sediment exhibited a bimodal distribution. Resuspended and captured sediment sizes fell into two main factions grouped according to their sinking velocity: a coarse and a fine mode with settling velocities of about 2.6 x 10^{-4} m s^{-1} and 2.5 x 10^{-5} m s^{-1}, respectively. Both contributed about one-quarter to two-thirds and one quarter, respectively, to the total amount of suspended material. The reminder of the total suspended sediment constituted a third group with sinking velocity smaller than 1.5 x 10^{-5} m s^{-1}. Thirdly, it is assumed that sediment is transported as suspended load. No account for a bed load is made which is primarily coarse sediment that moves or rolls along the seafloor (e.g. Smith, 1977). Another difference between suspended and bed load is the grain to grain contact in the latter case. In contrast, fine mode suspended material and its individual particles are supported by turbulence (Nittrouer and Wright, 1994).

There are uncertainties associated with the choice of particular formulations and parameters chosen in this specific model application. These primarily affect the exact amount of sediment resuspended and deposited, and computed net fluxes. Choosing different parameters such as critical stresses or sinking velocities will modify the exact quantity of sediment kept in suspension or deposited. In a series of sensitivity tests, the impact made by particular parameter choices is explored.

Based upon the equation for temperature or salinity in a σ-coordinate system (Mellor, 1996), the sediment transport equation solved in this model to describe the concentration *C* of a particular sediment class or size suspended within the water column is given by:

$$\frac{\partial CH}{\partial t} + \frac{\partial CUH}{\partial x} + \frac{\partial CVH}{\partial y} + \frac{\partial C(W+Ws)}{\partial \sigma} = \frac{\partial}{\partial \sigma}\left[\frac{K_z}{H}\frac{\partial C}{\partial \sigma}\right] + \frac{\partial}{\partial x}\left[HK_x\frac{\partial C}{\partial x}\right] + \frac{\partial}{\partial y}\left[HK_y\frac{\partial C}{\partial y}\right] + R - D \tag{1}$$

where symbols are: sediment concentration C [kg m^{-3}]; time t [s]; horizontal velocities U and V [m s^{-1}] in direction x and y [m]; vertical velocity W [m s^{-1}],; sinking velocity W_s [m s^{-1}]; total water depth H [m]; vertical coordinate σ; vertical eddy diffusivity coefficient K_z [m^2 s^{-1}]; horizontal eddy diffusivity coefficients K_x and K_y; resuspension flux R [kg m^{-2} s^{-1}] and deposition flux D [kg m^{-2} s^{-1}]. R and D are source and sink terms and are only non-zero at the seabed.

The derivation of the sediment transport equation and its boundary conditions has been described in the literature in detail (e.g. Mellor, 1996). In the following sections, only the formulation used to specify sediment fluxes at the seabed is discussed in some detail. The resuspension fluxes were formulated in direct dependence on BBL shear stresses which are computed by POM. The friction velocity u_* [m s^{-1}] and BBL shear stress τ [N m^{-2}] are related to each other through the relationship: $\tau = \rho \cdot u_*^2$ (Soulsby, 1983), where ρ [kg m^{-3}] is water density. This approach is similar, for example, to that by Pohlmann and Puls (1994) or Clarke and Elliott (1998). The resuspension term defining a sediment flux R [kg m^{-2} s^{-1}] into the water is:

$$R = M \cdot \left(\frac{\tau - \tau_{crs}}{\tau_{crs}} \right) \quad \text{for } \tau > \tau_{crs} \qquad (2)$$

where symbols are: BBL shear stress τ [N m^{-2}]; critical BBL shear stress for resuspension τ_{crs} [N m^{-2}], resuspension constant M [kg m^{-2} s^{-1}]. The resuspension flux is zero for τ smaller than τ_{crs}. The resuspension constant M may depend upon properties such as sediment type, composition, density, and thickness. It can also be considered as the product of total mass of sediment on the seabed available for resuspension M_t [kg m^{-2}] and an entrainment rate β [s^{-1}]

(Raaphorst et al., 1998). Values of M may be determined from laboratory experiments, and Clarke and Elliott (1998) applied constants of 2 x 10^{-4} kg m^{-2} s^{-1} and 5 x 10^{-5} kg m^{-2} s^{-1}.

There seems to be considerable uncertainty associated with the choice for M or the entrainment rate β (McLean, 1985). As it is intended to model the effect of internal tides upon the distribution of suspended sediment, the exact sediment quantities are only of secondary interest and therefore, M was set to 1 kg m^{-2} s^{-1} and the seabed provides an infinite source for the resuspended load. Hence, computed sediment quantities such as suspended sediment concentration and flux are only relative quantities in this study. It is possible to re-scale these quantities according to the exact choice of M to obtain an estimate of actual magnitudes as quantities are linearly proportional to M. This approach was also adopted by McLean (1985).

The BBL shear stress dependent deposition of resuspended sediment was again formulated following previous methods used by, for example, Buller et al. (1975), Pohlmann and Puls (1994), and Clarke and Elliott (1998). The flux or deposition rate D [kg m^{-2} s^{-1}] of sediment settling back to the seabed is computed as the product between the suspended sediment concentration C [kg m^{-3}] and the sinking velocity W_s [m s^{-1}]. A fraction of this flux is being deposited if the BBL shear stress τ is less than then the critical shear stress for deposition t_{crb} [N m^{-2}]. The deposition is defined as:

$$D = C \cdot W_s \cdot \left(\frac{\tau_{crb} - \tau}{\tau_{crb}} \right) \qquad \text{for } \tau < \tau_{\text{crb}} \qquad (3).$$

Sinking velocities for suspended sediment depend upon many factors such as particle size, shape, composition, ability to aggregate, and the physical environment. For different classes of suspended material found in the North Sea, for example, settling velocities are $O(10^{-4}\text{-}10^{-6}\ \text{m s}^{-1})$ (Pohlmann and Puls, 1994). The in-situ measurements by Hill et al. (1994)

exhibited a bimodal distribution of suspended sediment and sinking velocities ranged from 2.6 x 10^{-4} m s^{-1} to 2.5 x 10^{-5} m s^{-1}. Kachel and Smith (1989) listed settling velocities for various sediment classes collected at the Washington shelf in O(10^{-2}-10^{-5} m s^{-1}). The values given by Sternberg (1972) for quartz are O(10^{-2}-10^{-4} m s^{-1}). Wiberg et al. (1994) listed settling velocities in 10^{-2}-10^{-4} m s^{-1} and McLean applies settling velocities in O(10^{-2}-10^{-5} m s^{-1}). This study aims to investigate the impact of internal tides upon resuspended sediment which belongs to the fine mode and model runs are carried out using relative slow sinking velocities of 10^{-4} and 10^{-5} m s^{-1}.

The BBL shear stress used to compute resuspension (Eq. 2) and deposition (Eq. 3) fluxes, is obtained from a quadratic friction law (e.g. Solsby, 1977):

$$\tau_x = \rho \cdot C_d \cdot \sqrt{\left(U_b^{\,2} + V_b^{\,2}\right)} \cdot U_b\,; \quad \tau_y = \rho \cdot C_d \cdot \sqrt{\left(U_b^{\,2} + V_b^{\,2}\right)} \cdot V_b \qquad (4)$$

where τ_x and τ_y [N m^{-2}] are BBL shear stress in *x* and *y* direction, C_d bottom drag coefficient, U_b and V_b [m s^{-1}] bottom layer velocity in *x* and *y* direction. As the model uses a σ-coordinate, the thickness of the bottom layer depends on the total water depth. The drag coefficient C_d is defined as the larger of 0.0025 and $\left\{ \kappa \big/ \ln\left(h / z_o\right) \right\}^2$, where κ = 0.4 is the Von Karman constant, h is the distance between the seabed and the grid point where U_b and V_b are calculated, and z_o is the roughness length set as 0.02 m. Pingree and Griffiths (1979), for example, used the same formulation to investigate sand transport rates in the North Sea.

Both resuspension and deposition occur when the BBL shear stress is larger (resuspension) or smaller (deposition) than an assumed critical shear stress. Much of the present knowledge on critical shear stress data for resuspension and deposition dates back to laboratory studies carried out by Shields (1936). Cacchione et al. (1994) applied Shields' empirical findings and computed critical shear stresses of about 0.08 - 0.25 N m^{-2} for sediment grain sizes of 4 x 10^{-3} - 3 x 10^{-5} m.

The critical shear stresses for resuspension based upon in-situ observations, laboratory studies, and applied in previous modeling studies are all within the range of those determined initially by Shields (1936) (Kachel and Smith, 1989; Krone, 1993, Pohlmann and Puls, 1994; Friedrich and Wright, 1995). A value of about 0.1 N m^{-2} seems to be representative for fine quartz which is one of the most mobile sediment classes (Sternberg, 1972; Sternberg and Larsen, 1975; Kachel and Smith, 1989; Wiberg et al, 1994). Therefore, a critical shear stress for both resuspension and deposition of 0.1 N m^{-2} was chosen for this study.

3 Idealized Topography

The results from nine computational experiments are reported, the experiments falling into three groups. The first group uses idealized topography, the second group are sensitivity experiments, and the third group is composed of two experiments which use realistic topography, density, and forcing. The main features of the computational experiments are summarized in Table 1 and are discussed below. Common parameters are listed in Table 2.

A schematic of the idealized model topography is shown in Fig. 1a. The model design is the same to that chosen by Holloway and Barnes (1998). Experiments 1, 2, and 3 were carried out for a critical, subcritical, and supercritical seabed slopes. There are 110 grid points in x-direction, and 5 grid points in y-direction with a resolution of 1 km and 20 km respectively. In the vertical direction, the grid is non-uniform with 61 σ-levels. The maximum depth is 200 m in the west and 40 m on the shelf. The open boundary is in the west and the coast is located in the east.

The definition of the three idealized seabeds is based upon the ratio (α/s) between internal wave slope characteristics and the seabed slope. Parameters chosen are ω (1.41 x 10^{-4} s^{-1}), f (-5.0 x 10^{-5} s^{-1}, corresponding to 20^o S), and N (0.0016 s^{-1}). Ratios with α/s = 1, α/s < 1, and α/s > 1 define critical, subcritical, and supercritical conditions. At a supercritical condition, much of the internal wave energy is reflected off the shelf slope seaward, while at subcritical conditions, the shelf slope is transmissive and most the of internal wave energy propagates onto the shelf (e.g. Cacchione and Drake, 1986).

In all three experiment using idealized topography, a linear temperature profile is specified with 30 ^oC at the surface and 10 ^oC at 200 m depth. The salinity is set to 35.0 throughout the water column and no heat and fresh water fluxes are specified. At the western boundary, the model is forced by specifying a first mode internal tide propagating into the domain by prescribing the velocity in *x*-direction as: $u(z,t) = u_o \cdot cos(\pi z \cdot H^{-1})\ sin(\omega t)$. The velocity amplitude at the M_2-tidal period is set at u_o = 0.3 m s^{-1}. This produces velocities of magnitude similar to those observed on numerous continental slope and shelf break regions, e.g. on the NWS (Holloway, 1994).

Of particular interest are instantaneous across shelf suspended sediment fluxes (q_i) [kg m^{-2} s^{-1}], net suspended sediment fluxes integrated for a tidal cycle (q_t) [kg m^{-2} s^{-1}], and with depths (q_z) [kg m^{-1} s^{-1}]. For comparison between individual model experiments, these quantities were computed according to the following definitions:

$$q_i(x,z,t) = U(x,z,t) \cdot C(x,z,t)$$
$$q_t(x,z) = \frac{1}{T}\int_T q_i dt \qquad (5)$$
$$q_z(x) = \int_H q_t dz$$

The horizontal velocity U and the suspended sediment concentration C are computed by the internal wave and sediment transport model. T is the M_2 period of the tidal cycle (12.42 h), H (m) the local water depth, dt and dz are time step (s) and depth interval (m).

The model is integrated for a period of 4 days. After about 1.5 days the dynamic state of the model approached a quasi-equilibrium and results from the model runs are shown after this spin-up phase of the model. The temporal and spatial distribution of the cross-topography bottom layer velocity U, bottom layer shear stress τ, resuspension R, deposition D, and relative sediment concentration C are shown in Figs. 2a - 2e for the critical, subcritical, and supercritical slopes. Bottom current intensification over the sloping topography is strongest when the bathymetry is at a critical slope. Bottom currents are about 0.05 m s^{-1} and 0.1 m s^{-1} for the upslope and downslope direction, respectively (Fig. 2a). For the supercritical slope experiment, velocities are found to be smallest along the shelf slope with values of about 0.05 m s^{-1} for both up- and downslope flow. In this case, much of the internal tide energy is being reflected at the slope generating largest velocities of about 0.1 m s^{-1} just seaward of the slope. The slope of the regions of constant velocity (Fig. 2a) indicates the phase propagation of the internal tide. This is seen to be mainly upslope, but some off-shore propagation is seen for the super-critical slope.

Bottom layer shear stresses are largest for the critical shelf slope and during the intensification of the downslope bottom current (Fig. 2b). Maximum values of about 0.6 N m^{-2} are simulated. Maximum values for bottom layer shear stresses during the upslope current phase are about half the magnitude of those simulated during the downslope phase. During the reversal of down- to upslope current and vice verse, the computed stresses are smaller than the critical shear stresses of 0.1 N m^{-2} specified for both resuspension and deposition. For subcritical conditions (Fig. 2b), bottom shear stress maxima are reduced by about a third compared to those values for critical conditions. Similar to bottom layer velocities for supercritical slopes (Fig. 2a), maxima in bottom layer shear stresses of about 0.4 N m^{-2} are found to the west of the slope at supercritical conditions. In contrast to both critical and subcritical slopes, bottom layer velocities and shear stresses are at a maximum during upslope flow between about 20 and 30 km.

The time series of total mass of resuspended and deposited sediment are shown in Figs. 2c and 2d. It needs to be kept in mind that these quantities are only relative values because the resuspension constant was set to 1 kg m^{-2} s^{-1}. Both properties depend upon the computed bottom layer shear stress. Deposition is about 100 times smaller than resuspension and this results in a continuous increase of simulated suspended sediment concentration.

Resuspension occurs during both up- and downslope flows and is largest during downslope motion for both critical and subcritical conditions. No resuspension occurs during the switch from down- to upslope current or vice verse, while deposition is largest during this period. Resuspension is relatively weak over the super-critical slope.

The difference between resuspension and deposition is dependent on the actual sediment concentration within the water column. With resuspension approaching a quasi steady state similar to the dynamic state of the system during the spin up phase of the model, the sediment flux into the water column varies only within a tidal cycle. The amount of material entering the water column is constant for each tidal cycle and the suspended sediment concentration, for example shown for the bottom layer, increases with each tidal cycle (Fig. 2e). Both the deposition of material and the bottom layer suspended sediment concentration are not in an equilibrium during the integration.

In experiments with critical and subcritical slopes, resuspension occurs only over the slope (Fig. 2c). Resuspended material is deposited on the shelf and in deeper regions of the model domain (Fig. 2d), indicating a transport of material away from the slope. Both regions are characterized by very small bottom layer shear stresses (Fig. 2b) which are smaller than the critical values for resuspension and deposition.

For the critical slope, the highest bottom layer suspended sediment concentrations (Fig. 2e) are simulated at slope locations with the largest bottom shear stress (Fig. 2b) and during upslope flow at about at distance of 40 km. This is near the center of the slope region. For subcritical conditions, this maximum concentration is found near the shelf break at about 60 km. The maximum bottom layer suspended sediment concentration seems to coincide with a time of upslope or the change from up- to downslope current.

Time series of vertical profiles of velocity, suspended sediment concentration, and suspended sediment flux are represented in Fig. 3. The profiles are shown for a mid-slope location at 40 km with a water depth of about 130 m and computed with bathymetry at the critical slope. Profiles are only shown for 25 m above the seabed where the suspended sediment concentrations are at highest levels. The downslope flow is more intense and confined to a narrower region than upslow flow as described by Holloway and Barnes (1998). Maximum velocities are about 0.5 m s^{-1} and 0.3 m s^{-1} for down- and upslope flows at height above the seabed of about 10 m and 25 m, respectively. Suspended sediment is confined to a thin layer of about 15 m above the seabed. The model results indicate, that resuspended sediment is trapped within the BBL where vertical eddy diffusivity is very large (see Holloway and Barnes, 1998) homogenizing the suspended sediment. Outside the BBL, vertical eddy diffusivity is equivalent to the background value (Table 2) which is about 1000 times smaller than BBL values computed with MY2.5 turbulence closure scheme. Holloway and Barnes (1998) found peak values of 0.035 m s^{-2} for vertical mixing at the critical slope which approached the background value at a height of 10 m above the seabed. Due to the assumed infinite sediment source, continuous resuspension of material increases the suspended sediment concentration throughout the water column during the simulation period. A maximum of more than 6 x 10^4 kg m^{-3} is simulated toward the end of day 4.

The suspended sediment concentration (Fig. 3b) is largest during upslope flow (Fig. 3a), although the resuspension is more intense during downslope flow. This is in particular the case during day 2. With the start of day 3, this relationship is changing and concentration maxima are found when the upslope flow shifts to downslope flow. The direction of instantaneous sediment fluxes (Fig. 3c) is directly determined by the direction of the flow. In contrast, the direction of net sediment fluxes depends upon the relationship between velocity and suspended sediment concentration and how those vary over a tidal cycle. This is discussed in Section 4.

The instantaneous resuspended sediment flux is largest during downslope flows (Fig. 3c). This may establish a net sediment transport downslope for this particular location which will be discussed in detail in some of the following sections. Most of the up- and downslope sediment flux is confined to the BBL approximately 15 m thick. The maximum downslope suspended sediment flux coincides with a maximum in downslope velocity (Fig. 3a). However, the maximum upslope flux is still confined to a layer 10-15 m thick above the seabed, but the maximum upslope velocity is now found above this layer.

The erosion, defined as the difference between total resuspended and total deposited material, integrated over a tidal cycle is shown in Fig. 4. All three slopes are characterized by a predominance of sediment loss. Most of the eroded material remains in suspension and is advected into regions where bottom layer shear stresses are very small and material is deposited. In all cases there is a on-shelf deposition and this is largest for the critical slope.

The directions of the depth integrated net sediment transports integrated over a tidal cycle (Fig. 5) are predominantly downslope for all three slopes and largest for the critical slope. There is a parting point between down- and upslope transport and this is close to the shelf break, but for both the critical and subcritical slope it is about 5 km seaward of the shelf break. The parting point only coincides with the shelf break for the supercritical slope. The net sediment fluxes are about 100 times smaller than the instantaneous fluxes (Fig. 3). The processes determining the net resuspended sediment fluxes and the position of the parting between down- and upslope sediment fluxes are discussed in Section 5.

4 Sensitivity Experiments

There is a vast choice of possible sensitivity experiments. A review of the literature indicates that a critical shear stress for resuspension of 0.1 N m^{-2} for the fine sediment seems to be agreed upon. No sensitivity test for this parameter is carried out. More uncertain are choices for deposition rates and sinking velocities and the dependence on vertical eddy viscosity. The sensitivity experiments give some indication of the impact upon the model results made by changing these values.

Fig. 6a shows the distribution of sediment flux integrated over a tidal cycle as a function of depth and distance along the slope for the critical slope model discussed in Section 3. Downslope fluxes dominate the depth range closest to the seabed at most locations. A maximum of about 70 kg m^{-2} s^{-1} (negative values indicate downslope flux) is simulated at about 15 m above the seabed. The on-shelf advection of suspended sediment is at a maximum of 30 kg m^{-2} s^{-1} at 15 m off the seabed and at a distance of 50 km, i.e. near the shelf break. Net suspended sediment fluxes are about a 1000 times smaller than instantaneous fluxes (Fig. 3c), but both exhibit maximum suspended sediment fluxes at about the same depth.

The non-linearity of the velocity field associated with the internal tides will produce a residual flow that is expected to influence the magnitude and direction of net suspended sediment fluxes. These residual velocities are computed to range from about -1.4 x 10^{-3} m s^{-1} downslope to about 1.6 x 10^{-3} m s^{-1} upslope (Fig. 6b). Both net suspended sediment fluxes and residual velocities are computed for the last tidal cycle of the model integration. During this period, the typical sediment concentration (Fig. 3b) is O(10^{4} kg m^{-3}). With a typical residual velocity of O(10^{-3} m s^{-1}), net sediment fluxes are estimated to be O(10 kg m^{-2} s^{-1}), similar to the computed values shown in Fig. 6a. This suggests that the net sediment transport is largely determined by the residual flow.

Experiments are carried out for the critical slope with the vertical eddy diffusivity coefficient kept constant in time and throughout the water column (see Table 1). Values of 1 x 10^{-3} m^{2} s^{-1} (Experiment 6) and 2 x 10^{-3} m^{2} s^{-1} (Experiment 7) are used and these are about a tenth of the maximum values computed for the BBL by the MY2.5 turbulence closure scheme (see Holloway and Barnes, 1998).

The vertical structure of the residual velocity field does not vary significantly between the experiments (Fig. 6b, 6d, and 6f). In all three cases, the boundary (i.e. the zero contour line) between up- and downslope residual velocity is at about the same location, close to the shelf break and at a height of 11-12 m above the seabed. The magnitude of the residual velocity is slightly increased in both sensitivity experiments with a maximum of -1.8×10^{-3} m s^{-1} for downslope velocities at 48 km. Upslope flows are slightly reduced to values of 1.4×10^{-3} m s^{-1} (Fig. 6d) and 1.2×10^{-3} m s^{-1} (Fig. 6f). However, the vertical structure of net suspended sediment fluxes varies distinctively between the individual model runs. In applying a vertical eddy diffusivity coefficient which is constant throughout the water column, resuspended sediment is mixed more efficiently upward and away from the BBL. It then experiences a different advective environment than in the case of the MY2.5 mixing scheme. Instantaneous sediment concentrations and fluxes outside the BBL in both experiments with constant eddy coefficients (not shown) are larger than those computed using the MY2.5 turbulence closure scheme.

In all three experiments, the maximum in upslope net suspended sediment transport occurs at the shelf break (Fig. 6). This location is independent of the vertical mixing parameterization which seems to control only the height above the seabed of the maximum in transport. The location of the maximum in resuspended sediment transport is most likely a results of the bottom current intensification and an abrupt decrease in the magnitude of bottom layer velocity at the shelf break.

The changes made to vertical eddy viscosity and diffusivity coefficients, results in changes in bottom layer velocities, bottom layer shear stresses, and resuspension (Fig. 7). With constant vertical eddy diffusivity, more sediment is kept in suspension and advected onto the shelf, in comparison to the control experiment, and an increased instantaneous suspended sediment deposition is simulated (Fig. 7). This is true for both the sensitivity experiments. Despite reduced bottom shear stresses and subsequently net erosion along the shelf slope (Fig. 8), on-shelf net sediment fluxes are increased in both experiments using constant vertical mixing coefficients (Fig. 9).

The deposition of material is primarily controlled by a balance between vertical eddy diffusivity and the settling velocity. Within the BBL, eddy diffusivity is of O(10^{-2} m^2 s^{-1}) (Holloway and Barnes, 1998). The sinking velocity is increased by a factor of 10 for the critical slope model to a value of 10^{-4} m s^{-1} in experiment 4 (Table 1). However, it is found that this impacts only marginally upon the deposition of suspended material and the resulting net erosion (Fig. 8). Sinking velocities are small and within the velocity range associated with vertical turbulent motion in the BBL.

In contrast to varying the settling velocity, the effect of specifying deposition as being independent of the critical shear stress, i.e. $\tau_{crb} = \infty$ in (3), is more significant (Fig. 8). The net erosion is much reduced along the shelf slope in comparison to the control run (Fig. 8). The resuspension of sediment is largest here. Subsequently, a large fraction of the resuspended material is deposited at the same location and at each time step if the deposition is independent of a critical bottom layer shear stress. During the control experiment, deposition occurred only during periods when the computed bottom layer shear stress falls below a critical shear stress of 0.1 N m^{-2}. This occurred when currents switch from down- to upslope flow or vice verse (Fig. 2). On-shelf deposition just shoreward of the shelf break is also increased when deposition is stress independent (Fig. 8). While local resuspension is small shoreward of the shelf break (Fig. 2), material resuspended at the slope is advected onto the shelf. With a critical shear stress dependent deposition, that material remains in suspension and is advected downslope again. In the case of shear stress independent deposition, material transported just shoreward of the shelf break settles out immediately at each time step.

A comparison is made between depth integrated net suspended sediment transports computed for the control experiment and all four sensitivity experiments (Fig. 9). It was previously indicated (Fig. 6) that net suspended sediment fluxes are larger for runs with constant vertical mixing compared to the control experiment. The constant vertical eddy diffusivity coefficient allowed removal of more sediment from the BBL than for the MY2.5 turbulence closure scheme. The increased settling velocity produced changes in the depth integrated fluxes. The smallest on-shelf depth integrated net suspended sediment fluxes are found when deposition is stress independent.

Fig. 9 shows that the location of the parting point between up- and downslope net sediment transport depends upon the parameterization of vertical eddy diffusivity. For weak vertical mixing, there is a higher resuspended sediment concentration within the BBL, and the parting point moves closer toward the shelf break. Using a coefficient of 2 x 10^{-3} m^2 s^{-1} constant throughout the water column, the parting point is located at about mid-slope and a distance of 40 km. With the smaller coefficient of 1 x 10^{-3} m^2 s^{-1}, downslope suspended sediment fluxes are increased (Fig. 6c), and the parting point moves toward the shelf break. For the control experiment which applies the MY2.5 turbulence closure scheme, resuspended sediment remains within the BBL. Downslope fluxes are largest and the parting point between down- and upslope transports is closest to the shelf break.

5 Realistic Topography

In a realistic application, the model was integrated for an across shelf section of the NWS. The dynamics of internal tides for this application were previously investigated by Holloway (1996). The bathymetry of the model is shown in Fig. 1b. There are 300 intervals in x-direction with a grid spacing of 1 km and the origin of the coordinate system located in the west. The maximum depth is 1600 m which is resolved in 61 σ-levels. In contrast to Holloway (1996), temperature and salinity data recorded during a summer (January, 1995) RV Franklin cruise at the NWS is used.

Instead of forcing the model directly with an internal tide, realistic barotropic tidal forcing was applied. Holloway (1996) found that if the model is forced with an M_2 barotropic tidal elevation of amplitude 0.95 m (representative of Spring Tides) at the off-shore boundary, internal tides are generated which are sensitive to stratification and bathymetry. The simulated internal tide velocity field is consistent with observations although tended to underestimate observed velocities (Holloway, 1996). Different locations across the shelf produce subcritical and approximately critical conditions with associated intensification of bottom slope currents.

A time series of the horizontal bottom layer velocity distribution is shown in Fig. 10a. These velocities are a combination of barotropic and baroclinic tides. The velocity field is in a quasi-equilibrium with a maximum flow of about 0.08 m s^{-1} at the center of the across shelf section at about 160 km. Velocities around 0.06 m s^{-1} are generated between about 120 km to 180 km. There is no indication that the bottom layer current intensification is significantly stronger during down- or upslope flow in this experiment which employs the MY2.5 turbulence closure scheme.

Bottom layer shear stresses are larger than the critical shear stress for resuspension in the region between about 120 km to the coast, i.e. for water depths less than about 170 m (Fig. 10b). The maximum shear stress of about 0.5 N m^{-2} is produced for both down- and upslope flow at a distance of about 180 km (near the shelf break) during day 3 of the integration. At this particular location the water depth is about 60 m (Fig. 1b). Similar to bottom layer velocity and shear stress, no significant difference exists between the total mass of resuspended sediment during down- and upslope flow (Fig. 10c). The location of maximum resuspension coincides with that of maximum bottom layer shear stress at about 180 km. Both, the magnitude of resuspension and deposition fluxes is of the same order to those fluxes simulated for the experiments using idealized topography (see Figs. 2c and 2d).

The vertical distribution of suspended sediment at the end of the model integration exhibits a maximum in the relative suspended sediment concentration just shoreward of the shelf break between about 160 - 180 km (Fig. 11). In both directions, i.e. on- and off-shore, the suspended sediment concentration decreases. Similar to the results obtained from experiments employing idealized slopes (Fig. 3), resuspended sediment is trapped within a layer of about 15 - 20 m thickness above the seabed. There is a sharp suspended sediment concentration gradient toward the interior of the water column. Along the whole across shelf section, the direction of the net suspended sediment transport integrated for the last tidal cycle of the model experiment is downslope (Fig. 12).

In order to determine the contribution of the barotropic tide to the sediment dynamics, the model is run for constant temperature and salinity. Therefore, no internal tides are generated. The global integral of mass eroded during the runs with internal tides is about 2.3 x 10^{12} kg (Experiment 8) and without internal tides is about 2.1 x 10^{12} kg (Experiment 9). The integral is computed form the data shown in Fig. 13. The difference in the total amount of resuspended sediment of about 10 % is attributed to the bottom current intensification due to internal tides. The particular phase relationship between BBL currents (and bottom layer shear stresses) generated by internal and barotropic tides (Holloway and Barnes, 1998) leads to either an enhanced or reduced erosion at particular locations along the shelf at intervals of one internal tide wave length, about 10-15 km. This is most noticeable between 120 - 180 km (Fig. 13a), i.e. the region where internal tides are being generated. The phase lag between barotropic and internal tide can subsequently lead in some locations to an enhanced erosion and net downslope transport of suspended sediment (Fig. 14), for example, between about 160 - 170 km. Running the model for an additional 4 days showed no change in the locations of the peaks in erosion.

The internal tide generation is most intense to the east of the shelf break region located between about 160 - 180 km. In turn, internal tide generated bottom shear stresses are very efficient in enhancing or reducing sediment resuspension by barotropic tides (Fig. 13a). Shoreward of the shelf break, internal tide generation is small, and the difference in erosion between the stratified and unstratified experiment is only small. The ratio (α/s) of bathymetry to internal wave characteristics at the NWS (Fig. 13b) exhibits some correlation with erosion pattern (Fig. 13 a). Both distributions exhibit maxima at distances of about 120 km and 160 km although the peaks are offset by about 3 to 4 km. These are regions with steep bottom topography (Fig. 12) approaching ‘critical’ slope characteristics where internal tide generation is expected to be most intense. Note that the large peak on the shelf at 175 km is caused by the internal tide generated over the slope, propagating onto the shelf and being in-phase with the barotropic tide at this location.

6 Summary and Discussion

A simple model for the resuspension of sediment was incorporated into an internal wave model which is based upon the Princeton Ocean Model (POM). The assumptions made in the design of the sediment model limited the study to the fraction of resuspended sediment which belongs to the fine, non-cohesive mode and which is predominantly transported as suspended sediment. The model employs the MY2.5 turbulence closure scheme and therefore, it allows to prognose the BBL shear stress as an independent variable computed from a quadratic friction law.

The model was used to investigate the impact of internal tides and the associated bottom layer current intensification upon the resuspension and transport of fine sediment. For this purpose, three experiments employed idealized topography with the ratio between internal wave characteristics and bathymetry close to critical, subcritical, and supercritical conditions. Four additional experiments with bathymetry close to a critical condition demonstrated the sensitivity of the results to sinking velocity, critical shear stress for deposition, and vertical eddy diffusivity. Finally, two experiments were carried out with observed topography, temperature and salinity stratification, and barotropic tidal forcing for a cross shelf section on the NWS.

The modeling demonstrates the ability of internal tides to create and maintain a nepheloid layer. This supports Cacchione and Drake’s (1994) suggestions that internal tides are important in suspending sediment in the absence of wind generated currents and waves at the continental shelf and slope. Their findings were based upon theoretical and laboratory studies as well as observations. In the sediment transport model described here, the MY2.5 mixing scheme generates a large turbulent diffusivity within the BBL. This results in a homogenization of the suspended sediment within the BBL, and a limitation of the suspended sediment layer thickness to that of the dynamic BBL. Bottom layer shear stress, resuspension, deposition, and on-shelf sediment transport are largest for the critical slope.

The application of constant values for turbulent mixing results in a reduction of bottom layer shear stress and resuspension, but enhances on-shelf net sediment transport and deposition. Using a two layer model, Heathershaw et al. (1987) found that the shelf break is a location of suspended sediment parting. In the present study, it was found that the location of the parting point is sensitive to the choice of the eddy diffusivity parameterisation. The more suspended sediment is mixed into the upper levels of the model, the further the parting point moves downslope from the shelf break. This also results in an increase in the net on-shelf transport of suspended material and subsequent deposition. In-shore of the parting point, there is an increase in the build up of sediment. These results highlight the importance of correct specification of vertical mixing in predicting sediment resuspension and transport.

The modeling results of Adams and Weatherly (1981) revealed that the density of suspended sediment affects the BBL stratification. This reduces the strength of BBL currents which in turn, results in a shear stress decreased by about 45 %. This effect has not been considered in the present study, as only relative sediment concentrations are simulated. However, even with a reduction of bottom layer shear stresses by about 45 %, the shear stress computed in all experiments are well above the critical values assumed for resuspension. The reduction in shear stress will lead to lower suspended sediment concentration and a reduced magnitude of net sediment fluxes, but would not affect the actual direction of net fluxes.

The application of the sediment transport model to the NWS indicates that internal tides may lead to the resuspension of sediment and a net-downslope sediment transport. With barotropic tidal forcing, internal tides result in an enhancement in some locations or a reduction in other locations of the net erosion. This depends on the phase relationship between the barotropic and baroclinic tides. Particularly, over the slope, the bottom stress can be dominated by the contribution from the internal tide.

Acknowledgment

The authors would like to thank Dr. G. L. Mellor for making available to the ocean modeling community the Princeton Ocean Model applied in this work. Dr. C. R. Sherwood is thanked for helpful discussions and getting us underway with this work. The project received financial support through a grant from the Australian Research Council.

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Figure Captions

Figure 1: Schematic representation of the model designs used for nine experiments with (top) idealized topography [ratio between bathymetry and internal wave characteristics creating (a) critical, (b) subcritical, and (c) supercritical conditions], and (bottom) realistic across shelf topography of the Australian North West Shelf. Data from the latter model design are only shown for the region between 100 km and 280 km.

Figure 2a: Time series of bottom layer velocity [x 0.01 m s^{-1}] computed for experiments with an idealized bathymetry at the (a) critical, (b) subcritical, and (c) supercritical slope. Contour interval is 0.05 m s^{-1} and upslow current velocities are shaded.

Figure 2b: Time series of computed bottom layer shear stress [N m^{-2}] for experiments with an idealized bathymetry at the (a) critical, (b) subcritical, and (c) supercritical slope. Contour interval is 0.1 N m^{-2}. Only bottom shear stresses larger than the critical shear stress for resuspension (0.1 N m^{-2}) are contoured.

Figure 2c: Time series of total amount of resuspended material [x 10^9 kg] computed for experiments with an idealized bathymetry at the (a) critical, (b) subcritical, and (c) supercritical slope. Contour interval is 1 x 10^9 kg.

Figure 2d: Time series of total amount of deposited material [x 10^7 kg] computed for experiments with an idealized bathymetry at the (a) critical, (b) subcritical, and (c) supercritical slope. Contour interval is 1 x 10^7 kg.

Figure 2e: Time series of suspended sediment concentration [x 10^2 kg m^{-3}] in the bottom layer and computed for experiments with an idealized bathymetry at the (a) critical, (b) subcritical, and (c) supercritical slope. Contour interval is 1 x 10^2 kg m^{-3}.

Figure 3: Time series of near bottom profiles: (a) velocity contoured in intervals of 1 x 10^{-2} m s^{-1} , (b) suspended sediment contoured in intervals of 1 x 10^4 kg m^{-3}, and (c) suspended sediment flux contoured in intervals of 1 x 10^4 kg m^{-2} s^{-1}. Results are only shown for Experiment 1 with a bathymetry at the critical slope and at 40 km. Downslope velocities and suspended sediment fluxes are contoured in dashed lines.

Figure 4: Total mass [x 10^{12} kg] of eroded material computed during a tidal cycle from the difference between resuspended and deposited material. Gray shading indicates the location of idealized slopes with bathymetry close to the (a) critical, (b) subcritical, and (c) supercritical slope.

Figure 5: Depth integrated net sediment transport [kg m^{-1} s^{-1}] during tidal cycle and computed for the experiments with (a) critical, (b) subcritical, and (c) supercritical slope.

Figure 6: Net sediment fluxes (6a, 6c, and 6e) contoured in intervals of 10 kg m^{-2} s^{-1} and residual velocities (6b, 6d, and 6f) contoured in intervals of 1 x 10^{-2} m s^{-1}. Results are from Experiment 1 (a, b) applying the MY2.5 turbulence closure scheme, Experiment 6 (c, d) and Experiment 7 (e, f). The latter experiments assume isotropic eddy diffusivity of 1 x 10^{-3} m^{-2} s^{-1} and 2 x 10^{-3} m^{-2} s^{-1} respectively. All experiments were carried out for a bathymetry at the critical slope. Its location is indicated in gray shading. Downslope transports are dashed. The vertical axis indicated the distance of the seabed. Data are

only shown for the model domain with large sediment fluxes which are in proximity to the slope region.

Figure 7: Time series of bottom layer shear stress (top), resuspension (middle), and deposition (bottom) computed during sensitivity Experiment 6. Contour intervals are 0.1 N m^{-2}, 1 x 10^9 kg, and 1 x 10^7 kg for shear stress, resuspension and deposition, respectively. Experiments are carried out with a bathymetry at the critical slope and its location is indicated in gray shading.

Figure 8: Total mass [kg] of eroded material (resuspension minus deposition) computed for the last tidal cycle of Experiment 1, and the sensitivity Experiments 4 - 7.

Figure 9: Depth integrated net sediment fluxes [kg m^{-1} s^{-1}], for Experiment 1 (control experiment) and all four sensitivity experiments. All experiments were carried out for a bathymetry at the critical slope. The location of the slope is indicated in gray shading.

Figure 10: Time series of (a) bottom layer velocity [m s^{-1}], (b) bottom layer shear stress [N m^{-2}], (c) total mass resuspended [kg], and (d) total mass deposited [kg] computed for the model experiment using realistic topography and forcing (Experiment 9). Contour intervals are 0.01 m s^{-1} for velocity (dashed lines for downslope velocity), 0.1 for shear stress, 1 x 10^9 kg for resuspension, and 1 x 10^7 kg for deposition. Results are only shown for the mid-shelf region characterized by bottom shear stresses larger than that specified for initiation of resuspension.

Figure 11: Relative suspended sediment concentration [kg m^{-3}] computed during Experiment 8 using realistic topography and forcing. Data are shown at the last time step of the model integration.

Figure 12: Net suspended sediment flux [kg m^{-2} s^{-1}] computed during Experiment 8 using realistic topography and forcing. Data are integrated for the last tidal cycle.

Figure 13: (a) Total mass [kg] of eroded material (resuspension minus deposition) for experiments using realistic topography and forcing. Data are plotted for both Experiment 8 (with stratification) and Experiment 9 (without stratification); (b) ratio (α/s) between slope characteristics and internal wave characteristics (see Introduction for details).

Figure 14: Depth integrated net sediment flux [kg m^{-1} s^{-1}] for experiments using realistic topography and forcing. Data are plotted for both Experiment 8 (with stratification) and Experiment 9 (without stratification).

Table 1: List of computational experiments. Experiment 1 to 3 comprise model experiments carried out for idealized topography with standard parameter settings shown in Table 2; Experiment 4 to 7 are sensitivity experiments; and Experiment 8 and 9 were carried out for realistic topography and forcing.

Experiment	Topography	Forcing	Parameter
1	critical slope	linear T, constant S, internal tide at western boundary	Table 2 values
2	subcritical slope	as experiment 1	as experiment 1
3	supercritical slope	as experiment 1	as experiment 1
4	critical slope	as experiment 1	$w = 10^{-4}$ ms^{-1}
5	critical slope	as experiment 1	$t_{crb} = \infty$ N m^{-2}
6	critical slope	as experiment 1	$K_v = 0.001$ m^2 s^{-1}
7	critical slope	as experiment 1	$K_v = 0.002$ m^2 s^{-1}
8	real topography	real T/S forcing, sea level elevation in the west	Table 1 values
9	as above	T=25.0 ^{o}C, S= 35.0 sea level elevation in the west, no IW	Table 1 values

Table 2: Parameter settings which were used in experiments 1-3, 8, and 9 (see Table 1). Sensitivity experiments were carried out with different settings for all except the critical shear stress for resuspension.

Parameter	Setting
critical shear stress for resuspension	0.1 $N\,m^{-2}$
critical shear stress for deposition	0.1 $N\,m^{-2}$
sinking velocity	$1 \times 10^{-5}\,m\,s^{-1}$
vertical background diffusivity	$2 \times 10^{-5}\,m^2\,s^{-1}$