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The role of sea water viscidity in modeling the vertical movement of internal tides

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ABSTRACT

A numerical closure scheme has been developed to introduce dissipation processes in particular for the vertical movement of internal tides. This scheme is based on the assumption that a vertically oscillating water mass disturbs the pressure field and feels the viscosity from its neighborhood at the same time. The horizontal viscosity term, referred in this paper as the internal-tide viscosity (ITV) term, is retained in the vertical movement equation, which introduces a quasi-hydrostatic assumption. Therefore, a new expression of the total perturbation pressure has been derived. By applying this expression in a $5' \times 5'$ *z*-coordinate regional ocean model, the results show great improvements. With consideration of the ITV-term, the numerically enhanced vertical movement locally near a ridge has been damped in a *z*-coordinate system, and the propagation of internal tides away from the ridge has been converted into a more reasonable dissipative mode. With the tunable parameter C_w equals to 0.2, the values of the simulated vertical velocity have been reduced to approximately 50%. And the simulated thermocline structure has been preserved, as well.

1. Introduction

From a modeling perspective, the three main factors demanded to generate internal tides: the free surface tide, the stratified sea water and the rapidly changing topography, can be satisfied by well designed three-dimensional baroclinic ocean models. However, whether the internal-tide dissipation processes can also be reproduced is doubted. In *z*-coordinate primitive equation ocean models, the problem is obvious: the vertical movement of internal tides, both at the sites of generation and during the propagation, is very often overestimated. On one hand, due to the step-like representation of the topography in a *z*-coordinate system, a rough topography looks mostly like a 'wall' rather than a sloping bottom. Therefore, to maintain the continuity of incompressible sea water, horizontally back and forth tidal currents against the 'wall' are reproduced into vertical oscillations. And the turbulent bottom layer (TBL) which should have attached to the slopping bottom disappears because of the wall-like reconstruction. Hence, the dissipation occurring in the TBL in these special areas cannot be well simulated. In tide-forcing case, all the above distortions in a *z*-coordinate system may induce enhanced vertical oscillations over rough topography, which could generate abnormally strong internal-tide signals in the stratified ocean interior.

On the other hand, in the stratified ocean interior, where the strongest internal-tide oscillation probably occurs, the energy loss of internal tides during their propagation includes multi-scale energy transfer processes, such as nonlinear wave–wave and wave–current interactions. Some of these procedures are still theoretically unclear (Fang and Du, 2005), and the corresponding numerical closure schemes are underdeveloped (Du and Fang, 1999). In primitive equation ocean models, most of the commonly applied parameterizations, to compute the coefficients of both vertical and horizontal eddy viscosity terms, are designed to solve turbulent boundary layer problems. Considering the characteristics of internal-tide oscillation, which can vertically govern the entire water column with the strongest of which occurs in the interior where the most intense density stratification locates, we argue that an important numerical closure scheme that estimates the energy loss of internal tides during their propagation is missing. In the present work, the dissipative effect of sea water viscidity in modeling the vertical movement of internal tides, both at the sites of generation and during the propagation, are discussed by introducing a quasi-hydrostatic assumption and its corresponding numerical closure scheme.

The problem mentioned above leads us to re-think about the vertical momentum equilibrium in viscous sea water, which is usually reduced to the hydrostatic assumption in primitive equation ocean models. The pressure disturbed by the vertical movement of a water mass and the neighborhood viscosity felt by the water mass are considered. The horizontal viscosity term, referred in this paper as the internal-tide viscosity (ITV) term, is retained in the

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vertical movement equation. Thus a quasi-hydrostatic assumption is suggested. Differing from the Smagorinsky (1963) scheme, a parameterization that takes the Brunt–Väisälä Frequency into consideration to emphasize the role of stratification in modeling sea water viscidity in the ocean interior has been suggested to close the ITV-term. Therefore, a new expression of the total perturbation pressure is derived. Comparing the results of two parallel numerical experiments, those with consideration of the ITV-term have been significantly improved, i.e., the modeled co-phase and co-amplitude distributions of the barotropic M_2 constituent become more realistic, the simulated vertical velocity locally near a ridge has been damped, the propagation of internal tides away from the ridge has been converted into a more reasonable dissipative mode, the thermocline structure has been well preserved, and from the residual results of the computed ITV-term that averaged over two M_2 periods, additional dissipation occurs in a kind of 'turbulent lateral boundary layer' attached to the wall-like reconstruction of the rough topography, which is functionally similar to the TBL that disappears in the z-coordinate system.

In the following sections, we will firstly discuss the ITV-term and its numerical closure scheme in detail, then a description of the parallel numerical experiments will be given in Section 3. The model results of barotropic tides, vertical velocity, as well as the vertical temperature structure along a section, are shown in Section 4. Further discussions that indicate some characteristics of the newly implemented numerical closure scheme will be presented at the end.

2. Parameterization

We focus on the pressure disturbed by the vertical movement of a water mass, which feels the viscosity from its neighborhood at the same time. First of all, we have divided the total pressure, p, into four parts, written as:

$$p = p_0 - \langle\rho\rangle gz + p' + p_w \tag{1}$$

Here p_0 is the sea level pressure (SLP), $\langle\rho\rangle$ is the reference density at a certain depth, g is the gravity acceleration, z is the upward distance calculated from sea surface, p' is the perturbation pressure determined by the stratification of sea water, which can be calculated from:

$$p' = -g\int_0^z \rho' \mathrm{d}z \tag{2}$$

where ρ' is the perturbation density.

The last term in Eq. (1), p_w, is the pressure disturbed by the vertical movement of a water mass. Therefore, the vertical momentum equilibrium in viscous sea water can be written as:

$$\frac{\partial w}{\partial t} + \left[u\frac{\partial w}{\partial x} + v\frac{\partial w}{\partial y} + w\frac{\partial w}{\partial z}\right] = -g\frac{\rho'}{\rho} - \frac{1}{\rho}\frac{\partial p'}{\partial z} - \frac{1}{\rho}\frac{\partial p_w}{\partial z} + A_w\nabla^2 w + \frac{\partial}{\partial z}\left(A_v\frac{\partial w}{\partial z}\right) \tag{3}$$

Here we neglected the molecular viscosity term. $\rho = \langle\rho\rangle + \rho'$ is the in situ density calculated from the equation of state for sea water (Gill, 1982). The last two terms, $A_w\nabla^2 w$ and $A_v\frac{\partial^2 w}{\partial z^2}$ are Reynolds stresses. In particular, $A_w\nabla^2 w$ represents the neighborhood viscidity acting on the vertically oscillating water mass. A_w is the coefficient needs to be parameterized and $\nabla^2 = \frac{\partial^2}{\partial x^2} + \frac{\partial^2}{\partial y^2}$ is the Horizontal Laplace Operator.

Here we suppose a balance reads 'the vertical acceleration induced by p_w to a water mass, should balance the neighborhood viscosity felt by the water mass'. That is:

$$0 = -\frac{1}{\rho}\frac{\partial p_w}{\partial z} + A_w\nabla^2 w \tag{4}$$

and from Eq. (2), we get

$$0 = -g\frac{\rho'}{\rho} - \frac{1}{\rho}\frac{\partial p'}{\partial z} \tag{5}$$

Altogether, Eqs. (4) and (5) indicate a vertical momentum equilibrium in viscous sea water simplified under a new quasi-hydrostatic assumption. That is:

$$0 = -g\frac{\rho'}{\rho} - \frac{1}{\rho}\frac{\partial p'}{\partial z} - \frac{1}{\rho}\frac{\partial p_w}{\partial z} + A_w\nabla^2 w \tag{6}$$

Hereafter we will refer $A_w\nabla^2 w$ in Eq. (6) as the internal-tide viscosity (ITV) term. Since the internal-tide oscillation almost disappears in the surface mixing layer, we assume $p_w|_{z=0} = 0$. Therefore, the total perturbation pressure can be computed from:

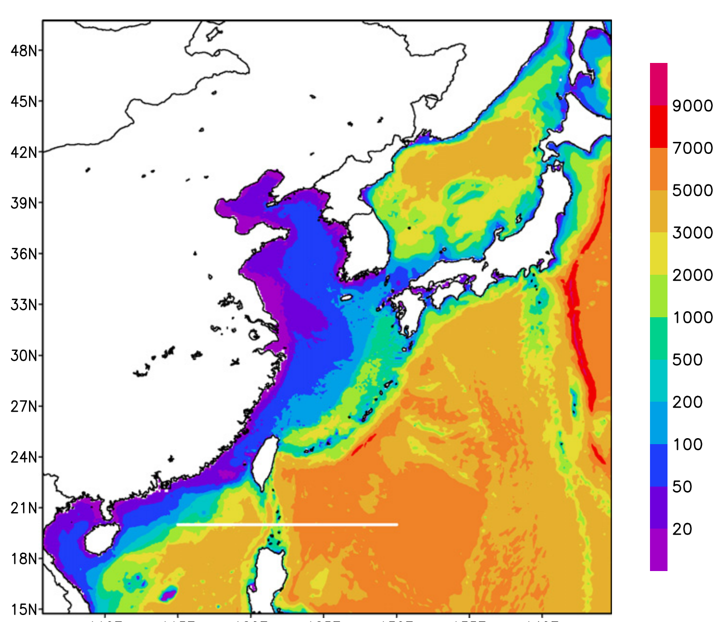

Fig. 1. The topography (m) of the model region. The thick white line indicates the position of the section discussed in this paper. X axis: latitude. Y axis: longitude.