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Generation of baroclinic tide energy in a global three-dimensional numerical model with different spatial grid resolutions

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ABSTRACT

We examine the global distribution of energy conversion rates from barotropic to baroclinic tides using a hydrostatic sigma-coordinate numerical model with a special attention to the dependence on the model grid resolution as well as the model topography resolution. A series of numerical experiments shows that the baroclinic tidal energy conversion rate increases almost exponentially with the decrease of the horizontal grid spacing, namely, from 1/5° to 1/20°. The baroclinic tidal energy conversion rates for the semidiurnal tidal constituents (M_2,S_2) are more sensitive to the horizontal grid spacing than those for the diurnal tidal constituents (K_1,O_1), reflecting the difference of their horizontal wavelengths. The sensitivity of the baroclinic tidal energy conversion rate to the horizontal grid spacing is also dependent on the generation sites of baroclinic tides; it becomes very sensitive in the regions characterized by geologically young seafloor having numerous small-scale rough topographic features such as the Mid-Atlantic Ridges, the eastern Pacific Ridges, and the Mid-Indian Ocean Ridges, whereas it is less sensitive in the regions such as the Indonesian Archipelago, and the western Pacific Ocean. The difference of the sensitivity can be best explained in terms of the value of the forcing function that is proportional to the square of the vertical velocity caused by barotropic tidal currents interacting with high-pass filtered bottom topography. Using the extrapolated value of the forcing function that takes into account all the topographic features generating baroclinic tides, we present the global distribution of the baroclinic tidal energy conversion rates in the limit of zero horizontal grid spacing.

1. Introduction

Baroclinic tides are internal gravity waves generated in a stratified ocean by the interaction of barotropic tidal currents with ocean bottom topography. When barotropic tidal currents flow over bottom topographic features, isopycnals are displaced up and down at the tidal frequencies and part of the barotropic tidal energy is converted to the baroclinic tidal energy. Analyses of satellite altimetry data (Egbert and Ray, 2000) have suggested that roughly 1 TW (1 TW = 10^{12} W) or 25–35% of the global total dissipation of the barotropic tidal energy (∼3.5 TW) is attributed to the conversion from barotropic to baroclinic tidal energy in the deep ocean around the regions of rough bottom topography.

One of the most important roles of baroclinic tides is that they transfer their energy through nonlinear wave–wave interactions to small dissipation scales inducing turbulent mixing. The turbulent mixing thus induced is thought to be in balance with the global upwelling of dense deep water to maintain the density stratification and the meridional overturning circulation (Munk and Wunsch, 1998; Webb and Suginohara, 2001). Actually, ocean general circulation models (OGCMs) (e.g. Hasumi and Suginohara, 1999; Oka and Niwa, 2013) have demonstrated that the magnitude and distribution of turbulent mixing strongly control the pattern and intensity of the global overturning circulation and the associated transport of heat and chemical tracers. Clarification of the global distribution of the baroclinic tidal energy conversion rates is, therefore, essential for an adequate parameterization of turbulent mixing as well as an improved modeling of the global overturning circulation.

Many studies have been conducted to predict the global distribution of the baroclinic tidal energy conversion rates. For example, Morozov (1995) applied a vertical/horizontal two-dimensional analytical model to all the major submarine ridges in the global ocean, but his model ignores the presence of three-dimensional

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irregularities along the bottom topography. The calculations by Jayne and St. Laurent (2001) and by Nycander (2005) are both based on a linear internal wave theory, which is valid so long as the bottom slope is subcritical (the topographic slope is less steep than the characteristic slope of radiated baroclinic tidal wave). Since these previous studies are based on simplified analytical models under idealized assumptions whose applicability to the real ocean is questionable, their derived estimates of the global baroclinic tidal energy conversion rate vary widely. Kantha and Tierney (1997) presented the global distribution of the M_2 baroclinic tide by analyzing the TOPEX/Poseidon altimetric data. The altimetric observation, however, has the serious problem that only a fraction of the baroclinic tide energy can be estimated, because high vertical modes have very weak sea surface expressions. For the direct calculation of the baroclinic tidal energy conversion rate, therefore, numerical simulations based on the primitive Navier–Stokes equations are most desirable.

Simmons et al. (2004) performed the first global numerical simulation of baroclinic tides to attain the global estimate of the baroclinic tidal energy conversion rate, although the grid resolutions of their model (1/8° horizontal grid spacing) are too coarse to resolve the short-wavelength baroclinic tides. Actually, very high-resolution regional models (horizontal grid spacing less than 1/100°) incorporating fine-scale bathymetric data obtained from multi-beam echo soundings (Robertson, 2006; Carter et al., 2008; Zilberman et al., 2009; Iwamae and Hibiya, 2012) have indicated that the information of fine-scale bottom topography of the order of 1 km is important for an accurate estimate of the baroclinic tidal energy conversion rate. Considering insufficient computer capacity as well as limitation of the multibeam bathymetric data at the present stage, we have to bridge the gap between the results of relatively coarse-resolution global models and those that would be obtained by very high-resolution numerical models.

In order to fill this gap, Niwa and Hibiya (2011) carried out a series of global numerical simulations using horizontal grid spacing ranging from 1/15° to 1/5° to find that the calculated baroclinic tidal energy conversion rate integrated over the global ocean increases exponentially as the horizontal grid spacing is decreased. This empirical relationship was then extrapolated to the limit of zero grid spacing $\delta x = 0°$ to estimate the global amount of the baroclinic tide energy available for the deep ocean mixing. However, Niwa and Hibiya (2011) did not show how the global distribution of the baroclinic tidal energy conversion rates varies with the increase of the model grid resolution as well as the model topography resolution.

In the present study, a similar series of the global numerical simulations is carried out to investigate the sensitivity of the global distribution of the baroclinic tidal energy conversion rates to the horizontal grid spacing ranging from 1/20° to 1/5°. It is found that the difference in the sensitivity can be best explained in terms of the value of a forcing function that is proportional to the square of the vertical velocity caused by barotropic tidal currents interacting with high-pass filtered bottom topography. Then using the extrapolated value of the forcing function that takes into account all the topographic features generating baroclinic tides, we present the global distribution of the baroclinic tidal energy conversion rates in the limit of zero horizontal grid spacing.

2. Numerical simulation

Fig. 1 shows the model domain that covers the global ocean from 80°S to 80°N, excluding the central part of the Arctic Ocean and the shallow shelf regions near the Antarctic continent.

In the present study, numerical simulations are carried out using the Princeton Ocean Model (Blumberg and Mellor, 1987) to solve the three-dimensional, free-surface primitive equations under the hydrostatic and Boussinesq approximations. The model uses a terrain-following, sigma-coordinate defined by $\sigma = \frac{z-\eta}{D}$ with z a vertical Cartesian coordinate positive upward, and D the total water depth ($D \equiv H + \eta$ where H is the time-mean water depth and η the perturbation of sea surface elevation). The sigma-coordinate thus ranges from $\sigma = 0$ at the surface ($z = \eta$) down to $\sigma = -1$ at the bottom ($z = -H$). The governing equations are then given by

$$\frac{\partial UD}{\partial t} + \frac{\partial UUD}{\partial x} + \frac{\partial UVD}{\partial y} + \frac{\partial U\Omega}{\partial \sigma} = +fVD - \frac{D}{\bar{\rho}_0}\frac{\partial P'}{\partial x} + \frac{\sigma}{\bar{\rho}_0}\frac{\partial P'}{\partial \sigma}\frac{\partial D}{\partial x} - \alpha g D\frac{\partial \eta}{\partial x} + \beta g D\frac{\partial \xi}{\partial x} - r(U - \bar{U})D + F_U \quad (1)$$

$$\frac{\partial VD}{\partial t} + \frac{\partial VUD}{\partial x} + \frac{\partial VVD}{\partial y} + \frac{\partial V\Omega}{\partial \sigma} = -fUD - \frac{D}{\bar{\rho}_0}\frac{\partial P'}{\partial y} + \frac{\sigma}{\bar{\rho}_0}\frac{\partial P'}{\partial \sigma}\frac{\partial D}{\partial y} - \alpha g D\frac{\partial \eta}{\partial y} + \beta g D\frac{\partial \xi}{\partial y} - r(V - \bar{V})D + F_V \quad (2)$$

$$\frac{\partial P'}{\partial \sigma} = -gD\rho' \quad (3)$$

$$\frac{\partial UD}{\partial x} + \frac{\partial VD}{\partial y} + \frac{\partial \Omega}{\partial \sigma} + \frac{\partial \eta}{\partial t} = 0 \quad (4)$$

$$\frac{\partial T'D}{\partial t} + \frac{\partial T'UD}{\partial x} + \frac{\partial T'VD}{\partial y} + \frac{\partial T'\Omega}{\partial \sigma} = -\frac{\partial T_0 D}{\partial t} - \frac{\partial T_0 UD}{\partial x} - \frac{\partial T_0 VD}{\partial y} - \frac{\partial T_0 \Omega}{\partial \sigma} - rT'D + F_{T'} \quad (5)$$

$$\frac{\partial S'D}{\partial t} + \frac{\partial S'UD}{\partial x} + \frac{\partial S'VD}{\partial y} + \frac{\partial S'\Omega}{\partial \sigma} = -\frac{\partial S_0 D}{\partial t} - \frac{\partial S_0 UD}{\partial x} - \frac{\partial S_0 VD}{\partial y} - \frac{\partial S_0 \Omega}{\partial \sigma} - rS'D + F_{S'} \quad (6)$$

where t is time; x and y are horizontal coordinates positive eastward and northward, respectively; U and V are the velocity components in the x and y directions, respectively; Ω is the velocity component normal to the σ = constant surfaces, which is related to the Cartesian vertical velocity W via $W = \Omega + U(\sigma\frac{\partial D}{\partial x} + \frac{\partial \eta}{\partial x}) + V\left(\sigma\frac{\partial D}{\partial y} + \frac{\partial \eta}{\partial y}\right) + \sigma\frac{\partial D}{\partial t} + \frac{\partial \eta}{\partial t}$; $\bar{U}$ and $\bar{V}$ are the depth-averaged components of U and V, respectively, such that $(\bar{U}, \bar{V}) = \frac{1}{D}\left(\int_{-H}^{\eta} Udz, \int_{-H}^{\eta} Vdz\right)$; $f = 2\omega_{earth}\sin\theta$ is the Coriolis parameter with ω_{earth} the rotation frequency of the earth and θ the latitude; T' and S' are the deviations of the potential temperature and salinity from their background basic values T_0 and S_0, respectively; ρ' is the density deviation defined by $\rho' = \rho - \rho_0$ where ρ is the sea water density determined from the potential temperature ($T = T' + T_0$) and salinity ($S = S' + S_0$) using the equation of state and ρ_0 is the background density field determined from T_0 and S_0; $\overline{\rho_0}$ is the constant reference density; P' is the pressure perturbation associated with the density deviation ρ'; g is the acceleration due to gravity; the factor α accounts for the effect of load tides and is assumed to be 0.9 following Ray and Cartwright (1998); ξ is the equilibrium tidal potential; the factor β multiplying ξ is the effective earth elasticity assumed to be 0.69 following Kantha (1995); (F_U, F_V) and $(F_{T'}, F_{S'})$ represent the viscosity and diffusivity terms, respectively.

The horizontal eddy viscosity and diffusivity coefficients are determined following the formulation of Smagorinsky (1963), whereas the vertical eddy viscosity and diffusivity coefficients are parameterized following the Richardson number formulation of Pacanowski and Philander (1981). At the lowest sigma level, bottom friction is applied through a quadratic friction law with a constant drag coefficient assumed to be 0.0025. Furthermore, taking into account the decay of propagating baroclinic tides due to their