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Tidal effect on the dense water discharge, Part 2: A numerical study ☆

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ABSTRACT

In Part 1 of this two-part paper, an analytical model is presented to examine the tidal effect on the dense water discharge. It is hypothesized that the tide-induced shear dispersion would augment the benthic-layer thickness to significantly enhance the spread of dense water on the shelf and its descent down the continental slope. Here in Part 2, we carry out a numerical process study to assess the analytical model and to aid the interpretation of the observed phenomenon in the western Ross Sea.

In this process study, we conduct numerical experiments for both passive and active tracer, and demonstrate the sharp contrast between the tidal and non-tidal cases whether on a flat or sloping bottom. In particular, on a slope as steep as the western Ross Sea, the model result from the non-tidal case shows that the dense water cannot descend much beyond the shelf break due solely to Ekman advection. When tides are included, however, the dense benthic layer would span several times the Ekman depth, which reduces the diabatic mixing across the density interface that would otherwise dilute the density anomaly, hence allowing the dense water to be more efficiently propelled by the Ekman flow and tidal diffusion. The model results are consistent with the analytical model, and also corroborated by the observations from AnSlope.

1. Introduction

The High-Salinity Shelf Water (HSSW) is formed on the Antarctic shelves, and as it descends down the continental slope, it entrains the ambient less-dense Circumpolar Deep Water (CDW), and forms a benthic outflow into the adjacent deep ocean, providing the kernel for Antarctica Bottom Water (AABW) formation.

Since the western Ross Sea is important for the AABW formation, an NSF-funded project AnSlope was carried out there to address the physical processes on the Antarctic slope. CTD measurements were taken near Drygalski Trough, which showed that the dense water on the shelf descends all the way down to the lower slope in the benthic layer of 100–250 m thick. This thickness is several times the Ekman depth at that latitude, and hence cannot be due solely to the Ekman dynamics. The mean temperatures of the HSSW and the CDW are -1.6 and $1.2\,^{\circ}$C, respectively, corresponding to a density difference of $0.4\,\mathrm{kg\,m^{-3}}$.

Several mechanisms have been proposed for the descent of the HSSW. One prevailing idea is that the Ekman flow generated by the bottom friction would propel the dense water down the slope (Baines and Condie, 1998; Nagata et al., 1993; Condie, 1995; Shapiro and Hill, 1997). Other mechanisms also have been proposed, including channeling by the submarine canyons (Chapman and Gawarkiewicz, 1995; Jiang and Garwood, 1995), and the transport by eddies (Jiang and Garwood, 1996; Gawarkiewicz and Chapman, 1995; Kikuchi et al., 1999), as both would break the rotational constraint on the down-slope motion.

The previous efforts to model the dense water discharge down the continental slope have sometimes adopted the streamtube approach (Smith, 1975), which considered only the bulk properties, but not the internal structure of the plume (Price and O'Neil Baringer, 1994). More sophisticated two- and three-dimensional numerical models also have been applied in the studies of dense water discharge. Jungclaus and Backhaus (1994) used a two-dimensional reduced-gravity model to study the topographic effect on the Denmark Strait Overflow. Ezer and Weatherly (1990) employed both a second-order turbulence closure scheme and an eddy-viscosity scheme in their two-dimensional z-coordinate primitive-equation model, and showed that the horizontal and vertical scales of the down-slope cold tongue are governed by the Ekman veering and Ekman depth, respectively. Moreover, complicated 3-D numerical models with different coordinate systems have been used to study the problem as well (Jiang and Garwood, 1995, 1996; Jungclaus and Mellor, 2000; Ezer and Mellor, 2000; Ezer, 2005, 2006). These models could resolve the instability features of the dense benthic layer such as eddies and sub-plumes, shedding further light on its dynamics.

Nevertheless, most of the previous numerical models (with the exception of Padman et al., 2009) have neglected the tides, an important feature in the AnSlope observations. The effectiveness

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of the tidal shear dispersion in passive tracer transport (Okubo, 1967), together with the prominent tidal current in the benthic layer as observed in the AnSlope (Gordon et al., 2004), motivated our process study. We carry out numerical experiments to elucidate the tidal rectified effect based on analytical study in Part 1 of this paper (Ou et al., 2009, referred henceforth as Part 1) and, specifically, to evaluate our hypothesis that the tides could augment the dense layer thickness and significantly enhance its descent down the slope.

Previous three-dimensional numerical studies have shown that eddy formation through baroclinic instability might be important for the dense water discharge, among others that would break the rotational constraint. The AnSlope observations have not indicated the prevalence of eddies in the benthic layer, since the descent angle of the dense water is relatively stable (Gordon et al., 2004). We do not exclude the possibly important role of eddy formation, but our purpose of this process study is to examine whether the tidal effect may significantly enhance the dense water descent on the slope. And since the proposed tidal effects are fully operative in the two-dimensional cross-shore plane, we employ in this paper, a two-dimensional configuration of a hydrostatic z-coordinate primitive-equation model. The tidal effect is demonstrated by the contrast between simulations with and without introducing tides into the model domain, which should be sufficient to evaluate our hypothesis in Part 1.

There have been many previous studies on the advantages and drawbacks of numerical models with different resolutions and vertical coordinates in simulating the dense water discharge. z-Coordinate models are not particularly favored against terrain and isopycnal-following models in the capability of simulating sufficient down-slope penetration of the dense plume, but if the vertical resolution is fine enough to resolve the benthic layer and the horizontal grid does not exceed its vertical grid divided by the maximum slope (Winton et al., 1998; Legg et al., 2006), z-coordinate models work well in propelling the down-slope transport. Ezer and Mellor (2004) also showed that with a 2.5 km horizontal resolution, the bottom boundary layer structure starts to converge to the observed overflows. z-Coordinate system is preferred in our simulations due to the steep bottom slope in AnSlope observations (0.06), and we adopt very fine vertical and horizontal resolutions that satisfy the above criteria to guarantee the effectiveness in simulating the dense water descent.

2. Model description

The numerical model we use to explore the dense water discharge is a primitive-equation, free-surface, coastal ocean model originally developed by Wang (1982, 1985). It has been applied to different coastal studies, such as the simulations of the upwelling front off the northern California coast (Chen and Wang, 1990), the tidal front in the Celtic Sea (Wang et al., 1990), and the internal tides in a coastal frontal zone (Chen et al., 2003), among others. In this model configuration, we use the Mellor and Yamada (1982) level 2.5 turbulence closure for vertical mixing, which has been well tested and discussed in many numerical studies of dense plume dynamics (Jiang and Garwood, 1995, 1996; Jungclaus and Mellor, 2000; Ezer and Mellor, 2000), and Smagorinsky's (1963) parameterization for horizontal mixing, which is essential for minimizing nonphysical mixing and maintaining a sharp front in the model (Chen et al., 2003). As we pointed out before, the model is configured on a two-dimensional cross-shore section, since our purpose is to isolate and elucidate the tidal effect, but not to simulate unstable eddies.

The resolution is quite high, with horizontal and vertical grid spacings of 600 and 10 m, respectively, in the sloping bottom cases, which is adequate to resolve the Ekman layer about 30 m. Tides are imposed through the oscillation of the surface elevation at the offshore boundary of the model domain to produce a cross-shore tidal current amplitude of $20\,\mathrm{cm\,s^{-1}}$, which corresponds to a tidal excursion distance around 5 km for diurnal tides. For all the experiments below (except for the semi-diurnal case) the model domain is placed at 65 °N, representative of the high-latitude coastal oceans, where the semi-diurnal tides approach the inertial frequency, so the domain is placed at mid-latitude only for the semi-diurnal case hereinafter. At the inshore boundary, the temperature is fixed at $-2\,^{\circ}\mathrm{C}$ with zero normal-gradient; the velocity satisfies the Orlanski open boundary condition. At the offshore boundary, Orlanski condition is imposed on temperature, while velocity has zero normal-gradient. The quadratic law friction is applied to the bottom boundary, with a drag coefficient of 0.0025.

Since the temperature effect dominates the density variation, we neglect for simplicity the salinity difference. The traditional "dam-break" technique is adopted: a vertical barrier that separates two water columns with different temperatures on the shelf is removed at time $t = 0$ to allow the density front to adjust. This technique has been widely used in numerical studies of coastal and continental shelf circulation in shallow seas (Wang, 1984; Oey and Mellor, 1993; Jiang and Garwood, 1995). The temperatures of the two water columns are -2 and $1\,^{\circ}\mathrm{C}$, respectively, which results in a sigma-t difference of $0.3\,\mathrm{kg\,m^{-3}}$, close to the observations, and a baroclinic deformation radius of about 8 km.

To evaluate the analytical model results in Part 1, we first treat temperature as a passive tracer to estimate how the benthic-layer properties change with varying forcing frequency in Section 3, in which both the bottom slope and density variation are excluded. We then include the density effect in Section 4 and bottom slope in Section 5 to address the additional physics which are important for the dense water discharge in the oceanic settings. In Sections 4 and 5, a surface heat loss of $100\,\mathrm{W\,m^{-2}}$ is imposed on a coastal strip of 20 km to simulate cooling in a coastal polynya.

3. Passive tracer on flat bottom

In Part 1, we extended the well-known physics of tidal shear dispersion to the genesis of the benthic layer of passive tracer, whose depth turns out to be several times the Ekman depth. We showed that apart from the singularity at the inertial frequency, both the benthic depth h_b (non-dimensionalized by Ekman depth) and the mean tidal diffusivity within the benthic layer k (non-dimensionalized by u'^2/f where u' is the tidal amplitude) generally increase with decreasing forcing frequency, as shown by the thick solid and dashed curves in Fig. 2.

Numerical experiments are carried out to validate the non-dimensionalization and to assess the results of h_b and k with varying forcing frequencies in the analytical model. As described in Section 2, we set an initial vertical front separating two homogeneous regions with different temperatures -2 and $1\,^{\circ}\mathrm{C}$. Since temperature is treated as a passive tracer by removing the temperature dependence from the equation of state, the two water masses have no density difference. A weak background stratification is nonetheless imposed to curtail the unphysical vertical mixing, and we set the upper bound of vertical diffusivity coefficient υ_{max} to be $500\,\mathrm{cm^2\,s^{-1}}$, and the background υ to be $10\,\mathrm{cm^2\,s^{-1}}$ unless otherwise stated.

The front would remain intact until we introduce the oscillatory forcing, which broadens the front near the bottom except for the inertial frequency, indicating a locally significant tidal diffusivity as predicted by Part 1. The benthic-layer depth h_b