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On the role of bottom roughness in overflows

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Abstract

Overflows play an important role in the downwelling limb of the oceanic thermohaline circulation. In light of the recognition that some overflows are not homogenous but exhibit a vertical density structure, and details of topography influence the pathways of some overflows, the impact of topographic roughness on the product property distribution is explored using the 3D non-hydrostatic spectral element model Nek5000. Numerical experiments are carried out by varying bottom roughness amplitude and ambient stratification parameters, in a regime where equilibrated product water masses are formed in a non-rotating environment.

Our main finding is that bottom roughness can influence the overflow product distribution such that the highest salinity classes are removed and neutral buoyancy level is attained higher up in the stratified ambient water column. It is also shown that the form drag coefficients in overflows over rough bottom can be much larger than the skin drag coefficient over smooth bottom. To our knowledge, form drag has never been measured in oceanic overflows. As such, these numerical experiments imply that such measurements would be useful for a better understanding of overflow dynamics. It is also found that the ratio of source and product overflow mass transports is robust to changes in bottom roughness. This appears to happen because the distribution of entrainment is totally different in the case of rough bottom. Entrainment tends to initiate earlier (due to vertical motion induced by topography) and terminate earlier (due to development of form drag) than that over smooth topography.

1. Introduction

The thermohaline circulation in the ocean can be broadly divided into three components. The first component is the deep-water mass formation, or downwelling. It is generally accepted that deep-water mass formation takes place in polar and marginal seas by convection (Spall and Pickard, 2001). The resulting dense water masses are introduced into the circulation via localized processes, such as overflows (Price and Baringer, 1994) that feed the bottom (e.g., Antarctic Bottom Water) and deep (e.g., North Atlantic Deep Water) water masses in the ocean. The second component is the transport of these water masses inside the various basins, not only through large-scale gyres, jets and meso-scale eddies, but also by flows through gaps and passages as some of

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the dense flows are in direct contact with bottom topography (Schmitz, 1995). The third component is the upwelling of bottom and deep water masses to complete the circulation.

Prediction of the Earth's future climate conditions necessarily relies on the accuracy of global circulation models, and it is known that numerical simulation of the ocean's thermohaline circulation is a central problem regarding this task. Thus, ocean general circulation models (OGCMs) must be constantly improved to capture these three components as realistically as possible. Clearly, ocean observations are critical to achieve this objective. The main challenge regarding observations is the large separation of scales in the ocean, or more formally the high Reynolds number $Re = UL/v$. Using a characteristic speed scale $U \approx 10^{-1}$ m s^{-1}, horizontal length scale $L \approx 10^5$ m and kinematic viscosity $v = 10^{-6}$ m^2 s^{-1}, an upper limit can be obtained as $Re \approx 10^{10}$. Since the number of degrees of freedom in the case of homogenous, isotropic turbulence is $\sim Re^{9/4}$ (Lesieur, 1997), we obtain $\approx O(10^{22})$ as an upper limit for the needed spatial sampling points, at a given time. Therefore, it is difficult at the present time to collect observations describing the state of the oceanic velocity and tracer fields in all spatial and temporal coordinates. The sampling problem is particularly pronounced regarding the observation of the upwelling limb of the thermohaline circulation in that it is not clearly known neither by which mechanisms this takes place (Munk and Wunsch, 1998; Wunsch and Ferrari, 2004), nor exactly where the upwelling takes place. Ocean observations point to the role played by rough topography of the oceanic ridges (Poltzin et al., 1997; Ferron et al., 1998; Ledwell et al., 2000), isolated topographic obstacles such as seamounts (Nabatov and Ozmidov, 1988; Gibson et al., 1993; Lueck and Mudge, 1997; Kunze and Toole, 1997; Lavelle et al., 2004), mixing over the continental slope (Moum et al., 2002), and inside the canyons of mid-ocean ridges (Thurnherr and Speer, 2003; Thurnherr et al., 2004; Thurnherr, 2006). Given the large number of such topographic features, upwelling appears to be widely distributed in space, and further observations are needed to map the spatial and temporal distribution of ocean upwelling.

Regarding the downwelling limb of the thermohaline circulation, both the observation and modeling of the deep-convection process remain challenging due to the small spatial and temporal scales of mixed patches (Marshall and Schott, 1999). Nevertheless, the state of knowledge is significantly better than the upwelling limb of the thermohaline circulation, because most of the deep and intermediate water masses are released into the general circulation from a few main overflows, namely those from the Mediterranean Sea, Denmark Strait, Faroe Bank Channel, Red Sea and the Antarctic (Warren, 1981; Price and Baringer, 1994). Subsequently, many aspects of these overflows have been explored as a result of observational programs (Baringer and Price, 1997; Girton and Käse, 2001; Girton and Sanford, 2003; Gordon et al., 2004; Peters et al., 2005; Peters and Johns, 2005).

The dynamical model for overflows based on the initial observations, namely the Mediterranean Sea and the Denmark Strait overflows, is so-called stream-tube model (Smith, 1975; Killworth, 1977; Price and Baringer, 1994). In this model, the overflow is considered as a homogeneous entity in which the gravitational force is balanced by shear stress at the bottom and entrainment stress at the top, and the Earth's rotation acts to steer the overflow along topographic contours. Since the overflow transport increases between the source and the location of neutral buoyancy, entrainment of ambient fluid is considered as the critical process in the dynamics of all overflows (Price and Baringer, 1994). Much of our present understanding of entrainment in overflows is based on laboratory experiments of bottom gravity currents (Ellison and Turner, 1959; Simpson, 1987; Cenedese et al., 2004; Baines, 2001, 2005). Recent advances in computer power and computational techniques also make possible another approach, so-called large-eddy simulation, in which the turbulent coherent structures that carry out most of the mixing, namely Kelvin–Helmholtz billows and secondary shear instabilities, are explicitly computed (Özgökmen and Chassignet, 2002; Özgökmen et al., 2003, 2004a,b, 2006, 2007). Generally speaking, the bottom drag has been traditionally viewed as a simple frictional mechanism and most of the attention has been paid to the process of entrainment. One of the main contributions of the laboratory and high-resolution numerical experiments is the development of simple representations of entrainment in gravity currents in relation to the gradient Richardson number, $Ri = N^2/S^2$ the ratio of the square of buoyancy frequency and vertical shear, or even simpler, to the bottom slope angle.

The importance of the representation of overflows in OGCMs led to a significant collaborative effort and formation of so-called climate process team on gravity current entrainment (`http://cpt-gce.org/index.htm`). In the context of this collaboration, a number of idealized gravity current experiments (Chang et al., 2005; Ezer, 2005; Legg et al., 2006; Xu et al., 2006) and regional simulations in comparison