ISENTROPIC MIXING AND FORMATION OF THE ANTARCTIC BOTTOM WATER

Kenzo TAKANO

Institute of Biological Sciences, University of Tsukuba, Sakura-mura, Niihari-gun, Ibaraki 305

Abstract: A possible effect of the isentropic mixing on the formation of the Antarctic Bottom Water is explored by using a numerical model of oceanic general circulation. The isentropic mixing increases the rate of deep and bottom water mass formation. The meridional circulation is waxed. The meridional heat transport is correspondingly increased.

1. Introduction

The meridional heat transport by the ocean is one of the key factors of climate. In most of the global ocean models so far developed, the meridional heat transport is much smaller than the recent estimates based on satellite and radiosonde observations (VONDER HAAR and OORT, 1973; OORT and VONDER HARR, 1976). The meridional heat transport has three components; transport by the horizontal circulation, transport by the meridional (vertical) circulation and transport by the eddy diffusion. The first is achieved by bringing warm western boundary waters poleward and relatively cold waters equatorward in the central and eastern portions of the ocean basins. The meridional circulation gives rise to a net poleward heat transport by bringing warm surface waters poleward and cold deep waters equatorward. The eddy diffusion brings heat from warmer regions to colder regions.

The previous results with the numerical models show that the heat transport by the meridional circulation is of the greatest importance. An analysis of hydrographic data also suggests it (BRYAN, 1962). The meridional circulation is driven mainly by sinking of dense surface waters in the northern region of the North Atlantic Ocean and in the Antarctic Ocean. Therefore, sinking in the Antarctic Ocean is an important issue to be understood before sophisticated ocean-atmosphere coupled models are used for climate research.

The estimates by observations are still crude. If the numerical models really yield too a small meridional heat transport, it might be partially attributed to a wrong simulation of the meridional circulation, which in turn might be attributed to a wrong simulation of the deep and bottom water formation in the Antarctic Ocean. The formation of the Antarctic Bottom Water depends on various processes such as water freezing, ice melting, surface cooling, precipitation, surface layer mixing. The mesoscale eddies might come into the picture, too. All these processes are not yet well understood to be correctly imbedded into the numerical models. Leaving aside them here, the present study is concerned with a somewhat different aspect: to what extent the isentropic mixing affects the bottom water formation and the intensity of the meridional circulation, and finally the meridional heat transport.

Although it has been known that the mixing is strong along the isopycnal surfaces and very weak across them, the numerical models have conventionally used, as pointed out by SOLOMON (1971), the horizontal and vertical mixing, regardless of the slope of the isopycnal surfaces. Simple numerical experiments are done to have some insight into the effect of the isentropic mixing.

2. Governing Equations

With conventional symbols, the momentum equations are given by

$$\frac{\partial u}{\partial t} = -\frac{1}{\rho R \cos\varphi}\frac{\partial p}{\partial \lambda} + fv + \mathrm{A}_m \nabla^2 u + \mathrm{k}_m \frac{\partial^2 u}{\partial z^2}\,,$$

$$\frac{\partial v}{\partial t} = -\frac{1}{\rho R}\frac{\partial p}{\partial \varphi} - fu + \mathrm{A}_m \nabla^2 v + \mathrm{k}_m \frac{\partial^2 v}{\partial z^2}\,,$$

$$o = -\frac{1}{\rho}\frac{\partial p}{\partial z} - g\,,$$

$$\nabla^2 = \frac{1}{R^2 \cos\varphi}\left\{\frac{1}{\cos\varphi}\frac{\partial^2}{\partial \lambda^2} + \frac{\partial}{\partial \varphi}\left(\cos\varphi \frac{\partial}{\partial \varphi}\right)\right\},$$

where R is the radius of the earth, z is the vertical coordinate positive upward, A_m the coefficient of horizontal eddy viscosity and k_m the coefficient of vertical eddy viscosity.

The momentum advection is neglected, because, as will be mentioned later, the grid size, 2° in both longitude and latitude, is too large for the momentum advection to be significant.

The continuity equation is

$$\frac{1}{R\cos\varphi}\left\{\frac{\partial u}{\partial \lambda} + \frac{\partial}{\partial \varphi}(v\cos\varphi)\right\} + \frac{\partial w}{\partial z} = 0\,.$$

The thermal equation is

$$\frac{\partial \mathrm{T}}{\partial t} = -\frac{1}{R\cos\varphi}\left\{\frac{\partial}{\partial \lambda}(u\mathrm{T}) + \frac{\partial}{\partial \varphi}(v\mathrm{T}\cos\varphi)\right\} - \frac{\partial}{\partial z}(w\mathrm{T}) + Q\,. \quad (1)$$

In upper layers Q is given by

$$Q=A_h\nabla^2T+\frac{k_h}{\delta}\frac{\partial^2 T}{\partial z^2}\,. \tag{2}$$

Here A_h is the coefficient of horizontal eddy diffusivity, k_h the coefficient of vertical eddy diffusivity, and the factor δ for the convective adjustment is defined by

$$\delta=\begin{cases}1 & \text{for } \dfrac{\partial\rho}{\partial z}<0\\[2ex] 0 & \text{for } \dfrac{\partial\rho}{\partial z}\geqq 0\,,\end{cases}$$

so that the stable stratification is restored by strong vertical mixing whenever the unstable stratification takes place. A formula by FRIEDRICH and LEVITUS (1972) is used for calculating the density ρ. In lower layers where the isentropic mixing is specified, Q is given by

$$Q=A_h\nabla_1{}^2T+k_h\frac{\partial^2 T}{\partial \zeta^2}+\delta_c\,, \tag{3}$$

where $\nabla_1{}^2$ and ζ refer to the isopycnal surface and δ_c represents the convective adjustment.

The equation for salinity S is readily obtained provided that the temperature T in eqs. (1) to (3) is replaced by the salinity S.

The model ocean is bounded by two parallels 70° apart and two meridians 48° apart. The northern boundary is the equator. Symmetry is assumed with respect to the northern boundary. The depth is constant (4000 m).

The conditions at the surface are

$$\rho k_m\frac{\partial u}{\partial z}=\tau_\lambda\,,$$

$$\rho k_m\frac{\partial v}{\partial z}=0\,,$$

$$c\rho k_h\frac{\partial T}{\partial z}=d(T_A-T_S)\,,$$

$$k_h\frac{\partial S}{\partial z}=S(E-Pr)\,,$$

$$w=0\,.$$

The external forcing, with eastward component of the surface wind stress τ_λ, difference between the evaporation and precipitation (E-Pr), and reference atmospheric temperature T_A, is steady and variable with latitude only. The surface water temperature is denoted by T_S.

There is no friction along the bottom and the northern boundary, no slip along the western, southern and eastern boundaries. There is neither heat flux nor salinity flux through the bottom and the lateral boundary.

3. Solutions

The grid spacing is 2° in longitude and latitude. Five levels are set up in the vertical. The horizontal velocity, temperature and salinity are calculated at depths of 20, 120, 640, 1280 and 2760 m, while the vertical component of the velocity is calculated at depths of 70, 380, 960 and 2020 m. The eddy coefficients are assumed as follows,

$$A_m = 2 \times 10^8 \, cm^2 s^{-1} ,$$
$$A_h = 2.5 \times 10^7 \, cm^2 s^{-1} ,$$
$$k_m = k_h = 1.5 \, cm^2 s^{-1} .$$

The time and space differencings are the same as those described in a previous note (TAKANO, 1974).

Neglect of the momentum advection decouples the barotropic component from the baroclinic component of the velocity. To begin with, therefore, the barotropic component is calculated by numerical integration of the vorticity equation for about 5 years. Then, the time integration of the momentum, heat and salinity equations is forwarded to get the baroclinic component and the temperature and salinity field. The time step is 8 hours in both phases of integration.

Two experiments are carried out. One is the case of the conventional horizontal and vertical mixing and the other is the case of the isentropic mixing. In the second case, the conventional horizontal and vertical mixing is specified in the uppermost layer 80 m thick and the isentropic mixing is specified in deeper layers. The final state of the first case is used as the initial case for the second case. The simulated time is about 200 years in the first case and about 100 years in the second case.

The meridional circulation defined by the transport stream function waxes in the second case. Its maximum value, considered as a deep and bottom water mass formation rate, is 40×10^{12} cm^{3}s^{-1} compared with 27×10^{12} cm^{3}s^{-1} in the first case. The meridional heat transport is correspondingly increased from 16×10^{13} cal s^{-1} in the first case to 21×10^{13} cal s^{-1}. This increase is mostly due to the increase of the heat transport by the meridional circulation.

The isentropic mixing in the present model does work everywhere except in a shallow surface layer. This should be unrealistic and would overestimate the effect of the isentropic mixing on the meridional heat transport. However, its importance could not be minimized as compared with other processes relating to the deep and bottom water mass formation.

References

BRYAN, K. (1962): Measurements of meridional heat transport by ocean currents. J. Geophys. Res., **67**, 3403–3414.

FRIEDRICH, H. and LEVITUS, S. (1972): An approximation to the equation of state for sea water, suitable for numerical models. J. Phys. Oceanogr., **2**, 514–517.

OORT, A. H. and VONDER HAAR, T. H. (1976): On the observed annual cycle in the ocean-atmosphere heat balance over the Northern Hemisphere. J. Phys. Oceanogr., **6**, 781–800.

SOLOMON, H. (1971): On the representation of isentropic mixing in ocean circulation models. J. Phys. Oceanogr., **1**, 233–234.

TAKANO, K. (1974): A general circulation model for the world ocean. Tech. Rept. No. 8, Numerical Simulation of Weather and Climate. Los Angeles, Dept. Meteorol., Univ. California, 46 p.

VONDER HAAR, T. H. and OORT A. H. (1973): New estimate of annual poleward energy transport by Northern Hemisphere oceans. J. Phys. Oceanogr., **3**, 169–172.

(Received April 10, 1981)