Bull. Nov. Comp. Center, Num. Model. in Atmosph., etc., 6 (2000), 73–78

On the deep water formation in the World Ocean model*

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This paper reports the results from coarse-resolution World Ocean numerical experiments on simulation of the mechanism of the deep water formations in high latitudes of the North Atlantic and in the Southern Ocean. The global large scale geostrophic ocean circulation model with 5 degree resolution on latitude and longitude and 24 levels in the vertical coordinate is used. The surface temperature, salinity, and wind stress with annual cycle are specified from the climatological observed data. In this work, the model of global circulation of the World Ocean from [1] is added by the parameterization of the Mediterranean high saline water and including the corrections of the seasonal change surface temperature, salinity as in [2–4].

In previous works [1], a series of numerical experiments with a three-dimensional numerical model of a global ocean climate with real topography and different vertical resolution was described. The temperature of the deep ocean at course resolution has appeared higher, than it should be. With increase resolution on the vertical coordinate the temperature of the deep ocean and the average temperature was near the observed data. Unfortunately, the increase of resolution does not improve the salinity field. In [2–4], it was shown that the feature of realistic water masses depends on the convection in the polar and subpolar regions, where these water masses are formed. In these regions, the largest density can be formed only under enough cold and saline values of the surface data. In [2–4], these surface temperature and salinity data from Levitus [5] manually adjust to obtain the largest density.

Let us consider a problem of formation of large-scale climatic temperature, salinity, and currents fields in the World Ocean including Arctic Ocean with real geometry and bottom relief. Consider an initial system of equations where the hydrostatic, Boussinesq's, and rigid lid approximation is employed and nonlinear terms in the equations of motions are omitted [1, 6]:

$$R_1 u + \ell v = -\frac{1}{a\rho_0 \sin\theta}\frac{\partial P}{\partial \lambda} + \frac{\partial}{\partial z}\nu\frac{\partial u}{\partial z}, \tag{1}$$

$$-\ell u + R_1 v = -\frac{1}{a\rho_0}\frac{\partial P}{\partial \theta} + \frac{\partial}{\partial z}\nu\frac{\partial v}{\partial z}, \tag{2}$$

*Supported by the Russian Foundation for Basic Research under Grant 00-15-98543.

$$\frac{1}{a\sin\theta}\left(\frac{\partial u}{\partial\lambda}+\frac{\partial v\sin\theta}{\partial\theta}\right)+\frac{\partial w}{\partial z}=0, \tag{3}$$

$$P=-g\rho_0\zeta+g\int_0^z\rho\,dz, \tag{4}$$

$$\frac{\partial\Phi}{\partial t}+\frac{u}{a\sin\theta}\frac{\partial\Phi}{\partial\lambda}+\frac{v}{a}\frac{\partial\Phi}{\partial\theta}+w\frac{\partial\Phi}{\partial z}=\frac{\partial}{\partial z}æ\frac{\partial\Phi}{\partial z}+\frac{\mu}{a^2}\Delta\Phi, \tag{5}$$

where

$$\Delta\Phi=\frac{1}{\sin^2\theta}\frac{\partial^2\Phi}{\partial\lambda^2}+\frac{1}{\sin\theta}\frac{\partial}{\partial\theta}\sin\theta\frac{\partial\Phi}{\partial\theta},\quad \Phi=(T,S);$$

$$\rho=\rho_0+10^{-3}[0.802(S-35)-T(0.0735+0.00469T)]. \tag{6}$$

Boundary conditions for $z=0$:

$$\nu\frac{\partial u}{\partial z}=-\frac{\tau_\lambda}{\rho_0},\quad \nu\frac{\partial v}{\partial z}=-\frac{\tau_\theta}{\rho_0},\quad w=0,$$
$$T=T^*(t,\lambda,\theta),\quad S=S^*(t,\lambda,\theta). \tag{7}$$

Boundary conditions for $z=H$:

$$\nu\frac{\partial u}{\partial z}=-R_2\int_0^H u\,dz,\quad \nu\frac{\partial v}{\partial z}=-R_2\int_0^H v\,dz,$$
$$w(H)=\frac{u(H)}{a\sin\theta}\frac{\partial H}{\partial\lambda}+v(H)\frac{\partial H}{\partial\theta},\quad æ\frac{\partial T}{\partial z}=0, æ\frac{\partial S}{\partial z}=0. \tag{8}$$

at the lateral wall Γ:

$$\mu\frac{\partial T}{\partial n}=0,\quad \mu\frac{\partial S}{\partial n}=0,\quad u_n=0. \tag{9}$$

at the initial moment $t=0$:

$$T=\tilde{T}(z),\quad S=\tilde{S}(z). \tag{10}$$

The equations are presented in the spherical coordinate system (λ is latitude, θ is the addition of longitude up to 90°, the axis z is directed vertically downwards); u, v, and w are velocity vector components, t is time, ρ_0 and ρ are the mean value and the anomaly of density, $\zeta=\xi-\frac{P_{atm}}{g\rho_0}$ is the reduced level, P_{atm} is atmospheric pressure, $z=\xi(\lambda,\theta)$ is the equation of ocean surface, R_1u and R_1v are the parameterization of horizontal turbulent viscosity, ν is the vertical turbulent viscosity coefficient, æ and μ are the vertical and horizontal turbulent temperature and salinity diffusion coefficients, $\ell=2\omega\cos\theta$ is the Coriolis parameter, a, ω, and g are the radius, the angular velocity, and acceleration of gravity of Earth, respectively, τ_λ and τ_θ are the wind stress, T^* and S^* are the known climatic distribution of temperature and salinity at the surface of ocean, R_2 is coefficient of bottom friction, n is the normal to lateral cylindrical wall Γ, H is the constant depth of ocean.

The method of solution of problem (1)–(10) is described in detail in [6], here we shall only mention that for equation (5) on the uniform five-degree grid the horizontal operator is approximated by the nine-dot difference scheme, received by the Richardson extrapolation, and the vertical operator, after introduction of new variable, condensing a grid at a surface of the ocean, is approximated by the second up-wind scheme.

The uniform five-degree grid shifted by 2.5° relatively the equator and the lateral boundary Γ is inserted in the polygonal area approximating the World Ocean from 72.5°S to 87.5°N. The model includes a realistic smoothed bottom topography. Two grids condescending at the surface of ocean under the square law are entered in the vertical coordinate. On one grid with integer-valued meanings of index k

$$z_k = (\eta_N - \eta_0)^2[(k - 1/2)/N]^2, \tag{11}$$

where $\eta_N^2 = H + \alpha, \alpha = 120$ sm, the variables u, v, T, S, and ρ are defined. The dots of the second grid lay in the middle of intervals

$$\frac{z_{k+1} + z_k}{2}, \quad z_{1/2} = 0, \quad z_{N+1/2} = H, \tag{12}$$

where $N = 24$ is the number of vertical levels.

In all the experiments, the horizontal diffusion coefficients $\mu_\lambda = \mu_\theta = 3.6 \cdot 10^7$ everywhere, except the equator belt of 3 grid step wide, and with a depth less than 500 m, where μ_λ increased to $3.6 \cdot 10^9$. The value of the coefficient R_1 near the equator was multiplied by 10^3. Other parameters used in the numerical experiment were:

$$R_1 = 4.5 \cdot 10^{-6}, \quad R_2 = 0.5 \cdot 10^{-7}, \quad \nu = 75, \quad æ = 0.2,$$

$$a = 6.4 \cdot 10^8, \quad \omega = 0.73 \cdot 10^{-4}, \quad \rho_0 = 1.02541, \quad g = 980.$$

In all the experiments, the global ocean model was forced by the seasonally varying climatic temperature, salinity, and wind stress at the ocean surface. In each experiment, the model was run until equilibrium with the time step 10 days.

In the first experiment on a surface of ocean, the climatological data on temperature for winter of northern hemisphere were corrected. The corrections were made for a north-east part of Northern Atlantic. If in the previous experiments in a grid point the mean value from a five-degree trapezoid was set, the value with minimum temperature was now selected.

Besides, salinity were corrected on large values in places indicated in [3]. These places are close to Greenland, and also Ross and Weddell seas near

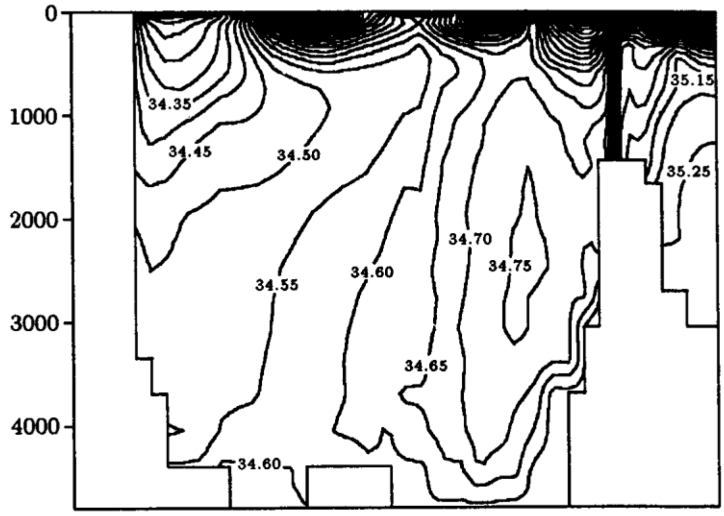

Figure 1. Zonally averaged salinity of the global ocean (in psu)

Antarctida. Starting from the state obtained in the previous experiment in [1] the pattern was integrated for 5000 years until stable state. Mean temperature of ocean became equal to 4.08 grades. The mean salinity of World Ocean was increased from 34.484 psu up to 34.596 psu.

The distribution of the zonally average temperature field on a meridional plane has varied a little. The large modifications have taken place with salinity (Figure 1). The intermediate antarctic tongue of gentle salt waters has appeared on the depths below 500 m and has penetrated to the equator. The salinity of the deep ocean increased.

In the second experiment, the influence of the higher saline Mediterranean waters was taken into account according to [3, 7]. On each time step in one grid point lying near a channel of Gibraltar at each grid level on thickness of a channel, that is on depths from 49 up to 464 m, salinity and temperature of the Mediterranean waters from the Levitus data were instantaneously mixed in a direction across a channel. Besides, in this experiment, the coefficients of vertical and horizontal diffusion were functions of the depth according to [3].

As well as in the previous experiment, a field of heat practically has not varied. Though mean temperature increased up to 4.2 grades. The variations have taken place in a field of salinity. The mean field of salinity obtained after 2000 years of integrating is shown in Figure 2. The penetration of the height saline deep waters from northern hemisphere to southern one was obtained. The low saline antarctic intermediate waters become more appreciable. The mean salinity increased up to 34.628 psu. The average vertical profiles of temperature and salinity are shown in Figure 3. If the

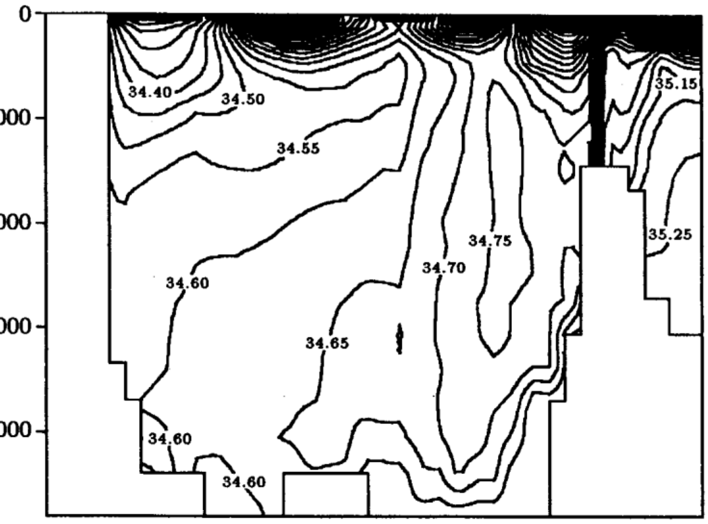

Figure 2. Zonally averaged salinity of the global ocean (in psu)

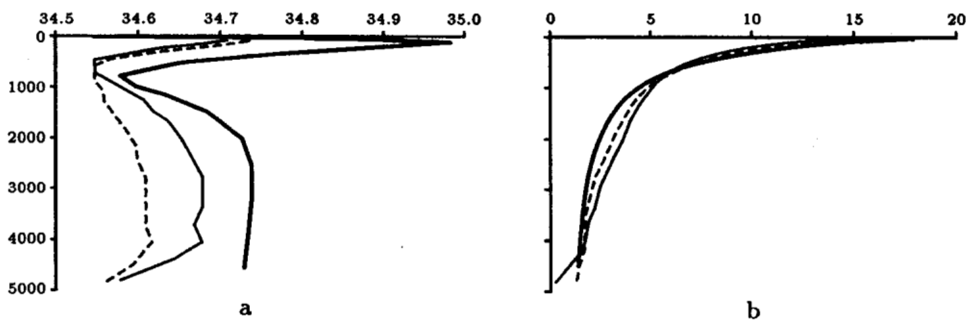

Figure 3. The vertical salinity (a) and temperature (b) profiles in the first experiment (dashed line), in the second experiment (thin line), and from Levitus' climatology (thick line)

modelled fields of salinity were qualitatively improved, the vertical profiles of temperature are worse. In high layer up to 700 m, waters became warmer as to the observational data, and on depths from 700 up to 4000 m they become more cold.

The use of constant coefficients of eddy diffusion both on a horizontal, and on a vertical, perhaps, is the reason of it. Probably, it is necessary to utilize isopycnic mixing, as in [2–4].

There is an intrusion of Antarctic Intermediate Water northwards which penetrates at the depth of 500–1500 m till equator and north tropic's latitude and corresponds well to the observation (see Figure 3). Analogous result was received in [2–4], where the Bryan model was used.

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