The Application of one-and-a-half-layer Ocean Model

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Introduction

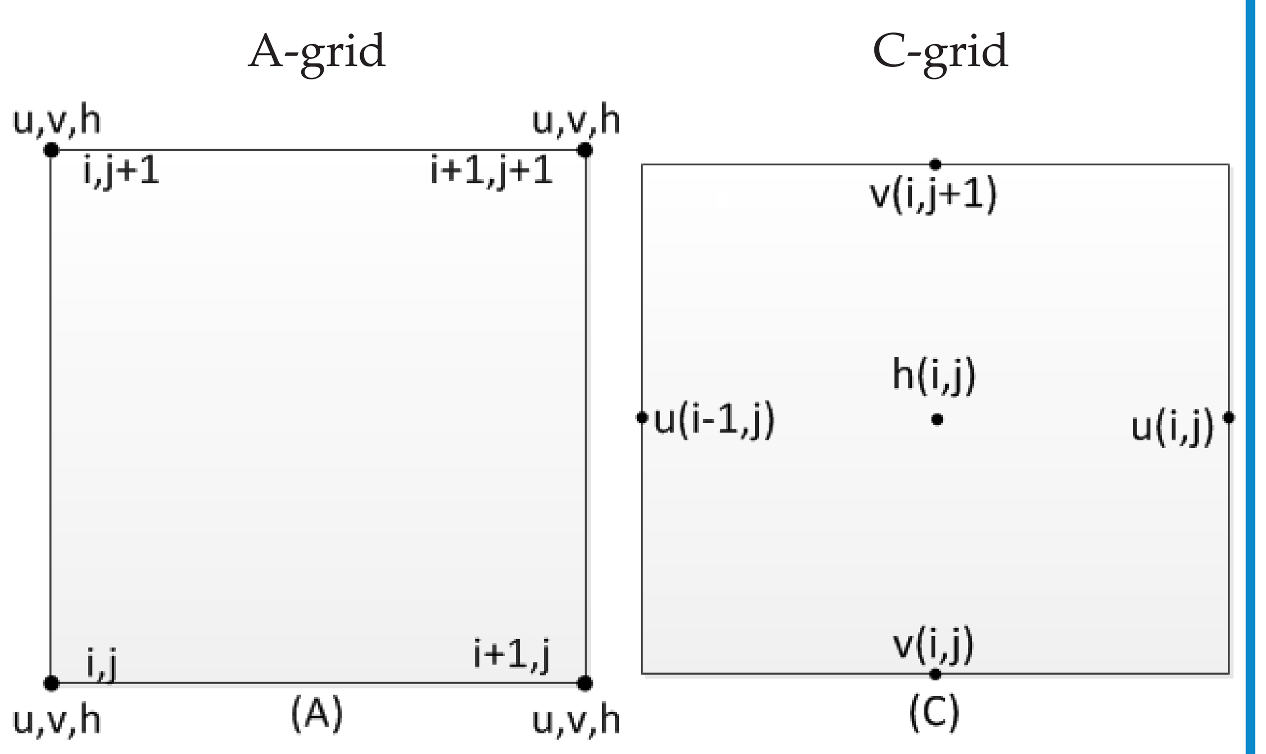
Ocean circulation is an independent cyclic system, covered in a large-scale basin. Especially, it is an important form of the water vapor interaction and the energy transport. So, the ocean circulation plays a vital role in regulation climatic of local and global area.

The oceanic general circulation model is a core element of the climate simulation, the construction of this model tends to be built on the geophysical fluid dynamics framework. Such as the nonlinear one-and-a-half layer ocean model we used. In this model, we include wind stress, coefficient of horizontal eddy viscosity and coriolis force to achieve the response of upper-layer ocean.

Through central difference scheme in Arakawa C-gird, we simulate the condition of wind-driven circulation, the direction of main flow, the development of energy, and the upper-layer thickness in North Pacific and North Atlantic.

Arakawa C-grid

As shown bellow, we introduce the A-grid and C-grid:



Arakawa A-grid:
The "unstaggered" Arakawa A-grid evaluates all quantities at the grid corners. It is the only unstaggered grid type.

Arakawa C-grid:
The "staggered" Arakawa C-grid separates evaluation of all quantities compared to the Arakawa A-grid. In this work, the water high is in the middle of the cell, the u component at the centers of the left and right grid faces, and the v component at the centers of the upper and lower grid faces.

Consider the efficiency of computation, we prefer the Arakawa C-grid. Because under the C-grid, we can do the same work within less information.

Model Equation

One-and-a-half layer ocean model equation:

$$\begin{cases} h_t + (hu)_x + (hv)_y = 0 \\ u_t + uu_x + vu_y - fv + gh_x = A_H \nabla^2 u + \frac{\tau^x}{\rho h} \\ v_t + uv_x + vv_y + fu + gh_y = A_H \nabla^2 v + \frac{\tau^y}{\rho h} \end{cases} \quad (1)$$

where h denotes the water height, u and v are the velocity of the fluid, f is the coriolis force, A_H is the coefficient of horizontal eddy viscosity, τ^x and τ^y are the surface wind stresses, g is the gravitational constant, and ρ is the average density.

Data

Table 1: topography data

region	Coordinates	resolution
North Pacific	$100^o E \sim 140^o W$ $0 \sim 140^o N$	$1/6^o$
North Atlantic	$100^o W \sim 30^o E$ $0 \sim 66.5^o N$	$5/12^o$

Wind Stress: NCEP Monthly Mean of u and v Component

Numerical Method

For the continuous equation, we have

$$\frac{\partial h_{i,j}}{\partial t} = -\delta_x^+ (hu)_{i-1/2,j} - \delta_y^+ (hv)_{i,j-1/2}$$

For the u momentum equation, we have

$$\frac{\partial u_{i,j}}{\partial t} = f\xi_x^+ v_{i-1/2,j} - g\delta_x^+ h_{i,j} + A_H(\Delta_x^2 + \Delta_y^2) u_{i,j}$$
$$-u_{i,j}\Delta_x u_{i,j} - \xi_x^+ v_{i-1/2,j} \Delta_y u_{i,j} + \frac{\tau^x}{\rho \xi_x^+ h_{i,j}}$$

For the v momentum equation, we have

$$\frac{\partial v_{i,j}}{\partial t} = -f\xi_y^+ u_{i,j-1/2} - g\delta_y^- h_{i,j} + A_H(\Delta_x^2 + \Delta_y^2) v_{i,j}$$
$$-v_{i,j}\Delta_y u_{i,j} - \xi_y^+ u_{i,j-1/2} \Delta_x v_{i,j} + \frac{\tau^y}{\rho \xi_y^+ h_{i,j-1}}$$

Remark:

$$\delta_x^+ u_{i,j} = \frac{u_{i+1,j} - u_{i,j}}{\Delta x} \qquad \delta_y^+ u_{i,j} = \frac{u_{i,j+1} - u_{i,j}}{\Delta y}$$
$$\delta_x^- u_{i,j} = \frac{u_{i,j} - u_{i-1,j}}{\Delta x} \qquad \delta_y^- u_{i,j} = \frac{u_{i,j} - u_{i,j-1}}{\Delta y}$$
$$\Delta_x u_{i,j} = \frac{u_{i+1,j} - u_{i-1,j}}{2\Delta x} \qquad \Delta_y u_{i,j} = \frac{u_{i,j+1} - u_{i,j-1}}{2\Delta y}$$
$$\xi_x^+ u_{i,j} = \frac{u_{i+1,j} + u_{i,j}}{2} \qquad \xi_y^+ u_{i,j} = \frac{u_{i,j+1} + u_{i,j}}{2}$$
$$\delta_x^+ \delta_x^- = \Delta_x^2 \qquad \delta_y^+ \delta_y^- = \Delta_y^2$$

Test for the North Pacific

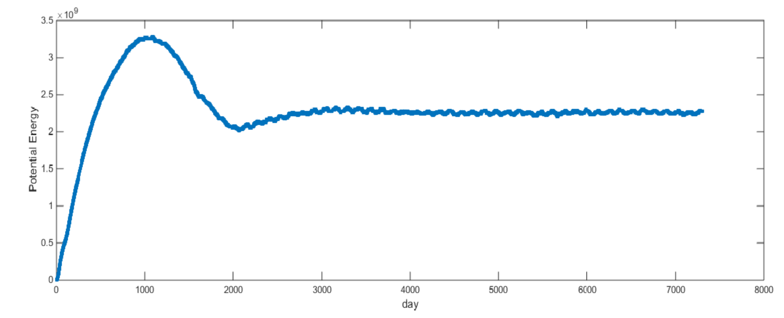


Figure 1: North-Pacific potential energy

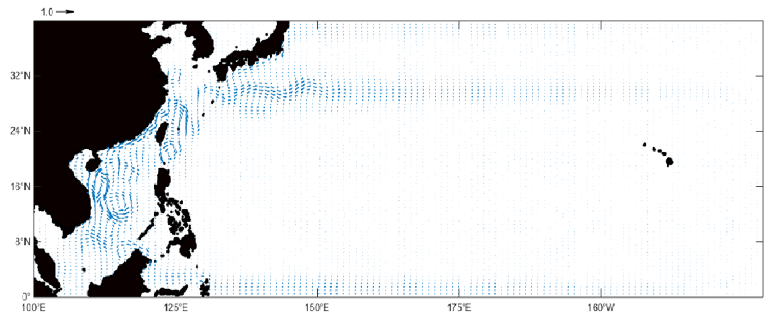


Figure 2: North-Pacific quiver

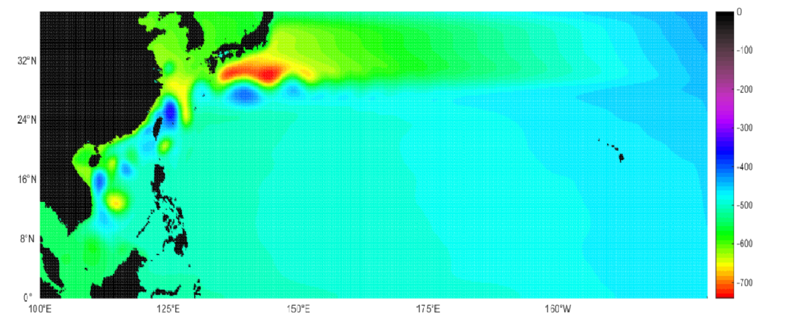


Figure 3: North-Pacific thickness

In this case, we use constant wind to drive the model towards the flat ocean where the initial depth $h = 500m$, $A_H = 6500m^2s^{-1}$.

In fig.1, the change in potential energy has stabilized for about 20 years, this means the system have reached a steady status. In fig.2, under the mean state of the climate, we simulate the clockwise wind circulation of the North Pacific in middle and low latitudes which flows from $0 \sim 30^\circ N$. Conjunction with fig.2 and fig.3, we can observe the increase of velocity and the upper-layer thickness form west to east which reach the maximum value in the Japan Sea. This confirms the phenomenon of westward.

Test for the North Atlantic

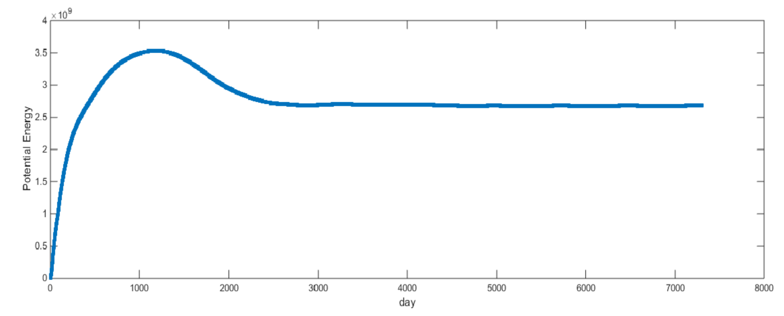


Figure 4: North-Atlantic potential energy

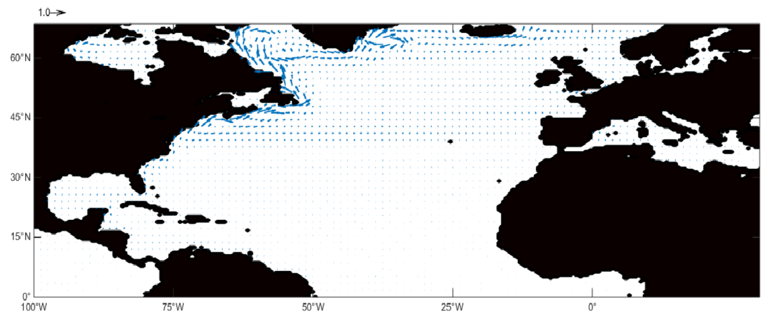


Figure 5: North-Atlantic quiver

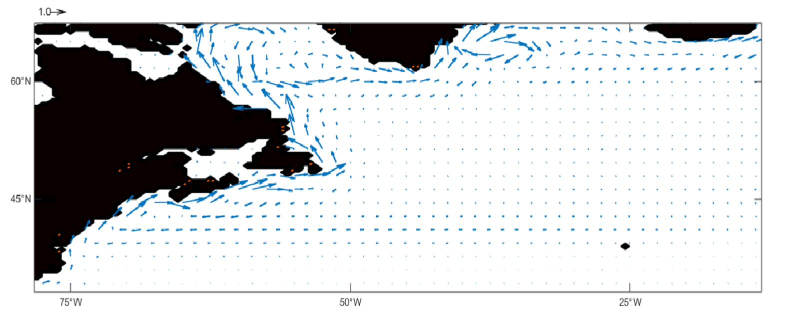


Figure 6: Local quiver

In this case, we also use constant wind to drive the model towards the flat ocean where the initial depth $h = 350m$, $A_H = 3500m^2s^{-1}$.

As well as the case in North Pacific, the system reach the mean state of the climate. In fig.5, we simulate the Mexico Gulf in North Atlantic which flows from $0 \sim 60^\circ N$. Conjunction with fig.5 and fig.6, we can observe the scale of velocity which is larger than North Pacific, this confirms the powerful Mexico Gulf. In fig.6, the Mexico Gulf is divided into two currents in 60°N, the north one flows into the Arctic Ocean.

Future Work

- Use two-and-a-half ocean model with a realistic open boundary and high-resolution method for simulation.
- Lead into thermohaline circulation in vertical.

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