Bull. Nov. Comp. Center, Num. Model. in Atmosph., etc., 1 (1993), 1–16

The ocean response to the tropical cyclone over the Kuroshito region

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The ocean response to the tropical cyclone over the Kuroshio region is studied to obtain some information about the 3-D structures of temperature and current tendencies. The data for temperature and salinity used in this investigation were obtained at Far East Hydrometeorological Institute. Diagnostic and adaptational calculations of currents were made on the basis of 3-D regional ocean circulation model (ROCM).

To simulate interaction between atmosphere and ocean ROCM was completed by the coupled atmosphere-ocean boundary layer model. In addition to the ocean mixed layer parameterization this model contains the turbulent closure model for the lower atmosphere up to 2 km. This model includes parameterization of turbulent length scale, solar and infrared radiation, phase transformation of water vapor, cloudiness, rainfall, convective adjustment. The bulk formulas are used to calculate heat, moisture (salinity) and momentum fluxes between the two media at the sea surface. The boundary conditions at the upper level are taken to simulate the tropical cyclone crossing this region in the north-east direction with the speed 56 km/h.

The resulting fields of temperature, salinity and currents show a very regular structure on the cyclone track. It consists of the mixed layer, deepened and cooled by the value of 0.18^0, the very deep Ekman pumping cell, the tendency of the streams to flow round the Idzu islands and the deep horizontal recirculation structure.

1. Introduction

The forecasting of tropical cyclones and investigation of their influence on the global atmosphere-ocean structures are both important problems. It is common knowledge that tropical hurricanes and typhoons are able to bring large financial damage for many countries. The theoretical and practical facilities are now fair enough to give results comparable with the reality.

According to recent advances the intensive cyclonic pressure minimum arises as a result of non-uniform geographical patterns [19] or is generated by atmospheric easterly waves [18]. After that the cyclone is fed through the sea surface temperature anomalies of strong western boundary currents such as Gulf Stream and Kuroshio [19]. The basic energy source is the latent heat released by phase transformations of water vapor.

This study is generally concerning to the Kuroshio region that is also known as one of the most cyclogenetically active zones. Among the features contributing to the predominance of cyclogenegis are enhanced coastal baroclinic zones maintained by the strong thermal contrast between the cold continent to the west and the relatively warm ocean to the east [20].

The influence of the tropical cyclones on the internal structure of the ocean is interesting to understand both the direct response to the intensive storm forcing and the general ocean circulation changes. There is a hypothesis stated that the tropical cyclones are able to intensify the anticyclonic large-scale circulation in middle latitudes followed by the development of strong barotropic instability of jet stream and generation of barotropic eddy-waves propagating to the neighbouring regions of open ocean [17]. As a direct response it is known some observations confirming that long-time stable position of the cyclone leads to the ring circulation with cold water lifted in the center [5]. This provides with a chain of negative adverse effect: sea surface cooling – evaporation decrease – decrease of heat escaped after condensation and as a result cyclone relaxation. The negative temperature anomaly rested after that can affect further atmosphere processes.

The main purpose of the paper is to reveal the direct response of the ocean to the tropical cyclone that moving fast through the region of the Kuroshio jet stream and strong subpolar front. As a secondary purpose we shall look after the Kuroshio stream position as it may deal with the problem of bimodality [15].

2. Regional ocean circulation model (ROCM)

2.1. System of equations

The model governing equations are the primitive ones written for the sphere of radius a in spherical coordinates (λ, θ, z) (z is directed from surface to bottom):

$$
\begin{aligned}
&\frac{du}{dt} - (f-\gamma)v = -\frac{m}{\rho_0}\frac{\partial p}{\partial \lambda} + \frac{\partial}{\partial z}\nu\frac{\partial u}{\partial z} + F^{\lambda},\\
&\frac{dv}{dt} + (f-\gamma)u = -\frac{n}{\rho_0}\frac{\partial p}{\partial \theta} + \frac{\partial}{\partial z}\nu\frac{\partial v}{\partial z} + F^{\theta},\\
&\frac{\partial p}{\partial z} = g\rho,\\
&m\Big(\frac{\partial u}{\partial \lambda} + \frac{\partial}{\partial \theta}\big(v\frac{n}{m}\big)\Big) + \frac{\partial w}{\partial z} = 0, \qquad (1)
\end{aligned}
$$

$$\frac{dT}{dt} = \frac{\partial}{\partial z}\nu_T\frac{\partial T}{\partial z} + F^T,$$
$$\frac{dS}{dt} = \frac{\partial}{\partial z}\nu_S\frac{\partial S}{\partial z} + F^S,$$
$$ho = r(T, S).$$

Initial and boundary conditions are:

$$t = 0: \quad u = u^0, \quad v = v^0, \quad T = T^0, \quad S = S^0;$$

at the ocean surface $z = 0$:

$$-\rho c_p \nu_T \frac{\partial T}{\partial z} = H_s + \mathcal{L}E_s - I,$$
$$-\rho\nu_S\frac{\partial S}{\partial z} = S(P_s - E_s),$$
$$-\rho\nu\frac{\partial \vec{U}}{\partial z} = \vec{\tau},$$
$$w = 0;$$

at the ocean bottom $z = H(\lambda, \theta)$:

$$\frac{\partial T}{\partial z} = \frac{\partial S}{\partial z} = 0, \quad \nu\frac{\partial \vec{U}}{\partial z} = -R_H\vec{\bar{U}}, \quad w = mu\frac{\partial H}{\partial \lambda} + nv\frac{\partial H}{\partial \theta},$$

at the solid boundary G_1:

$$\frac{\partial T}{\partial \vec{N}} = 0, \quad \frac{\partial S}{\partial \vec{N}} = 0, \quad \vec{U}\cdot\vec{N} = 0, \quad \frac{\partial \vec{U}\cdot\vec{K}}{\partial \vec{N}} = 0,$$

at the liquid boundary G_2:

$$T = T_G, \quad S = S_G, \quad \vec{U} = \vec{U_G},$$

where $\vec{N}$ and $\vec{K}$ are normal and tangent unit vectors to the $G = (G_1 \cup G_2)$ surface accordingly.

$$\vec{U} = (u, v), \quad H\vec{\bar{U}} = \int_0^H \vec{U}dz,$$

u, v, w – the velocity vector components by the (λ, θ, z) coordinate directions respectively,

$n = \frac{1}{a}$, $m = \frac{1}{a\sin\theta}$, $f = -2\omega\cos\theta$, $\gamma = m\cos\theta$,

ω – angular velocity of Earth rotation,

p – pressure,

ρ – density deviation from averaged ρ_0 =const,

T – temperature,

S – salinity,

$\vec{\tau}$ – wind stress vector,

R_H – drag coefficient,

$r(T,S)$ – sea water state function,

F^λ, F^θ – forms of parameterizing subgrid turbulent momentum exchange and

$$F^{T,S} = m\Big(\frac{\partial}{\partial\lambda}\mu m\frac{\partial(T,S)}{\partial\lambda} + \frac{\partial}{\partial\theta}\mu\frac{n^2}{m}\frac{\partial(T,S)}{\partial\theta}\Big).$$

In the system (1) the common notation is used for the time derivative

$$\frac{d}{dt} = \frac{\partial}{\partial t} + mu\frac{\partial}{\partial\lambda} + nv\frac{\partial}{\partial\theta} + w\frac{\partial}{\partial z}.$$

This model is the adaptation for the Kuroshio region of the finite-element World ocean circulation model worked out in the Novosibirsk Computing Center [9]. In the version presented here the calculations are carried out in the physical coordinate z instead of transformed coordinate z/H as it was in the main version.

2.2. Numerical algorithm

The governing equations of the model are transformed by separation of the external and internal modes. The equations for the external mode are reduced by rotor operation to the vorticity and stream function equations:

$$\zeta_t + [\vec{\nabla} \times (\zeta + f)\vec{\nabla}\psi] + R_H\zeta - \mu\Delta\zeta = 0, \tag{2}$$

$$\Big(\vec{\nabla}\cdot(\frac{1}{H}\vec{\nabla}\psi)\Big) = \zeta. \tag{3}$$

The numerical realization of the model is based on the combination of splitting and finite-element methods.

Let us itemize some of the essential points of this algorithm:

1. the transformation of the advective terms of the model from the gradient form to some divergent one;
2. preliminary splitting of the equation;

3. finite element discretization by space;
4. splitting of the finite-element method (FEM) operators to 1-D ones;
5. using the implicit and semi-implicit schemes by the time.

Brief remarks are to be made to comment the listed items. For the equations (2) and (3) the splitting by physical processes was used. It means that for each time interval two problems are solved sequentially: equation for the forced Rossby waves and equation for relative vorticity advection-diffusion. Those equations are solved by FEM discretization with piece-wise coordinate function. In this case there are two invariants which are conserved in discrete form in the absence of dissipation at the general time-step: averaged vorticity $\int_\Omega \zeta d\Omega$ and kinetic energy $\int_\Omega (\nabla\psi)^2 d\Omega$. The averaged square of relative vorticity is conserved only for the step of advection-diffusion. The scheme of the Jacobian approximation is analogous to [1]. At the regular grid linear combination of finite element Jacobians give the seven point modification of Arakawa Jacobian. The realization of the vorticity advection-diffusion equation by time is carried out by splitting method [13].

For heat and salt equations double splitting was used. After the transformation of advection terms to divergent form the result is

$$\frac{\partial T}{\partial t} + \sum_{i,j} \left(a_{ij} T_{x_j}\right)_{x_j} = 0 \quad (a_{ij} \neq a_{ji}), \tag{4}$$

$$(x_1, x_2, x_3) \equiv (\lambda, \theta, z).$$

Then equation (4) can be reduced to the series of 2-D equations at the cross-sections of the basin by means of weak approximation method [4]. So the difficulties of using FEM in 3-D domain are avoided. The conservation laws for averaged temperature and squared temperature are valid. Each of 2-D equations is discretized by FEM with piece-wise linear coordinate functions. Then the finite element grid operators are split to the 1-D ones, which include coordinate directions. The 1-D operators are of three points and positively semidefined.

2.3. Mixed-layer parameterization

After each step of integration of ROCM vertical profiles of temperature, salinity and velocity components are corrected to eliminate static and dynamic instability. The resulting distribution is taken as initial condition for next time step.

The corrections are made if the discretized Richardson number

$$Ri = \frac{g}{\rho}\frac{\partial \rho}{\partial z} \Big/ \left|\frac{\partial \vec{U}}{\partial z}\right|^2 \tag{5}$$

becomes less than critical number that is taken to be equal 0.5. If this happens, the mixed layer will be assumed to have homogeneous vertical distribution of all characteristics. The new values of this characteristics ($\bar{T}$, $\bar{S}$, $\bar{u}$, $\bar{v}$) are calculated with taking into account some of the conservation laws: heat and salt content and total momentum vector. The water density of the mixed layer is calculated from the state equation $\bar{\rho} = r(\bar{T}, \bar{S})$. Then the mixed layer depth can be determined from the kinetic energy conservation equation:

$$\Big[\frac{1}{2}\int\limits_0^h \rho|\vec{U}|^2 dz - \frac{1}{2}\bar{\rho}|\vec{\bar{U}}|^2 h\Big] - \Big[\int\limits_0^h g\rho z dz - g\bar{\rho}\frac{h^2}{2}\Big] + D = 0. \qquad (6)$$

The term in the first square brackets of (6) represents source of kinetic energy escaped after mixing. The second square brackets include the term representing energy seek for the destruction of steadily stratified layer or energy source if stratification is unstable. The last term of (6) is some kind of the energy dissipation parameterization. Thus, from five conservation laws we have four thermodynamic characteristics of the mixed layer and its depth h.

3. Atmospheric boundary layer

3.1. General equations and boundary conditions

For the purpose of this work the model of lower atmosphere is needed to simulate the transformation of the tropical cyclone structure in the atmospheric boundary layer and to serve as a transmitting system between the cyclone and the ocean. This model is to be three-dimensional one. So the basic equations are taken in the following form

$$\begin{aligned}
\frac{du}{dt} &= -\frac{\partial \overline{u'w'}}{\partial z} + f(v - V_g) + \frac{\partial U_g}{\partial t}, \\
\frac{dv}{dt} &= -\frac{\partial \overline{v'w'}}{\partial z} - f(u - U_g) + \frac{\partial V_g}{\partial t}, \\
& m\Big(\frac{\partial u}{\partial \lambda} + \frac{\partial}{\partial \theta}\big(v\frac{n}{m}\big)\Big) + \frac{\partial w}{\partial z} = 0, \\
\frac{d\theta}{dt} &= -\frac{\partial \overline{\theta' w'}}{\partial z} + \epsilon_r + \epsilon_p, \\
\frac{dq}{dt} &= -\frac{\partial \overline{q'w'}}{\partial z} - \epsilon_c + \epsilon_e, \qquad (7)
\end{aligned}$$

$$\frac{dq_l}{dt} = -\frac{\partial \overline{q_l' w'}}{\partial z} - \epsilon_e + \epsilon_c + \epsilon_{pr},$$
$$\frac{\partial}{\partial t}\begin{pmatrix} U_g \\ V_g \end{pmatrix} = f\begin{pmatrix} 0 & 1 \\ -1 & 0 \end{pmatrix}\begin{pmatrix} U_g \\ V_g \end{pmatrix} - \frac{1}{\rho}\vec{\nabla}\tilde{p},$$
$$\frac{\partial p}{\partial z} = -g\rho,$$
$$p = \rho RT(1 + 0.61q + q_l),$$
$$\theta = T\left(\frac{p_0}{p}\right)^{\gamma}.$$

In the system of equations (7):

u, v, w – wind velocity components according to coordinate system axes (Oz is in upper direction);

θ – potential temperature of the air, related with the absolute temperature T by means of latter equation of the system;

q – specific humidity;

q_l – specific liquid water content;

U_g, V_g – geostrophic wind components;

p – pressure;

$\vec{\nabla}\tilde{p}$ – horizontal gradient of the sea surface pressure $\tilde{p}$;

ϵ_r, ϵ_p – radiation and phase heat income;

ϵ_c, ϵ_e, ϵ_{pr} – rates of the water condensation, evaporation and precipitation;

$\overline{u'w'}$, $\overline{v'w'}$ – components of the vertical Reynolds stress;

$\overline{\theta' w'}$, $\overline{q' w'}$, $\overline{q_l' w'}$ – vertical turbulent fluxes;

R – gas constant for air;

γ– adiabatic power (=0.29);

p_0 – reference pressure level (=1000mb).

The solution of the system (7) is to be found in the area of lower atmosphere from the sea level to 2 km. It can be divided by two layers. First, the layer of constant fluxes is distributed from the sea level up to 50 m. The characteristics of this layer are defined by bulk parameterization formulas [7]. In the second layer the system of equations (7) is solved

numerically. The boundary conditions for this area are
$z = 50$ m:

$$\rho\overline{\vec{U}'w'} = \vec{\tau}, \quad \rho c_p\overline{\theta' w'} = H_s,$$
$$\rho\overline{q'w'} = E_s, \quad \rho\overline{q_l' w'} = 0,$$
$$p = \tilde{p}(x,y,t);$$

$z = Z = 2$ km:

$$u = U_g, \quad v = V_g, \quad \theta = \theta_H(x,y), \quad \frac{\partial q}{\partial z} = \frac{\partial q_l}{\partial z} = 0.$$

The functions $\tilde{p}$ and θ_H are supposed to be externals, we shall use $\tilde{p}$ to define space distribution of pressure analogous to cyclonic minimum. As a background for these functions the climatological distribution is used, and the deviation to it for pressure cyclonic minimum is defined as

$$\tilde{p}' = -\tilde{p}_0 \exp\left(\frac{-(x-x_0(t))^2 - (y-y_0(t))^2}{R^2 - (x-x_0(t))^2 - (y-y_0(t))^2}\right)$$

for the internal area of the circle of radius R

$$(x - x_0(t))^2 + (y - y_0(t))^2 < R^2,$$

where $\tilde{p}_0$ is the magnitude of relative pressure minimum; $x = a\lambda\sin\theta$, $y = a\theta$, $x_0 = a\lambda_0\sin\theta_0$, $y_0 = a\theta_0$; λ_0 and θ_0 – the position of the cyclone center.

The lateral boundary conditions for u, v, θ, q and q_l are the fluxes of this characteristics equal to the climatological ones in case of inflow, and the fluxes from inside region otherwise.

3.2. Turbulent closure

To calculate the turbulent fluxes $\overline{\phi' w'}$ the parameterization of diffusion approximation is used

$$\overline{\phi' w'} = -\alpha_\phi k\frac{\partial\phi}{\partial z},$$

where k – coefficient of turbulent viscosity and α_ϕ – proportionality coefficient specific to ϕ. Turbulent coefficient is defined by the Kolmogorov relation [8]: $k = l\sqrt{e}$, where l – turbulent length scale, e – turbulent kinetic energy calculated from the equation

$$\frac{\partial e}{\partial t} = \alpha_e\frac{\partial}{\partial z}k\frac{\partial e}{\partial z} + k\left(\left(\frac{\partial u}{\partial z}\right)^2 + \left(\frac{\partial v}{\partial z}\right)^2\right) - \alpha_T k\frac{g}{\theta}\left(\frac{\partial\theta_v}{\partial z}\right) - \frac{c_e e^{\frac{3}{2}}}{k} \qquad (8)$$

and the turbulent length scale parameterized by the Blackadar formula

$$l = \frac{\kappa z}{1 + \frac{\kappa z}{l_\infty}}$$

with

$$l_\infty = 0.01 \frac{\int_0^Z \sqrt{e} z dz}{\int_0^Z \sqrt{e} dz}.$$

The boundary conditions for equation (8) are
$z = 50$ m: $e = e_0 = \left(\frac{k^*}{l}\right)$
$z = Z$: $\frac{\partial e}{\partial z} = 0$,
where k^* – turbulent coefficient given by bulk parameterization of the constant flux layer.

3.3. Other parameterizations

In addition to turbulent closure the model includes some other parameterizations that we shall not detail but mention them here:

- dry and wet convection [12];
- solar and infrared radiation [3][10];
- hydrological cycle [16];
- storm intensification of turbulent exchanges [6].

4. Numerical experiment

In the numerical experiment we consider the tropical cyclone as it passes through the region in north-east direction with the speed of 56 km/h. It is set by determination of x_0 and y_0 in a proper way. The radius of the cyclone is $R = 550$ km and the relative minimum value $\tilde{p}_0 = 10$ mb. The cyclone displacement in 24 hours are presented by Figure 1.

The beginning state of the ocean is set by use of the Nelezin and Man'ko data [14] of the Far East Hydrometeorological Institute. The data consist of temperature and salinity fields obtained by approx. 40000 monthly averaging measurements for the period from 1936 to 1988 for the depth to 1500 m. The resulting set of data is placed on the $1^0 \times 1^0$ grid for the Kuroshio region. For the horizons deeper than 1500 m the data from Levitus atlas [11] are used.

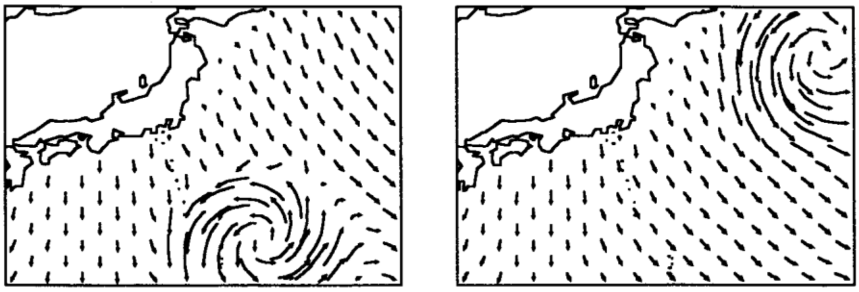

Figure 1. The tropical cyclone displacement in 24 h shown by means of geostrophic wind fields.

The preliminary preparation for the main numerical experiment includes the diagnostic calculations for every month (the temperature and salinity were set by the data and velocity was found numerically) and short period of integration of the whole model, starting from the diagnostic fields, to obtain some adaptation of temperature and salinity to the currents. This procedure gives the insignificant changes in the distribution of main characteristics of the ocean, and adjusts the atmosphere turbulence field. After that two experiments are carried out: one includes the cyclone and another don't. The deviation fields are obtained by subtraction of the second experiment fields from the first ones.

It was stated out previously that some chain of negative adverse effect appears when the cyclone has a long-time stable position. Another features arise when it moves fast. The sea surface temperature deviation presented by Figure 2 shows clearly the negative anomaly approaching the value of -0.18^0 on the left side of the cyclone trajectory. The left displacement of anomaly minimum is caused by the cold-air outbreak over the ocean. The generation of this anomaly forced by the intensive heat and humidity exchanges between the atmosphere and the ocean during storming-time. The mixed layer depth increases abruptly from 80–100 m up to 130–150 m. Because of deep mixed layer the amount of temperature decreasing is not very large, but the amount of heat lost by the layer of upper 100 m is significant and approximately equals to $\delta Q = \rho c_p \times 100\ m \times 0.18^0 = 77.4 \times 10^6 J/m^2$. It is comparable with the value $86 \times 10^6 J/m^2$ estimated by Gregg for the period of cold-air outbreak off New England in January 1983, when a deep convective mixed layer was observed [2].

The cross-section of temperature deviation field (Figure 3) indicates that negative temperature anomaly approaches the depth 100 m (the initial mixed layer depth). The heat lost by this layer may supply the lower atmosphere by sensible and latent heat and the ocean layers underlying the

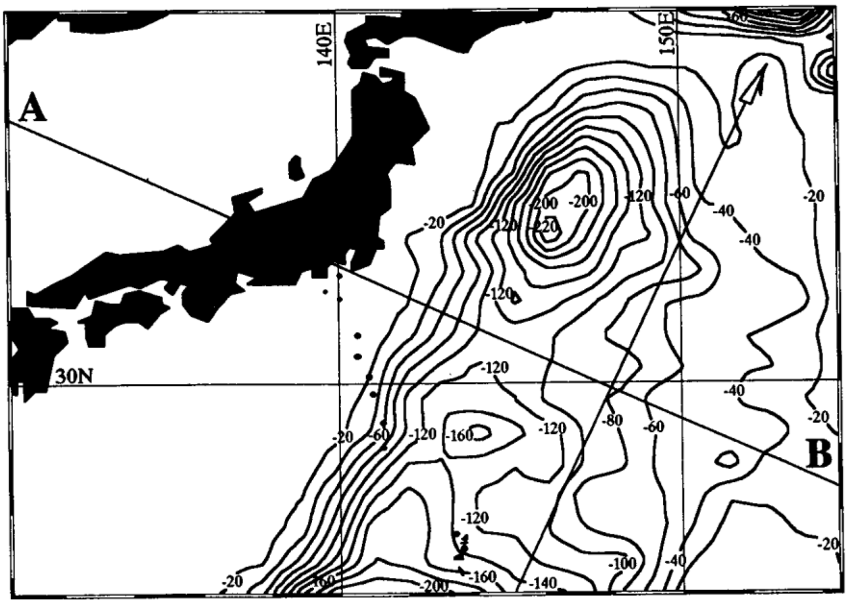

Figure 2. The sea surface temperature deviation (multiplied by 10^3) field, caused by the cyclone. The arrow displays the trajectory and position of cyclone.

mixed layer. So directly under this anomaly a layer of positive maximum is placed that is caused by the growing of the mixed layer depth and penetration of relatively warm near-surface water into the cold one of thermocline. This process seems to be able to generate internal waves packet, but because of model capability it can't be properly resolved. Nevertheless for the short-term period under consideration the results can't be significantly distorted by these waves. So they may be neglected.

Under the maximum the three-dimensional thermal structure caused by the Ekman pumping is located. It is generated by the near-surface current divergence, produced by the strict action of wind stress. The center of this thermal structure is formed at the level of 600 m. This is a very high depth value of cyclone influence on the internal ocean layers, and besides this influence is not acting through the turbulence but through the three-dimensional dynamics. The temperature deviation field at the level 600 m (Figure 4) displays central zone of upwelling and two peripheral zones of downwelling. The temperature difference is about 0.04^0 between these areas.

Figure 5 represents the deviation of sea surface currents after the cyclone moving. The direct action of the wind stress produces a divergent zone a little bit behind the cyclone center. It is seen that except the direct

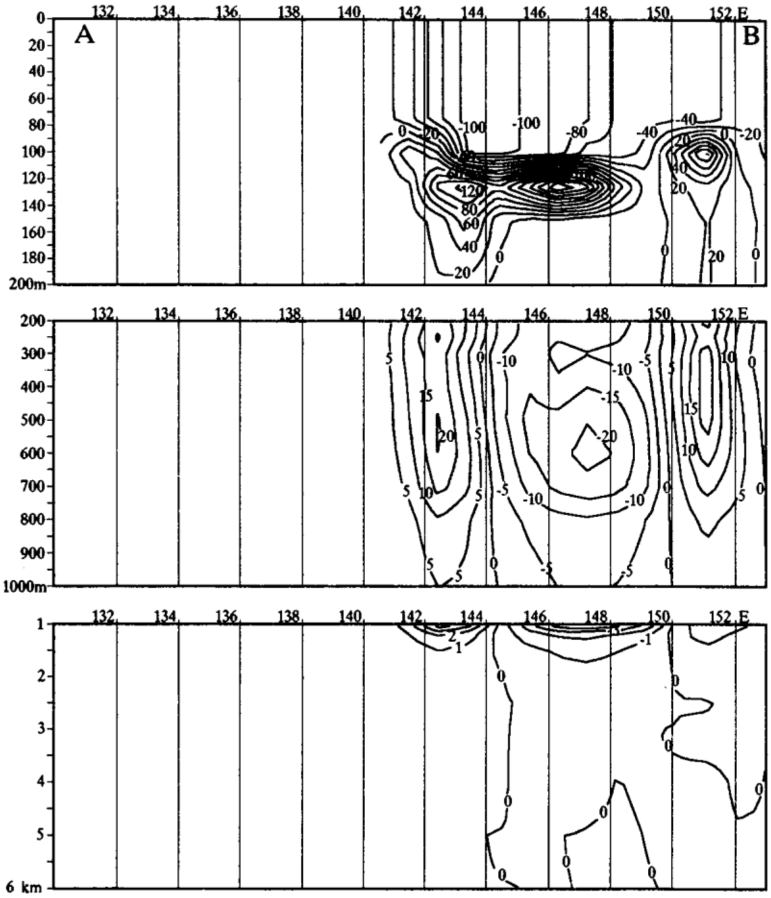

Figure 3. The cross-section of the temperature deviation (multiplied by 10^3) field in the direction shown in Figure 2.

cyclone forcing near its center some new feature arises south and east to the Idzu ridge. The convergent zone arises there followed by the current deviations directed to the north. From the left side of this zone the inflow of water moving round the Idzu islands occurs. It means that cyclone moving increases the tendency of the streams to flow round the islands, forcing the Kuroshio current in offshore direction and allowing the near-coastal penetration of northern cold water to the south along the continental slope. This tendency is strong enough as one can see in the absolute velocity field represented by Figure 6, but no penetration of cold water exists there. It seems to be just because of very rough resolution of our model.

The temperature gradients on the depth 600 m marked previously, causes the proper horizontal dynamics shown by Figure 7.

Figure 4. The temperature deviation (multiplied by 10^3) field caused by the cyclone on the depth 600 m

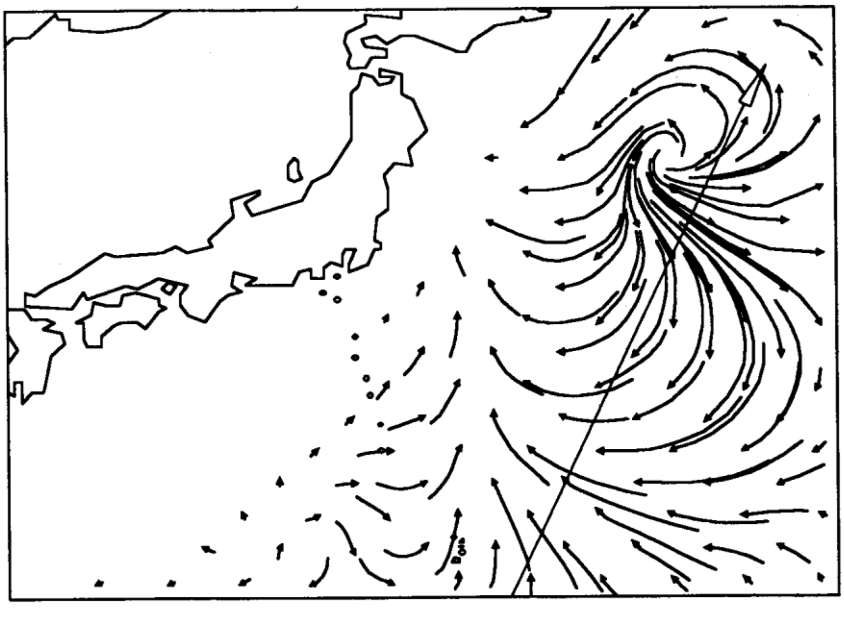

Figure 5. The sea surface currents deviation field

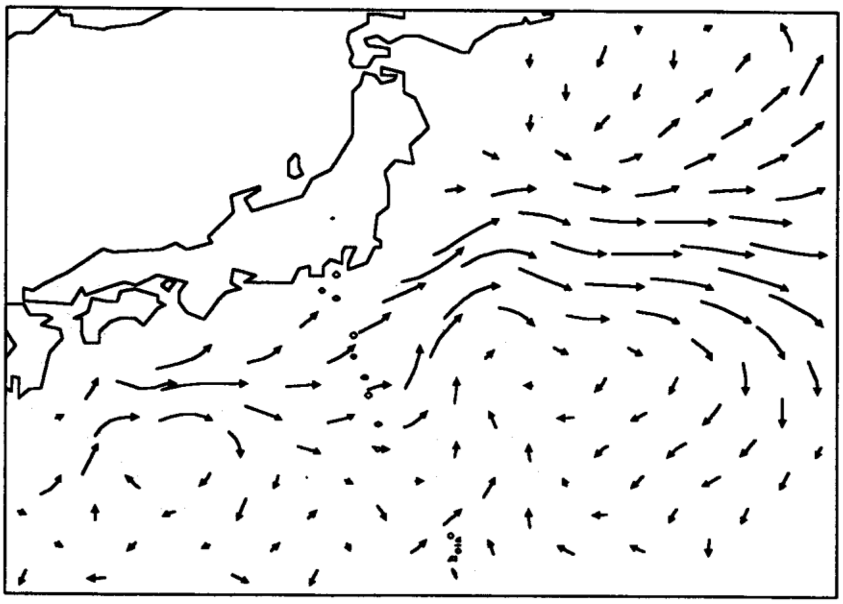

Figure 6. The sea surface currents field

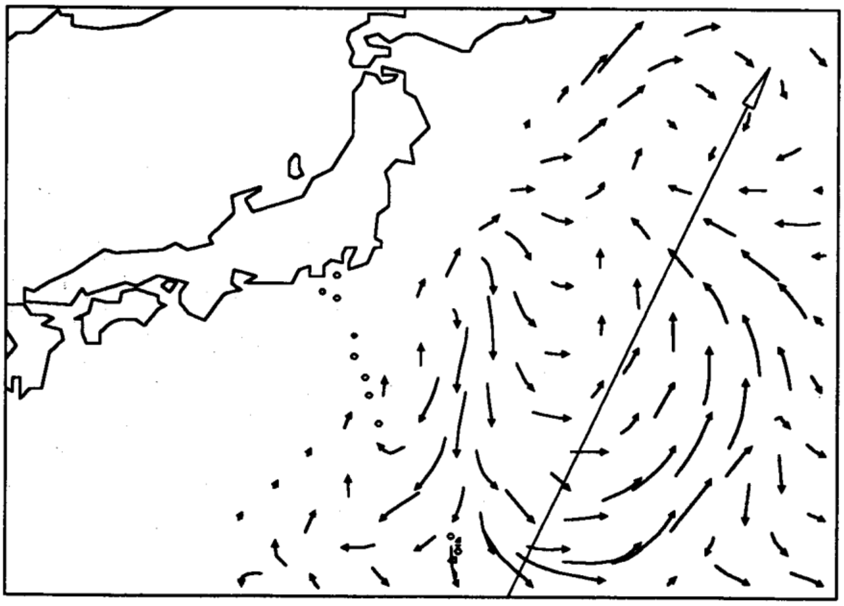

Figure 7. The currents deviation field on the depth 600 m

This deviation field is not a direct consequence of the cyclone only but the secondary feature of ocean adaptation to the external forcing. This anticyclonic circulation is a compensational one to the near-surface wind-driven currents.

5. Conclusion

The results of the numerical experiment make us capable to formulate the main features of ocean response to the tropical cyclone when it passes through the Kuroshio region:

- mixed layer deepening from 80–100 m to 130–150 m and cooling by the value of 0.18^0;
- formation of the deep Ekman pumping cell approaching horizons of 1 km or more which gives temperature deviation maximum at the depth 600 m;
- arising the tendency of the streams to flow round the Idzu islands;
- appearance of deep (on 600 m and deeper) horizontal recirculation structure in the cyclone track.

It is possible to achieve in future experiments more detailed information about the transformation of the near-coastal currents and some changes in the frontal zone. The near-coastal currents seem to be more intensive because of cyclone action. The model of fine resolution is to be designed to obtain any results about it.

Acknowledgements. We are very grateful to Dr.Nelezin for his courtesy of giving us a possibility to work with a set of temperature and salinity data prepared for the Kuroshio region.

References

[1] J.T. Fix, Finite element models for ocean circulation problems, SIAM J.Appl.Math., Vol. 29, 1975, 371–387.

[2] M.C. Gregg, Structures and fluxes in a deep convecting mixed layer, Proc. Hawaiian Winter Workshop, University of Hawaii at Manoa, 1987, 1–24.

[3] V.I. Hvorostianov, On the schematization of the atmospheric long-wave radiation spectrum for the cloud and smog models based on the vertical fine resolution spectral calculation, Izv. Akad. Nauk SSSR, Vol. 17, No. 10, 1984, 1022–1030 (in Russian).

[4] N.N. Janenko, The Peace-steps Method of the Solution of Multidimensional Studies in Mathematical Physics, Nauka, Novosibirsk, 1967 (in Russian).

[5] V.M. Kamenkovich, M.N. Koshljakov, A.S.Monin, Ocean Synoptical Eddies, Gidrometeoizdat, Leningrad, 1987 (in Russian).

[6] A.L. Kazakov, V.N. Lykossov, On the heat and moisture fluxes parameterization for storming period applied for the atmosphere-ocean interaction problems, Meteorologija i gidrologija, No. 8, 1980, 58–64 (in Russian).

[7] A.L. Kazakov, V.N. Lykossov, On the parameterization of the atmosphere surface interaction in numerical modelling of atmospheric processes, Tr. Zap. Sib. NII, Vol. 55, 1982, 3–20 (in Russian).

[8] A.N. Kolmogorov, The equation of the turbulent motion of uncompressible liquid, Izv. Akad. Nauk SSSR, Vol. 6, No. 1–2, 1942, 56–58 (in Russian).

[9] V.I. Kuzin, E.N. Golubeva, Numerical modelling of the temperature and currents in the World ocean using finite element method. Numerical modelling of the World ocean climate, Vychisl. Tsentr SORAN, Novosibirsk, 1986, 137–150 (in Russian).

[10] K.-N. Liou, G.D. Wittman, Parameterization of the radiative properties of clouds, J. Atmos. Sci., Vol. 36, No. 7, 1979, 1261–1273.

[11] S. Levitus, Climatological atlas of the World ocean, Environmental res.lab., Geophys. fluid dynamics lab., Princeton, N.J., Rockville, 1982.

[12] G.I. Marchuk, V.P. Dymnikov, V.B. Zalesny, V.N. Lykossov, V.Ja. Galin, Mathematical Modelling of General Circulation of the Atmosphere and Ocean, Gidrometeoizdat, Leningrad, 1984 (in Russian).

[13] G.I. Marchuk, V.I. Kuzin, On the combination of finite element and splitting-up methods in the solution of parabolic equations, J. Comp. Phys., Vol. 52, No. 2, 1983, 237–272.

[14] A.D. Nelezin, A.N. Man'ko, Calculations of the climatic fields of temperature, salinity and current velocity in Kuroshio region, (to be published) (in Russian).

[15] H. Nitani, Process of the occurrence, continuation and disappearance of the large meander of the Kuroshio off Enshu-Nada, Marine Sciences Monthly (Kaiyo Kagaku), Vol. 9, 1977, 29–36.

[16] G.A. Platov, Atmosphere-ocean interaction investigation by means of numerical modelling, Dissertation, Novosibirsk, Vychisl. Tsentr SORAN, 1993 (in Russian).

[17] D.G. Seidov, Synergetics of Ocean Processes, Gidrometeoizdat, Leningrad, 1989 (in Russian).

[18] L.J. Shapiro, Tropical storm formation from easterly waves: a criterion for development, J. Atmos. Sci., Vol. 37, No. 7, 1977, 1007–1021.

[19] V.V. Shuleykin, Calculation of tropical hurricanes development, motion and relaxation and main waves caused by hurricanes, Gidrometeoizdat, Leningrad, 1978 (in Russian).

[20] J.V. Zapotocny, Distribution of available potential and kinetic energy budget quantities associated with wintertime cyclone activity along the eastern coast of Asia and North America, Tellus, Vol. 42A, No. 5, 1990, 508–519.