Numerical Experiments of Tropical Cyclones on a Plane with Variable Coriolis Parameter

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Abstract

Numerical experiments of tropical cyclones are performed using a moving variable grid scheme proposed by Kitade (1979). The formulation for parameterization of cumulus convection is based on the concept of penetrative convection. The physical processes, such as the surface friction, latent and sensible heat supplies from sea surface, eddy dissipation and diffusion, dry convective adjustment and large-scale condensation, are also contained with conventional ways in present model.

The initial vortex develops into a mature disturbance in about 2 days. Its structure is similar to that of the tropical cyclone in the real atmosphere and that in numerical experiments performed by many researchers. The simulated tropical cyclone has some asymmetric features due to the variable Coriolis parameter in spite of the adoption of an initial nearly symmetric vortex. Furthermore, the developed tropical cyclone has dynamically unstable regions in the upper layer, which may also contribute to the asymmetry of the simulated vortex.

The simulated vortex moves north-north-westward, which is due to the effect of variable Coriolis parameter pointed out by Rossby (1948). The moderate tropical cyclone moves at a speed of about 3 km hour^{-1} for westward and at about 7 km hour^{-1} for northward components. The meandering about the mean path occurs with a period of about 25 hours, which seems to be due to the Magnus effect suggested by Yeh (1950).

When the sea surface temperature is changed, the path of the vortex center is deviated a little from that in the basic experiment. The deviation is partly explained for the following reason. The strength of vortex is strongly influenced by sea surface temperature. According to Rossby (1948), on the other hand, the northward acceleration due to variable Coriolis parameter increases with the increase of strength of vortex. Therefore the sea surface temperature affects the path of vortex. The variation of speed of vortex movement through its life cycle is also understandable from such point of view. It is confirmed by an experiment with non-uniform sea surface temperature that the asymmetric heating of cumulus convection affects the path of the vortex center.

1. Introduction

Many works have successfully simulated the development of tropical cyclones. Though axisymmetric models were developed by many authors (Ogura, 1964; Kuo, 1965; Yamasaki, 1968, 1977; Ooyama, 1969; Rosenthal, 1970, 1978; Sundqvist, 1970; Kurihara, 1975; Peng and Kuo, 1975; and Wada, 1979) and some of the observed features of real tropical cyclones were simulated, they could not simulate asymmetries such as the confluence and diffluence lines in the streamline fields, the cloud bands and the motion

of the tropical cyclones. Anthes *et al.* (1972), Kurihara and Tuleya (1974) and Jones (1977a) simulated an asymmetric tropical cyclones on a f -plane using a balanced symmetric vortex as the initial condition. They successfully simulated some of observed features, such as cloud bands and the asymmetries in the outflow layer.

Mathur (1974) simulated an asymmetric moving tropical cyclone on a plane with variable Coriolis parameter. The initial conditions for the model were obtained from the real data for hurricane Isbel, 1964, when it was at a tropical depression stage. The model simulated the asym-

metries in the wind, temperature and pressure fields, and also the movement of the tropical cyclone.

Madala and Piacsek (1975) developed a model to simulate moving and asymmetric tropical cyclones on a β -plane. A comparison of the results with f -plane calculations showed that the vortex on the β -plane intensified at a lower rate before the storm stage, but at the same rate thereafter. The vortex moved with a velocity of 4.3 km hour^{-1} for the westward and 3.3 km hour^{-1} for the northward components.

Jones (1977c) has also integrated a numerical model of tropical cyclone with a variable Coriolis parameter. He has found that the vortex moves north-north-westward in the developing stage. The northward movement was attributed to the acceleration force of the variable Coriolis parameter pointed out by Rossby (1948). After the mature stage is reached the tropical cyclone moves north-westward, and the moving speed increases. He speculated that the north-westward turn in the mature stage is primarily a result of the influence of the outflow layer mean current. The speed of vortex movement in Jones (1977c) is larger than that in Madala and Piacsek (1975), which was attributed to the difference of the intensities of initial vortices.

In order to predict tropical cyclone movement over a large area and yet resolve finer details near eye, moving multi-grid networks were utilized by Mathur (1974), Madala and Piacsek (1975), Jones (1977a, b, c), Ookochi (1974) and Hovermale (1976). Although they are advantageous in terms of saving computing time, some small-scale noise is generated at the interface because of the difference of resolvable scales in nets with different grid size.

The interface must be carefully treated to avoid the excitation of noise. An appropriate smoothing operation is indispensable. The center of the fine grid is restored to the vortex center whenever it moves a few grid intervals from the center of the fine grid network. When the fine grid network is translated, the new data at the leading edge of the grid net are calculated by an interpolation. Since these new data are not dynamically balanced, a shock noise may contaminate the model results. Kurihara and Teluya (1974) used a variable grid mesh, which does not have any explicit interface. Since they were interested in the development of tropical cyclone on a f -plane and thus the movement of the tropical cyclone was not large, the grid network

was fixed. When we are interested in the movement of tropical cyclone, the high resolution region in the grid network should be moved following the center of tropical cyclone. Kitade (1979) developed a method of numerical time integration using a moving variable grid. The grid movement is continuously carried out based on the Lagrangian concept and, therefore, the shock noise does not occur by the movement of grid. In this article the method is applied to a 4-layer primitive equation model of the tropical cyclone.

It is well-known that the heating by cumulus convection plays an essential role in the development of tropical cyclones. However, we don't know whether the heating of cumulus convection has a crucial effect on the movement of tropical cyclone or not. In Ookochi's model (1978) the heating effect by cumulus convection was given by a circularly symmetric function. Mathur (1974), Madala and Piacsek (1975), Hovermale (1976) and Jones (1977b) have adopted some variations of Kuo's parameterization scheme of cumulus convection. From such previous works we can not clearly obtain the insight for the influence of cumulus activity to the movement of tropical cyclone.

In the present work the behaviors of tropical cyclone are studied under a simplified situation without basic current to isolate the influence of cumulus heating on the movement of vortex. The behavior of tropical cyclone with a variable Coriolis parameter is reexamined with a different parameterization scheme of cumulus convection from those by Madala and Piacsek (1975) and Jones (1977c). Then we carry out some numerical experiments by changing the sea surface temperature, which severely controls the activity of cumulus convection. A divergent barotropic model is also integrated with a similar initial condition and the same horizontal grid net as the baroclinic model. The paths of vortices in such experiments are compared one another. Those comparisons may make the effect of the cumulus convection on the path of vortex clear.

2. The numerical model

2.1 Basic equations

Since the non-hydrostatic convective scale of motion is parameterized in this model, the primitive set of equations appropriate to the study of tropical cyclones may be written in Cartesian coordinates using the mercator projection,

$$\frac{Du}{Dt} = -\omega \frac{\partial u}{\partial p} + fv - mg \frac{\partial z}{\partial x} + E_x + F_{uh} + F_{up} \quad (1)$$

$$\frac{Dv}{Dt} = -\omega \frac{\partial v}{\partial p} - fu - mg \frac{\partial z}{\partial y} + E_y + F_{vh} + F_{vp} \quad (2)$$

$$\frac{D\theta}{Dt} = -\omega \frac{\partial \theta}{\partial p} + Q + F_{\theta h} + F_{\theta p} \quad (3)$$

$$\frac{Dq}{Dt} = -\omega \frac{\partial q}{\partial p} + M + F_{qh} + F_{qp} \quad (4)$$

$$\frac{Dz_0}{Dt} = \omega_0 \frac{R}{pg} \theta_v \left(\frac{p}{p_0} \right)^{R/c_p} \quad (5)$$

$$g \frac{\partial z}{\partial p} = -\frac{R}{p} \theta_v \left(\frac{p}{p_0} \right)^{R/c_p} \quad (6)$$

$$\frac{\partial \omega}{\partial p} = -m \left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right) \quad (7)$$

$$\theta_v = T(1 + 0.61q) \left(\frac{p_0}{p} \right)^{R/c_p} \quad (8)$$

where

$$\frac{D}{Dt} = \frac{\partial}{\partial t} + mu \frac{\partial}{\partial x} + mv \frac{\partial}{\partial y}$$

and (u, v) is wind velocity, z height, (x, y) horizontal coordinate, p pressure, T temperature, θ potential temperature, θ_v virtual potential temperature, q mixing ratio of water vapor, ω vertical p -velocity, Q heating rate by phase change of water, M rate of accession of moisture by the phase change, (E_x, E_y) momentum transport by cumulus convection, $(F_{uh}, F_{vh}, F_{\theta h}, F_{qh})$ horizontal diffusion, $(F_{up}, F_{vp}, F_{\theta p}, F_{qp})$ vertical diffusion, R gas constant, c_p specific heat at constant pressure, f Coriolis parameter, m map factor and g acceleration of gravity. The subscript 0 represents the value at 1,000 mb.

Eqs. (1) and (2) are the equations of motion in the east and north directions, respectively. Eq. (3) is the first law of thermodynamics and (4) is a water vapor conservation equation. Eq. (5) is the tendency equation for height of the 1,000 mb surface. Eqs. (6) and (7) represent the hydrostatic equation and the continuity equation, respectively.

2.2 Grid system

As shown in Fig. 1, the numerical model contains four layers in the vertical. The lowest layer is the boundary layer, which is regarded as the mixed layer. The remaining three layers represent the free atmosphere in which the cumulus convection may develop. The variables are staggered in the vertical direction. The values of u, v and z appear at 1,000, 900, 650, 400 and 150 mb,

while θ, q and ω appear at intermediate levels of 950, 775, 525 and 275 mb. This staggering simplifies vertical integration of the hydrostatic and continuity equations. The difference formulae for the vertical advection terms are described in Appendix.

The horizontal grid mesh at the initial time is represented in Fig. 2. The abscissa and ordinate are longitude and latitude, respectively. All variables are defined at the same point. The configuration of the mesh is continuously changed on the Lagrangian concept so as to locate the center of simulated vortex on the high resolution part of mesh throughout time integration. This moving grid scheme was described in detail by Kitade (1979). The velocity of grid movement V_g is determined as follows:

$$V_g = \bar{V} + V_m$$

where $\bar{V}$ is a space averaged velocity in the domain $S = \{|x - x_c| < 750 \text{ km}, |y - y_c| < 750 \text{ km}, 1,000 \text{ mb} > p > 150 \text{ mb}\}$ around the vortex center

P	VARIABLE
mb	
25	----- $\omega - q - \theta, \theta = \theta_0$
150	----- u, v, z
275	----- θ, ω, q
400	----- u, v, z
525	----- θ, ω, q
650	----- u, v, z
775	----- θ, ω, q
900	----- u, v, z
950	----- θ, ω, q
1000	----- u, v, z, ω

Fig. 1 Vertical structure of the model.

Fig. 2 Initial grid network. The abscissa and ordinate denote longitude and latitude, respectively.

(x_c, y_c) and V_m is given to restore the coincidence between the centers of vortex and high resolution mesh during following 180 time steps if the difference appears.

This variable grid net consists of the minimum grid interval $\Delta\lambda = \Delta\theta' = 0.5^\circ$ ($\Delta x \approx 50$ km) and the maximum one is given by $\Delta\lambda = \Delta\theta' = 3^\circ$, where $\Delta\lambda$ and $\Delta\theta'$ are increments of longitude and latitude, respectively. This resolution is insufficient to represent the detail of tropical cyclone, in particular, such as the eye and spiral rain bands. However, it may give some useful insights concerning the asymmetry and movement of tropical cyclones.

2.3 The time integration and lateral boundary conditions

A quasi-Lagrangian advective scheme (Mathur, 1970) is used for the time integration of the basic equations. The detail of the time integration scheme using the moving variable grid was described in Kitade (1979). A time step of 100 sec is used throughout the integration.

The lateral boundary conditions are set to be cyclic at western and eastern boundaries ($x=0$, $x=L_x$) and $v=0$ at the southern and northern boundaries ($y=y_0$, $y=L_y$). The influence of boundary conditions may be small since the domain of integration is sufficiently large, compared with the size of vortex.

2.4 Initial conditions

We set a basic state by the horizontally uniform potential temperature θ_p in unit of $^\circ\text{K}$ and relative humidity r_p in unit of %.

$$\theta_p = \begin{pmatrix} \theta_{25} \\ \theta_{275} \\ \theta_{525} \\ \theta_{775} \\ \theta_{950} \end{pmatrix} = \begin{pmatrix} 380.0 \\ 335.7 \\ 318.6 \\ 303.3 \\ 300.4 \end{pmatrix}, \quad r_p = \begin{pmatrix} r_{25} \\ r_{275} \\ r_{525} \\ r_{775} \\ r_{950} \end{pmatrix} = \begin{pmatrix} 0 \\ 60 \\ 60 \\ 60 \\ 95 \end{pmatrix}.$$

This vertical profile of potential temperature is a little more unstable than that of mean soundings over tropical Pacific ocean in September (Bell and Kar-sing, 1973). An initial vortex is given by the stream function in unit of $\text{m}^2 \text{sec}^{-1}$,

$$\phi = \frac{4.5 \times 10^{17} \phi_p \sin \pi(y-y_0)/(L_y-y_0)}{(x-x_c)^2 + (y-y_c)^2 + 9 \times 10^{10}},$$

$$\phi_p = \begin{pmatrix} \phi_{150} \\ \phi_{400} \\ \phi_{650} \\ \phi_{900} \\ \phi_{1000} \end{pmatrix} = \begin{pmatrix} -0.4 \\ 1.1 \\ 1.6 \\ 1.8 \\ 1.8 \end{pmatrix} \quad (10)$$

The vortex center (x_c, y_c) is put on the center of fine grid network (140° , 30°).

The non-divergent part of velocity (u_ϕ , v_ϕ) and vorticity ζ are as follows,

$$u_\phi = -m \frac{\partial \phi}{\partial y}, \quad v_\phi = m \frac{\partial \phi}{\partial x}, \quad \zeta = \nabla^2 \phi$$

The geopotential height z at each level is obtained by the balance equation.

$$g \nabla^2 z = f \zeta - \frac{\partial f}{\partial y} \frac{u_\phi}{m} - 2 \left(\frac{\partial v_\phi}{\partial x} \cdot \frac{\partial u_\phi}{\partial y} - \frac{\partial u_\phi}{\partial x} \cdot \frac{\partial v_\phi}{\partial y} \right) \quad (11)$$

The lateral boundary condition for z is $z=0$ at $p=1,000$ mb. For other levels it is obtained from the hydrostatic relation using the values of θ_p and r_p .

The virtual potential temperature θ_v is obtained by use of the hydrostatic equation, i.e.,

$$\theta_v = -\frac{p g}{R} \left(\frac{p_0}{p} \right)^{R/c_p} \frac{\partial z}{\partial p}$$

The mixing ratio q and the potential temperature θ at each level are obtained from

$$q = \frac{r q_s}{100}, \quad \theta = \frac{\theta_v}{1 + 0.61 q} \quad (12)$$

A simple ω -equation, similar to the quasi-geostrophic equation except for inclusions of convective heating rate Q and the map factor m and variable Coriolis parameter, is used to obtain the initial field of the divergent wind.

$$-\frac{R}{p} \left(\frac{p}{p_0} \right)^{R/c_p} \frac{\partial \bar{\theta}}{\partial p} \nabla^2 \omega + \frac{f^2}{m^2} \frac{\partial^2 \omega}{\partial p^2} = f \frac{\partial}{\partial p} \left\{ u_\phi \frac{\partial}{\partial x} \left(\frac{\partial v_\phi}{\partial x} - \frac{\partial u_\phi}{\partial y} \right) + v_\phi \frac{\partial}{\partial y} \left(\frac{\partial v_\phi}{\partial x} - \frac{\partial u_\phi}{\partial y} \right) + \frac{\partial f}{\partial y} \frac{v_\phi}{m} \right\} - \frac{R}{p} \left(\frac{p}{p_0} \right)^{R/c_p} \nabla^2 \left(-m u_\phi \frac{\partial \theta}{\partial x} - m v_\phi \frac{\partial \theta}{\partial y} + Q \right) \quad (13)$$

We use the boundary condition

$$\omega = 0 \quad \text{at } p = 25 \text{ mb}$$

$$\omega = \omega_{950} \quad \text{at } p = 950 \text{ mb}$$

The frictionally induced vertical velocity at 950 mb, ω_{950} , is given by

$$\omega_{950} = -\frac{\Delta p g \rho C_D m}{f} \times \left\{ (u_\phi^2 + v_\phi^2)^{1/2} \left(-\frac{\partial v_\phi}{\partial x} - \frac{\partial u_\phi}{\partial y} \right) \right\}_0 \quad (14)$$

where we have $\Delta p = 50$ mb. The heating rate

Q in Eq. (13) is given by use of the cumulus parameterization scheme described in section 3. If the term Q is not contained in Eq. (13) and the boundary value of ω is set to be zero, the value of ω is nearly zero at all points in our initial condition. We expect that the excitation of gravity waves may be reduced by the inclusion of the term Q . The divergent part of the wind is obtained from the velocity potential χ , which satisfies the following relation;

$$m\nabla^2\chi = -\frac{\partial\omega}{\partial p} \quad (15)$$

The boundary condition for χ is

$$\begin{aligned} \chi &= 0 \quad \text{at } x=0, L_x \\ &\quad \text{at } y=y_0, L_y \end{aligned}$$

The initial wind field is obtained as a sum of the non-divergent and the divergent part of the wind.

Fig. 3 Streamlines and isotachs at $p=1000$ mb and $p=150$ mb for the initial time. The solid line and dashed line represent the streamlines and isotachs in unit of m sec^{-1} , respectively. The positions of grid are assigned on the bottom and left hand side.

$$u = u_\phi + \frac{\partial\chi}{\partial x}, \quad v = v_\phi + \frac{\partial\chi}{\partial y} \quad (16)$$

Fig. 3 shows the streamlines and isotachs at the initial time by the solid and dashed lines, respectively.

3. Parameterization of cumulus convection

Latent heat released is the primary source of energy for the development and maintenance of tropical cyclones. The release of latent heat of condensation in a conditionally unstable atmosphere is accomplished by the vertical motions associated with deep convective clouds and not through the mean vertical motion.

We adopt the following requirements for the parameterization scheme of cumulus convection.

1. The approximate energy budget equation $d/dt(C_p T + Lq) = 0$ is not violated through the inclusion of the effect of cumulus convection.

2. The heating rate by cumulus convection is assumed to be nearly proportional to the vertical p -velocity ω^* at the top of mixed layer.

3. We consider that the heating by cumulus convection should be larger under the unsaturated situation of the free atmosphere than that in Kuo's scheme (1965). In Kuo's scheme the converged moisture seems to be excessively consumed for the moistening of the atmosphere. The additional conversion process from moisture to heating should be introduced.

4. The cumulus activity is assumed to depend upon the static stability and, furthermore, the vigorous cumuli take moisture from the mixed layer, which is called 'suction effect'.

The cumulus convection is assumed to exist if the vertical velocity ω^* is negative at the top of boundary layer and if $T_c - T$ is positive, where the cloud temperature T_c is assumed to be equal to the temperature which an air parcel would have if it ascends moist-adiabatically from the mixed layer. The cloud base p_B is assumed to be 900 mb. The cloud top p_T is determined as the highest level which $(T_c - T)$ is positive.

The effect of cumulus convection is parameterized as follows:

$$\begin{aligned} Q &= \omega^* \frac{\partial\theta}{\partial p} + (H_1 + H_2 + H_3) \\ &\times \left\{ \frac{c_p}{L} \int_{p_B}^{p_T} (T_c - T) dp + \int_{p_B}^{p_T} \beta(q_s - q) dp \right\} \\ &\times \left(\frac{p_0}{p} \right)^{R/c_p} \quad (17) \end{aligned}$$

$$M = \omega^* \frac{\partial q}{\partial p} + (H_1 + H_2 + H_3) \times \left\{ \frac{c_p}{L} \int_{p_B}^{p_T} (T_c - T) dp + \int_{p_B}^{p_T} \beta(q_s - q) dp \right\} - H_3 \quad (18)$$

$$E = \gamma \omega^* (V_c - V) / (p_T - p_B) \quad (19)$$

$$H_1 = - \int_{p_B}^{p_T} \omega^* \left\{ \frac{\partial q}{\partial p} + \frac{\partial \theta}{\partial p} \left(\frac{p}{p_0} \right)^{R/c_p} \frac{c_p}{L} \right\} dp \quad (20)$$

$$H_2 = -\alpha \omega^* q_{950} I \quad \text{for } I < 1 \\ = -\alpha \omega^* q_{950} \quad \text{for } I \geq 1 \quad (21)$$

$$I = \frac{1}{\Delta T (p_T - p_B)} \int_{p_B}^{p_T} (T_c - T) dp \quad (22)$$

$$V_c = \frac{1}{p_T - p_B} \int_{p_B}^{p_T} V dp$$

where L is latent heat, q_s saturation mixing ratio, q_{950} mixing ratio at 950 mb and $\Delta T = 5^\circ\text{C}$.

In this formulation it is assumed that a part of effect of cumulus convection is represented by the cumulus mass flux formulation by Ooyama (1971) and Arakawa and Schubert (1974) and the cloud mass flux is given by $-\omega^*/g$. Then the first terms in Eqs. (17) and (18) are interpreted as a virtual source or sink when the large-scale budget is formulated in terms of the mean vertical velocity. The terms give apparent heating and drying effects in cloud region.

The term H_1 is introduced to satisfy the requirement (1). The first terms in Eqs. (17) and (18) imply that the moisture supplied from the boundary layer to the clouds

$$-\omega^* q^* = \int_{p_B}^{p_T} \omega^* \frac{\partial q}{\partial p} dp$$

is condensed and the heating

$$\int_{p_B}^{p_T} \omega^* \frac{\partial \theta}{\partial p} dp$$

occurs in cloud. The difference of these quantities H_1 is consumed to raise the temperature and moisture in cloud, whose effect is included in the second terms in eqs. (17) and (18).

Furthermore we assume that the deep cumulus cloud has a root in the boundary layer and take the moisture into the cloud by strong cumulus updraft. The moisture taken into cloud from the mixed layer H_2 depends on the activity of cumulus convection, which is assumed to be proportional to the vertical velocity at the top of the boundary layer and the stability factor I .

If the difference between in-cloud and environmental temperatures is zero, H_2 is zero because the cumulus activity vanishes. Though H_2 increases with the increase of the difference $(T_c - T)$, a limitation is set at the vertically averaged temperature difference 5°C . This assumption is introduced to take account of the increase of entrainment with the increase of $(T_c - T)$. The amount of moisture H_2 is subtracted from the boundary layer, that is, $M_B = -H_2/\Delta p$. This may be regarded as the drying effect in the mixed layer by cumulus-induced subsidence. We call the process "convective suction" because the cumulus convections take out the moisture from the mixed layer by cumulus updraft.

The water vapor over the saturation value may gather in the neighborhood of center of disturbance. We assume that the excess moisture H_3 is consumed to raise the temperature and moisture in cloud. The effect is contained in the second terms in Eqs. (17) and (18) and the last term in Eq. (18).

The scheme for the parameterization of cumulus convection by Kuo (1965) is mainly modified in two points in order to satisfy the requirement (3). One is an introduction of the first terms in Eqs. (17) and (18). The other is an introduction of $\beta(q_s - q)$ in place of $(q_c - q)$, where q_c is mixing ratio in cloud and β is given as follows.

$$\beta = 0 \quad \text{for } T_h \geq 0.8 \\ = 5(0.8 - T_h) \quad \text{for } T_h < 0.8 \quad (23)$$

$$T_h = \frac{\int_{p_B}^{p_T} q dp}{\int_{p_B}^{p_T} q_s dp} \quad (24)$$

The quantity in the large parentheses in Eqs. (17) and (18) gives the partitioning of the condensed water vapor $(H_1 + H_2 + H_3)$ into the heating and moistening. When the environmental air is dry, the condensed moisture in cloud re-evaporates and the environment is cooled, that is, the partition to the moistening increases against the heating. From the qualitative consideration the modification factor β is given.

The cumulus convection transports the horizontal momentum in the vertical direction. Even though the horizontal scale of cumulus convection is much smaller than that of the large-scale circulation, the vertical scales of two phenomena are almost the same. Therefore we cannot regard the cumulus convection as small-scale eddy as Peng and Kuo (1975) did. Gray (1967) and

Matsumoto, Ninomiya and Akiyama (1967) showed that a nearly uniform vertical profile of wind was often found in the area with vigorous cumulus activity. The horizontal momentum of the large-scale circulation seems to be exchanged in the vertical column by the cumulus ensemble. We assume a simple formulation (19) for the effect of the momentum transport by cumulus convection. The effect makes the horizontal wind uniform in the vertical direction and does not change the total horizontal momentum in the vertical column.

Ooyama (1964) and Syono and Yamasaki (1966) have studied the linear stability problem of tropical cyclone by giving the heating rate of cumulus heating as follows.

$$Q = -\eta \frac{\partial \theta}{\partial p} \omega^*$$

where η is a non-dimensional heating parameter. The typhoon-like disturbance develops when η is larger than 1. We have at the initial stage in our experiment

$$\eta \sim 1 + \alpha$$

Then we adopt $\alpha = 0.5$ in our experiment. As for the value of γ , we have no appropriate information. When the convective momentum transport is assumed to be nearly equal to the momentum flux by large-scale upward motion in case of the linear vertical shear, we have

$$\omega \frac{\partial V}{\partial p} = 2\omega \frac{V_c}{p_B - p_T} \sim \gamma \omega^* \frac{V_c - V}{p_B - p_T}$$

Thus we adopt $\gamma = 2$.

4. Other physical processes

4.1 Subgrid-scale diffusion and surface friction

The horizontal and vertical exchange by sub-grid scale eddies are (written in terms of a general variable S):

$$\begin{aligned} F_{sh} &= K_h m^2 \left(\frac{\partial}{\partial x^2} + \frac{\partial}{\partial y^2} \right) S \\ F_{sp} &= K_p \frac{\partial^2 S}{\partial p^2} \end{aligned} \quad (25)$$

The coefficients of horizontal and vertical diffusion in our experiment are given as follows.

$$\begin{aligned} K_h &= \text{Max}(2 \times 10^5, D \times 10^9) \text{ m}^2 \text{ sec}^{-1} \\ K_p &= 10^{-2} \text{ mb}^2 \text{ sec}^{-1} \end{aligned} \quad (26)$$

where

$$D = \left\{ \left(\frac{\partial u}{\partial x} - \frac{\partial v}{\partial y} \right)^2 + \left(\frac{\partial v}{\partial x} + \frac{\partial u}{\partial y} \right)^2 \right\}^{1/2}$$

The non-linear viscosity scheme is adopted for the vigorous wind field after Smagorinsky (1963). The value of $2 \times 10^5 \text{ m}^2 \text{ sec}^{-1}$ may be regarded as rather large for the smallest grid interval 50 km or as rather small for the largest grid interval 300 km. The introduction of dependence of K_h on grid interval may be preferable for the variable grid system. The use of the constant value was, however, guided by numerical experimentation.

At 1,000 mb, the vertical stress terms in Eqs. (1) and (2) are

$$\begin{aligned} F_{up} &= -C_D \rho |V_0| u_0 \frac{g}{\Delta p} \\ F_{vp} &= -C_D \rho |V_0| v_0 \frac{g}{\Delta p} \end{aligned} \quad (27)$$

where ρ is air density and $V_0 = (u_0, v_0)$ wind velocity at 1,000 mb. A constant value of 0.0034 for the drag coefficient C_D is assumed.

4.2 Air-sea exchanges

Sensible heat and latent heat fluxes are expressed by the following relations, respectively.

$$\begin{aligned} F_s &= 3.29 \times 10^{-2} (T_w - T_a) |V_0| \frac{g}{\Delta p} \\ &\quad \text{mb} \cdot \text{m} \cdot \text{sec}^{-1} \\ F_l &= 1.71 \times 10^{-5} (q_w - q_a) L |V_0| \frac{g}{\Delta p} \\ &\quad \text{mb} \cdot \text{m} \cdot \text{sec}^{-1} \end{aligned} \quad (28)$$

where T_w is sea surface temperature, T_a surface air temperature, q_w saturation mixing ratio at sea surface temperature and q_a mixing ratio of air at sea surface. The values of F_s and F_l are given at the 1,000 mb surface and $\Delta p = 100$ mb. The values of K_h , K_p , C_D and coefficients of F_s and F_l are referred to the tropical prediction model by Krishnamurti *et al.* (1973).

The surface air temperature T_a are determined under the assumption that the boundary layer is a mixed layer, i.e.,

$$\frac{\partial \theta}{\partial p} = 0$$

Then we have

$$\begin{aligned} P_a &= \left(\frac{P_s'}{1000} \right)^{R/c_p} \theta_{950} \\ P_s' &= \text{max}(p_0 + \rho_0 g z_0, 970) \end{aligned} \quad (29)$$

The surface air mixing ratio q_a is obtained under the assumption that the relative humidity at the surface is the same as that at 950 mb. The

formulation of surface values plays an essential role for the development of the simulated disturbance. The deepening of the surface pressure brings the lowering of surface air temperature through Eq. (29) and the increase of supply of sensible heat from the sea through Eq. (28). Then the relative humidity at 950 mb has a tendency to decrease by the warming effect. It causes the increase of moisture supply from the sea since the relative humidity at the surface is assumed to be the same as that at 950 mb. The limitation for P_s' suppresses further increase of the sensible heat flux from the sea and, therefore, further development of the disturbance. We adopt the limitation value of 970 mb for P_s' to avoid larger truncation errors which might increase if a deeper depression is represented with the coarse grid in our model.

4.3 Large-scale condensation

Latent heat is also released by the large-scale condensation process in addition to that by cumulus convection. The warming rate ΔT by condensation and the condensed moisture Δq are obtained from the following two relations:

$$c_p \Delta T + L \Delta q = 0$$

$$q_s(T + \Delta T) \doteq q_s'(T) + \left(\frac{\partial q_s}{\partial T} \right)_T \Delta T = q + \Delta q$$

4.4 Dry convective adjustment

In long term numerical prediction, removal of super-adiabatic lapse rates is desirable to avoid small scale gravitational instabilities. Therefore the process of dry convection is parameterized in our model. If the relation $-\partial \theta_v / \partial p < 0$ holds, we adjust the potential temperature field between the neighboring two levels ($i, i+1$) to satisfy the relation $\partial \theta_v / \partial p = 0$, but for convenience, we do not adjust the moisture field. Assuming that the total potential temperature is conserved through the dry convective adjustment process, we have

$$\theta_i = \bar{\theta} - \delta$$

$$\theta_{i+1} = \bar{\theta} + \delta$$

$$\bar{\theta} = \frac{0.61 \bar{\theta} (q_i - q_{i+1})}{2(1 + 0.61 \bar{q})}$$

$$\bar{\theta} = (\theta_i + \theta_{i+1})/2.0$$

$$\bar{q} = (q_i + q_{i+1})/2.0$$

This procedure is iteratively repeated from the lowest layer to the upper layer so that the situation $-\partial \theta_v / \partial p < 0$ may disappear at all levels.

5. Time variation and path of simulated vortex

We carried out three cases of numerical experi-

Fig. 4 The distribution of sea surface temperature T_w in cases I, II and III.

ments with variations of sea surface temperature. Fig. 4 shows the sea surface temperature profiles in three cases I, II and III. The basic experiment I was carried out with uniform sea surface temperature 27.4°C. In order to examine the influence of sea surface temperature to the motion of vortex, the cases II and III are carried out by changing the sea surface temperature from the state at 30 hours in the basic experiment I. In cases II and III the sea surface temperature without latitudinal variation is fixed after 30 hours. Fig. 5 shows the time variations of the minimum surface pressure and the maximum wind speed in cases I, II and III. The solid, dashed and dash-dotted lines represent the cases I, II and III, respectively. The dotted line shows the results from the barotropic model, which is described in Appendix B. It is seen in the case I that the lowest surface pressure is 955 mb at Hour 52. The maximum wind speed at the surface attains an extreme of 44.5 m/sec at Hour 60. The tendency of decreasing intensity of tropical cyclone after rapid development is similar to those by Ooyama (1969) and others. The development is slower in case II than that in the basic experiment I since the sea surface temperature is lower in case II. Therefore we can know the influence of strength of vortex to the movement of vortex by comparing case II with case I. There is little difference of development between cases I and III. However the horizontal variation of sea surface temperature induces an asymmetry of vortex, which may affect the movement of vortex. Fig. 6 shows the paths of vortex center, which is defined as a position with the

Fig. 5(a) The time variation of the minimum surface pressure in cases, I, II, III and B.
(b) The time variation of the maximum wind speed in cases I, II, III and B.

maximum vorticity. We can obtain some informations about the influence of sea surface temperature to the motion of vortex from the comparison among the results.

Anthes and Hokes (1975) and Jones (1977c) pointed out that the northward force due to the variable Coriolis parameter induces the mean northward motion of vortex. Kitade (1980) also carried out the numerical time integration of the barotropic model. He has shown that the circulation in the vortex on a plane with the variable Coriolis parameter causes a mean wind ($\bar{U}$, $\bar{V}$) and the cyclonic vortex moves northward. The westward movement is also induced through such process. The movement depends upon the size and strength of vortex. The larger the size and the stronger the cyclonic vortex, more increases

the north-westward velocity. Since the sea surface temperature is a controlling factor of the development of tropical cyclone, the movement of tropical cyclone may be affected through the changes of strength and size of vortex by the sea surface temperature.

A mean velocity around the vortex is defined as follows.

$$\bar{U} = \frac{1}{100} \int_{\theta_0-5}^{\theta_0+5} \int_{\lambda_0-5}^{\lambda_0+5} u \, d\lambda \, d\theta$$
$$\bar{V} = \frac{1}{100} \int_{\theta_0-5}^{\theta_0+5} \int_{\lambda_0-5}^{\lambda_0+5} v \, d\lambda \, d\theta$$

where λ_0 and θ_0 represent the longitude and latitude at the position of vortex center. Fig. 7 shows ($\bar{U}$, $\bar{V}$) at each level as a function of time in case I. Fig. 8 shows (U , V) at 650 mb in cases

Fig. 6 The path of vortex center in cases I, II, III and B. The numeral denotes the time in unit of hour after the start of time integration.

Fig. 8 The mean wind (U , V) at 650 mb as a function of time in cases I, II and III. The barotropic case B is denoted by the dotted line.

Fig. 7 The mean wind (U , V) as a function of time at each vertical level in case I.

I, II and III. The barotropic case is denoted by the symbol B. The correspondence of vortex motion to ($\bar{U}$, $\bar{V}$) in the lower layers is remarkable, comparing Fig. 6 with Figs. 7 and 8. It is noticed that the mean wind at the top level has the inverse direction against vortex motion. Jones (1977c) has supposed that the north-westward deflection at the mature stage of tropical cyclone is due to the mean flow at the upper outflow layer. The concept of outflow layer steering current does not hold in our case.

In the case of Madala and Piacsek (1975) the northward motion was much smaller than that in our case and the westward displacement exceeded northward one. It is due to the differences of size and strength between the simulated vortices, as pointed out by Jones (1977c). Kitade (1980)'s experiments have given a confirmation for this reasoning.

Jones (1977c) have employed an initial vortex with almost the same amplitude as ours. He gained a path of vortex similar to that in case I.

The northward displacements for 80 hours are about 450 km in Jones's experiment and 480 km in our case I, respectively. The westward component is much smaller than the northward one in the developing stage and the northwestward deflection occurs at the mature stage in both experiments. This deflection may be due to the increases of strength and size of vortices at the mature stage. In Jones's experiment the strength of tropical cyclone is maintained and the northwestward motion continues after the attainment of the mature stage. On the other hand, the tropical cyclone in case I gradually decays and, thus, the westward movement decreases after the mature stage.

Our result also shows a meandering of tropical cyclone. The periods of meandering are about 25 hours in case I and 17 hours in case B. The oscillation is not regular in cases II and III, perhaps, due to the sudden change of sea surface temperature at Hour 30. The meandering with a period of about 20 or 15 hours appeared in the numerical experiments with the mean current by

Madala and Piacsek (1975) and Jones (1977b). Yeh (1950) examined analytically the motion of a rotating cylinder in the mean current. He found that the cylinder undergoes an oscillatory motion about its mean path, and the period of this oscillation T' is given by $2\pi/\bar{\omega}$, where $\bar{\omega}$ is an angular velocity of rotating cylinder. When the value of T' is estimated for the vortex in our experiment using the relation $\bar{\omega}=\bar{\zeta}/2$, we have $T'=17$ hours for $\bar{\zeta}=2\times10^{-4}\text{ sec}^{-1}$, where $\bar{\zeta}$ is an averaged vorticity.

Furthermore the convergence and divergence strengthened by cumulus convection may affect the movement of the vortex. Therefore we examine the relation between the movement of vortex and the distribution of rainfall amount, which implies the existence of an averaged convergence in the lower layers. Figs. 9, 10 and 11 represent the distribution of the rainfall amount in cases I, II and III, respectively.

5.1 Comparison between cases I and B

It is seen from Fig. 6 that there are large

Fig. 9 The distribution of rainfall intensity in unit of mm/hour by cumulus convection in case I. The area surrounded by the dashed line shows the appearance of the large-scale condensation. The abscissa and ordinate represent the longitude and latitude. The vortex center is denoted by the symbol s.

Fig. 10 Same as Fig. 9 except for case II.

Fig. 11 Same as Fig. 9 except fro case III.

differences between paths of two cases until 30 hours and after 50 hours. The eastward meandering until 20 hours is larger in case I than that in case B. We can not understand it from the mean wind variation in Fig. 8. We speculate that it is due to an initial unbalance excited by cumulus heating in the basic experiment I. After 50 hours the north-westward velocity increases in case I. This is due to the increase of strength and size of vortex in case I. The westward movement in case I seems to be mainly due to the increase of vortex size in the mature stage of tropical cyclone. There is also a possibility that the westward movement after 50 hours in case I is affected by the increasing rainfall amount in the western part of vortex.

5.2 Comparison between cases I and II

The north-westward velocity of vortex motion is smaller after 50 hours in case II than that in case I. This is accounted from the smallness of intensity and size of the vortex in case II. It is noticed that the asymmetry of rainfall pattern is smaller in case II before 60 hours than in case I.

5.3 Comparison between cases I and III

The strength of vortex is almost the same each other. However the path in case III is east side of case I. We can interpret the difference of paths as a result of the difference of the distributions of cumulus heating. Fig. 11 shows that the amount of heating is larger in the east side than in case I. The difference of path between cases I and III is consistent with that of the mean wind in Fig. 8. Therefore, we may say that the difference of the distribution of cumulus heating also causes that of mean wind and indirectly changes the path of vortex.

The meandering may be also influenced by the asymmetry of simulated disturbance. The maximum position of rainfall amount rotates cyclonically in cases I and III. However, its rotation is not synchronized with the meandering of path.

6. Structure of the simulated disturbance

We describe the structure of the simulated disturbance at Hour 70 in case I. The structures of disturbances in cases II and III are similar to that in case I. Fig. 12 shows the height fields at the upper level $p=150$ mb. The asymmetry at the upper level is more remarkable than that at the lower levels. The center of the upper high is to the west of the low-level vortex center indicated by the symbol s .

Fig. 12 Height field at $p=150$ mb in unit of m . The position of the minimum surface pressure is denoted by the symbol s .

Fig. 13 Vorticity field at $p=150$ mb in unit of 10^{-5} sec^{-1} .

Fig. 13 shows the vorticity pattern at the upper level ($p=150$ mb). The cyclonic vorticity dominates at the lower levels (not shown). The asymmetry of vortex is more remarkable at the upper level than those at the lower levels. The negative minimum value of vorticity appears in the north-east quadrant at the upper level. The criterion for inertial instability is given by

$$\zeta_a(f+2V_\theta/R') < 0$$

where ζ_a , V_θ and R' are absolute vorticity, tangential velocity and radius of curvature, respectively. The values of ζ_a and $(f+2V_\theta/R')$ at 150 mb are represented in unit of 10^{-5} sec^{-1} by the dashed and solid lines, respectively, in Fig. 14. The inertially unstable area is shaded, which can be found only at the upper level. Since

the gradient of the absolute vorticity vanishes at some place, the barotropic instability may also occur. Such dynamical instabilities may excite

Fig. 14 $(f+2V_0/R')$ and absolute vorticity ζ_a in unit of 10^{-5} sec^{-1} are shown by the solid and dashed lines, respectively. The shaded area is inertially unstable.

Fig. 15 Same as Fig. 3 except for $t=70$ hours.

a local circulation. However, it is not clearly perceptible in the flow pattern. Fig. 15 shows the streamlines and isotachs by the solid and dashed lines, respectively. The cyclonic circulation dominates except at the upper layer, where anticyclonic flow dominates outside the radius 100 km. The momentum transport by cumulus convection contributes to the cyclonic circulation in the neighborhood of the vortex center at the upper layer. This was confirmed by an experiment without the momentum exchange by cumulus convection. The inflow is almost restricted to the lower levels and the outflow is to the upper level. The isotachs clearly represent the asymmetry of the velocity field. The maximum speed of wind appears on the east side at all levels.

Fig. 16 shows the vertical p -velocity in unit of $10^{-3} \text{ mb sec}^{-1}$. The eye of the tropical cyclone is not sufficiently simulated because of the coarseness of the grid net. However, the weak updraft in the center of vortex in Fig. 16 may

Fig. 16 Vertical p -velocity in unit of $10^{-3} \text{ mb sec}^{-1}$ at $p=950 \text{ mb}$ and $p=775 \text{ mb}$.

Fig. 17 Potential temperature field at $p=275$ mb in unit of $^{\circ}\text{K}$.

Fig. 18 Relative humidity at $p=775$ mb and $p=275$ mb.

be an indication of an eye. The remarkable up-draft region exists around the center of vortex, which may be regarded as a eyewall. The pattern is very similar to that of rainfall amount shown in Fig. 9. The remarkable asymmetry of these patterns may be regarded as an indication of inner spiral rain band. Such rain band cyclonically rotates around the vortex center at a speed of $1\sim 3$ m/sec. The speed is much smaller than the wind speed at the point in the lower layer.

The vertical velocity at 950 mb is weaker than that in the free atmosphere. The patterns and strengths of vertical velocity in the free atmosphere are almost same one another. The vast zone with the weak upward velocity appears in the outer area. It is a gravity wave. Diercks and Anthes (1976) and Kurihara and Tuleya (1974) examined the spiral bands that appeared in their hurricane simulations. They concluded that the bands behaved like gravity waves. Though present result gives a similar gravity wave, no outer spiral rain band is found. The insufficient appearance of spiral bands in our result may be due to either to the coarseness of the grid net or to the formulation of the parameterization scheme of cumulus convection.

The disturbance has a warm core structure at all levels. The largest warming takes place in the upper layer. The potential temperature deviations from the horizontal means are about 3, 8, 9 and 13°K for 950, 775, 525 and 275 mb, respectively. Fig. 17 represents the potential temperature fields at the upper level ($p=275$ mb). The warmer region extends to the west in the upper layers. Such an asymmetric pattern of potential temperature is consistent with that of height in the upper

layer as described in Fig. 12.

The relative humidity is presented in Fig. 18. The mid-levels are saturated only in a narrow core. In the upper layer, it is covered by larger saturated area corresponding to the advection of moist air by the outflow. The mixed layer is almost saturated over the whole area (not shown).

7. Summary

Some numerical experiments were carried out in order to examine the characteristics of tropical cyclones on a plane with variable Coriolis parameter. A moving variable grid scheme was applied to save the computation time. In the present preliminary experiment we examined the behavior of the vortex in the case without basic current. The simulated disturbance is similar to tropical cyclones observed in the real atmosphere and those simulated in numerical experiments by many authors. With the variable Coriolis param-

ter the simulated disturbance has pronounced asymmetric features in spite of the adoption of an initial nearly symmetric vortex. Such asymmetric features are summarized as follows.

1. The contour lines of height and temperature expand westward in the upper layer.

2. The vorticity field at the upper level has an area with the minimum value in the north-east quadrant, where the atmosphere is barotropically unstable. Inertially unstable areas also appear as the vortex develops.

3. The isotach field has the maximum value on the east-side at all levels, as in the case of Madala and Piacsek (1975) and Jones (1977c).

4. The maximum positions of the vertical velocity and rainfall amount rotate cyclonically round the storm center.

5. The relative humidity in the upper layer is lower in the northwest quadrant.

The simulated tropical cyclone moves north-north-westward due to the effect of the variable Coriolis parameter. The speed of vortex movement depends on the strength of the disturbance. The mean speed for the moderate tropical cyclone is about 3 km hour^{-1} for westward and 7 km hour^{-1} for northward components. Furthermore, the vortex meandered about the mean path with a period of about 25 hours.

When the sea surface temperature is changed, the path of vortex is influenced a little. The deviation of the path occurs mainly through the changes of the strength and size of vortex. It was confirmed by an experiment with non-uniform sea surface temperature that the asymmetric heating of cumulus convection affects the path of the vortex center.

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Appendix A

Difference scheme of vertical advective terms

Most of derivatives in the vertical advective terms are replaced by the following finite difference (written with a general variable S):

$$\left(\frac{\partial S}{\partial p}\right)_i = \left\{ \frac{S_{i-1} - S_i}{p_{i-1} - p_i} (p_i - p_{i+1}) + \frac{S_i - S_{i+1}}{p_i - p_{i+1}} (p_{i-1} - p_i) \right\} \frac{1}{p_{i-1} - p_{i+1}} \quad (\text{A-1})$$

where the suffix represents the level. The level number increases with decreasing pressure and p_i is defined at the same level as that of the variable S_i . As for $(\partial q / \partial p)$ the following formula is adopted to avoid the appearance of a negative value of q .

$$\left(\frac{\partial q}{\partial p}\right)_i = \frac{q_i - q_{i+1}}{p_i - p_{i+1}} \quad \text{for } \omega_i > 0$$

In case of $\omega_i < 0$ the formula (A-1) is utilized.

The upper and lower boundaries are treated as follows.

For u and v

$$\left.\frac{\partial S}{\partial p}\right|_{p=150 \text{ mb}} = \frac{1}{2} \left(\frac{S_5 - S_5}{p_5 - p_5} \right)$$

$$\left.\frac{\partial S}{\partial p}\right|_{p=1000 \text{ mb}} = \frac{1}{2} \left(\frac{S_1 - S_2}{p_1 - p_2} \right)$$

For θ and q

$$\left.\frac{\partial S}{\partial p}\right|_{p=950 \text{ mb}} = 0 \quad \text{for } \omega_1 < 0$$

$$\left.\frac{\partial S}{\partial p}\right|_{p=950 \text{ mb}} = \frac{2}{7} \left(\frac{S_1 - S_2}{p_1 - p_2} \right) \quad \text{for } \omega_1 > 0$$

The values of $\partial \theta / \partial p|_{p=275 \text{ mb}}$ and $\partial q / \partial p|_{p=275 \text{ mb}}$ are calculated with use of Eq. (A-1) by setting to be $\theta = 380^\circ \text{K}$ and $q = 0$ at $p = 25 \text{ mb}$.

Appendix B

The outline of the divergent barotropic model

The basic equations are given as follows.

$$\frac{Du}{Dt} = fv - mg \frac{\partial z}{\partial x}$$

$$\frac{Dv}{Dt} = -fu - mg \frac{\partial z}{\partial y}$$

$$\frac{D}{Dt}(\ln z) = -m \left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right)$$

The initial disturbance is given by the same stream function as Eq. (10) except for $\phi_p = 1.8$. The initial height field is obtained by use of Eq. (11) assuming $z = 3,000 \text{ m}$ at the northern and southern boundaries. The lateral boundary conditions, horizontal grid net and time integration scheme are the same as the three dimensional case described in the present paper.

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一様でないコリオリパラメーターを持つ平面上での 台風の数値実験

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北出(1979)によって提案された移動する可変格子計算スキムを使って台風の数値実験を行った。CISKの概念に基づいた積雲対流のパラメタリゼーションの方式が導入された。その他海面摩擦、渦散逸、渦拡散、潜熱顕熱の海面からの補給、乾燥対流調節、大規模凝結等の物理過程が慣行の方法で導入された。

初期に与えた渦は約2日で台風に似た構造を持つ擾乱に発達した。再現された台風はコリオリ係数が緯度によって異なるため、ある種の非対称性を持つことが見い出された。さらに発達した台風は最上層に力学的不安定領域を持っていた。

再現された台風は北北西に移動する。これは Rossby (1948) が指摘した β 効果によると考えられる。その移動スピードは台風の強さが強くなるほど速くなった。この台風は西方向にほぼ 3 km hour^{-1} 北方向にほぼ 7 km hour^{-1} で移動した。そしてこの平均経路のまわりに約25時間の周期を持って蛇行した。これは Yeh (1950) が指摘した Magnus 効果による様に思われる。

さらに海面温度を変えた実験を行なった所、台風の経路は少し変化した。これは海面温度の変化によって台風の強さが変化してその結果経路が変化したものと理解された。又水平方向に一様でない海面温度を与えた実験において積雲対流による加熱の非対称性が渦の移動に影響を与えることを確かめた。