Some Examples of Numerical Weather Prediction, with the Special Emphasis on the Development and Maintenance of Relatively Small Scale Cyclones

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Abstract

The non-adiabatic heating due to the latent heat released by the condensation of water vapor through the simply parameterized convective activity is incorporated in the quasi-geostrophic four-level numerical prediction model which is in operational use at Japan Meteorological Agency. This effect is so formulated that the heating takes place only over the cyclonic domain and the amount of heating is assumed to be proportional to the intensity of the relative vorticity at the lowest level. An air column resolved in the four level model is heated, at first over its lowest layer (700-900 mb) alone (Version B), secondly over its lower two layers (500-700 and 700-900 mb) at a certain distributive proportion of heating (Version C) and at the third over its whole three layers (300-500, 500-700 and 700-900 mb) at some rate of distribution of heat (Version D).

Comparing a few examples of numerical weather prediction by means of the non-adiabatic model mentioned above with the observed patterns and with results of the operational routine model which has no parameterized convective heating (Version A), we observe that

1. the relatively small scale cyclones and the cyclonic circulation of typhoons are well reproduced in surface prognostic charts by Versions B, C and D, though the predicted deepening of the disturbances does not cease, but that
2. the retardation of the predicted displacements of typhoons is common in all the versions, and also
3. when we cannot successfully forecast the upper level pattern, the inclusion of the heating exaggerates the error, and that
4. Version B distorts the vertical structure of the disturbances, that Version C well describes the development of cyclones in the lower troposphere and that Version D seems to be appropriate for simulating the behavior of typhoons.

1. Introduction

Recently, computational models of numerical weather prediction have been improved in several points. One of the significant improvements is incorporation of the effect of the non-adiabatic heating in the lower troposphere into the baroclinic model. (Gambo, 1963). A quasi-geostrophic four level model of Japan Meteorological Agency (JMA), which is now in routine operation, already undergoes the influence of the diabatic heating in the atmosphere such as the transport of the sensible heat from the warmer ocean surface and the release of the latent heat through the process of raining associated with a large scale disturbance. Results of recent forecast show successfully remarkable skill, for example, in predicting the movement and development of large scale cyclones, especially in winter, spring and autumn near Japan, and also in predicting the outburst of the continental anticyclone over the Asian coastal area, in comparison with the results of the adiabatic model used before.

However, mature typhoons or relatively small scale cyclones are not well followed by our operational model. As is well known,
mature typhoons frequently invade the forecast domain and attack the Japanese Islands in late summer and particularly in fall. The relatively small scale cyclones, on the other hand, often appear with heavy rain over the ocean off the Pacific coast of Japan in spring and autumn, and sometimes develop violently in short time. In our prognostic charts produced by means of the present operational model, the typhoons and the relatively small scale cyclones decay or disappear as the forecasting time advances.

The typhoons and the relatively small scale cyclones have their characteristic horizontal scale around 1000 km. As we use 304.8 km as a grid interval and these small scale depressions have rather larger positive vorticity in lower layer than in upper layer, we can easily presume that the discrepancy between the prediction and the observation is mainly due to the truncation error. Therefore the pressure pattern seen in the prognostic charts at the lower level looks more or less vague and crude in comparison with the actual pattern.

The supply of driving heat energy for a mature typhoon is due to the latent heat released by condensation of water vapor through convective cumulus clouds or hot towers (Riehl and Malkus, 1958). Accumulate various observational reports emphasize that the macro-motion such as the mature typhoon is considered as a cooperating system with the micro-motion such as the cumulus convection.

There has not appeared so far any satisfactory answer how we should parameterize and formulate the interaction between the micro-motion such as the convective cumulus clouds and the macro-motion such as the tropical cyclone. Concerning the formation of typhoons or hurricanes, however, Ooyama (1963), Charney and Eliassen (1964), Syōno et al. (1964)*, and Ogura (1964) have presented a model or an idea by treating a tropical depression as a forced circulation driven by the latent heat released in organized deep cumulus convections.

Unfortunately, on the other hand, there are little observational analyses as to the structure of the small extratropical lows and the physical mechanism of their intensive development. Almost all of these small lows are generated over the ocean, following approach of upper short waves. The rapid intensification of the small scale cyclones is, however, remarkably observed in the lower troposphere, whereas the upper trough does not develop so much. Reports on the radar observation, in recent days, say that some small scale cyclones of this type contain convective cumulus clouds within their systems (Imakado, 1965; Ishizuri, 1965).

Although its physical structure may be more or less different from the typhoon, the relatively small scale cyclone also develops presumably, at least in part, due to the non-adiabatic heating by means of the release of latent heat through the cumulus convection. The role played by the frictional surface boundary layer in the typhoon may be played by the warm wet tongue in the small low over the ocean off the Pacific coast of Japan.

The purpose of the present study is to improve the numerical prediction of the relatively small scale cyclones and of the cyclonic circulation of typhoons, in our operational quasi-geostrophic baroclinic model of JMA, by incorporating the effect of the latent heat released by the parameterized convective activity. Concerning the quasi-geostrophic approximation, we may need more exact scale considerations for every developing stage of the cyclone. However, as we are mainly concerned with the macro-motion which is indirectly influenced by transportation of heat through the microscale turbulent convection, we use *a priori* the quasi-geostrophic approximation for the large scale tropospheric motion (small Rossby number and large Richardson number) to prevent the prognostic chart for the macromotion from the distortion by noises (Ooyama, 1963; Charney and Eliassen, 1964; Ogura, 1964). As to the too large grid interval compared with the characteristic horizontal scale of the disturbance of our concern, we cannot but adopt it in our operational model because of the limited capacity of the electronic computer IBM 704 now in use at JMA.

The present author (1964) studied the simulation of the behavior of the relatively small scale cyclones and the typhoons near Japan, following the Ooyama's model, made numeri-

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* Private communication.
cal experiments to examine the role of the supply of the water vapor on the development of the disturbances. Okochi (1964)* made the preliminary application of the non-adiabatic heating of the same type to the quasi-geostrophic four level baroclinic model of JMA. Although he calculated only two cases and heated the lowest layer (700-900 mb) alone, he observed that
(1) the incorporation of the heating mentioned above markedly emphasized the existence of the small scale disturbances in the lower atmosphere, then
(2) the prognostic map at ground surface showed such a fine configuration as the actual one, but
(3) the influence of this heating action exaggerated the wrong results of forecast, when the prediction of the vorticity field was failed, because the heating is parameterized in proportion to the vorticity at the lowest level, and
(4) as the rate of the supply of the water vapor was fixed, cyclones did not cease to develop.

The third point may be related closely to the accuracy of forecasting the large scale motion at the upper level and the fourth point should be solved by a more advanced method of the parameterization for the cooperating system of the macro- and micromotions than that used here.

In this paper we extend the non-adiabatic heating of this type to the upper layers, and see extensively and in more detail the effect, of the heating applying it to the observed patterns, particularly in connection with the development of the relatively small scale cyclones and the maintenance of the cyclonic circulation in the typhoons which invades near Japan.

2. Modelling assumptions

The computational model used here is the four level baroclinic model (Fig. 1). The model is the same as that used in routine operation at JMA (Staff Members, 1964) except the term of the non-adiabatic heating due to the latent heat released by the parameterized convective activity. Main assumptions are as follows:

(1) Motion is hydrostatic and quasi-geostrophic. Since the horizontal dimension of the disturbance (1,000 km) is much larger than the vertical dimension (10 km), the hydrostatic relation is considered to be valid. The balance of the characteristic magnitude of terms in a quasi-geostrophic system of prognostic equations requires that the non-adiabatic heating term should be equal to or less than $10^{-1} \text{m}^2 \text{sec}^{-3}$ (Phillips, 1963)

(2) The atmosphere is heated by the following non-adiabatic heating.

a) There is a transport of sensible heat from the ocean surface to the atmosphere where the air temperature at the sea surface is lower than the water temperature, i.e., $T_a < T_s$, or from the atmosphere to the ocean with smaller rate where the situation is reverse, i.e., $T_a > T_s$ where $T_a$ and $T_s$ are temperatures of the air and water at sea level, respectively. $T_a$ is extrapolated from temperatures at 600 mb and 800 mb.

b) The atmosphere is always fully saturated under the 500 mb level where an upward current exists. The condensation and the release of latent heat are to occur where an upward vertical motion exists. However, an artificial limit of the heating is put so as not to break the ellipticity condition of the $\omega$-equation. No different treating is assumed

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* Private communication.
over the continent and the ocean.

c) There is the release of latent heat due to the condensation through the organized convective activity. The parameterization is the same as shown by the present author in the previous paper (1964), i.e., the heating is placed only over the cyclonic region and the amount of heating is assumed to be proportional to the intensity of the relative vorticity at the lowest level. Syono (1949, 1950) was the first who recognized the close relation between the rate of precipitation associated with a typhoon and the vertical component of the relative vorticity at the lowest level. He proposed a mathematical expression for the amount of precipitation as a product of the surface mixing ratio of water vapor and the vertical velocity which is given in terms of the relative vorticity at the ground surface and at the top of the Ekman boundary layer. Syono et. al. (1951) showed that the radial distribution of the amount of precipitation thus calculated agrees fairly well with that observed.

(3) While no radiative effect is incorporated, we assume that the space-mean thickness or temperature should be conserved as a constant during calculation. Thus, the strong differential heating in some place requires the remarkable differential cooling in the remained area.

(4) The effect of surface dissipation is incorporated in proportion to the relative vorticity at the lowest level.

3. Prognostic equations

The quasi-geostrophic vorticity equations for the 300, 500 and 700 mb levels are

\[
\frac{\partial \zeta_k}{\partial t} = - \mathbf{V}_k \cdot \nabla \zeta_k + f_0 \left( \frac{\partial \omega}{\partial p} \right)_k, \quad (k=3, 5, 7), \quad (3.1)
\]

and that for the 900 mb level is

\[
\frac{\partial \zeta_k}{\partial t} = - \mathbf{V}_k \cdot \nabla \zeta_k + f_0 \left( \frac{\partial \omega}{\partial p} \right)_g + F_g, \quad (3.2)
\]

where \( f_0 \) is the constant Coriolis parameter (at 45°N) and \( F_g \) the frictional dissipation. We assume that

\[
F_g = g \left( \frac{\partial \zeta}{\partial p} \right)_g \sim - \frac{g}{\Delta p} \tau_{10} = - \kappa \tau_{10} < - \kappa \zeta_0,
\]

where \( \Delta p = 200 \) mb, and \( \kappa \) and \( \kappa' \) are constants. Here we assume that \( \omega_1 = 0, \tau = 0 \), and the relative vorticity at 1000 mb, \( \zeta_10 \), is proportional to \( \zeta_0 \). We use \( 7.5 \times 10^{-6} \) sec\(^{-1} \) as \( \kappa' \). Other notations are as usual.

The equations for thickness change with time are as following:

\[
\frac{\partial h_1}{\partial t} = - V_1 \cdot \nabla h_1 + \frac{\Delta p}{g} S_1 \omega_1 + G_1 - G_4, \quad (3.3)
\]

\[
\frac{\partial h_6}{\partial t} = - V_6 \cdot \nabla h_6 + \frac{\Delta p}{g} S_6 \omega_6 - \frac{\partial \rho}{\partial \rho} \left( S_6^* \omega_6 - S_6^* \omega_6 \right) + G_6 = G_6, \quad (3.4)
\]

\[
\frac{\partial h_8}{\partial t} = - V_8 \cdot \nabla h_8 + \frac{\Delta p}{g} S_8 \omega_8 - \frac{\partial \rho}{\partial \rho} \left( S_8^* \omega_8 - S_8^* \omega_8 \right) + H - H + G_8 - G_8, \quad (3.5)
\]

where \( h_1 = Z_3 - Z_5 \) etc., \( S_k = - \left( \alpha \theta \frac{\partial \varphi}{\partial p} \right)_k, \quad (k = 4, 6, 8) \), the mean static stability within the forecast area which is fixed as a constant during calculation, \( \alpha \) the specific volume, \( \theta \) the potential temperature. \( S^* \) is defined as

\[
-S^* \omega = \frac{R}{c_p} \frac{1}{\bar{\rho}} \frac{dQ}{dt}
\]

where \( c_p \) is the specific heat at constant pressure, \( R \) the gas constant, and \( dQ/dt \) the heating per unit time and unit mass, which is associated with the assumption (2b) in the previous section. \( \varphi = 1 \) for the upward velocity \( (\omega_k < 0, k=6, 8) \) and \( \varphi = 0 \) for the downward current \( (\omega_k > 0) \). \( H \) denotes the transport of sensible heat between the atmosphere and the ocean, i.e.,

\[
H = \frac{\Delta p}{g} \frac{R}{c_p} \bar{\rho} \kappa' v_{10}(T_s - T_i)
\]

where \( \kappa'' \) is a constant, whose numerical value is 0.0015 m sec\(^{-2} \) deg\(^{-1} \) when \( T_s < T_i \), (the similar magnitude as that suggested by Jacobs (1951)) and one tenth of this value is used when \( T_s > T_i \). \( v_{10} \) is approximated to be equal to 0.7 \( \times v_g \). The monthly mean values \( T_s \) are used and assumed to be invariably with time. \( G_8, \quad (k=4, 6, 8) \), stands for the parameterized non-adiabatic heating due to the latent heat released by condensation of water vapor through the organized convective cumulus clouds.

Quantities with bar in (3.3)–(3.5) signify
their space averaged values over the computational domain. \( \omega_{x_k} \) \((k=4, 6, 8)\) is obtained from the so-called \( \omega \)-equation.

The following four versions of \( G_k \) are examined.

Version A: \( G_k \equiv 0 \), (which corresponds to the operational model).

Version B: \( G_4=G_6=0 \),

\[
G_8 = -\frac{S_8 dp}{g} (-\eta \omega_{10})
\]

where \( \eta \) is the proportionality constant relating the entropy obtained in the frictional boundary layer to the entropy required to transport the air parcel upward. The vertical \( p \)-velocity at the top of the boundary layer, \( \omega_{10} \), is obtained from \( \zeta \) as

\[
\omega_{10} = -k''\zeta,
\]

where \( k'' \) is the constant coefficient denoting the continuity of mass in the frictional boundary layer, and connecting the frictional convergence in the boundary layer with the relative vorticity in lower layer (Syôno, 1949; Charney and Eliassen, 1949). In this version, only the lowest layer (700-900 mb) is heated.

Version C: \( G_4=0 \)

\[
G_6 = a_6 \times \frac{S_6 dp}{g} (-\eta \omega_{10}) ,
\]

\[
G_8 = a_8 \times \frac{S_8 dp}{g} (-\eta \omega_{10}) ,
\]

where \( a_6 \) and \( a_8 \) are taken for convenience as \( a_6 = a_8 = 0.5 \), for the relatively small scale cyclones, and \( a_6 = 0.7 \), \( a_8 = 0.3 \), for typhoons.

There is no specific reason why we use these numerical values for partitioning the heating. The two lower layers (500-700 and 700-900 mb) are heated in this case.

Version D: \( G_4=a_4 \times \frac{S_4 dp}{g} (-\eta \omega_{10}) ,
\]

\[
G_6 = a_6 \times \frac{S_6 dp}{g} (-\eta \omega_{10}) ,
\]

\[
G_8 = a_8 \times \frac{S_8 dp}{g} (-\eta \omega_{10}) ,
\]

where \( a_4 = 0.3 \), \( a_6 = 0.4 \) and \( a_8 = 0.3 \) are adopted, though again the rate of partition of heating is temporary.

Whole three layers (300-500, 500-700 and 700-900 mb) are heated in this version.

For all the versions, we use \( \eta = 3 \) where \( \omega_{10} < 0 \) and \( \eta = 0 \) where \( \omega_{10} > 0 \). The total amount of the non-adiabatic heating in the whole air column is the same, though the partitioned rate of heating for each layer is different.

As for the upper and the lower boundary conditions, we assume that the vertical \( p \)-velocity \( \omega \) should vanish. It should be noted that \( \omega_{10} = 0 \) is assumed at the base of the boundary layer for solving the \( \omega \)-equation, whereas \( \omega_{10} \) is defined at the top of the frictional boundary layer. Calculation is made in the domain covered by 28 \( \times \) 20 grid points placed at 304.8 km interval. Time increment is 1 hour. The surface isobaric lines are converted from the output by the barometric formula.

4. Numerical examples and discussions

In Fig. 2, the initial and verification maps for 12 GCT 27 May 1964 are shown. At the initial time (Fig. 2 a), a moving anticyclone covers the Japanese Islands and a very weak trough appears over the East China Sea on the ground surface (solid line). A minor trough is coming from the northwest part of the computational domain to Japan at the 500 mb level (dashed line). 24 hours later, a cyclogenesis is observed on the East China Sea off the southern coast of the Kyûshû Island, where already at the initial time precipitation was observed. The high over Japan moves into the Pacific Ocean and a distinct surface trough is covering the Korean Peninsula and the western part of Japan. The upper minor trough moves eastward. (Fig. 2 b).

The prognostic charts are shown in Fig. 3. Version A gives a poor forecast for the cyclogenesis over the East China Sea and we see only the weak surface trough here. (Fig. 3 a). On the other hand, in the calculated chart made by Version B we can see a distinct and closed cyclone although the location of the predicted cyclone is slightly behind the actual one. Concerning the low in the Japan Sea, the forecast is rather poor. (Fig. 3 b). Although the surface prognostic patterns are quite different between Figs. 3 a and 3 b, both models give similar forecast at 500 mb. The
Fig. 2. (a) The observed initial 500 mb contours (dashed lines) and the surface isobars (solid lines) on 12 GCT 27 May 1964. Thick straight lines coincide with the horizontal cross lines upon the ground surface by the plane of the vertical cross sections in Fig. 5.
(b) Same as (a) but for 24 hours later (The verification map).

Fig. 3 (a) The calculated 500 mb contours (dashed lines) and the surface isobars (solid lines) after 24 hours in Version A, starting from 12 GCT 27 May 1964. Thick straight lines coincide with the horizontal cross lines upon the ground surface by the plane of the vertical cross sections in Fig. 6.
(b) Same as (a) but for Version B (after Ōkōchi).
(c) Same as (a) but for Version C.
(d) Same as (a) but for Version D.
upward extension of the heating shallows the central pressure of cyclone on the surface map but again makes little change on the 500 mb charts. (Figs. 3c and 3d). If we increase the total amount of the heating, the predicted central pressure of the low may be modified and intensified. Thus, in this simple treatment of the development of the cyclone, the absolute value of the central pressure of the cyclone in the forecast map may be temporary, because we cannot specify the absolute value of the heating. However, the concentrated intensification of the cyclone in Figs. 3b, 3c and 3d in comparison with that in 3a is the essential feature of the present heating model.

The vertical cross sections of the relative vorticity field across the central area of the cyclone clearly show the difference among the forecasts calculated by the various ways. Before proceeding to view the cross sections, it may be necessary to mention briefly about the accuracy of making the cross sections. In Fig. 4 upper sounding stations in the forecast domain are indicated by the black spots. The horizontal cross lines upon the ground surface by the planes of the vertical cross sections. Fig. 4. Data coverage over the forecast domain. Making the vertical cross section through the central part of cyclone seems to be approved as being reasonable and reliable. Thick straight lines show the horizontal cross lines upon the ground surface by the planes of the vertical cross sections. Attached figures show the date of making the cross sections, i.e., I: the cross section of the verification map for 12 GCT 24 September 1964, II: same as I but for 12 GCT 28 May 1964 and the cross section of the prognostic charts starting from 12 GCT 23 September 1964, III: that of the initial map for 12 GCT 27 May 1964, and IV: same as III but for 12 GCT 23 September 1964.

![Fig. 4. Data coverage over the forecast domain. Making the vertical cross section through the central part of cyclone seems to be approved as being reasonable and reliable. Thick straight lines show the horizontal cross lines upon the ground surface by the planes of the vertical cross sections. Attached figures show the date of making the cross sections, i.e., I: the cross section of the verification map for 12 GCT 24 September 1964, II: same as I but for 12 GCT 28 May 1964 and the cross section of the prognostic charts starting from 12 GCT 23 September 1964, III: that of the initial map for 12 GCT 27 May 1964, and IV: same as III but for 12 GCT 23 September 1964.]

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Fig. 5. The vertical cross sections of the vorticity field through the center or the neighborhood of disturbances in Fig. 2. The vertical component of the relative vorticity is labeled in $1.0 \times 10^{-5} \text{sec}^{-1}$ and the dashed lines show the negative vorticity. (a) and (b) correspond to Figs. 2 (a) and (b), respectively.
sections are identified with the thick straight lines. Making the vertical cross sections in the area along the thick straight lines seems to be approved as being reasonable and reliable.

Now, let us return to and look at the vertical cross sections of the relative vorticity field thus obtained. In Figs. 5 and 6, the vertical cross sections are shown which are made from the vertical component of the relative vorticity field through the center or the neighborhood of the disturbance in Figs. 2 and 3, along the thick straight lines illustrated in Figs. 2, 3 and 4. The vertical component of the relative vorticity is labeled in 1.0×10^{-5} sec^{-1} and the dashed lines show

Fig. 6. Same as Fig. 5 but for the calculated results illustrated in Fig. 3. (a) through (d) correspond to Figs. 3 (a) through (d), respectively.
the negative vorticity. Figs. 5a, 5b, and Figs. 6a through 6d correspond to Figs. 2a, 2b, and Figs. 3a through 3d, respectively. At the initial time a weak positive vorticity area extends up to the 600 mb level (Fig. 5a) and 24 hours later the actual vorticity is intensified and the positive vorticity area broadens horizontally (Fig. 5b). The operational model (Version A) gives the poor description about the change of the vertical structure of vorticity both in magnitude and in size (Fig. 6a). Concerning three methods of incorporating the parameterized 'convective heating', we can easily observe that Version B overly exaggerates the intensification of vorticity at the lowest layer but the other two methods, i.e., Versions C and D, depict moderately the variation. The actual pattern seems to be simulated by the method between these two versions, though the absolute value of the relative vorticity is temporary in these treatments. So far as a resemblance of the calculated pattern to the observed one is concerned, this may be one of the most successful examples.

Next is an example of typhoon forecasting. Here one of the problems to be attacked is how to simulate by numerical model the maintenance of the circulation in the typhoon, particularly in the lower layer, against the reduction of the amplitude due to the truncation error and the stabilizing baroclinic effect.

In Fig. 7, the typhoon No. 20 in 1964 is heading for the Kyūshū Island. At the initial time, on 12 GCT 23 September, the central sea-level pressure is 930 mb (Fig. 7a) and during 24 hours the central pressure shallows to 950 mb, while the typhoon is moving almost northward (Fig. 7b). In the calculation we filled up the central pressure to 996 mb at the initial time so as to decrease the truncation error. The prognostic map of Version A, however, shows exaggerated shallowing of the pressure at the center of the typhoon, i.e., 1002 mb. (Fig. 8a). In contrast with the predicted central pressure in Version A, those in Versions B, C and D rather deepen 24 hours later and the area surrounded by closed isobars around typhoon broadens, though the sharp concentration of isobars around the center can not be duplicated (Figs. 8 b through d).

The predicted movement of the typhoon in all the methods is slower in comparison with the observed one. Thus, we may say that the present modification of the prognostic map due to the non-adiabatic heating of the 'convective type' does not alter the general result of numerical prediction by means of the operational model but keeps the central pressure of the typhoon as it is at the initial time.

In Fig. 9, shown are the vertical cross sections of the relative vorticity around the typhoon which is described in Fig. 7. The legend of figure is the same as that in Fig. 5. The cross sections in Figs. 9 and 10 strongly suggest the important role of the warm core in the typhoon. Versions B (Fig. 10 b) and

![Fig. 7. Same as Fig. 2 but for 12 GCT 23 September 1964. Thick straight lines are the horizontal cross lines upon the ground surface by the plane of the vertical cross sections in Fig. 9. (a) and (b) correspond to Fig. 2 (a) and (b), respectively.](image-url)
Fig. 8. Same as Fig. 3 but for 12 GCT 23 September 1964. (a) through (d) correspond to Figs. 3 (a) through (d), respectively.

Fig. 9. Same as Fig. 5 but for Fig. 7. (a) and (b) correspond to Fig. 7 (a) and (b), respectively.
C (Fig. 10c) distort the vertical structure of the typhoon, while they seemingly retain the initial intensity of the central pressure. On the other hand, the vertical structure simulated by Version D (Fig. 10d) may be the most similar to the observed one (Fig. 9b) among all the methods. The result obtained by the operational model, i.e., Version A (Fig. 10a) indicates the false weakening of the typhoon.

In Figs. 11 and 12, we can see the similar results to those of Figs. 7 and 8. It should be noted that we are remarkably impressed in this case with the fact that the present numerical forecasting model fails in tracing the rapid movement of the typhoon and gives slower displacement compared with the observed one. By our experience, almost all the operational numerical predictions result in
the similar retardation of the movement of the typhoon. The primary reason for this delay is presumably the truncation error, because the present resolution can cover the typhoon with only a few grid points. The fact, that the retardation becomes larger as we get the deeper central pressure by the non-adiabatic heating due to the parameterized convective activity, supports this presumption. However, the delay of the movement of the typhoon seems to be much more than that attributable to the truncation error alone. Retaining the deep central pressure of the typhoon in one hand and forecasting the fast movement in other hand are difficult problems in numerical prediction, which remain.

As was mentioned before, this method of simulating the diabatic heating over the

Fig. 11. Same as Fig. 2 (b) but for 12 GCT 24 September 1964. The vertical cross section of the vorticity field is not presented here in this case. The initial pattern of this instance is equal to Fig. 7 (b) and not reproduced here.

Fig. 12. (a) Same as Fig. 3 (a) but for 12 GCT 24 September 1964. (b) Same as Fig. 3 (b) but for 12 GCT 24 September 1964.

Fig. 13. Same as Fig. 2 but for 12 GCT 30 September 1964. (a) and (b) correspond to Figs. 2 (a) and (b), respectively.
cyclone emphasizes the vorticity at the lowest level, i.e., at the 900 mb level. Then if the prediction of the vorticity is not succeeded, errors may be exaggerated.

The example illustrated in Figs. 13 and 14 is one of the failed cases. On 12 GCT 30 September 1964 we observe a small cyclone over the Pacific Ocean off the southern coast of the Japanese Islands. An upper trough is observed from the Japan Sea to the East China Sea (Fig. 13 a). 24 hours later, this upper trough is strengthened and passes through the Japanese Islands straightly toward east. The development of the cyclone occurs actually over the ocean off the Pacific coast of Japan on the surface chart (Fig. 13 b). The upper trough predicted by Versions A and B is almost the same and the southern part of trough remains at the Kyūshū Island, though the northern part of trough moves rather accurately toward east. On the surface prognostic charts we see the weak tendency for making depression over the Pacific Ocean off Japan (Fig. 14 a and b). However, as the predicted movement of the upper trough is too slow particularly in its southern part, the location of the predicted low is also different than that in the observed map. Version B exaggerates this failure.

Although the reason for obtaining an inaccurate prognostic pattern at the upper level is uncertain in this case, a better treatment of this type of the heating may contribute at least partially to the improvement of the forecast, even for the upper level.

5. Summary

From the examples presented above, we may summarize the effect of the non-adiabatic heating due to the latent heat released by the condensation of water vapor through the parameterized convective activity as follows:

(1) The development of the relatively small scale cyclone which appears over the ocean off the Pacific coast of the Japanese Islands is well depicted by the method between Versions C and D, but that made by Version B exaggeratedly shows the warming of the lowest layer.

(2) The maintenance of the cyclonic circulation of the typhoon is simulated most skillfully by Version D, while the phase error, i.e., the retardation of the movement of the typhoon is little reduced.

(3) As the supply of the water vapor is not controled directly nor indirectly in this formulation of the diabatic heating due to the latent heat released by the condensation of water vapor through the parameterized convective activity, the disturbance does not cease to develop with time, though a prognostic chart beyond 24 hours is not illustrated in the present paper. The limit of the amount of the non-adiabatic heating which is appropriate to the quasi-geostrophic approximation (e.g., $10^{-1} \text{m}^2\text{sec}^{-3}$ by Phillips (1963)) seems to be not exceeded by the present non-adiabatic model during 24 hours.

(4) When the forecast of an upper pattern is not succeeded, this effect exaggerates an error.
Generally speaking, this simple way of incorporating the non-adiabatic heating gives interesting results. However, more accurate and advanced treatments of the problem may be needed.

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References
Staff Members, 1964: Periodic report on numerical weather prediction, IV, 15 pp, Electronic Computation Center, Japan Meteorological Agency.
小規模低気圧の発達と維持に重点をおいた数値予報例

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気象庁でルーチンに使っている4層傾斜モデルには既に海面との熱のやりとりとか、大規模な場にみある上昇流によって解放された凝結熱の効果といった非断熱効果がとり入れられ、春秋の移動性低気圧の発達と進路予報の改善や冬の大気高気圧の張り出しの予報に著しい成果を示した。しかし、台風が日本附近にやってきた場合や、本邦南岸沿いに発生、発達する小規模な低気圧の予報はうまくゆかず、こうした小規模乱れは予報と共に衰えたり消滅したりする。このひとつの主要な原因は、水平の代表的規模が1,000キロメートルといった領域に対し、304.8キロという現在の格子間隔がいかに大きく、そのための切断誤差や平均操作によるものと考えられる。更には、この種の低気圧や台風では、普通の移動性低気圧と逆に下層にゆく微弱なうつ引いた垂直構造であるため、運動エネルギーが位置のエネルギーに移るという意味の領土の安定化のための振巾が減少すると考えられる。

台風の構造を維持する上で、対流性凝結が主要な役割を担い、台風というマクロな運動に対して、協力系を形成することは近年明らかにされてきた所である。他方、残念ながら、本邦南部洋上にあらわれる小規模低気圧についても、その物理的構造に関する解析的研究に乏しくはっきりしたことはわからない。しかし上層の小さい谷の接引に伴って、主に対流層下部で短時間に急速に発達することはよく知られており、最近のレーダーの観測では低気圧の中に雄大蒸発のエコーを認めているものもある。台風とは多かれ少なかれ違っているかもしれないが、マクロな運動とマクロな運動の協力系を一応なし、そうした方向の取扱いをこころめた。台風の場合に重要な役割を果たす摩擦境界層は、小規模低気圧の場合は湿潤とみなせよう。

現在のルーチン・モデルで考慮されている大規模な運動に関する降水中解される凝結熱だけでは不十分であることは先述の通りなので、本論文ではこれを主として対流性凝結による潜熱の非断熱効果という観点から考え、簡単にパラメータ化して準地衡風近似の4層モデルに導入した。準地衡風近似に係るかどうかの妥当性については、発達の各発達段階に応じたスケール・アナリシスが必要である。ここではマクロな運動を対象とし、それが単独で乱されないうようにするため先鋭的に準地衡風近似を用いた。その範囲内で許される非断熱効果の大きさは約 10⁻¹ m²/s⁻³（Phillips, 1963）である。ここに用いた例では24時間予報の間、ほぼこの制限は守られていたようである。

モデルに含まれる主要な仮定は次の通りである。

(1) 運動は静力学的及び準地衡風的起る。
(2) 非断熱項として(i)海面との顯熱のやりとり、(ii)大規模援乱に対する準地衡的う状程式から求まるw による凝結熱の解放、(iii)上層のパラメータ化した凝液対流型の加熱（但し上昇流の上でのみ生じ、加熱量は下層の相対うつ引いた値に比例すると考えること）の3つの効果を考える。
(3) 気圧場を時間と共に変らない。
(4) 地面摩擦によるエネルギーの消費を考慮する。


このような効果を考慮した場合の予報結果を、ルーチンの結果（対流の効果を考慮していない）と比較するために、2, 3 の少数例について数値予報を行なった。その結果次のことがわかった

(1) Version B, C 及び D を用いると、切断誤差及び傾向の安定効果で衰えがらの小規模低気圧の発達や台風内循環の維持が再現された。
(2) しかし依然として台風進路の顕著なおくれは改良されなかった。
(3) 又、上層のう状度予報がうまくゆかなかった場合には、この効果の、今回の加きパラメータ化はかえって誤差を拡大する傾向がある。
(4) 台風の相似には、Version D が最も適しているようである。

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