The Effects of Different Cumulus Parameterization Schemes on the Intensity Forecast of Typhoon Flo (1990)

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Abstract

The impact of two cumulus parameterizations on intensity of a simulated tropical cyclone was examined using a limited area numerical model. A moist convective adjustment scheme and Arakawa-Schubert scheme were used as a cumulus parameterization for each numerical experiment. Typhoon Flo (1990) was selected as a case study.

Difference in the cumulus parameterizations had drastic influence on a difference in sea-level pressure averaged over the core region (within the 250 km radius from the storm center) of the simulated storm from an early stage of the integration. That is, compared with the experiment with moist convective adjustment, the averaged sea-level pressure with Arakawa-Schubert scheme was relatively low. Central pressure in the latter was also lower than the former, which showed that the latter produced a more intense tropical cyclone consistent with the observation. The difference in central pressure, however, appeared at a later stage, compared with the difference in the averaged sea-level pressure.

Temperature tendency due to a moist convective adjustment was considerably negative in the lower troposphere and so was actual temperature tendency, which resulted in a rapid drop of temperature. Consequently, the total air mass within the inner-core region became greater in the experiment with moist convective adjustment than with the Arakawa-Schubert scheme. On the other hand, the difference in the low-level temperature field between the two experiments results in the difference in radial profile of rainfall, and the radius where the eyewall is located. The difference probably brings about the delayed appearance of the central pressure difference.

Another numerical experiment was conducted with modified moist convective adjustment not to be cool in the lower troposphere. The result confirmed that cooling in the lower troposphere within the inner core-region greatly inhibits intensification of the simulated storm.

1. Introduction

The tropical cyclone is one of the most life threatening and destructive natural phenomena on earth. Therefore, forecasting of tropical cyclones is one of the most important components in numerical weather prediction. As far as the track prediction is concerned, it has been largely improved in forecast skill in recent years based on the advancements in understanding of various factors associated with tropical cyclone motion such as environmental wind field, symmetric and asymmetric vortex structures (Elsberry 1995). For instance, the hurricane model developed at the GFDL (Geophysical Fluid Dynamics Laboratory) has a hurricane-like vortex, which includes asymmetries, in the initial field (Kurihara

et al. 1998). The GFDL model has been adopted as an operational hurricane forecast model since 1995 at NCEP (National Centers for Environmental Prediction). This model outperformed the previously operational hurricane model of NCEP in forecast skill.

Intensity prediction using numerical models has been attempted (e.g., Bender et al. 1993; Nagata et al. 1998). However, the tropical cyclone intensity prediction, that is, the prediction of the central pressure or maximum wind speed, prove to have much less skill than track prediction. One of the reasons for this problem is that resolutions of prediction models are too coarse to appropriately represent tropical cyclone structure within the region where the center and eyewall are included. Another reason is that intensity of the tropical cyclones is influenced by many unknown processes such as internal processes (e.g., eye subsidence, spiral band's downdrafts), and environmental forcings (e.g., ver-

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tical shear, eddy angular momentum flux convergence) (Elsberry et al. 1992).

It is well known that latent heat release of water vapor is crucial for development of tropical cyclones. Many researchers have investigated the relationship between latent heat release and tropical cyclone development from the observational points of view (e.g., Alliss et al. 1992; Rodgers et al. 1994a, b), and from the theoretical ones (e.g., Yamasaki 1968a, b). Using the satellite microwave data, Alliss et al. (1993) showed that timing of the intensification of Hurricane Florence (1988) was coincident with that of an increase in total latent heat release. Rodgers and Pierce (1995) suggested that an increase in the precipitation within the inertially stable region near the storm center precedes intensification of a tropical cyclone.

Some studies addressed the relationship between the vertical profile of latent heat release and tropical cyclone development. On the basis of numerical experiments with an axisymmetric primitive-equation model, Yamasaki (1968a) showed that a simulated tropical cyclone could develop when latent heat was released in the lower troposphere, and that the time period required for the development was shortened as convective heating increased there. There are studies with the Eliassen balanced vortex model (e.g., Shapiro and Willoughby 1982; Schubert and Hack 1982). Hack and Schubert (1986) showed that the lower the height of maximum heating became, the more the simulated vortex developed. On the other hand, Rodgers et al. (1998) estimated the vertical distribution of latent heat release in a tropical cyclone using microwave data from satellites, and hypothesized that the latent heat release in middle- and upper-troposphere intensified the system.

Recently, Numerical Weather Prediction (NWP) models have been widely used for tropical cyclone forecasting. Since cumulus convection appears to play a key role in the tropical cyclone intensification processes, it is desirable that cumulus clouds are resolved in the models. However, it is beyond current computer capabilities to regularly compute the high resolution model which can explicitly represent cumulus convection. Thus, it is necessary to use cumulus parameterization that represents the effects of cumulus clouds on grid scale phenomena.

As mentioned above, development of a tropical cyclone appears to be sensitive to the magnitude and the vertical profile of heating produced by cumulus parameterization. Several studies examined the sensitivity of tropical cyclone forecast to the formulation of cumulus parameterization. On the basis of analysis-forecast experiment with ECMWF (European Center for Medium-Range Weather Forecasts) operational global model, Puri and Miller (1990) mentioned that they could obtain more intense tropical cyclones by the Betts-Miller (1986) convective

adjustment scheme than by the Kuo scheme (Kuo and Anthes 1984). Baik et al. (1991) compared Betts (1986) and Kuo scheme (Kuo 1974; Anthes 1977), through numerical simulations of tropical cyclone with an axisymmetric model. They showed that the storms begin to intensify when grid-scale heating extended through a deep layer, and that later development of the storm with the Betts scheme resulted from a delayed onset of grid-scale heating. They also found that the storm evolution with both cumulus parameterization schemes was sensitive to inclusion of the cooling effect of rain-drop evaporation. Recently, Tsutsui et al. (1998) conducted forecast experiments for tropical cyclones with a limited area model. They investigated the impact of two kinds of cumulus parameterization, which were a relaxed Arakawa-Schubert scheme proposed by Moorthi and Suarez (1992) and Kuo (1974) scheme, on tropical cyclone prediction. Their result was that the former appeared more suitable than the latter for the intensity prediction of developing tropical cyclones, whereas the use of the latter produced better scores in mean displacement errors.

The purpose of this study is to investigate the influence of the vertical heating profile on the intensity of a simulated tropical cyclone through the comparison of numerical experiments with two different cumulus parameterization schemes. One of them is a moist convective adjustment scheme by Gadd and Keers (1970), and another is the scheme proposed by Arakawa and Schubert (1974). In the present study a high-resolution limited-area model is used targeting Typhoon Flo (1990) so that the structure of the tropical cyclone such as eyewall would be reproduced well.

In Section 2, the observational aspect of Typhoon Flo is summarized briefly. Section 3 provides an outline of the experiment. From Section 4 to Section 6, results of the numerical experiment are presented. The tropical cyclone intensity obtained by the forecast experiments is shown in Section 4. Section 5 describes the main difference in thermodynamic field between two sets of model integration with each cumulus parameterization. In Section 6, the results from an additional experiment conducted in order to confirm a hypothesis introduced in Section 5 are compared with those from the previous ones. Section 7 discusses the results as compared with other studies. Finally, Section 8 gives a summary and conclusion.

2. Typhoon Flo

Typhoon Flo, which is assigned a serial typhoon number 19 at 1990 by the Regional Specialized Meteorological Center (RSMC)-Tokyo, is used for the forecast experiments. This typhoon began as a distinct tropical depression southeast of Guam on 12 September 1990 (Fig. 1). Tropical Storm inten-

Fig. 1. The track of Typhoon Flo (1990) by RSMC Tokyo Typhoon Center during September 1990. Positions indicated by an open circle and a closed circle are at 0000 UTC and 1200 UTC respectively. A plotted number denotes each day and the bold line is the period used for forecast experiments described in this paper. Solid lines indicate that Flo is at the stage of tropical storm intensity or higher, while dashed lines indicated tropical depression or extratropical cyclone.

sity (17 m/s or 34 kt) was attained at 0000 UTC on 13 September. Flo followed a basically north-westward track towards the Japanese island of Okinawa and recurved to the northeast at 1200 UTC 17 September. It crossed the main Japanese island of Honshu from 19 to 20 September and damaged a lot of materials. The transition into an extratropical cyclone occurred at 0600 UTC on 20 September.

From 14 to 16 September, Flo underwent a period of explosive intensification. A minimum central pressure of 890 hPa was attained from 0000 UTC to 0600 UTC on 17 September (Fig. 2). The model integrated over the range of 72 hours starting at 0000 UTC 14 September, which coincides with the period that Flo went through rapid intensification.

Flo was generated during special observations called TCM-90 (Elsberry 1990), SPECTRUM (Spe-

Fig. 2. The central pressure of Typhoon Flo (1990) by RSMC Tokyo Typhoon Center during September 1990. Data are plotted every 6 hours from 0000 UTC 12 September to 0000 UTC 22 September. The period for forecast experiments described in this paper is indicated by the bold line.

cial Experiment Concerning Typhoon Recurvature and Unusual Movement) and Typhoon-90. Thus, many researchers chose this typhoon as a target of their numerical studies on tropical cyclones (e.g., Yamasaki 1992; Murata et al. 1996; Aonashi et al. 1996; Kasahara et al. 1996; Tsutsui et al. 1998).

3. Outline of experiment

A higher resolution version of the Japan Spectral Model (JSM) is used for the forecast experiments. The model is a hydrostatic, 23-level spectral limited-area model that was used as one of the operational forecast models of the Japan Meteorological Agency (JMA) from March 1988 to February 1996 (Segami et al. 1989; Nakamura 1992).

A domain for the forecast experiments is a part of the western North Pacific (6°N–35°N, 116°E–155°E) and has a regular $203^{\circ} \times 163^{\circ}$ transform grid with a grid distance of 20 km at 22.5°N on a Mercator's projection plane (Fig. 3). The initial field and lateral boundary data are obtained from interpolating the JMA Global Analyses (GANAL) to the model grids. A typhoon bogus insertion procedure is used to produce correctly positioned tropical cyclone-like vortices within the initial field. This typhoon bogus, which is the same one as in the JMA operational NWP system, contains both symmetric and asymmetric components of wind and geopotential height (Ueno 1995).

As mentioned above, two kinds of cumulus parameterization are used for the forecast experiments. One of them is a moist convective adjustment scheme proposed by Gadd and Keers (1970), which was operationally used in the JSM. In this scheme,

Fig. 3. A domain for forecast experiments. Tracks of the typhoon in forecast experiments with a moist convective adjustment (CA) and Arakawa-Schubert scheme (AS) are shown every 6 hours from 0000 UTC 14 September to 0000 UTC 17 September. The track of Typhoon Flo by RSMC Tokyo Typhoon Center is also shown.

when the temperature lapse rate grows greater than the critical value Γ_c , the temperature and specific humidity are adjusted so that the lapse rate becomes Γ_c without changing the relative humidity. Γ_c is a function of relative humidity. The other is a version of the economical prognostic Arakawa-Schubert scheme (EPAS), which was developed by Kuma et al. (1993) and is currently in operational use at the JMA. The cloud mass flux is prognostically determined in the EPAS.

4. Tropical cyclone intensity obtained by the forecast experiment

4.1 The central pressure

The tracks of a simulated tropical cyclone in the experiment with the moist convective adjustment (CA), and with the Arakawa-Schubert scheme (AS) are shown in Fig. 3. The latter is slightly better than the former, although they both have a northward bias compared with the observation (the best track data archived by the RSMC-Tokyo). Figure 4 shows the time series of central pressure. The central pressure in the AS run is lower than, or almost equal to, that of the CA throughout the integration time, although deepening is still not sufficient compared with the observation. It should be noted that the difference between two simulations in terms of central pressure becomes noticeable at around 27 hours

Fig. 4. The time series of the central pressure in forecast experiments with moist convective adjustment (CA) and Arakawa-Schubert scheme (AS). The central pressure of Typhoon Flo by RSMC Tokyo Typhoon Center is also shown.

Fig. 5. The time series of sea-level pressure averaged within a circular area in forecast experiments with moist convective adjustment (CA) and Arakawa-Schubert scheme (AS). Lateral boundary of the circular area is at 250 km from the simulated storm location.

and the magnitude exceeds 10 hPa around 36 hours.

4.2 Areal mean pressure near the tropical cyclone center

Central pressure is often used as a measure of tropical cyclone intensity. However, central pressure stands for the total air mass of a single air column only at the storm center. Therefore, instead of one value there, we adopt areal mean sea-level pressure as an indicator of mass field near the tropical cyclone center. The quantity of mean sea-level pressure, P , within an annular area is defined as follows:

$$P(r_1, r_2) = \sum p/n$$

where p is grid point values of sea-level pressure, and n is the number of grid points included within an annular area bounded by a circle with radii of r_1 [km] and r_2 [km], centered on the forecast storm location.

Figure 5 shows the time series of $P(0, 250)$, that is, a mean value within a circular area with the 250 km radius. $P(0, 250)$ in the case of AS is consistently

Fig. 6. Vertical distribution of difference in areal mean temperature between two forecast experiments with moist convective adjustment (CA) and Arakawa-Schubert scheme (AS). The value of AS is subtracted from that of CA at each level and forecast time. Lateral boundary of the area is at 250 km from the simulated storm location. Contour interval is 0.1 K. The positive values are hatched.

smaller than the CA throughout the integration. Unlike the central pressure, however, the difference in terms of areal mean pressure appears at 3 hours.

Weatherford and Gray (1988) defined the 0–2.5 degrees radius of the tropical cyclone of the western North Pacific as the core region. For the purpose of present study, we use the approximately same definition as theirs. That is, the 0–250 km radius area is defined as the core region.

5. Difference in temperature distribution between two cumulus parameterizations

5.1 Vertical distribution of temperature

In order to examine the contribution of air mass at each pressure level to the areal mean sea-level pressure, the vertical distribution of temperature difference (i.e., CA-AS) is drawn (Fig. 6). It is found from the figure that from lower to upper troposphere the values are generally negative, which means that simulated air in the CA is colder than that in the AS. In particular, below 700 hPa the difference is relatively large. It is also found that the temperature difference in the lower troposphere increases rapidly from initial time to 6 hours.

5.2 Temperature at low levels

Figure 7 shows the time series of temperature at 950 hPa that is horizontally averaged within the same region as used for the calculation of $P(0,250)$.

Fig. 7. The time series of areal mean temperature at 950 hPa in forecast experiments with moist convective adjustment (CA) and Arakawa-Schubert scheme (AS). The area has the same boundary as in Fig. 5.

In the figure, values are plotted for both simulations of the CA and AS.

Through the whole forecast time, temperature in the CA is lower than that in the AS. From the time series for the CA, we see that the temperature goes down rapidly from initial time to 6 hours. In particular, between the initial time and 3 hours, the temperature falls from 24.4 to 22.8 degrees centigrade. This temperature drop occurs during the period when the temperature difference between the CA and AS in the lower troposphere increases rapidly. On the other hand, the initial temperature fall is

considerably small in the AS simulation. Consequently, temperature difference between the two becomes significantly large during the early stage of the integration.

As mentioned already, the difference in $P(0,250)$ appears soon after the integration starts (Fig. 5). Therefore, one can say that this difference in the areal mean sea-level pressure correlates with temperature difference in the low levels at least during the early stage of the integration.

5.3 Analysis of temperature tendency

To investigate the reason for the rapid temperature drop in the CA run in the lower troposphere, we divide the temperature tendency [K/hour] into several components, that is, horizontal advection, vertical advection, large-scale condensation, and cumulus parameterization. Vertical advection includes adiabatic cooling or heating. Vertical profiles for each component of temperature tendency at the beginning of the forecast are shown in Fig. 8. Vertical profile for the sum of these four components and "actual" temperature change are shown in Fig. 9. The "actual" temperature change, ΔT [K/hour], is defined as follows:

$$\Delta T(t) = (T(t+1) - T(t-1))/2$$

where T is temperature, and t is forecast time in hours.

In both the CA and AS, strong cooling occurs owing to vertical advection (Fig. 8). On the other hand, temperature change due to large-scale condensation and cumulus parameterization is positive on the whole, although the ratio of the former to the latter is fairly different between the two model runs. At 2 hours, all four components of the temperature tendencies cancel out to near zero value in the middle troposphere in the CA and from the lower to the middle troposphere in the AS (Fig. 9).

As mentioned above, the CA temperature falls rapidly in the low levels at the early stage of the integration. Close correspondence between "actual" temperature change and summation of the four components of temperature tendency enable us to interpret the temperature change in terms of the components. A small difference between "actual" temperature change and "extrapolated" one based on the instantaneous tendencies, might be partly explained by the neglect of contributions from eddy mixing and radiative processes to the latter calculation.

From Fig. 8 (a), it should be noted that temperature tendency due to cumulus parameterization has negative values at 950 hPa in the CA, while it has positive or slightly negative ones in the AS. In the CA scheme, if heating due to vertical instability occurs in a layer, specific humidity there increases simultaneously because the scheme puts the restriction that grid-scale relative humidity there does not

Fig. 8. Vertical profiles of each component in temperature tendency [K/hour] in forecast experiments with (a) moist convective adjustment (CA) and (b) Arakawa-Schubert scheme (AS) from 1 hour to 3 hours of time integration. These components are the temperature tendency due to horizontal advection (triangle), vertical advection (circle), large-scale condensation (cross), and cumulus parameterization (plus), where vertical advection includes adiabatic cooling or heating. They are averaged values within a circular area bounded by a circle with a radius of 250 km centered on the forecast storm location.

change during the adjustment is active. Since total moist static energy in the air column has to be conserved, cooling and specific humidity decreasing have to occur somewhere. The negative values at 950 hPa in the CA appear in this way. It should be concluded, from what has been described above, that negative temperature tendency due to the CA cumulus parameterization at 950 hPa influences "actual" temperature change in the low levels.

The difference in the amount of cumulus-scale heating between two schemes depends on different method in cumulus parameterization. In the CA scheme, moisture in an air column tends to be transported above when the column has vertical instabil-

Fig. 9. Vertical profiles of the sum of the four components [K/hour] mentioned in Fig. 8 (circle) and "actual" temperature tendency [K/hour] (triangle) in forecast experiments with (a) moist convective adjustment (CA) and (b) Arakawa-Schubert scheme (AS) from 1 hour to 3 hour of time integration. They are averaged values within the same area as Fig. 8. For the definition of the term "actual", see the text.

ity. Therefore, residual moisture provided for precipitation is less than the AS scheme. These characteristics in the CA scheme lead to less cumulus-scale heating. Actually, the CA cumulus heating has much smaller values at 2 and 3 hours compared with the AS one (Fig. 8).

5.4 Influence of cumulus parameterization on sea-level pressure

It is necessary to examine how the low-level cooling in the CA influences sea-level pressure. Figure 10 shows the difference in areal mean (within the 250 km radius) geopotential height between the CA and AS. It is found that the quantity is remarkably negative below 600 hPa from the initial to 3 hours. The result means that the low-level cooling in the CA lowers geopotential height from low to the middle of the troposphere. The difference between the two in azimuthally averaged radial wind velocity below 600 hPa, which is regarded as a measure

of mass flux, at the 250 km radius from the storm center (not shown) is also negative (i.e., the CA has stronger inflow or weaker outflow) between the initial to 3 hours. Therefore, the low-level cooling in the CA brings about less decrease in areal mean sea-level pressure $P(0,250)$, compared with the AS through the changes in geopotential height and the lateral mass flux.

To summarize, negative temperature tendency in the lower troposphere caused by the CA algorithm leads to temperature drop there in contrast with the AS simulation. The difference in temperature between the two simulations also accounts for the difference in mass field around the storm center.

6. Forecast experiment with a modified moist convective adjustment

6.1 Modified moist convective adjustment

On the basis of the results mentioned, we hypothesize that lower tropospheric temperature just before the developing stage of a simulated tropical cyclone, is one of the important factors to control development of the tropical cyclone. In order to obtain evidence that supports the hypothesis, another forecast experiment is performed within the framework of moist convective adjustment. The experiment is the same as previous two experiments except for the cumulus parameterization scheme. The experimental framework is chosen so that we can focus our discussion on the vertical heating profile irrespective of various differences between two cumulus parameterization schemes based on the largely different algorithms as the CA and AS. We modify the moist convective adjustment as follows (from now on we refer to it as MS).

In the CA, if the lapse rate of temperature grows greater than a critical value Γ_c , which is a function of relative humidity (RH[%]), temperature is adjusted so that the lapse rate is equal to Γ_c . Γ_c is determined as follows:

$$\Gamma_c = (RH - 50)(\Gamma_m - \Gamma_d)/50 + \Gamma_d \quad (RH > 50 \%) \quad (1)$$

where Γ_d and Γ_m is the dry and moist adiabatic lapse rate respectively. On the other hand, in the modified scheme Γ_w is used instead of Γ_m , where Γ_w is given by

$$\Gamma_w = 0.9\Gamma_m + 0.1\Gamma_d.$$

This modification to the critical lapse rate intends that the degree of the adjustment becomes weaker than that of the CA. One can expect that cooling by cumulus parameterization in the lower troposphere in the CA is suppressed with the modification. The modification implies that the lapse rate in an unstable layer is not adjusted to the critical one immediately as in the original CA scheme. Therefore,

Fig. 10. Vertical distribution of the difference in areal mean geopotential height between two forecast experiments with moist convective adjustment (CA) and the Arakawa-Schubert scheme (AS). The value of AS is subtracted from that of CA at each level and forecast time. Lateral boundary of the area is at 250 km from the simulated storm location. Contour interval is 1 gpm. The positive values are hatched.

convective instability remains there more than the original scheme.

6.2 Column test of the MS

In order to examine the performance of the MS, we have tested it under the environment of the single column experiment. A vertical sounding used by Kurihara (1973) for the similar purpose is adopted for the column test. Figure 11 compares change of temperature and specific humidity due to the MS and CA.

In the CA, temperature change is positive above 950 hPa and negative below this level. In particular, it should be noted that fairly large cooling occurs at the lowest two pressure levels in Fig. 11. Vertical profile of specific humidity change is similar to that of temperature change. This result is due to the assumption of cumulus parameterization that relative humidity at each level is not changed by invoking the cumulus parameterization scheme and total moist static energy in an air column is conserved. On the other hand, in case of the MS, cooling at the lower two levels is negligible. The disappearance of prominent cooling layers is a direct consequence of the modification of the critical lapse rate as expressed in (1).

6.3 Results of the experiment

The time series of 950 hPa temperature averaged within a circular area bounded by a circle with a

radius of 250 km, centered on the forecast storm location, is shown in Fig. 12 for the MS simulation. In addition to that, the time series of areal mean sea-level pressure averaged over the same area is shown in Fig. 13. In both figures, results from the previous experiments (i.e., the CA and AS) are also given for comparison.

As shown in Fig. 12, temperature decrease during the early stage of the MS integration is significantly smaller than the CA. The mean sea-level pressure $P(0,250)$ in the MS are less than in the CA after 6 hours (Fig. 13). The result of the experiment with the MS supports the aforementioned hypothesis that temperature in the lower troposphere just before the developing stage of Flo is one of the important factors controlling the subsequent evolution of the storm intensity.

7. Discussion

7.1 Static stability

Radial distribution of azimuthally averaged static stability in the lower troposphere from the initial time to 5 hours is shown in Fig. 14. The static stability is calculated as follows. First, moist static energy, $gz + c_p T + Lq$, at 950 hPa is calculated, where g is gravity acceleration, z is geopotential height, c_p is specific heat capacity, T is temperature, L is latent heat release constant from water vapor to liquid water, and q is the specific humidity. Secondly, satu-

Fig. 11. Change of (a) temperature and (b) specific humidity due to a modified moist convective adjustment (MS), and the original one (CA).

Fig. 12. Same as Fig. 7, but in a forecast experiment with modified moist convective adjustment (MS). Values already shown in Fig. 7 for experiments with a moist convective adjustment (CA) and with the Arakawa-Schubert scheme (AS) are also shown.

rated moist static energy, $gz + c_p T + Lq_s$, at 700 hPa is calculated where q_s is the saturation specific humidity. Thirdly, the saturated moist static energy at 700 hPa is subtracted from the moist static energy at 950 hPa. Finally, an azimuthal mean of these values is calculated at an interval of 50 km radius. If the value is positive (negative) at a radial point, then the point is assumed to be unstable (stable) in the low levels.

There is a pronounced difference between the CA and AS as clearly seen in Fig. 14. It is also found

Fig. 13. Same as Fig. 5, but in a forecast experiment with modified moist convective adjustment (MS). Values already shown in Fig. 5 for experiments with a moist convective adjustment (CA), and with the Arakawa-Schubert scheme (AS) are also shown.

Fig. 14. Radial distribution of azimuthally averaged vertical stability in lower troposphere in forecast experiments with moist convective adjustment (CA) and Arakawa-Schubert (AS) convective parameterization from initial time to 5 hours. The vertical stability is calculated by subtraction of saturated moist static energy at 700 hPa from moist static energy at 950 hPa.

Fig. 15. The time series of rainfall rate in forecast experiments with moist convective adjustment (CA) and Arakawa-Schubert scheme (AS). The rainfall rate includes contributions from both large-scale condensation and cumulus parameterization. The area has the same boundary as in Fig. 6.

that the difference tends to be confined within the area of 300 km radius or so. The relatively large static stability in the CA atmosphere as compared with the AS one is manifestation of the aforementioned result that temperature in the lower troposphere falls rapidly in the CA run.

7.2 Vertical profile of latent heating

Many researchers have pointed out so far that latent heat release, particularly within the region which corresponded to the core we defined, is important for development of tropical cyclones (e.g., Yamasaki 1968a, b; Hack and Schubert 1986; Rodgers et al. 1994a, b). Rodgers et al. (1994a) found that diabatic heating within the region correlated with the subsequent intensity change from the analyses of 18 named western North Atlantic tropical cyclones with satellite microwave data. Rodgers and Pierce (1995) also suggested that the increased precipitation in the inertially stable core preceded intensification of tropical cyclones. Figure 15 shows the time series of rainfall rate within the radius of 250 km from the tropical cyclone center, where rainfall rate includes contributions from both large-scale condensation and cumulus parameterization. During the early period of the integration (say, up to 5 hours), difference in the rainfall rate between the CA and AS is negligibly small. Thus, it seems that total latent heat release within the core region in this case does not account for the difference in the storm intensity at early times.

As shown previously, from the initial to 3 hours there are considerable differences between the CA and AS in vertical profile of latent heat release within the core region (Fig. 8). We have seen that the difference in latent heat release between two cumulus parameterizations accounts for the difference in total mass within the core region (Fig. 5). On

the basis of a numerical model with primitive equations, Yamasaki (1968a) showed that model tropical storm developed when 43 % of released latent heat was distributed to the lower troposphere, and the remaining 57 % to the upper troposphere. However, when the distribution to the upper was increased up to 70 %, a model storm did not develop. He also showed that the central pressure of the model storm dropped more rapidly as heating in the lower troposphere increased although the tropical cyclone-like structure was broken when the rate of released latent heat in the low level rose to 70 %. Similarly, by an axisymmetric primitive-equation model and an Eliassen balanced vortex model, Hack and Schubert (1986) showed that a model vortex developed more rapidly as the heating maximum was placed in the lower levels. Our result that diabatic cooling in the low levels suppress the development of the simulated tropical cyclone is consistent with their studies.

7.3 Central pressure difference

As shown in Fig. 4 and Fig. 5, difference in central pressure between the CA and AS becomes noticeable about 30 hours, i.e., 27 hours after the appearance of areal mean sea-level pressure difference. In this section, in order to explain the delayed appearance of the central pressure difference, we discuss the relationship between radial profile of some quantities at a pre-developmental stage, and the subsequent storm development.

Before the discussion, several quantities are defined. Firstly, the azimuthally averaged low-level moisture flux convergence (MFC) is defined as

$$\text{MFC} = - \int_P \{ \partial(r\bar{q}\bar{u}) / r \partial r \} dp / g,$$

where q is the specific humidity, u is the radial wind velocity (we define u to be positive for inflow), r is radius, dp is the vertical pressure increment, and g is the gravity acceleration. The integration is carried out with respect to pressure over the depth of 350 hPa just above the sea surface. The over-bar represents an azimuthal mean. Secondly, temperature deviation is defined as deviation of temperature from a reference temperature, where the reference is calculated at each pressure level by making average within a circular area, bounded by a circle with a radius of 700 km, centered on the forecast storm location.

Figure 16 shows the time-averaged (from initial to 24 hours during which the central pressure difference between the two simulation is not apparent) radial profile of temperature deviation at 950 hPa, sea-level pressure gradient, inflow at 950 hPa, MFC, and rainfall rate. The CA temperature deviation is always less than that in the AS and has negative values outside the 150 km radius, which is mainly caused by the low-level cooling with the CA scheme. Radial

Fig. 16. The radial profile of time averaged (from initial to 24 hours) (a) temperature deviation at 950 hPa, (b) sea-level pressure gradient, (c) inflow at 950 hPa, (d) moisture flux convergence in the lowest 300 hPa, (e) rainfall rate, in forecast experiments with moist convective adjustment (CA) and Arakawa-Schubert scheme (AS).

profile of the low-level temperature probably influences that of sea-level pressure. In fact, the peak location of the sea-level pressure gradient in the AS is farther from the storm center than that in the CA. Therefore, the CA inflow is stronger than the AS one inside the 100 km radius. On the contrary, the AS inflow is stronger than the CA one from 125 to 300 km radius. Simultaneously, the AS MFC is larger than the CA one from about 100 to 200 km radius. It should be noted that the radial profile of rainfall rate is quite similar to MFC's (i.e., AS's are larger than CA's from around 100 to 200 km radius, and vice versa outside the 225 km radius.) The result suggests that the radial profile of precipitation within the core region is mainly accounted for by that of MFC. The result also implies that contribution from the middle and high-level moisture flux is very small. The rainfall in the AS is rapidly decreased with radius whereas the CA one is relatively flat. The peaks of the rainfall in the CA and

AS are located on the 75 and 100 km radii, respectively. The result suggests that the eyewall in the AS is more distant from the storm center than that in the CA (from now on we refer to the distance as "eye size" for convenience).

The difference in "eye size" between the CA and AS probably results in the subsequent appearance of the difference in central pressure between the two. One possible mechanism for that is associated with the moisture supply from outside the core region. The larger "eye size" affords more room for obtaining moisture to the storm. Another possible mechanism is related to eyewall contraction. The larger "eye size" affords more room for shrinking to the eyewall. Actually the eyewall of the AS storm significantly contracts after 24 hours assuming that radius of maximum tangential wind and eyewall are approximately equal (Fig. 17), followed by deepening of the storm and manifestation of the central pressure difference between the two integrations. Willoughby (1990) showed that the mechanism of tropical cyclone intensification involved a contracting eyewall, on the basis of aircraft observations collected over 13 years for 19 Atlantic hurricanes and tropical storms.

On the basis of the results mentioned above, one can say that the early-stage difference in thermal structure at the low levels, which is mainly caused by the low-level cooling with the CA scheme, accounts for the following two things. One is the quick appearance of areal-mean pressure difference, which is directly explained by the effect of cumulus parameterization on mass field described in Section 5. Another is the delayed appearance of central pressure difference, which is ultimately explained by the effect of cumulus parameterization on temperature field. That is, a different radial profile of the low-level temperature between the two accounts for the "eye size" difference. The difference in the "eye size" during the pre-developmental stage (i.e., until 24 hours) influences the difference in subsequent (i.e., after 30 hours) central pressure deepening.

7.4 Verification against observational data

So far the discussion have been confined to the comparison between the two simulations. To confirm the reliability of the simulations, it is to be desirable to compare them against observations as much as possible. The number of reliable sounding data available near the integration period is only one. This is drop-sonde data (Elsberry 1990) valid at 0400 UTC 17 September near the center of Flo.

Warm core structures, which is one of the typical phenomena associated with tropical cyclones, are compared between the sonde data and simulated storms in terms of temperature deviation. The temperature deviation is defined as the deviation of the sonde datum from a reference temperature,

Fig. 17. Radial profile of azimuthally averaged 950 hPa tangential wind velocity at 24, 30, and 36 hour in forecast experiments with moist convective adjustment (CA) and Arakawa-Schubert scheme (AS).

Fig. 18. Vertical profile of temperature deviation at the storm center from reference temperature. OBS: Temperature at the center and reference temperature are obtained by drop-sonde data (0400 UTC 17 September) and GANAL (0600 UTC 17 September), respectively. CA, AS: Both temperature at the center and reference temperature are obtained by JSM forecast fields (72 hours valid at 0000 UTC 17 September) with moist convective adjustment (CA) and Arakawa-Schubert scheme (AS), respectively.

where the reference temperature is calculated using GANAL (0600 UTC 17 September) at each pressure level, by making an average within a circular area bounded by a circle with a radius of 1000 km, centered on the storm location. The temperature deviations are also calculated for the simulated storms at 72 hours, valid at 0000 UTC 17 September.

Figure 18 shows the considerably large temperature deviations from the middle- to upper-troposphere in the sounding profile. In both simulated profiles, however, the temperature anomalies between 700 to 400 hPa are about 5 K less than for the sonde data. The result is probably because of rough representation of the simulated core structure, due to insufficient model resolution. We speculated that downward motion within the eye is not sufficiently represented in the simulations because the simulated quantity coincides well with the observed one at 850 and 300 hPa. Willoughby (1998) showed that an inversion layer existed at around 750 hPa within the eye of Flo. The subsidence is located above the layer in his conceptual model for the tropical cyclone eye. The deficiency of the temperature anomalies in the simulations could be explained by

weaker subsidence within the eye, although we cannot know the magnitude of actual subsidence.

It is also found that the degree of the deficiency on the temperature anomalies in the AS is less than that in the CA. The result suggests that the AS thermal structure is more realistic than the CA, consistent with the results for the central pressure.

8. Summary and conclusion

Numerical experiments on intensity forecasts of tropical cyclones have been conducted using a limited-area numerical model. Typhoon Flo (1990), which developed rapidly during SPEC-TRUM, TCM-90, and Typhoon-90, was selected as a target of our study. It has been demonstrated by many numerical studies that the intensity of simulated tropical cyclone strongly depends on the choice of cumulus parameterization. In our numerical experiments, the model storm corresponding to Flo intensified more rapidly with an Arakawa-Schubert scheme (AS) than with a moist convective adjustment scheme (CA). In the present study the reason why the different intensity forecasts are obtained according to different formulations of cumulus parameterization is explored.

According to the results of our numerical experiments, cumulus parameterization has a strong influence on total air mass within the core region from the early stage of the integration, although it has no apparent impact on the central pressure during the early period. The main reason for the result for the total air mass is that cooling due to moist convective adjustment is remarkable in the lower troposphere. The cooling results in rapid temperature drop in the low levels. It follows that the total air mass within the core region of the CA storm becomes greater than that of the AS one. The result, that the low-level cooling inhibits the intensification of the simulated tropical cyclone, is consistent with other numerical studies.

On the other hand, the delayed appearance of the central pressure difference between the CA and AS is related to the "eye size" difference, which is influenced by the different radial profile of the low-level temperature. The different "eye size" during the pre-developmental stage brings about the difference in subsequent central pressure deepening.

An additional forecast experiment was carried out to confirm that cooling in the lower troposphere is a dominant factor to suppress the intensification of the simulated tropical cyclone. In the experiment, we modified the moist convective adjustment so that it does not cool the air at the low levels. The modified version (MS) produces lower values of areal mean sea-level pressure $P(0,250)$ compared with the CA one from the early period of the forecast. The result supports the hypothesis that temperature in the low levels just before the developing stage of the storm

is one of important factors to control the subsequent evolution of the storm intensity.

It is not obvious, however, how long the inhibition effect of cold air masses formed at the early stage of the integration lasts and how much it affects subsequent storm evolution. The regionally averaged sea-level pressure in the AS lowered at a greater rate in comparison to the CA. Cooling by the moist convective adjustment in the lower troposphere occurs only at the early part of the 72 hours forecast. From that time on, heating rates at the low levels produced by the two cumulus parameterization schemes are not so different from each other. On the basis of these results, it is speculated that the low-level cooling suppresses the early-stage storm intensification and delays the subsequent storm intensification. However, other factors must be sought for the reason why the difference in total air mass within the core region between the two experiments increases with forecast time. It still remains to investigate more clearly diabatic heating (or cooling) and drying (or moistening) effects of cumulus parameterization on the subsequent model-storm evolution.

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台風 Flo (1990) の強度予報に対する積雲対流のパラメタリゼーションの影響

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領域モデルを用いて、二種類の積雲対流のパラメタリゼーションが台風の強度に及ぼすインパクトを調べた。使用した積雲対流のパラメタリゼーションは湿潤対流調節と荒川-シューバートスキームである。事例として台風 Flo (T9019) を用いた。

台風のコア領域 (中心から半径 250 km 以内) で平均した海面気圧は、湿潤対流調節と荒川-シューバートスキームを用いた場合とで予報初期から大きく異なり、前者より後者の方が低くなった。中心気圧についても、前者より後者の方が深まる度合が大きく、観測により近かった。しかし、両者の中心気圧の違いはコア領域平均海面気圧の違いよりかなり遅れて出現した。

コア領域平均海面気圧の違いは、予報初期の対流圏下層における気温時間変化量の違いに起因する。湿潤対流調節を用いた場合、下層での冷却が気温の急降下をもたらし、荒川-シューバートスキームを用いたときよりもコア領域平均海面気圧が高くなった。一方、中心気圧の違いは壁雲の位置の差に起因すると考えられる。パラメタリゼーションに影響された下層温度場によって半径方向の降水分布、すなわち壁雲の位置に差が生じ、この差が後に中心気圧の違いをもたらしたと考えられる。対流圏下層での冷却を抑制するように湿潤対流調節を変更して同様な数値実験を行った。その結果は台風のコア領域での冷却が台風の発達を妨げていることを支持するものであった。