A Representation of Cumulus-scale Effects in a Mesoscale-Convection-Resolving Model for Tropical Cyclones

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Abstract

A new mesoscale-convection-resolving model (MCRM) for tropical cyclones (TCs) is proposed along a line of Yamasaki's (1986) model in which mesoscale organized convection (MC) is resolved by the grid, but cumulus-scale convection (CC) is treated as the subgrid-scale. The most significant difference from Yamasaki's model is that Kuo's (1965) formulation is used to represent the CC-effects.

Some important modifications from Kuo are made: (1) it is assumed that part of moisture converged in an air column is used for CC and that the ratio depends on low-level convergence and the degree of latent instability; (2) the ratio of moisture consumed for heating to moisture redistributed by CC is assumed to be significantly larger than that given by Kuo; (3) turbulent and dynamic entrainment and detrainment are taken into account so that the vertical profile of heating, which is crucial to MC, may be effectively controlled; (4) cloud water and rainwater of the subgrid-scale are treated with prognostic equations.

Numerical experiments are performed with an axisymmetric model having a horizontal grid size of 10 km. The results are discussed compared with those from a 1 km grid model that can resolve CC. It is shown that the new model can simulate important features of MC and a TC, provided that the values of model parameters are specified properly.

The CC-effects are made clear by comparison with a case that does not include them. It is shown that CC has two different effects, depending on the low-level relative humidity. One is to give realistic growth of MC through the upward transports of heat and moisture under a very humid condition, and the other is to enhance the formation and growth of MC when the air is not very humid.

This study suggests that though Kuo's parameterization was not proposed as that representing the CC-effects, a reasonable model can be developed when Kuo's formulation is adopted in the framework of the MCRM. Important aspects of the model that contribute to realistic simulation of MC and a TC are discussed in comparison with those of Yamasaki's model.

1. Introduction

As is well known, the energy source of a tropical cyclone (TC) is latent heat of condensation of water vapor. The latent heat is directly used as the energy of convection, known as cumulonimbus and cumulus clouds. An ensemble (a group)

of these convective clouds creates the cyclone-scale temperature gradient and available potential energy, part of which is converted into kinetic energy, and thereby a TC forms and develops. In numerical models of TCs, convective clouds have often been treated as the subgrid-scale, and the effects of an ensemble of the clouds have been incorporated by parameterization techniques. Modeling studies of this type started in the 1960s (Ooyama 1964, 1969; Kuo 1965; Yamasaki 1968; Rosenthal 1970; Anthes 1972; Kurihara and Tuleya 1974 and others), and have been still made in recent years. Although these models have been quite useful for

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understanding many aspects of TCs, one can also argue that the use of such models has often led to inappropriate understanding of the important mechanisms of TCs.

On the opposite side of these models, there are explicit models that resolve individual clouds without parameterization. In these models, the use of a small horizontal grid size of 100 m~1 km is required. In an axisymmetric model it is quite easy to use such a fine grid. In fact, studies with a 500 m grid (Yamasaki 1977, 1983), and with a 1 km grid (Nasuno and Yamasaki 1997), have been made to reveal some of the essential mechanisms of TCs and mesoscale organization of convection. The impacts of the inclusion of the ice phase could be also discussed with this type of models (Lord et al. 1984; Willoughby et al. 1984a). Though the application of an axisymmetric model is limited, many important mechanisms of TCs can be fairly represented with it. On the other hand, the use of such high resolution is not easy for a three-dimensional model because of computational constraints. Recently, it has become possible to use 2~5 km grids for numerical simulation under restricted conditions (Liu et al. 1997, 1999). However, these grid sizes are not necessarily fine enough to resolve individual clouds appropriately. A three-dimensional model, with a grid size of 1 km or less, can be used only for a quite limited domain for the present.

The models that will be widely used in the forthcoming decade as a powerful tool not only for understanding of TCs, but also for numerical prediction, will be those with 5~20 km grids. In such coarse grid models it is quite important to take account of the effects of subgrid-scale convection. Subgrid-scale convection here means convection with horizontal scales of several hundred meters to several kilometers, and time scales of a few tens of minutes to one hour. In this paper, it is referred to as cumulus-scale convection, which is abbreviated to CC¹.

The CC is very often organized, and it forms an ensemble of convective clouds that lasts for a few hours or more. Such an ensemble is commonly seen in TCs and other convective phenomena. In the case of TCs, each convective echo cell observed in

spiral rainbands (e.g., Powell 1990a; Barnes et al. 1991) can be considered to be an organized form of CC. The representative time scale is a few hours (about 3 hours) under usual conditions. Such organized convection is referred to as mesoscale organized convection MC, which is emphasized as one of the important modes of convection (Yamasaki 1983). The significance of MC seems to be supported by many observational studies in the past.

With the use of 5~20 km grids, MC can be resolved to a certain extent, while CC should be implicitly taken into account. A model of this type is referred to as a mesoscale-convection-resolving model, which is called MCRM in this paper. In contrast, models that resolve CC will be referred to as cumulus-scale-convection-resolving model (CCRM)².

A MCRM for TCs was developed by Yamasaki (1986, 1987). Since the horizontal scale of MC is not so large, truncation errors are significant for 5~20 km grids. Nevertheless, it is important to resolve MC by the grid. In models that treat both CC and MC implicitly, some of the essential mechanisms of TCs are not properly simulated. This recognition had been obtained from the studies with a CCRM (Yamasaki 1977, 1983), and was a basis of the MCRM of Yamasaki (1986), which is referred to as Y86, hereafter. Though scale separation between CC and MC are not necessarily clear in many cases, as was mentioned in Y86, the importance of treating MC explicitly was emphasized by Y86. The present study is made along the same line as Y86.

One of the most important features of the MCRM of Y86 is that grid-scale (grid-resolved) cloud water and rainwater were taken to be predicted variables. This had not been done in TC models with cloud parameterization. It is also characteristic of Y86 that not only grid-scale, but also cumulus-scale (subgrid-scale) cloud water and rainwater were treated as important variables. This is because the effects of evaporation of subgrid-scale rainwater and the downdraft in the subcloud layer should be incorporated to describe MC.

Another important aspect of Y86 is that an as-

1 The term that has been widely used for more than forty years is 'cumulus convection'. The CC (cumulus-scale convection) used in this paper is not essentially different from cumulus convection, but this term is used to emphasize the importance of it as a component of mesoscale organized convection (MC) rather than as isolated convection.

2 MCRM can be abbreviated to MRM, and CCRM to CRM (cumulus-scale-resolving model). The term CRM is now widely used as an abbreviation of 'Cloud Resolving Model'. In order to avoid confusion we use CCRM. These terms CCRM and MCRM are used, based on the supposition that both CC and MC have their own significant implications in modeling TCs.

sumption for heat release by CC was made so as to reproduce the properties of MC. Equations for potential temperature and water vapor mixing ratio are similar to those in Ooyama (1971) and Arakawa and Schubert (1974) in that cloud mass flux is utilized. However, the assumption for the cloud mass flux (heating) is quite different. In Y86 it was based on the results from linear theories (Syono and Yamasaki 1966; Yamasaki 1969), and many studies on TCs and convection in the past. Needless to say, the assumption in Y86 should not be satisfactory. We have to find a better assumption by any different idea. It is important to develop a different model and compare experimental results.

In the present study we construct a model which adopts Kuo's (1965) idea as an assumption concerning the effects of CC. The essence of his idea is that part of moisture converged in an air column is used for heating, and the vertical distribution of the heating depends on the temperature difference between the grid-scale (environment) and the implicit cloud. Such an assumption has been criticized to be less physical than that adopting cloud mass flux, because the effects of mass displacement by convection are not properly incorporated (Ooyama 1971). However, it does not necessarily mean that the performance of Kuo type models is inferior to that of the cloud mass flux type model (Y86) because the latter also contains its own uncertain assumptions. Better understanding of TCs with Kuo type models should also be very useful to improve models of the cloud mass flux type.

It should be emphasized that, though the present model utilizes Kuo's (1965) idea, it is fundamentally different from previous Kuo type models (Rosenthal 1970; Anthes 1972, 1977 and others). The most important difference is the recognition that MC has to be resolved by the grid, as mentioned above. In addition, since evaporative cooling of rainwater in the subcloud layer plays an essential role in MC in many situations, cloud water and rainwater are predicted not only for the grid-resolved but also for the cumulus-scale, which has not been done in the Kuo type models.

As an extension of Y86, we make two major improvements in this study. One is the prognostic treatment for cumulus-scale cloud water³. It was diagnostically determined in Y86, as in Arakawa and Schubert (1974), though a prognostic equation was used for cumulus-scale rainwater. Because of the diagnostic treatment, conservation of the total water was only approximately satisfied in Y86 (also

in Arakawa and Schubert 1974). In the present model the total water is strictly conserved in a simple way. Another merit of the prognostic treatment is that evaporation of the cumulus-scale cloud water can be taken into account after CC is terminated. In Y86 it was allowed only through the detrainment from implicit clouds.

The second improvement is the use of a nonhydrostatic model (though an assumption of the hydrostatic equilibrium is justified for the MCRM). A nonhydrostatic model also has an advantage of the continuity to CCRM when the grid size is reduced. As the grid size is taken smaller, subgrid-scale effects should be reduced and the grid-resolved convection should possess the properties similar to CC. In the present study, however, we take the stand point that CC is treated as the subgrid-scale, regardless of the grid size. A merit of the use of a smaller grid size is to make truncation errors smaller.

As a step to develop a three-dimensional model, we use an axisymmetric model in the present study. We will present the results from three numerical experiments. The first is the result from a CCRM with a 1 km grid, which is used as a reference for other cases. The second experiment is performed with a 10 km grid without cumulus-scale parameterization⁴ to clarify the importance and impacts of parameterization. In the third, the result from a case with parameterization is presented.

2. Model

In this study an axisymmetric nonhydrostatic CCRM of Yamasaki (1983) is modified to be used for horizontal grid sizes of 5~10 km. In order to incorporate (or parameterize) the effects of CC (the

3 The whole of this Introduction including this paragraph was written on the premise that this paper was to be published before Yamasaki (2001, *J. Met. Soc. Japan*, 79, No.2) in which a similar treatment is made. Correspondingly, the Introduction of Yamasaki (2001) is written on the supposition that the present paper has been already published.

4 The term 'parameterization' is not necessarily appropriate because scale separation between CC and MC is not very clear. However, this term may be used to imply a representation of the cumulus-scale effects that are treated as the subgrid-scale. It should be also remarked that the meaning of the term 'cumulus-scale parameterization' is quite different from that of 'cumulus parameterization' which has been widely used in large-scale modeling studies, in that the latter parameterizes not only the cumulus-scale effects but also mesoscale effects.

CC-effects), cumulus-scale cloud water and rainwater as well as those for the grid-resolved are treated, as done in Y86. Different from Y86, the CC-effects in the present model are parameterized by a modified Kuo's (1965) method.

2.1 Governing equations

a. equations for grid-resolved variables

Equations for grid-resolved variables are based on Yamasaki (1983), but with the CC-effects included, as indicated by underlines:

$$\frac{dv_r}{dt} - \left(f + \frac{v_\lambda}{r}\right)v_\lambda = -C_p\theta_0 \frac{\partial \pi}{\partial r} + F_H(v_r) + F_V(v_r), \quad (1)$$

$$\frac{dv_\lambda}{dt} + \left(f + \frac{v_\lambda}{r}\right)v_r = F_H(v_\lambda) + F_V(v_\lambda), \quad (2)$$

$$\frac{dw}{dt} = -C_p\theta_0 \frac{\partial \pi}{\partial z} + g \left(\frac{\theta - \theta_0}{\theta_0} + 0.608(q_v - q_{v0}) - q_c - q_r - \underline{q_c^c} - \underline{q_r^c} \right) + F_H(w) + F_V(w), \quad (3)$$

$$\frac{\partial(\rho_0 r v_r)}{r \partial r} + \frac{\partial(\rho_0 w)}{\partial z} = 0, \quad (4)$$

$$\frac{d\theta}{dt} = \frac{L}{c_p \pi} (C - E_r) + \frac{Q^c}{c_p \pi} - \frac{L}{c_p \pi} (\underline{E_c^c} + \underline{E_r^c}) + \underline{D_w(\theta)} + Q_R + F_H(\theta) + F_V(\theta), \quad (5)$$

$$\frac{dq_v}{dt} = -C + E_r + \underline{\Delta q_v} + \underline{E_c^c} + \underline{E_r^c} + \underline{D_w(q_v)} + F_H(q_v) + F_V(q_v), \quad (6)$$

$$\frac{dq_c}{dt} = C - A - C_o + \underline{D_w(q_c)} + F_H(q_c) + F_V(q_c), \quad (7)$$

$$\frac{dq_r}{dt} = A + C_o - E_r + \frac{1}{\rho_0} \frac{\partial}{\partial z} (\rho_0 q_r V_T), \quad (8)$$

where

$$\frac{d}{dt} = \frac{\partial}{\partial t} + v_r \frac{\partial}{\partial r} + w \frac{\partial}{\partial z}.$$

Equations (1)~(3) are equations of motion for radial velocity v_r , tangential velocity v_λ , and vertical velocity w . Equation (4) is the continuity equation for deep convection. The thermodynamic vari-

ables are potential temperature θ , mixing ratios of water vapor q_v , cloud water q_c and rainwater q_r , and these are predicted by Eqs.(5)~(8). Another dependent variable is pressure deviation (Exner function) π . The variables with subscript 0 (θ_0 , π_0 , q_{v0} , and density ρ_0) indicate those for the basic state. Independent variables are r : radius (distance from the cyclone center), z : height and t : time.

The effects of eddy diffusion in the horizontal and vertical directions are indicated by F_H and F_V , respectively. The effect of radiative cooling Q_R is included in the thermodynamic Eq.(5). Cloud microphysical processes are represented by C : condensation rate (negative value indicates the rate of evaporation of cloud water), A : the rate of auto-conversion from cloud water to rainwater, C_o : the rate of collection of cloud water by rainwater, E_r : the rate of evaporation of rainwater and V_T : the terminal velocity of rainwater. In Eq.(5) L and C_p are latent heat of condensation and specific heat of dry air at constant pressure, respectively.

As for the cumulus-scale variables, q_c^c and q_r^c indicate mixing ratios of cumulus-scale cloud water and rainwater, respectively. The total cloud water (rainwater) in a grid area is represented by the sum of q_c and q_c^c (q_r and q_r^c).

In Eqs.(5) and (6), the cumulus-scale quantities Q^c and Δq_v indicate the combined effects of condensation and vertical transport of heat and moisture by CC, respectively. If conditions for the occurrence of CC are not satisfied, these are omitted. Drag forces by cumulus-scale cloud water q_c^c and rainwater q_r^c are included in Eq.(3), together with those by grid-resolved cloud water q_c and rainwater q_r . The evaporation rate of cumulus-scale cloud water is denoted by E_c^c and that of rainwater by E_r^c . While CC occurs, the evaporation of q_c^c and q_r^c is inhibited except for the boundary layer. When CC is terminated, q_c^c and q_r^c are assumed to evaporate at a finite rate in an unsaturated area. At this stage, q_c^c and q_r^c represent remnants of clouds that have already become non-convective. The notation D_w in Eqs.(5)~(7) indicates the effect of vertical mixing of low-level air by cumulus-scale downdraft. Formulations for these are described later.

b. Equations for cumulus-scale variables

In Y86 cumulus-scale rainwater q_r^c was prognostically treated, but cumulus-scale cloud water q_c^c was diagnostically determined. In the present study, both q_c^c and q_r^c are taken as prognostic variables. Equations for q_c^c and q_r^c are assumed to be

$$\frac{dq_c^c}{dt} = C^c - A^c - C_o^c - E_c^c + D_w(q_c^c) + F_H(q_c^c) + F_V(q_c^c), \quad (9)$$

$$\frac{dq_r^c}{dt} = A^c + C_o^c - E_r^c + \frac{1}{\rho_0} \frac{\partial}{\partial z} (\rho_0 q_r^c V_T^c), \quad (10)$$

where cloud microphysical processes for q_c^c and q_r^c are denoted by superscript c .

As seen in Eq.(9), q_c^c is generated from C^c , the water vapor condensation by CC. Since the vertical profile of condensation in clouds was not specified in Kuo's (1965) model, we use Q^c/L as an approximation of C^c , though the vertical profile of the heating rate $Q^c(z)$ includes not only C^c , but also the vertical heat transport by CC.

Another uncertainty exists in advection terms in Eqs.(9) and (10). The vertical velocity in the advection terms should be of the cumulus-scale, but the grid-resolved velocity is used as a crude approximation, as done in Y86. This treatment is acceptable to some extent because q_c^c and q_r^c are mixing ratios averaged over a grid area.

Cumulus-scale rainwater q_r^c is produced by A^c (autoconversion of q_c^c to q_r^c) and C_o^c (collection of q_c^c by q_r^c), and it falls relative to the air at the terminal velocity V_T^c . The evaporation of cumulus-scale cloud water E_c^c , and that of rainwater E_r^c are also included in Eqs.(5) and (6). That is, when the subgrid-scale evaporation occurs, q_c^c and q_r^c are converted to grid-resolved water vapor q_v .

2.2 Cumulus-scale effects

a. Cumulus-scale heating

(modified Kuo's assumption)

The CC-effects on potential temperature Q^c , and on water vapor Δq_v , are determined with a modified Kuo's (1965) method. The moisture convergence in an air column is given by

$$I = - \int_{z_B}^{z_W} \nabla \cdot (\rho \vec{v} q_v) dz, \quad (11)$$

where z_B is the cloud base height, z_W is the level where vertical motion turns downward, and $\vec{v}$ is the three-dimensional wind vector. The use of z_W instead of the cloud top height z_T is to avoid underestimation of I when z_W is at a low-level. In such a situation, drying due to subsidence above z_W reduces the net moisture convergence in the air column, even if low-level convergence is considerably large. When the level of z_W is higher than z_T ,

the value of z_T is used for z_W .

One of the important modifications from Kuo (1965) is that some part of moisture convergence I is assumed to be used for CC. Its ratio is given by a parameter F_I . An assumption of F_I will be described in subsection 2.3. The use of unity for F_I is identical with Kuo's assumption. The moisture used for CC ($F_I I$) is partly consumed by condensation, while the rest is vertically redistributed to moisten the air column. In this respect an important modification is made on the ratio of water vapor storage b (Kuo 1974), which is given by

$$b = \frac{F_b M_V}{M_Q + F_b M_V}, \quad (12)$$

where

$$M_Q = \frac{C_p}{L} \int_{z_B}^{z_T} \rho (T^c - T^*) dz, \quad (13)$$

$$M_V = \int_{z_B}^{z_T} \rho (q_v^c - q_v) dz. \quad (14)$$

The quantities M_Q and M_V indicate the amounts of moisture required when the grid-resolved temperature T^* and moisture q_v increase to be the cloud temperature T^c and moisture q_v^c , respectively, where T^* is the temperature in the absence of the CC-effects and grid-resolved condensation. The formulation (12)~(14) is the same as those in Kuo (1965), but a parameter F_b is introduced in (12) because convective heating is generally weak under Kuo's (1965) assumption. When F_I and F_b are unity, the formulation is identical with that of Kuo (1965). Based on numerical experiments, the value of F_b is chosen to be a constant of 0.3 in this study.

As in Kuo (1965), the vertical distributions of heat and moisture are assumed to be proportional to the difference between the variables of the implicit cloud and environment (approximated by the grid-resolved). Then cumulus-scale heating Q^c and moisture modification Δq_v are given by

$$Q^c = C_p (T^c - T^*) \frac{F_I I}{M_t}, \quad (15)$$

$$\Delta q_v = F_I \left(v_r \frac{\partial}{\partial r} + w \frac{\partial}{\partial z} \right) q_v + F_b (q_v^c - q_v) \frac{F_I I}{M_t}, \quad (16)$$

where

$$M_t = M_Q + F_b M_V. \quad (17)$$

The use of T^* instead of T in (13) and (15) has an advantage that the effect of adiabatic cooling can be taken into account in determining the cumulus-scale heat profile; more heat may be released at levels of stronger adiabatic cooling.

b. Cloud model

In Arakawa and Schubert (1974), the effects of convective clouds on the large-scale are represented by an ensemble of clouds of various types with different top levels. On the other hand, in Y86 and in the present model, cumulus-scale clouds with a single top level are treated, though the time variation in the cloud top is taken into account.

The cloud properties T^c , q_v^c and z_T are determined by the same cloud model as Y86. The use of the cloud mass flux, balance condition of moist static energy and the saturation condition are the same as those in Arakawa and Schubert (1974). As for the balance of moist static energy, considering a layer between l (top) and $l+1$ (bottom) in a numerical model, it is given by

$$-\eta_l h_l^c + \eta_{l+1} h_{l+1}^c + (\eta_l - \eta_{l+1}) \tilde{h}_l - (D_T)_l \frac{h_l^c + h_{l+1}^c}{2} + (E_T)_l \frac{h_l + h_{l+1}}{2} = 0, \quad (18)$$

where

$$\tilde{h}_l = \begin{cases} \frac{h_l + h_{l+1}}{2} & (\eta_l > \eta_{l+1}) \\ \frac{h_l^c + h_{l+1}^c}{2} & (\eta_l < \eta_{l+1}), \end{cases}$$

and η indicates the non-dimensional cloud mass flux normalized by the amount at the cloud base. Moist static energy in the environment is defined by $h = C_p T_h + gz + Lq_v$, where $T_h = T(p_0/p)^{R/C_p}$ (p : pressure and p_0 : basic state pressure or pressure at the undisturbed state) is used instead of T , as done in Y86 because pressure effect has to be taken into account in the TC models. The quantity h^c is moist static energy in the implicit clouds.

The first and second terms of Eq.(18) indicate the moist static energy flux through the top and bottom of the layer, respectively. The third term is called dynamic entrainment/detrainment, which is owing to vertical convergence of mass. The fourth and fifth terms represent turbulent detrainment and entrainment, respectively. These are assumed to exist at all levels. Turbulent detrainment and entrainment in the l -th layer is given in the same form as in Y86:

$$(D_T)_l = (E_T)_l = E_D \eta_{l+1} (\Delta z)_l, \quad (19)$$

where $(\Delta z)_l$ is the depth of the l -th layer and E_D is a parameter concerning the turbulent detrainment/entrainment rate. It has a critical influence on the vertical distributions of the cloud properties and Q_c , as shown later.

The normalized cloud mass flux η is assumed to be of the same form as in Y86:

$$\eta = 1.0 + C_\eta \{T^c(1 + 0.608q_v^c - q_c^c) - T(1 + 0.608q_v - q_c)\}, \quad (20)$$

where C_η is a parameter that relates the cloud mass flux to buoyancy. By this assumption dynamic detrainment or entrainment can occur at all cloud levels. In Y86 assumption (20) was of great importance because cumulus-scale heating was assumed to be closely related with the cloud mass flux. In the present model this assumption is related to the heating only through the determination of T^c and q_v^c .

An air parcel is assumed to rise from a certain level z^* in the boundary layer. The moist static energy at the cloud base $h^c(z_B)$ is assumed to be equal to $h(z^*)$. Cloud temperature and water vapor at the cloud base are given by dry or moist adiabatic ascent of the air. Thus, the vertical distributions of T^c , q_v^c and η are determined from the three assumptions; (18), (20) and the saturation condition. The cloud top z_T is approximately determined as the level of zero buoyancy.

c. The effect of cumulus-scale downdraft

The evaporation of rainwater in downdrafts is considered to be an important factor to MC in a TC (Yamasaki 1977, 1983), which has been supported by observational studies (Barnes et al. 1983, 1991; Ishihara et al. 1986; Powell 1990a, b). Therefore, in the MCRM of Y86 the effect of drying due to the cumulus-scale downdraft was incorporated so that the evaporation may occur. The effect of cumulus-scale downdraft is formulated analogous to turbulent diffusion, following Y86:

$$D_w(s) = \frac{\partial}{\rho_0 \partial z} \left(\rho_0 K_C \frac{\partial s}{\partial z} \right), \quad (21)$$

where s represents θ , q_v , q_c or q_c^c , and K_C is a coefficient, which is assumed to be related with vertical velocity and cumulus-scale rainwater at the cloud base:

$$K_C = \begin{cases} K_D \sqrt{q_r^c(z_B)} w(z_B) & (w(z_B) > 0, \quad q_r^c(z_B) > 0) \\ 0 & (\text{otherwise}), \end{cases} \quad (22)$$

$$F_I = \begin{cases} 1 & (R_I \geq R_2), \\ \frac{R_I - R_1}{R_2 - R_1} & (R_1 < R_I < R_2), \\ 0 & (R_I \leq R_1), \end{cases} \quad (24)$$

where K_D is a parameter denoting the value of K_C for $q_r^c(z_B) = 1 \text{ g kg}^{-1}$ and $w(z_B) = 1 \text{ m s}^{-1}$. The above assumptions, particularly (22) are not necessarily appropriate and should be improved in future studies. However, it seems to be reasonable in the sense that cumulus-scale downdrafts become significant as rainwater near the cloud base increases.

2.3 Conditions for the occurrence of the cumulus-scale heating and moistening

In this subsection several conditions for the occurrence of CC will be described.

(c1) moisture convergence

As mentioned in 2.2.1, moisture convergence ($I > 0$) is one of the necessary conditions for the cumulus-scale heating and moistening. This assumption is the same as in Kuo (1965).

(c2) latent instability

In common with Kuo (1965), latent instability (positive buoyancy for a rising air parcel, or CAPE, convective available potential energy) is checked for the occurrence of the cumulus-scale heating. If the air parcel ascending from the boundary layer has positive buoyancy above the cloud base, CC may occur. In the present study the buoyancy B is simply defined as

$$B = T^c - T.$$

The conditions (c1) and (c2) are not sufficient for proper representation of formation and evolution of MC. Additional conditions (c3)~(c5) are introduced.

(c3) low-level moisture convergence

It is assumed that the activity of CC is closely related with low-level moisture convergence I_L , defined as

$$I_L = - \int_0^{z_{LT}} \nabla \cdot (\rho \vec{v} q_v) dz, \quad (23)$$

where z_{LT} is the upper level of the integration (an altitude of 2.4 km is assumed in this study). Therefore, F_I (the ratio of moisture used for CC to total convergence I) is determined from R_I , the ratio of I_L to the total convergence I .

where R_1 and R_2 are thresholds of R_I . The ratio R_I is usually large at the formative stage of MC, and it decreases as MC evolves. When low-level moisture convergence accounts for more than a certain portion of I , F_I is assumed to be unity, that is, all the I is used for CC. When low-level convergence considerably decreases, F_I is zero, that is, CC does not exist even if total moisture convergence is positive. For smooth transition between the two situations the value of F_I is assumed to be continuously varied. The condition (24) qualitatively incorporates the idea that the boundary layer convergence is important in tropical cyclones (Ooyama 1964, 1969; Yamasaki 1968, 1986).

(c4) deviation from a moist neutral state

As the stratification of the air column approaches a state that the grid-resolved temperature is close to the cloud temperature T^c , the effects of CC diminish. Therefore, F_I is assumed to decrease, depending on the deviation from the neutral state, in the following form:

$$F_I = \begin{cases} 1 & (\Delta T_m \geq \Delta T_2) \\ \frac{\Delta T_m - \Delta T_1}{\Delta T_2 - \Delta T_1} & (\Delta T_1 < \Delta T_m < \Delta T_2) \\ 0 & (\Delta T_m \leq \Delta T_1), \end{cases} \quad (25)$$

where ΔT_m is the cloud-layer mean difference between cloud temperature T^c and grid-resolved temperature T , and ΔT_1 and ΔT_2 are the thresholds of ΔT_m . The parameter F_I is determined in the same way as in (c3). When a considerable amount of ΔT_m exists, all the I is consumed by CC. When a certain degree of stabilization is attained, no CC is assumed. Since two assumptions (24) and (25) are imposed on the specification of F_I , the smaller value is used.

(c5) vertical velocity at low-levels

The existence of a certain degree of upward motion at the cloud base is introduced as the fifth condition for occurrence of CC when q_c^c at z_B is zero or small:

$$\begin{aligned} w(z_B) &> w_c \quad \text{for} \quad q_c^c(z_B) < q_{cc}^c \\ w(z_B) &> 0 \quad \text{for} \quad q_c^c(z_B) > q_{cc}^c, \end{aligned} \quad (26)$$

where w_c and q_{cc}^c are the thresholds of $w(z_B)$ and q_c^c , respectively. A similar condition was introduced in a revised version of Y86 (Yamasaki 1989),

based on the idea that relative humidity in the boundary layer rises and grid-resolved condensation occurs at low-levels prior to the initiation of MC. The threshold q_{cc}^c is introduced because at the decaying stage of MC, when a considerable amount of q_c^c still remains, it seems rather realistic to assume that CC occurs, provided that $w(z_B)$ is positive.

2.4 Cloud microphysical and other processes

Since cumulus-scale cloud water q_c^c is prognostically treated, evaporation of q_c^c can be taken into account. It is assumed that the cloud water evaporates only after CC is terminated. The evaporation does not occur instantly but at a rate given by

$$E_c^c = \begin{cases} \tau^{-1} q_c^c & (q_v < q_s \text{ and } Q^c = 0) \\ 0 & (\text{otherwise}), \end{cases} \quad (27)$$

where τ represents a time constant for evaporation of q_c^c , and q_s is the saturation mixing ratio for the grid-resolved temperature. In the same way, the evaporation of rainwater q_r^c above the boundary layer occurs only when there is no CC. In the boundary layer, the evaporation of q_r^c can always take place. The evaporation rate determined by Kessler (1969) is used.

As for autoconversion of cloud water to rainwater A^c and collection C_0^c , Berry's (1968) and Kessler's (1969) parameterization are used, respectively. The formula used for E_r^c , A^c , C_0^c and V_T^c are given in Yamasaki (1983).

Cloud microphysical processes for the grid-

resolved variables are also formulated in the same way. Although the use of the same values for a 10 km grid is not the best way, these are employed as a crude approximation.

The eddy diffusion terms F_H and F_V are given in the same form as those in Yamasaki (1983). Surface friction and supply of latent and sensible heat from the sea surface are incorporated by the use of the bulk method. Coefficients for surface friction C_D and heat and moisture fluxes C_E are both quantified as 0.0015. The prescribed sea surface temperature is 302 K. Sensible and latent heat fluxes from the sea surface are indispensable to the development and maintenance of TCs (Ooyama 1969).

As to radiation effect, uniform cooling (Q_{RC}) and Newtonian cooling (Q_{RN}) are considered in the form of:

$$Q_R = -Q_{RC} + Q_{RN}, \quad (28)$$

where Newtonian cooling is written as

$$Q_{RN} = -\alpha_R(\theta - \theta_0).$$

The values of Q_{RC} and α_R are taken to be 0.5 K day⁻¹ and 0.1 K day⁻¹, respectively.

2.5 Design of the numerical experiments

The outline of the numerical model is almost the same as Yamasaki (1983), and Nasuno and Yamasaki (1997), except for the horizontal grid size. It is taken to be 10 km inside the radius of 1,200 km. Though a 10 km grid is used in this study, the

Table 1. Values of parameters for the cumulus-scale effects. Numbers (characters) in parenthesis indicate equations (conditions for the occurrence of CC).

cumulus-scale heating		
F_b	factor of M_V in b (12)	0.3
C_η	factor of buoyancy in η (20)	0.2
E_D	turbulent detrainment/entrainment (19)	3×10^{-4}
cloud microphysical processes		
τ	evaporation of q_c^c (28)	12 min
conditions for the cumulus-scale (threshold values)		
R_1, R_2	(c3) ratio of I_L to I	0.01, 0.07
$\Delta T_1, \Delta T_2$	(c4) air column mean ($T^c - T$)	0.2, 0.5 K
w_c	(c5) $w(z_B)$	0.15 ms^{-1}
q_{cc}^c	(c5) q_c^c	0.1 g kg^{-1}
coefficients for cumulus-scale downdraft		
$K_D(z = 0.9 \text{ km})$	(22)	$10 \text{ ms}^{-1} \text{ g}^{-0.5} \text{ kg}^{0.5}$
$K_D(z = 0.3 \text{ km})$	(22)	$3 \text{ ms}^{-1} \text{ g}^{-0.5} \text{ kg}^{0.5}$

results are qualitatively similar for grid sizes of 5~10 km. The radius of the outer boundary and the height of the top of the domain are 2,760 km and 22.6 km, respectively. The vertical grid size of 600 m is used below 12 km.

In this paper, we will present the results from numerical experiments with three kinds of models. The first experiment is performed with a CCRM using a 1 km grid (underlined terms in Eqs.(1)~(8) omitted). The second experiment is performed with a 10 km grid, other conditions being the same as those in the first. That is, the CC-effects are not incorporated despite using a 10 km grid. The third is performed with the MCRM that includes the CC-effects. The first case is useful as a basis for the development of a better MCRM (together with the observational facts). The comparisons of the second case with other cases would reveal the necessity of the inclusion of the CC-effects (parameterization) in a 10 km grid model, and their impacts on the behavior of convection.

The values of parameters used for the MCRM are given in Table 1. Specification of the param-

eters is carefully done so that simulated convection may represent the main features of MC in the CCRM, and those in observed TC-rainbands. For this purpose, a number of experiments were performed, though we will present one typical case in this paper. As will be shown in the next section, and more fully in a separate paper, parameters concerning the cumulus-scale heating have large impacts on the behavior of MC and TCs.

As for a basic state, tropical mean profiles of temperature and relative humidity are used, as in Yamasaki (1983). The tangential velocity and temperature perturbations at the initial time are given in the same way as Nasuno and Yamasaki (1997). The tangential velocity has a maximum of 12 m s^{-1} at a radius of 180 km, and temperature perturbations are imposed around a radius of 280 km. All the experiments are performed with a time increment of 10 seconds.

3. Results

3.1 The whole development of the tropical cyclone

The general features of the TC obtained from the

Fig. 1. Radius-time cross-sections of (a) rainfall intensity at the surface and (b) tangential velocity at a 900 m height for case CR ($\Delta r = 1 \text{ km}$). Isolines in (a) are 1, 5, 10 mm (10 min)^{-1} . Three different shades in (b) denote tangential velocities more than 18, 25 and 60 m s^{-1} .

Fig. 2. Same as Fig. 1, but for case NP ($\Delta r = 10$ km, without cumulus-scale effects). An isoline of $0.1 \text{ mm (10 min)}^{-1}$ is added in (a).

1 km grid case (referred to as case CR) are essentially the same as those in Yamasaki (1983). Figure 1 shows the time-radius sections of rainfall intensity at the surface and tangential velocity at 900 m height. Several continuous rainfalls correspond to TC-rainbands. A rainfall that lasts for more than a few days at 300~400 km may correspond to a principal band (Willoughby et al. 1984b). As suggested from the increase of tangential velocity, convective activity contributes to the intensification of a TC-scale circulation. Tangential velocity peaks are located at rainfall areas. The rainfalls tend to propagate outward when the tangential wind is weak. As the wind is intensified, they begin to propagate inward owing to the intensification of frictional inflow. When the wind becomes strong near the center, an eyewall is formed by frictional convergence. Rainbands are maintained by successive formation of mesoscale organized convection (MC) with a time scale of several hours. Such processes of tropical cyclone development and organization of MC were more fully discussed in Yamasaki (1983) and Nasuno and Yamasaki (1997).

When a 10 km grid is used without including the CC-effects (case NP), the behavior of the TC is greatly modified. Figure 2 shows the results from case NP. Compared with case CR, some distinct differences are noted. One is the weakness of convection in the outer area. As a result, the tangential velocity is not so intensified as that in case

CR. In contrast to convection in the outer area, vigorous convection occurs near the center, which causes an earlier formation of an eyewall. The time scale of individual rainfalls associated with outward propagating convection is only 1.5~2 hours. The phase velocity of the outward propagation is about 1.4 m s^{-1} , which is much larger than in case CR. It is noteworthy that an inward propagation as seen in case CR does not occur. This is closely related to the non-intensification of the tangential velocity in the outer area. It can be concluded that the intensity and lifetime of convection in case NP are significantly different from those of MC in case CR.

Results from the case with the CC-effects (case P) are shown in Fig. 3. The essential features of the TC and convection are qualitatively similar to those in case CR, though smaller-scale rainfalls are not described inevitably by the use of a 10-km grid (as well as cumulus-scale parameterization). Intense rainfalls due to MC occur with a time scale of several hours, and MC is successively formed on the inner or outer sides of the previous one. The location of the formation depends on the strength of frictional inflow. The intensification of tangential velocity is also similar to that in case CR. Although the formation of an eyewall is somewhat earlier than that in case CR, general features are significantly improved from those in case NP, particularly in the outer area.

The eyewall in case NP forms from local intense

Fig. 3. Same as Fig. 2, but for case P ($\Delta r = 10$ km, including cumulus-scale effects).

circulation close to the center, because convection is scarcely enhanced in the outer area. In case CR and case P, the formation of the eyewall is associated with the TC-scale circulation that has developed by successive formation of MC at various radii. An improvement made by cumulus-scale parameterization is mainly due to the enhancement of MC in the outer area and the resulting suppression of extremely intense convection near the center. These differences in the formation process cause different radii of the eye at a nearly-steady state. The radius is extremely small in case NP, and it is not so small in case CR and case P. As expected from the time evolutions of tangential velocity, surface pressure at the center rapidly decreases in case NP. It is about 960 hPa at 48 h, whereas it is still about 990 hPa in case CR and in case P. At the nearly-steady state it is below 900 hPa in all three cases.

3.2 Life cycle and structure of mesoscale organized convection

We discuss the life cycle and structure of MC under a weak frictional effect in the three cases. In

particular, time evolution of MC in case P will be examined in some detail.

Figure 4 shows vertical cross-sections of the mixing ratio of cloud water in case CR. A life cycle of MC indicated by an arrow in Fig. 1a is shown. Low-level clouds that are inactive at the first stage grow to higher levels (Fig. 4a–b). Cumulus-scale cells (cloud elements) ascend one after another, and form an organized mesoscale cloud at the mature stage (Fig. 4c). Well-developed MC induces intense rainfall and subsequent development of a cold pool, and MC decays (Fig. 4d). The lifetime is about 5 hours, which is a little longer than 3 hours typically seen in Yamasaki (1983). Outside decaying MC, a low-level cloud forms, and grows into new MC.

The vertical structure at the mature stage (corresponding to Fig. 4c) is shown in Fig. 5. At this stage peaks of warming and updraft reach to a 10 km height, and a large amount of rainwater is produced. Near the surface, cooling occurs owing to the evaporation of rainwater. The cold divergent outflow contributes to the successive formation of low-level clouds by lifting warm moist air coming from the outside. At the outer edge of MC

Fig. 4. Vertical cross-sections of cloud water in the life cycle of MC in case CR: (a) formation stage, (b) growing stage, (c) mature stage, and (d) decaying stage. Iso-lines are drawn every 0.3 g kg^{-1} .

a tangential velocity peak (extending upward) is produced by radial, and vertical transport of angular momentum. These characteristics are consistent with those from observational studies (Barnes et al. 1983, 1991; Barnes and Stossmeister 1986; Ishihara et al. 1986; May 1996; Powell 1990a, b; Tabata et al. 1992; Tatehira 1961, 1962).

A life cycle of typical convection in case NP is shown in Fig. 6. As shown in Fig. 2, the lifetime of convection in case NP is about 1.5~2 hours, which is significantly shorter than that of MC in

Fig. 5. Vertical cross-sections at the mature stage corresponding to Fig. 4c; rainwater (Iso-lines: 0.2, 2 and 4 g kg^{-1}), temperature deviation (every 1 K; darkest shade: negative), vertical velocity (every 2 m s^{-1} with -1 m s^{-1}), radial velocity (every 2 m s^{-1} ; shades denote inflow [leftward]), tangential velocity (every 2 m s^{-1}).

Fig. 6. Vertical cross-sections of cloud water in case NP. Isolines and shades are the same as in Fig. 4. (The scale of the abscissa is different from that in Fig. 4.)

Fig. 7. Same as Fig. 6, but for radial velocity corresponding to Fig. 6c (isolines: every 2 m s^{-1}). Shades denote inflow (leftward).

Fig. 8. Vertical cross-sections of cloud water (sum of cumulus-scale and grid-resolved) in the life cycle of MCP: (a) formation stage, (b) growing stage, (c) mature stage and (d) decaying stage. Isolines and shades are the same as in Fig. 4.

case CR. Compared with Fig. 4, we can see that a single cloud (convective cell) grows and ascends, but no subsequent cell is formed below the old cells. This is in contrast to the successive formation of convective cells in case CR. The vertical scale of the circulation is small, and low-level convergence is not maintained in case NP. Thus, convection decays within a short time.

The radial wind velocity at the mature stage (corresponding to Fig. 6c) is shown in Fig. 7. Horizontal convergence is produced near the surface 1

Fig. 9. Vertical cross-sections of MCP at the mature stage (Fig. 8c); rainwater (iso-lines: every 0.5 g kg^{-1} and 0.1 g kg^{-1}), temperature deviation (every 0.5 K), vertical velocity (every 1 m s^{-1} ; upward shaded), radial velocity (every 2 m s^{-1}), tangential velocity (every 2 m s^{-1}).

hour before. However, the low-level convergence is not maintained, and divergence occurs very soon below the cloud, as seen in this figure. Therefore, cloud water is not successively formed at low-levels. This outflow propagates as a gravity wave, and its level goes upward (not shown).

In case P, the behavior of convection greatly differs from that in case NP. A life cycle of a typical MC indicated by an arrow in Fig. 3a (referred to as MCP) is shown in Fig. 8. The presented variable is total cloud water ($q^c + q_c^c$); the sum of grid-resolved and cumulus-scale cloud water. Similar to MC in case CR, low-level cloud grows to upper-levels. In contrast to convection in case NP, cloud water is continuously generated at low-levels, and the growth of MC is maintained for a longer time until the rainfall is considerably intensified. Such a behavior is simulated by the inclusion of the CC-effects. After the mature stage MCP decays by the same mechanism as in case CR; by the development of a cold pool. The lifetime of 5 hours in this case is comparable to that in case CR.

The vertical structure at the mature stage of MCP (Fig. 8c) is shown in Fig. 9. We can find main features shown in case CR. For example, the peak levels of cloud water, rainwater, warming and updraft are similar to those in case CR. A maximum of tangential velocity is also noted. A cold pool begins to form at low-levels by the evaporation of rainwater, though it is not as pronounced as in case CR.

Our next interest is how the introduced CC-effects and grid-resolved condensation contribute to the simulated features. Figure 10 shows the time-height sections of various quantities for MCP at 170 km radius. First, we will see the gross features shown in Fig. 10a–d. The updraft (Fig. 10a) is intensified from low-levels ($1\sim 3 \text{ km}$ height) after 27h10min. It takes about 3 hours to attain its maximum intensity in the upper troposphere. Downdraft is generated below the updraft at the mature stage (after 30h10min). Cloud water (total cloud water in Fig. 10c) is produced as the upward motion is intensified, and its peak is located at higher levels than the updraft peak. Rainwater (Fig. 10d) is produced by conversion of cloud water, and its evaporation induces a cold pool near the surface at the later stage (after 30h30min), as shown in the temperature deviation (Fig. 10b). All the four variables have double peaks. A peak at the earlier stage is owing to the CC-effects, and the other is to the grid-resolved, as will be shown later. (A pos-

Fig. 10. Time-height cross-sections of MCP ($r = 180$ km): (a) vertical velocity (isolines: every 0.2 m s^{-1} [$< 1.2 \text{ m s}^{-1}$], 0.5 m s^{-1} [$> 1.5 \text{ m s}^{-1}$]), (b) temperature deviation (every 0.5 K and $0.25, 0.75 \text{ K}$), (c) total cloud water (every 0.2 g kg^{-1} and 0.1 g kg^{-1}), (d) total rainwater (every 0.25 g kg^{-1} and 0.1 g kg^{-1}), (e) $\Delta T_c (= T^c - T)$ (isolines: every 0.2 K , and 0.1 K), (f) $q_v^c - q_c$ (every 0.2 g kg^{-1} [$< 1.0 \text{ g kg}^{-1}$], 0.5 g kg^{-1} [$> 1.0 \text{ g kg}^{-1}$] and 0.1 g kg^{-1}), (g) cumulus-scale heating (every 1), (h) cumulus-scale cloud water, (i) cumulus-scale rainwater, (j) grid-resolved condensational heating (isolines: every 0.2 [< 4], 2 [> 4]), (k) grid-resolved cloud water, (l) grid-resolved rainwater.

itive temperature anomaly around 28 h is caused by downward motion associated with a preexisting gravity wave.)

The processes at each stage of the life cycle are described as follows. At the initial stage the initiation of cloud water (at 27h30min) is due to the grid-resolved process. This is important because the low-level updraft makes relative humidity higher and causes grid-resolved saturation. Production of grid-resolved cloud water (Fig. 10k) due to grid-resolved condensation (Fig. 10j), which occurs below 2 km, is considered a typical feature in MC formation process. The condensation causes warming at low-levels, and further intensification of the low-level updraft. The cumulus-scale heating and moistening are inhibited until 28 h, though the temperature difference ($T_c - T$; hereafter, ΔT_c) (Fig. 10e) is positive (latently unstable). This is because updraft is still weak at this stage, and the occurrence of the CC-effects is prevented by the condition (c5).

After 28 h, when the condition (c5) is satisfied, production of cumulus-scale cloud water (Fig. 10h) and cumulus-scale heating (Fig. 10g) take place. These extend to higher levels, as expected from the ΔT_c profile. At 29 h, cumulus-scale heating has its peak at 4~5 km. (The top level exceeds 9 km.) Grid-scale-condensation (hereafter, grid-condensation) and cloud water also increase. The moistening for grid-condensation is due to the CC-effects. This can be understood from $(q_v^c - q_v)$ (Fig. 10f). A peak of grid-condensation is located at 3 km. (The top level is at 5~6 km.) As a result of these heating, an updraft continues to intensify, and the level of the maximum updraft becomes higher. At 29h20min, another peak appears at 9~10 km owing to the CC-effects. Thus, the evolution of updraft and distribution of cloud water are greatly affected by the cumulus-scale condensation and upward heat transport.

The effect of cumulus-scale heating is not so distinctly seen in the temperature deviation as in the updraft. This is partly because a certain degree of warming is already caused by subsidence, and partly because a considerable part of the cumulus-scale heating is compensated by the adiabatic cooling associated with the updraft.

Cumulus-scale heating reaches its maximum intensity at 29h30min. Although the peak level remains at altitudes of 4~5 km, the peak level of cumulus-scale cloud water becomes higher (10 km height) because cumulus-scale cloud water has been

Fig. 11. Time-height cross-sections of MCPO ($r = 350$ km): (a) vertical velocity (iso-lines: every 0.2 m s^{-1} [$< 1.2 \text{ m s}^{-1}$], 0.5 m s^{-1} [$> 1.5 \text{ m s}^{-1}$]), (b) temperature deviation (every 0.5 K and 1.8 K), (c) total cloud water, (d) total rainwater.

transported by the updraft. Cumulus-scale cloud water is converted into cumulus-scale rainwater (Fig. 10i), and it falls from the cloud. Downdraft due to CC takes place after about 28 h by the assumption described in subsection 2.2.3, and evaporation of the rainwater begins below the cloud base, though cooling is still weak before 30 h.

As MCP develops, the buoyancy ΔT_c diminishes. The reduction of ΔT_c begins from low-levels, and by 30h20min ΔT_c becomes negative at all levels. The stabilization is owing to the rise of the grid-resolved temperature T , and the decrease in T^c due to cooling in the boundary layer.

When ΔT_c decreases, cumulus-scale heating is reduced (the condition (c4)). Instead, grid-condensation considerably increases, and heating is entirely due to grid-condensation after 30h20min. The grid-condensation and updraft take their maxima

at 30h40min at a height of 7~8 km.

The increase of grid-condensation in the upper-level at the mature stage is one of the characteristics in the life cycle of MCP. Cloud water produced by grid-condensation is converted to grid-resolved rainwater (Fig. 10l) (a peak at 30h50min), and contributes to the intensification of the cold pool and downdraft through the evaporative cooling in the subcloud layer. The downdraft, which extends to higher levels, leads to the decay of MCP and formation of new MC.

MCP is well-developed MC and its lifetime is relatively long. However, weaker MC with a lifetime of 2~3 h also exists in case P (and in case CR). Figure 11 shows the time-height sections of MC located in the outermost convective area (350 km radius, 26h40min~31h40min). It is referred to as MCPO. In this area the relative humidity is lower than that in the case of MCP. The main features of the evolution is the same as those of MCP, but the updraft is less intense (1.2 m s^{-1} at maximum) with a lower peak level (at 7 km) and with a shorter time scale (about 3 h). Total rainwater has a single peak, different from double peaks in MCP. This is because MCPO grows mainly owing to the cumulus-scale heating, and the grid-condensation is considerably less than that in MCP. Thus, another peak at the later stage due to grid-condensation is not found. The CC-effects play a more critical role in the growth of MCPO.

4. Further experiments on the cumulus-scale effects

To further examine the impacts of the CC-effects on MC, numerical experiments are performed by terminating the CC-effects at a certain time just before the formation of MC. Two cases with different low-level humidity are presented. One is the case under a fairly humid condition (MCP), and the other is the case in which low-level relative humidity is well below saturation (MCPO). The behavior of MC significantly changes when the CC-effects are removed, and the impacts are quite different depending on the low-level relative humidity, as shown below.

The results from the experiment on MCP are shown in Fig. 12. The CC-effects are turned off at 27 h. Since the air is fairly moist (exceeding 90%) at low-levels, grid-condensation at low-levels causes too rapid intensification of updraft, and the time scale of convection becomes shorter than that in Fig. 10. The peak velocity is 4 m s^{-1} (at 10 km at

Fig. 12. Time-height cross-sections of the case in which cumulus-scale effects are terminated from the beginning of MCP ($r = 180 \text{ km}$).

29h50min), which is much stronger than in MCP. Another remarkable point is that gravity oscillation is notable in the vertical velocity and temperature fields. Corresponding to this, other peaks of cloud water appear. These may be rather unrealistic in a TC. The above-mentioned features are similar to those in case NP, and entirely different from those of MC in case CR. When CC is resolved, as in case CR, heat and moisture are transported upward by the cumulus-scale updrafts and not accumulated at low-levels. It is concluded that when the CC-effects are not included for the 10 km grid, too much heat is released at low levels and clouds grow very rapidly under the condition of very humid low-level air.

In the second case in which the low-level relative humidity is well below saturation ($< 80\%$) (MCPO), the impact of the cumulus-scale effects on MC significantly differs from that described

Fig. 13. Same as Fig. 15, but for MCPO ($r = 350$ km).

above. Figure 13 shows the result obtained when the CC-effects are terminated at 26h40min for MCPO. Compared with Fig. 11, it is seen that convection is localized in a thin layer. The cloud top does not reach higher than a 8 km height (12 km in MCPO). The total amount of heat is not large enough for the growth of MC.

These results suggest that the CC-effects greatly affect the behavior of MC through the heat release and vertical heat transport. In a humid situation the vertical heat transport by CC tends to inhibit the unrealistically rapid growth of convection. In contrast, in a less humid condition the heat release by CC enhances convection.

5. Discussion

Numerical experiments indicate that significant modifications from Kuo's formulation are required for the proposed MCRM to simulate MC and a TC realistically. Three important parameters, F_I , F_b and E_D , are introduced.

As mentioned in Section 2, the ratio F_I is assumed to be reduced as R_I decreases or the stratification approaches a neutral state. The aim of these treatments is to assure that the vertical distribution of moisture, heat release by CC, and grid-condensation should be continuous even when CC is terminated. In fact, when these conditions are not imposed, abrupt increase of grid-condensation occurs, particularly at low-levels, and discontinuity emerges in the distributions of heating, cloud water and rainwater (not shown). Smooth transition from convective heating to grid-resolved heating was also intended in previous Kuo type models (Krishnamurti et al. 1983; Molinari 1982).

The second modification is concerned with Kuo's b -parameter. Previous studies (Krishnamurti et al. 1983; Anthes 1977; Kuo and Anthes 1984) pointed out that Kuo's (1965) assumption for b often results in excessive moisture storage, and too weak heating. They proposed several modifications to produce stronger heating. However, these were intended for large-scale simulations, and not for the description of mesoscale systems like MC. (Mesoscale was rather treated as the subgrid-scale.) In the present study, a parameter F_b is introduced to make the ratio b smaller than that in Kuo (1965). The sensitivity experiments in which Kuo's (1965) quantification of b is used (corresponding to $F_b = 1$) show that heating is insufficient for the growth of MC when the low-level relative humidity is well below saturation. On the other hand, when it is nearly saturated, the dependence of MC on F_b is not so large; MC easily develops even with a large value of F_b .

Another improvement is concerned with the vertical distributions of heating and moistening. A turbulent detrainment and entrainment parameter E_D in the cloud model has a large impact on the heat distribution through the profiles of T^c and q_v^c . As is well known, small entrainment causes higher cloud top and peak level of heating, whereas large entrainment causes lower cloud top and heat release concentrated at low-levels. In the experiments without turbulent entrainment, heat distribution tends to have an upper-level peak, and the low-level heating is not strong enough for MC to form. On the contrary, when excessive entrainment is assumed, MC behaves as if no cumulus-scale effects were incorporated (as in case NP). The sensitivity of MC on these parameters, F_b and E_D , will be discussed in a separate paper.

As to the occurrence of CC, we introduce con-

dition (c5). Additional numerical experiments indicate that this condition is especially important when the moisture in the air column is well below saturation. In this situation $F_I I$ is stored as moisture, and a sufficient amount of latent instability is produced so that MC may be active. In other words, without condition (c5), the cumulus-scale transport of heat and moisture rather acts to reduce the humidity near the cloud base, and suppress the generation of MC.

With these modifications and by proper specification of the parameters, the MCRM that adopts Kuo's assumption as a representation of the CC-effects can reproduce realistic evolutions of MC and a TC. Since the framework of the present model is based on Y86, the main features and mechanisms of MC and a TC in case P, are qualitatively similar to those from Y86. To make further improvements on the MCRM, it is important to clarify why MC and a TC can be simulated in spite of the difference between Y86 and the present MCRM, though results obtained in this study can be only qualitatively compared with those from the three-dimensional model of Y86.

In Y86 the cumulus-scale heating was formulated in terms of cloud mass flux M^c , as done by Ooyama (1971) and Arakawa and Schubert (1974), but M^c was assumed to be proportional to w^* , the grid-resolved vertical velocity at the top of the boundary layer ($z = z^*$):

$$\frac{Q^c}{C_p \pi} = \frac{M^c}{\rho} \frac{\partial \theta}{\partial z} + D(\theta^c - \theta), \quad (29)$$

$$\frac{M^c}{\rho} = Y w^*, \quad (30)$$

where Y is the most important quantity that determines the properties of MC, and the second term on the right hand side of Eq.(29) represents the effect of detrainment from the implicit cloud, as in Arakawa and Schubert (1974).

Though the above assumption (except for detrainment) has a form similar to that used in the TC model of Ooyama (1964), Y86 argued that this assumption should be applied to MC in a TC, rather than to a TC itself. Since mesoscale organization of convection is considered to be closely related to the low-level convergence, this assumption seems to be reasonable to a fair degree. The quantity Y was written in the following form, using Y^* , the value of Y at a height of z^* :

$$Y = Y^* \eta, \quad (31)$$

where η is the normalized cloud mass flux ($\eta = 1$ at z^*), and Y^* represents the ratio of the cloud mass flux M^c to the grid-scale (mesoscale) mass flux ρw at z^* . One of the important features of the MCRM of Y86 is that Y^* is assumed to depend on the degree of buoyancy that a rising air possesses in a layer above the cloud base (at 2~4 km altitudes), and it is taken to be larger than unity for positive buoyancy in that layer. Another important feature is that η is assumed to be of the form given by Eq.(20).

These assumptions were based on the results from a CCRM (Yamasaki 1977, 1983) and those from the linear theory (Syōno and Yamasaki 1966; Yamasaki 1969). The theory, which was originally applied to a TC, was later recognized as useful to understand the properties of mesoscale motion in a TC. The properties of mesoscale motion such as MC are determined primarily by a non-dimensional quantity Y in a context of the linear theory. If Y is large in the upper troposphere and small in the lower (particularly around and just above z^*), unrealistic gravity waves amplify. If Y is too large in the lower layer, a mode similar to CIFK (conditional instability of the first kind) becomes dominant. When the vertical profile of Y is within a certain range, mesoscale motion occurs with a reasonable growth rate similar to that of MC. Needless to say, warming at low-levels is one of the necessary conditions for the growth of MC. Low-level latent instability (positive buoyancy) was considered an important condition for low-level warming in Y86; it is the values of Y^* and η at low-levels that control the growth of MC. On the other hand, the values of η in the upper layer are important to the vertical extent of MC, and its structure. These suggestions implied by the linear theory have been confirmed by numerical experiments with non-linear models (Yamasaki 1986, 1987, 1988, 1989).

In the present model, the cumulus-scale heating is given in the form of Eq.(15), and the heat amount around the cloud base z^* depends on the moisture convergence I , the value of F_b , vertical profile of T^c and so on. With such a formulation cooling may occur around the cloud base level under some conditions because the adiabatic cooling may exceed the heating. Under other conditions, particularly when low-level air is nearly saturated, the heating might be strong enough to cause a CIFK mode. Thus, the amount of low-level heating that is important for the growth of MC is not so systematically controlled as in Y86, in which the heating is

closely related to low-level latent instability (positive buoyancy).

Nevertheless, in case P, the amount of low-level heating seems to be fairly appropriate in most situations. The formation and growth of MC are rather similar to those in case CR. This is mainly because the value of E_D is selected so that the cloud top may be lower at early stages and a considerable amount of heat may be released at low-levels, and because F_b is specified to avoid too much moisture storage. In this way, the amount of low-level heating can be controlled by the specification of E_D and F_b to some extent.

The vertical velocity w^* at the cloud base, which strongly controls the amount of heating in Y86, roughly corresponds to moisture convergence I in the present MCRM (referred to as NY). This is because moisture supplied from the boundary layer top generally takes a large part of I . As for the vertical profile of heating, it depends on η in Y86 and on $(T^c - T^*)$ in NY. In Y86 the first term on the right hand side of Eq.(20) corresponds to the amount of heating that nearly cancels out the adiabatic cooling, and the net warming is nearly proportional to the second term, which is related to buoyancy $(T^c - T)$. In NY the use of T^* instead of T in Eqs.(13) and (15), qualitatively implies that the heating which compensates the adiabatic cooling is taken into account. Since the effects of E_D on η and on $(T^c - T)$ are qualitatively the same in both models, its influence on the heat profile is expected also to be similar. The most distinct difference between Y86 and NY may be the assumption for Y^* , that relates the heat amount (particularly around the cloud base) to the low-level buoyancy in Y86, as mentioned above.

Another significant difference between Y86 and NY is the sensitivity to the relative humidity. Relative humidity directly modifies the amount of moisture convergence I in NY. Not only relative humidity in the boundary layer, but also that in the free atmosphere (especially, at low-levels) affects the heat amount to a considerable extent. In Y86, though high relative humidity in the boundary layer induces strong heating through larger values of Y^* and η , relative humidity in the free atmosphere affects the heating in a different way; it has an influence on T^c and q_{tc}^c through entrainment of the environmental air to the implicit cloud. It is expected that the sensitivity of the heating to relative humidity is less than that in NY.

In both models, cumulus-scale cloud water and

rainwater are produced from cumulus-scale condensation, and the evaporation of cumulus-scale rainwater produces a cold pool. The importance of the evaporation of cumulus-scale rainwater to simulation of MC, which has been stressed by Y86, is again emphasized by the present study. This is confirmed by additional experiments in which cumulus-scale heating is included, but cumulus-scale cloud water and rainwater are not treated. In such experiments the time scale of MC is not clear, and heating causes rapid intensification of frictional inflow, which results in too early formation of an eyewall. In connection with the evaporation of rainwater, the effect of cumulus-scale downdraft should be remarked. The sensitivity experiments in which the effect of cumulus-scale downdraft is terminated after a certain time in case P, show that MC is not successively generated in an intermediate area (at radii of 150~200 km) because of weak cold pools.

The cold pools in case P are generally weaker than those in Y86 and in the CCRM. The evaporation of rainwater is relatively suppressed until the grid-resolved downdraft is generated. This is partly because Kuo's assumption generally moistens the low-level air, and tends to cause grid-resolved saturation. For a more enhanced evaporation of cumulus-scale rainwater, an improvement in the formulation of cumulus-scale downdraft may be also required.

6. Concluding remarks

A new mesoscale-convection-resolving model (MCRM) is proposed in this study. The model differs from Y86 in that it adopts Kuo's (1965) formulation as a representation of the cumulus-scale effects (CC-effects), with significant modifications. The results from a numerical experiment using a horizontal grid size of 10 km (case P) are discussed, compared with those from a cumulus-scale-resolving model (CCRM) with a 1 km grid (case CR), and those from a 10 km grid model without the CC-effects (case NP). It is shown that the use of Kuo's formulation can reproduce realistic evolution of a TC and MC, provided that we make several modifications and choose proper values for model parameters.

One of the modifications from Kuo's formulation is partly intended to avoid abrupt changes of the cumulus-scale and grid-resolved condensational heating. It is assumed that the ratio F_I (the ratio of the moisture available for CC to converged mois-

ture I) depends on the ratio of I_L (moisture converged at low-levels) to I and the cloud-layer mean value of the temperature difference ($T^c - T$); the buoyancy of rising air. In previous studies which employed Kuo's scheme in the 'large-scale' models, efforts have been made to avoid abrupt changes by taking account of the relative humidity as well as the adiabatic effect. In the present model, we consider the above two quantities important. The introduction of the ratio of I_L to I is based on the importance of low-level convergence for MC in a TC.

In some of the previous studies that used Kuo's (1965) assumption the effect of entrainment has been taken into account. However, the importance of *low-level* heating, which was emphasized by Y86, has not been necessarily noticed. In the present study, by assuming proper turbulent entrainment and detrainment rate E_D , sufficient heat can be released at *low-levels* at the formative and growing stages of MC. Sensitivity of MC to E_D is remarkable. Small entrainment, which seems to be favorable to deep convection, often leads to too weak low-level heating, which does not give rise to the growth of MC. On the other hand, too large entrainment tends to induce too rapid growth of shallow convection (similar to the conditional instability of the first kind, CIFK). An introduction of a parameter F_b (related to Kuo's b parameter) also makes an important contribution to realistic simulation of MC and a TC. In addition, we impose a condition that prevents the occurrence of CC according to the intensity of the updraft at the cloud base. This condition contributes to produce latent instability which is large enough for MC to form successively.

The impacts of the CC-effects are examined by comparison of case P with case NP and by additional experiments in which the CC-effects are terminated at a certain time in case P. Without the CC-effects, the behavior of convection is considerably different from that in case CR and case P. It greatly depends upon low-level relative humidity. When low-level air is well below saturation, grid-resolved saturation and convection less easily occur in a coarse grid model. Convection has weak intensity and decays with a short lifetime (about 1h30min). In this situation, heating by CC often plays an important role in the growth of MC. On the other hand, when low-level air is very humid, once some amount of grid-condensation takes place, condensational heating is concentrated to a thin

layer, and unrealistically intense vertical circulation develops around the condensation level. Strong low-level heating induces rapid growth of excessively vigorous MC. In this situation inclusion of the vertical transports of heat and moisture by CC prevents the concentration of low-level heating, and suppresses the unrealistic growth of MC.

A prognostic treatment for cumulus-scale cloud water q_c^c is considered an improvement from Y86 in which q_c^c was diagnostically determined. This treatment enables us to discuss not only the distribution of grid-resolved cloud water, but also that of total cloud water. By this modification, the evaporation of q_c^c which remains after CC is terminated (MC becomes non-convective) can also be treated. Evaporative cooling due to this portion of cloud water contributes to the growth of another MC which exists in its vicinity. It is obviously desirable for a more proper simulation of MC and water budget though the impact of the prognostic treatment is not demonstrated in this paper.

In this paper, we are concerned with the early stages of TCs when the effect of surface friction is not very significant. As the wind velocity increases and the frictional forcing becomes dominant, the impact of the CC-effects on MC is not so large as that in the earlier stages, and MC more easily forms because low-level moisture convergence is also maintained by the dynamical forcing. When an eyewall is formed, the organization of convection by frictional convergence can be qualitatively represented even without the CC-effects. Since MC under weak forcing is more difficult to be simulated than those under strong forcing, it is important for a TC model to simulate MC realistically, particularly at the stage of weak wind velocity. This is why we intensively discuss MC at the early stage of the TC in this paper. The CC-effects on the eyewall will be reported in a separate paper.

In this study, discussions are confined to the two-dimensional model. The next step of this study is to discuss the problem by use of a three-dimensional model. For this purpose, the formulation of the CC-effects presented in this paper has been incorporated in the three-dimensional MCRM of Y86; in other words, the cloud mass flux type formulation of the CC-effects was replaced by the Kuo type. Numerical experiments and comparison between the results from the two MCRMs have been made (Nasuno and Yamasaki 2000⁵). Comparisons

5 presented at the 24th Conference on Hurricanes and Tropical Meteorology

with other models such as 2~5 km grid models⁶ without the CC-effects (Liu et al. 1997, 1999; Yau et al. 1999⁷) are also needed to clarify to what extent MC-like convection is simulated in these models, and to get insight into how the CC-effects should be incorporated to describe MC properly. Studies with 3D CCRMs having a grid size of about 1 km or less⁸ are also important to improve MCRMs through better understanding of the mesoscale features of TCs.

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