A Tropical Cyclone Model Resolving Mesoscale Organized Convection with Prognostic Treatment of Subgrid-scale Cloud Water

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Abstract

A new version of a tropical cyclone model is presented as an extension of Yamasaki's (1986) model that intended to resolve mesoscale organized convection by a grid and treat the cumulus-scale motion as the subgrid-scale. Two important improvements are made in this study. One is a prognostic treatment of cumulus-scale cloud water, which was treated diagnostically in the old version. The other is that the ratio of the cumulus-scale ascending area is not assumed to be sufficiently small compared to unity.

Results from numerical experiments indicate that the features of tropical cyclones (including rainwater distribution) simulated by the new and old versions are essentially similar, but the cloud water distribution is simulated more realistically by the new version, as evident from the model formulation. Clarification of the impact of the finite area ratio and an improvement of its assumption remain to be studied.

1. Introduction

This paper presents an improved version of a tropical cyclone model of Yamasaki (1986, 1987), which can be called mesoscale-convection-resolving model in that mesoscale organized convection is resolved by a grid (with a grid size of 5~20 km) and cumulus convection (basic mode of moist convection) is implicitly treated as the subgrid-scale (or parameterized). The most important concept on which Yamasaki's (1986) model is based is that cumulus convection (time scale: a few tens of minutes) is usually organized into mesoscale convection with a time scale of a few hours (up to more than 10 hours under some conditions) and mesoscale convection is one of physically significant modes of convection. Observationally, these have been known since the 1950s; mesoscale organized convection constitutes rainbands in tropical cyclones; in other words, each rainband is composed of many

mesoscale organized clouds (known as mesoscale echo cells in radar observations). However, the significance of mesoscale convection as one of the important modes of convection has not been necessarily widely recognized, particularly by researchers of numerical modeling. This recognition was made in a study of Yamasaki (1983) with the use of a non-hydrostatic cumulus-scale-resolving model, which led to the development of a mesoscale-convection-resolving model of Yamasaki (1986) (referred to as Y86 hereafter).

Tropical cyclone models with the hydrostatic approximation before Y86 were such that convective clouds were not resolved by the grid but their effects were incorporated by way of parameterization (Ooyama 1964, 1969; Yamasaki 1968; Rosenthal 1970; Anthes 1972; Kurihara and Tuleya 1974; Wada 1979; and many others). Although many aspects of the essential features of tropical cyclones (TCs) were understood by these models, some important features and mechanisms of TCs (particularly mesoscale features) were not appropriately treated or understood. On the other hand, studies with nonhydrostatic models that could resolve cumulus convection were also performed (Yamasaki 1977, 1983; Willoughby et al. 1984; Lord et al.

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1984; Nasuno and Yamasaki 1997 and others). Such studies with cumulus-scale-resolving models were very useful to understand some important mechanisms of TCs, but were limited to axisymmetric TCs mainly because of computational resources. Even today, the use of a three-dimensional model should be restricted to TCs under highly idealized conditions because a very small grid size such as 1 km or less has to be used. A merit of a mesoscale-convection-resolving model (abbreviated to MCRM) is that it can have a wide application (including formation and movement of a TC interacting with its environment) because a coarse grid size such as 5~20 km can be used. Instead, we have to overcome a difficulty of appropriately incorporating the cumulus-scale effect.

In the MCRM of Y86 the cumulus-scale effect was incorporated based on the results from Yamasaki (1983). One of the important points was inclusion of cooling due to rainwater evaporation (based on the importance of a mechanism similar to the cold pool mechanism typically exemplified in a squall cluster). For this purpose, not only the mixing ratio of rainwater for the grid-resolved but also that for the subgrid-scale (cumulus-scale) were included as prognostic variables. In this connection, the effect of cumulus-scale downdraft was also taken into account. These were important to simulate the decay of mesoscale convection (abbreviated to MC) and formation of new MC or successive formation of MC, particularly in the situation of weak frictional inflow.

The second important point was concerned with the growing stage of MC and its vertical structure. The condition for the growth of MC and its growth rate and structure should be given primarily by the vertical profile of the cumulus-scale heating. Linear theory for the TC (Syono and Yamasaki 1966; Yamasaki 1968) was helpful to this. The linear theory can be applied to MC in a TC rather than a TC itself. The heating profile in Y86 was formulated so that MC might behave realistically throughout its life cycle.

The thermodynamic equations used in Y86 were similar to those by Ooyama (1971) and Arakawa and Schubert (1974) in that the cloud mass flux M^C was used. However, assumptions for M^C were very different. Since it was the cumulus-scale effect (not the effect of the whole cloud) that was parameterized in Y86, the cloud mass flux M^C means cumulus-scale mass flux, which was determined based on the arguments above. Its detail is

described in Section 2.

The MCRM of Y86 has been used for several problems such as TC development and the mesoscale structure of the TC (Y86; Yamasaki 1987), TC formation associated with the intertropical convergence zone (Yamasaki 1988, 1989), interaction of a TC with an upper tropospheric cold low (Yamasaki and Hara 1990, Yamasaki 1996) and TC motion prediction with observed data (Yamasaki 1992). The performance of the model predicting the motion of several typhoons in 1990 was better than those of some models including pre-existing parameterizations (Ueno, 2000). One of the results obtained from applications to typhoon Flo (T9019) was also reported at a workshop of the Compare III in 1999. The model has also been applied to some phenomena other than TCs such as extratropical cyclones (Nodaguchi 1997), cloud clusters associated with a baiu front (Isoya 2000) and equatorial super cloud clusters (Yamasaki 1994 UCLA-CCSR Joint Workshop).

It is obvious that the model has to be improved by comparing model results with observations. Despite of efforts to find model shortcomings, it seems that observational data have not been necessarily available to an extent that unrealistic features of the model behavior are clarified. There have been no suggestions that indicate unrealistic features. On the other hand, as was recognized in Y86, the model itself has two obvious shortcomings (although these were not essential compared with improved aspects of the model). One is an assumption that the ratio of the ascending area with the cloud mass flux is very small compared to 1. Since the cloud mass flux in the MCRM refers to the cumulus-scale, this assumption is not appropriate in a quantitative sense. Any assumption in determining the ratio introduces another uncertainty. For this reason, an assumption of a small ratio, which had been used widely in cloud parameterizations for the large-scale models, was adopted in the MCRM of Y86 as a crude approximation. However, in order to improve the MCRM, we have to avoid this assumption. In the present study, the ratio is given by a simple assumption, which should be improved in future studies.

The second shortcoming is a diagnostic treatment of subgrid-scale (cumulus-scale) cloud water. The temperature, water vapor and cloud water inside an area of the cumulus-scale updraft were determined by the saturation condition and budgets of energy and water substance in a way similar to

that by Arakawa and Schubert (1974). In such a model subgrid-scale cloud water is a diagnostic variable. Therefore, cloud water that remains after the cumulus-scale motion ceases and its evaporation are not taken into account, and thereby cloud water distribution associated with the subgrid-scale is not simulated properly. In the present model subgrid-scale cloud water is treated with the use of a prognostic equation. This improvement was made recently in a MCRM of Nasuno and Yamasaki (2000) which applied Kuo's (1965) formulation to parameterization of the cumulus-scale, with several improvements. By treating subgrid-scale cloud water in a prognostic way, cloud water distribution can be better simulated than in the MCRM of Y86.

It should be remarked that the scale separation between cumulus convection and MC is not very clear in many situations, as mentioned in Y86. Nevertheless, it is important to intend to resolve MC by the grid. This is particularly important to the problem of TC formation in which dynamical forcing such as frictionally controlled flow is not very strong. The horizontal grid size for an appropriate simulation of MC may be less than 5 km. Truncation errors should be large for a 10~20 km grid. However, a 5~20 km grid MCRM should be useful until computational resources become sufficiently available.

In Section 2 an improved version of the MCRM is presented. Although the aim of this paper is model description, an example of numerical experiments is presented simply to show the mesoscale structure of a TC at the developing stage, with a focus on the cloud water distribution which was not described well by the MCRM of Y86.

2. Model description

2.1 Equations with finite area ratio of cumulus-scale ascending motion

We consider the thermodynamic equation for potential temperature θ .

$$\frac{d\theta}{dt} = \frac{Q}{c_p \pi}, \quad (1)$$

where Q is the non-adiabatic heating per unit mass and unit time, π is Exner function and c_p specific heat at constant pressure. The Exner function is defined by $\pi = (p/p_0)^{R/c_p}$, where R is the gas constant for dry air, p is pressure, and p_0 is a reference pressure. In order to derive an equation for the grid-resolved quantity $\bar{\theta}$, each dependent variable is divided into two parts; the grid-resolved

field (describing MC explicitly) and the deviation from it (contribution of the cumulus-scale treated implicitly).

$$\left. \begin{aligned} \theta &= \bar{\theta} + \theta' \\ w &= \bar{w} + w' \\ V &= \bar{V} + V' \end{aligned} \right\}, \quad (2)$$

where w is the vertical velocity, V is the horizontal wind vector, bar indicates the grid-resolved field, and dash indicates the deviation. Substituting (2) into (1) and taking average, the equation for $\bar{\theta}$ may be written as

$$\frac{d\bar{\theta}}{dt} = \frac{\bar{Q}}{c_p \pi} - \frac{\partial(\rho \bar{w}' \bar{\theta}')}{\rho \partial z}, \quad (3)$$

where

$$\frac{d\bar{\theta}}{dt} = \frac{\partial \bar{\theta}}{\partial t} + \bar{V} \cdot \nabla \bar{\theta} + \bar{w} \frac{\partial \bar{\theta}}{\partial z},$$

and ρ is the density. When (3) is derived, a contribution of $\bar{V}' \bar{\theta}'$ is neglected because it is small. In order to express the second term on the right hand side of (3) in a more tractable form, we consider two areas of upward and downward motions associated with the cumulus-scale. Denoting the representative values of the vertical velocity and potential temperature in the ascending area by w^c and θ^c , and those in the descending area by w^e and θ^e , respectively, $\bar{w}$ and $\bar{\theta}$ can be written as

$$\begin{aligned} \bar{w} &= \sigma w^c + (1 - \sigma) w^e \\ \bar{\theta} &= \sigma \theta^c + (1 - \sigma) \theta^e, \end{aligned} \quad (4)$$

where σ is the area ratio of the ascending motion of the cumulus-scale. It should be remarked that the term downward motion associated with the cumulus-scale does not necessarily mean that the air motion itself is downward ($w^e < 0$). When $\bar{w}$ is positive, w^e may be also positive in some situations. The term means the negative deviation from the grid-resolved motion as a result of the cumulus-scale motion.

It is difficult to determine the area ratio σ realistically. In the present model σ is introduced as a quantity which may take a range from 0 to 0.4. Using w^c and θ^c , the quantity $\bar{w}' \bar{\theta}'$ can be expressed as

$$\begin{aligned} \bar{w}' \bar{\theta}' &= \sigma(w^c - \bar{w})(\theta^c - \bar{\theta}) \\ &\quad + (1 - \sigma)(w^e - \bar{w})(\theta^e - \bar{\theta}) \\ &= \frac{\sigma}{1 - \sigma}(w^c - \bar{w})(\theta^c - \bar{\theta}). \end{aligned} \quad (5)$$

In Y86, as well as models that treat the large-scale motion, it is assumed that $\sigma \ll 1$ and $w^c \gg \bar{w}$. In this case $\overline{w'\theta'}$ can be approximated as

$$\overline{w'\theta'} = \sigma w^c (\theta^c - \bar{\theta}).$$

However, this assumption is not used in the present model.

Equation (3) may be rewritten, using (5), as

$$\frac{d\bar{\theta}}{dt} = \frac{\bar{Q}}{c_p \pi} - \frac{1}{1-\sigma} \frac{1}{\rho} \frac{\partial}{\partial z} [(M^C - \rho \sigma \bar{w})(\theta^c - \bar{\theta})], \quad (6)$$

where

$$M^C = \rho \sigma w^c, \quad (7)$$

is the mass flux associated with upward motion of the cumulus-scale. It is referred to as cumulus-scale mass flux. (In contrast, $\rho \bar{w}$ can be called mesoscale mass flux.

We consider the effect of condensation C^C in ascending area of the cumulus-scale. (A possible contribution of w^e , which depends on the relative magnitude of $\bar{w}$ and σw^c , is ignored. This problem remains to be treated in the future.) Replacing $\bar{Q}$ in (6) by C^C , (6) may be written as

$$\begin{aligned} \frac{d\bar{\theta}}{dt} = & \frac{L}{c_p \pi} C^C \\ & - \frac{1}{1-\sigma} \frac{1}{\rho} \frac{\partial}{\partial z} [(M^C - \rho \sigma \bar{w})(\theta^c - \bar{\theta})], \end{aligned} \quad (8)$$

where L is the latent heat of condensation. Similarly, equations for the mixing ratios of water vapor $\bar{q}_v$ and cloud water $\bar{q}_c$ are as follows:

$$\begin{aligned} \frac{d\bar{q}_v}{dt} = & -C^C \\ & - \frac{1}{1-\sigma} \frac{1}{\rho} \frac{\partial}{\partial z} [(M^C - \rho \sigma \bar{w})(q_v^c - \bar{q}_v)], \end{aligned} \quad (9)$$

$$\begin{aligned} \frac{d\bar{q}_c}{dt} = & C^C \\ & - \frac{1}{1-\sigma} \frac{1}{\rho} \frac{\partial}{\partial z} [(M^C - \rho \sigma \bar{w})(q_c^c - \bar{q}_c)]. \end{aligned} \quad (10)$$

The grid-resolved quantities $\bar{q}_v$ and $\bar{q}_c$ are given by

$$\begin{aligned} \bar{q}_v &= \sigma q_v^c + (1-\sigma)q_v^e, \\ \bar{q}_c &= \sigma q_c^c + (1-\sigma)q_c^e, \end{aligned} \quad (11)$$

where q_v^c and q_c^c are the mixing ratios of water vapor and cloud water, respectively in the ascending area of the cumulus-scale, and q_v^e and q_c^e are those in the descending area. The second terms on the right hand side of (8)~(10) represent the effects

of the vertical transports of heat, water vapor and cloud water associated with the cumulus-scale motion. It is one of features of the present model to use the equations that express condensation of water vapor and vertical transports of heat, water vapor and cloud water explicitly, though these are qualitative. These equations (8)~(10) are formally the same as those known many years ago, but its physical meaning is different. (The effects of evaporation, the grid-resolved condensation and rainwater will be added later.)

2.2 Properties of the cumulus-scale ascending area

The next problem is how to determine the thermodynamic properties in the ascending area of the cumulus-scale, θ^c , q_v^c and q_c^c . As for q_v^c , it is assumed to be equal to the saturation mixing ratio of water vapor corresponding to θ^c . That is,

$$q_v^c = q_s(T^c, p), \quad (12)$$

where q_s is the saturation mixing ratio for temperature T^c and pressure p . In order to determine θ^c and q_v^c we use the following two conditions in addition to (12).

$$\theta^c = T^c / \pi, \quad (13)$$

$$\frac{\partial(M^C \theta_e^c)}{\partial z} - \tilde{\theta}_e \frac{\partial M^C}{\partial z} = 0. \quad (14)$$

The condition (14) is derived from the budget consideration for equivalent potential temperature for the ascending area of the cumulus-scale. The quantity $\tilde{\theta}_e$ in (14) is chosen to be θ_e^c or θ_e^e according to detrainment or entrainment.

$$\tilde{\theta}_e = \begin{cases} \theta_e^c & (\partial M^C / \partial z < 0 \text{ for detrainment}) \\ \theta_e^e & (\partial M^C / \partial z > 0 \text{ for entrainment}) \end{cases}$$

where θ_e^c and θ_e^e are equivalent potential temperature in the ascending and descending areas of the cumulus-scale, respectively. In the case of saturation, the equivalent potential temperature θ_e is usually given by

$$\theta_e = \theta \exp \left(\frac{L q_s}{c_p T} \right). \quad (15a)$$

We can use an approximate form

$$\theta_e = \theta \left(1 + \frac{L q_s}{c_p T} + \frac{L^2 q_s^2}{2 c_p^2 T^2} + \cdots \right). \quad (15b)$$

However, the accuracies of the approximations (15a) and (15b) are not good. In the present model we

approximate θ_e by the following form:

$$\theta_e = \theta + F^* \left(\frac{Lq_v}{c_p \pi_s} + \frac{L^2 q_{vB}^2}{2c_p^2 \pi_B T_B} \right), \quad (15c)$$

where π_B , T_B and q_{vB} are the values of the basic state, π_s is the value of π at the lifting condensation level in the case of unsaturation and F^* is a correction quantity, which is taken to be 1.1. This correction gives more accurate conservation quantity than θ_e given by (15a). It should be remarked that moist static energy gives much less accuracy than θ_e given by (15a), particularly in the case of low pressure in tropical cyclones and therefore another form was used in Y86. In the present study we compared several quantities and found that (15c) was the best conservative quantity among them. Some quantities of the basic state such as q_{vB} are used in (15c). We also use π instead of π_s even in the unsaturated subcloud layer. These approximations are used to determine the saturation state with much less computer time without any significant loss of accuracy. The four conditions (12), (13), (14) and (15c) determine T^c , θ^c , θ_e^c and q_v^c when M^C , θ_e and p are known.

As for q_c^c , the mixing ratio of cloud water in the ascending area, it is assumed that q_c^c can be determined by the following equation:

$$\frac{\partial q_c^c}{\partial t} = \frac{C^C}{\sigma} - \frac{1}{\rho\sigma} \left[\frac{\partial(M^C q_c^c)}{\partial z} - \tilde{q}_c \frac{\partial M^C}{\partial z} \right], \quad (16)$$

where

$$\tilde{q}_c = \begin{cases} q_c^c & (\text{for detrainment}) \\ q_c^e & (\text{for entrainment}) \end{cases}.$$

Equation (16) expresses the time change of q_c^c due to the production of cloud water owing to condensation in the cumulus-scale ascending area, the vertical flux of cloud water and entrainment from the environment or detrainment. The horizontal advection term due to the grid-resolved motion will be added later. It should be remarked that q_c^c was determined in a diagnostic way by use of budgets of water vapor and cloud water in Y86 (by assuming that the time change of (16) is 0).

The next problem is how to determine C^C . This is determined from the budget of water vapor.

$$C^C = -\frac{1}{\rho} \left[\frac{\partial(M^C q_v^c)}{\partial z} - \tilde{q}_v \frac{\partial M^C}{\partial z} \right], \quad (17)$$

where

$$\tilde{q}_v = \begin{cases} q_v^c & (\text{for detrainment}) \\ q_v^e & (\text{for entrainment}) \end{cases}$$

Equation (17) is obtained from an idea similar to (16), but with an assumption of a steady state for q_v^c . This assumption corresponds to the diagnostic determination of q_v^c , as mentioned before.

Equation (16) and (17) can be rewritten as follows:

$$\frac{\partial q_c^c}{\partial t} = \frac{1}{\sigma} \left[C^C - \frac{M^C}{\rho} \frac{\partial q_c^c}{\partial z} - (q_c^c - q_c^e) E \right], \quad (16a)$$

$$C^C = -\frac{M^C}{\rho} \frac{\partial q_v^c}{\partial z} - (q_v^c - q_v^e) E, \quad (17a)$$

where E indicates entrainment, which consists of dynamic entrainment E_{dyn} and turbulent entrainment E_T .

$$E = E_{dyn} + E_T \quad (18)$$

$$E_{dyn} = \begin{cases} \frac{1}{\rho} \frac{\partial M^C}{\partial z} & (\partial M^C / \partial z > 0) \\ 0 & (\partial M^C / \partial z \leq 0) \end{cases} \quad (18a)$$

$$E_T = E_D \frac{M^C}{\rho}. \quad (18b)$$

The coefficient E_D is a parameter concerning turbulent entrainment which is assumed to be proportional to the cumulus-scale mass flux. It may take values ranging from 0.5 to $2.0 \times 10^{-4} \text{ m}^{-1}$. It is very important how to assume the value of this parameter. The assumption on E_T is the same as that in Y86. (The definition of E_T in Y86 is different from (18b) by a factor of the thickness of the layer in the numerical model.) Equation (14) that expresses the budget of equivalent potential temperature is modified to include turbulent entrainment and detrainment. It is assumed that the magnitude of turbulent detrainment is equal to that of turbulent entrainment. Equation (14) may be written as

$$\begin{aligned} \frac{\partial(M^C \theta_e^c)}{\partial z} - \tilde{\theta}_e \frac{\partial M^C}{\partial z} \\ + E_D M^C (\theta_e^c - \theta_e^e) = 0. \end{aligned} \quad (14a)$$

As is well known, the entrainment acts to make C^C and q_c^c smaller because $q_v^c > q_v^e$ and $q_c^c > q_c^e$ in (16a) and (17a). When the entrainment is absent ($E = 0$), (16a) and (17a) can be written using w^c

as

$$\frac{\partial q_c^c}{\partial t} = \frac{C^C}{\sigma} - w^c \frac{\partial q_c^c}{\partial z}, \quad (16b)$$

$$C^C = -\sigma w^c \frac{\partial q_v^c}{\partial z}. \quad (17b)$$

The expression (17b) for C^C is similar to a well-known form for condensation rate of the air rising moist adiabatically. The time change of q_c^c is due to condensation and vertical advection of cloud water.

Needless to say, how to determine C^C is very important to realistic simulation of MC and a TC. The assumption (17a) includes $\partial q_v^c / \partial z$, which is an ambiguous quantity around the cloud base level z^* . Therefore, it is assumed that C^C is determined from $\partial \bar{\theta} / \partial z$ as follows:

$$\frac{L}{c_p \pi} C^C = \frac{M^C}{\rho} \frac{\partial \bar{\theta}}{\partial z} - \frac{L}{c_p \pi} (q_v^c - q_c^c) E. \quad (19)$$

Since the second term on the right hand side of (8) is generally one order smaller than the first term, the use of (19) means that we use essentially the same assumption as that in Y86. Above the level z^* , (17a) should be used, but numerical experiments indicate that an application of (19) to all cloud layers gives a better result. Therefore, (19) is used in a numerical experiment in Section 3. When $-(L/c_p \pi)(\partial q_v^c / \partial z)$ is equal to $\partial \bar{\theta} / \partial z$, the two assumptions are exactly the same. The reason why the slight difference between the two quantities causes somewhat significant difference for MC has not been understood yet. This problem remains to be studied.

As evident from (11), the value of q_c^c must satisfy the following condition:

$$\sigma q_c^c \leq \bar{q}_c.$$

Therefore, the area ratio σ should not be larger than a ratio of $\bar{q}_c$ to q_c^c .

$$\sigma \leq \bar{q}_c / q_c^c. \quad (20)$$

In this study, σ is tentatively assumed to be as follows:

$$\sigma = \begin{cases} \sigma_0 & (\bar{q}_c / q_c^c > \sigma_0) \\ \bar{q}_c / q_c^c & (\bar{q}_c / q_c^c \leq \sigma_0) \end{cases} \quad (20a)$$

where σ_0 is an upper limit of σ , which may take a value ranging from 0.2 to 0.4. In a numerical experiment presented in this paper, the value of σ_0 is chosen as 0.4. When the value of σ_0 is smaller than

0.2, the vertical velocity associated with MC is very strong, its horizontal scale is small, and more MC forms. Since (20a) is not necessarily reasonable, a more appropriate assumption should be used in future studies.

2.3 Assumption on the cumulus-scale mass flux

The cumulus-scale mass flux M^C is assumed to be essentially the same as that in Y86. Since the assumption on the mass flux is very important, it is summarized below.

$$\frac{M^C}{\rho} = \begin{cases} Y(\bar{w}^* - w_0) & (\bar{w}^* > w_0) \\ 0 & (\bar{w}^* \leq w_0) \end{cases}. \quad (21)$$

It was assumed in Y86 that the cloud mass flux was proportional to $\bar{w}^*$, the vertical velocity at a certain level z^* (around the top of the boundary layer), but a modification was made since Yamasaki (1988) to include w_0 , which is a positive parameter ($0.5 \sim 1 \text{ cm s}^{-1}$). That is, the vertical velocity should be intense to some extent for the occurrence of the cumulus-scale. This modification was based on the idea that the occurrence of the cumulus-scale is usually associated with the formation of the stratiform low-level clouds. Without this modification, the cumulus-scale occurs too frequently. A similar assumption was also made by Nasuno and Yamasaki (2000).

A proportional coefficient Y in (21) is an essential quantity that determines the properties of MC, as expected from linear theory mentioned in Section 1. When Y is written in the following form

$$Y = Y^* \eta, \quad (22)$$

η is the normalized cumulus-scale mass flux which is taken to be unity at z^* , and Y^* is a value of Y at z^* . The physical meaning of Y^* is that it is the ratio of the cumulus-scale mass flux to the mesoscale mass flux when $w_0 = 0$. As for η , it is assumed to depend on B as

$$\eta = 1 + C_\eta B, \quad (23)$$

where B is a measure of buoyancy of the rising air parcel (latent instability), which is given by

$$B = T^c(1 + 0.608q_v^c - q_c^c) - \bar{T}(1 + 0.608\bar{q}_v - \bar{q}_c). \quad (24)$$

The value of a parameter C_η in (23) is assumed to be 0.2 for positive values of B and 1.0 for negative values. The latter allows an overshooting of convection into a layer where $-1 < B < 0$. By

definition, C_η is 0 at z^* . According to numerical experiments, model results are not so sensitive to the value of C_η in (23) as expected, provided that it takes 0.2~0.3 for positive B . Large values of C_η cause too rapid growth of MC and somewhat unrealistic behavior of spiral rainbands.

As emphasized in Y86, the value of Y^* contributes strongly to temperature change at z^* and is closely related to the growth of MC. (The values of η just above the level z^* are also very important.) The growth of MC should be sensitive to the distribution of B in a layer just above the level z^* . Therefore, Y^* is assumed to depend on η^{**} in the following form:

$$Y^* = \begin{cases} 1 + C^{**}(\eta^{**} - 1) & (\eta^{**} > 1) \\ \eta^{**} & (1 \geq \eta^{**} \geq 0) \end{cases} \quad (25)$$

where η^{**} is the (averaged) value of η in a layer of 2~3 km heights and C^{**} is a parameter, which is taken to be 0.5~1.5. According to many numerical experiments performed so far, the properties of MC and a TC are not very different provided that the value of C^{**} takes this range. The most significant impact of larger values of C^{**} is to make shallow clouds more enhanced in the outer area. In a numerical experiment in Section 3, C^{**} is taken to be 1.5. A smaller value such as 1.0 or 0.5 may give more realistic result. The most reasonable determination of C^{**} (and C_η) remains to be studied. The assumption (25) assures that the growth of MC tends to cease as the stratification in the lower layer becomes moist neutral. (The values of Y^{**} and η approach 1.0 as B in the lower layer becomes 0.)

2.4 Equations with grid-resolved processes and other effects

Now we take account of the effects of the grid-resolved condensation (in the case of oversaturation at the grid-resolved scale), evaporation of grid-resolved cloud water (in the case of unsaturation), conversion of cloud water to rainwater, fall of rainwater relative to the air and evaporation of rainwater. In addition, the effects of the cumulus-scale downdraft D_w , subgrid-scale turbulence F and radiation Q_R are also included. Equations (8)~(10) may be written as

$$\frac{d\bar{\theta}}{dt} = \frac{L}{c_p \pi} (\bar{C} - E_r + C^C) + Q_R$$

$$- \frac{1}{1-\sigma} \frac{1}{\rho} \frac{\partial}{\partial z} [(M^C - \rho \sigma \bar{w})(\theta^c - \bar{\theta})] + D_w(\bar{\theta}) + F_\theta, \quad (26)$$

$$\begin{aligned} \frac{d\bar{q}_v}{dt} &= -\bar{C} + E_r - C^C \\ &- \frac{1}{1-\sigma} \frac{1}{\rho} \frac{\partial}{\partial z} [(M^C - \rho \sigma \bar{w})(q_v^c - \bar{q}_v)] \\ &+ D_w(\bar{q}_v) + F_q, \end{aligned} \quad (27)$$

$$\begin{aligned} \frac{d\bar{q}_c}{dt} &= \bar{C} + C^C - (A + C_o) \\ &- \frac{1}{1-\sigma} \frac{1}{\rho} \frac{\partial}{\partial z} [(M^C - \rho \sigma \bar{w})(q_c^c - \bar{q}_c)] \\ &+ D_w(\bar{q}_c) + F_c, \end{aligned} \quad (28)$$

where $\bar{C}$ is the production rate of cloud water $\bar{q}_c$ due to the grid-resolved condensation (evaporation of cloud water for $\bar{C} < 0$), E_r evaporation rate of rainwater, A autoconversion of cloud water to rainwater and C_o collection rate (increase rate of rainwater due to collection of cloud droplets by raindrops). The effects of the cumulus-scale downdraft and subgrid-scale turbulence are assumed to be the same as those in Y86.

In the following, we include an equation for the mixing ratio of the grid-resolved rainwater $\bar{q}_r$.

$$\begin{aligned} \frac{d\bar{q}_r}{dt} &= -E_r + (A + C_o) + V_r \\ &- \frac{1}{1-\sigma} \frac{1}{\rho} \frac{\partial}{\partial z} [(M^C - \rho \sigma \bar{w})(q_r^c - \bar{q}_r)], \end{aligned} \quad (29)$$

where q_r^c is the mixing ratio of rainwater in the cumulus-scale ascending area, and V_r indicates the effect of fall of rainwater relative to the air.

Since the values of $\bar{\theta}$, θ^c and θ^e ($\bar{q}_v$, q_v^c and q_v^e) are generally different, it is desirable to determine $\bar{C}$ by imposing saturation conditions for the ascending and descending areas separately. However, whether it is saturated or not is determined only from $\bar{\theta}$ and $\bar{q}_v$ in this study. When the following condition is satisfied:

$$\bar{q}_v > q_s(\bar{T}, p),$$

we determine $\bar{C}$, $\bar{T}$ and $\bar{q}_v$ so that it may be just saturated. When it is unsaturated, evaporation may occur. In this situation, cloud water other

than σq_c^c (cloud water in the cumulus-scale ascending area); that is, the amount $(\bar{q}_c - \sigma q_c^c)$ or $(1 - \sigma)q_c^e$ evaporates instantaneously (under the condition that it does not become over-saturated).

Similarly, evaporation of rainwater is allowed for the amount $(1 - \sigma)q_r^e$, where q_r^e is determined from

$$\bar{q}_r = \sigma q_r^c + (1 - \sigma)q_r^e. \quad (30)$$

The assumption for the evaporation rate is the same as that used in Y86.

$$E_r = 4.32(q_{vs} - \bar{q}_v)(1 - \sigma)(\rho q_r^e)^{0.65}. \quad (31)$$

As for q_r^c , its prognostic equation will be described later.

The rates of autoconversion and collection are given as the sum of these quantities in the cumulus-scale ascending and descending areas. In the case when Kessler's (1969) parameterization is used for autoconversion, it can be written as

$$A = \sigma A^c + (1 - \sigma)A^e, \quad (32)$$

$$A^c = \begin{cases} \alpha(q_c^c - q_{c0}) & (q_c^c > q_{c0}) \\ 0 & (q_c^c \leq q_{c0}) \end{cases} \quad (32a)$$

$$A^e = \begin{cases} \alpha(q_c^e - q_{c0}) & (q_c^e > q_{c0}) \\ 0 & (q_c^e \leq q_{c0}) \end{cases}$$

where α and q_{c0} are autoconversion rate and a threshold for autoconversion, respectively. Similarly, the collection rate is given by

$$C_o = \sigma C_o^c + (1 - \sigma)C_o^e \quad (33)$$

$$C_o^c = 929q_c^c(\rho q_r^c)^{0.875} \quad (33a)$$

$$C_o^e = 929q_c^e(\rho q_r^e)^{0.875}.$$

The effect of the fall of rainwater is

$$V_r = \sigma V_r^c + (1 - \sigma)V_r^e, \quad (34)$$

$$V_r^c = \frac{1}{\rho} \frac{\partial(\rho V_T^c q_r^c)}{\partial z},$$

$$V_r^e = \frac{1}{\rho} \frac{\partial(\rho V_T^e q_r^e)}{\partial z}, \quad (34a)$$

where V_T^c and V_T^e the terminal velocities (positive: downward) of rainwater in the ascending and descending areas, respectively. These are given as functions of q_r^c and q_r^e , respectively.

2.5 Cloud water and rainwater for the cumulus-scale

An equation (16a) for q_c^c can be written, adding the effects of autoconversion, collection and one

more term, as follows:

$$\frac{\partial q_c^c}{\partial t} = \frac{1}{\sigma} \left[C^c - \frac{M^c}{\rho} \frac{\partial q_c^c}{\partial z} - (q_c^c - q_c^e)E \right] - (A^c + C_o^c) - \delta q_c^c. \quad (35)$$

On the other hand, an equation for q_r^c is similar to (35) concerning the terms including M^c and E . Adding the effects of autoconversion, collection and fall of rainwater, an equation for q_r^c may be written as

$$\frac{\partial q_r^c}{\partial t} = \frac{1}{\sigma} \left[-\frac{M^c}{\rho} \frac{\partial q_r^c}{\partial z} - (q_r^c - q_r^e)E \right] + (A^c + C_o^c) + V_r^c - \delta q_r^c. \quad (36)$$

The last terms δq_c^c and δq_r^c in (35) and (36) indicate the effects of decreasing q_c^c and q_r^c at the stage after the cumulus-scale mass flux becomes 0. These do not contribute to the changes of $\bar{q}_c$ and $\bar{q}_r$ directly, but the decreases of q_c^c or q_r^c contribute to evaporation that can occur afterwards in unsaturated areas. The evaporation modifies $\bar{q}_c$ and $\bar{q}_r$ through $\bar{C}$ and E_r in (28) and (29), respectively. The most significant effect of δq_c^c is to reduce q_c^c which is advected from the cumulus-scale ascending area, particularly in the upper troposphere. It is assumed that, in the absence of the cumulus-scale mass flux, this reduction occurs at a finite rate:

$$\delta q_c^c = \beta q_c^c, \quad (37)$$

where β is taken to be $0.2 \sim 1.0 \times 10^{-3} \text{ s}^{-1}$ (typically $0.5 \times 10^{-3} \text{ s}^{-1}$, about 30 min in relaxation time). In the presence of the cumulus-scale mass flux, β is taken to be 0. As for δq_r^c above the subcloud layer, it is assumed to be in the same form as (37);

$$\delta q_r^c = \beta q_r^c. \quad (38)$$

In the case of the subcloud layer, δq_r^c expresses the effect that makes q_r^c zero. That is, rainwater which falls into the subcloud layer is not treated as q_r^c .

Finally, the effects of the horizontal advection of q_c^c and q_r^c are added to (35) and (36), respectively. Although the horizontal advection terms cannot be accurately represented in a numerical model, it is necessary to include these effects. In addition, the term V_r^c in (36) is expressed by the advective form (but with an approximation).

$$\frac{\partial q_c^c}{\partial t} + \bar{V} \cdot \nabla q_c^c = \frac{1}{\sigma} \left[C^C - \frac{M^C}{\rho} \frac{\partial q_c^c}{\partial z} - (q_c^c - q_c^e) E \right] - (A^c + C_o^c) - \delta q_c^c, \quad (35a)$$

$$\frac{\partial q_r^c}{\partial t} + \bar{V} \cdot \nabla q_r^c = - \left(\frac{M^C}{\rho \sigma} - V_T^c \right) \frac{\partial q_r^c}{\partial z} - (q_r^c - q_r^e) E / \sigma + (A^c + C_o^c) - \delta q_r^c. \quad (36a)$$

In connection with (36a) an equation (29) for grid-resolved rainwater $\bar{q}_r$ may be written with the term V_r in the advective form, as follows:

$$\begin{aligned} \frac{\partial \bar{q}_r}{\partial t} + \bar{V} \cdot \nabla \bar{q}_r = & -(\bar{w} - V_T^e) \sigma \frac{\partial q_r^c}{\partial z} \\ & - (\bar{w} - V_T^e)(1 - \sigma) \frac{\partial q_r^e}{\partial z} \\ & - \frac{1}{1 - \sigma} \frac{1}{\rho} \frac{\partial}{\partial z} [(M^C - \rho \sigma \bar{w})(q_r^c - \bar{q}_r)] \\ & - E_r + (A + C_o). \end{aligned} \quad (29a)$$

3. An application to a TC at the developing stage

In this paper results from one of numerical experiments of TCs at the developing stage are presented. The primary objective is simply to show that the present model is capable of simulating the mesoscale structure of a TC, particularly the cloud water field, which should be improved by prognostic treatment of the subgrid-scale cloud water (instead of diagnostic one in Y86). Since many efforts are necessary to clarify the impact of the different treatment and more observations are also required to evaluate the degree of validity of the model, the present study should be considered preliminary.

In the numerical experiment we consider an initial vortex in the tropical atmosphere at rest. The maximum wind velocity is taken to be 5 m s^{-1} and it is located at the 200 km radius. The temperature and water vapor fields are chosen to be similar to those of the tropical atmosphere in summer, as done in Yamasaki (1983). The sea surface temperature (SST) used is 301 K in a latitudinal belt of 10~20N, 300 K in the southern hemisphere, 299K to the north of 30N, and linearly interpolated values are used for 20~30N and 10N~equator. This profile of SST is considered to be essentially uniform over a wide area. Although the SST is too

high to the north of 30N, this situation is chosen to avoid the baroclinic instability and only to simulate the development of the initially given vortex to a TC and its mesoscale structure under a simplified environmental condition. In addition to a weak vortex, a north-south gradient of the relative humidity field is given at the initial time. To the north of 30N and to the south of the equator, the relative humidity is taken to be 65% at 900 hPa, and a maximum of 85% is imposed at 15N.

The numerical model used in this study is a hydrostatic σ -coordinate model of Y86. The thermodynamic aspect of the model is replaced by the model described in Section 2, which is written using a z -coordinate but modified for the σ -coordinate. Since the cumulus-scale is treated as the subgrid-scale, the hydrostatic approximation is fairly justified. A triply-nested grid is used, as in Y86. As for the horizontal coordinate, the latitude-longitude coordinate is used, as in Yamasaki (1992). The grid sizes are taken to be 1/2, 1/6 and 1/18 degrees for the coarse, medium and fine grid areas, respectively. The grid interval in the fine grid area is about 5 km. A cyclic condition is used in the west-east direction for the coarse grid area, which covers an area enclosed by 115E and 160E longitude and 40N and 10S latitude. The sizes of the whole area (coarse grid area) with 45 longitude and 50 latitude distances are large enough to simulate a TC under a simplified condition. The fine grid area (135~143E, 20~13N) are contained in the medium grid area (129~149E, 28~8N). The numbers of grid points of the coarse, medium and fine grid areas are 90×100 , 120×120 and 144×126 , respectively. As for the vertical grid, we use a ten-layer model where the horizontal wind velocity is defined at 10 levels. The vertical σ -velocity is defined at 11 levels. Their basic state pressures are 100, 150, 200, 280, 400, 550, 700, 820, 900, 955 and 1010 hPa. In the time integration a leap frog scheme with an Euler backward scheme is used.

The numerical experiment was performed by a coarse-grid model for the first 24 hours. Based on the 24 hour data, medium and fine grid data were constructed by linear interpolation. The central surface pressure is 1002.8 hPa (1008.5 hPa at the initial time), and the maximum wind velocity is 18 m s^{-1} , which indicates that the vortex has got the typhoon intensity at 24h.

At first, the distributions of the surface rainfall intensity (determined from the mixing ratio of rainwater near the surface) are shown at an interval of

Fig.1. Surface rainfall intensities (or rainwater) obtained from the old version of Yamasaki (1986). Isolines are drawn for 1, 4, 16 and 64 mm h⁻¹. The area shown is the fine grid area with a grid size of 1/18 degrees (about 5 km). The intensities are shown at an interval of 6 hours from 30h to 60h.

Fig. 2. Same as Fig. 1 except for the present model.

6 hours. For comparison, the result from the Y86's model is shown in Fig. 1 and that from the present model is in Fig. 2. The area shown is the fine grid area. We can see that an eyewall and spiral rainbands are simulated in both models. Significant differences are found in the position of each rainband and the properties of MC that constitutes the rainband, but it seems to be difficult to say which one simulates the rainbands and MC more realistically. The most significant difference can be found in the direction of the movement of the TC center. The present model predicts the TC center more westward. The reason for this has not been understood yet, but this problem is very important to understanding of the dependence of TC motion on the cumulus-scale parameterization and evaluating of the model validity.

Figure 3 shows the surface pressure and temperatures at 400 and 900 hPa at 48h. A large gradient of the surface pressure is concentrated to the inner portion of the eyewall (at smaller radii inside the eyewall). The central surface pressure is 952 hPa at this time, and it continues to fall (940 hPa at 66h). Corresponding to the low surface pressure, a notable warm core is created. The temperature anomaly at 400 hPa exceeds 10 K. The west-east asymmetry of the temperature field is also one of the features of TCs. Positive anomalies associated with the spiral rainbands are more significant at 900 hPa than at 400 hPa. The positive but small anomaly is important for MC (and the rainbands) to grow at a moderate rate. This is controlled by an appropriate choice of the values of Y^* and η just above the level z^* .

Figure 4 shows the surface rainfall intensity (same as the upper right of Fig. 2) and the buoyancies at 400 and 700 hPa. The buoyancy at each level is determined from the difference between the value of θ_e of the boundary layer air (at a level of the basic state pressure of 930 hPa, or at about 800 m height) and the saturated value of θ_e . The values are calculated as those of θ_e multiplied by c_p . Therefore, the numerical values are similar to the difference of the moist static energy. Since Y^* is determined by (25) and η^{**} is chosen to be a value of η at a level of the basic state pressure of 700 hPa (at a height of about 3 km), the buoyancy estimated at this level is particularly important to the growth of MC and the

Fig. 3. Surface pressure (deviation from 1,000 hPa) and temperatures (°C) at 400 and 900 hPa isobaric surfaces at 48h.

rainbands. A comparison with the rainwater distribution indicates that relatively large buoyancy (latent instability) is generally associated with the rainbands. It is also found in the outer areas of the rainbands located at the outer portion of the domain. The buoyancy is negative in the eye and some descending areas between the inner rainbands mainly because of drying due to downward motion in the subcloud layer.

The buoyancy at 400 hPa has wider areas of negative values. On the whole, the stratification tends to be stabilized in the inner portion of the TC area. On the contrary, strong latent instability is fairly maintained in the outer portion. As shown later, the grid-resolved condensation due to ascending motion in the stable area plays an important role at this level (and above).

Figure 5 shows the distributions of $Y(\overline{w^*} - w_0)$ at four levels of the basic state pressures of 200, 280, 400 and 700 hPa. The cumulus-scale mass fluxes at 200 and 280 hPa can be found only in a small portion of the eyewall and the inner rainbands, but a large portion of the areas having the cumulus-scale mass flux at 700 hPa can be found even at 400 hPa. In many situations of observed TCs, more convective clouds tend to be confined below 500–600 hPa in the outer area. The present situation corresponds to the developing stage of the TC, and a large latent instability is given initially over a wide area.

As far as the physical quantities shown in Figs. 2–5 are concerned, notable differences between the Y86's model and the present model are not found. As already mentioned, the most important difference can be found in the distribution of cloud water because the cumulus-scale cloud water was not treated as a prognostic variable in Y86. Figure 6 shows the mixing ratio of cloud water $\overline{q_c}$ at four levels with the basic state pressures of 200, 400, 700 and 900 hPa. It should be remarked that $\overline{q_c}$ includes q_c^c . At lower levels such as 700 and 900 hPa, most of cloud water is largely due to the cumulus-scale clouds. This can be understood from com-

Fig. 4. Surface rainfall intensity (or rainwater) and buoyancies that an air parcel rising from the subcloud layer acquires at the basic state pressures of 400 and 700 hPa (a measure of latent instability). Positive buoyancies are indicated by three different shades, depending on its magnitude.

Fig. 5. Cumulus-scale mass fluxes at the basic state pressures of 200, 280 (left), 400 and 700 hPa (right).

parison with Fig. 5. Compared with $\bar{q}_r$ in Fig. 4, it is seen that the surface rainfall (or rainwater) in the central area of the TC are caused by an eyewall cloud and a few banded clouds. Spiral rainbands are due to not only banded clouds but also scattered clouds. At 400 hPa, cloud water is distributed in wider areas, particularly in the central area of the TC. Comparison with Fig. 5 indicates that some of the clouds in the eyewall are due to grid-resolved condensation. At 200 hPa, the fine grid area is entirely covered by cloud water. We can see a notable CDO (central dense overcast) that is also typical of TCs.

Now we will show the cumulus-scale cloud water q_c^c separately. The upper panel of Fig. 7 shows this quantity at 700 hPa. Compared with total cloud

water $\bar{q}_c$ (upper right of Fig. 6) it can be seen that the difference between these two distributions is not noticeable. That is, most of the cloud water at 700 hPa is due to the cumulus-scale. The middle panel shows three areas of different cloud types. The darkest shade indicates that q_c^c exists in the grid-saturated area. The area of this type can be seen only in the eyewall and very small portions of the rainbands. The medium shade, which covers most portion of the shaded areas, indicates that q_c^c exists in the grid-unsaturated area. The thin shade indicates that q_c^c does not exist and only grid-resolved cloud water exists. This area can be seen only in several small areas such as (17.5N, 139.5E). The lower panel shows the relative humidity (and cloud water in areas of grid-resolved saturation).

Fig. 6. Total cloud water content $\rho \bar{q}_c$ at the basic state pressures of 200, 400, 700 and 900 hPa. Isolines are drawn every 0.1 gm^{-3} . At 200 hPa, areas of water content larger than 0.3 and 0.4 gm^{-3} are indicated by the medium and darkest shades, respectively. At other levels, areas of water content larger than 0.2 and 0.3 gm^{-3} are indicated by these shades.

Comparison with the distribution of q_c^c indicates that a number of the cumulus-scale clouds exist in grid-resolved unsaturated areas. The saturated areas are illustrated by the darkest and thin shades in the middle panel. It should be remarked about q_c^c that some of the medium shaded areas are relatively wide. Some of q_c^c in such an area are due to inactive clouds that were of the cumulus-scale before and are evaporating at a finite rate. This can be seen from comparison with Fig. 5 which shows active cumulus-scale cloud areas. The size of the medium shade depends on the evaporation rate β

in (37). Cooling due to evaporation of cloud water associated with an inactive cloud tends to enhance the growth of convective clouds which exist in its vicinity. This property was also noted in Nasuno and Yamasaki (2000).

In contrast to 700 hPa (3 km height), most of cloud water at 900 hPa (1 km height) exist in the grid-resolved saturated area (not shown), though the cumulus-scale cloud water coexists in most of the areas. This is one of the features of the present model (as well as the Y86's model), as mentioned in Section 2.

Fig. 7

Fig. 8

Figure 8 shows cumulus-scale cloud water q_c^c , cloud type and relative humidity at 400 hPa. Compared with those at 700 hPa, it can be seen that there exist much more grid-resolved saturated areas. A large part of the cumulus-scale cloud water exists in the saturated areas. The grid-resolved condensation without the cumulus-scale also takes place in many portions of the domain. A very dry area inside the eye is clear. Dry areas between spiral rainbands are also seen. At 200 hPa, cloud water (upper left of Fig. 6) is mostly due to the grid-resolved condensation (not shown). This can be understood from the cumulus-scale mass flux shown in Fig. 5.

Some remarks are made here concerning the difference between cloud water fields obtained from the present model, and Y86. Since the subgrid-scale cloud water was determined diagnostically from a budget equation for the implicit cloud in Y86, the total cloud water (as well as the subgrid-scale cloud water) was not obtained. The grid-resolved cloud water field at 200 hPa in Y86 (not shown) is similar to the total cloud water field $\bar{q}_c$ shown in Fig. 6 because the amount of the subgrid-scale (cumulus-scale) cloud water q_c^c is small. However, it is important to remark that a prognostic treatment of q_c^c results in increased amount of $\bar{q}_c$ at 200 hPa through upward transport of q_c^c in the cumulus-scale ascending area, leading to a better simulation of the CDO in the present model. The pattern of grid-resolved cloud water at a height of 1 km (basic state pressure of 900 hPa) obtained from Y86 is also similar to that of $\bar{q}_c$ in Fig. 6 because most of the cumulus-scale take place in grid-resolved condensation areas (though quantitative comparison is difficult to make). At a height of 3 km (basic state pressure of 700 hPa), grid-saturated areas are much limited, as shown in Fig. 7.

Fig. 7. Top: cumulus-scale cloud water content ρq_c^c , middle: cloud types; the darkest and the medium shades indicate that q_c^c exists in grid-resolved saturated and unsaturated areas, respectively. The thin shade indicates that q_c^c does not exist but grid-resolved cloud water $\bar{q}_c$ exists. bottom: relative humidity (unit: 10%) and cloud water for grid-saturated areas. These are quantities at the basic state pressure of 700 hPa.

Fig. 8. Same as Fig. 7 except for 400 hPa.

This is also true for Y86. There was some expectation that drying at this level in Y86, which may be somewhat excessive, might be avoided by a prognostic treatment of q_c^c . However, Fig. 7 indicates that much cloud water exists in cumulus-scale ascending areas and grid-resolved relative humidities are not high in most areas. In spite of similar patterns of the cloud distribution in both models, the present model obviously gives a better simulation of cloud water field, particularly in a quantitative sense.

4. Concluding remarks

As an extension of Yamasaki (1986, 1987), an improved version of a tropical cyclone model that resolves mesoscale organized convection is presented in this paper. An emphasis is placed on the prognostic treatment of cumulus-scale cloud water that is treated as the subgrid-scale. Results from only one numerical experiment are described simply to show an example of the mesoscale features of a TC obtained from the new version. Since the essential aspects of the old model are retained, the features of TCs obtained from both models are similar except that the distribution of cloud water, which was treated in a prognostic way only for the grid-resolved in the old model, is better simulated in the new version. This is an obvious improvement of the model. Another important improvement is concerned with the area ratio of cumulus-scale ascending motion. Since any assumption of the ratio should not be satisfied at the present knowledge of this problem, the present study does not give any definite information on the impact of the finite ratio on the properties of TCs. It should be remarked that many mechanisms of the TC simulated by the new version are essentially similar to those discussed in Yamasaki (1986, 87, 88, 89) with the old model and Yamasaki (1977, 1983) with a cumulus-scale resolving model. Therefore, these are not mentioned specifically in this paper.

The present mesoscale-convection-resolving model as well as Yamasaki's (1986) is based on recognition that inclusion of the effects of the cumulus-scale is important in a numerical model with a horizontal grid size of 5~20 km. Numerical experiments with coarse grid models that do not include the cumulus-scale effects have been made by Rosenthal (1978), Jones (1977) and recently Liu et al. (1997). The present author has also performed such numerical experiments for the past 15 years in order to understand the impact of the

cumulus-scale parameterization, and two opposite effects of the cumulus-scale under different situations have been recognized clearly. For an axisymmetric case these are demonstrated in Nasuno and Yamasaki (2000). Results from numerical models that do not include the cumulus-scale effects are significantly different (not shown). The degree of its importance depends on how mesoscale organized convection is controlled by frictional flow, and also on the grid size used. The eyewall that is strongly controlled by frictional convergence can be simulated to a fairly degree even without the cumulus-scale effects, particularly for a very strong TC, as also seen in a numerical experiment of Liu et al. (1997). However, spiral rainbands, particularly those embedded in weak frictional flow in the outer area cannot be simulated well without the cumulus-scale effects. The magnitude of the cumulus-scale impact also depends on the grid size. An eyewall and inner rainbands close to the eyewall were simulated much better for a 2 km grid than for a 6 km grid in a numerical experiment of a strong hurricane by Yau et al. (1999). Some mesoscale features obtained for a 2 km grid, which were presented at the 23rd Conference on Hurricanes and Tropical Meteorology, were surprisingly similar to those obtained from Yamasaki's (1986) and the present model for grid sizes of 5~10 km.

Since the present model includes many uncertain assumptions and indefinite parameters, efforts should be continued to clarify the mechanisms of these physical processes and thereby to improve the model. Numerical experiments of TCs as well as basic studies on individual processes will be necessary to improve uncertain assumptions and specify appropriate values of parameters. Comparisons with results from the old version and observational studies in the past remain to be made in the forthcoming papers. In particular, it is very important to compare the model results with those from a cumulus-scale resolving model with a grid size of 1 km or less. Although this is not easy owing to computer resources for a moment, basic researches in this direction is particularly needed for better understanding of the mechanisms of some important features of TCs and an improvement of the mesoscale-convection-resolving model. Finally it should be remarked that inclusion of the ice phase remains to be done. In order to improve both cumulus-scale and mesoscale resolving models an improvement of existing parameterization schemes of cloud microphysical processes has been made

with an axisymmetric model (Sato and Yamasaki 1999).

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