A Study of Cloud Clusters Associated with a Baiu Front by Use of a Mesoscale-Convection-Resolving Model

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(Manuscript received 17 December 2001, in revised form 24 April 2002)

Abstract

Numerical experiments are performed to simulate and understand cloud clusters associated with a Baiu front observed over Kyushu and the East China Sea on 16 July 1993. The clusters treated in this study are those that are not associated with any synoptic-scale low. A numerical model used is a meso-scale-convection-resolving model (MCRM) that resolves mesoscale organized convection by a grid, and treats cumulus convection as the subgrid-scale (Yamasaki 2001). The primary objective of this study is to investigate to what degree the MCRM can describe cloud clusters, and mesoscale organized convection that constitutes the clusters. In this study the grid size is taken to be 5/36 degrees (about 15 km) in the finest grid area of the triply-nested grid model. A global objective analysis data of JMA is used as the initial condition (00UTC, 16 July).

A numerical experiment is performed by use of the MCRM referred above. For comparison, an old version of the MCRM (Yamasaki 1986) is also used, and an additional experiment is made for the case without parameterization of cumulus convection. The rainfall distributions at 12 hours after the initial time are compared with AMeDAS data, and with those from Peng and Tsuboki (1997) in which four cloud parameterization schemes are used.

It is emphasized that the most important factor to prediction and understanding of the cloud cluster and rainfall over Kyushu at 12 hours is the eastward movement of a latently unstable area that exists at the initial time. A comparison of the results from the MCRM with those from the case without cumulus-scale parameterization shows that the effects of cumulus convection are not essential to the eastward movement of the unstable area and rainfall over Kyushu at 12 hours, but time evolution (behavior) of mesoscale organized convection and cloud clusters are quite different, depending on the inclusion of the cumulus-scale effects.

The performance of the model as to how cloud clusters and mesoscale organized convection behave realistically under the given initial condition is discussed, based on the studies in the past and physical considerations.

1. Introduction

The Baiu (or Meiyu) front is a quasi-stationary front that is formed between the tropical air mass over the western north Pacific (or tropical monsoon air mass) and the polar (marine and continental) air mass in a period from June (or May) to July. It brings a rainy season and sometimes causes heavy rainfalls in some countries in the Far East including Japan and China. Therefore, understanding and prediction of this phenomena have been one of the most important problems in meteorology in these countries.

It is well known that not only a large gradient of temperature but also that of moisture play an important role in the Baiu front. To the south of the front a large amount of moisture in
the lower troposphere is supplied by the southwesterly or southerly flow associated with the subtropical high. The stratification in this area is characterized by conditional instability, latent instability and/or convective instability. On the contrary, it is usually stable to the north of the front. This marked contrast as well as the vertical shear associated with the large-scale flow (including the mid-latitude westerlies) is one of the important features that distinguish the Baiu front from tropical disturbances and tropical cyclones with respect to the hierarchical property makes it somewhat difficult to construct an appropriate model. The Baiu front is embedded in a field of the large-scale such as 5,000 km. On the other hand, the horizontal scale of cumulus convection is only a few kilometers. In order to resolve cumulus convection a fine grid size such as 1 km~100 m has to be used. The use of such a fine grid model (cumulus-convection-resolving model) is restricted to an area that is too small to properly treat the front interacting with the large-scale. Even when we study cloud clusters, the use of it is still expensive though it serves as a very important tool to understand and predict the phenomena under restricted conditions (with somewhat larger grid size), as done by Kato (1998), Seko et al. (1999) and others.

Historically, the grid size of numerical models used for the studies of the Baiu front has decreased; a grid size of 100~50 km was used many years ago, and 40~20 km in recent years. Now we can use a 10~5 km grid to treat many problems. In developing such numerical models it has been one of the most difficult but important problems how we treat the subgrid-scale convection. Several schemes of parameterization, which is a technical way of incorporating the effects of subgrid-scale convective clouds on the larger-scale motions, were proposed for various purposes. The schemes include those by Manabe and Strickler (1964), Kuo (1965, 1974), Arakawa and Schubert (1974), Anthes (1977), Fritsch and Chappell (1980), Kuo and Anthes (1984) and others. Since these were proposed mainly for the large-scale phenomena and tropical cyclones, some modified versions were also proposed to treat mesoscale phenomena (Kain and Fritsch 1990; Grell 1993 and others).

In the studies of tropical cyclones the present author developed a model that intended to resolve mesoscale organized convection by the grid and treat cumulus convection as the subgrid-scale (Yamasaki 1986, 1987). Here,
mesoscale organized convection has a specific meaning; it is an organized form of cumulus convection. In contrast to the time scale of cumulus convection (a few tens of minutes), mesoscale convection (referred to as MC) takes a time scale of a few hours (3 hours) under many conditions and it takes even a longer time scale (up to more than 10 hours) under certain conditions.1 The time scale of a few hours corresponds to that of mesoscale phenomena (known as, for instance, thunderstorms since the late 1940s) and that of the mesoscale cell in the Baiu frontal zone, as recognized from the observational studies referred above2. In the tropical atmosphere, ensembles of mesoscale convective cells constitute spiral rainbands in tropical cyclones and cloud clusters (including squall clusters). From numerical modeling studies, a recognition of MC as one of important modes of moist convection was made in those studies of tropical cyclones (Yamasaki 1983) and tropical disturbances and squall clusters (Yamasaki 1984) that used the cumulus-scale-resolving model. Based on such recognition a mesoscale-convection-resolving model of Yamasaki (1986) was developed, and it has been used to study many aspects of tropical cyclones (e.g., Yamasaki 1989, 1992).

This paper presents the results from numerical experiments of a cloud cluster associated with a Baiu front with the mesoscale-convection-resolving model (hereafter, referred to as MCRM). The model essentially differs from those that include parameterization schemes of Kuo (1965), Arakawa and Schubert (1974) and others in that the former intends to resolve MC by the grid and to implicitly treat (or parameterize) only the effects of cumulus convection3. In view of the importance of cooling due to rainwater evaporation in the subcloud layer in many situations, an equation for subgrid-scale (cumulus-scale) rainwater is treated. The vertical distribution of cumulus-scale heating is formulated so as to simulate the whole stage (life cycle) of MC realistically. The detail of the formulation and its physical meaning are given in Yamasaki (1986, 1987, 2001) and Nasuno and Yamasaki (2001).

In this study we treat a cloud cluster observed in Kyushu district on 16 July 1993. The reason why this cluster is chosen is that this case was studied by Peng and Tsuboki (1997). They performed numerical experiments with four schemes of cloud parameterization and made comparison among the four results. The model used was a 1988 version of Japan Spectral Model (JSM), and the grid size used was about 40 km. Their results are used as a reference in this paper, but comparison of the results from the MCRM with their results does not necessarily give sufficiently meaningful discussion because the grid size used in this study (about 15 km) is different from theirs, and because some different formulations are used with respect to not only the cloud process but also other physical processes. The primary objective of this study is, as a basis of finding appropriate directions of model improvements, to discuss to what extent the present MCRM can simulate cloud clusters associated with the Baiu front, based on considerations as well as the general views obtained from many studies on cloud clusters and mesoscale organized convection in the past three decades.

2. Outline of the model

As mentioned in section 1, we use a MCRM that was developed with an intention of resolving MC and treating cumulus convection as the subgrid-scale. The original version of the MCRM is given in Yamasaki (1986, 1987). In the present study an improved version described in Yamasaki (2001) is used in the con-

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1 The time scale depends on the intensities of the cold pool and the environmental low-level flow, its vertical shear, frictional effects and so on (Yamasaki 1984).

2 Mesoscale convection (MC) may correspond to most of meso-β-scale and meso-γ-scale convective systems described in Ninomiya and Akiyama (1991).

3 The term cumulus convection has been often used to mean what is different from that used before the 1970s. That is, the term has been used to mean not only a basic mode of moist convection having a time scale of a few tens of minutes, as originally defined, but also its ensemble. Correspondingly, the term cumulus parameterization has been often used to mean that the effects of not only cumulus convection but also its organized convection are parameterized. It appears that the term ‘parameterization of cumulus convection’ was originated in the 1960s when the concept (or the importance) of the mesoscale was not accepted widely. To avoid confusion a different term cumulus-scale parameterization is used in the MCRM in that only the effects of cumulus convection as a basic mode of moist convection are parameterized.
control case. For comparison, a model of Yamasaki (1986) is also used. Hereafter, Yamasaki (1986) and Yamasaki (2001) are referred to as Y86 and Y01, respectively. In both models, subgrid-scale rainwater as well as grid-resolved cloud water and rainwater is predicted with a prognostic equation. Ice phase is not taken into account. The amount of subgrid-scale heating due to cumulus convection is assumed to be nearly proportional to the cloud mass flux (referred to as cumulus-scale mass flux in the MCRM). It is assumed that the mass flux depends on the grid-resolved vertical velocity at the top of the boundary layer and the buoyancy of an air parcel rising from the subcloud layer. In determining the thermodynamic properties of the rising air, not only turbulent entrainment and detrainment but also dynamic entrainment or detrainment are taken into account.

The model of Y01 is different from that of Y86 in two respects. One is that subgrid-scale cloud water is treated with a prognostic equation. In the old version as well as many parameterization schemes such as Arakawa and Schubert (1974), subgrid-scale cloud water is a variable that is determined diagnostically. In this case, cloud water that remains after the termination of cumulus-scale heating, and its evaporation are not treated. Therefore, total cloud water distribution cannot be simulated or discussed properly. The other model difference is that the fractional ratio of the cumulus-scale updraft area is not assumed to be sufficiently small compared to unity. An assumption of a small fractional ratio has been used in many schemes of cloud parameterization, as in Ooyama (1971) and Arakawa and Schubert (1974). In the MCRM of Y86 this assumption was used as a crude approximation because of uncertainty in determining the ratio. In the new version of Y01, a finite ratio is introduced using a simple assumption. (Numerical experiments for tropical cyclones have indicated that the properties of spiral rainbands as well as tropical cyclones are not essentially different between the two versions.)

The second modification mentioned above leads to the use of those equations for potential temperature, water vapor and cloud water which include subgrid-scale condensation and vertical transports of heat, water vapor and cloud water. The equations described in Y01 are reproduced below:

$$\frac{d\bar{\theta}}{dt} = \frac{L}{c_p \bar{\varpi}} (\bar{C} - E_r) + Q_R + F_\theta + \frac{L}{c_p \bar{\varpi}} C^C - \frac{1}{1 - \sigma \rho \bar{\varpi} \bar{z}} [(M^C - \rho \sigma \bar{w})(\theta^c - \bar{\theta})] + D_w(\bar{\theta}),$$  

(1)

$$\frac{d\bar{q}_v}{dt} = \bar{C} - E_r + F_q - C^C - \frac{1}{1 - \sigma \rho \bar{\varpi} \bar{z}} [(M^C - \rho \sigma \bar{w})(q_v^c - \bar{q}_v)] + D_w(\bar{q}_v),$$  

(2)

$$\frac{d\bar{q}_c}{dt} = \bar{C} - (A + C_O) + F_c + C^C - \frac{1}{1 - \sigma \rho \bar{\varpi} \bar{z}} [(M^C - \rho \sigma \bar{w})(q_c^c - \bar{q}_c)] + D_w(\bar{q}_c),$$  

(3)

where $\bar{\theta}$, $\bar{q}_v$, and $\bar{q}_c$ are grid-resolved (and grid-averaged) potential temperature, water vapor mixing ratio and cloud water mixing ratio (total cloud water including cloud water in the cumulus-scale ascending area), respectively. Each grid area is assumed to consist of cumulus-scale ascending and descending areas when $\sigma$ (fractional ratio of the cumulus-scale ascending area) is not zero. Other notations in (1)~(3) are $\bar{C}$: the production rate of cloud water due to the grid-resolved condensation (evaporation of cloud water for $\bar{C} < 0$), $E_r$: evaporation rate of rainwater, $A$: rate of autoconversion of cloud water to rainwater and $C_O$: rate of collection of cloud droplets by raindrops, $Q_R$: effect of radiation and $F$ ($F_v$, $F_q$ and $F_c$): effects of subgrid-scale turbulence. The last three terms in these equations represent the effects of subgrid-scale cumulus convection; $\theta^c$, $q_v^c$ and $q_c^c$ are potential temperature, water vapor mixing ratio and cloud water mixing ratio in the cumulus-scale ascending area, respectively, $C^C$: cumulus-scale condensation rate, $M^C$: cumulus-scale mass flux, $\bar{w}$: grid-resolved vertical velocity, and $D_w$: the effects of the cumulus-scale downdraft.

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4 A MCRM of this type has been also developed by Nasuno and Yamasaki (2001) in which Kuo’s (1965) formulation proposed to parameterize the effects of the whole clouds is used to represent the cumulus-scale effects, with several important modifications.
Other notations are used in a conventional way. The terms that include $M^c$ represent the effects of the cumulus-scale vertical transports of heat, water vapor and cloud water. Equations (1)–(3) are formally similar to those that were known in the 1960s. However, these are quite different as to what is resolved by the grid, and what is parameterized.

An equation for the mixing ratio of total rainwater $\bar{q}_r$ may be written as

$$\frac{dq_r}{dt} = -E_r + (A + C_0) + V_r$$

$$-\frac{1}{1 - \sigma} \frac{1}{p} \frac{\partial}{\partial z} [M^c - \rho \sigma \bar{w})(q^c_r - \bar{q}_r)],$$

where $q^c_r$ 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. The terms $V_r$, $A$ and $C_0$ are expressed as the sum of the corresponding quantities in cumulus-scale ascending and descending areas.

Cumulus-scale condensation rate $C^c$ is assumed to be of the following two forms in Y01;

$$C^c = -\frac{M^c}{\rho} \frac{\partial}{\partial z} (q^c_r - q^c_v) E, \quad (5a)$$

$$\frac{L}{c_p} C^c = \frac{M^c}{\rho} \frac{\partial}{\partial z} \frac{L}{c_p} (q^c_v - q^c_r) E, \quad (5b)$$

where $E$ indicates the effects of entrainment (dynamic and turbulent entrainment) and $q^c_v$ is water vapor mixing ratio in the cumulus-scale descending area (calculated from $\bar{q}_r$, $q^c_v$ and $\sigma$). The assumption (5a) is obtained from a balance condition based on the water vapor budget for the cumulus-scale ascending area. When $-\frac{L}{c_p} (\frac{\partial q^c_v}{\partial z})$ is equal to $\frac{\partial}{\partial z} (q^c_v - q^c_r)$ (5a) is identical to (5b). The use of (5b) makes the model behavior more similar to that described by Y86, particularly when the vertical transport term of (1) and the entrainment term of (5b) are very small. In the present study the following assumption is used instead of (5a) or (5b):

$$\frac{L}{c_p} C^c = \frac{M^c}{\rho} \frac{\partial}{\partial z} + (\theta^c - \theta^s) E, \quad (5c)$$

where $\theta^c$ is the potential temperature in the cumulus-scale descending area. This is obtained from a balance condition based on the heat budget for the cumulus-scale ascending area. That is, warming due to condensation is assumed to take place so as to cancel cooling due to adiabatic motion and entrainment. The amount of $C^c$ obtained from (5c) is between those from (5a) and (5b) (largest for (5a)) except in the uppermost cloud layer.

The cumulus-scale mass flux $M^c$ is the most important quantity to determine the properties of MC. The assumption used in Y86 and Y01 is as follows:

$$M^c = \begin{cases} Y(\bar{w}^* - w_0) & (\bar{w}^* > w_0) \\ 0 & (\bar{w}^* \leq w_0) \end{cases}, \quad (6)$$

$$Y = Y^* \eta, \quad (7)$$

$$B = T^c(1 + 0.608 q^c_v - q^c_v - q^c_r) - T(1 + 0.608 \bar{q}_r - \bar{q}_r - \bar{q}_r), \quad (8)$$

where $\bar{w}^*$ is the grid-resolved vertical velocity at a certain level $z^*$ (around the top of the boundary layer), $w_0$ a threshold value, which is taken to be 1 cm s$^{-1}$ in this study, $\eta$ is the normalized cumulus-scale mass flux which is taken to be unity at $z^*$, and $Y^*$ is the ratio of the cumulus-scale mass flux at $z^*$ to the meso-scale mass flux $\rho \bar{w}^*$ at the same level when $w_0 = 0$, $B$ a measure of buoyancy of a rising air parcel (latent instability), $T^c$ temperature in the cumulus-scale ascending area and $T$ grid resolved temperature. The terms including $q^c_v$ and $\bar{q}_r$ in (9) indicate drag forces due to rainwater. (These terms are missing in (24) of Y01, though these were included in the numerical experiments.) The value of $C_q$ in (8) 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 of $\eta$, $C_q$ is 0 at $z^*$. According to numerical experiments, model results are not sensitive to the value of $C_q$, provided that it takes 0.2–0.3 for positive values of $B$.

The value of $Y^*$ contributes strongly to temperature change at $z^*$ and therefore, to the growth of MC. (The values of $\eta$ just above the level $z^*$ are also very important.) The growth of MC should be sensitive to the value of $B$ in the lower atmosphere just above the level $z^*$. Therefore, $Y^*$ was assumed to depend on $\eta^*$ in the following form in Y86:

$$Y^* = \begin{cases} 1 + C^{\eta^*} (\eta^* - 1) & (\eta^* > 1) \\ \eta^* & (1 \geq \eta^* \geq 0) \end{cases}. \quad (10)$$
where $\eta'''$ is the (averaged) value of $\eta$ 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 TC are not very different provided that the value of $C'''$ takes this range. In the present study $C'''$ is taken to be 0.5.

The formulation of the mass flux described above is very different from that assumed by Arakawa and Schubert (1974). This is partly because the mass flux used in the MCRM means cumulus-scale mass flux. The assumption of the mass flux is formulated so as to simulate the properties (time scale and life cycle) of MC realistically to a considerable extent, based on a linear theory. For further detail of the formulations of the thermodynamic aspect of the model, the reader is referred to Yamasaki (1986, 1987, 2001).

As for the fractional ratio $\sigma$, it was assumed to be as follows in Y01:

$$
\sigma = \begin{cases} 
\sigma_0 & (q_c/q_c^c > \sigma_0) \\
(q_c/q_c^c) & (q_c/q_c^c \leq \sigma_0), 
\end{cases} \tag{11}
$$

where $\sigma_0$ is an upper limit of $\sigma$. The value of $\sigma_0$ may take a value ranging from 0.2 to 0.4. It is chosen as 0.3 in this study. When it is smaller than 0.2, the vertical velocity associated with MC is very strong, its horizontal scale is small, and more MC forms. The assumption (11) is tentatively used, and it should be improved in future studies.

As mentioned in section 1, it is one of important features of Y86 to predict subgrid-scale rainwater $q_c^c$. On the other hand, subgrid-scale cloud water $q_c^c$ is treated diagnostically in Y86. A prognostic treatment of $q_c^c$ is made in Y01. Prognostic equations for these variables are not reproduced in this paper. Briefly, cumulus-scale condensation rate $C^C$ given by (5c) is a major source for $q_c^c$, which evaporates in the unsaturated area after cumulus-scale parameterization is terminated. Autoconversion and collection in the cumulus-scale ascending area produce $q_c^c$, which falls relative to the air.

The dynamical aspect of the model is almost the same as described in Y86 except that the spherical coordinate (Yamasaki 1992) is used. The vertical coordinate is sigma, and the hydrostatic equilibrium is assumed. Since cumulus convection is implicitly treated, this assumption is justified to a fairly degree. Equations of motion and mass continuity equation (an equation for the surface pressure and the vertical sigma-velocity) are used.

A triply-nested grid is used with the so-called two-way nesting (Y86). In this study the grid size is taken to be 5/4, 5/12 and 5/36 degrees in the coarse, medium and fine grid areas, respectively. The grid size of the fine grid area is about 15 km. The domains of the three grids are chosen as (70 N~10 S, 40 E~170 W), (55~5 N, 100~150 E) and (45~25 N, 120~140 E). Under this setup, the numbers of the coarse, medium and fine grids are 121 $\times$ 65, 121 $\times$ 121 and 145 $\times$ 145, respectively.

As for the vertical grid, the vertical sigma-velocity is defined at 11 levels; their basic state pressures are 100, 150, 200, 280, 400, 550, 700, 820, 900, 955 and 1010 hPa. The horizontal velocities are predicted for 10 layers. Potential temperature and mixing ratios of water vapor, cloud water and rainwater are predicted at the same levels as the vertical velocity except that the uppermost and lowest levels are 112.5 and 996.3 hPa, respectively. (Results from a 20-layer model are essentially similar to those from the present 10-layer model. This is closely related to the adopted scheme of parameterization of cumulus convection which redistributes heat and moisture in the vertical direction. Needless to say, stratiform clouds in the upper troposphere should be better described by the former. Because of the use of the latter, the numerical experiments have been performed more easily since Yamasaki (1992), without losing the essence of the problems to be discussed.)

3. Observed cloud clusters and the initial condition

The initial time for the numerical experiment is 00UTC, 16 July 1993. The infrared satellite imagery from GMS-4 at the initial time (Fig. 1a) shows that a small cloud cluster (referred to as cluster ‘A’) was observed about 300 km to the west of Kyushu (the westernmost part of the Japan Islands). In addition, two large clusters are observed in the northwestern part of the East China Sea and over the sea to the south~southwest of Kyushu. Figure 1b shows the visible imagery corresponding to Fig. 1a. According to Peng and Tsuboki (1997; referred
to as PT), cluster ‘A’ took a mature stage at 03UTC, and a new cluster called cluster ‘B’ formed at (32 N, 128.5 E) at 08UTC. At 12UTC (Fig. 1d), clouds are seen over Kyushu and off the west coast of Kyushu. According to the rainfall rate observed by AMeDAS (Automated Meteorological Data Acquisition System), which is shown in Fig. 2 of PT, cluster ‘B’ brought much rainfall over Kyushu during a 12-hour period beginning around 12UTC. (The rainfall rate at 12UTC, adapted from PT, is shown in Fig. 4a.) At 00UTC, 17 July, the center of cluster ‘B’ moved over the sea to the southeast of Kyushu, as seen in Fig. 1 of PT.

Figure 1c is shown for 06UTC, 16 July. The descriptions in PT referred above indicate that a cluster to the southwest of Kyushu is neither cluster ‘A’ nor cluster ‘B’. However, this statement does not seem to be appropriate. This is suggested from the satellite imageries at 03UTC and 09UTC (not shown) and radar observations. Figure 2 shows radar echoes at a 1-hour interval in a period from 00UTC to 06UTC, and at 09 and 12UTC. A banded echo corresponding to cluster ‘B’ at 12UTC (21JST) can be traced back to that in the western portion of a large echo, which consists of three echoes at 00UTC. The cluster corresponding to the echo in the western portion can be referred to as cluster ‘B’. Cluster ‘B’ can be clearly identified, in the satellite imagery of Figs. 1a and 1b, as a cluster in the western portion of the large cluster to the south-southwest of Kyushu. The visible imagery also indicates that four convective clouds are located in the northeastern edge of cluster ‘B’. It is also seen that cluster ‘A’ can be identified as banded convective clouds and that the banded clouds are connected with those in cluster ‘B’. The radar echo corresponding to the southeastern portion of cluster ‘A’ can be seen at the left edge of Fig. 2a. As suggested from the radar observation, the middle portion (cluster M) of the large cluster (Fig. 2a) decays rapidly after 00UTC. The south-
eastern portion (cluster N) moves in the east-southeast direction, and it is seen over the ocean to the southeast of Kyushu at 06UTC in Fig. 1c (and also at 12UTC in Fig. 1d). The cluster to the southwest of Kyushu at 06UTC is referred to as cluster 'B' in this paper, though it may include part of cluster 'A'. Our interest will be concentrated on cluster 'B', the existence of which is now recognized throughout the period under consideration.

An initial condition used for the numerical experiment is taken from the data of the JMA global objective analysis (GANAL). The data for 1993 is given at grid points at an interval of 15/8 degrees. Grid point values are calculated, by linear interpolation, for the three different grid areas in the model. Any special initialization technique is not used except that the horizontal wind is adjusted so as to make the total horizontal divergence zero in each grid area. The initial conditions for $\bar{q}_c$, $\bar{q}_r$, $q_c^e$ and $q_r^e$ are such that these quantities are zero.

The initial fields of the surface pressure (with the wind at a height of about 250 m) and relative humidity at 900 hPa are shown in Figs. 3a and 3b, respectively. In this paper the height...
A low associated with a Baiu front is located over the China Continent, nearly 1,000 km to the west-northwest of Kyushu. This low is synoptic-scale one (called a mesoscale cyclone in PT). The surface pressure and relative humidity fields as well as the satellite imagery indicate that the Baiu front extends from the low center to another low over the Pacific Ocean just to the south of Kanto district (eastern part of the Japan Islands) through the southern edge (31°N, 131°E) of Kyushu.

Figure 3c shows the wind velocity at 930 hPa. It is seen from this figure as well as Fig. 3a that, to the south of the Baiu front, southwesterly flow prevails over the East China Sea with wind velocities of more than 8 ms⁻¹. This relatively strong wind enhances surface fluxes and contributes to convective activity, as seen later. In the present case a low-level Baiu frontal jet is located around 30°-31°N below 500 hPa though it is weak (less than 15 ms⁻¹). In the upper layer of 100-250 hPa, northerly flow associated with a Tibetan high prevails over Kyushu (not shown).

Figure 3d shows the difference between \( \theta^B_e \) (equivalent potential temperature at a certain level in the boundary layer) and \( \theta^e(700) \) (equivalent potential temperature of the hypothetically saturated air with the same temperature at 700 hPa). Hereafter, the difference is denoted by \( \Delta \theta_e(700) \). Areas of positive values indicate that an air parcel rising from the boundary layer without entrainment has positive buoyancy at 700 hPa. The level of the parcel origin is taken to be about 750 m. The sign of \( \Delta \theta_e \) is not necessarily in accord with that of \( B \) given by (9), but \( \Delta \theta_e \) can be considered an important measure of latent instability. Needless to say, the top level of convective clouds...
depends on $\Delta \theta_c$ in the upper troposphere. However, $\Delta \theta_c$ estimated for a layer around 700 hPa can be considered to be a good index for convective activity in many situations. In the figure, the value of $c_p \Delta \theta_c$ is shown (unit: 0.1 cal g$^{-1}$), where $c_p$ is the specific heat at constant pressure (0.24 cal g$^{-1}$ K$^{-1}$ or 1004 J kg$^{-1}$ K$^{-1}$). The value is nearly equal to that of the difference between the corresponding moist static energy. It is seen from the figure that a latently unstable area is present to the west of Kyushu. It is located at the frontal zone and in an area to the south of the front. A maximum instability is located about 500 km far from the low center. It should be noted that clusters 'A' and 'B' are situated just to the northeast and east of (and very close to) the location of maximum instability, respectively.

As mentioned above, it is assumed in the numerical experiment that cloud water and rainwater do not exist at the initial time. It should be also remarked that the initial field shown in Fig. 3 does not include the mesoscale (at least, meso-$\beta$-scale and meso-$\gamma$-scales). As shown in Figs. 1 and 2, clouds are observed at the initial time. Therefore, it is not expected that observed behavior of cluster 'B' can be simulated in the present setup of the numerical experiment, at least in early hours of the time integration. Any initialization with an intention of simulating the mesoscale in the early hours is not made in this study, as in most of past numerical experiments of this type including PT. Since the rainfall over Kyushu at 12UTC and its location are roughly simulated in the four schemes of cloud parameterization in PT, numerical experiments with the MCRMs can be expected to give any significant results even when special initialization is not made. One of important objectives of this study is to see what happens in the present model for the given initial condition, and how we can get better understanding of cloud clusters and the model properties.

4. Results from numerical experiments

A numerical time integration was performed up to 24 hours, but the main discussions are devoted to the results from the first 12 hour integration to avoid a too lengthy paper. Our first interest is to what extent the model can simulate cluster 'B' and rainfall associated with it over Kyushu, as seen in Fig. 1d and Fig. 4a for 12UTC, 16 July; 12 hours after the initial time of the numerical experiment. It is also of interest to know what are important factors to understand, and predict, the rainfall over Kyushu.

4.1 Comparison among predicted rainfalls

The rainfall intensity at 12 h is shown in Fig. 4b as Case P. The location of the rainfall, over the western part of Kyushu, observed by AMeDAS (the uppermost panel) is simulated to a fair degree, though two peaks in the rainfall intensity are located somewhat east of those from AMeDAS. The rainfall intensity maximum in Case P exceeds 32 mm h$^{-1}$, which is comparable to an observed value (between 20–40 mm h$^{-1}$). In addition to this rainfall, two weak rainfalls are seen over the ocean to the southwest of Kyushu, which are outside the area observed by AMeDAS. (Weak rainfalls are observed to the south of Kyushu (Fig. 2i), though the predicted location is somewhat different.)

For comparison, results from other experiments are presented. Figure 4c shows the rainfall intensity from Y86’s model (Case Y86). Though the intensity is less than 16 mm h$^{-1}$ and weaker, the location of the rainfall is closer to the observed than in Case P. It is not clear whether the present model can simulate the rainfall better than Y86 (though it is obvious that it simulates the cloud water field much better). Figure 4d shows the result from Case NP in which the cumulus-scale effects are not included. The rainfall is much stronger than in Case P and Case Y86. Rainfalls stronger than 32 mm h$^{-1}$ are found over a large area, and a maximum intensity exceeds 64 mm h$^{-1}$. In general, such stronger rainfall is a property that is very often found in the case without the cumulus-scale effects. However, it is noteworthy that the location of the rainfall is not so different from the observed. Its implication will be mentioned later.

The right panels in Fig. 4 reproduce the four results from PT with the use of cloud parameterizations of Arakawa and Schubert (referred to as AS), Kuo and Anthes (KA), Kain and Fritsch (KF) and moist convective adjustment (MA). The above three models simulate the rainfall over Kyushu. A maximum rainfall intensity in AS and KF exceeds 20 mm h$^{-1}$. The most significant difference among the seven
The orientation of the axis of the band-shaped rainfall area. The orientation is west-east in AS and MA, northeast-southwest in KF and northwest-southeast in other four cases. The rainfall pattern from AMeDAS seems to have the northwest-southeast orientation. Radar observation (Fig. 2) clearly supports this feature. As far as the orientation of the rainband is concerned, the results from P, Y86, NP and KA seem to simulate it better than the other three cases. Since the orientation of convection is closely related to the environmental field, it is of interest to discuss, in this regard, the reality of the calculated orientations. This problem will be mentioned later.

4.2 Formation process of cloud clusters

The next concerns are why the rainfall takes place over the western part of Kyushu at 12 h, how we can understand the rainfall from the initial field as shown in Fig. 3, or what factor is important to prediction, or understanding of the rainfall. In the present case, any low is not associated with the rainfall. Examination of the properties of the rainfall (convective or non-convective) and time changes of many fields during the 12 hour period suggests that proper prediction of the eastward propagation of the latently unstable area and the location of the most unstable area is particularly important to the prediction of the rainfall over Kyushu.

Fig. 4. Surface rainfall intensity at 12 hours. (a) AMeDAS, (b) Case P (control case), (c) Case Y86 (d) Case NP (without cumulus-scale parameterization) (e)~(h) four results adapted from Peng and Tsuboki (1997) with parameterizations of (e) Arakawa and Schubert, (f) Kuo and Anthes, (g) Kain and Fritch, and (h) moist convective adjustment.
Figure 5 shows a measure of latent instability, which is given by $c_p\Delta \theta_e(700)$. Hereafter, this quantity is denoted by LI. The left panels show LI at the initial time, 6, 9 and 12 h in Case P. At the initial time, a latently unstable area exists in some part of the East China Sea (also shown in Fig. 3). A maximum of LI is located at 125–126 E, about 500 km to the west-southwest of Kyushu. The unstable area and the location of its maximum move eastward. The maximum value of LI is largest at about 9 h. At 12 h, the maximum is just off the west coast of Kyushu. The unstable area extends to the northern part of Kyushu. Comparison with Fig. 4 indicates that the major rainfall at 12 h is located to the northeast–east of the location of maximum LI, and to the southwest of the northern zero line of LI. This relative location of the rainfall and maximum LI is similar to that of clusters ‘A’ and ‘B’, and maximum LI at the initial time.

In order to understand the rainfall at 12 h in Case NP the predicted LI is shown in the right panels. (A figure at 3 h is included.) As antici-
pated, LI in Case NP is larger than in Case P. (The darkest shade is used differently between the left and right panels.) However, the eastward movements of the unstable area and the location of maximum LI are quite similar for both cases.

It is well known that latent instability (or unstable thermal stratification) is affected by the large-scale flow (including vertical motion), and the fluxes of sensible and latent heat from the sea surface as well as convective activity. The present study suggests that the predicted field of latent instability is not very sensitive to the different schemes of parameterization including a case without the cumulus-scale effects, though the cumulus-scale effects are quantitatively important. This relatively small contribution of the cumulus-scale effects to the pattern and location of latent instability partly explains why the predicted locations of the rainfall at 12 h do not differ much between Case P and NP. However, this happens to hold for 12 h. The behavior of convection, and the process during the 12-hour period are quite different between the two cases.

Figure 6 shows the surface rainfall intensity, which also indicates rainwater mixing ratio $q_r$ near the surface, in Case NP at several selected hours. The result at 12 h shown in Fig. 4 is reproduced, but with different isolines and shading. Several mesoscale rainfall cells (or rainwater cells) are identified at 12 h. A cell in the northwestern portion can be traced back to one at 5 h. The synoptic-scale upward motion that is induced by the initial wind and pressure fields makes the lower atmosphere moister than that shown in Fig. 3. Grid-resolved condensation, and cloud formation which take place before 5 h, are closely related to the initial field and the model physics that does not include the cumulus-scale effects. An ensemble of rainfall cells (cloud system) moves eastward, though the surface flow is southerly in the early stage and southwesterly in the later stage. This is because cloud water is produced successively around the eastern (or southeastern) edge primarily by synoptic-scale convergence associated with the Baiu front.

The formation process of the rainfall cells at 12 h over Kyushu in Case P is quite different from that in Case NP. Figure 7 shows the surface rainfall intensity (rainwater mixing ratio)
at a 1-hour interval for 7~12 h. The rainfall cells at 12 h over Kyushu can be traced back to rainfall cells B1 and B2 at 8 h. Cell B2 forms at 7 h at the southern side of B1 (not seen near the surface but 700 hPa). The location of B2 formation is just at the upstream side of the surface southerly flow (not shown). It grows with time, moving toward the north-northwest. Cell b1', which forms at the eastern edge of cell b1, becomes a part of B2 at 9 h. (Each cell is traced based on figures at a 10-minute interval.) A few peaks appear in B2 after 9 h, and two peaks are pronounced after 11 h. Cell B3 forms at the southern edge of B2 just after 9 h and it is separated from B2 afterwards. Similarly, cell B4 forms at the southern side of B3.

In order to trace back B1 and to see other rainfall cells in the earlier hours, rainwater mixing ratio at 700 hPa is shown in Fig. 8 at an interval of 30 minutes for 3~7 h. The minimum value of the isolines is taken to be 0.03 g kg⁻¹ (0.06 in Figs. 4, 6 and 7). This makes it easier to trace each cell. In the figure for 7 h, cell B2 can be seen in the southern side of B1 (not seen at the surface in Fig. 6). Cell B1 forms at 6 h in the vicinity of the western edge of cell a2. To the north of cell a2 we see cell a1, which can be traced back to a cell located near the west coast of the southern Kyushu at 3 h. Probably cell a1 plays an important role in the formation of B1 and B2. This will be explained later.

At 3 h, several rainfall cells A1, A2, A3 and A4 exist to the west of cell a1. These cells disappear after 5 h. Since the initial condition used for the numerical experiment is not specially devised so as to simulate the observed clusters at early hours of the time integration, behaviors of clusters 'A' and 'B' shown in Fig. 1a and Fig. 2 are not simulated. However, a few cells such as A1, A2 and A3 are formed in an area of clusters 'A' and 'B'. As for cells b1~b5, which begin to form at about 5 h, they form a rainband oriented in the north-south direction. At 6 h, only b3~b5 are high cells that exceed a

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Fig. 7. Surface rainfall intensity at an interval of one hour for 7~12 hours in Case P. Three different shades indicate rainfall intensities stronger than 1, 4, and 16 mm/hour. This figure also indicates rainwater mixing ratio near the surface, and individual rainwater cells are denoted by ai, bi and Bi. The northern boundary of the latently unstable area is indicated by the solid line.
level of 400 hPa. This rainband takes a maximum intensity at 9~10 h (Fig. 7), and it is seen even at 11~12 h to the west of B2 and B4.

Now we compare Figs. 8 and 7 with Fig. 6. At 5 h, when a cell in Case NP exists around (33 N, 128 E), we can see many cells in Case P. Among them, cell A3, which is located about 60 km to the southeast of the cell in Case NP, is the only one in which the grid-resolved relative humidity at 900 hPa level is 100%. In Case P, several cells form already at 3 h in a wide area to the south-southeast of cell A1. (The first cell forms before 1 h.) The solid line crossing Kyushu in Fig. 8 indicates the zero line of LI. It can be seen that all cells are located in the latently unstable area. As known since many years ago, inclusion of parameterization of subgrid-scale convection in the latently unstable area have two different effects; one is to enhance formation and growth of convection which may not occur owing to a coarse grid size, and the other is to suppress unrealistically rapid growth of convection by upward heat transport, in case when convection grows to a

Fig. 8. Rainwater mixing ratio at 700 hPa at an interval of 30 minutes for 3~7 hours in Case P. Isolines are drawn for 0.03, 0.06, 0.12 and 0.22 g kg$^{-1}$. The solid line indicates the northern boundary of the latently unstable area.
certain extent. The latter is understood from comparison of Case NP and Case P in Fig. 4, and the former is from comparison of Fig. 8 with Fig. 6.

As mentioned above, in Case NP, subsequent cells form in the eastern edge of the existing cells, and as a result, a single cloud system consisting of several cells is created. On the other hand, in Case P, cells a1~a3 and cells b1~b5 form to the east of cells A1~A5, and cells B1~B4 (and B1') to the east of cells b1~b5. These can be identified as different cloud systems. The effects of parameterization of cumulus convection are evident.

Our next problem is why a group of cells Bi (i = 1, 2, ...) becomes most intense among several groups including Ai, ai and bi. The most important factor in this regard may be the temperature field near the surface (at the lowest level, 120 m), which is shown in Fig. 9. At 5 h, an area of temperatures lower than 24.0°C extends southeastward to the southern Kyushu, where cell a1 exists (Fig. 8). Since isotherms at the initial time is directed nearly in west-east (not shown), this southeastward extension is due to evaporation of rainwater associated with cell a1 (and partly A1). Since a1 moves eastward and it is at the decaying stage at 5 h, a minimum of temperatures is located to the northwest of a1. Another minimum produced by cell A1 before 4 h is also seen. This cold belt weakens afterwards, but a new minimum is produced at 7 h by cell b1. The formation and growth of cells B1, B1' and B2 appear to be closely related to the presence of this cold area. Although the temperature contrast is not so large as in case of typical squall-lines, its effect should be significant to convective activity. With the growth of these cells, the cold pool becomes pronounced, and the positive feedback makes cell B2 intense and long-lasting.

The formation of cells A1 and a1, which contributes to the formation of the cold belt, is related to the upward motion that is initially of the synoptic-scale in the framework of the model. The synoptic-scale upward motion creates many cells in the latently unstable area. Cells A1 and a1 are situated in the most northeastern portion of the area among these cells. Though strong upward motion is confined inside these cells, weaker upward motion exists in a relatively wide area to their northeast (not shown). Such upward motion field as well as cell a1 moves eastward. Cells bi and Bi form and grow, under the existence of the cold belt, in an area to the southwest of the line of relatively strong upward motion that is located at the southwestern edge of the wide ascending area.

Cells bi are arranged in the north-south direction at the early hours (5~7 h). Generally, each cell is elongated in the west-east direction (Fig. 8). This feature is particularly pronounced in the upward motion and cloud water fields (not shown). This west-east elongation is probably due to the mid-layer westerly flow. As cells grow, the strongest portions of the cells are connected, and a band-shaped convective system oriented in north-south is formed as seen in Fig. 7. A similar feature can be seen for cells Bi (though not distinctly).
4.3 Structure of cluster 'B'

The structure of cluster 'B' obtained in Case P is shown in Fig. 10, which indicates the horizontal distributions of predicted variables at several levels. The selected time is 12 h when cluster 'B' is just entering the mature stage. Figure 11 shows the vertical cross-section on the straight line in Fig. 10h.

Rainwater at 700 hPa (Fig. 10h) corresponds to that at the surface (Fig. 7f). The patterns are quite similar at the two levels. The vertical cross-section of rainwater (Fig. 11a) indicates that rainwater field does not tilt much with height, as expected. The predicted location is close to the observed one shown in Figs. 2i and 4a. Total cloud water $q_c$ at 900 hPa (Fig. 10f)
indicates that cloud water at low-levels extends to the southwest of the rainwater area. This suggests that cloud formation takes place at the southwestern side of the rainfall area. The relative humidity at 900 hPa (Fig. 10c), and its vertical cross-section (Fig. 11b) clearly shows the southwestward extension of the saturated area. The temperature field at the lowest level of the model (Fig. 10g), and the vertical cross-section of temperature perturbation (Fig. 11f) indicate that the coldest area near the surface (called cold pool) is located at the northeastern side (rather than the central portion) of the rainfall area. Relatively warm (and moist) low-level air is located to the southwest of the rainfall area. This is also true for equivalent potential temperature or moist static energy (Fig. 11h). Wind fields at 900 hPa (Fig. 10f) and at the surface (Fig. 10i) suggest that the warm and moist low-level air enters the rainband from its southwestern side.

The wind velocity of the southwesterly flow near the surface (Fig. 10i) is about 9 ms⁻¹, and that of the west-southwesterly~westerly flow in a layer of 900~700 hPa is 12~15 ms⁻¹ (Figs. 10f and 10h). The line-parallel (normal to the rainband, southwesterly) and line-normal (parallel to the rainband, northwesterly) components of the wind velocity are shown in Figs. 11d and 11e, respectively. Under this low-level
flow to the southwest of the rainband, the rainband and mesoscale cells move eastward (or east-northeastward) at speeds of 5–8 m s$^{-1}$ (not shown). These speeds are significantly smaller than those of the low-level air. This supports the above suggestion as to the direction of low-level inflow. Another important feature is that the low-level wind is relatively weak to the northeast of the rainband (Figs. 10f,h,i and 11d). (The wind difference is 6–8 m s$^{-1}$ over a distance of 80 km across the band.) On the contrary, above 500 hPa, the wind is most intense on the northeastern side (not shown). These features indicate that the air flowing from southwest or west converges below 600–700 hPa and ascends inside the cloud. The air diverges above 500 hPa, and the divergent flow (particularly at 400 hPa) strengthens the westerly flow that prevails in the mid-troposphere below 300 hPa. (Above 300 hPa, the flow is northwesterly, and it is not much affected by convective activity.) In conclusion, the wind relative to the rainband movement indicates that the air which supplies the cloud system with water vapor comes from the south (or south-southwest) near the surface and from the west-southwest around a level of 900 hPa. The simulated and observed orientation of the rainband favors energy supply. This problem will be later described in relation to the vertical shear.

One of the important features of the present model is that cumulus-scale cloud water $q^c$ is treated with a prognostic equation. Therefore, not only total cloud water but also cumulus-scale cloud water fields can be discussed. The relative humidity at 700 hPa (Fig. 10b), and cumulus-scale cloud water at the same level (Fig. 10e) indicate that cumulus-scale cloud water is present over a wide area, but a considerable portion of the rainband area over Kyushu is not saturated at 700 hPa. Grid-resolved condensation is localized in two small areas where intense rainfalls take place. In contrast, at 280 hPa, cumulus-scale cloud water (Fig. 10d) can be seen only for an area of the most intense rainfall B2' (Fig. 7f). At this level, grid-resolved condensation and cloud water can be seen over a wide area (Fig. 10a). At 200 hPa (not shown), cumulus-scale cloud water does not exist, and grid-resolved condensation and cloud water have their peaks corresponding to rainfall cells B2 and B2'. Cloud water extends in the south-southeast direction owing to the north-northwesterly flow at this level.

As mentioned in section 3, an observed Baiu front extended through the southern edge of Kyushu at 00UTC, 16 July, and it was moving northward at 12UTC. (It may be more appropriate to say that the Baiu front, which has an orientation from west-northwest to east-southeast, was moving eastward.) Compared with Fig. 3a it can be mentioned that the predicted surface pressure field in Fig. 10i (northward shift of a low pressure axis and also a shear line) is qualitatively in good agreement with observations.

Fig. 11g shows the buoyancy of rising air, calculated as $c_p \Delta \theta_e$. It is seen that the buoyancy in the rainband area is positive up to about 250 hPa (in the absence of entrainment and drag force of water substance). As expected, the latent instability inside the rainband over Kyushu is weaker than in an area to its southwest, but it still remains. The saturated values of $\theta_e$ are shown in Fig. 11i. The values are given by $c_p \theta_e^*$, which corresponds to the saturated value of moist static energy. This figure shows that conditional instability $(\partial \theta_e^*/\partial z < 0)$ still remains even in the rainband area, but it is very weak. As for $\theta_e$ (Fig. 11h), larger values of $\theta_e$ is seen in the mid-troposphere of the rainband area owing to the ascent of low-level air with higher $\theta_e$. Higher $\theta_e$ of low-level air to the southwest of the rainband, and its importance, as mentioned above, can be confirmed by this figure.

The effect of the vertical shear on the orientation and persistence of the cloud band has been studied by a number of researchers these three decades (e.g., Klemp and Wilhelmson 1978; LeMone et al. 1998). It is well known that a squall-line in which the cold pool plays an important role is, in general, a transverse mode. In the present case, the wind field mentioned above indicates that the shear vector below 700 hPa is directed from northwest to southeast. This is more clearly seen from Figs. 11d and 11e. To the southwest of the rainband, vertical shear of the line-parallel component of the wind is very weak below 450 hPa (Fig. 11d). On the contrary, vertical shear of the line-normal component is strong in the whole layer (Fig. 11e). This means that the orientation of
the rainband over Kyushu is nearly longitudinal (parallel to the shear vector). Since the cold belt that exists on the northeastern side of the rainband (Fig. 10g) plays a significant role in the formation and persistence of the rainband, the simulated orientation differs from that expected from the general view on the effect of the vertical shear on convective systems. In this connection, the importance of energy (water vapor) supply from the southwestern side of the rainband as mentioned above should be noted. This problem will be remarked in section 5.

4.4 Cluster ‘B’ and synoptic-scale field

This subsection is supplementary in view of the primary objective of this study. The subsequent evolution of cluster ‘B’ after 12 hours and the predicted mesoscale and synoptic-scale fields at 24 hours are described briefly.

The clusters treated in this study are such that they are not associated with any synoptic-scale low and are formed in a latently unstable area (just to the south of the Baiu front), and in an area which is far from the synoptic-scale low center. Figure 12 shows the surface rainfall intensity at 12, 18 and 24 hours. The figures cover an area much larger than those of Fig. 10 (similar to those of Fig. 3). Fig. 13 shows various fields at 24 hours. It is seen from Fig. 12 that cluster ‘B’ moves eastward and it is located over the sea to the east of Kyushu (south of Shikoku) at 24 hours. The existence of a cluster in this area is demonstrated by satellite imagery (see PT). A very small depression can be found at the surface (Fig. 13i) and 700 hPa (Fig. 13h). The horizontal scale of the low is only about 200 km. No observational evidence of this low is available, but it is likely that convective activity creates a small and weak low in this area.

One of the important features found in Figs. 12 and 13 is that a large area of rainfall exists to the northwest of cluster ‘B’ (Fig. 12). The large cloud system (Fig. 13b~e) is associated with a synoptic-scale low as well as the Baiu front, and it is located to the east~southeast of the low center. Comparison of rainfall area (Fig. 12e) and a latently unstable area (Fig. 13g) indicates that most of the cloud area is latently stable (also absolutely stable). This means that clouds are formed by the large-scale ascent associated with a system of the synoptic-scale low and the Baiu front. Convective clouds and rainfalls are found only in small areas at the southwestern side (and also to the south) of this area. (The cumulus-scale cloud water field is not shown in Fig. 13.) The location of the Baiu front that extends in the east-southeast direction from the synoptic-scale low can be determined from the fields of the surface pressure and low-level vorticity (Fig. 13f). The area of large gradient of latent stability (just to the north~northeast of the unstable area) also sug-
gests the location of the Baiu front. It is appropriate to consider that the Baiu front extends to (34 N, 135 E), corresponding to positive vorticity area. A strong vorticity to the south of Shikoku corresponds to cluster ‘B’. It may be reasonable to infer that this vorticity peak and the cluster are located to the south of the Baiu front.

According to the surface weather map, a low (with the central pressure of 998 hPa) is formed at (35 N, 131 E) on the Baiu front. In the numerical experiment, this low is not simulated (Fig. 13i). This is probably because condensational heating is too weak to produce a low, though a pronounced vorticity peak can be found at (36 N, 130 E). The predicted surface pressure is too high in the southwestern part of the Japan Sea.

The cloud system associated with stable stratification is not discussed in PT. Predicted...
surface rainfall intensities with the four parameterization schemes are shown only for an area south of 35 N. Comparison among the four results from PT, the result from Y86 (Isoya 2000), and the present one indicates that rainfall centered at (34.5 N, 132 E) at 24 hours, which is shown by AMeDAS data in PT, is simulated only by Y86. The present model predicts rainfalls just to its north and west.

It is also shown that there exists a cloud band that extends in the southwest direction in the upper troposphere (Figs. 13a and 13b). This is in agreement with satellite observations (see PT). Corresponding to this, a somewhat moist area extends similarly in the lower troposphere (900 hPa), and a large gradient of latent stability is also found just to the north of this area.

Since the cloud system that is directly associated with the synoptic-scale low is primarily intensified or maintained by non-convective heating (as well as dynamical forcing), a similar result is obtained without cumulus-scale parameterization. In this sense, this cloud system is not the primary concern of the present study. In order to pursue this problem further, more layers of the numerical model may be necessary. However, even when the number of the layer is taken to be 20, the properties of the cloud system are not significantly modified in the present case (except for the upper-tropospheric cloud band that extends in the southwest direction). That is, the formation of a low to the east-southeast of the synoptic-scale low is not simulated. The initial condition and some physical processes parameterized in the model should be responsible for this problem, which remains to be studied.

5. Discussion and remarks

This study is the first attempt to simulate and understand cloud clusters associated with a Baiu front by use of a mesoscale-convection-resolving model (MCRM). Results from only one case study are presented in this paper; cloud clusters observed over Kyushu and the East China Sea on 16 July, 1993 are treated. Comparison with the results from the numerical experiments with the four cloud parameterization schemes by Peng and Tsuboki (1997) indicates that the present model simulates cloud clusters better as far as the location and pattern of the cloud clusters at 12UTC, 16 July are concerned. It is probable that this better result is primarily due to parameterization of subgrid-scale convection (and partly due to the grid size used). A more important difference should be found in behavior (time evolution) of cloud clusters and mesoscale organized convection (MC) during the early 12-hour period of the time integration. Unfortunately, no comparison can be made for the processes because no description is made in PT. Since the four parameterization schemes do not intend to treat MC explicitly and a larger grid size is usually used for these schemes, any comparison of MC (or meso β-scale convective cells) between the models of the two types has little significance. The degree of validity of the present model in this respect can be made only by comparison with observations, and by physical considerations.

In general, it is not easy to evaluate validity of the model by comparison with observations. There are three factors responsible for difficult evaluation. The initial condition used in this study is adopted from the global objective analysis data with a grid size of 15/8 degrees. Even if the synoptic-scale and large-scale fields are given appropriately, mesoscale fields are not necessarily represented. In particular, meso β-scale disturbances are not included in the initial data. The model has to produce these disturbances from the given initial data. Fortunately, the numerical experiments indicate that part of meso α-scale, and even meso β-scale fields, can be simulated from the initial synoptic-scale field when latently unstable stratification and wind (and pressure) fields are given appropriately. However, part of mesoscale fields should not be simulated. The second factor is shortage of detailed observations including information of meso β-scale convection. In general, available data is not sufficient enough for model validity to be evaluated. The third one is the problem of predictability. Even if observations are made to a considerable extent and the model is nearly complete, the time scale of predictability of mesoscale motions should not be long. The time scale is shorter for cloud clusters that are not strongly controlled by dynamical forcing (such as synoptic-scale convergence).

The behaviors and properties of individual cloud clusters and, in particular, those of meso-
scale organized cells that constitute these clusters, as described in Figs. 7~9, are not in agreement with the observed ones in the early period of the time integration. The primary reason is that the initial condition used does not include the mesoscale field corresponding to cloud clusters that already exist. Although it is an important problem to use, as the next step of this study, an initial condition that is closer to observation, the concern of the present study was, as the next step of PT, to see what happens in the MCRM under the given initial condition. The behaviors and properties of cloud clusters and mesoscale organized cells seem to be consistent with those general views (on cloud clusters and meso $\beta$-scale convective cells) that have been constructed from a number of observational (and numerical) studies made for the past five decades. It is probable that there do not exist any behaviors and properties that are apparently unrealistic. This argument, however, does not imply that the model is nearly satisfactory. There is no doubt that the model should have shortcomings. In particular, quantitative aspects of the model are by no means justified. This is also obvious from the fact that the model includes many uncertain assumptions and indefinite parameters. An important point is how we can improve the model by clarifying inappropriate aspects. This problem is beyond the scope of the present study. Any observations that are useful to clarify model shortcomings, and numerical experiments with adequately designed initialization as well as numerical studies with a cumulus-convection-resolving model (CCRM) will be helpful for improvements of the MCRM.

As described in section 4, one of the most distinguished differences of the cloud clusters obtained by the present model and PT is the orientation of the rainband over Kyushu (except for a numerical experiment with Kuo-Anthes scheme). Which is more realistic may be inferred from the past numerical studies (with CCRMs) on the effects of the environmental flow on mesoscale organized convection. However, any general agreement among researchers has not been necessarily made for this problem. Some of the important concepts are not necessarily accepted widely. In order to get a better understanding of the cloud clusters treated in this study, the present author performed numerical experiments on the effects of environmental flow with the use of a CCRM. Numerical experiments indicate that a longitudinal rainband (pararell to the shear vector) is the most preferred mode under low-level shear flow without a jet. This can explain the observed and simulated orientation of the rainband over Kyushu. The numerical experiments also indicate that convective activity creates low-level inflow at the side(s) of the longitudinal rainband, and that this inflow (not the inflow associated with the shear flow) plays a primary role in maintaining the rainband. (These results were presented at a fall meeting of the Meteorological Society of Japan in 2001). The properties of the clusters obtained from the MCRM are consistent with those from the CCRM. Further studies with the CCRM are needed for better understanding of the cloud clusters and MC, and for more appropriate evaluation of the MCRM.

The author’s concerns before performing the numerical experiments were why the cloud cluster and rainfall were observed over Kyushu at 12 UTC, 16 July 1993; 12 hours after the initial time, and what is important to prediction of the rainfall and its understanding. This problem was not discussed in PT. As already mentioned in section 4, the most important factor seems to be an eastward movement of the latently unstable area and its maximum. Although no analysis for clarification of the mechanism of the eastward movement is made, the significance of considering $\Delta \theta_e = \theta_e^p - \theta_e^v$ as an important quantity is emphasized in this study. Recent studies have tended to place their emphasis on $\theta_e$ rather than $\Delta \theta_e$. Larger values of $\theta_e$ in the boundary layer are certainly important to convective activity, but $\Delta \theta_e$ is much more important to discuss this problem. It should be also remarked that convective instability, which is characterized by negative values of $\partial \theta_e / \partial z$, has often been taken to be an important quantity. However, the cloud clusters in this study are not necessarily convection in a stratification destabilized by the large-scale (synoptic-scale) ascent. Therefore, latent instability is more appropriate one to explain the clusters treated in this study.

The degree of latent instability is generally estimated from a value integrated over a layer of positive buoyancy that a rising air possesses.
This corresponds to CAPE (Convective Available Potential Energy) that has been used widely. However, the integrated value is not necessarily a good measure for convective activity. Since the development of a MCRM of Y86, $\Delta \theta_e$ in a layer of 700~1800 hPa has been considered a particularly important quantity (Y86, Y01). This was mentioned in section 2, as seen in Eq. (10). Because of this, the fields of $\Delta \theta_e$ in Figs. 3 and 5 are shown for 700 hPa. This is based on an assumption that convective activity is strongly controlled by low-level buoyancy rather than upper-level one, although there is no doubt that the properties of convection (particularly, cloud top) are also affected by buoyancy at the upper levels, as seen in Eq. (8). It should also be remarked that latent instability is estimated from buoyancy of the air rising from the upper layer of the boundary layer; 900~950 hPa. That is, $\theta_e^B$ is used to estimate $\Delta \theta_e$. The level of the rising air origin chosen in this study is different from the level (850 hPa) which has been usually used.

The horizontal grid size used in this study is somewhat too coarse to describe MC properly. A grid size smaller than 5~10 km is desirable in the MCRM. However, it may be important to mention that even a grid size of about 15 km can describe cloud clusters and MC to such an extent as shown in this paper. There remain many other problems to be discussed. These are left for future studies. Numerical experiments are also needed to apply the model to cloud clusters of various types and to understand the mechanisms. A study for cloud clusters which were found in the GAME/HUBEX observation period in June and July in 1998 is being made.

Acknowledgments

The author wishes to thank Dr. Yasushi Fujiyoshi for his support of this study and Dr. Kozo Ninomiya for reading the original manuscript and making valuable suggestions. He also thanks Dr. Kazuhisa Tsuboki for providing SST data for the numerical experiment, and the anonymous referees and Dr. T. Nakazawa for valuable suggestions. This study was made possible by use of the global analysis data produced by the Japan Meteorological Agency. Information from radar observation was obtained from CD-ROM produced by the Fukuoka District Meteorological Observatory of JMA.

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