Research Article

Toward an Understanding of the Madden-Julian Oscillation: With a Mesoscale-Convection-Resolving Model of 0.2 Degree Grid

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This paper describes results from numerical experiments which have been performed as the author's first step toward a better understanding of the Madden-Julian oscillation (MJO). This study uses the author's mesoscale-convection-resolving model that was developed in the 1980s to improve parametrization schemes of moist convection. Results from numerical experiments by changing the SST anomaly in the warm pool area indicate that the period of the MJO does not monotonously change with increasing SST anomaly. Between the two extreme cases (no anomaly and strong anomaly), there is a regime in which the period varies in a wide range from 20 to 60 days. In the case of no warm pool, eastward-propagating Kelvin waves are dominant, whereas in the case of a strong warm pool, it produces a quasi-stationary convective system (with pronounced time variation). In a certain regime between the two extreme cases, convective activities with two different properties are strongly interacted, and the period of oscillations becomes complicated. The properties and behaviors of large-scale convective system (LCS), synoptic-scale convective system (SCS), mesoscale convective system (MCS), and mesoscale convection (MC), which constitute the hierarchical structure of the MJO, are also examined. It is also shown that cloud clusters, which constitute the SCS (such as super cloud cluster SCC), consist of a few MCS, and a new MCS forms to the west of the existing MCS. The northwesterly and southwesterly low-level flows contribute to this feature. In view of recent emphasis of the importance of the relative humidity above the boundary layer, it is shown that the model can simulate convective processes that moisten the atmosphere, and the importance of latent instability (positive CAPE), which is a necessary condition for the wave-CISK, is emphasized.

1. Introduction

This paper describes results from a study which is made as the author's first step toward a better understanding of the Madden-Julian oscillation (MJO) with a mesoscale-convection-resolving model. As is well known, the MJO was discovered in the early 1970s by Madden and Julian [1, 2], and its understanding has been advanced by a number of observational, theoretical and numerical studies in these 40 years.

The author intends to better understand the following three problems at the early phase in a series of his studies. The first is the period (or time scale) of the MJO. The period was first identified as 40–50 day by Madden and Julian [1, 2]. In the 1980s, a period of 30–50 day was preferably used by Krishnamurti and Subrahmanyam [3], and 30–60 day by Nakazawa [4, 5]. In these 25 years, the latter period has been

very often used by many researchers. Although the period of this intraseasonal oscillation certainly takes a wide range from about 30 days to 60 days, the author prefers to refer to the oscillation as a 40–50 day oscillation and intends to understand why the period is about 40–50 days rather than 30 days. The former period was still used in a review paper by Madden and Julian [6].

The second problem in which the author is interested is the so-called super cloud cluster (or super cluster), which is a synoptic-scale (2,000–4,000 km) cloud system observed in the MJO. In this paper, the two terms super cloud cluster (SCC) and super cluster (SC) are used as having the same meaning (used interchangeably). The term super cluster (SC) was first used by Hayashi and Sumi [7] in their numerical study. Observational evidence of SC was given by Hayashi and Nakazawa [8], and an extensive study of SC was made by

Nakazawa [9], referred to as N88. The SCs moved eastward, and its phase speed was about 15 m s^{-1} in the numerical model of Hayashi and Sumi [7], and $5\text{--}15 \text{ m s}^{-1}$ in the observational study of N88, which used the GMS IR data.¹ In the observational study by Sui and Lau [11], the phase speeds of two observed SCCs were $4\text{--}6 \text{ m s}^{-1}$ in the western Pacific area. They indicated that the SCCs tend to slow down and intensify when they approach the warm pool in the western Pacific. Subsequent studies have confirmed this feature of SCC. The recognition of this feature was an important basis of a numerical study of Oouchi and Yamasaki [12], referred to as OY01. The present study is made along this line.

The second problem is closely related to the first problem. In the case of the OLR data used by Lau and Chan [13], the propagation speed of the most dominant mode of tropical convection (associated with the 40–50 day oscillation) was $4\text{--}5 \text{ m s}^{-1}$ over the equatorial Indian/western Pacific Ocean. Using nine year data of Northern Hemisphere summer, Knutson et al. [14] showed that the phase speeds of the OLR and the upper-level zonal wind were $4\text{--}6 \text{ m s}^{-1}$ in the eastern hemisphere and the latter was 15 m s^{-1} in the western hemisphere. Hendon and Salby [15] also indicated that the phase speeds of the OLR anomaly in the eastern and western hemispheres were 5 m s^{-1} and 10 m s^{-1} , respectively. It has been increasingly recognized that the warm pool in the Indian/western Pacific Ocean plays an important role not only in the propagation speed of SCC but also in the period of the MJO. The important role of the warm pool is one of the major concerns of the present study.

The third problem of the author's interest is how convection behaves in the MJO. Convection associated with the MJO circulation was referred to as "large-scale convection" by Madden and Julian [2]. The observational evidence that synoptic-scale convection (such as SC) is identified in the MJO was given by Hayashi and Nakazawa [8], as mentioned above. Furthermore, N88 showed that SC, which propagates eastward, consists of several mesoscale cloud clusters which move westward. (It was later shown by Lau et al. [16] that there are other two types in the combination of the moving directions of SC and cloud cluster.) The time scale of the cloud cluster is about 2 days (N88; [16]), and that of SCC is 10–15 days [16]. It is equally important to remark that cloud clusters in the tropical atmosphere consist of mesoscale convective cells. It is also important to remark that the basic mode of moist convection has been referred to as cumulus convection. The present author [17, 18] recognized a basic organized form of cumulus convection in his numerical studies of tropical cyclones and tropical convection, and referred to it as mesoscale convection (MC). With the term MC, a cloud cluster can be considered as an ensemble of MC in many cases.

Thus, the hierarchical structure of the MJO can be described in terms of the large-scale convection (LSC), synoptic-scale convection (SSC), mesoscale cloud cluster (MSCC), mesoscale convection (MC), and cumulus convection. Convective systems corresponding to LSC, SSC, and MSCC can be referred to as large-scale convective system (LCS), synoptic-scale convective system (SCS), and mesoscale convective system (MCS), respectively. SCC belongs to SCS. Although

the term MCS has been usually used for an isolated mesoscale system, it can be used as indicating one component of the hierarchical structure of SCS. The author is interested in each of the behaviors of these five classes of convection that constitute the hierarchical structure of the MJO, particularly in the behaviors of SCS, MCS and MC.

This paper describes results from numerical experiments that have been performed with an intention of understanding the three problems listed above. Before describing the results, a review is necessary concerning theoretical and numerical studies in the past.

Intensive theoretical and numerical studies on the MJO started in the middle of the 1980s by Hayashi and Sumi [7] and Lau and Peng [19]. It was suggested that the MJO should be essentially excited and maintained by the mechanism of the so-called wave-CISK (conditional instability of the second kind), and it appears that many subsequent studies have supported this.

The concept of wave-CISK dates back to the late 1960s just before the MJO was discovered. Before this concept emerged, a concept of CISK was proposed in the early 1960s by Ooyama [20] and Charney and Eliassen [21] in their studies of tropical cyclones (TCs). One of linear stability analyses for the problem of TCs was made by Syono and Yamasaki [22]. After nonlinear numerical experiments of TCs by Yamasaki [23–25], the linear stability analysis was applied to wave disturbances in the tropical atmosphere (Yamasaki [26], referred to as Y69) with an intention of understanding easterly waves in the troposphere and a westward propagating large-scale wave in the stratosphere, which was discovered by Yanai and Maruyama [27] and was later identified as the mixed Rossby-gravity wave studied by Matsuno [28]. In Y69, a two-dimensional model was used as the first step, and it was suggested that three types of unstable waves might exist, depending on the vertical profile of the parameterized convective heating,² the so-called beta-effect, vertical shear of the environmental wind, and surface friction. The instability of this type, which is different from CISK applied to TCs, was later referred to as wave-CISK by Lindzen [29].

The two-dimensional linear analysis of Y69 was extended to the three-dimensional (Hayashi [30], referred to as H70; Yamasaki [31], Y71). In the absence of vertical shear of the environmental wind and surface friction, separation of variables can be made in a set of linearized equations (unless the parameterized heating parameter depends on the latitude). In this case, the vertical structure equation, which determines the properties of the instability, takes the same form as that in the two-dimensional model. Therefore, the parameterized heating condition for instability is also the same. The three-dimensional model provides us with the different properties for various types of equatorial waves such as Kelvin wave, mixed Rossby-gravity wave and other waves through the horizontal structure equation. The major concern of Y69, H70 and Y71³ was directed to a planetary-scale stratospheric wave corresponding to the mixed Rossby-gravity wave, but not to Kelvin wave, although Kelvin wave in the stratosphere had been observed [32]. The MJO, which is a tropospheric phenomena and should be an interesting target of wave-CISK⁴ studies, was discovered just after that time.

As described in H70, gravity waves are most preferred, and the growth rate increases with decreasing horizontal scale under the parameterized heating used by Y69. The instability of gravity waves and the preference of small-scale gravity waves were first noted in a numerical experiment of TCs by Syono and Matsuno (unpublished) and later examined by linear stability analysis of Syono and Yamasaki [22]. The unstable gravity waves⁵ were interpreted as unrealistic modes that arose from inappropriate parameterization of moist convection. This interpretation was an important basis for studies of Y69 and Y71, and it was considered that only the linear stability analysis of Rossby wave and mixed Rossby-gravity wave which have low frequency should be informative. As for Kelvin wave, the author felt that the preference of small-scale Kelvin wave (H70) was also a result of inappropriate parameterization because Kelvin wave is essentially similar to gravity waves with respect to the stability property, as evident from the vertical structure in the longitude-height cross section. As for planetary-scale (and large-scale) Kelvin waves with low frequency, the problem remained to be studied.

In addition to the problem of the unrealistic gravity waves, it was strongly recognized in Y69 and Y71 that the vertical profile of the parameterized heating is one of the very important factors to determine the stability properties of various waves and their structure. Recognizing these important problems (including TCs) and intending to get a basis for an appropriate parameterization of moist convection, the author started his studies, in the 1970s, with the use of a cumulus-convection-resolving model⁶ with a horizontal grid size of 500 m–1 km (e.g., [17, 18, 39–42]). Based on the results from such studies, the author developed a TC model [43, 44], which is referred to as mesoscale-convection-resolving model (abbreviated as MCRM). This model was developed to study not only TCs but also other phenomena in which moist convection plays an important role. The development of the MCRM in the 1980s is one of the important basis for the present study.

Now we will return to the review on the numerical studies of the MJO in the 1980s. Hayashi and Sumi [7] performed numerical experiments using the aqua-planet (and zonally homogeneous sea surface temperature) version of a general circulation model (GCM). The most important result is that synoptic-scale convection (such as SC) which moves eastward, and Kelvin-like wave with wavenumber 1 are simulated by the model. The latter means that planetary-scale Kelvin wave is obtained as one of the most preferred modes, which was not predicted from the wave-CISK studies in the early 1970s. On the other hand, the calculated SC may not be necessarily realistic in view of the coarse grid size (T42, triangular truncation at wavenumber 42) and the parameterization scheme used.

Lau and Peng [19] performed numerical experiments with parameterized heating which is similar to that used in the previous wave-CISK studies but without cooling in regions of low-level divergence, positive-only heating parameterization in their terminology. This assumption was previously used in the stability analysis for TCs by Syono and Yamasaki [22] in which the term conditional heating

(as well as unconditional heating) was used. (The terms “conditional” and “unconditional” were also used in Lau et al. [45] and others.) Conditional heating (positive-only heating) assumption alters the vertical motion field such that ascending motion area (convective area) is confined to a relatively small area and descending motion occurs in a much wider area. This contrast of the features for conditional and unconditional heating cases is quite similar to that of moist convection in the conditionally unstable atmosphere and dry convection in the absolutely unstable atmosphere. Keeping this difference in mind, the stability analyses of Y69 and Y71 were made only for the unconditional heating case because the essence of the stability properties can be understood from this case.⁷ One of important results obtained from numerical experiments by Lau and Peng [19] is that the eastward propagating mode is much more enhanced than the westward, as also seen in Hayashi and Sumi [7]. However, only a single ascending area (convective area) with a “relatively small” horizontal scale was obtained. (In the general case, ascending motion is not necessarily confined to a single area.) It is unlikely that the LSC associated with the MJO can be simulated by the parameterization used. In addition, the description of SCC may not be realistic, as in Hayashi and Sumi [7], because of the coarse resolution (R15, rhomboidal truncation at wavenumber 15) and the parameterization scheme used.

The dominance of Kelvin waves with wavenumber 1 was also found in a GCM of Geophysical Fluid Dynamics Laboratory (GFDL) that included realistic land, orography and some other physical processes. N. C. Lau and K. M. Lau [46] showed this feature from a GCM with R15 although SCC was not simulated because of the coarse resolution used. Hayashi and Golder [47] showed that a clear peak of the space-time power spectrum is also found at wavenumber 1⁸ in a GCM with higher resolution of R30. Since these studies used GCMs that have zonally inhomogeneous sea surface temperature (SST) and land-sea distribution, it was not clear that the dominance of wavenumber 1 is realized without zonal inhomogeneity. Other GCM studies with zonal homogeneity such as N. C. Lau et al. [45], Swinbank et al. [48], and Hayashi and Golder [49] have strongly suggested that the Kelvin wave with wavenumber 1 should be one of the most preferred modes that arise from the wave-CISK mechanism and that this mechanism should explain the essence of the MJO.

Now, we proceed to a review in connection with the first and second problems of the author’s interest mentioned above. It is well known that the period of the MJO has been underestimated in most of theories and numerical experiments in the past. For example, the period of the eastward propagating Kelvin wave was about 30 days rather than 40–50 days in Hayashi and Sumi [7]. Although the period probably depends on many factors, some researchers considered a possibility that the shorter period should be due to the parameterization scheme of moist convection. When Hayashi and Sumi’s [7] result emerged, the author considered that the stability analysis of Syono and Yamasaki [22] for gravity waves was informative, because the Kelvin wave is quite similar to the gravity wave with respect to the stability property.

According to the analysis, the phase velocity of the unstable gravity wave is very sensitive to parameterized heating rates (nondimensional parameter used in Y69) in the lower troposphere, particularly, in a layer of 900–800 hPa (or 900–700 hPa). As the low-level heating rates increase, the phase velocity decreases. (When the rates exceed critical values, a gravity wave becomes a stationary unstable wave (CIFK) except for longer waves in which inertial stability is important.) A similar stability analysis was made by Takahashi [50], and the conditional heating case of Syono and Yamasaki [22] was studied by Miyahara [51]. On the other hand, Lau and Peng [19], Sui and Lau [52], and Lau et al. [10] suggested that the propagation speed is sensitive to the level of maximum heating. Tokioka et al. [53] also referred to the level of maximum heating. However, the author has recognized that it is much more appropriate to understand this problem in terms of the low-level heating rather than the level of maximum heating.

Although the period of the MJO can be simulated by artificial modification of the vertical distribution of the parameterized heating, what is important is how the vertical distribution is realized in nature. As mentioned already, the author's study with a cumulus-convection-resolving model (CCRM) in the 1970s and 1980s was a step toward a better understanding of this problem, and the development of the MCRM by the author [43, 44] in the 1980s was based on such a study. Some details of the significance and intention of the MCRM development are described in these papers and Yamasaki [54–56], and it is not repeated here. However, the following remark is important in connection with the parameterized heating used in the wave-CISK studies in the late 1960s–1980s. In the MCRM, it is intended to resolve MC, which is the basic organized form of cumulus convection, and the effects of cumulus convection are incorporated as the subgrid-scale (or parameterized). The heating rate due to cumulus convection is assumed to be of the same form as that used in TC studies of Ooyama [20, 57] and wave-CISK studies of Y69, Y71, and many others (except for the use of the conditional heating). The values of unknown heating parameters are determined so that MC may be realistically simulated. An important point in this respect is that realistic simulation of MC completely prevents unrealistic growth of gravity waves, which was the most serious difficulty in the wave-CISK studies (as well as TC studies). In the MCRM, TCs and other phenomena (such as wave disturbances) can be simulated through description of many ensembles of MC. One of the typical examples of the ensemble of MC is a rainband in TCs. In the case of the MJO, an ensemble of MC is manifested as a cloud cluster, which is a constituent of SCC (SCS in general). It should be emphasized that the development of the MCRM was based on the recognition of the importance of describing MCS (cloud clusters or mesoscale cloud systems including rainbands) through resolving MC. One of the author's major concerns in this study is whether or not the MCRM can simulate the observed period of the MJO and the phase velocity of SCC. Another concern, which is the central topic of the third problem, is how MC and cloud clusters behave in SCCs described by the MCRM.

The MCRM, which was developed in the middle of 1980s, has been used for studies of TC structure [43, 44, 58], TC formation [59–63], TC motion [64], and for studies of cloud clusters associated with Baiu-Meiyu fronts [65–68]. An application of the MCRM to Kelvin wave-CISK was also made.⁹ Because of computer restrictions, its application to the MJO, however, was not made until recently.

Instead, the author started his study of the synoptic-scale and large-scale gravity wave-CISK (as a basis for Kelvin wave-CISK) with the use of a two-dimensional CCRM in the middle of the 1990s [33]. These studies, which were made as extensions of Yamasaki [17, 18], included discussions of the important role of the cold pool and gravity waves (of the small-scale and mesoscale) in the successive formation of MC and cloud clusters, as in CISK of TCs and easterly waves. These studies were extended by Oouchi [69].

It should be mentioned here that the first study of the gravity wave-CISK with a CCRM was made by Nakajima [70]. Although the author [18] studied another type of wave-CISK (corresponding to the easterly wave in the tropical troposphere), whether or not gravity wave-CISK could be simulated by a CCRM remained to be studied. Nakajima [70] was the first to show that the gravity wave-CISK is simulated by a two-dimensional CCRM although correspondence to observed phenomena was not discussed. One of the important concerns of these studies [33, 70] with CCRMs was the successive formation of new cloud clusters to the east of the existing clusters, as observed in SCC studied by N88. In the absence of vertical shear of the environmental wind and low-level wind, both eastward and westward propagations of an envelope of cloud clusters are obtained in the two-dimensional model, whereas an eastward propagation is much more enhanced when the westerly shear [18] or low-level easterly flow [33] exist. In the latter case, the west-east asymmetry of the wind-induced surface heat exchange (WISHE) is one of the important factors, as pointed out by Emanuel [71] and Neelin et al. [72]. It should be again emphasized that the cold pool and gravity waves (of the small-scale and mesoscale) play important roles in successive formation of cloud clusters.

The successive formation of cloud clusters to the east of the existing clusters was also simulated by Chao and Lin [35] with a two-dimensional model that included the parameterized heating in a coarse-grid model. They emphasized the importance of simulating cloud clusters for successful simulation of SCC and the MJO. Needless to say, the model results were very sensitive to the parameterization schemes used. They succeeded in simulating not only successive formation of cloud clusters, but also the slow eastward phase speed of SCC in the presence of the low-level easterly flow. However, description of MCs that constitute each cloud cluster, as done in the MCRM, was beyond the scope of that study partly because the horizontal grid size was taken to be large (about 100 km). Although the author's numerical experiments with the MCRM at that time simulated successive formation of cloud clusters, the results were not submitted for publication, and studies with a CCRM started, as mentioned above. The basic study of Yamasaki [33] and the subsequent study of

Oouchi [69] led the first attempt to study the MJO-like wave with a CCRM by OY01 [12].

The present study is made as an extension of OY01 [12] in which a two-dimensional CCRM was used. In this study, a three-dimensional MCRM is used. Although a review of studies in these 10 years since OY01 [12] may be desirable here, this will be described in the following sections when it is necessary, to avoid a lengthy introduction.

However, the following remark concerning the definition of the term MJO may be necessary to avoid some readers' misunderstanding of the descriptions in this paper. Some researchers have used this term as implying convective activity that occurs primarily over the warm pool area, and a circulation associated with it. Lin et al. [73] defines the MJO as the eastward-propagating mode with periods 30–70 days and zonal wavenumbers 1–6, based on the observational study of Wheeler and Kiladis [74] in which the MJO is distinguished from convectively coupled Kelvin waves (nondispersive) in view of the dispersion relation. The author defines the MJO essentially as an eastward-propagating Kelvin wave whose phase speed is relatively small over the Indian/western Pacific warm pool area and large in the western hemisphere, based on Madden and Julian [2] and many studies in the 1980s and 1990s. The author has also recognized that many convectively coupled Kelvin waves may be dispersive, in contrast to the nondispersive property of the convectively coupled Kelvin waves described by Takayabu [75], Wheeler and Kiladis [74], and Wheeler et al. [76], and neutral Kelvin waves discussed by Matsuno [28]. This recognition is based on a study of Oouchi and Yamasaki [38] and the results (described later) from the author's recent numerical experiments with the MCRM.

In Section 2, brief descriptions of the model used and experimental design are given. In Section 3, the period of the Kelvin waves (MJO) obtained from numerical experiments and the hierarchical structure of convection are discussed. In Section 4, additional discussions concerning wave-CISK and other problems are given. Concluding remarks are given in Section 5.

2. Model

2.1. A Brief Description of the MCRM. In the MCRM, it is intended that MC is resolved by the grid of a numerical model, and the effects of cumulus convection are included as the subgrid scale (or parameterized), as mentioned in Section 1. The horizontal grid size for properly describing MC is, ideally speaking, about 1–5 km. However, it is considered that use of the MCRM is most efficient when the grid size is taken to be 5–20 km. Although a 20-km grid is somewhat too large to describe MC with smaller horizontal scales, MC with large horizontal scales can be described to some extent, and qualitative simulation and understanding of MCS and SCS can be made with this grid size. In this study, a 0.2 degree grid is used for the equatorial region in the spherical coordinate model. When a larger computer becomes available, it can be expected that a 0.1 degree grid will be easily used. Although it is possible to use the latter grid even at the present time, use of the former grid enables us to

make research much more efficient. A number of numerical experiments for TCs and tropical disturbances, which have been performed in these 25 years, have suggested that use of such a coarse grid is, to a fair degree, justified for better understanding of various phenomena.

The original version of the MCRM was developed in the middle of the 1980s [43, 44], as mentioned already. A revised version was developed later [54] with two major improvements. One is that the subgrid-scale cloud water was treated with a diagnostic equation in the original version, as in the most parameterization schemes, whereas it is treated with a prognostic equation. This modification has improved the cloud water field to a considerable extent. Another modification is that the fraction of parameterized (implicitly treated) clouds (cumulus-scale ascending area) is not assumed to be sufficiently small compared to unity, as done in the past parameterization schemes, but assumed to take finite values (such as 0.2). This is because the horizontal grid size in the MCRM is taken to be so small that mesoscale motions such as MC can be simulated. In addition to these two improvements, the determination of the condensation rate in the cumulus-scale ascending area (implicitly treated cloud area) is modified in Yamasaki [65] although its effect is not large.

These MCRMs are hydrostatic models, because most of TC models, GCMs, and numerical weather prediction (NWP) models in the 1980s when the original MCRM was developed were hydrostatic models. Very recently, a nonhydrostatic version has been developed [56]. The ice phase has not been taken into account in the hydrostatic MCRM yet, whereas it is incorporated in the nonhydrostatic MCRM. In the present study, we use the hydrostatic MCRM of Yamasaki [65], because essential discussions of the MJO can be made by the hydrostatic MCRM and because the hydrostatic MCRM is much more efficient (much less computer time) than the nonhydrostatic MCRM.

The model behavior (or performance) of the hydrostatic MCRM of Yamasaki [65] has been described in Yamasaki [58, 62, 63, 66–68] for studies of TCs and cloud clusters associated with Baiu-Meiyu fronts. Results from its application to mesoscale cloud systems over a large island in the equatorial area show a more realistic diurnal variation of rainfalls (presented at the spring meeting of the Meteorological Society of Japan in 2007). The author believes that even the original version of the MCRM gave better results for TCs [33, 43, 44, 59–61, 64] than other models.

One of the important model features that contribute to significant improvements is that cloud water and rainwater mixing ratios are included as prognostic variables despite the fact that it is a hydrostatic model, and the subgrid-scale rainwater is also predicted. The prediction of cloud water and rainwater, which had not been made in other hydrostatic models (hydrostatic TC models and GCMs and NWP models) before the middle of the 1980s, was based on recognition that rainwater evaporation and the resulting cold pool play important roles in successive formation of MC and MCS and, thereby, more realistic behavior of larger-scale disturbances such as TCs. This recognition is just what was obtained from the studies with a CCRM in the 1970s and 1980s. Although the importance of rainwater

TABLE 1: Values of σ at 11-levels where vertical σ -velocities are predicted, the corresponding basic state pressure P_B , basic state temperatures T_B , relative humidity at the center of a moist area R_{H0} , maximum values of the basic state zonal velocity in the middle latitudes, and those of the zonal velocity of a wave given at the initial time.

σ	P_B (hPa)	T_B (K)	R_{H0} (%)	$U_{\max}$ (m s ⁻¹)	$U_{K \max}$ (m s ⁻¹)
0.0	100 (125)	200	50	16	3
0.055	150 (175)	206	50	28	7
0.110	200 (240)	219	55	28	10
0.198	280 (340)	237	60	25	6
0.330	400 (475)	256	65	20	3
0.495	550 (625)	271	70	12	0
0.695	700 (760)	282	75	7	-3
0.791	820 (860)	289	80	4	-6
0.879	900 (927)	293	85	2	-10
0.940	955 (982)	296	83	1	-8
1.0	1010	300	80		

evaporation and the cold pool has been recognized from observational studies, this had not been taken into account in the hydrostatic models before Yamasaki [43, 44].

The importance of convective momentum transport was emphasized by Grabowski and Moncrieff [77]. In the MCRM, the momentum transport by subgrid-scale convection is not taken into account, because a large portion of the momentum transport is accomplished by MC.

The hydrostatic MCRM uses sigma (σ) as the vertical coordinate. A ten-layer model has been used since Yamasaki [64], because the author has believed that qualitative simulation and understanding have been successfully made with only ten layers. It should be mentioned in this respect that the author has used Charney-Phillips grid with respect to the vertical arrangement of the predicted variables. This is based on the author's recognition that a very small vertical grid size (such as 20 m or 2 hPa) is required for Lorenz grid when we study CISK problems with the use of a numerical model that includes parameterization of moist convection. The author recognized this in the 1960s when he performed numerical experiments of TCs with a multiple-layer model [25]. Although the required vertical grid size should be increased by the effects of nonlinear advection and eddy diffusion processes, it should be still very small. There is a possibility that numerical solutions should be largely distorted (owing to computational modes) even if

30 layers are used for the troposphere. On the contrary, qualitatively correct numerical solutions can be obtained for Charney-Phillips grid even when a ten-layer model is used. The values of σ at 11 levels at which the vertical σ -velocities are predicted, and the corresponding basic state pressure P_B are shown in Table 1.

2.2. Experimental Design. The goal of a series of this study is to understand the MJO under the most realistic conditions as observed in the real atmosphere. That is, the land-sea distribution, orography, the diurnal and seasonal changes of solar insolation, seasonal changes of SST as well as the ground temperature, and other factors have to be taken into account at the final stage of this study. In this paper, results from numerical experiments under the most idealized and simplified conditions are presented. Since the behavior and mechanism of the observed MJO are affected by many factors as mentioned above, it is a reasonable first step toward a satisfactory understanding to examine the MJO-like phenomena obtained under simplified conditions. Because of many factors that are not included in the model used as the first step, it is anticipated that there should be many differences (or discrepancies) between the calculated MJO and the observed MJO. The primary objective of this first-step study is to understand what occurs under simplified conditions.

TABLE 2: Sea surface temperature (SST) of the basic state. The SST is linearly interpolated between two latitudes shown. It is taken to be symmetric with respect to the equator.

Latitude	0	5	15	25	30	35	40	45	55	70
SST	301	301	300	299	298	294	288	283	280	280

As is well known, the most important factors for the MJO are the latitudinal variation of SST and the warm pool in the Indian/western Pacific Ocean. In this study, an aqua-planet model (only covered with the sea) is used. Although the latitude of the maximum SST shows seasonal change in nature, we consider the situation such that SST is constant with respect to time and a maximum of SST is located at the equator. The north-south gradient of SST in the middle latitudes is also taken into account. This means that baroclinic instability occurs in the model. The subtropical highs are produced, and the easterlies exist in the equatorial sides of the subtropical highs. Although this easterly flow does not control the MJO directly, it should have some indirect effects. The most important effects of the easterly flow are to enhance convective activity (through the sensible and latent heat flux at the sea surface) and to produce stronger vortices and TCs in this area. The relation of the MJO and TC formation, which has been one of the interesting subjects in MJO studies, will be discussed at the later stage of this series of studies. The easterly flow also contributes to westward movement of vortices and TCs. Under the existence of the subtropical high, TCs tend to move into the middle latitudes, and TCs are removed from the subtropical and equatorial areas effectively. The easterly flow also contributes to frictional convergence in the equatorial area, which provides a favorable condition for convective activity in this area.

The latitudinal distribution of SST used in this study is shown in Table 2. It is taken to be symmetric with respect to the equator. The magnitudes of the latitudinal gradient of SST in the equatorial area and in the middle latitudes are important. If results obtained in this study are qualitatively modified by the use of more realistic distributions of SST, it is worthy to discuss such cases. In this paper, only results obtained from the SST in Table 2 are presented.

A warm pool corresponding to the Indian/western Pacific Ocean is imposed in the following form:

$$T_{\text{sea}} = \begin{cases} T_{\text{warm}} \cos^2\left(\frac{\pi r}{2}\right) & (r < 1) \\ 0 & (r > 1), \end{cases} \quad (1)$$

$$r = \left[\left(\frac{\lambda - \lambda_{T0}}{\Delta\lambda_T} \right)^2 + \left(\frac{\varphi - \varphi_{T0}}{\Delta\varphi_T} \right)^2 \right]^{1/2},$$

where λ is longitude (deg), φ is latitude, T_{sea} is the SST anomaly, and T_{warm} is a model parameter, which is taken to be 0, 1.0, 1.5, and 2.0 K in this study. When T_{warm} is 0, SST is uniform in the longitudinal direction. The horizontal width and shape of the warm pool area are given by two parameters $\Delta\lambda_T$ and $\Delta\varphi_T$. In this study, we take $\Delta\lambda_T = 80$ deg

(or 120 deg) and $\Delta\varphi_T = 20$ deg. The location of the center of the warm pool is taken to be (120E, equator); $\lambda_{T0} = 120$ deg and $\varphi_{T0} = 0$. Although a cold pool exists around or just to the south of the equator in the eastern Pacific, it is not taken into account in this study.

The initial condition is given by the sum of the basic state which does not depend on longitude and the perturbation. The basic state westerly flow (westerly jet) in the middle latitudes in the northern hemisphere is given by

$$U_B(\varphi, p) = \begin{cases} U_{\text{max}}(p) \cos^2 \left[\frac{\pi(\varphi - \varphi_C)/(\varphi_N - \varphi_S)}{2} \right] & (\varphi_N > \varphi > \varphi_S) \\ 0 & (\text{otherwise}), \end{cases} \quad (2)$$

where φ_C is the latitude of the center of the westerly jet and φ_N and φ_S are latitudes of the northern and southern boundaries of the westerly flow, respectively. In this study, we take $\varphi_N = 48^\circ\text{N}$, $\varphi_C = 35^\circ\text{N}$, and $\varphi_S = 22^\circ\text{N}$. In the southern hemisphere, the same westerly flow is given at the same latitudes (symmetric with respect to the equator). Since the SST gradient is imposed, the westerly flow is produced after a long time even if it is not imposed at the initial time. The specification of the westerly flow at the initial time shortens the necessary integration time.

The geopotential, surface pressure and temperature fields at the basic state are determined so that the geostrophic and hydrostatic balances may be satisfied. The values of $U_{\text{max}}(p)$ as well as the basic state temperatures $T_B(p)$ used in this study are given in Table 1. Since the westerly flow changes with time so that it may adjust with the imposed SST, specification of more appropriate values of $U_{\text{max}}(p)$ given at the initial time is not very important.

The initial relative humidities R_H are given in the following form:

$$R_H = \begin{cases} \left[C_{RH} + (1 - C_{RH}) \cos^2\left(\frac{\pi r}{2}\right) \right] R_{H0}(p) & (r < 1) \\ C_{RH} R_{H0}(p) & (r > 1), \end{cases}$$

$$r = \left[((\lambda - \lambda_{R0})\Delta\lambda_R)^2 + \left(\frac{\varphi - \varphi_{R0}}{\Delta\varphi_R} \right)^2 \right]^{1/2}. \quad (3)$$

The relative humidity at each level takes a maximum $R_{H0}(p)$ at $(\lambda_{R0}, \varphi_{R0})$. The values of $R_{H0}(p)$ are shown in Table 1, and λ_{R0} and φ_{R0} are taken to be 180E and 0 (equator), respectively. The width and shape of the moist area are given by $\Delta\lambda_R$ and $\Delta\varphi_R$. We take $\Delta\lambda_R = 50$ deg and $\Delta\varphi_R = 20$ deg. A parameter C_{RH} indicates the ratio of the relative humidity outside the moist area to that at its center. In this study, C_{RH} is taken to be 0.8.

It should be mentioned in connection with Table 1 that the basic state pressures of the lowest and uppermost layers where the potential temperature and mixing ratios of water vapor, cloud water, and rainwater are predicted are taken to be 982–1,010 hPa, and 100–125 hPa, respectively.

In order that large-scale convection may be initiated in the model tropics, a perturbation (zonal-vertical circulation) similar to Kelvin wave is imposed at the initial time. The zonal velocity is given in the following form:

$$U_K(\lambda, \varphi, p) = U_{K\max}(p) \exp\left(-\frac{\beta a^2 \varphi^2}{(2C_g)}\right) \cos\left(\frac{k\pi(\lambda - \lambda_K)}{180}\right), \quad (4)$$

where $U_{K\max}(p)$ is the maximum zonal velocity associated with the wave. The values of $U_{K\max}(p)$ are given in Table 1. The maximum westerly at upper levels and the maximum easterly at low levels are imposed at 0E ($\lambda_K = 0$ deg). Other notations are a : the radius of the earth, β : Rossby parameter ($2 \times 10^{-11} \text{ s}^{-1} \text{ m}^{-1}$), C_g : phase speed of the wave, and k : wavenumber. In this study, we take $k = 1$ and $C_g = 10 \text{ m s}^{-1}$.

The surface pressure and geopotential and potential temperature fields are determined so that the geostrophic balance may be satisfied with respect to the latitudinal direction (in addition to the hydrostatic balance). As for the longitudinal direction, the structure of the wave is similar to that of the eastward-propagating gravity wave. The amplitude of the surface pressure and temperature at 700 hPa corresponding to the imposed $U_{K\max}$ are about 1.0 hPa and 1.0 K, respectively. The center of the initial low-level convergence is located at 90W. The temperature in the lower layer is lowest at 180E, which corresponds to the surface high centered at this longitude. The buoyancy of air rising from the boundary layer is positive (latently unstable) around this area. It can be inferred that two centers of induced convection are found at 90W (initiated by low-level large-scale convergence) and 180E (latent instability) at the very early stage of the time integration. Although the behaviors of convection and other fields which are caused by the initial condition are also interesting, the primary objective of this study is to understand the behaviors of these fields at the later stage when the effect of the artificial initial condition becomes small. Long-time integrations (about 500 days) are made in this study.

The horizontal grid size of the numerical model is taken to be 0.2 deg in the equatorial region (about 20 km), as mentioned already. This grid size covers only an area of 20S–20N. A grid size of 0.6 deg is used for other regions. The northern and southern boundaries are placed at 70N and 70S, respectively, because inclusion of the polar areas is not necessary although the MCRM is designed so that the polar areas can be included. The number of grid points is $1,800 \times 201$ in the equatorial region and 600×81 in each of other two regions. A time increment for the time integration is taken to be 15 sec. The values of other model parameters are taken to be the same as those used in Yamasaki [54, 65] except for the Newtonian cooling rate. The coefficient Q_{RADN} is taken to be somewhat large (0.3 day^{-1}) in the first series of the numerical experiments. The value of 0.2 day^{-1} is also used. As mentioned by Bony and Emanuel [78], theoretical models of the tropical atmosphere had long represented radiative processes as a Newtonian cooling. A model that includes cloud-radiation interaction should be used at the later stage of this series of studies in the future.

TABLE 3: Specification of five numerical experiments which are performed to examine the effects of the warm pool.

Case	T_{warm} (K)	$\Delta\lambda_T$ (deg)	Initial	Integration period (day)
Case (N)	0			0–320
Case (S)	1.0	80	80 Day of case (N)	80–560
Case (M)	1.5	80	160 Day of case (S)	160–640
Case (L)	2.0	80	240 Day of case (M)	240–800
Case (W)	2.0	120	320 Day of case (L)	320–800

In the first portion of the next section, results from five numerical experiments are presented. The specifications of the five experiments are given in Table 3. The SST anomaly corresponding to the warm pool is changed to examine its impact. Other several experiments have also been performed to understand the dependency of the model behavior on some parameters used in the model. Results from two numerical experiments among them are presented in this paper.

3. Results

3.1. The Period of Calculated Kelvin Waves. As mentioned in Section 1, the author's first interest is to understand the period of the MJO. At first, results from case (N) in which SST is uniform in the longitudinal direction are shown. The longitude-time sections (Hovmöller diagrams) for rainwater mixing ratio at the lowest level (or surface rainfall intensity), surface pressure, and zonal wind speed at 925 hPa are shown in Figure 1. These physical quantities are those averaged from 5S to 5N. For the first 15 days, two eastward-propagating peaks of low surface pressure, easterly flow and rainfall can be seen. The western peak (located at 0–90E) is associated with the initially given planetary-scale convergence, which is centered at 90W. The eastern peak is produced by convective activity due to the initial latent instability (positive buoyancy of rising air) and gravity waves that are excited after the initial time. The maxima of the easterly flows are located slightly to the east of the low-pressure centers in these two systems. This feature indicates the structure of an eastward-propagating Kelvin wave which is amplified or maintained against frictional dissipation.

Our major interest is directed to the model behavior at the stage (after 30 days) when the effects of the artificial initial condition become small. The most important result from Figure 1 is that the period of the eastward-propagating wave (with wavenumber 1), which is Kelvin wave, is about 30 days. The phase speed is about 15 m s^{-1} . This period (phase speed) is (or happens to be) very similar to that obtained by Hayashi and Sumi [7]. Although it is important to understand this period obtained by their model and the present model, it remains to be studied (not clarified in this study).

One of other pronounced features seen in Figure 1 is a westward-propagating mode (referred to as WPM) whose phase speed is about $1\text{--}2 \text{ m s}^{-1}$ (200 deg/100–200 days). The WPM modulates convective activity associated with

FIGURE 1: Longitude-time sections (Hovmöller diagrams) of rainwater mixing ratio at the low level (the lowest level of the model, $P_B = 996$ hPa), surface pressure, and zonal velocity at $P_B = 925$ hPa in case (N). Values averaged for 5S–5N are shown.

the Kelvin wave. The surface pressure and the zonal wind are also modulated by WPM and the modulated convective activity. An example of rainfall systems modulated by WPM is indicated by red ellipses (left panel).

The eastward propagation of the Kelvin wave is nearly continuous until about 150 days. Afterwards, new peaks of the easterly wind speed are produced to the east of the existing one. The rainfall systems that contribute to the formation of the new peaks are indicated by two blue ellipses. The formation of these two rainfall systems is closely related to the westward-propagating, low-level westerly and easterly flows (white ellipses) which were excited by convective activity 20–30 days before. Although the phase speed of the Kelvin wave after 120 days is close to 10 m s^{-1} , the formation of these convective systems shortens the period of the oscillation; about 30 days (not 40 days) is also seen after 120 days.

In addition to the WPM, eastward-propagating modes (with similar phase speeds) are also seen (particularly, after 240–360 days, although it is not shown). No physical interpretation for the westward- and eastward-propagating modes can be made in this study, although it is certain that cooling due to rainwater evaporation in the subcloud layer plays an important role.

The second numerical experiment, case (S) is performed with inclusion of some effect of the warm pool, but the maximum of the SST anomaly is taken to be only 1.0 K. This choice is made with an intention of better understanding the results for the case with realistically large anomaly.

Figure 2 shows the longitude-time sections of surface rainfall intensity, surface pressure, and zonal wind speed at 925 hPa. The deviations of the surface pressure and zonal velocity from their time averaged values are also shown. The SST anomaly is imposed after 80 days of case (N) with an abrupt increase of the SST. Comparison of the uppermost portion of the left panel of Figure 2 with the middle portion of the left panel of Figure 1 indicates that the effects of the SST anomaly can be clearly seen after several days and long-lasting rainfalls associated with the warm pool appear, as indicated by the uppermost red ellipse. Somewhat, long-lasting rainfalls are also seen in the warm pool area in a period of 180–310 days.

The patterns of the rainfall and the zonal wind after 320 days are significantly different from those before that time. The time when the maximum easterly flow is located at 0E is indicated by the red arrows. The period is about 25 days in a period of 160–240 days, and afterwards, it takes 50–60 days. In the latter period, the Kelvin wave does not propagate continuously around the globe, but a new peak of the easterly wind speed is produced around 180E (indicated by three blue arrows). The corresponding rainfalls are indicated by three red ellipses in the lower portion of the left panel. The phase speed of many eastward-propagating convective systems is about 15 m s^{-1} , which is similar to that in case (N), but the above-mentioned feature (formation of eastward propagating cloud systems and resulting Kelvin wave) is responsible for the longer periods of 50–60 days. The termination (or weakening) of the eastward-propagating, strong easterly flow is closely related to strong westerly flow (indicated by white

FIGURE 2: Same as Figure 1 but for case (S). Deviations from time-averaged values are also shown for surface pressure and zonal velocity.

circles), which was produced by strong convective activity in the warm pool area 15–20 days before. Since new convective clouds are formed just to the east of the warm pool area, the period is 25–30 days in the western hemisphere, which is contrasted with that (50–60 days) in the eastern hemisphere to the west of about 150E. Although it is of interest to see what happens after 560 days, the time integration has been terminated, because it can be considered that the major objective of the numerical experiment has been achieved; it is suggested that two types of rainfall patterns may occur in case of the weak warm pool, as indicated by those in the upper and lower portions of the figure.

Results for case (M) are shown in Figure 3. Since the SST anomaly (1.5 K) is taken to be larger than in case (S), convective activity in the warm pool area is stronger and more long-lasting, whereas it is much weaker in the western hemisphere. The eastward propagation of convection, surface pressure and low-level wind is not so clear as in case (S); that is, the amplitude of Kelvin wave is much smaller than in case (S) although more distinct anomaly of the surface pressure can be seen in the warm pool area in case (M). The low-pressure peaks are indicated by the red arrows, and the time intervals

are given by numerals. It is important to note that the time interval of the anomaly peaks in case (M) takes a wide range from 20 days to 60 days. The propagation of Kelvin wave is most irregular among the four cases (N), (S), (M), and (L) despite the intensity of the SST anomaly is between those in case (S) and in case (L), which will be shown in the following.

Figure 4 shows results from case (L) in which the maximum of the SST anomaly is taken to be 2.0 K after 240 days of case (M). As expected, convection in the warm pool area is very active and long-lasting. The longitudinal scale of rainfall is also larger. The anomaly of the surface pressure is very pronounced. The strong convective activity produces more notable Kelvin wave, as seen in the zonal wind speed. It may be important to remark again that the amplitude of Kelvin wave is smallest in case (M) among the four cases. In case (N), Kelvin wave is maintained by convective activity which is fairly uniform in the longitudinally uniform SST field. In case (L), Kelvin wave is enhanced by convective activity in the warm pool area, and somewhat maintained by convective activity to the east of about 150E. The eastward-propagating convection is stronger in case (L) than in case (M) owing

FIGURE 3: Same as Figure 2 but for case (M).

to stronger convection in the warm pool area and resulting stronger vertical circulation.

Peaks of negative anomaly of the surface pressure are indicated by the red arrows. The time intervals are 50–60 days and about 80 days. (Inclusion of a weak peak at 430 days indicates addition of 30 and 50 days instead of 80 days.) The phase speed of eastward-propagating easterly anomaly in the western hemisphere ranges from about 15 m s^{-1} to 22 m s^{-1} . The faster phase speed appears to be related to stronger convection in the warm pool area and resulting weaker convection in the western hemisphere.

In order to understand the time interval of the negative anomaly peaks mentioned above, the deviations of the surface pressure and the low-level zonal wind speed from their time averages are also shown in Figure 4 (as well as in Figures 2 and 3). Since the time averaged field indicates the stationary component of disturbances produced by the warm pool (stationary vertical circulation similar to Walker circulation), the subtraction of the time averaged value from the total value makes the propagating component more distinct. As for the propagating easterly anomaly in the western hemisphere, the phase speeds are nearly the same as those mentioned above. On the other hand, propagations in an area of 0–120E can also be seen clearly, but the phase

speeds in this area are very small. The small phase speeds of the low-level easterly anomaly are closely related to the persistence of the low-level westerly flow that contributes to long-lasting convection in the warm pool area. The smallest phase speed is about $2\text{--}3 \text{ m s}^{-1}$, which can produce a long period of even 80 days.

Probably it is correct to say that the phase speed of the Kelvin wave (in terms of the easterly peak) become small by the effects of strong convective activity in the warm pool area, and therefore, the period becomes longer. Some researchers may argue that the long period in case (L) is not related to Kelvin wave, but it is primarily determined by strong convective activity associated with the warm pool. However, as seen in Figure 4, the amplitude of the eastward-propagating Kelvin wave is still large even just to the west of the warm pool area. This suggests the importance of the role of the Kelvin wave in determining the period. Only the concept of the standing oscillation induced by strong convective heating does not appear to be appropriate.

Westward propagations of cloud clusters can be seen in the rainfall pattern. However, the eastward propagation of SCS, which is an ensemble of cloud clusters, is very slow. The long period of Kelvin wave, persistence of convection, and the slow eastward propagation of the SCS in the warm

FIGURE 4: Same as Figure 2 but for case (L).

pool area suggest that it is of interest to perform another numerical experiment by changing the longitudinal scale $\Delta\lambda_T$ of the warm pool.

Figure 5 shows results from case (W) in which $\Delta\lambda_T$ is taken to be larger (120 deg) after 320 days of case (L). It is more important to remark that the longitudinal gradient of SST is smaller in case (W) than in case (L). It appears that the magnitude of the gradient is more important than the horizontal scale of the warm pool area, although the maximum anomaly of SST is also very important. As clearly seen from comparison of Figure 5 with Figure 4, strong convections in the warm pool area in case (W) are separated into two or three groups in the longitudinal direction. This feature is the most notable result in case (W) compared with other cases (L) and (M). Another significant difference can be found in the period of the oscillation, which is slightly shorter in case (W). The period is 50–70 days (50–80 days in case (L)) before 560 days, and it is about 50 days (50–60 days in case (L)) after that time. As expected, the amplitude of the surface pressure in the warm pool area is weaker in case (W) than in case (L).

In the above, the results from the five cases listed in Table 1 have been described to show the effects of the intensity and the size of the warm pool (centered on 120E). An additional numerical experiment has been performed.

The value of Q_{RADN} , a coefficient concerning Newtonian cooling is taken to be 0.2 day^{-1} . This case is referred to as case (R), which should be compared with case (W). Results in case (R) are shown in Figure 6. Qualitatively, the three fields of the surface rainfall, surface pressure, and low-level zonal wind in case (R) are similar to those in case (W). The time intervals of the MJO-scale in case (R) are 40–45 days in many cases and 25–35 days and 60–65 days in some cases. The average time interval is about 40 days, which is shorter than that (about 50 days) in case (W). A more notable difference between the two cases is that small-amplitude oscillations with shorter time intervals (10–30 days), which are seen in case (W), are much suppressed in case (R). This difference can be understood because stronger Newtonian cooling in case (W) acts to suppress convective activity and to produce more unstable stratification in a shorter time, which leads to shorter time scales of convective activity. Although Newtonian cooling is probably assumed to be somewhat too strong in the five cases, it can be inferred that the above-mentioned results concerning the effects of the intensity and the size of the warm pool area are not qualitatively modified for a reasonable value of Newtonian cooling. A more reasonable result should be obtained for a realistic formulation of radiative cooling. The effects of radiation

FIGURE 5: Same as Figure 2 but for case (W).

interacting with moisture (water vapor and clouds), as studied by Grabowski and Moncrieff [79] and Bony and Emanuel [78] for the organization of convection, remain to be studied with the MCRM after better understanding of the model MJO under realistic conditions (such as land-sea distribution).

3.2. Convective Activity in the Warm Pool Area. This subsection describes results concerning the second and third problems among the three listed in Section 1 although the observed phase speed (about 5 m s^{-1}) of super cloud cluster (SCC) is not well simulated. As mentioned in Section 1, SCC belongs to synoptic-scale convective system (SCS). More definitely, SCC is defined as a slow-moving system in the warm pool area in this paper, as in many other papers. As an example, results from case (R) are presented. At first, Hovmöller diagram for the surface rainfall intensity (Figure 6) is reproduced in Figure 7, but only for 480–640 days to make the rainfall pattern much clearer. It can be seen from the figure that most of the eastward-propagating rainfall systems (named A–K), which are somewhat similar to SCC, consist of several westward-propagating rainfall systems which correspond to the mesoscale cloud cluster. The clusters form at a time interval of about 2–4 days, which is somewhat longer than the observed interval of about 2 days. The lifetime

(10–30 days) of the eastward-propagating systems is also longer than that of the observed SCC. The phase speed of the eastward propagation is about 2 m s^{-1} , which is significantly smaller than that of the observed SCC (about 5 m s^{-1}) although it seems to the author that some envelopes of low OLR studied by Weickmann and Khalsa [80] move eastward at speeds of less than 4 m s^{-1} . Physical quantities used in the model and experimental conditions which cause these two features (differences from observations) will be examined in future studies. In this respect, a remark is given here. In a two-dimensional CCRM of OY01 [12] with a 1-km grid, the eastward speed of the SCCs is about $1\text{--}2 \text{ m s}^{-1}$. The physical significance of this similarity should be examined in future studies with an attempt to simulate the observed speed.

In addition to the westward-moving cloud clusters, two cloud clusters that move eastward are seen in Figure 7. These clusters have the property of the squall-line in that its propagation direction is different (opposite) from that of the low-level wind. These are referred to as S1 and S2.

Now, we will see horizontal distributions of the mixing ratio of rainwater at the lowest level of the model (surface rainfall intensity). Figures 8(a) and 8(d) show rainwater fields from 578 day 12 h to 602 day 00 h at a time interval of 12 hours. These figures describe the behaviors (time evolution) of SCC G and H, and squall clusters S1 and S2.

FIGURE 6: Same as Figure 2 but for case (R).

Mesoscale cloud clusters that constitute G and H are named G1, G2, G3, H1, H2, H3, and H4. The westward movement of these seven clusters and the eastward movement of S1 and S2 are clearly seen although S1 begins to move westward after 583 day. An unrealistic feature seen in the figure is that the lifetime of the cloud clusters is too long. It takes a range from 5 days to 10 days (even more). As seen in Figure 7, the lifetimes of cloud clusters in SCC C, E, K, and J are not so long as those in SCC G and H. The lifetimes of the former are 2–5 days.

Although several important features seen in SCC and mesoscale clusters are simulated qualitatively, it has an important problem quantitatively, as mentioned above. The large grid size of about 20 km is certainly responsible for this problem. If a smaller grid size is used, a cloud cluster can be easily replaced by a new cluster that forms in its vicinity. In this case, the time interval of the formation of new clusters and the lifetime of the clusters will be shorter than those shown in Figures 7 and 8. It appears that the formulations of the model and the values of the parameters used are also responsible for these problems. However, it should be remarked that the model (MCRM) has an ability of simulating realistic time scale of cloud clusters (rainbands in the case of TC), as has been seen in these many years. As the next step of this study, the author will seek for other

possibilities: the effects of the SST distribution and the land-sea distribution. The diurnal variation over the land has a strong effect of producing convective systems with a period of 2 days. In the author's idealized numerical experiments of the diurnal variation of rainfall over a large island over the equatorial area, the model could simulate not only realistic phase of maximum rainfall (most intense rainfall time in LST) in the diurnal variation but also convective activity with a period of 2 days (presented at a meeting of the MSJ in 2007). This feature is closely related to consumption of water vapor due to strong convection and its slow recovery due to the surface flux. This has also been discussed by some other authors. It can be argued that the long lifetimes of SCC and cloud clusters obtained in this study are not necessarily due to shortcomings of the model although it is certain that the large grid size is partly responsible for the too long time scale. It can also be expected that the results described in this paper will be useful for better understanding as a basic research and further studies under observed conditions.

Since Figure 8 covers a large area, and it is somewhat hard to see details of the behavior of each cloud cluster, a smaller area is shown for cloud cluster H1 in Figure 9. Compared with Figure 8, details of the rainwater distributions are much more clearly seen although finer resolution in creating Figure 8 would represent clearer distributions. The rainwater

FIGURE 7: Longitude-time section (Hovmöller diagrams) of the low-level rainwater mixing ratio in case (R). It is reproduced from Figure 6, but only a period from 480 to 640 days is shown. Each of synoptic-scale convective systems is labeled by A–K (except I). Squall-line systems are labeled by S1 and S2.

distributions in Figure 9 are shown at a time interval of 2 hours from 588 day 02 h (588:02) to 590 day 00 h (599:00). The zonal scale of cluster H1 is 800–1500 km, which is much larger than that of isolated mesoscale clusters ordinarily observed in the tropics. Although the author describes the hierarchical structure of the MJO in terms of LCS, SCS, MCS (corresponding to the mesoscale cluster), MC, and cumulus convection in Section 1, Figure 9 shows that cluster H1 consists of some MCSs. At 588:02, it can be considered that cluster H1 consists of MCS H1A and H1B. MCS H1A moves eastward, and does not constitute H1 after 589:00. A small cluster, which is referred to as MCS H1C, is located to the west of H1 at 588:06, moves eastward and joins cluster H1. MCS H1D forms around the western edge of H1 and joins MCS H1E, which is located to the west of H1 at the early stage, moves eastward, and constitutes the western portion of cluster H1. After 589:14, cluster H1 consists of two MCSs H1B and H1E, and a large cluster with a zonal scale of more than 1,000 km is seen at this stage.

As mentioned in Section 1, MCS is an organized form of MC. Rainwater mixing ratio peaks of calculated MC correspond to peaks which can be seen in Figure 9. Although the MCRM can be efficiently used for horizontal grid sizes of 5–20 km, the grid size of 20 km used in this study is somewhat too large to properly describe MC. (The most desirable grid size for simulation of MC is 1 km or so.) The lifetime of the calculated MC is too long. Therefore, further discussion

of MC behavior is not made in this paper. An important question is whether the unsatisfactory property of the calculated MC greatly affects the behavior of cloud clusters (MCSs) and SCC. This problem will be studied in the future when a finer resolution can be used. It should be added that the model has an ability of simulating the time scale of MC for a 20-km grid to some extent in cases of TCs and Baiu-Meiyu fronts, as was shown in the author's previous studies. Numerical experiments with other experimental conditions as well as finer resolution may answer the present problem.

The second example of the behavior of cloud clusters is shown in Figure 10 for cluster G3. This cluster consists of three MCSs G3A, G3B, and G3C at 587:02. MCS G3A joins G3B and decays in the eastern portion of G3. MCS G3C joins G3B after 588:06 and decays (or lose its identity) in the western portion of G3. Cluster G3 is a single system of MCS G3B after 588:10–16. It takes a nearly circular shape rather than a band shape after 588:16, and its horizontal scale is about 400 km at this stage.

Figure 11 shows the result for cluster G1 at a time interval of 4 hours. The areas shown in the left and right of Figure 11(b) are different, and also different from that in Figure 11(a). A small MCS G1A at 579:04 grows while it moves southward. It is matured in a period of 580:04–580:16, and afterwards decays rapidly. This MCS constitutes the eastern portion of cluster G1. MCS G1B, which forms around 579:20, begins to lose its identity after 581:00. MCS

FIGURE 8: Continued.

FIGURE 8: Horizontal distribution of the low-level rainwater mixing ratio (g kg^{-1}) or surface rainfall intensity (mm h^{-1}) in case (R) at a time interval of 12 hours: (a) from 578 day 12 hour to 584 day 00 hour, (b) 584 day 12 hour–590 day 00 hour, (c) 590 day 12 hour–596 day 00 hour, and (d) 596 day 12 hour–602 day 00 hour. Each of cloud clusters, which constitute the synoptic-scale convective system, is labeled by F4, G1, G2, ..., and H4. Squall line clusters are labeled by S1 and S2. The ordinate ranges from 8S to 8N.

(a) rain water

(b)

FIGURE 9: Horizontal distribution of the low-level rainwater mixing ratio in case (R) at a time interval of 2 hours: (a) from 588 day 02 hour to 589 day 00 hour and (b) 589 day 02 hour–590 day 00 hour. Each of mesoscale convective systems, which constitute cloud cluster H1, is labeled by H1A, H1B, . . . , and H1E.

(a) rain water

(b)

FIGURE 10: Horizontal distribution of the low-level rainwater mixing ratio in case (R) at a time interval of 2 hours; (a) from 587 day 02 hour to 588 day 00 hour, and (b) 588 day 02 hour–589 day 00 hour. Each of mesoscale convective systems, which constitute cloud cluster G3, is labeled by G3A, G3B, and G3C. The ordinate ranges from 1S to 5N.

(a) rain water

(b)

FIGURE 11: Horizontal distribution of the low-level rainwater mixing ratio in case (R) at a time interval of 4 hours; (a) from 579 day 04 hour to 581 day 00 hour, and (b) 581 day 04 hour–583 day 00 hour. Each of mesoscale convective systems, which constitute cloud cluster G1, is labeled by G1A, G1B, G1C, and G1D.

FIGURE 12: Horizontal distribution of the low-level rainwater mixing ratio (lower panel), and the upper-level (200 hPa) cloud water mixing ratio (upper panel) at 580 day in case (R). Wind vectors are also shown. Cloud clusters F4, G1, and S1 are labeled.

G1C begins to grow just before 580:16. MCS G1D forms at almost the same time as G1C to its west (around 107E, outside the area shown). It is a single MCS that constitutes cluster G1 after 582:12. The horizontal scale is about 400 km at this stage. This case shown in Figure 11 is one of the most typical examples that new MCS forms to the west of the old one.

The author has considered that a typical organized form of MC is MCS that has horizontal scales of 200–500 km, which correspond to the smaller portion of the so-called meso- α -scale. This has been increasingly recognized in the author's studies of TCs, and cloud clusters associated with tropical disturbances and Baiu-Meiyu fronts. As in TCs and other tropical disturbances, the present numerical experiments indicate that two or more MCSs very often constitute a large (larger portion of the meso- α -scale) cloud cluster in the case of SCC (generally, SCS).

Our next concern is why new MCS tends to form and grow to the west of old MCS in the case of cloud clusters which constitute SCC although a new cloud cluster tends to form to the east of old one (eastward propagation of SCC as an envelope of cloud clusters). In order to understand this problem, the low-level wind field as well as the rainwater

mixing ratio near the surface at 580 day is shown in the lower portion of Figure 12. Clusters F4, G1, and S1 are seen. This rainwater field corresponds to that shown for a smaller area in the left of Figures 8(a) and 11(a). At this stage, cluster G1 consists of MCS G1A and G1B. It can be seen that westerly, northwesterly, and southwesterly flows contribute to cluster G1. The vertical shear in the lower troposphere is easterly shear (vertical profile: not shown), and this contributes to the formation and growth of new convection to the west of existing convection, which has been known in these many years (since the 1970s). The air of the northwesterly flow comes, in many cases, from the northern area where easterly flow prevails.

The upper portion of Figure 12 shows the cloud water (cloud ice in nature) mixing ratio and the wind field at 200 hPa. As a matter of course, cloud water produced by convection is advected by the upper-level outflow. In this figure, southwesterly~westerly and northeasterly~northerly flows are pronounced.

Another example is shown in Figure 13. The selected time is 588 day when four clusters G2, G3, H1, and H2 exist. The rainwater field corresponds to that shown in the right of Figure 8(b). Cluster H1 consists of MCS H1A and H1B

FIGURE 13: Same as Figure 12 but at 588 day. Cloud clusters G2, G3, H1, and H2 are labeled.

(uppermost left of Figure 9(a); nearly the same time), and cluster G3 consists of MCS G3A, G3B, and G3C (lowest right of Figure 10(a)). In the lower panel of Figure 13 for the low-level field, it can be clearly seen that the air in the northern area has easterly component, it turns its direction, and it has northwesterly component when it enters cluster H1. As in other cases, southwesterly flow from the southern hemisphere also contributes to cluster H1 (also, G2 and G3). In the upper troposphere, outflow is seen with pronounced southerly and northerly components, and the zonal flow does not have pronounced westerly and easterly components in most of the area shown in this figure.

4. Additional Discussion

4.1. Wave-CISK. As mentioned in Section 1, the author has considered the term wave-CISK as an instability implying cooperative interaction between moist convection (of various types) and a large-scale (including synoptic-scale and planetary-scale) wave. It appears that inappropriate descriptions concerning wave-CISK, which might lead to misunderstanding, have often been made, depending on how one defines it (e.g., Chao [34]). The author's present understanding of wave-CISK is common with that envisaged

at the early 1970s in some respects and significantly different from that in other respects. As Ooyama [81] stated, one should view wave-CISK (CISK in general) in terms of the conceptual content that has grown and matured with advances in research.

The discharge-recharge mechanism was proposed by Blade and Hartmann [82]. This mechanism appears to correspond to consumption (due to convection) and recovery (due to the flux at the sea surface) of water vapor, which have been considered essential to wave-CISK. For instance, the problem of the recovery of water vapor and its time scale was discussed by Ooyama [81]. Kemball-Cook and Weare [83] discussed the importance of building and discharge of the low-level moist static energy in determining the period of the MJO. (However, they also mentioned that it is not necessary to invoke large-scale wave motions to explain the observed oscillation of convection.) Benedict and Randall [84] discussed the importance of this mechanism in terms of low-level moistening and heating by shallow convection. As mentioned above, the mechanism included in these discussions can be considered as one of important components of wave-CISK. The time evolution of the vertical profiles of the moisture and heating, which should be one of interesting results, is not shown in this paper but remains to be reported.

The WISHE mechanism was proposed by Emanuel [71] and Neelin et al. [72], as mentioned in Section 1. Hayashi and Golder [85] argued that intraseasonal oscillations are maintained primarily through the evaporation-wind feedback (EWF: similar to WISHE) mechanism. The present author has considered that the dependence of surface heat and moisture fluxes on the wind speed is one of the important components of CISK (including wave-CISK). This has been important basis for TC studies since the 1960s. This was clearly shown by Ooyama [57]. More surface fluxes in stronger wind areas, which are the general property of the turbulent process in the boundary layer, are favorable or important to CISK in that water vapor consumed by convective activity can be compensated by more surface fluxes due to stronger wind that has been produced by convective activity.

Another aspect of WISHE effects is that a wave due to wave-CISK tends to propagate eastward (westward) in the presence of the environmental, low-level easterly (westerly) flow because of the west-east contrast of the surface flux as a result of superposition of the environmental flow and the flow associated with the wave, as was suggested by Emanuel [71] and Neelin et al. [72]. A study of OY01 [12] with a two-dimensional CCRM was based on this effect of WISHE. However, in the case when the Coriolis parameter depends on latitude, as in nature and also in the present study, the eastward propagation of the planetary-scale wave is primarily caused by this effect, as manifested as Kelvin waves (dominance of Kelvin waves in the usual case). The WISHE may play a significant role but not important to wave propagation.

Another problem related to wave-CISK is the role of surface friction (friction between the atmosphere and the sea surface). An instability that arises when surface friction is included in a model has often been referred to as frictional wave-CISK. The instability of this type was found in linear analysis made in a TC study of Syono and Yamasaki [22], which showed the instability of gravity waves of two types with and without surface friction. The Kelvin-wave CISK as well as gravity wave-CISK was studied by Hayashi [86], and later by Wang [37] and others (e.g., Wang and Rui [87]; Xie and Kubokawa [88]; Wang and Li [89]; Salby et al. [90]; Wang and Schlesinger [91]; Kemball-Cook and Weare [83]). This problem was also discussed by Oouchi and Yamasaki [38]. One of the most important results from the latter is that Kelvin wave induced by convective heating in the presence of surface friction is dispersive, whereas neutral Kelvin wave is nondispersive Matsuno [28].

Although the role of surface friction in wave-CISK may be understood to a fairly degree as far as the results from linear stability analyses are concerned, our important question should be directed to the role of surface friction in a nonlinear model (and in nature). The author has tried to understand this problem to some extent. However, no definite answer has been obtained. In the case of the numerical experiments in this study, surface friction strongly affects the environmental flow such as not only subtropical easterlies associated with mid-latitude baroclinic waves but also the flow in the equatorial area in which Kelvin waves

are embedded. Without surface friction, the flow is very different from that obtained in the numerical experiments. Only the direct effects of surface friction on Kelvin waves have to be extracted, keeping the environmental flow. Only idealized numerical experiments have been performed, without environmental flow, for various zonal domains with the cyclic condition. The numerical experiments have confirmed the dispersive property of Kelvin waves which was suggested by Oouchi and Yamasaki [38]. That is, in the case when the speed of the eastward propagation is 22 m s^{-1} for a zonal wavelength of 40,000 km, the speeds are 9 m s^{-1} and 6 m s^{-1} for zonal wavelengths of 10,000 km and 5,000 km, respectively. Although these speeds are modified according to experimental conditions, it can be suggested that the dispersive property would not be changed. Only this role of surface friction is what the author can suggest at the present stage. It is certain that surface friction plays a significant (favorable) role in convective activity in the equatorial area through frictional convergence because subtropical easterlies are usually present. However, direct effects (other than dispersive property) of surface friction on Kelvin waves remain to be studied in the future.

The vertical profile of convective heating is very important, particularly in the case of models in which the heating directly controls the planetary-scale (or synoptic-scale) wave, as mentioned in Section 1. In this connection, our concern is to what extent the heating in the upper stratiform clouds (stratiform heating) and cooling associated with stratiform precipitation play important roles in Kelvin wave-CISK. The author has not recognized yet the importance of the stratiform precipitation that may produce the top-heavy heating profile (e.g., Lin et al. [92]) as far as the instability (not structure) of Kelvin waves is concerned. In addition, the author has not understood that stratiform instability Mapes [93] and moisture-stratiform instability Kuang [94] play important roles in the MJO.

4.2. Tropospheric Humidity and Latent Instability. Wave-CISK can occur only in the presence of latent instability, which is characterized by positive CAPE (convective available potential energy). In recent years, some researchers have discussed the importance of moistening above the boundary layer in the MJO (e.g., Maloney and Hartmann [95]; Maloney [96]; Bony and Emanuel [78]). The sensitivity of moist convection to mid-tropospheric humidity was also discussed (e.g., Derbyshire et al., [97]). The importance of the humidity in convective activity and rainfall was studied using observational data (e.g., Bretherton et al., [98]). Zhu et al. [99] suggested that precipitation is stronger as the column integrated relative humidity increases and that it should be an exponentially increasing function of the column saturation fraction to better simulate the MJO. The importance of the humidity in convective activity has been well known, at least, since the 1960s. What is important in the discussion of wave-CISK (CISK in general) is how the tropospheric relative humidity behaves as a result of interaction of moist convection and larger-scale motions. In numerical models with a coarse-grid, it depends on how

the effects of moist convection are treated (or parameterized). Some studies have imposed a relative humidity threshold for parameterized convection in GCMs to get better results (e.g., Wang and Schlesinger [91]; Zhang and Mu [100]). Thayer-Calder and Randall [101] suggested the importance of convective moistening; that is, a model has to realistically represent convective processes that moisten the entire atmosphere in order to simulate the MJO.

In the MCRM of Yamasaki [43, 54], active convective region is more humid as a result of the combined effects of the mesoscale (and large-scale) ascending motion and convective activity. The author has seen such a feature in tropical cyclones and Baiu-Meiyu fronts computed with the MCRM in these 25 years. This is also true for the present numerical experiments of the MJO.

The right panels of Figures 14(a) and 14(b) show the Hovmöller diagrams of the relative humidity at 700 hPa and at 900 hPa in the same period as that shown in Figure 7. The mixing ratio of cloud water (cloud ice in nature) at 200 hPa is shown in the left panel of Figure 14(a). These quantities are those averaged for 3S–3N (not 5S–5N). The mixing ratio of low-level rainwater, which corresponds to Figure 7, is shown in the left of Figures 14(c) (averaged for 3S–3N). For convenience, the surface pressure and the zonal velocity at 925 hPa (Figure 14(d)) are reproduced from part of Figure 6 (averaged for 5S–5N).

It can be seen from these figures that active convective and rainfall areas are generally more humid than other areas. The upper troposphere is also more humid (not shown). This result means that the MCRM satisfies the suggestion by Thayer-Calder and Randall [101]. As mentioned above, this feature has been a general property of the moisture field from the MCRM in these 25 years.

Another important feature is that the area to the east of the warm pool area is more humid than the warm pool area at 900 hPa, whereas the former is drier than the latter at 700 hPa although it is somewhat too dry in the model. This feature means that downward motion associated with planetary-scale circulation makes the atmosphere drier at 700 hPa outside the warm pool area and that more shallow clouds exist to the east of the warm pool area (not shown). The warm pool area at 900 hPa is less humid than the area outside it because of compensating downward motion due to strong convection in the warm pool area.

The left panel of Figure 14(b) shows the Hovmöller diagram of $B(700)$, a measure of the buoyancy which the air rising from the boundary layer acquires at 700 hPa. Although the vertical profile of the buoyancy is also important, the author has presented this quantity in his papers when only one figure concerning the buoyancy is shown. This is because he has considered that this quantity should be most important. In other words, the author has not considered CAPE as the most important quantity. The formulation of the effects of subgrid-scale cumulus convection in the MCRM is based on this consideration (Yamasaki [43, 54]). Since latent instability (positive CAPE) is one of the necessary conditions for convection, $B(700)$ is positive in almost all areas of rainfall in the mesoscale sense, as seen in the figures. Needless to say, more humid air in the upper

portion (around 925 hPa) of the boundary layer and lower temperature air at 700 hPa contribute to positive $B(700)$. The right panel of Figure 14(c) shows the temperature anomaly at 980 hPa. The importance of the cold pool associated with active convection can be confirmed.

Raymond and Fuchs [102, 103] explored the hypothesis that the MJO is driven by “moisture mode” instability. This term was derived from their view on numerical studies with the past parameterization schemes that did not incorporate the effects of moist convection appropriately. It seems to the author that this is not a new type of instability, but it corresponds to part of CISK (such as stationary CISK and slowly moving Kelvin wave-CISK). Some of their results and the present results concerning the behavior of the MJO over the warm pool area appear to have some common properties although the models used and the terms for instability are different.

Although some authors have not mentioned the importance of latent instability (positive CAPE) but rather emphasized the importance of relative humidity in recent years, the present author has considered that it is most important to discuss wave-CISK problems in terms of latent instability. Latent instability is the most important and necessary condition for wave-CISK. The author has emphasized it in his previous papers. In the MCRM, not relative humidity but the positive buoyancy of the air that rises from the boundary layer is one of the most important quantities for subgrid-scale convective heating although relative humidity significantly affects the buoyancy of the rising air.

4.3. Other Problems. One of the important concerns in the MJO is the relative dominance of eastward-propagating Kelvin waves and quasi-stationary component that may usually behave like a localized standing oscillation in the warm pool area. It can be argued that the former is enhanced or excited by the latter although the latter is strongly affected by the former (more specifically, by the low-level wind associated with the former). Some features of the localized standing oscillation and the relative dominance were described by Zhang and Hendon [104]. As also shown in the present numerical experiments, the relative dominance is determined primarily by the SST anomaly and its gradient (Figures 1–6). It appears that the stationary component (or standing oscillation) is too pronounced in cases (L) and (W) in which the maximum SST anomaly is 2 K. One example that the intraseasonal oscillation is dominated by strong stationary oscillation was shown by Hsu et al. [105]. They also showed that the upper tropospheric circulation did not complete the cycle around the globe. This is different from the result of many other studies (e.g., Knutson and Weickmann [106]). It is remarked that the dominance of the standing oscillation may indicate that the zero wavenumber is dominant. The latter is not pronounced in the wind field but the pressure (geopotential) and temperature fields, as was mentioned by Itoh and Nishi [107]. The temperature (not shown) and surface pressure fields in cases (L), (W), and (R) clearly show this feature.

Closely related to the above problem, it is also our important concern to clarify whether eastward-propagating

FIGURE 14: Continued.

FIGURE 14: (a) Longitude-time sections of the upper-level (200 hPa) cloud water mixing ratio, and relative humidity at 700 hPa, (b) a measure of buoyancy (in unit of temperature) which the air rising from the boundary layer (about 750 m height) acquires at 700 hPa, and relative humidity at 900 hPa, (c) the low-level rainwater mixing ratio, and temperature anomaly at the lowest level of the model ($P_B = 996$ hPa), and (d) surface pressure and the zonal velocity at 925 hPa in case (R). The period corresponds to that in Figure 7, but the values are averaged for 3S-3N.

Kelvin waves strongly affect the standing oscillation in the warm pool area. Hu and Randall [108] argued that dynamically induced convection is not needed to explain the observed oscillation and that it is a side effect. They also suggested that the long time scale of radiative cooling is an important factor for the low-frequency oscillation. The present author considers that the time scale of recovery of water vapor field due to surface flux which compensates consumed water vapor is much more important than that of radiative cooling and that the eastward-propagating wave (as well as surface flux) plays an important role to determine the time scale of the oscillation in the warm pool area. In order to answer this problem clearly, it may be the best way to perform numerical experiments by adopting a sufficiently wide zonal area instead of about 40,000 km, or a condition in which convection outside the warm pool area is suppressed.

A remark is made for the relation between low-level rainwater (surface rainfall) and upper-level cloud water fields. The author has usually examined (or has been interested in) the low-level rainwater field at first when a numerical experiment is performed, because he has considered that description of low-level rainwater is much more important for better understanding (not for comparison with observations) than that of the upper-level cloud water (cloud ice in nature) in many problems. However, we are also concerned with the upper-level cloud water field, particularly because most of observational studies have described results from satellite data, and because better simulation of the upper-level cloud water field is important to the radiation budget in future studies. Our concern here is whether any important difference between distributions and behaviors of low-level rainwater and upper-level cloud water exists. Comparison of the Hovmöller diagrams of the two fields (left panels of Figures 14(a) and 14(c)) indicate that these

are essentially similar although upper-level clouds are more widely spread, as is well known. An eastward-propagation speed of about 5 m s^{-1} is clearly seen in both fields (averaged for 3S-3N). This is not clear in Figure 7 (averaged for 5S-5N). This speed corresponds to the observed phase speed of SCS (including SCC) which many authors have obtained from satellite data. Since the present author considers that low-level rainwater field is much more important than the upper-level cloud water for better understanding of wave-CISK, it is hoped that much more observational data of low-level rainwater will be obtained in the future. It is important to clarify which aspects of the results obtained in this study will be validated or invalidated from such observations.

The final numerical experiment presented in this paper is performed with a maximum SST anomaly of 1.5 K, which is the same as that in case (M). Other conditions are taken to be the same as those in case (R). The longitude-time sections corresponding to Figures 3 (case M) and 6 (case R) are shown in Figure 15(a), but the zonal velocity at 175 hPa is shown instead of the surface pressure deviation. The variability of the time interval of the minimum surface pressure is notable, as in case (M). The time interval takes a range from 20 days to 55 days. On the other hand, the low-level zonal velocity has only six major propagations in the eastward direction. This is somewhat in contrast to more propagations (shorter time scales) in cases (M) and (R). This is primarily due to the combined effects of smaller gradient of the SST and weaker anomaly. As is well known, the upper-level zonal flow is nearly out of phase with the low-level flow. Very strong westerly flow exceeding 20 m s^{-1} can be seen at the upper-level. The low-level westerly flow and the upper-level easterly flow prevail over the active convection and rainfall in the warm pool area, as has been usually observed. Relatively slow propagation of the zonal wind speed around the warm

FIGURE 15: Continued.

FIGURE 15: (a) Same as Figure 2 but for case (MWR). The zonal velocity at 175 hPa is shown instead of the surface pressure deviation, (b) upper-level cloud water, a measure of buoyancy (in unit of temperature) at 700 hPa, relative humidity at 700 hPa and 900 hPa, and temperature anomaly at the lowest level of the model ($P_B = 996$ hPa), and (c) the upper-level and low-level meridional velocities averaged for 5S–5N and 10N–15N are shown.

pool area and the relatively fast propagation in the western hemisphere can also be seen, as in other two cases. The surface pressure field, which has large amplitude of zero-wavenumber, exhibits much faster eastward propagation.

Figure 15(b) shows the upper-level cloud water, a measure of latent instability $B(700 \text{ hPa})$, and relative humidities at 700 hPa and 900 hPa, corresponding to Figures 14(a) and 14(b). The descriptions made for case (R) are also valid in this case qualitatively. The major convection area over the warm pool is more humid at 700 hPa (also in the upper troposphere, not shown). At 900 hPa, it is more humid to the east of the warm pool area. The temperature near the surface (right panel) is lower in active convective area owing to evaporative cooling, and higher in the cloud-free area, where adiabatic compression due to downward motions occurs, as also seen in Figure 14(c).

Figure 15(c) shows the meridional component of the wind at the low and upper levels. The two different values averaged for 5S–5N and 10N–15N are shown. The most pronounced feature seen in these figures is that the synoptic-scale waves propagate westward at the low level. Superimposed on this, eastward propagations of the envelope of westward propagating and somewhat stagnant modes can also be seen. These propagation speeds are significantly slow compared with that of the zonal wind. At 10–15N, strong vortices (TCs) are often seen (not shown), as suggested from strong (even for the 10–15N average) northerly flow just

to the west of southerly flow. At the upper-level, nearly stationary modes prevail. The phase speed of eastward propagations to the east of the warm pool area is only about 1 m s^{-1} . The physical mechanism remains to be studied. This final experiment is a step toward numerical experiments under more realistic (observed) conditions, which will be reported in the future.

5. Concluding Remarks

This paper describes the results from numerical experiments which have been performed as the author's first step toward a better understanding of the MJO. One of the main features of this study is that it uses the author's mesoscale-convection-resolving model (MCRM) which has been used, in these 25 years, for his studies on several phenomena such as tropical cyclones (including the formation process) and cloud clusters associated with Baiu-Meiyu fronts. A somewhat large grid size of about 20 km is used for more efficient research, although a grid size of 10 km or 5 km is more desirable. One of the primary objectives of this study was to examine to what degree the MCRM can describe the properties (behavior) of the observed MJO (and large-scale convective system), SCS (synoptic-scale convective system including SCC), and MCS (mesoscale convective system) and MC (mesoscale convection; a basic organized form of cumulus convection). Another feature of this study is that the author intends to

understand the observed MJO by considering simplified and idealized experimental conditions as the first step. Therefore, good simulation of the MJO is somewhat beyond the scope of this study. The most important points in this study are to understand what happens in the model and to infer what conditions are important to the observed MJO. This study suggests that numerical experiments should be performed by taking account of the land-ocean distribution as the next step of this study.

One of the results which the author had not necessarily expected (or inferred) before performing the numerical experiments is that the period of the MJO does not monotonously change with increasing SST anomaly in the warm pool area. Between the two extreme cases (uniform SST in the longitudinal direction and large SST anomaly corresponding to the Indian/western Pacific Ocean), there is a regime in which the period varies in a wide range from 20 to 60 days. In the case of longitudinally uniform SST, eastward-propagating Kelvin waves are dominant, whereas in the case of a strong warm pool, a quasi-stationary convective system (with a pronounced time variation; standing oscillation) is formed in the warm pool area, and it strongly enhances Kelvin waves that propagate eastward around the globe. In a certain regime between the two extreme cases, convective activities with two different properties coexist, and these are strongly interacted. Therefore, the period of oscillations becomes complicated.

Another notable result from the numerical experiments is that mesoscale cloud clusters, which constitute SCS (including SCC), very often consists of two or three meso-scale convective systems (MCSs), each of which has the meso- α -scale of the smaller portion, and that a new MCS tends to form to the west of the existing MCS. The northwesterly and southwesterly low-level flows of the air, the origin of which is the air in the easterlies on the equatorial side of the subtropical highs, contribute to this feature.

The most notable difference of the model results from observations is that the lifetimes of many MCSs, cloud clusters, and SCCs are too long. Whether this is inevitable in the case of a 20-km grid and whether the experimental conditions used in this study are responsible remain to be studied. As mentioned in Section 3, the author's idealized numerical experiments of the diurnal variation of rainfall over a large island over the equatorial area showed that the MCRM could simulate convective activity with a period of 2 days. Some authors have also shown that the diurnal variation over the land has a strong effect of producing convective activity with a period of 2 days. Inclusion of the land-sea distribution as the next step of this study will be important, particularly with respect to the lifetimes of MCSs, cloud clusters, and SCCs (in addition to distributions of tropical cyclones).

In this connection, a remark is made here. Although some authors (e.g., [109, 110]) interpreted the 2-day period cloud cluster in SCC as a westward-propagating inertio-gravity wave, the author has interpreted the cloud cluster as an organized form of MCs, although gravity waves (of small scale and mesoscale) contribute to the organization of convection; the cluster is not an inertio-gravity wave. As for the importance of SCC in the MJO, Hendon and Liebmann

[111] stated that SCC does not appear to be a salient feature of the MJO and that the role of SCC in the excitation and propagation in the MJO is questioned. As suggested from observations and model results in these many years, it appears that SCC is a natural consequence of convective organization in the warm pool area of the Indian/western Pacific Ocean and that it plays an important role in the MJO as its component.

Inclusion of the land-sea distribution in the future study will also modify other aspects of convective activity. It has been pointed out that convective activity is weaker over the Maritime Continent, for example, [104, 112, 113]. Convective activities over the America and Africa Continents have to be simulated. A more realistic SST distribution (such as inclusion of the cold pool around and to the south of the equator in the eastern Pacific) is also important. It is our strong concern to see how the MJO and various types of convective systems behave in the model under realistic conditions.

As mentioned in Section 2, the baroclinic instability occurs in the middle latitudes in the numerical experiments. The behaviors of the model MJO and convective systems should be indirectly affected by what occurs in the middle latitudes and more strongly affected by the behavior of the subtropical highs, which are closely associated with the baroclinic instability. As for the extratropical forcing that may contribute to the initial excitation of convection in the Indian Ocean (or onset of the MJO), some authors have discussed the role of subtropical Rossby wave train (e.g., Hsu et al. [105]). The importance of midlatitude baroclinic eddies in the excitation of the equatorial CISK mode was pointed out in a discharge-recharge theory of Blade and Hartmann [82]). These problems are also interesting, but are beyond the scope of this study. The problem of tropical cyclones associated with the MJO also remains to be reported in the future.

It may be useful to refer to the cloud resolving convection parameterization (CRCP, Grabowski [114]). An early attempt to study the MJO with the CRCP was made by Grabowski [115]. Some recent GCMs (SP-CAM) have used super parameterization (SP: the same as CRCP) for studies including the MJO e.g., [101, 116–118]). Since a coarse horizontal resolution (such as T42) has been used, the objectives of their studies with the SP-CAM and the present study with the MCRM should be different in some respects. The author's interest is how the SP-GCM will behave when the resolution becomes fine in the future.

As often mentioned in the author's previous papers, the MCRM was developed in the 1980s to study various phenomena in which moist convection plays an important role, with an intention of improving existing parameterization schemes of moist convection. Recent studies of the global warming effect on tropical cyclones still uses parameterization schemes which are essentially similar to those in the 1980s. In the near future when a finer grid (such as a 20 km grid) can be used, the author expects that the MCRM will be useful for studies of the global warming effect. Also, for this purpose, further studies of the MJO with the MCRM should be important.

As mentioned in Section 1, the MCRM used in this study is a hydrostatic model. In recent years, studies of the MJO and/or SCC have been made with a nonhydrostatic model e.g., [119–121]. Although the horizontal grid size used is not fine enough to resolve cumulus convection, the effects of the subgrid-scale convection have not been taken into account. The author's interest is to understand to what degree the model can properly treat the MJO and various types of convective systems. Comparison of results from the MCRM and the coarse-grid nonhydrostatic model will answer this problem. Numerical experiments under the same conditions and grid size have not been performed, however.

There is no doubt that a nonhydrostatic global model with a fine grid (such as a 1 km grid) will be (easily) used to study many problems in the near future. Even in that situation, a coarse-grid, hydrostatic and nonhydrostatic models will be still useful for efficient research. A nonhydrostatic MCRM has also been developed for this purpose (Yamasaki [56]). Since the hydrostatic MCRM is much more efficient than the nonhydrostatic MCRM, the former is used in this study. However, the latter has different merits, and it will be more useful for other objectives of research. Comparison of three results from hydrostatic and nonhydrostatic MCRMs, and a coarse-grid nonhydrostatic model without subgrid-scale effects will help us better understand observed phenomena, and thereby, improve these models. Under some restricted conditions, a cumulus-convection-resolving (global) model (with a grid size of 1 km or less) will be used, and it should also be important for better understanding of observed phenomena and improvement of the two MCRMs.

Needless to say, interactions among researchers are very important. Although this study is by no means satisfactory in understanding the MJO and related convective systems, and a number of necessary studies (modeling studies as well as observational studies) remain to be made, it is hoped that this study will be useful for interactions among researchers in this research field and that it will contribute to advances in meteorology through this open access Journal "Advances in Meteorology".

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Endnotes

1. The phase speed is described as $10\text{--}15\text{ m s}^{-1}$ in N88, and SCCs with slower phase speeds ($5\text{--}10\text{ m s}^{-1}$) in N88 are referred to by Lau et al. [10].
2. It was assumed that the heating rate was proportional to the vertical velocity at the top of the boundary layer, as was proposed by Ooyama [20] in his TC study. In this respect, a remark is necessary. Some subsequent researchers have referred to this type of parameteriza-

tion as "CISK-type parameterization". However, this term is not appropriate because other parameterization schemes have also been (and should also be) used for studies of CISK.

3. In Y69 and Y71, the easterly wave in the troposphere was also one of the major concerns.
4. The author has used the term wave-CISK as indicating an instability due to cooperative interaction between a wave and moist convection, as mentioned in Yamasaki [18, 33], and OY01 [12]. Some researchers (e.g., Chao [34]) have used it as implying the instability of the type discussed in the 1970s and 1980s (including the instability of unrealistic, small-scale Kelvin waves and gravity waves). The author's definition is also different from that of Chao and Lin [35] in which it is stated that wave-CISK is responsible for the growth of a cloud cluster. The author has considered that it is appropriate to use the term wave-CISK as instabilities of Kelvin wave, and SCC that is an ensemble of cloud clusters. In the case of TCs, what has been called CISK phenomenon is not a rainband (corresponding to a cloud cluster) but a TC that is an ensemble of rainbands and various-shaped convective systems including an eyewall.
5. The instability of gravity waves is a result of the interaction between two vertical modes, as was suggested by Syono and Yamasaki [22]. It was also shown that a stationary wave occurs when parameterized heating rates exceed critical values in the lower troposphere. This instability, which corresponds to conditional instability of the first kind (CIFK), is also due to inappropriate parameterization of moist convection because the parameterization used intended to avoid CIFK. These instabilities were later discussed by Chang and Lim [36] with the equatorial beta-plane model. When surface friction is taken into account, another type of an unstable gravity wave is obtained (e.g., Syono and Yamasaki [22]; Wang [37]; Oouchi and Yamasaki [38]).
6. Cumulus-convection-resolving model (CCRM) corresponds to the cloud resolving model (CRM) that is used for a horizontal grid size of 1 km–100 m. Some researchers have used the term CRM for a larger grid size such as 5 km. It should also be mentioned that the term cloud resolving is rather a general term that can be used for a wide range of clouds. The model used in this study (MCRM) also belongs to CRM in this sense. Therefore, the author has used the term CCRM instead of CRM. This is also based on the recognition that it is important to resolve cumulus convection which is the basic mode of moist convection, unless the effects of cumulus convection are taken into account (or parameterized).
7. Some researchers have argued that the unconditional heating case is unrealistic. However, the discussion from this case was very important as the first step because the essence of the stability properties was obtained from the unconditional heating case, and important properties for realistic conditions were inferred.

8. Since no Kelvin wave is distinct but a Rossby wavelike pattern is distinct in the lower troposphere, it seems to the author that this wave may not correspond to the MJO.
9. Presented at a workshop at UCLA in 1993, unpublished.

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