Considerations for the Structure of the Tropical Intraseasonal Oscillation

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Abstract

The 30–60 day tropical intraseasonal oscillation (ISO) has a quite complicated structure. For example, different behavior is observed between the wavenumber-one zonal wind and the wavenumber-zero temperature (or geopotential). An attempt is made to understand such a structure from a unified view, using a linear response model based on the primitive equations. The thermal forcing, which is externally prescribed, has a 40 day period with variable amplitudes and moves eastward over a distance of 120° of longitude. This model simulates well the observed structure.

The key to understanding the structure of the ISO lies in the contrast between the slow moving speed of the forcing and the fast group velocity of Kelvin modes. Regions with upward motion move slowly with the forcing, while areas with downward motion promptly extend to the entire longitude owing to the Kelvin modes. Thus, the vertical velocity and divergence fields have the same wavenumber-one structure as the heating. The zonal wind is the integral of the divergence, so that it must exhibit a wavenumber-one structure. On the other hand, the temperature must have a wavenumber-zero structure. The reason is that high temperatures result from diabatic heating in the forced region and adiabatic heating due to downward motion in the unforced region, and the maximum of diabatic heating and that of downward motion in the unforced region occur almost simultaneously. The geopotential shows similar behavior via the hydrostatic equation. Thus, the wavenumber-one structure of the zonal wind and the wavenumber-zero structure of the temperature (geopotential) necessarily correspond to each other.

Marked structural changes occur in one cycle of the ISO. The slowly moving heat source excites a Matsuno-Gill (MG) pattern. When the heating amplitudes become small in the eastern portion of the forced region, a free Kelvin mode separates and moves eastward. Thus, a MG pattern having a slow speed is observed in the forced region, while a fast Kelvin mode is formed in the unforced region. Further, it can be shown that the vertical structure of the geopotential should be different from that of the zonal wind.

The effects of the values of the imposed parameter on the structure of the model ISO are also examined. Moreover, five time scales which characterize the ISO are proposed, and the relationships which they must satisfy in order to reproduce the observed structure are discussed.

1. Introduction

The 30–60 day tropical intraseasonal oscillation (hereafter referred to as ISO) observed by Madden and Julian (1971) has a quite complicated structure. Among the structural characteristics of the ISO are: 1) The zonal wind exhibits a wavenumber one and baroclinic structure (Madden and Julian, 1972, hereafter referred to as MJ72). The eastward phase

velocity is as slow as about $5 \text{ m} \cdot \text{sec}^{-1}$ over the western Pacific, while it is as fast as $15 \sim 35 \text{ m} \cdot \text{sec}^{-1}$ over the eastern Pacific or the Atlantic (MJ72; Knutson *et al.*, 1986; Nishi, 1989, hereafter referred to as N89; Gutzler and Madden, 1989). Thus, the phase speed undergoes drastic changes in the longitudinal direction.

2) The phase velocity of the surface pressure is very fast (MJ72, N89). In other words, the surface pressure shows an almost standing oscillation,

i.e., a wavenumber-zero oscillation. Its phase leads the lower-level zonal wind by about 90° over the maritime continent, while they are approximately in phase over the eastern Pacific (MJ72, N89).

3) The upper-level geopotential and temperature also show an almost standing oscillation (N89). When the zonal mean is subtracted, the deviation of the geopotential has a wavenumber-one structure (Hayashi and Golder, 1986).

4) The upper-level geopotential is not necessarily out of phase with the surface geopotential, but leads it by $70^\circ\sim 120^\circ$ over the western and mid Pacific (N89). In this sense, the geopotential does not have a complete baroclinic structure in these regions, in contrast to the zonal wind.

5) The so-called twin cyclones and twin anticyclones associated with the ISO are frequently observed in the subtropics and *always* move slowly eastward (Weickman *et al.*, 1985; Knutson *et al.*, 1986; Murakami, 1987). The twin anticyclones are located slightly west of the center of the upper-level divergence.

6) Outgoing longwave radiation (OLR) associated with the ISO is observed between about 60°E and 180° , and moves eastward at a speed as slow as about $5\text{m}\cdot\text{sec}^{-1}$ (Weickman, 1983; Weickman *et al.*, 1985; Lau and Chan, 1985; Nakazawa, 1986). The center of low values of OLR, *i.e.*, strong convection, coincides with the upper divergent region (Weickman *et al.*, 1985; Murakami, 1987; Nogués-Paegle *et al.*, 1989). Also, it is in phase with the low surface pressure over the western Pacific, as is shown in Appendix A.

There are many questions found in the above characteristics: a) Why do the phase speed and structure of the ISO change at different longitudinal locations? b) Why do the zonal wind and mass fields (the surface pressure, geopotential and temperature) have such a different structure? Is this accidental or is it necessary? c) Why do the twin eddies *always* move eastward, even though they can be regarded as Rossby modes? d) Why is the amplitude of the surface pressure so large, although the ISO, as a whole, is associated with a major baroclinic component and the surface pressure is associated with a barotropic component?

These characteristics and questions have not been understood from a unified view, although there have been several studies in which some aspects of the structure described above are reproduced (*e.g.*, Yamagata and Hayashi, 1984; Hayashi and Miyahara, 1987; Miyahara, 1987; Hendon, 1988). The main objective of this paper is to understand these characteristics and to answer these questions.

When a simple model is used here for this purpose, we encounter a difficulty in determining the magnitudes of various parameters, such as damping time constants. These parameters are more or less

arbitrary. Nevertheless, it will be shown later that these magnitudes must be severely restricted in order to reproduce the observed structure of the ISO in the numerical model. The determination of these magnitudes is one contribution to the understanding of the ISO. This is the second objective of this paper.

Many of the magnitudes of the above-stated parameters can be redefined as time scales. One obvious time scale is 30~60 days, but the ISO has many other time scales. As will be shown, but only when these time scales satisfy several restricted relationships, the observed structure of the ISO can be simulated by the numerical model. The third objective of this paper is to propose the time scales which characterize the ISO and to discuss their relationships.

The model used in this study will be described in Section 2. The given parameters are also explained in this section. In Section 3, the results from the experiment with the standard parameter values are shown, which reproduce the observed ISO rather well. Some considerations for the structure of the ISO are also made in this section. The sensitivity of the model ISO with respect to the parameters will be examined in Section 4. As a result, the adequacy of the standard values of the parameter to reproduce the ISO can also be shown. The various time scales associated with the ISO and their relationships will be proposed in Section 5.

2. Description of a two-level model

a. Basic equations

In order to understand the characteristics of the ISO in the simplest terms, we employ a two-level model based on the linear primitive equations on an equatorial β -plane. It is assumed that there is no basic flow and the surface is flat. The model top and bottom are set at 100 mb (p_t) and 1000 mb (p_s), respectively. The vertical p -velocity (ω) is specified at these two boundary levels and 550 mb, which is named level 2. Other variables, u (zonal wind), v (meridional wind), T (temperature), and ϕ (geopotential), are defined at the 325 and 775 mb levels, which are termed level 1 and 3, respectively. Motions are assumed to be symmetric about the equator. Lateral boundaries are set at the equator and at the latitude of 40° , where the boundary conditions are

$$v = 0, \quad \partial u / \partial y = \partial T / \partial y = \partial \phi / \partial y = 0. \quad (1)$$

The boundary condition at the model top is $\omega = 0$.

The governing equations are written as

$$\frac{\partial u_i}{\partial t} = f v_i - \frac{\partial \phi_i}{\partial x} - \alpha u_i, \quad (2)$$

$$\frac{\partial v_i}{\partial t} = -f u_i - \frac{\partial \phi_i}{\partial y} - \alpha v_i, \quad (3)$$

$$\frac{\partial T_i}{\partial t} = \omega_i S_i + \frac{Q_i}{C_p} - \alpha T_i, \quad (4)$$

$$\omega_1 = (\omega_2 + \omega_0)/2 = \omega_2/2, \quad (5)$$

$$\omega_3 = (\omega_s + \omega_2)/2, \quad (6)$$

$$\omega_2 = - \left(\frac{\partial u_1}{\partial x} + \frac{\partial v_1}{\partial y} \right) \Delta p, \quad (7)$$

$$\omega_s = \omega_2 - \left(\frac{\partial u_3}{\partial x} + \frac{\partial v_3}{\partial y} \right) \Delta p, \quad (8)$$

$$\frac{\partial \phi_s}{\partial t} = \frac{\omega_s}{\rho_s}, \quad (9)$$

$$\phi_3 = \phi_s + \frac{R \Delta p}{2 p_2} T_3, \quad (10)$$

$$\phi_1 = \phi_3 + \frac{R \Delta p}{2 p_2} (T_3 + T_1). \quad (11)$$

The notations are conventional, that is, t represents time, x the zonal coordinate, y the meridional coordinate, p the pressure, S the static stability, Q the diabatic heating rate, α the Rayleigh damping and Newtonian cooling rates, f the Coriolis parameter ($=\beta y$ where β is the Rossby parameter which is taken as $2.28 \times 10^{-11} \text{ m}^{-1} \text{ s}^{-1}$), C_p the specific heat at constant pressure ($1004 \text{ J} \cdot \text{kg}^{-1} \text{ K}^{-1}$), ρ_s the surface density ($1.225 \text{ kg} \cdot \text{m}^{-3}$), and R the gas constant for air ($287 \text{ J} \cdot \text{kg}^{-1} \text{ K}^{-1}$). Subscripts denote the given level, and Δp is the pressure difference between the top (or bottom) and the middle level (450 mb). Subscript i in Eqs. (2)~(4) is either 1 or 3. The surface geopotential is explicitly computed by Eq. (9). In this respect, the present model differs from many other two-level models (*e.g.*, Lim and Chang, 1983).

Explanations are necessary for the hydrostatic equations, (10) and (11). Equation (11) can be represented by an alternate equation, $\phi_1 = \phi_3 + R \Delta p (T_3/p_3 + T_1/p_1)/2$, which is more commonly used. However, since the understanding of the model dynamics is easier in the model employing Eq. (11) than the alternate equation, Eq. (11) is adopted here. Equation (10) is directly derived through energy conservation when Eq. (11) is employed. Differences between the two systems were calculated, and were found to be small.

An important problem is the formulation of the diabatic heating rate, Q . It is assumed that Q results from the action of cumulus clusters. For simplicity, it is also assumed that Q is prescribed with a given period and does not interact with any air motions. However, this does not exclude the possibility that the atmospheric ISO results from the interaction between air motions and cumulus convection. Even in this model, low-level convergence occurs in regions where cumulus convection is prescribed, as will be shown later. In this respect, the

structure appearing in this model is similar to that in a model having cumulus-air motion interactions. S and α , as well as Q , affect the structure of the model ISO. These parameters are described in the next subsection.

Numerical integrations are performed using Eqs. (2)~(11).

The initial conditions are taken to be the atmosphere at rest. In the horizontal, an Arakawa C-grid (staggered grid) with grid lengths of 4° latitude and 6° longitude is used. The time interval used in the integrations is 600 sec. Data are stored for one cycle of the ISO, usually the fourth cycle, after the oscillation reached a steady state.

In order to remove any suspicion that the vertical truncation of the two-level model is too severe, a twelve-level model was also designed. Results were similar to those for the two-level model; the existence/nonexistence of the planetary boundary layer and small changes in the vertical profiles of Q and S did not affect the structure of the model ISO. We can say that the vertical resolution of the two-level model is not coarse for the simulation of the ISO.

b. Parameters characterizing the model ISO

There are 3 parameters in Eqs. (2)~(4) which exert a great influence on the model ISO, namely, the diabatic heating rate Q , the static stability S , and the damping and cooling rate α . In practice, their values, however, cannot be specified from observations alone. In the present study, their values are determined in the following manner. The structure of the model ISO changes as the values of these parameters change. The model is adjusted by changing the values so that the model structure coincides with the observed one. The magnitudes of the parameters might be regarded as in agreement with those of the real atmosphere. The dependence of the model ISO on these parameters will be extensively examined in Section 4. In this subsection, only standard values of these parameters will be given.

We assume that S is horizontally uniform and functions of only pressure. This assumption is required from the system of linear equations. Following the result in Appendix A, $S_1 + S_3$ is selected as $10^{-3} \text{ K} \cdot \text{Pa}^{-1}$, and the standard ratio of S_1/S_3 is set to 1.4. The standard value of α is $1/(8 \text{ days})$.

On the basis of observations for OLR (see item 6 in the Introduction), the heating rate, Q , is formulated as

$$\frac{Q_i}{C_p} = q_i \cdot \sin^n \left(\frac{\pi}{T_c} t_1 \right) \cdot \cos \frac{\pi}{R_c} \left(x - \frac{L}{T_c} t_1 \right) \cdot f_0(y), \quad (12)$$

when $\sin(\pi/T_c)t_1 > 0$ and $\cos\{(\pi/R_c)(x - Lt_1/T_c)\} > 0$, otherwise,

$$Q_i/C_p = 0.$$

Here, $t_1 = t - T_o[t/T_o]$, where $[t/T_o]$ represents a maximum integer which does not exceed t/T_o . Therefore, sub-parameters in this formulation are q_i , n , T_c , R_c , L , $f_0(y)$, and T_o . The parameter T_o is the period of the ISO. Note that the period of the ISO is thus defined as the period of reoccurrence of cumulus clusters, as observed by Weickman *et al.* (1985) and Nakazawa (1986). This period should not be confused with the time taken for wave-CISK modes to propagate around the globe. T_c represents the duration of a cumulus cluster, and the time variation of heating amplitudes is regulated by n , which is hereafter called the amplitude factor. R_c and L are the half length of a cumulus cluster and the length of the forced region in the longitudinal direction, respectively. The parameters q_i and $f_0(y)$ are the structure of the heating in the vertical and latitudinal direction, respectively.

Among these sub-parameters, q_i and especially n are found to be difficult parameters in the sense of determining their magnitudes from observations alone. The standard value of n is 1.5, although there is no observational support to choose this value. Owing to the linearity of the equation system, amplitudes of the forcing are arbitrary, and a value of $q_1 + q_3 = 10 \text{ K} \cdot \text{day}^{-1}$ is selected. The ratio of q_1/q_3 has a great influence on the model ISO, and the standard value is 1.4. Both the standard values of T_o and T_c can be set to 40 days without any problem. R_c is of the order of 1000 km (Nakazawa, 1988; Hayashi and Nakazawa, 1989), so that 30° longitude is given as the standard value. Since the standard forced region is defined from 60°E to 180° longitude on the basis of observations (*e.g.*, Weickman *et al.*, 1985; Nakazawa, 1986), the standard value for L is 120° . $f_0(y)$ is not a sensitive parameter and is formulated as $f_0(y) = \exp(-\beta y^2/(2\sqrt{gh}))$, where $h = 100 \text{ m}$, unless otherwise stated.

Figure 1 shows time-longitude sections of the heating employed in this study, where parameter values are standard and the maximum amplitude is unity. It should be pointed out in Fig. 1b, which illustrates the deviation from the time mean, that negative heating appears in the forced region. Further, note that longitudes where the heating attains its maximum at a given time in Fig. 1b differ from those in Fig. 1a, resulting in the speed of the heating maximum being faster in Fig. 1b than in Fig. 1a. This occurs owing to larger (smaller) time-mean values being subtracted in east sides than in west sides over the region west (east) of 120° .

c. Considerations for basic characteristics of the two-level model

In order to gain some physical insights, barotropic and baroclinic equations in non-dimensional forms are developed from Eqs. (2)~(11). The length, time and temperature scales are $(c/2\beta)^{1/2}$, $(2\beta c)^{-1/2}$,

and $2c^2 p_2/(R\Delta p)$, respectively, where $c^2 = R\Delta p^2 S/(4p_2)$ and $S = (S_1 + S_3)/2$ (*i.e.*, $0.5 \times 10^{-3} \text{ K} \cdot \text{Pa}^{-1}$). The symbol c denotes the phase (group) velocity of baroclinic Kelvin waves. By substituting values of the parameters, c takes a value of $36.3 \text{ m} \cdot \text{sec}^{-1}$, while the three scales measure $8.92 \times 10^5 \text{ m}$, $2.45 \times 10^4 \text{ sec}$ (0.284 day), and 11.25 K. For simplicity, non-dimensional variables are expressed using the same notations as dimensional variables, with the exception of the heating term, whose notation is changed from Q/C_p to H . The system of equations then becomes

$$\frac{\partial u_-}{\partial t} = \frac{1}{2} y v_- - \frac{\partial \phi_-}{\partial x} - \alpha u_-, \quad (13)$$

$$\frac{\partial u_+}{\partial t} = \frac{1}{2} y v_+ - \frac{\partial \phi_+}{\partial x} - \alpha u_+, \quad (14)$$

$$\frac{\partial v_-}{\partial t} = -\frac{1}{2} y u_- - \frac{\partial \phi_-}{\partial y} - \alpha v_-, \quad (15)$$

$$\frac{\partial v_+}{\partial t} = -\frac{1}{2} y u_+ - \frac{\partial \phi_+}{\partial y} - \alpha v_+, \quad (16)$$

$$\begin{aligned} \frac{\partial T_+}{\partial t} = & \left(\frac{\partial u_-}{\partial x} + \frac{\partial v_-}{\partial y} \right) \\ & - \left(\frac{\partial u_+}{\partial x} + \frac{\partial v_+}{\partial y} \right) (1 + S_3) + H_+ - \alpha T_+, \end{aligned} \quad (17)$$

$$\begin{aligned} \frac{\partial T_-}{\partial t} = & \left(\frac{\partial u_-}{\partial x} + \frac{\partial v_-}{\partial y} \right) S_- \\ & - \left(\frac{\partial u_+}{\partial x} + \frac{\partial v_+}{\partial y} \right) (S_- + S_3) + H_- - \alpha T_-, \end{aligned} \quad (18)$$

$$\gamma \frac{\partial \phi_s}{\partial t} = - \left(\frac{\partial u_+}{\partial x} + \frac{\partial v_+}{\partial y} \right), \quad (19)$$

$$\phi_- = -T_+, \quad (20)$$

$$\phi_+ = \phi_s + 2T_+ + T_-. \quad (21)$$

In these equations, $u_+ = (u_3 + u_1)/2$ and $u_- = (u_3 - u_1)/2$, with the same format employed for the other variables. Note that the barotropic components (u_+ , v_+ , ϕ_+ and ϕ_s) necessarily appear through interaction with baroclinic modes, even for the case of $H_- = S_- = 0$.

The parameter, $\gamma = \rho_s c^2/(2\Delta p) \sim 0.018$, can be interpreted as the equivalent depth of the baroclinic mode normalized by the scale height. The small value of γ leads to the following important conclusion. Although both u_+ and ϕ_s belong to a barotropic component, their amplitudes are very different. Amplitudes of u_+ are negligibly small, meaning that the zonal wind has a complete baroclinic

Fig. 1. (a) Hovmöller diagram of the normalized heating with a maximum amplitude of unity and an amplitude factor of 1.5. (b) As in (a) except that the time mean at each longitude has been subtracted. Contour interval is 0.2, and regions of positive values are shaded.

structure, as observed. On the other hand, amplitudes of ϕ_s are large, which means that the surface pressure emerges as a marked signal of the ISO. This is an answer to Question d) presented in the Introduction.

Next, an estimation of the order of Eqs. (13)~(21) is conducted at the equator, *i.e.*, under the condition of $y=v=0$. In the forced region, H_+ has the largest value and must be balanced by $\partial u_-/\partial x$. Therefore,

$$O(1) \text{ eq. } H_{+0} + \frac{\partial u_{-0}}{\partial x} = 0, \quad (22)$$

$$O(\varepsilon) \text{ eq. } \frac{\partial u_{-0}}{\partial t} = \frac{\partial T_{+1}}{\partial x} - \alpha u_{-0}, \quad (23)$$

where ε is a small parameter and the subscripts 0 and 1 denote the orders of magnitude $O(1)$ and $O(\varepsilon)$, respectively. These equations indicate that the behavior of T_+ differs from that of u_- , as has been found in the observed ISO. This is a partial answer to the previously posed Question b).

On the other hand, in the unforced region, u_- and T_+ show the same behavior, since

$$\frac{\partial T_{+0}}{\partial t} = \frac{\partial u_{-0}}{\partial x} - \alpha T_{+0} \quad \text{and} \quad \frac{\partial u_{-0}}{\partial t} = \frac{\partial T_{+0}}{\partial x} - \alpha u_{-0}. \quad (24)$$

Thus, the variables characterizing the ISO can be divided into two groups based on the difference in their behavior. The first group consists of H_+ and u_- , with the ω field also being included in this group. Since H_+ has a wavenumber-one structure, u_- and ω should also have this structure. The second group is composed of T_+ , ϕ_- and ϕ_s . From Eqs. (22)~(24) it can be predicted that the second group may not have a wavenumber-one structure.

3. Results of the experiment with the standard parameter values

Data from the standard experiment are analyzed, using the Fourier transform with respect to time, after being averaged between the equator and 10°N . In Fig. 2, the phase and amplitude of the 40 day period as a function of longitude are shown for several variables. Day 20 (phase of 180°) is the time when the forcing has its maximum value. Positive slopes in this figure indicate eastward propagation. When phase lines intersect the ordinate for phase or time at only one point, such variables have a wavenumber-one structure.

The phase structure is similar to those found in MJ72, N89 and others. That is, Characteristics 1), 2), 3), 4), and 6) mentioned in the Introduction are simulated well, if the forcing is made to correspond with the OLR.

However, minor discrepancies are found. The geopotential and temperature have large amplitudes even in the unforced region. These are not found in the observed ISO. Amplitudes of the upper-level zonal wind are very small over unforced longitudes. This is also contradicted to observations (*e.g.*, Gutzler and Madden, 1989). These discrepancies will be discussed later.

Figure 3 illustrates synoptic maps of the horizontal structure at the upper level for every 5 days. From day 5 through 30, the structure exhibits a Matsuno-Gill pattern (Matsuno, 1966; Gill, 1980; hereafter referred to as the MG pattern). After day 30 the Kelvin and Rossby modes separate, although the geopotential field associated with the Kelvin mode does not show definite features. The Kelvin mode moves rapidly eastward as a free first-baroclinic mode. The Rossby mode, although it does not show distinct twin anticyclones, moves westward. The direction of movement does not ap-

Fig. 2. Phase and amplitude of several variables with a 40 day period in the experiment using the standard parameter values. The meaning of each character is as follows: P , surface geopotential height; U , upper-level zonal wind; L , lower-level zonal wind; T , upper-level temperature; H , upper-level geopotential height. The values have been averaged between the equator and 10°N . The thick dashed line represents the heating with amplitudes larger than $1/e$ of the maximum. Amplitude of the lower-level zonal wind is omitted, since it is almost the same as the upper-level one.

pear to be consistent with observations, which show an eastward movement as stated in the Introduction. This “contradiction” will be solved in Fig. 4.

Figure 4 is the same as Fig. 3 except that the time mean at each longitude is subtracted from the field. The notable difference from Fig. 3 is that the MG pattern having the small values of geopotential appears with the cool anomaly. The time when the Kelvin and Rossby modes separate is about day 30, which is earlier than in Fig. 3. The geopotential associated with the Kelvin mode is identical to the theoretical structure. The twin cyclones and anticyclones are clearly detectable and *always* move eastward, captured by the heating and cooling regions. The eastward movement is consistent with observations, because time-mean fields or seasonal trends are always subtracted in observational studies. Thus, an answer to Question c) and an understanding of Characteristic 5) are obtained. The key to why the Rossby mode moves “eastward” is the operation of subtracting the time-mean.

Figure 5 displays time-longitude sections of the temperature at 325 mb and the vertical p -velocity at 550 mb. We can see that the signal of ω radiates rapidly from the forced region eastward. This speed appears to coincide with the group velocity of free Kelvin waves. This will be further verified in Subsection 4.g. Westward propagation, although very weak, can also be found. This speed is probably identical to the group velocity of Rossby waves. The temperature in the unforced region reaches its maxima slightly after ω attains its maxima. The minimum temperature also propagates rapidly eastward. It can easily be understood that this is caused by the upward motion associated with the trailing

part of the separated Kelvin wave (see Fig. 3).

At this point, Characteristics 1), 3) and 6) can be explained. The MG pattern is formed by the slow-moving forcing. When the amplitude of the forcing becomes small, the large amplitude MG pattern collapses, and the separation into Kelvin and Rossby modes necessarily occurs. Characteristic 1) is brought about by the MG pattern moving slowly in the forced region and by the fast Kelvin mode in the unforced region (see also Hendon, 1988). Characteristic 3) can be understood by examining Fig. 5. Signals of ω propagate fast, except just west of the forced region. Maxima of heating and those of ω in the unforced region occur almost simultaneously. The former causes the temperature maxima in the forced region while the latter creates those in the unforced region. Thus, the temperature shows an almost standing oscillation. The upper-level geopotential also has the same structure as the temperature via the hydrostatic equation. Characteristic 6) is obvious.

In order to understand Characteristics 2) and 4), Fig. 6 is shown, in which the phases and amplitudes are displayed for several variables in non-dimensional form to compare their relative magnitudes. As has already been predicted, amplitudes of u_+ are negligibly small, while $T_+ (= -\phi_-)$ and ϕ_+ have large amplitudes. Further, we can see that the amplitude (phase) of T_- is small (out of phase), compared with T_+ . One noticeable feature in this figure is the structure of ϕ_+ . Its amplitude is large, contrary to u_+ , and the amplitude and phase are almost constant with longitude. Furthermore, the phase almost coincides with the phase of T_+ when averaged longitudinally.

Fig. 3. Horizontal structure at 325 mb every 5 days for the same experiment as given in Fig. 2. Refer to the lower-right hand side of each chart for wind scales and contour intervals. Regions where geopotential height is greater than 20 m are shaded. The thick solid lines represent heating contours with $1/e$ of the maximum.

Fig. 4. As in Fig. 3 except that the time means have been subtracted. Thick dashed lines show heating contours of -0.2 times the maximum.

Fig. 5. Hovmöller diagrams for (a) the temperature at 325 mb, and (b) the vertical p -velocity at 550 mb over 2°N for the same experiment as in Fig. 2. Two cycles are depicted. Contour intervals are 0.5 K for (a), and 0.05 $\text{Pa}\cdot\text{sec}^{-1}$ (solid line) and $1/4 \times 0.05 \text{ Pa}\cdot\text{sec}^{-1}$ (dashed line) for (b). Shading: $> 1 \text{ K}$ for (a) and > 0 for (b).

Fig. 6. As in Fig. 2 except for several barotropic and baroclinic components. Amplitudes are nondimensionalized. The meaning of each character is as follows: P , ϕ_s ; U , u_+ ; H , ϕ_+ ; T , T_+ ; and M , T_- .

By using Eqs. (13)–(21), these features can be proven as follows. Let δ denote the ratios of S_1/S_3 and q_1/q_3 which were assumed to be 1.4 in this model. Equation (18) can be rewritten as

$$\frac{\partial T_-}{\partial t} = \frac{1-\delta}{1+\delta} \left\{ \left(\frac{\partial u_-}{\partial x} + \frac{\partial v_-}{\partial y} \right) + H_+ \right\} - \left(\frac{\partial u_+}{\partial x} + \frac{\partial v_+}{\partial y} \right) (S_- + S_3) - \alpha T_- \quad (25)$$

Since $(\partial u_+/\partial x + \partial v_+/\partial y)$ in Eqs. (17) and (25) is negligible,

$$T_- \sim (1-\delta)/(1+\delta)T_+ = -0.17T_+ \quad (26)$$

This means that T_- has a small amplitude and is out of phase with T_+ . Next, in order to prove the

properties of ϕ_+ , area means and time derivatives of Eq. (21) are taken. Then, the equation,

$$\frac{\partial}{\partial t} [\phi_+] \sim \frac{\partial}{\partial t} [\phi_s] + 1.83 \frac{\partial}{\partial t} [T_+],$$

can be obtained, where $[\]$ represents an area mean and $[T_-] \sim -0.17[T_+]$ is used. In this equation, $\partial[\phi_s]/\partial t = 0$ from Eqs. (19) and (1), and initially $[\phi_s] = 0$. Further, since $\partial\phi_+/\partial x$ and $\partial\phi_+/\partial y$ are found to be negligibly small from Eqs. (14) and (16), respectively, ϕ_+ is nearly constant at any locations. Thus, we can obtain

$$\phi_+ \sim [\phi_+] \sim 1.83[T_+] \quad (27)$$

This indicates that ϕ_+ has large amplitudes; its amplitude and phase are almost constant with longitude and ϕ_+ is in phase with $[T_+]$. The phase of $[T_+]$

Fig. 7. (a) Amplitude ratio of the surface geopotential height to the upper-level zonal wind (unit; sec), and (b) phase difference between the minimum surface geopotential and the maximum forcing (unit; degree). They are shown as a function of S_1/S_3 and q_1/q_3 , calculated from the experiment with steady forcing and averaged between the equator and 10°N .

is mainly determined by T_+ in the tropics, because T_+ is much larger there than in the mid-latitudes or subtropics. This results from the fact that H_+ is predominantly large in the tropics. Therefore, ϕ_+ is also in phase with T_+ when averaged in the tropics, as seen in Fig. 6.

At this time Characteristics 2) and 4) can be understood. The fact that T_+ near the equator is large also indicates that $1.83T_+$ is much larger than ϕ_+ in equatorial regions and, therefore, ϕ_s has an amplitude on the same order as T_+ . This, together with the zonal derivative of Eq. (21), *i.e.*,

$$\frac{\partial\phi_+}{\partial x} \sim \frac{\partial\phi_s}{\partial x} + 1.83\frac{\partial T_+}{\partial x} \sim 0, \quad (28)$$

produces the following conclusions. The structure of T_+ is wavenumber zero, so that ϕ_s also has a wavenumber-zero structure. Since amplitudes of ϕ_s are not larger than those of $1.83T_+$, ϕ_s must exhibit larger phase variations in the longitudinal direction than T_+ . Thus, the geopotential field cannot have a complete baroclinic structure, unlike the zonal wind. Although the above proof was derived under the assumption that q_1/q_3 and S_1/S_3 have the same value, needless to say the proof would still be valid whenever T_- is small.

4. Dependence of the model ISO on various parameters

In this section, we will examine the dependence of the model ISO on various parameters. The dependence on each parameter will be examined in each subsection. As a result, the validity of the standard parameter values will be demonstrated. Parameter values are changed over wide ranges, including even unrealistic values, since it will be attempted to understand the ISO from a general point of view in

the next section. Except for the parameter under consideration, the standard values are employed for the remaining parameters.

a. Vertical profiles of the forcing and static stability

In this subsection, the standard values of q_1/q_3 and S_1/S_3 will be determined. The model forcing differs from the formulation previously described: The forcing propagates around the globe with a constant amplitude. We confirmed that the structure from this model can approximate that in the forced region of a model with time-dependent forcing.

These magnitudes are determined as follows. N89 showed that $z_s/u_1 \sim 2$ sec, where z_s is the geopotential height at 1000 mb. Appendix A shows that the OLR minima are in phase with the surface geopotential minima over the western Pacific. These properties depend on the vertical profiles of the forcing and static stability, and can be reproduced in a numerical model only by adopting the proper vertical profiles. Thus, q_1/q_3 and S_1/S_3 can be objectively derived.

Figure 7 shows the amplitude ratio of z_s to u_1 , and the phase difference between forcing maxima and surface geopotential minima, as a function of q_1/q_3 and S_1/S_3 . According to Fig. 7a, q_1/q_3 and S_1/S_3 must be nearly equal, in order that the value of z_s/u_1 equals about 2 sec. The phase relation can be satisfied, when $S_1/S_3 \geq q_1/q_3$.

On the other hand, the observed value of S_1/S_3 is from 0.9 to 1.8, as shown in Appendix A. It is difficult to obtain observed values of q_1/q_3 , but they can be estimated as being about 1.2~1.5 from Yanai *et al.* (1973), although the heating profile associated with the ISO might be different.

From the above considerations, 1.4 was adopted as the standard value of both q_1/q_3 and S_1/S_3 .

b. Time variation of the magnitude of the forcing

Six experiments are performed for the amplitude factors, n , of 0.25, 0.5, 1, 2, 4, and 8. Note again that, when n is small, large-amplitude forcing persists for a long period of time.

Results are as follows, although figures are not shown. The surface geopotential for $n=0.25$, 0.5, and 1 indicates a wavenumber-one structure, which differs from observations. The zonal wind for $n=4$ and 8 shows a movement from 60° westward to 300° , which is also different from observed zonal winds. The horizontal structure indicates that the westward movement is associated with a Rossby mode. Thus, from among the 6 values of the amplitude factor, the only proper value is 2.

Table 1 summarizes whether or not the model ISO satisfies several important characteristics. Time scales of large-amplitude forcing are defined as the duration when the forcing amplitude is greater than $1/e$ of its maximum. The surface geopotential and the upper-level temperature (or the upper-level geopotential) are defined as having a wavenumber-zero structure, when their phase variations in the longitudinal direction are within 135° and 90° , respectively. In this judgment, phases with amplitudes smaller than $1/5$ of the maximum amplitude are neglected. A Rossby mode appears west of the forced region in many experiments, as has already been shown, although it is not observed in the real ISO. Therefore, the characteristic that it does not appear is important, and is added to this table. If these characteristics, which are considered to be independent of each other, are all satisfied, then it can be stated that an experiment reproduces the observed structure rather well.

c. Period of the ISO

Experiments are carried out for the period of 20, 30, 60, and 90 days. As a matter of course, the moving speed of forcing and the duration of large-amplitude forcing also change in each experiment (see Table 1). The experiments for 60 and 90 day period reproduce all the 7 characteristics, while the experiments for 20 and 30 day period do not reproduce a wavenumber-zero structure of the temperature and geopotential. Further, the results for 20 day period exhibit neither MG patterns in the forced region nor "fast" phase speeds in the unforced region, compared with the 20 day period.

d. Longitudinal length of the forced region

Four experiments for forced region lengths of 60° , 90° , 150° , and 180° are performed. The results for the length of 90° are similar to those for the original length (120°). The influence of the westward movement of Rossby modes appears for the length of 60° . The surface geopotential no longer has a wavenumber-zero structure for the length of 150° or 180° . In addition, the temperature does not exhibit

a wavenumber-zero structure for the length of 180° .

e. Longitudinal scale of cumulus clusters

The half longitudinal scales for cumulus clusters are changed from 10° to 50° with an increment of 10° . The phases for the longitudinal scale of 20° or more are similar to those for the original scale (30°). However, for the scale of 10° , the temperature and geopotential do not exhibit a wavenumber-zero structure.

f. Time constant of Rayleigh damping and Newtonian cooling

Experiments are carried out for the time constants of 2, 4, 16, and 32 days. The temperature and geopotential no longer have a wavenumber-zero structure for the time constant of 2 or 4 days. Further, Kelvin modes are not generated for the time constant of 2 days. For the 32 day time constant, a Rossby mode with large amplitudes moves westward and MG patterns are not excited. Even for 16 days a Rossby mode seems to exist. Thus, none of the 4 time constants appear appropriate for the simulation of the ISO.

g. The thickness of the troposphere

In this series of experiments, the thickness of the troposphere $2\Delta p$ is kept small, fixing $p_s = 1000$ mb. The range of Δp in these experiments is from 350 mb to 50 mb with an increment of 100 mb. p_2 and p_t are defined in the same manner as in the original model; $p_2 = p_s - \Delta p$, $p_t = p_2 - \Delta p$. The measure of the latitudinal scale of the forcing, h , is also changed, being proportional to c^2 .

Figure 8 shows the propagation of the vertical p -velocity in a time-longitude section at p_2 for $\Delta p = 150$ mb. The propagation speed east of the forced region is fairly slow, which coincides with the group velocity of the free Kelvin mode. Similar results are obtained for other values of Δp . From these results it can be stated that signals of ω propagate eastward at the speed of the free Kelvin mode.

5. Discussion

a. Considerations of various time scales

The time scale which characterizes the ISO ranges from 30 to 60 days, with the standard value taken as 40 days in the present model. In addition, it can be seen from previous experiments that other time scales play important roles: one related to the moving speed of the forcing, another to the duration of large-amplitude forcing, a third to the phase speed of the free Kelvin wave, and the damping time constant. Only when the relationships among these time scales are appropriate, does the model structure of the ISO appear to be similar to that observed. In order to understand the ISO more fully, these relationships are examined.

Before considering these relationships, the above-

subsection			3	4.b					4.c				4.d		4.e		4.f				4.g	
name of experiment			A	B	C	D	E	F	G	H	I	J	K	L	M	N	O	P	Q	R	S	
parameters used in each experiment	period of ISO (T_o ; day)		40					20	30	60	90											
	amplitude factor (n ;)		1.5	0.5	1.0	2.0	4.0															
	damping time constant (T_d ; day)		8													2	4	16	32			
	forcing period (T_e ; day)		40					20	30	60	90											
	length of forced region (L ; °)		120									60	180									
	half length of cumulus cluster (R_c ;°)		30											10	50							
	thickness of troposphere ($2\Delta p$; mb)		900																	700	300	
	time scales (T_f ; day)		120					60	90	180	270	240	80									
		(T_p ; day)	26	37	30	23	17	13	20	39	59											
		(T_k ; day)	13																18	48		
satis- faction of the 7 charac- teristics observed in the ISO	A	slow u in the forced region	○	○	○	○	○	○	○	○	○	○	○	○	○	○	○	○	○	○	○	
	B	fast u in the unforced region	○	○	○	○	×	△	○	○	○	△	○	○	○	○	○	△	×	△	×	
	C	wavenumber 1 of u	○	○	○	○	×	○	○	○	○	×	○	○	○	○	○	○	×	○	×	
	D	MG pattern in the forced region	○	○	○	○	○	△	○	○	○	○	○	○	○	○	○	○	×	○	○	
	E	Kelvin in the unforced region	○	○	○	○	×	○	○	○	○	○	○	△	○	×	○	△	×	○	○	
	F	nonappearance of Rossby modes	○	○	○	○	×	○	○	○	○	×	○	○	○	○	○	△	×	○	○	
	G	wavenumber 0 of T and ϕ	○	×	×	○	○	×	×	○	○	○	×	△	○	×	×	○	×	×	×	
unsatisfied conditions						Ⓕ	Ⓒ F	Ⓖ D				F	Ⓓ			E		Ⓓ F	D F		B Ⓒ E	

Table 1. Characteristics of 19 representative experiments. The first line represents the subsection in which the results of each experiment are described.

Explanations of the parameters in the column of “parameters . . .” are given in Section 2. Blanks in the column denote that the standard values are adopted. See Section 5 for explanations of T_f , T_p and T_k in the column headed “time scales”. Meanings of the symbols in the column of “satisfaction . . .” are as follows; ○, satisfaction; ×, nonsatisfaction; and △, uncertain. When one condition or more among the 6 conditions (A~F) proposed in Section 5 are not satisfied, the unsatisfied conditions are written in the last line. Letters encircled mean that the conditions are on a border between the satisfaction and nonsatisfaction.

Fig. 8. Hovmöller diagram of the vertical p -velocity at p_2 (850 mb) over 2°N from the experiment of $\Delta p = 150$ mb. Other values are the same as in Fig. 5b.

mentioned time scales should be more rigorously defined. First, the time scale of the ISO is obvious and has been abbreviated as T_o . The time scale characterizing the speed of movement of the forcing is defined as the time required to move around the earth at this speed. It will be abbreviated as T_f . The duration of the large-amplitude forcing is defined as the time when the forcing amplitude is greater than $1/e$ of its maximum, and will be abbreviated as T_p . As for the time scale related to the Kelvin wave, it is also defined as the time required to propagate around the earth, being noted as T_k . Finally, the time scale of the damping is obvious, and will be written as T_d . There is, however, one additional factor concerning this time scale. Although there is no problem in the use of T_d when simply estimating the damping scale, it is appropriate to use $2\pi T_d$ in some cases. For example, it should be used when comparing the damping time scale with time scales of periodic oscillations (since a frequency is $2\pi/\text{period}$). For this purpose, it will be abbreviated as pT_d . These 5 time scales are summarized in Table 2.

In addition to these scales there is a time scale of adjustment; that is, the time required for the formation of the MG pattern after forcing has been introduced, and the time required for the separation of the MG pattern into Kelvin and Rossby modes when forcing becomes weak or disappears. However, since this scale is very short (Silva Dias *et al.*, 1983; Matsuda and Takayama, 1989) when compared with the other time scales, it will not be taken into consideration.

Next, the relationships among the 5 time scales are considered. We propose that the 7 characteristics listed in Table 1 can be reproduced when the

following conditions are satisfied, although a condition for Characteristic G (for a wavenumber-zero structure of the temperature and geopotential) is incomplete.

A). $T_f \gg T_o$ for Characteristic A (the slow phase speed of the zonal wind in the forced region).

The zonal wind in the forced region is closely related to the forcing, according to Eq. (22). Therefore, a slow phase speed of the zonal wind results from the slow movement of the forcing.

B). $T_o \gg T_k$ for Characteristic B (the fast phase speed of the zonal wind in the unforced region).

This is due to the fast phase speed being related to the free Kelvin wave.

C). $1.5T_o > T_p + (T_f - T_p)T_k/T_f > 0.5T_o$ for Characteristic C (the wavenumber-one structure of the zonal wind).

Here, $(T_f - T_p)T_k/T_f = (L_0 - c_f T_p)/c_K$, where L_0 is the distance around the globe, c_f is the velocity of the forcing ($c_f = L_0/T_f$), and c_K is the phase velocity of the free Kelvin wave ($c_K = L_0/T_k$). The phase of the zonal wind propagates with the forcing for the period of T_p , and with the free Kelvin mode over the remaining longitudinal section ($L_0 - c_f T_p$), when the contribution from Rossby modes is neglected. Taking this into consideration, the time required for the phase of the zonal wind to move around the earth is $T_p + (L_0 - c_f T_p)/c_K$. For a wavenumber-one structure, this must be less than $1.5T_o$ and larger than $0.5T_o$.

D). $T_f \gg pT_d$ for Characteristic D (the appearance of MG patterns).

The reason for this is obvious. MG patterns originally are a horizontal structure in response to steady forcing (see also Yamagata, 1987).

E). $pT_d \gg T_k$ for Characteristic E (the appearance of Kelvin waves).

The reason for this condition is also obvious. If the damping is too large, a mode appears in which the zonal wind and the geopotential are not in phase but in quadrature.

F). $3T_p T_k/T_f > T_d$ or $3T_p T_k/T_f > T_o - T_p$ for Characteristic F (the nonappearance of Rossby modes west of the forced region).

This is explained in Appendix B. Condition G) is also considered in this Appendix.

Each experiment is checked as to whether or not it satisfies the above conditions, employing the 5 time scales adopted for each experiment. However, there is one problem concerning T_f . As stated in the explanation of Fig. 1, the deviation of heating from the time mean moves faster than the total heating, and such deviation fields are suitable for demonstrating the structure such as shown in Fig. 2. Taking this into consideration, we use $0.9T_f$, instead of T_f , as the first time scale. The factor, 0.9, is used for simplicity, since strict estimation of the factor would be difficult. The condition of "much larger" appearing

Table 2. Five time scales and their typical magnitudes. All units are given in days. References for T_f in the real atmosphere are Weickman *et al.* (1985), Lau and Chan (1985), Nakazawa (1986) and Knutson *et al.* (1986).

	Definition	model	real
T_f	time taken for forcing to propagate around the earth	120	90–150
T_o	period of the intraseasonal oscillation	40	30–60
T_p	duration time of large-amplitude forcing	26	
T_k	time taken for Kelvin modes to propagate around the earth	13	10–20
T_d	damping time constant	8	

in A), B), D) and E) was checked by comparing the left hand side with 1.5 times the right hand side.

The results are listed in Table 1. Overall, the experiments which display all the characteristics of the ISO satisfy all the conditions, except for Characteristic G. Experiments which do not exhibit a given characteristic do not satisfy the corresponding condition; in some of these experiments, other characteristics do not appear, either. For example, in Experiment J, Characteristic C, in addition to F, cannot be found although the unsatisfied condition is only F. This is obviously due to Rossby modes masking a wavenumber-one structure of the zonal wind. The correspondence is not complete only in a few experiments.

Thus, it can be concluded that the correspondence between the characteristics and the conditions is excellent. It is necessary that the 5 time scales satisfy these 6 conditions (and a condition for Characteristic G), in order to reproduce the observed structure of the ISO in a numerical model.

Now it is possible to interpret various structures appearing in previous studies, from the general view obtained above. Yamagata and Hayashi (1984) employed a localized heating, which means $T_f = \infty$ when the present notation is used. This necessarily violates Condition F). In fact, their model shows a conspicuous westward movement of the zonal wind associated with a Rossby mode, west of the forced region. Lau and Lau (1986) obtained an ISO of about 30 day period in their GCM. According to their results, the structure of the geopotential is not wavenumber zero but one. This may correspond to Experiment G in this study. In the actual atmosphere, N89 reported that there were a few cases in which the surface geopotential exhibited eastward-moving characters. Although we have had no information about T_f , T_o and T_p in these cases, the combination of these time scales is possibly related to this behavior of the geopotential.

b. Discrepancies between the model results and observations

We will discuss the two discrepancies between the model results and observations. First, our model shows that amplitudes of the upper-level tempera-

ture and geopotential are large even over the unforced region, while observed amplitudes exhibit marked signals only over active cumulus regions. This discrepancy may be interpreted as follows. In the present model, cumulus convection is prohibited in the unforced region, with high temperatures being produced by downward motion. However, in the actual atmosphere, cumulus convection occurs even over “unforced regions” (from 180° eastward to 60°E), which results in high temperatures. Therefore, areas with average upward motion (*e.g.*, South America or Africa) have high temperatures (Newell *et al.*, 1972). When the ISO cycle is added over these regions, cumulus convection is generally suppressed in the phase of strong downward motion. Thus, downward motion associated with the ISO may not create high temperatures in the real atmosphere, and the temperature field may exhibit no signal of the ISO over “unforced regions”.

The second discrepancy is that amplitudes of the upper-level zonal wind in our model are small over the unforced region, while observed ones are large. Gutzler and Madden (1989) suggested that large amplitudes over “unforced regions” might be caused by the energy propagation from the extratropics through a wave guide of the westerly wind. If this is true, it is natural that this model does not exhibit large amplitudes over the unforced region, because there is no interaction between the tropics and extratropics in this model.

6. Summary

An attempt has been made to understand the complicated structure of the 30–60 day intraseasonal oscillation (ISO) from a unified point of view, using a simple numerical model. The model is based on the linear primitive equations on an equatorial β -plane. The heating is imposed, having a 40 day period with variable amplitudes, and moves eastward over a distance of 120°. The following structure of the heating is also assumed: one group of cumulus clusters in the zonal direction (*i.e.*, wavenumber one), and latitudinally symmetric about the equator. Motions are excited as the response to this external forcing.

This model simulates well the observed structure.

Results show that the variables characterizing the ISO can be divided into two groups according to the difference in their behavior: The heating, zonal wind, divergence, and vertical velocity comprise the first group, while the temperature, geopotential, and surface pressure make up the second group. The first group exhibits a wavenumber-one structure, while the second group has a wavenumber-zero structure. The difference between the two groups, together with other features of the ISO, can be understood as follows.

The clue to understanding the structure lies in the contrast between the slow moving speed of the forcing and the fast group (or phase) velocity of free Kelvin waves. The latter promptly transfers signals occurring in the forced region to the unforced region. Downward motion (lower-level divergence and upper-level convergence) occurs everywhere in the unforced region, when the heating is strong to produce strong upward motion in the forced region. Therefore, the vertical velocity and divergence fields must have a wavenumber-one structure. Since the zonal wind is the integral of the divergence in the longitudinal direction when the meridional wind is small, it must also exhibit a wavenumber-one structure. On the other hand, the temperature field must have a wavenumber-zero structure, since high temperatures are created by strong heating in the forced region and by strong downward motion in the unforced region, which occur almost simultaneously. The geopotential has behavior similar to the temperature via the hydrostatic equation. Thus, the relation between the wavenumber-one zonal wind and the wavenumber-zero temperature (geopotential) is not accidental but necessary.

Marked structural changes occur in one cycle of the ISO. The slowly-moving heating excites a Matsuno-Gill pattern. Then, free Kelvin and Rossby modes separate in the eastern portion of the forced region, where heating amplitudes become small. The Kelvin mode moves eastward at a fast speed. The Rossby mode moves westward in the forced region, but its amplitude is small, compared with that of the MG pattern. As a result, the MG pattern can be seen in the forced region and the Kelvin mode in the unforced region. The phase of the zonal wind moves slowly in the forced region, being out of phase with the geopotential, while it propagates at a fast speed in the unforced region, where it is in phase with the geopotential. Thus, the zonal wind propagates around the globe within the period of the ISO. Further, it can be understood from structural changes that the behavior of the geopotential is very different from that of the zonal wind.

Moreover, it can be proven that the barotropic mode of the zonal wind has a negligibly small amplitude, while the barotropic component of the geopotential has an amplitude of the same order magni-

tude as the baroclinic component. This difference in turn causes the contrast in the vertical structure. The zonal wind shows a complete baroclinic structure, while the geopotential does not show a completely baroclinic structure.

In the simulation of the ISO, parameter values are severely restricted. With the 40 day period of the ISO, the observed structure can be reproduced when the damping time constant and the duration time of large-amplitude heating are about 8 days and around 25 days, respectively. Furthermore, the longitudinal scale of the cumulus cluster must not be too small.

Five time scales were defined; the time required for the forcing to propagate around the earth (T_f), the period of the ISO (T_o), the duration of large-amplitude forcing (T_p), the time required for the free Kelvin wave to move around the earth (T_k), and the damping time constant (T_d). Six conditions were proposed which involve relationships among these time scales. These conditions must be satisfied to reproduce the typical characteristics of the observed structure. A typical case in which all the conditions are satisfied is $T_f=120$ days (forcing velocity is about $4 \text{ m}\cdot\text{sec}^{-1}$), $T_o=40$ days, $T_p=25$ days, $T_k=15$ days, and $T_d=8$ days.

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Appendix

A. Observed relation between OLR and the surface geopotential in the ISO, and the static stability in the tropics

The relationship of OLR to the surface geopotential is examined, since this relationship is important to determine whether or not a model simulates the ISO well. Bandpass-filtered data of OLR and the geopotential height at 1000 mb (H1000) over three stations are used for this purpose. As for the filtering procedure, refer to N89. The three stations are Kuala Lumpur, Koror and Kwajalein, which are representative of the maritime continent, the western Pacific, and the mid Pacific, respectively.

Figure A shows the time variations of H1000 at the three stations and the OLR during 1979. Similar

Fig. A. Time series of 1000 mb geopotential heights (H1000) at the three stations (dashed line) and OLR averaged near the stations (solid line) during 1979: (a) for Kuala Lumpur (3.12°N, 101.55°E), (b) Koror (7.33°N, 134.48°E) and (c) Kwajalein (8.73°N, 167.72°E). The OLR is averaged between 10°S~10°N and three different longitudinal ranges, *i.e.*, (a) 80°E~120°E, (b) 110°E~150°E, and (c) 140°E~180°. A band-pass filter for the 30~60 day period has been applied to these data. The abscissa is given in days from Jan. 1. Units are m for H1000 and $\text{W}\cdot\text{m}^{-2}$ for OLR.

results can be obtained for other years. H1000 at Koror is in phase with the OLR averaged over 110°E~150°E. This means that the low values of surface geopotential correspond to strong convection. On the other hand, H1000 at Kuala Lumpur lags behind the OLR averaged over 80°E~120°E, while H1000 at Kwajalein leads the OLR averaged over 140°E~180°. This analysis is consistent with the fact that the phase velocity of the surface geopotential is fast, while that of OLR is slow.

Next, the static stability is calculated. The data during 1979 from the 33 stations which were analysed by N89 are used. The stability of a lower layer is calculated at 850mb, and the average value of the all stations is found to be $4.50 \times 10^{-4} \text{ K}\cdot\text{Pa}^{-1}$. On the other hand, the representative upper layer is 200 mb, where the averaged stability is $5.95 \times 10^{-4} \text{ K}\cdot\text{Pa}^{-1}$. Therefore, the ratio, S_1/S_3 , is 1.32. Differences from station to station are not very large; the minimum ratio is 0.9 at Kuantan, and the maximum ratio is 1.8 at Fernando. Also, time variations are not very large, and do not seem to be related to the ISO (not shown).

B. Derivation of Condition F) and consideration of G)

Figure B is a schematic of a time-longitude sec-

tion of cumulus heating used to consider Conditions F) and G). T_o is the period of the ISO, and T_p is the persistent time of large-amplitude heating, as defined in Section 5. L.C. expresses the half length of a cumulus cluster. A~F (including E_1 and F_1) are "points" indicating both time and space, while X_L and X_R denote spatial points. The center of the cumulus cluster therefore moves along line AB, and the line between C and D corresponds to the center of strong heating. When the time mean of the heating is subtracted, cool anomalies appear. The regions of strong cooling are shaded.

First, consideration is given to the first part of Condition F). If a Rossby mode, which separates from the MG pattern at point D, passes through point X_L with large amplitudes, the influence of the Rossby mode can be seen in the region west of the forced region. This situation must be avoided in the simulation of the ISO. The condition that the Rossby mode has negligibly small amplitudes is that the time required to travel from the separation point, X_R , to X_L is longer than the damping time constant. Therefore, the condition should be $c_f T_p / c_R > T_d$, where c_R is the phase speed of the Rossby mode and $c_R \approx c_K / 3 = L_0 / (3T_k)$. Further, $c_f = L_0 / T_f$. Thus, we can obtain the condition as

Fig. B. Schematic figure of heating in a time-longitude section. See the text for detail.

$$3T_p T_k / T_f > T_d. \quad (\text{B1})$$

It is easy to obtain the second part of Condition F). If the forcing amplitude in the subsequent ISO cycle becomes large before the separated Rossby mode passes through X_L , the Rossby mode cannot be detected west of the forced region. The condition is therefore

$$3T_p T_k / T_f > T_o - T_p. \quad (\text{B2})$$

Next, we consider Condition G). For this purpose, Eq. (23) as well as Fig. B are used. Further, note that maxima of the zonal wind fall along line EE_1 , while minima (easterlies) fall along F_1F , as is easily understood from Eq. (22). For a wavenumber-zero structure of the temperature, maxima of T_+ must move around the equator within $T_0/2$.

First, we will consider the movement for maxima of T_+ in the forced region. In the case that forcing amplitudes are almost constant between C and D in Fig. B, it is easy to predict the behavior of T_+ . When $\partial u / \partial t \gg \alpha u$ in Eq. (23), it can be easily understood that the maximum of T_+ at point X_R occurs at F, since the sign of $\partial u / \partial t$ changes from negative to positive at F. Also, the time when T_+ reaches a maximum at point X_L is C, because $\partial u / \partial t$ is negligible before C, and becomes large at C. Maxima of T_+ therefore move along line CF. On the other hand, in the case of $\alpha u \gg \partial u / \partial t$, it can easily be speculated that maxima of T_+ exist along line CD. However, when forcing amplitudes largely vary with time as in most experiments in this study, it is difficult to specify the behavior of T_+ . In the case of $\partial u / \partial t \gg \alpha u$, the maximum of T_+ at point X_R occurs at F as in the previous case. On the other hand, the time of maximum T_+ at point X_L is ambiguous, since $\partial u / \partial t$ becomes large still after C. It is possible that

the maximum is attained after C. Since cooling appears after E, the latest possible time of maximum T_+ at point X_L is E. Therefore, the line of maximum T_+ will exist between CF and EF. When $\alpha u \gg \partial u / \partial t$, the amplitude of T_+ at D is about $1/e$ of the maximum which obtains near the center of CD. Therefore, T_+ may reach the maximum at point X_R before D. Presently, we have not been able to interpret the behavior of T_+ rigorously in the forced region.

In the unforced region, maxima of T_+ propagate at the speed of the free Kelvin wave. The time taken from X_R eastward to X_L is about $(L_0 - c_f T_p) / c_K = (T_f - T_p) T_k / T_f$ (see Condition C).

Thus, the condition for a wavenumber-zero structure of the temperature which we can obtain at the present time is (the time taken for maxima of T_+ to go across the forced region) + $(T_f - T_p) T_k / T_f < T_0/2$.

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季節内振動の構造に関する考察

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熱帯の季節内振動 (30–60 日周期振動) は非常に複雑な構造を持っている。例えば、東西風成分は波数 1 の構造を持つのにに対し、気温や高度場は波数 0 の構造を持っている。この論文の狙いは統一的な見地から季節内振動の構造を理解することにある。この目的のため、線形プリミティブ方程式系に基づく数値モデルを用いる。加熱は外部的な強制として与えた。その振幅は時間変動し、 120° の経度を周期 40 日で東進する。このモデルは季節内振動の構造をうまく再現する。

季節内振動の構造を理解する鍵は、加熱強制の遅い移動速度とケルビン波の速い群速度との対照にある。上昇流域は強制とともにゆっくり移動するが、下降流域はケルビン波の働きで速やかに全経度に広がる。すなわち、鉛直流場と発散場は波数 1 の構造を持つ。東西風は発散の経度方向の積分なので、これも波数 1 となる。一方、温度場は波数 0 の構造を持つことが分かる。なぜなら高温は、強制域では非断熱加熱によってもたらされるが、非強制域では下降流によってもたらされ、非断熱加熱の最大と下降流の最大はほぼ同時に起こるので。高度場も静力学の関係から温度場と同様の構造を持つ。このように、東西風成分における波数 1 の構造と気温・高度場における波数 0 の構造は必然的な関係にある。

季節内振動の 1 サイクルにおいてその構造は顕著な変動を示す。強制の移動速度は遅いので松野-Gill パターンを励起するが、強制の振幅が弱くなるとケルビン波が分離し東進する。このように強制域では遅い位相速度をもつ松野-Gill パターンが見え、非強制域では速いケルビン波を見ることになる。さらに高度場の鉛直構造は東西風のそれとは異なるべきであることも示すことができる。

モデルの種々のパラメータに対して季節内振動の構造がどのように依存するかについても調べた。最後に、季節内振動を特徴づける 5 つの時間スケールを提案した。観測される構造を再現するためには、これらの時間スケールの間でいくつかの関係式が満足されなければならない。