Part III: Monsoon Dynamics

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ABSTRACT

This is the third in a series of papers to study the origin of intraseasonal oscillations. In this paper, we address
the issue of the interaction of the monsoon large-scale circulation and intraseasonal oscillations. We show that
as a result of the interaction of the large scale monsoon flow with the near-equatorial intraseasonal oscillation,
unstable baroclinic disturbances are generated over the monsoon region. These disturbances have spatial scales
of approximately 3000-4000 km and periods of 3-6 days with the vertical wave axis tilting eastward with height.
The rapid development of these cycloonal disturbances along 15°-20°N and the concomitant weakening of the
equatorial disturbances are accompanied by the rapid northward shift of the rising branch of the local Hadley
circulation. They may also be identified with the observed sudden jump of the Mei-yu rainband over East Asia
and the inverse relationship between the monsoon ITCZ and the equatorial ITCZ over India and East Asia.

From a linear stability analysis of quasi-geostrophic motion in a two-level model, it is shown that the westward
propagating disturbances generated over the monsoon region are the manifestation of heat-induced unstable
Rossby waves. The instability is favored in the region with large vertical wind shear and reduced effective static
stability. The monsoon large scale circulation over India and southeast Asia and the plentiful supply of moisture
in the region appear to be favorable for the development of these unstable waves. It is argued that the prevailing
easterly waves found over the subtropical western Pacific during northern summer may also be due to the above
unstable Rossby wave mechanism.

1. Introduction

In Part I (Lau and Peng 1987) of this paper, we proposed a simple mechanism for the origin of intra-
seasonal oscillations in the tropical atmosphere. This mechanism referred to as mobile wave-CISK involves
the excitation of unstable internal modes arising from a positive feedback between low-level moisture con-
vergence and condensational heating. The most dominant unstable mode consists of an eastward propa-
gating interactive heat source and circulation pattern that resembles the 30-60 day oscillations. This mode
appears also to fit the description of “superclusters” or intrinsic circulation modes found in “aqua-planet”
GCM experiments of Hayashi and Sumi (1986) and many other similar GCM experiments (Swinbank et
al. 1988; Lau et al. 1988). In Part II (Sui and Lau 1989), the effect of evaporation and surface heating on
the basic mechanism is studied. It was found that differential surface heating and evaporation due to the
zonal sea surface temperature gradient is responsible for the large zonal variation in convection associated
with the 30-60 day oscillations. While a number of issues still need to be resolved, present theories (e.g.
Parts I and II of this paper; Wang 1988; Itoh 1988; Chang and Lim 1988; Lau et al. 1988; and many oth-
ers) all support the idea that the feedback between convergence and convection may be a key to the pre-
dominant eastward propagation and time scale of intraseasonal oscillations.

An important question not addressed by all the aforementioned studies is the meridional propagation
of the 30-60 day oscillations. It has been known for a long time that strong intraseasonal oscillations are
found over the monsoon regions, where the dominant mode of variation is a northward propagation or sud-
den northward shift of convection and rainfall associated with the onset and break of the northern summer
monsoon (Yasunari 1982; Krishnamurti 1985; Lau and Chan 1986; Murakami and Nakazawa 1985; Lau
et al. 1988; Chen et al. 1988). In fact, it is this feature that led earlier investigators to argue that the origin of
the 30-60 day oscillation is in the zonally symmetric circulation associated with monsoon physics, i.e., land-
sea contrast, soil moisture, radiation–moisture–cloud feedback, etc., (Webster 1984; Goswami and Shukla 1984; Anderson and Stevens 1987). In reality, intraseasonal oscillations over the monsoon region are much too complex to be described by a simple zonally symmetric circulation. For example, Murakami et al. (1984) showed that as the monsoon trough migrates northward, westward-propagating synoptic scale cyclones are frequently observed along the trough axis. Krishnamurti et al. (1985) noted that abrupt changes related to monsoon break and revival are associated with the interaction of eastward propagating 30–60 day waves and westward propagating synoptic scale disturbances. Nakazawa (1986) showed that while eastward propagating 30–60 day cloud clusters prevailed in the equatorial region, the cloud clusters generally move westward over the subtropics in the Pacific and East Asia/India during the summer monsoon season.

Recent theoretical works so far suggest that the origin of the 30–60 day oscillation need not be in the monsoon region because monsoon dynamics are not required to explain the basic time scales and eastward propagation of the oscillations which occur along the equator all year round (Lau and Chan 1986). Thus, present theories of the 30–60 day oscillation face a paradox. If the monsoon is not the origin of the oscillation, why is there such strong signal over the monsoon region? Until this paradox is resolved, a comprehensive theory of the 30–60 day oscillation is impossible. In this paper, this issue will be addressed.

Given that the large-scale monsoon mean flow is driven by thermal land–sea contrast, it is conceivable that an interaction of the basic large-scale monsoon flow with the near-equatorial, eastward propagating 30–60 day oscillation may excite new forced migratory modes of the system, which has the observed structure and propagation over the monsoon region. When averaged over a large zonal domain, or viewed in some time-averaged sense, the above will be manifested in the form of a northward shift of the axis of ITCZ or monsoon trough in some zonally symmetric sense. We shall investigate the above hypothesis using the basic theoretical framework developed in Parts I and II. In section 2, a brief description of the model is given. In section 3, we revisit the basic mobile wave-CISK concept to focus on features relevant to monsoon dynamics. The results of experiments designed to test the above hypothesis will be presented in sections 4 and 5. In section 6, a simple theoretical analysis is presented to interpret the numerical experiments. In section 7, we present the conclusions and discuss the implication of the present results on future work.

2. The model

The numerical model is the same as that used in Parts I and II. The model is a 5-level global spectral model with rhomboidal truncation to wavenumber 15. The only physics in the model consists of a positive-definite wave-CISK heating function with the total heating proportional to the 990 mb moisture convergence. It was shown in Part I, that the positive-definite heating is essential in maintaining the growth and development of intraseasonal oscillations in the model. In Parts I and II, no basic flow is considered. In this paper, we shall adopt the same approach as before but emphasize the dynamical interactions of the 30–60 day disturbance with the monsoon mean flow. For the present study, a specified vertical heating profile is used. For more detailed study including the interaction of the 30–60 day oscillation with the monsoon convection and surface conditions, an internal self-determining heating profile including boundary layer processes (cf. Part II) will be necessary. These are ongoing research outside the scope of this investigation.

3. Mobile wave-CISK modes and vertical heating profile

One of the key results of Parts I and II is the sensitivity of the phase speed of intraseasonal oscillations to the vertical heating profile. In the proposed theory, the time taken for the oscillation to go around the equatorial belt once sets the basic period of the oscillation. This period is strongly dependent on the vertical heating profile. It was shown in Part II that an eastward propagating “fast” or “slow” mode is excited respectively when the level of maximum heating is above or below the 500 mb level. While the phase speed of the fast mode (~20 m s$^{-1}$) and the slow mode (~10 m s$^{-1}$) differs substantially, the qualitative features of the circulation are similar, except that the slow mode has a shallower vertical circulation (for details see Part II). While recent observations tend to put a heating maximum higher in the tropical convection due to mesoscale effects (Hartmann et al. 1984), there are reasons to believe that diabatic heating directly relevant to large-scale low level convergence, i.e., heating in the core of the convection, may actually be lower than hitherto believed. For example, Johnson (1984) partitioned the total heating profile into that due to stratiform rain (mesoscale effect) and convective rain (wave-CISK effect) and found that the maximum of heating due to convective rain is between 700 and 500 mb. Given that there is still large uncertainty about the form of the heating profile appropriate for the 30–60 day oscillation, we have carried out identical experiments for both the fast mode (heating maximum at 300 mb) and for the slow mode (heating maximum between 700 and 500 mb). Since the qualitative nature of the results are essentially the same for both, we only discuss results for the slow mode, which corresponds to an approximately 50-day oscillation.

To aid the reader, the following is a simplified analysis of the mobile wave-CISK concept. Those who are familiar with the concept may proceed to section 4.
For simplicity we consider only moist Kelvin waves. The vertical modes of a multilevel system can be written as

\[ \frac{\partial u_i}{\partial t} = -\frac{\partial \phi_i}{\partial x} \]

\[ \frac{\partial \phi_i}{\partial t} + c_i \frac{\partial u_i}{\partial x} = Q_i \]  

(1)

where \( Q_i \) is a heating function and \( c_i \) is a phase speed related to the static stability for the \( i \)th mode. The static stability is prescribed as the mean condition of the tropical atmosphere. If we assume the internal heating is proportional to the dynamic convergence, then

\[ Q_i = - \sum_j m_{ij} \frac{\partial u_j}{\partial x} \]  

(2)

where \( m \) is the precipitation efficiency factor and \( \lambda_{ij} \) is a function of the vertical heating profile. Equation (1) can be combined to obtain a single equation for a system of coupled linear oscillators:

\[ \sum_j \left[ \delta_{ij} \frac{\partial^2 u_j}{\partial t^2} - (\delta_{ij} c_i^2 - m \lambda_{ij}) \frac{\partial^2 u_i}{\partial x^2} \right] = 0. \]  

(3)

From (3), the effect of the interactive heating is twofold. First, it effectively reduces the static stability of the system by reducing the phase speed of each of the original vertical modes. Second, it couples all of the original vertical modes to produce new vertical modes—the mobile wave-CISK modes. If we assume a solution of the form \( \exp \{ i(kx - ct) \} \), then the new phase speed of the system can be solved as a complex eigenvalue problem for the phase speed \( c \). In a linear analysis, the shortest scales are most unstable and will eventually cause the numerical integration to break up. However, as shown in Part I, this tendency is suppressed by nonlinear effect present in the positive-only heating function. The linear phase speed obtained above agrees very well with that obtained by numerical integration using the initial value approach (see Parts I and II for details). Tables 1a and 1b show the variation of the mobile wave-CISK modes as a function of the precipitation efficiency \( m \) for the fast and slow modes, respectively. For the deep heating profile shown in Table 1a, “fast” unstable modes occur at \( m = 0.6 \) when the third and the fourth modes coalesce to form a growing and a decaying mode with identical phase speed. The speed of the marginally unstable wave is about 22 m s\(^{-1}\), corresponding to a 24-day oscillation. For the “shallow” heating profile (Table 1b), the most unstable modes are formed from the interaction of modes 4 and 5. The phase speed of the marginally unstable mode which occurs at \( m = 0.4 \) is 10.5 m s\(^{-1}\), corresponding to a 44-day oscillation. From Tables 1a and 1b, it can be seen that the larger the precipitation efficiency (i.e., the stronger the interaction between convective heating and low level convergence), the slower the unstable mode propagates and the more unstable it becomes. This is a feature true for all reasonable ranges of heating profile we have so far tried. In addition, the mobile wave-CISK modes are more unstable and the phase speeds are more sensitive to \( m \), for the shallower heating profile.

4. The summer monsoon experiment

a. The mean summer monsoon flow

The three-dimensional mean summer monsoon flow is simulated by prescribing a fixed heat source (no wave-CISK) centered at \( \phi_0 = 22.5^\circ \)N and \( \lambda_0 = 180^\circ \):

\[ Q = \begin{cases} Q_0 \eta(p) \cos \left( \frac{\pi}{2} \left( \phi - \phi_0 \right) \right) \cos \left( \frac{\pi}{2} \left( \lambda - \lambda_0 \right) \right), & \text{for } |\phi - \phi_0| < 22.5^\circ \text{ and } |\lambda - \lambda_0| < 45^\circ, \\ 0, & \text{otherwise.} \end{cases} \]

The vertical profile \( \eta(p) \) corresponds to that of the “slow” mode given in Table 1. Figure 1a shows the steady state 300 mb wind field generated by the above heat source. The main “monsoon” feature is the large anticyclone centered slightly to the west of the heat source, the strong tropical easterly jet (TEJ) to the south of the heat source. The maximum wind in the TEJ is about 17 m s\(^{-1}\). A weak cross-equatorial flow (~ 1 m s\(^{-1}\)) is also noted in that region. Elsewhere, the flow is
rather weak. The flow at 700 mb (Fig. 1b) is almost an exact mirror image of the 300 mb flow. The most conspicuous feature is the strong cyclonic flow around the heat source with strong westerly flow in the region between longitude 120° to 180° near 0°–20°N. The maximum westerlies is about 10 m s⁻¹. For convenience, we shall refer to the area between 120°–200° longitude and 0°–35°N where the winds are strong as the model “monsoon region.” Figure 2 shows the meridional cross section of the mean monsoon flow near the center of the heat source at 180° longitude. There is strong rising motion over the heat source at 20°N and subsiding motion to the north and south, indicating a simple heat-induced Hadley-type circulation centered over the monsoon region with weak subsidence over the equator.

b. Interaction with equatorial intraseasonal disturbances

In the following, all quantities shown are for the perturbation field with the monsoon basic flow as shown in Figs. 1 and 2 held fixed.

1) TIME-LONGITUDE SECTION

With the mean flow and the monsoon heat source fixed, the initial disturbance is specified as an equatorial “supercluster” centered at 0° longitude, which is generated by the mobile wave-CISK mechanism without mean flow. The value of $m = 0.6$ is used and the heating profile used is shown in Table 1b. The “supercluster” has an eastward propagating speed of about 10 m s⁻¹.
(Fig. 3a) which is very close to that predicted by the linear analysis (Table 1b). Figure 3 shows the time-longitude section of the precipitation pattern associated with the equatorial disturbance where Day 0 corresponds to the time when the supercluster is at 60° longitude. At the equator (Fig. 3a), the precipitation is dominated by that associated with the supercluster. There are secondary but much weaker precipitation bands trailing behind the supercluster. These secondary precipitation patterns are associated with transient circulation features generated away from the supercluster.

As the supercluster approaches the monsoon region, it is seen to weaken considerably while new westward propagating precipitation patterns appear over the monsoon region. The temporary weakening of the equatorial disturbance is due to the development of the rising branch of the local Hadley circulation in the monsoon region, which draws equatorial low level air and hence reduces the equatorial convergence and destroys the symmetry (about the equator) of the circulation that favors Kelvin-wave type disturbances. At day 20, a new eastward propagating feature appears to re-emerge from the monsoon region.

There is a dramatic development over the monsoon region, as a result of the approach of the equatorial disturbance (Fig. 3b). At 16°N, successive precipitation centers are generated over the monsoon region as the equatorial supercluster approaches. While the generation of the new centers progresses eastward, each center intensifies and migrates westward with a speed approximately the same as the equatorial eastward disturbance. The amplification of these westward propagating precipitation patterns suggests an exponential increasing rate of e-folding time of a few days. By Day 20, the maximum rainfall has reached an unrealistic value of over 100 mm day⁻¹. This is because the present dynamical model is basically linear (except for the nonlinearity in the heating parameterization) so that finite amplitude oscillation cannot occur for unstable modes. Thus, in this and subsequent discussions, only the phase of the disturbance should be emphasized. The amplitude of the disturbance near the end of the integration is unrealistically large and should not be taken too seriously. Figures 4a and 4b shows the variation of the 300 mb zonal wind. The changes in the wind field are consistent with those for the rainfall.

**Fig. 2.** Latitude–height cross section showing the streamline pattern of the Hadley circulation over the monsoon region. The scale of the wind vector in the vertical is greatly exaggerated.

**Fig. 3.** Time–longitude section showing the pattern of latent heat release from the model-generated disturbances (a) at the equator and (b) at 16°N. Contour interval is in 1 mm day⁻¹.
southwesterly flow are found on the equatorward side of the leading disturbance. By Day 18, three distinct
cyclones are formed over the monsoon region. Each of the cyclonic center is accompanied by a rainfall cen-
ter slightly to its east (Fig. 5).

3) ZONAL-VERTICAL STRUCTURE

The vertical structure and the propagation of the above cyclonic disturbances are best depicted by the
height–longitude cross section of the meridional ve-
locity component. Such cross sections along 16°N for
the last eight days of the integration at 2-day intervals
are shown in Fig. 7. Each of the disturbances, denoted
by A, B, C has a distinct baroclinic structure with an
eastward tilt with height and a reversal of the meridi-
onal wind at upper and lower levels. This orientation
is favored for the disturbance to propagate westward
and gain energy at the same time from wave-CISK
heating. Each disturbance appears to develop in the
region near 180° longitude, amplifies and decays as it
propagates westward while a new disturbance is gen-
erated upstream near the source region. Within the
region of the steady monsoon heat source and strong
TEJ (150° to 180° longitude), the transient distur-
bance amplifies dramatically (disturbance C). Further
downstream, in the region 120°–150° longitude, the
disturbance decay rapidly (disturbances B and A). The
maximum amplitude of these disturbances are found
near the middle troposphere. From the westward
propagation speed and the spatial scale, it is estimated
that the periods of these disturbances are of the order
of 4–6 days.

4) MERIDIONAL-VERTICAL STRUCTURE

Figure 8 shows the latitude–height section of the
zonal wind at 160° longitude from Day 12 to 18 at
two day intervals. It is seen that large anomalies appear
to originate at upper levels near the region of the upper
TEJ. As the equatorial disturbance approach the mon-
soon heat source (Day 12–14), localized fluctuation with alternate easterlies and westerlies appear at both
upper and lower levels. This is associated with the vari-
ation of the cyclones in the middle and lower levels
noted earlier. By Day 16, the interaction at the TEJ
appears to be strongest. At Day 18, when the precipi-
tation anomaly is fully developed at 20°N, the westerly
and the easterly region associated with a cyclonic cir-
culation are well developed around the center of the
heat source. At this time, the zonal flow is very well
developed in the lower and middle troposphere near
the region of maximum heating while at 200 mb, the
flow is relatively weak. The rapid northward shift of
precipitation and development of the zonal wind pattern
around the heat source with the above vertical
structure again suggest the excititation of unstable mobile
wave-CISK modes in the monsoon region.

field. At the equator, the dramatic changes in the wind
field as the supercluster approaches the monsoon re-
region, occur rather abruptly. In about 4–5 days, the
westward propagating disturbances are fully developed.
At 16°N, the heavy rainfall is associated with westward
moving synoptic-scale disturbances induced by the
approaching equatorial disturbance.

2) HORIZONTAL STRUCTURE

Figure 5 shows the sequence of rainfall maps every
two days apart, starting from Day 8 when the equatorial
disturbance enters the monsoon region. From Day 10,
precipitation centers begin to emerge at about 20°N
in the monsoon region downstream of the equatorial
center, which begins to weaken. By Day 16, the center
of action has shifted to the region of the monsoon heat
source where two rainfall centers are observed. It
appears that these centers are formed at the eastern part
of the monsoon region near the ascending branch of the
mean Hadley circulation (near 180°) and then mi-
grate westward and dissipates as it moves out of the
monsoon region. By Day 18, rainfall over monsoon
heat source region has reached unrealistically large val-
ues. Similar maps for the horizontal wind field at 700
mb (Fig. 6) show that the initial disturbance is main-
tained by strong zonal wind convergence near the
equator in Day 8–10. As the equatorial disturbance
weakens, westward propagating cyclonic vortices are
generated over the monsoon region. Strong low level
Fig. 5. Sequence of rainfall maps at selected days indicating the approach of the equatorial disturbance and the onset of monsoon rainfall at 20°N. Due to large precipitation changes, different contour intervals are used for each map. They are 0.5, 1, 2, 4, 8 and 16 mm day$^{-1}$, respectively, in order of the display.
Fig. 6. Sequence of 700 mb wind, showing the monsoon cyclonic development associated with the rainfall pattern in Fig. 5. The wind vectors have been scaled in proportion to the maximum wind for each map. The maximum wind magnitudes are approximately 1.5, 2.0, 4.0, 10.0, 15 and 25 m s$^{-1}$ in order of the display.
The local Hadley circulation associated with the above zonal wind changes at 180° longitude is depicted in Fig. 9. The vertical circulation for Day 6 indicate sinking motion at the equator and all other latitudes. This corresponds to the descending branch of the equatorial Walker-type circulation leading the initial supercluster when it is far from the monsoon region. At Day 12, when the equatorial disturbance enters the monsoon region, a Hadley-type circulation is generated, with two ascending branches, over 5°S and 15°N, respectively. The northward shift of the heating center is complete by Day 18, when a single Hadley-type cir-

Fig. 7. Longitude–height cross section of the meridional wind along 16°N at selected times showing the vertical structure and propagation of the monsoon transient disturbances. Units: m s⁻¹.

Fig. 8. Latitude–height cross section of the zonal wind at selected times showing the vertical structure of the monsoon transient disturbance. Contour intervals in 0.5, 0.5, 1.0 and 2.0 m s⁻¹, respectively.
Fig. 9. Vertical streamline pattern before (Day 6), during (Day 12) and after (Day 20) the passage of the intraseasonal disturbance at 180 longitude. The scale of the vertical velocity is arbitrary and greatly exaggerated.
5. The winter monsoon experiment

a. Winter monsoon mean flow

In the following, we present the results of an identical experiment as discussed above but with the mean monsoon circulation exactly reversed. This reversed flow will be referred to as the "winter monsoon" mean flow and corresponds to that generated by a heat sink of the same absolute magnitude as (5) but with the sign reversed. It should be pointed out that the main objective here is not to study the features of the real winter monsoon flow, which is far from a mirror image of the summer monsoon flow, but rather to further delineate the effect of the mean flow on the heat-induced unstable modes.

b. Interaction with equatorial intraseasonal disturbances

1) TIME–LONGITUDE SECTION

Figures 10a and 10b show the time–longitude section of the precipitation associated with the equatorial 50-day disturbance along the equator and 16°N. Contrary to the summer monsoon case, as the disturbance approaches the winter monsoon region, it intensifies. At 16°N, an eastward propagating but rather weak precipitation center was generated over the monsoon region. The secondary precipitation centers appear only for a brief period from Day 12 to Day 15 and disappear as the equatorial heat source moves out of the monsoon region. The continuous propagation of the equatorial disturbance into and out of the monsoon region is also obvious from Fig. 11a showing the time evolution of the 300 mb zonal wind at the equator. The unchanged phase speed suggests that no new mobile wave-CISK modes are generated. The prevailing mode appears to be amplified by its interaction with the winter monsoon mean flow. Only very weak perturbations are noticed at 16°N (Fig. 11b).

2) HORIZONTAL STRUCTURES

Snap-shots of the precipitation field at 2-day intervals starting from Day 12 to 18 are shown in Fig. 12. The enhancement of the equatorial heat source at 180° longitude near the region of maximum vertical westerly shear is clearly seen. Most importantly, there is no new organized development of convection in the region of the heat sink. The horizontal flow around the propagating equatorial heat source remains largely zonal (Fig. 13). Although perturbation in the largely zonal flow are observed due to transient wave activities over the monsoon region, the Kelvin wavelike characteristic prevailing in the equatorial latitudes remain intact as the supercluster emerges out of the monsoon region at Day 18.
FIG. 12. As in Fig. 5, except for the winter monsoon basic flow. Contour intervals in 0.5, 1.5, 3.0, 6.0, 6.0 and 6.0 mm d$^{-1}$, respectively.
Fig. 13. As in Fig. 6, except for the winter monsoon basic flow. Magnitudes of wind vector is scaled by maximum wind speeds: 1.5, 2.0, 2.0, 2.5, 3.0, 4.0 m s\(^{-1}\), respectively.
3) **LONGITUDE-HEIGHT STRUCTURE**

The eastward propagating rainfall pattern at 16°N (Fig. 10b) appears to be part of the equatorial pattern as the latter expands somewhat laterally as it enters the monsoon region. The height–longitude section (not shown) generally shows a westward tilt with height. In Part I and II, it has been shown that the westward tilt with height is a characteristic feature of the Kelvin wave–CISK mode in the equatorial regions. Thus, vertical structure here may be an extension of the equatorial intraseasonal disturbance as its influence extends somewhat into the monsoon region. However, no new unstable modes are excited for the winter monsoon flow.

4) **LATITUDE-HEIGHT STRUCTURE**

For the winter monsoon case, the largest amplitude of the zonal wind is found within the equatorial region between 10°N and 10°S. The response at higher latitudes is weak and transient in nature (not shown). This is in contrast to the organized zonal wind distribution for the summer monsoon flow where the maximum easterlies are found at around 30°N (Fig. 8). More interesting are the changes in the Hadley-type circulation in the winter monsoon region due to the approach of the 50-day disturbance (Fig. 14). At the equator, initially descending motion prevails at 180° longitude as a result of the approach of the sinking branch of the Walker cell ahead of the equatorial heat source. At Day 12, as the heat source enters the monsoon region, two Hadley cells symmetric about the equator are developed. The circulation itself is basically trapped near the equator with very little amplitude outside of 15°N and S. By Day 18, when the equatorial disturbance just emerges out of the monsoon region, the equatorial Hadley cell is amplified, and secondary Hadley cells are induced north of the monsoon region with weak rising motion at 30°N and sinking motion around 45°N. However, the circulation is very weak and disappears subsequently. In contrast to the summer monsoon case, there is no indication of excitation of unstable modes.

6. **Instability analysis**

The results in section 4 suggest that the westward propagating disturbances generated over the summer monsoon region are unstable baroclinic modes induced by wave-CISK. Structure and evolution of the disturbance suggest they may be related to westward propagating Rossby waves destabilized by mobile wave-CISK. In Parts I and II we have shown moist Kelvin waves are most unstable at the equator and that in the absence of mean flow, Rossby waves are stable. But away from the equator, Kelvin waves are no longer effective and in the presence of mean wind shear, Rossby waves may become unstable. Further, from the baroclinic structure of the disturbances discussed in section 4, it is clear that the classical barotropic instability plays no major role in the development of the disturbances, even though there may be energy exchange between the monsoon mean flow and the finite amplitude disturbances. In order to understand the basic mechanism of the model baroclinic wave disturbance, we shall focus our attention on the moist baroclinic instability in the core region of the strong easterly vertical shear on the southern flank of the monsoon circulation.

a. **A simple, analytic moist atmosphere model**

We seek to carry out instability analysis of the growing modes shown in Figs. 7 to 9 in the simplest possible dynamical framework. Because of the deep vertical structure of the simulated growing modes, the low-level upward motion upon which the CISK parameterization is based is directly related to the midlevel vertical motion. This suggests that a two-level model may be adequate. The two-level model has the added advantage of allowing analytic solutions for the instability analysis. The linearized two-level quasi-geostrophic vorticity and thermodynamic equations with wave-CISK heating are

\[
\begin{align*}
\frac{\partial}{\partial t} + U_1 \frac{\partial}{\partial x} \left( \frac{\partial^2 \psi_1}{\partial x^2} + \beta \frac{\partial \psi_1}{\partial x} \right) &= \frac{f_0}{\Delta p} \omega_2 \\
\frac{\partial}{\partial t} + U_3 \frac{\partial}{\partial x} \left( \frac{\partial^2 \psi_3}{\partial x^2} + \beta \frac{\partial \psi_3}{\partial x} \right) &= -\frac{f_0}{\Delta p} \omega_2 \\
\frac{\partial}{\partial t} + U_m \frac{\partial}{\partial x} (\psi_1 - \psi_3) - U_T \frac{\partial}{\partial x} (\psi_1 + \psi_3) &= -\frac{f_0}{\Delta p} \omega_2 \frac{\lambda^2}{\Delta \xi^2}
\end{align*}
\]

where

\[
U_m = \frac{1}{2} (U_1 + U_3), \quad U_T = \frac{1}{2} (U_1 - U_3)
\]

\[
\lambda^2 = \left( \frac{\Delta p}{f_0} \right)^2 S - \frac{\Delta p}{\rho} \frac{L q_c}{C_p} \frac{\rho_0}{p_2} \frac{\rho^2}{f_0^2} \frac{\ell_0}{C_p}.
\]

The subscripts 1 and 3 denote the upper and lower level respectively, \(U\) is the basic zonal wind, \(\psi\) the perturbation streamfunction, \(\omega_2 = \delta p/\delta t\) at the midlevel, \(S\) the static stability, \(q_c\) the moisture condensation rate per unit mass of ascending air, \(L\) the latent heat of condensation, \(C_p\) the specific heat of air at constant pressure, \(p_2\) the pressure at the middle, \(\Delta p\) the pressure difference between the upper and the lower level and \(f_0\) the Coriolis parameter. The second term on the right-hand side of (8) is the contribution by the release of latent heat, which reduces the effective static stability of the atmosphere in the area of ascending motion. Notice that the form of the perturbation equations are exactly the same form as given by Holton (1972), ex-
Fig. 14. Vertical streamline pattern showing vertical structure of the local Hadley circulation before (Day 6), during (Day 12) and after (Day 22) the passage of the equatorial intraseasonal disturbance.
cept with the new definition of \( \lambda \). Assuming a wave solution of the form \( \exp[ik(x - ct)] \), the solution of the dispersion equation gives

\[
c = U_m - \frac{\beta (k^2 + \lambda^2)}{k^2 (k^2 + 2\lambda^2)} \pm \delta^{1/2}
\]

where

\[
\delta = \frac{\beta^2 \lambda^4}{k^4 (k^2 + 2\lambda^2)^2} - \frac{U_T^2}{2\lambda^2 + k^2} \frac{2\lambda^2 - k^2}{2\lambda^2 + k^2} \tag{10}
\]

For a given basic wind, static stability and condensation rate, the phase speed and growth rate of a wave can be readily obtained from (9)–(10) in terms of the wavenumber \( k \) and the vertical tilt of the wave is given by

\[
\alpha = \tan^{-1} \left[ \frac{c_0 U_T^{-1} \left( \frac{2\lambda^2 + k^2}{k^2} \right)}{2\lambda^2 + k^2} \right] \tag{11}
\]

positive westward, where \( c_0 \) is the imaginary part of \( c \) in (9).

Before we identify the unstable modes in the numerical model, let us briefly discuss the effects of latent heat release on instability. From (10), it can be shown that marginal instability occurs only if the magnitude of the vertical wind shear \( U_T \) is greater than a critical value \( U_{TC} \) \{=\beta/(2\lambda^2)\} which is a function of \( \beta \) and the static stability only. If the static stability remains unchanged, tropical disturbances will require a much higher vertical wind shear to become unstable compared with extratropical disturbances. The major difference between tropical and extratropical disturbances however, lies in the effect of latent heat on static stability. The weaker the effective static stability, the lesser the wind shear is required for instability and the faster the wave growths for a given supercritical wind shear. In other words, for tropical disturbances, the condensation rate \( q_e \) is the critical factor for instability.

Figure 15a shows the growth rate per day as a function of wavelength and \( q_e \) for basic-state static stability \( S = 1.8 \times 10^{-12} \text{ m}^4 \text{ s}^{-2} \text{ g}^{-2} \) and \( U_T = -11 \text{ m s}^{-1} \). If there is no interactive latent heating mechanism, i.e. \( q_e = 0 \), the critical wind shear \( U_{TC} \) required for marginal instability is 31 m s\(^{-1}\) (i.e. a difference of 60 m s\(^{-1}\) between the 750 and 250 mb zonal wind). This far exceeds any realistic vertical mean wind shear observed over the monsoon region. The release of latent heat corresponding to \( q_e > 4 \times 10^{-3} \) reduces the magnitude of the required critical shear to 11 m s\(^{-1}\) or less, thereby allowing selective growth of disturbances with intermediate scales (Fig. 15a). For realistic range of \( q_e \) \((6 \times 10^{-3} \text{ to } 4 \times 10^{-3})\), the unstable wavelengths are in the range of 2000 to 6000 km. As \( q_e \) increases, the scale of the most unstable disturbance decreases. The maximum growth rate decreases from a wavelength of 5000 to 2500 km corresponding to the above range of \( q_e \). A plot of the real part of the phase speed with \( q_e \) and \( k \) indicates that the phase speed of the waves are generally reduced as \( q_e \) increased (not shown). The reduction is most pronounced for the marginally stable waves in the short-wave range less than 4000 km. As soon as the wave becomes unstable, the phase speed remains almost constant. Figure 15b shows the maximum growth rate as a function of \( U_T \) and \( q_e \) for static stability \( S = 1.8 \times 10^{-12} \text{ m}^4 \text{ s}^{-2} \text{ g}^{-2} \). Clearly, the maximum growth rate increases with increasing \( U_T \) similar to that expected from classical baroclinic theory. More importantly, the
maximum growth rate increases with increasing $q_e$ while the critical wind shear $U_{TC}$ and the wavelength of the fastest growing wave decrease with increasing $q_e$.

In the summer monsoon experiment where instability occurs, the numerical values of various parameters used are $S = 1.8 \times 10^{-12} \text{ m}^4 \text{ s}^{-2} \text{ g}^{-2}$, $f_0 = 0.4 \times 10^{-8} \text{ s}^{-1}$, $\beta = 2.2 \times 10^{-11} \text{ m}^{-1} \text{ s}^{-1}$, $U_m = -1.4 \text{ m} \text{ s}^{-1}$, $U_T = -11.0 \text{ m} \text{ s}^{-1}$. In order to put the numerical results in context with the stability analysis we have to estimate the equivalent value of $q_e$ for the numerical model. To do this, we compute the average heating rate $\mathcal{H}$ (positive only) and the average sigma-velocity $\tilde{\sigma}$ (negative only) over the region $120^\circ$–$180^\circ$ longitude and $5^\circ$–$20^\circ$ latitude on day 12 when the disturbances are rapidly growing. We then assume $L_q = -0.5 \times \mathcal{H}/\tilde{\sigma}$, with the 0.5 factor taking into account the fact that CISK heating in the numerical model occurs only when the lowest layer is convergent while the latent heating in the 2-level model is strictly a linear function of the vertical velocity at the middle level. The value obtained for $q_e$ in this way is $5.6 \times 10^{-3}$ which gives $\lambda^{-2} = 0.29 \times 10^{12} \text{ m}^2$. This value is about one order of magnitude smaller than the value of $\lambda^{-2} = 2.8 \times 10^{12} \text{ m}^2$, when $q_e = 0$ in (8). Thus, the tropical Rossby wave instability can occur with relatively weak wind shear provided the latent heat effect is included. For the value of the effective stability given above, the most unstable wave has wavelength approximately 3500 km and a growth rate of 0.96 per day corresponding to an $e$-folding time of a little more than one day. This wave moves westward with a speed of 5.9 m s$^{-1}$ which is equivalent to a periodicity of about 6 days. The vertical tilt of the wave between the two layers from (11) is $\pi/3$ (or one-sixth of a wavelength) eastward. The point in parameter space corresponding to the numerical experiment is marked by a cross in Fig. 15a and 15b. These results are in remarkable agreement with the numerical results presented in section 4.

b. Comparison with previous theoretical studies

The above results on the effect of CISK heating on baroclinic stability are consistent with the result of the continuous model of Mak (1983), where cumulus heating is proportional to vertical velocity at the top of the surface moist layer, which in turn is determined by the large scale circulation through the $\omega$-equation. Mak found that CISK heating greatly increases the growth rate of a baroclinic unstable wave, causes the wave to contract in horizontal scale and reduces its phase speed, similar to the result of our analyses. Our analytic results are also in qualitative agreement with Moorthi and Arakawa (1985) using a 7-level quasi-geostrophic linear model. In the absence of vertical mean shear, no unstable modes resembling monsoon depressions are found. With nonzero vertical shear, it is found that unstable deep vertical modes with intermediate wavelengths from 2000 up to 7000 km are excited by cumulus heating. Similar to our analysis, they found that for a given shear and static stability, the growth rate of the deep mode increases and the horizontal scale decreases with increasing heating and moisture availability. However, there is a major difference between the present (two-level) analysis and the multilevel analysis of Moorthi and Arakawa. Our analytic results suggest that cumulus heating greatly facilitates development of unstable baroclinic modes independent of the sign of the vertical shear. Moorthi and Arakawa showed that as a result of strong cumulus heating, growth of deep unstable modes is favored in easterly shear compared with westerly shear. This is consistent with our numerical results for the winter monsoon simulations. The lack of a distinction between easterly and westerly shear is clearly a deficiency of the two-level model used here.

In summary, the two-level stability analysis has provided a useful and plausible mechanism for the model growing disturbances in the summer monsoon experiment. To a first approximation, these disturbances can be interpreted as heat-induced unstable baroclinic Rossby waves under vertical shear. The absence of instability in the winter monsoon experiment can be attributed to the much larger basic static stability associated with the strong subsidence in the winter mean flow and also to the presence of westerly shear that inhibits unstable deep mode development. However, the latter requires more sophisticated multilevel stability analysis (e.g. Moorthi and Arakawa 1985).

7. Further discussions

The presence of westward propagating synoptic scale disturbances over the monsoon region is well known (e.g. Murakami et al. 1984, Krishnamurti 1985). To provide a better comprehension of the model results relative to observation, Fig. 16 shows snapshots of observed 850 mb wind field adopted from Murakami et al. 1984 for the period 21–27 June 1979 at two-day intervals. A series of cyclones over the monsoon region at $15^\circ$–$20^\circ$N appeared to propagate westward along the latitude circle and dissipates while new cyclones are generated near $100^\circ$–$110^\circ$ longitude or further east. The centers of these cyclones were found in the vicinity of the region of maximum vertical easterly wind shear (around $12^\circ$–$15^\circ$N). The synoptic sequence and the position of the cyclones relative to the low level flow is quite similar to the numerical results and stability analyses discussed in this paper. The numerical results are also in agreement with recent observations of a pronounced 5–7 day westward-propagating precipitation pattern across the middle of the Indian Subcontinent along tracks of monsoon lows (Hartmann and Michelsen 1989).
Fig. 16. Observed 850 mb wind fields showing horizontal structure and propagation of synoptic scale disturbances ($C_1, C_2, \ldots$) over the monsoon region at two day interval during 21–27 June 1979. Full barb is 5 m s$^{-1}$. (Adopted from Murakami et al. 1984.)

Since the observed equatorial convection associated with intraseasonal oscillation reappears near the region $60^\circ$–$120^\circ$E in approximately 30–60 days (e.g. Lau and Chan 1986), they will lead to a major shift of the equatorial convection to the monsoon heat source region ($15^\circ$–$20^\circ$N) in similar time scales by the mechanism proposed in this paper. On the other hand, during the winter monsoon, when the major heating over the East
Asian/Indian monsoon region is replaced by cooling, no such abrupt migration is possible. Thus, there is a certain preferred time window in which the excitation of high frequency (3–5 days) subtropical disturbances and poleward transition of the ITCZ can occur as the intraseasonal oscillations evolve in tandem with the seasonal cycle. This means that only one to two major transitions can occur during the summer months (June to August). Rapid multiple monsoon onsets separated by periods of breaks are common occurrences in the Indian and East Asian monsoon and may be related to the aforementioned mechanism. In reality, a similar but less extensive southward migration of equatorial convection also occurs during the northern winter because of the onset of the Australian summer monsoon.

The present results can also be compared to those of Nakazawa (1986) who observed that during the northern summer intense convective systems associated with easterly waves tend to propagate westward at 15°–20°N in the region of the western Pacific, South China Sea to the Bay of Bengal. Since these regions are known to have strong vertical wind shear during the northern summer, it may be argued the pronounced development of easterly waves around 15°–20°N in the western Pacific region during northern summer may be a manifestation of the mobile wave–CISK unstable Rossby wave mechanism discussed in this paper. More recently, Nakazawa (1988) revealed that along the equator high frequency, well-organized westward propagating synoptic scale systems are embedded in supercloud clusters associated with eastward propagating 30–60 day oscillations. Lau et al. (1989) conjectured that this rather complex behavior of the equatorial convective system in equatorial regions is the result of mutual adjustment of the large circulation to latent heating involving both moist Kelvin and Rossby waves. In the presence of a favorable large scale vertical shear, it is likely that the equatorial westward propagating disturbances may be steered away from the equator and amplified as heat-induced unstable Rossby waves. Nakazawa’s observation that tropical cyclones or typhoons are often found to develop from the westward propagating equatorial disturbances seem to support this view.

8. Conclusions

We have shown that due to the interaction between diabatic heating and the summer monsoon mean flow, equatorial convection associated with the 30–60 day oscillation may lead to the rapid development of westward propagating synoptic-scale cyclonic vortices over the monsoon region. These cyclones have spatial scales of approximately 3000–4000 km and periods of 4–5 days with the vertical wave axis tilting eastward with height. They are associated with strong convection occurring along an axis immediately poleward of the region of maximum westerly flow. In addition, the intensification of convection in the monsoon region initiates rising motion that draws low-level air northward, consequently weakening the low-level moisture convergence and a concomitant weakening of the eastward propagating supercluster in the equatorial region. As the cyclonic activities develop further over the monsoon region, the local Hadley circulation changes from a double-cell structure with rising motion over the equator to a single cell circulation rising over the monsoon region. The rapid development of these cyclonic disturbances over the monsoon region and the concomitant weakening of the equatorial convective activities may be identified with the rapid northward shift or sudden jump of the Mei-yu rainband observed over East Asia and inverse relationship between the monsoon ITCZ and the equatorial ITCZ over India and East Asia (Lau and Chan 1986; Lau et al. 1988). An experiment using the same basic state flow, but with the sign reversed, shows no counterpart to the above results. In this case, the equatorial disturbance amplifies over the monsoon region but the lateral movement of the convection is small.

From a linear stability analysis of quasi-geostrophic system subject to interactive latent heating, it was shown that the westward propagating disturbances generated over the monsoon region are the manifestations of heat-induced unstable Rossby waves under the influence of vertical wind shear. The monsoon basic circulation over India and Southeast Asia is favorable for the development of these unstable waves. On the other hand, the absence of monsoon-like disturbances in the winter monsoon run may be due to the stronger static stability associated with subsidence in the winter mean flow and the presence of westerly vertical shear which tends to inhibit moist baroclinic growth (Moorthi and Arakawa 1985).

Finally, it is important to note that the stability analysis can only be applied to the initial stages of the development of the monsoon lows. The question of possible barotropic interaction of the monsoon disturbance with the large-scale monsoon flow which may contribute to the decay of the disturbances remain unanswered. Thus, the numerical experiments and the analyses discussed in this paper is at best a quasi-linear approximation to a possible mechanism of monsoon transition. This approximation is acceptable, provided the summer monsoon circulation can be described by a well-defined basic state circulation governed by external forcings due to insolation, large scale topography and land–sea contrast. However, given that the monsoon does not evolve smoothly as a direct function of the external forcings, internal dynamics and physics within the monsoon system must play an important role in the multi-scale processes of the monsoon. When the effects of these internal processes become as large as those governing the basic state, the separation be-
tween basic state and perturbation becomes impossible. In such cases a nonlinear model with full physics, i.e., a GCM, is necessary. Many physical mechanisms such as dynamic instability, cloud–radiation feedback, soil moisture–dynamic feedback, air–sea interaction etc., have been proposed for intramonsoon variabilities. In this work, we only focus on the dynamic/latent heat aspects but leave open the possibility of interaction of this mechanism with others. Most importantly, we have extended our theory of the origin of intraseasonal oscillation in the equatorial region to include a plausible mechanism for monsoon transition. Further modeling and observational work are needed to validate this theory.

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REFERENCES


