Annual Changes of Tropical Convective Activities as Revealed from Equatorially Symmetric OLR Data

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Abstract

How and why equatorial convections are activated in certain preferred regions during specific times of the calendar year, are investigated utilizing equatorially symmetric OLR data. In the equatorial African and American continents, semianual variability is predominant with two peaks of convective activities in boreal spring and fall due to the in-situ radiational heating.

Over the oceanic regions, the role of in-situ surface heating becomes insignificant and gives place to remote forcings in excitement of equatorial symmetric convections and associated Rossby-Kelvin wind responses within the equatorial duct between about 15°N and 15°S, that are determined by the Rossby deformation radius at the equator. The active convective phase in the equatorial western Pacific (EWP) lasts about five months from November to March in association with a systematic southward migration of the surface pressure trough. When the trough arrives at the equator in November, the zonal as well as meridional down-pressure gradient winds cause significant low-level convergence to enhance convections in EWP. Here, January is the month of most active convections, since southward down-pressure gradient winds become strongest due to equatorward penetration of the winter-time North Pacific high. During boreal winter, EWP corresponds to the updraft leg of the equatorial E-W overturning with wavenumbers 1 to 2. There exists a gigantic season-fixed clockwise phase rotation of low surface pressure across the Indian Ocean and western Pacific, namely, northward along 75°E in spring to summer, eastward at 10°N from summer to fall, southward along 155°E in fall to winter, and westward at 10°S from winter to spring, thus completing an annual journey. As such, equatorial convections in EWP are activated during the fall-winter phase of southward migration.

In the equatorial Indian Ocean (EIO), convections are not really activated before and during the South and Southeast Asian summer monsoon season (SEAM), since it persistently induces divergent northward down-pressure gradient winds in EIO. Here, the preferred period of active convections differs significantly with different longitudes. Between about 80° and 100°E (EIO1), October of the post SEAM season is the month of most active convections due primarily to the convergence via the so-called β-effect. The winter-time Indian Ocean high, which penetrates equatorward along the Kenya coast, is responsible for causing a substantial west-to-east pressure gradient and convergent equatorial westerlies in EIO1. Between about 100° and 120°E (EIO2), December represents the peak convective phase under the influence of the northern hemisphere winter monsoon bursting out of Siberia. The role of this winter monsoon system is two fold, i.e.; first, accelerating equatorward down-pressure gradient winds which meridionally converge into regions of heavy convections near Sumatra and Borneo, and second, enhancing the convergence due to the β-effect.
1. Introduction

The mapping of the climatological mean outgoing longwave radiation (OLR) fields (e.g., Janowiak et al. 1985) indicates that the April pattern is nearly symmetric with respect to the equator, with two distinct convective centers over equatorial Africa (EAF) and South America (EAM), respectively. April is the month of monsoon transition from the southern to northern hemisphere. The annual variation of precipitation for global continents and islands have been documented by a number of authors such as Kendrew (1961), Hsu and Wallace (1976). That of tropical convective activity derived from satellite data (e.g., OLR) is analyzed by Heddinghaus and Kruegar (1981), Horel et al. (1989), Matsumoto (1989), Mitchell and Wallace (1992) and Wang (1994a, b). All these former studies show that semi-annual cycle, or double seasonal peaks of precipitation and/or convective activities, are predominant over the near equatorial continents due primarily to seasonal migration of the overheaded sun. Changes in convection over these equatorial continents are often called as zenithal rain (Nieuwalt, 1977). However, the situation over the ocean is more complex. Mitchell and Wallace (1992) have shown that contrasting features are seen in the seasonal variation of zonally averaged OLR between the “western sector” (140°W–40°E) and “eastern sector” (40°E–140°W). Some studies (e.g., Horel 1982; Meehl 1987; Murakami and Wang 1993) have investigated the annual variation of convection and/or rainfall in the tropical Indian and Pacific oceans. However, minute regional differences of seasonal variation of equatorial convections have not been well documented.

Partitioning the all meteorological variables into symmetric and asymmetric components, Murakami and Nakazawa (1985) documented variations in equatorial convections and associated circulations near Sumatra and the Malaysian Peninsula before, during and after the April transition. The symmetric data set is well suited to monitor equatorial convections and equatorially trapped circulations in April, while the asymmetric data set provides impeccable judgment on the structure of southern hemisphere monsoon convections and associated circulations in March, as well as those in May for the northern hemisphere monsoon. As such, convections and associated circulations change their structural features when crossing the equator during the April transition.

A quick look at the climatological mean OLR maps constructed by Matsumoto (1992) also reveals somewhat active convections occurring over the equatorial Indian Ocean (EIO) in October, during which month monsoons retreat from the northern to southern hemisphere. Over oceanic regions, such as EIO, the in-situ sensible heating due to solar radiation contributes little, if any, to annual variations in equatorial convections. Activation of equatorial convections in EIO is believed to be accomplished by a remote influence of SEAM (South- east Asian Monsoon), as previously postulated out by Meehl (1987) and Yasunari (1991). Of particular interest in Matsumoto’s (1992) OLR maps is commencement of extremely strong convections over the equatorial western Pacific (EWP) in November, i.e., one month lagged behind activation of equatorial convections in EIO. Furthermore, EWP is characterized by prolonged active convections over five months from November to March, and possibly beyond. Observational information has been scanty pertaining to the nature and origin of active and persistent equatorial convections in EWP. The present study is an attempt to remedy this deficiency, by documenting the processes through which the winter-time Siberian High, as well as the North Pacific High, exerts a measure of control upon equatorial convections over EWP in boreal winter.

For each meteorological variable, the climatological pentad mean fields are decomposed into symmetric and asymmetric components. As will be shown in Section 2, the symmetric data set is introduced to clearly define the equatorial convections and associated basic equatorial flows, while no asymmetric monsoonal modes are present at and in the immediate vicinity of the equator. An intensity index and areal extent of equatorial convections are also defined in Section 2. Annual variability of convective indexes over EAF, EAM, EIO and EWP will be reported in Section 3. Fourier wavenumber analyses of convective indexes are performed in Section 4, while influence of the northern and southern hemisphere winter monsoons upon convectivities in EWP and EIO are documented in Section 5. Finally, a brief summary of the results and discussions are presented in Section 6.

2. Data and computational procedures—

Definition of asymmetric monsoon mode and symmetric equatorial mode—

Datasets utilized in the present study include twice daily OLR derived from NOAA satellite observations for the 12-year period 1975-1977 and 1979-1987 (Matsumoto 1992), and once daily (12 GMT) u and v at 850 hPa extracted from the ECMWF gridded global operational analyses for a 9-year pe-
cal area between 30°N and 30°S. Due to a some-
what short sample size, the pentad mean data is
perhaps subject to some statistical noises caused by
severe transient events. To reduce such statistical
noises, a (1-2-1) weighed mean of three consecu-
tive pentad data is computed at every three pen-
tads. These smoothed every three pentad data are
regarded as independent data. The spatial resolu-
tion is 2.5 x 2.5 degrees. For symbols used in this
paper, refer to the Appendix.

Similar computational procedures, as employed
by Murakami and Nakazawa (1985), Murakami
(1993) and Wang (1994b) are used in the present
study to partition all meteorological variables “A”
into two components, the first component being
symmetric about the equator (signified as A''), while
the second one being asymmetric with respect to the
equator (denoted by A'''); i.e.,

\[
A' = \frac{1}{2}(A_N + A_S),
\]

\[
A'' = \frac{1}{2}(A_N - A_S),
\]

where the suffix “N” (“S”) refers to the variable
at the northern (southern) hemisphere. One ad-
vantage of the data partitioning is to clearly sepa-
rate the monsoonal circulation of quasi-geostrophic
character in subtropical latitudes from the diver-
gent circulation regime prevailing in near-equatorial
latitudes. This is schematically shown in Fig. 1,
which demonstrates annual migration of active con-
vections between North and South Africa, where
the ground surface is relatively uniform and nearly
symmetric with respect to the equator. During bo-
real summer, the monsoon (signified as NAFM) is
active over North Africa in and around the mon-
soon trough. There exist surface pressure doublets
straddling the equator with a low (high) pressure
cell over North (South) Africa, reflecting asymme-
tric heat contrasts between heated and cooled land
masses. An immediate consequence is development
of a Hadley circulation directed northward, and is
responsible for releasing gravitational potential en-
ergy, eventually converging into a low pressure cen-
ter and contributing, at least partially, to accelera-
tion of quasi-geostrophic monsoon circulation with
active convections in its southwestern sector (Fig. 1,
left). Namely, the asymmetric monsoon circulation
can be identified as a gravity-Rossby mixed system.
A set of variables (u'', v'', T'', and w'') defines the
asymmetric monsoon circulation and satisfies the
continuity equation as well as the total energy con-
serving equations of motion, as shown by Murakami
(1993). During boreal winter, SAFM becomes ac-
tivated over South Africa and the asymmetric mode
defined by (u'', v'', T'', and w'') becomes the re-
verse of that in boreal summer (Fig. 1, right). As
such, the asymmetric mode can be regarded as defin-
ing the monsoon, since not only winds (u', v')
but also other variables (\(\phi''\), T'', and \(\omega''\)) alternate an-
ually.

No monsoonal circulation is present during the
spring and fall transition, when the sun is located
right at the equator. As schematically shown in
Fig. 1 (middle), the equatorial circulation is essen-
tially symmetric about the equator, and is trapped
within the equatorial duct between about 15°N and
15°S, that is determined by the Rossby deformation
radius at the equator, i.e., \(\sqrt{c/\beta}\) where \(\beta = \partial f/\partial y\)
and \(c\) is the phase speed of gravity waves (Young
1987). To the west of equatorial convections is a
Rossby response, which is shorter in longitudinal
extent with a pair of cyclonic cells straddling the
equator and stronger westerlies along the equator.
To the east of equatorial convections is a Kelvin re-
response, which is broader in longitudinal extent with
easterlies along the equator. The secondary vertical
circulation consists of strong rising motions in the
convective center with corresponding subsidence on
the equator to the east and slightly poleward to the
west, which is commonly referred to as the “Walker
circulation”. When frictional damping is induced,
one also finds that (1) the Hadley circulation in the
symmetric mode consists of rising motion right at
the equator, with down-pressure gradient \(v''\) con-
verging toward the equator (Fig. 1, middle), and
(2) there is a tendency for generating double ITCZ near
5°N and 5°S, respectively (refer to Fig. 15A).
Here, our point is that a simple model, as shown in
Fig. 1, serves as a useful building block for identi-
fying the difference between asymmetric monsoonal
modes and symmetric equatorial modes.

Over the equatorial continents, such as Africa,
the symmetric equatorial modes are of semi-annual
standing oscillation character, recurring twice a year
at the spring and fall equinoxes. Along the equator,
no meridional wind is present (Fig. 1, middle). As
such, the monsoon itself can not traverse the equa-
tor during the spring and fall transition. It is the
equatorial symmetric mode as defined by \(u', v', \phi',
T', \text{ and } \omega'\) that contributes to the cross-equatorial
migration of convections from one hemisphere to an-
other. In short, the annual migration of convections
between North and South Africa is completed by
consolidated efforts of two types of standing oscilla-
tions; one is the one-year standing oscillation with
the maximum amplitude away from the equator, and
the other being the semi-annual oscillation which is
strongest at the equator.

1 If one defines the positive meridional wind to be poleward
in both the northern and southern hemisphere, one can
signify the asymmetric mode as a simple form of \(u'', v'', \phi'',
T'', \text{ and } \omega''\). However, one weakness in this
definition is that \(v''\) discontinuously changes its sign at
the equator.
An advantage of the data partitioning into symmetric and asymmetric components can be further substantiated when comparing the observed structure of the asymmetric mode over the global subtropics (Fig. 2) with that of the symmetric mode over the global tropics (Fig. 3). During the boreal winter at P4 (Figs. 2A and 2B), NAFM, NAIM and SAMM are three of active monsoons over the southern hemisphere subtropics with the monsoonal westerly $u''$, and the northerly Hadley circulation $v'$. In contrast, P40 (Figs. 2C and 2D) is the pentad of active northern hemisphere monsoons of NAFM, SEAM, WNPM, ENPM and ENAM with the southerly Hadley circulation $v''$ that converges into the monsoonal domains. In other words, all variables alternate their sign annually and, hence, satisfy monsoon definition.
The structure of the symmetric mode is of different character, as evident in Fig. 3. Four convective areas with $\text{OLR}'$ less than 240 Wm$^{-2}$ are recognized in the near equatorial region; equatorial Africa (EAF), equatorial Indian Ocean (EIO), equatorial western Pacific (EWP), and equatorial South America (EAM). The symmetric wind fields ($u'$, $v''$) associated with these four equatorial convective centers are very much similar to those shown schematically in Fig. 1 (middle). For example, westerlies ($u' > 0$) are present over EIO all the year round, while EWP is characterized by permanent easterlies ($u' < 0$), thus defining the equatorial Walker circulation. Cross-equatorial Hadley circulation is absent since $v''$ is equal to zero at the equator. Of course, $v''$ changes its sign across the equator and is responsible, at least partially, for causing meridional convergence or divergence along the equator. The annual mean ($u'$, $v''$, $\phi'$, $T'$, and $\omega'$) defines the basic tropical circulation. Here, care must be exercised when interpreting the symmetric mode far away from the equator in Fig. 3. The symmetric mode becomes ill organized and loses its identity as one goes poleward of the equatorial duct (about $15^\circ$N–$15^\circ$S).

At any rate, the tropical circulation is subject to substantial annual variations, which makes it possible for convections to cross the equator from one hemisphere to another. Over the equatorial continents, such as EAF and EAM, convections become strongest twice a year when the sun crosses the equator during the spring and fall transition. In EIO, $\text{OLR}'$ becomes lowest near 90°–100°E with prominent westerly $u'$ near 80°E during mid-October, as will be elaborated on in Section 3. The early winter between mid-December to mid-January is the season of lowest $\text{OLR}'$ near 110°–120°E, as well as near 140°–150°E, with the strongest equatorial westerly $u'$ near 130°E and the most intensified equatorial easterly $u'$ near 130°W. The primary objective of the present study is to detail how and why the tropical basic flow over EIO and EWP, as defined by ($u'$, $v''$, $\phi'$, $T'$, and $\omega'$) varies annually without an apparent connection with annual migration of the sun.

The data partitioning of $A$ into symmetric $A'$ and asymmetric $A''$ is quite similar to that employed in the study of general atmospheric circulation by separating the annual varying zonal mean $\overline{A}$ from zonally asymmetric eddies $A'$ and investigate the zonal mean-wave interaction. The present study is an attempt to identify the structure, as well as the origin, of the annually varying basic tropical flow as defined by the symmetric data set of ($u'$, $v''$, $\text{OLR}'$). This data set appears to be quite appropriate, if not the best, to characterize the flow field in near-equatorial latitudes. More specifically, since the annual mean of $A''$ is approximately zero, the annual mean of $A'$ is identical to the annual mean of original data $A$. The mutual interaction between $A'$ and $A''$ will be reported elsewhere (Matsumoto and Murakami 2000) as a sequel to the present study. Such interaction is of vital importance when convections transit from one hemisphere to another during the course of monsoon alternation between the two hemispheres.

The annual variations of equatorial convections are monitored by $\text{OLR}'$ over EAF, EIO, EWP, and EAM, respectively. Each of the domain is subdivided by 2.5° lat. $\times$ 5.0° long. mesh. Next, the heavy rain area $S_i'$ in $i$-th mesh is defined as the area where the value of $\text{OLR}'$ is lower than 240 Wm$^{-2}$. The optimum value 240 Wm$^{-2}$ approximately corresponds to the daily precipitation of about 6 mm (Wang, personal communication). Convective intensity $C_i'$ is then defined by the anomaly from 240 Wm$^{-2}$ as follows:

$$C_i' = 240 - \text{OLR}_i'$$

Here, $\text{OLR}_i'$ is the area average of pentad mean $\text{OLR}'$ in the $S_i'$ area. Area weighted $S_i'C_i'$ is regarded as convective intensity in $i$-th mesh. The convective intensity index $C_i'$ in each convective domain of EAF, EIO, EWP and EAM is,

$$C_0' = \sum S_i'C_i' = S_0' \frac{\sum S_i'C_i'}{\sum S_i'}$$

$$S_0' = \sum S_i'$$

The values of $C_0'$ and $S_0'$ in each pentad were measured. In general, when $C_0'$ is large, $S_0'$ is also large, however, the relationship between the two values is not proportional. The area with $\text{OLR}'$ exceeding 240 Wm$^{-2}$ is eliminated, thus $S_0'$ is always smaller than the total monitoring area of each domain.

In the symmetric mode, the center of convective activity is always located at the equator. The longitudinal location of the convective center, $X_0'$, in a certain pentad is defined as follows;

$$X_0' = \frac{\sum X_i'S_i'C_i'}{\sum S_i'C_i'}$$

Here $X_i'$ is the relative distance from the reference longitude (for example, 20°E in EAF).

Before closing this section, an additional comment is given to the optimum criterion (3), which can be met (i.e., positive) even for relatively weak equatorial convections. Hence, caution must be exercised when interpreting the $C_0'$, $S_0'$ and $X_0'$ values computed from Eqs. (4) to (6). On a trial and error basis, a more tense criterion of $C_i' = 220 - \text{OLR}_i'$ was also tested. It turns out that this tense criterion is appropriate when investigating cross-equatorial migration of strong convections during the monsoon transitions from one hemisphere to another, as will be reported in the subsequent paper by Matsumoto and Murakami (2000). However, the primary objective of the present study is to monitor the nature of
equatorial convection only. To this end, regions of $OLR'$ less than 240 Wm$^{-2}$ are monitored over an extensive latitude band between 20°S and 20°N at every pentad. We have deliberately chosen such an extensive zone so that no computational error occurs even when monitoring relatively weak equatorial convections of $OLR'$ less than 240 Wm$^{-2}$ (refer to Fig. 3). Generally speaking, equatorial convections are confined within a narrow equatorial channel between 15°N and 15°S.

3. Annual cycle of equatorial convective activities

In the equatorial continents such as EAF and EAM, convective index $C_0'$ shows clear semiannual cycle (Figs. 4A and 4B). Several weeks after the vernal and autumnal equinoxes, convective activities are enhanced, while they are suppressed several weeks after the winter and summer solstices. This is, without doubt, due to the seasonal changes of surface heating by the seasonally migrating sun. The
phase of $C'_0$ and $S'_0$ is concurrent in EAF, while it is slightly different in EAM.

Four phases are signified in order to describe annual changes of $C'_0$. WI is the boreal winter phase between P70 and P10. In WI phase $C'_0$ is less than the annual mean value according to Fig. 4A. The term "boreal" will hereafter be omitted for brevity. SP corresponds to the period between P16 and P28. This is the season of activated convection over the equatorial continents. Similarly, SU phase is taken as P34-P49, while FA phase is P52-P64. In short, clear semiannual cycles are evident in convective activities over the equatorial continents.

Seasonal changes of $C'_0$ are also discernible in the EIO (Fig. 4C). Here, activated convection occurs in SU and FA, while convective activities in EIO become weakened in late WI and SP. As such, convective activity in the equatorial oceanic region does not seem to be directly influenced by the seasonal solar migration.

Temporal variations of $C'_0$ in Fig. 4C include not only seasonal changes, but also intraseasonal variability with a period approximately 30 day. The amplitude of this variability is especially larger in SU, which characterizes nature of convective activities in SU season. Predominance of intraseasonal variability may be due to the statistical noises owing to the relative short sample period. The amplitude, however, seems to be large enough to exceed the statistical noise level. Another possibility is the regular season dependent, and recurrent climatological intraseasonal oscillation (CISO). Wang and Xu (1997) investigated the CISO that brings about climatological active and break phases in the monsoon activity. To determine whether the intraseasonal oscillations in Fig. 4C are caused by a statistical noise, or indicating pure CISO, is difficult due to a relative short sample period.

Figure 5 presents sequence of $OLR'$ maps from P37 to P46. When paying attention to the EIO domain, convective activity is relatively weak in P37 and P43, while it is relatively strong in P40 and P46. Such systematic alternation of active-break phases implies evidence of CISO. One reason why CISO is clearly shown in Figs. 4C and 5 is that these diagrams are presented not by the original $OLR$ data, but by the equatorially symmetric $OLR'$ data. CISO is basically of equatorial origin and, thus, $OLR'$ is well suited to monitor such equatorial perturbations. Another reason is that the convective index $C'_0$ in this paper is based not only on $C'_i$ but also on $S'_0$ (Eq. 4). Multiplication of these two factors may amplify the variability of convective activities.

Figure 4D shows the annual variation of $C'_0$ in EWP. At a glance, it is evident that the seasonality is more systematically appeared in EWP than in EIO. The $C'_0$ value is slightly in excess of the annual mean in SU phase. The most prominent wet season in EWP occurs during the WI phase. As will be elaborated on later in Section 5, an exceptionally strong convection in WI is accomplished, by some way, due to a remote forcing of the Asiatic winter monsoon and the North Pacific High. SP and FA are relative dry phases in EWP. The reason why convection is suppressed in FA between P46 and P52, even though convection in WNPM is still active, will also be presented in Section 5. CISO can be seen in the time series of $C'_0'$ in Fig. 4D all the year round.

Since the seasonal changes are obscured by the CISO component, seasonal mean values of $C'_0$, $S'_0$, and $X'_0$ in WI, SP, SU and FA phases are shown in Table 1. Since the length of each phase is more than twice as long as the period of CISO (approximately 30 day), the season mean values are no longer influenced by CISO. Over the equatorial continents, wet phases in SP and FA alternate with dry phases in WI and SU. The value $C'_0$, however, is only around 90 W to 100 W, thus the equatorial continents contributing only minor part of the global tropical heat sources. The largest heat source in the equatorial zone emerges in EWP during WI phase. Here, the value of $C'_0$ attains 312 W, which is the largest value in $C'_0'$ of all the domains. In comparison, the $C'_0$ value in EIO during SU is only 208 W, i.e., EIO is
relatively smaller heat sources.

4. Harmonic analysis of convective index $C_o'$ near the equator

Figure 6 (top) presents the longitudinal distribution of annual mean $C_o'$ obtained at the equatorial channel between $10^\circ S$ and $10^\circ N$. Four major peaks are observed in 25°E, 100°E, 150°E and 70°W, respectively, which nearly coincide with the longitudes of annual mean $X_o'$ presented in Table 1. When harmonic analysis is applied to the annual mean $C_o'$, dominant waves turn out to be wavenumbers 1 and 2, and their amplitude and phase are (10.7 W, 111°E) and (11.2 W, 125°E), respectively (Table 2, top). For convenience, the phase is presented by the longitude where maximum of each wave emerges.

Symbol $\Delta$ in Fig. 6 and Table 2 represents the anomaly from the annual mean. For instance, the anomaly in WI phase $C_o'(WI)$ is defined as,

$$\Delta C_o'(WI) = C_o'(WI) - C_o'(MEAN). \quad (7)$$

The characteristics of $\Delta C_o'(WI)$ is dominant positive values in the EWP region, extending 60° in longitude between 120°E and 180°. Positive $\Delta C_o'(WI)$ anomaly, although small, also appears in the eastern coast of South America between 60°W and 40°W. Because of two positive anomaly peaks, the dominant wavenumber becomes 1 and 2 (Table 2). The amplitudes of these waves are both in excess of 6 W, indicating how anomalous convective activity in WI is predominant. Phases of both wavenumbers 1 and 2 are located near 150°E, thus producing strong heat sources there by mutual consolidation. While the waves with wavenumbers 1, 2 and 3 cancel out each other, thus no large heat sources emerge near the east coast of South America. In short, WI is the most important season for intensification of the equatorial ultra-longwave circulation, including the Walker circulation. The origin of this strong ultra-longwave heat sources is apparently the northern hemisphere winter monsoon circulation, which will be elaborated on in Section 5.

The amplitude of wavenumber 1 is much greater in SP than in FA. This is because positive $\Delta C_o'(SP)$ anomaly in EAF shifts westward, while that in EAM migrates most eastward in SP than in FA, which diminishes the distance between these two anomalous convective regions, and contributes to wavenumber 1 during SP.

One of the peculiarities in Table 2 is very small amplitude of only 1.6 W for wavenumber 1 during
SU phase. Two major peaks are apparent at 80°–90°E and 110°–90°W in $\Delta C_0'$ (SU) profile (Fig. 6). Both $\Delta C_0'$ (SU) anomalies are of nearly the same magnitude, and the distance between them is about the half of the earth’s circle. Thus, wavenumber 2 predominates over wavenumber 1. In short, due to the interference between two major monsoons in the northern hemisphere (SEAM and ENPM), ultra-longwave heat sources of wavenumber 1 do not dominate in SU. As such anomalous equatorial E-W circulation with wavenumber 1 does not appear in the 850 hPa $u'$ field either (not shown). The dominant wavenumber in SU is 2, but wavenumbers 4 and 5 are also significant (not shown).

5. Winter monsoons as a forcing of equatorial convective activities in EIO and EWP

The $C_0'$ values only in the near-equatorial channel confined between 5°N and 5°S are plotted in Fig. 7. Here, an 1-2-1 filter is once again applied to the every 3-pentad data in order to eliminate the effect of CISO. Therefore, the phenomena with less than 30 day period are no longer included in Fig. 7 and, thus, clearly demonstrate the seasonal variations of equatorial convections in EIO and EWP. Also, recall that the $C_0'$ value in Fig. 7 reflects equatorial convections of symmetric character only. In other words, it is not contaminated by tropical convections of asymmetric monsoonal character.

According to Murakami and Matsumoto (1994), SEAM onsets at around P29 and continues its activity approximately 5 months until it weakens by P57. A glance at Fig. 7 reveals that the life cycle of convective activities along the equator in EIO (80°–120°E) is quite different from those occurring in the SEAM domain. About 8 pentads after the SEAM onset, EIO begin to be activated near 90°E around P37. Once established, the center of the equatorial convections in EIO tends to adhere near 90°–95°E, while intensifying, until it reaches its maximum intensity of 50 units by P58 which is nearly coincident with the withdrawal phase of SEAM. Between about P61 and P67, the center of equatorial convections shifts eastward and arrives at its final destination near Borneo (100°–115°E) by mid-November. Borneo is the persistent convective center throughout the boreal winter, while gradually weakening and finally dissipating by mid-April. As outlined hitherto, there seem to be two preferred areas of convective activities in EIO. The first area extends from about 80° to 100°E, which is hereafter signified as “EIO1”, while the second area between about 100° and 120°E is marked by “EIO2”. The convectively active period of EIO1 extends about four months from the end of June to the end of October, in contrast to the

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five month active period of EIO from November to March. As will be shown later on in Subsection 5.2, the origin (or forcing) of equatorial convections in EIO exhibits a fundamental difference from that in EIO2.

A glance at Fig. 7 reveals that $C_0'$ is relatively small near 125°E. As such, this longitude was assigned as the boundary between EIO and EWP in Figs. 4C and 4D. The first activation of convection in EWP appears in P31–P37 in relatively narrow region between 140°E and 150°E (Fig. 7). Afterwards, equatorial convective activities in EWP diminish even though WNPM itself further develops. The reason why equatorial convection depresses for a long period between P37 and P61 will be elaborated on later. WNPM begins to retreat from its major active arena (130°–160°E, 10°–20°N) by P60 according to Murakami and Matsumoto (1994). After the withdrawal of WNPM, equatorial convective activities in EWP suddenly enhance with $C_0'$ value exceeding 10 units in a wide domain (125°–165°E). The peak of $C_0'$ appears near 140°–150°E (Fig. 7) around P7–P10. Equatorial convection in EWP continues to be active until early April of the following year. As stated above, convective activities in EWP show rather peculiar variations: (1) Why equatorial convection in EWP is inactive, while that in EIO is very active during the period between P37 and P61? (2) Why does equatorial convection in EWP become suddenly active, in a climatological sense, in mid-November and attain its peak in January–February? Subsection 5.1 attempts to answer these questions.

5.1 Origin of equatorial convections in EWP

Figures 8 and 9 are prepared to identify the origin of equatorial convections in EWP. Figure 8A depicts the longitude-month section of monthly mean SST along the equator based on Sadler et al. (1987b). The highest SST is located in 140°–160°E all the year round exceeding 29°C, thus SST cannot be a cause of annual convective variability. High SST can be one of the necessary conditions for convective activation but, certainly, it cannot be a sufficient condition. Figure 8B presents longitude-month section of monthly mean surface pressure at the equator, extracted from Sadler et al. (1987b). One immediately notes that the surface pressure in 140°E–180° domain is low during the period from November to April. This domain and period exactly coincide with those of active convection in EWP. The in situ SST effect contributes little, if any, to establishment of the equatorial low pressure system. In addition, the surface pressure in the same domain is high during the period from June to September, therefore, equatorial convection in EWP is depressed even when it corresponds to the active phase of WNPM.

Equatorial convective activities are primarily controlled by low-level convergence within the equatorial planetary boundary layer. Based on the climatological monthly mean data, Murakami et al. (1992) obtained an approximate set of equations of motions in the tropical boundary layer as follows:

$$-fv + \frac{1}{\rho} \frac{\partial p}{\partial x} = -\kappa u, \quad (8)$$

$$fu + \frac{1}{\rho} \frac{\partial p}{\partial y} = -\kappa v, \quad (9)$$

where $\rho$ is the air density (10⁻² kgm⁻³), and $k$ a frictional damping coefficient (10⁻⁵ s⁻¹), and $f$ the Coriolis parameter. Equations 8 and 9 can be rewritten as follows:

$$u = \frac{\kappa}{f^2 + \kappa^2} \frac{1}{\rho} \frac{\partial p}{\partial x} - \frac{f}{f^2 + \kappa^2} \frac{1}{\rho} \frac{\partial p}{\partial y}, \quad (10)$$
A similar approximate set of equations was also obtained by Meehl (1990). In Fig. 8B, the pressure gradient is eastward near 130°–140°E, while being westward to the east of the dateline. At the equator where $f = 0$, the zonal winds are directed down the $\partial p/\partial x$ gradient (refer to Eq. 10). Accordingly, the eastward pressure gradient near 130°–140°E approximately corresponds to 5 m s$^{-1}$ (westerly), while the westward pressure gradient near the dateline is indicative of $u$ of about 7–8 m s$^{-1}$ (easterly). This results in zonal convergence of approximately $2 \cdot 3 \times 10^{-6}$ s$^{-1}$ between these two longitudes.

There also exists significant meridional convergence within the equatorial planetary boundary layer. Figure 9B presents the November mean surface pressure field extracted from Sadler et al. (1987b). Although the $\partial p/\partial x$ gradient is quite small near 10°N (10°S), the $\partial p/\partial y$ counterpart is substantially large in that vicinity. Omitting the small terms related to $\partial p/\partial x$ in Eqs. (10) and (11), note that (1) $u$ is sub-geostrophic since $f^2$ is of the same magnitude as $\kappa^2$, and (2) $v$ is northerly near 10°N as contrasted with southerly near 10°S. The meridional convergence due to northerly $v$ near 10°N and southerly $v$ at 10°S is estimated to be about $5 \times 10^{-6}$ s$^{-1}$. Hence, the net convergence due both to $\partial v/\partial y$ and $\partial u/\partial x$ totals 7 to $8 \times 10^{-6}$ s$^{-1}$, which is quite substantial in a climatological sense and perhaps capable of exciting and maintaining strong equatorial convections in November over the EWP domain. Except for November, the surface pressure field is not symmetric with respect to the equator. Even so, an inspection of Figs. 9C and 9D reveals the presence of a strong planetary boundary layer convergence due to down-gradient, non-geostrophic $u$ and $v$ winds over the equatorial western Pacific throughout the boreal winter. More specifically, the eastward $\partial p/\partial x$ gradient near New Guinea becomes more prominent in December and January than in November. Furthermore, the southward $\partial p/\partial y$ gradient to the south of the winter-time Pacific High substantially sharpens in December and January. In fact, the three winter months of December to February are the most convectively activated phase of EWP (refer to Fig. 7).

Glancing at Fig. 9, one at once notes a systematic southward migration of the surface pressure trough from the northern hemisphere WNPM domain in September, crossing over the equator in November, to as far south as the southern hemisphere NAIM domain in January. This is an exceptional, and unique aspect of the North and South Western Pacific. When the surface pressure trough is located in northern latitudes, it is customarily referred to as the WNPM monsoon trough. Here, the horizontal convergence of mass in the planetary boundary layer due to the frictional secondary circulation is proportional to the geostrophic vorticity in and around the monsoon pressure trough (refer to Section 5.3 of Holton 1979). Equatorward of the monsoon trough is horizontal divergence due to anticyclonic vorticity. This is one reason why EWP experiences a dry phase between July and October (Figs. 4D and 7). When the monsoon trough approaches the equator, it changes its structure and the prevailing winds become no longer quasi-geostrophic, as already pointed out in Eqs. 8–11. These non-geostrophic winds are responsible for enhancement of strong equatorial low-level convergence and associated convections between November and February, during which period $\partial u/\partial x$ and $\partial v/\partial y$ are both significant over the EWP domain. In short, it is the WNPM monsoon in June...
to September that acts as a primary and time-lagged regulator to enhance equatorial convections in EWP during the WJ season.

The seasonal migration of the WNPM monsoon trough is most likely to be forced by the southward as well as eastward phase propagation of the Siberian High between September and November as manifested by significant pressure increase in the vicinity of the Philippines (see Figs. 9A and 9B). This facilitates an increase in the west-to-east pressure gradient near New Guinea, contributing to enhancement of zonal convergence and convections in EWP. In November, the east-to-west pressure gradient eastward of the dateline also becomes strong due to the arrival of the monsoon trough to the equator. In contrast, presence of an equatorial trough is not absolutely required for convections to be activated in EIO, as will be shown later.

The unique and peculiar aspects of the western Pacific are further exemplified by presenting Figs. 10A to 10D, which present the latitude-month sections of surface pressure at 30°E (Africa), 75°E (Indian Ocean), 120°E (South China Sea), and 155°E (western Pacific), respectively. The heavy full line traces the phase of minimum pressure (or trough) appearing on the y-t domain. Over the African continent, the minimum phase of surface pressure alternates between the northern and southern hemispheres by propagating northward (southward) during the spring (fall) transition, that follows the changes in surface heating by the seasonally migrating sun. The role of surface heating becomes insignificant over oceanic regions, such as the Indian and Pacific Oceans. The Indian Ocean, between about 60°E and 100°E, is characterized by a distinct northward shift during spring, in sharp contrast to an ill organized (or non-existing) southward phase propagation in fall. The reversal phase propagation between the Indian Ocean in spring and the western Pacific in fall is probably indicative of a clockwise, one-year rotation of minimum pressure phase across these two gigantic oceans. This possibility is examined by presenting Figs. 11A and 11B, which are for the longitude-month sections of surface pressure at 10°N and 10°S, respectively. Figure 11A reveals a systematic eastward progression of minimum pressure phase from the Arabian Sea in June to the western Pacific by November along 10°N. Conversely, the phase propagation of minimum pressure at 10°S is westward during boreal winter between December and March (Fig. 11B). This completes the one-year clockwise rotation of minimum pressure phase encircling along 75°E, 10°N, 155°E and 10°S. It is likely that such clockwise rotation is controlled by the state of land-sea coupled system.

Over the equatorial western Pacific (Fig. 10D), regions of high SST in excess of 29°C (dashed lines) alternate between the northern and southern hemispheres with maxima (29.5°C) at 10°N in September and also at 10°S in January. As such, the in-situ SST is not a controlling factor for the southward phase propagation of minimum pressure over the western Pacific. Here, the WNPM monsoon trough becomes strongest in mid-August. Kawamura and Murakami (1998) postulated that the WNPM monsoon trough is established not by a local SST, but by a remote continental forcing. The thermally induced heat low over northern China begins to shift southward due to commencement of surface cooling, followed by a counterclockwise penetration into the western North Pacific and contributing to full intensification of the WNPM monsoon trough by mid-August. Kawamura and Murakami's (1998) postulation can be reconfirmed from an inspection of Figs. 10C, 11A and 10D as follows: (1) The southward shift of minimum pressure along the east coast of China from 20°N to about 5°N (Fig. 10C), (2) the eastward phase progression across the western Pa-
specific along 10°N (Fig. 11A), and (3) the equatorward penetration along 155°E (Fig. 10D). When the minimum pressure phase reaches the equator by November and December, it brings about an extensive E-W equatorial trough (Figs. 9B and 9C), accompanied by widespread active convections dominating the equatorial western Pacific (Fig. 7).

The center of the clockwise rotation of minimum pressure phase appears to be located somewhere near 120°-125°E, which longitudes are characterized by the presence of both the northward propagation in spring and the southward progression in fall (refer to Fig. 10C). This is another supporting evidence of our assignment of 125°E as the boundary between EIO and EWP in Figs. 4C and 4D. How and when equatorial convections are activated in EIO are quite different from those encountered in EWP.

5.2 Origin of equatorial convections in EIO

As stated earlier in Fig. 7, convective activities in EIO1 (80°-100°E) are prominent between about June and October, while those in EIO2 (100°-120°E) are dominant from November to March of the following year. The manner in which convections are excited is also quite different between EIO1 and EIO2.

Figures 12A to 12C present the surface pressure maps in June, August and October, respectively, during which periods EIO1 is convectively activated. In June, we see a well-established SEAM monsoon trough extending off the east coast of India with a substantial northward pressure gradient poleward of about 10°N. Thus, the leading component in Eq. 11 is the \( \partial p/\partial y \) related term, which induces northward divergent \( v \) winds between the equator and 10°N. See also Fig. 5 of Murakami et al. (1999) for positive E-P (evaporation exceeding precipitation) over the equatorial Arabian Sea-the Bay of Bengal region. In contrast, in the vicinity of 10°S between 80°E to 100°E (Fig. 12A), the equatorward \( \partial p/\partial y \) gradient causes equatorward convergence of \( v \) winds, which are commonly referred to as the low-level Hadley circulation associated with the southern hemisphere winter-time anticyclonic system. The same is true in August (Fig. 12B), i.e., there exists cancellation (or overcompensation) of convergence due to the southern hemisphere winter monsoon by divergence due to the northern hemisphere summer monsoon (SEAM). This makes it difficult for equatorial convections in EIO1 to fully develop during the mid-SEAM season of June to August. By October (Fig. 12C), the SEAM monsoon trough drastically dissipates, while the southern hemisphere anticyclone still remains appreciably strong. Hence, October is expected to be the month of active convections in EIO1. This statement must be substantiated by evaluating divergence (or convergence) in EIO1 (80°-100°E, 10°N-10°S). The divergence equation can be derived by differentiating Eq. (8) with respect to \( x \) and Eq. (9) by \( y \), followed by adding up the resulting equations together. At the equator where \( f = 0 \), we obtain

\[
\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} = -\frac{1}{\rho k} \left( \frac{\partial^2 p}{\partial x^2} + \frac{\partial^2 p}{\partial y^2} \right) + \frac{\beta}{\kappa^2} \frac{1}{\rho} \frac{\partial p}{\partial x},
\]

where \( \beta = \partial f/\partial y \).

Table 3 (top) presents computed results in EIO1. Although negative (convergent), the term A is not significant at all months. As expected, the term B is positive (divergent) during the mid-SEAM season of June to September and changes its sign to negative after October. The term C, which indicates the so-called \( \beta \)-effect, is convergent and substantial throughout the period. In other words, the \( \beta \)-effect is the major contributor to equatorial convections in EIO1. Along the equator, the \( \partial p/\partial x \) gradient is eastward (Fig. 12) and tends to induce westerly \( u \) winds along the equator. The surface pressure is
constantly high near Kenya due to an equatorward invasion of the southern hemisphere high pressure system, while it is low over the maritime continent near Sumatra and Borneo, thus causing the eastward $\partial p/\partial x$ gradient along the equator. In fact, it is the southern hemisphere winter-time anticyclonic system that is primarily responsible for enhancement of equatorial convection in $E1O_{2}$, in particular during late boreal fall of October to November.

The remaining problem is to explain how and why equatorial convections become fully intensified during boreal winter from about November to February in $EIO_{2}$ (100°-120°E, 10°N-20°S). This is a key area of investigation into the role of cold air outbreaks emanating out of Siberia upon enhancement of equatorial convections in and around large equatorial landmasses of Sumatra, Borneo and Java Islands. A tiny anticyclonic cell over Indochina in Fig. 12C is a sign of the northern hemisphere winter monsoon. By November (Fig. 13A), the anticyclonic system becomes well organized with prominent equatorward pressure gradient near the southern tip of the Indochina Peninsula. The pressure gradient near Java is also equatorward. Undoubtedly, these equatorward pressure gradients favor low-level convergence and, in turn, contribute to enhance equatorial convections in $EIO_{2}$ in November. By December, the northern hemisphere anticyclonic system further penetrates deep into the South China Sea with an equatorward pressure gradient all the way to the equator. Although the pressure gradient near Java becomes poleward due to development of the NAIM monsoon trough, December is expected to be the month of most active convections in $EIO_{2}$ due to the dominant contribution of the northern hemisphere winter monsoon.

Computed results of each term in Eq. (12) over the $EIO_{2}$ domain are presented in Table 3 (bottom). Although these results are below our expectations, the estimated D agrees, at least qualitatively, with seasonal variations in the convective intensity index $C_{0}'$ in $EIO_{2}$, as previously observed in Fig. 7. Of course, the largest contribution is due to the northern hemisphere winter monsoon in November and December, which exhibits two ways of contributions to enhance equatorial convections, i.e.; first, intensifying the meridional convergence as expressed by term B and, second, augmenting the zonal convergence due to the $\beta$-effect by increasing the surface pressure near Sumatra-Borneo (for example, see Fig. 13B). Also, one should not overlook the role of NAIM upon low-

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Fig. 12. As in Fig. 9, but for monthly mean surface pressure over the tropical Indian Ocean in June (A), August (B) and October (C), respectively extracted from Sadler et al. (1987a).
ering the surface pressure to the east of Java and eventually facilitating convective activation in EIO\(_2\) via the \(\beta\)-effect.

6. Summary and discussion

Equatorially symmetric OLR' (Eq. 1) is introduced as an optimum parameter for describing annual variations of convective activities near the equator. Convective intensity index Co', defined by Eq. 4, is also an appropriate parameter to signify the equatorial convective activities. Semi-annual cycle in the convective activities is predominant over the equatorial continents, such as Africa and South America. The driving force of semi-annual convective variations is, of course, sensible heating at the ground surface due to the annually migrating sun. The role of sensible heating becomes insignificant as one approaches equatorial oceans, such as the Indian (EIO) and western Pacific (EWP) Oceans. In EWP (130\(^\circ\)E-160\(^\circ\)E), an active convective phase begins in November and ends around the end of March, with its peak between January and February. In November, when the surface pressure configuration is nearly symmetric with respect to the equator (Fig. 9B), the pressure trough extends zonally along the equator with substantial zonal as well as meridional down-pressure gradient convergent winds, that are responsible for enhancement of equatorial convections. The southward convergent flow near 10\(^\circ\)N becomes strongest in January in association with the intensification and equatorward invasion of the North Pacific high (Fig. 9D), congruent with most active equatorial convections in January.

The southward migration of the surface pressure trough is an integral part of the gigantic clockwise rotation of low pressure phase encompassing around the western Pacific and Indian Ocean (refer to Fig. 10). During the spring transition, the phase propagation is northward in EIO. However, the spring transition exhibits a peculiar feature and brings about a dry phase in EIO. This is schematically illustrated in Fig. 14. The surface pressure configuration in April is asymmetric about the equator with high pressure over the cooled Indian Ocean as contrasted with low pressure over the heated Southeast Asian continent. Near the southern tip of India, the northward down-pressure gradient winds are estimated to be about 7-8 m s\(^{-1}\), which are much stronger than the northward down-pressure gradient winds emanating out of the South Indian Ocean high. The net effect is a meridional divergence near the equator of EIO, in agreement with a relatively dry weather in the spring transition of April to May (Figs. 4C and 7). Only exception are equatorial land masses of the Malaysian Peninsula, Sumatra and Borneo, where the Laplasian of surface pressure becomes positive due to surface heating and, accordingly, brings about low level convergence and associated convections (refer to Eq. 12).

How and when convections are activated are quite different between EIO\(_1\) (80\(^\circ\)-100\(^\circ\)E) and EIO\(_2\) (100\(^\circ\)-120\(^\circ\)E). An active convective phase of EIO\(_1\) persists five months from about mid-June to mid-November with the peak in October. Contrary to our anticipation, SEAM does not contribute at all to convections in EIO\(_1\). Due to the presence of the dominant SEAM trough, there exists strong down-pressure gradient northward winds (generally referred to as the low-level Hadley circulation) over the Bay of Bengal. This northward wind accelerates poleward and, as a result, causes low-level divergence in the equatorial EIO\(_1\) domain. There is a tendency for such divergence to be compensated for, at least partially, by low-level convergence due to southerly flows emanating out of the cold southern hemisphere high pressure system. As the season advances from August to October, SEAM dissipates while the southern hemisphere high still remains strong (Fig. 12). This is one reason why October is the month of most active convections in EIO\(_1\). Between June and October, the surface pressure is constantly high near the east coast of equatorial Africa due to an invasion of the southern hemi-
sphere ridge system. The eastward pressure gradient induces westerly winds along the equator and contributes to equatorial convergence due to \( \beta \)-effect (Eq. 12). As such, the southern hemisphere winter monsoon is the primary contributor to enhancement of equatorial convections in EIO\(_1\).

Equatorial convections in EIO\(_2\) are of different character. Here, the major contributor is the northern hemisphere winter monsoon bursting out of Siberia. In Fig. 13B for January, we see three prominent ridges extending across Indochina, along the east coast of India, and also along the east coast of Somalia, all of which are associated with the Siberian high. It is the Indochina ridge system that causes strong down-pressure gradient northerly winds sweeping across the South China Sea, and contributing to strong convections in EIO\(_2\), in particular to those occurring over Borneo. December is the month of strongest convections in EIO\(_2\) (Table 3). At this time, the \( \beta \)-effect is also dominant due to a prominent E-W pressure gradient along the equator. Here, note that the E-W pressure gradient is primarily caused by an equatorward invasion of the northern hemisphere winter-time ridge through three preferred routes of Somalia, India and Indochina.

Associated with changes in equatorial convections are distinct variability in the tropical basic flow (\( u' \), \( v' \)). At P61, equatorial convections become fully established off the west coast of Sumatra (Fig. 15A). Further westward of Sumatra is a Rossby-type wind response with equatorial westerlies over the central Indian Ocean (Fig. 15B). Between P61 and P70, equatorial convections near Sumatra diminish, while they are greatly intensified over Borneo (Fig. 15C) due to an increase in the northeasterlies over the South China Sea that are originated from the Siberian High and eventually converging into the convections near Borneo (Fig. 15D). This undoubtedly indicates the contribution of the Asian winter monsoon on the enhancement of convections near Borneo in December. Between P61 and P70, the western Pacific also experiences a marked intensification of equatorial convections. Noteworthy changes emerged in the wind fields are: (1) extraordinary intensification of \( u' \) westerlies near the northern tip of New Guinea which exceed 6 ms\(^{-1}\) and converge into active convections over the western Pacific, (2) eastward of active convections are a Kelvin-type wind response with easterly \( u' \) increasing by about 2 ms\(^{-1}\) from P61 to P70 near 130\(^\circ\)W, (3) an increase of \( u' \) westerlies (easterlies) near 130\(^\circ\)E (130\(^\circ\)W) is indicative of acceleration of the equatorial Walker circulation at P70, which nearly coincides with the strongest annual phase of the E-W overturning, and (4) eastward of about 170\(^\circ\)E are double ITCZs stretching along about 5\(^\circ\)–10\(^\circ\)N, as well as along 5\(^\circ\)–10\(^\circ\)S (Fig. 15B). The southern convergence zone is associated with a NW-SE tilted trough from about (160\(^\circ\)E, 5\(^\circ\)S) to as far southeastern as (130\(^\circ\)W, 25\(^\circ\)S). This South Pacific trough is generally referred to as “SPCZ”. It is the SPCZ that first triggers the onset of NAIM over the western South Pacific (refer to Fig. 2A).

In short, a series of events that lead to the onset of NAIM are as follows: (1) Activated symmetric convections in EWP are responsible for intensification of a Kelvin response over the central Pacific, (2) SPCZ is then activated over the warmer South Pacific and contributes to establishment of NAIM in that vicinity, and (3) NAIM then expands westward into northern Australia and Indonesia, which occurs as a part of the gigantic clockwise phase rotation mentioned earlier. A glance at Fig. 15C, together with Fig. 2A, indicates different structural features between equatorial symmetric convections in EWP and monsoonal asymmetric convections in and around SPCZ. Also, their origin are of fundamental difference; namely, the former being enhanced by southward down-pressure gradient winds

![Fig. 14. As in Fig. 13, but for April with convections (shaded) and down-pressure gradient winds (arrows) over the EOI domain.](image-url)
from the winter-time northern hemisphere high pressure systems, while the latter being induced by low-level convergence within the summer-time southern hemisphere trough system. It is an important prerequisite to clearly separate the equatorial convective system from the monsoonal convective regime, in order to properly describe the monsoon transition between the two hemispheres. This point will be further elaborated on in the subsequent paper by Matsumoto and Murakami (2000).

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Appendix

List of Symbols

- $p_s$ surface pressure
- $u, v$ zonal and meridional winds
- $(\ )'$ symmetric component defined by (1)
- $(\ )''$ asymmetric component defined by (2)
- $OLR$ outgoing longwave radiation
- $C_0'$ convective index as defined by (4)
- $S_0'$ convective area as defined by (5)
- $X_0'$ convective center as defined by (6)
- $CISO$ climatological intraseasonal oscillation
- SST sea surface temperature
- NOAA National Oceanic and Atmospheric Administration
- ECMWF European Centre for Medium Range Weather Forecast
- EAF Equatorial Africa ($0^\circ-40^\circ E, 20^\circ S-20^\circ N$)
- EAM Equatorial America ($110^\circ W-30^\circ W, 20^\circ S-20^\circ N$)
- EIO Equatorial Indian Ocean ($60^\circ-125^\circ E, 20^\circ S-20^\circ N$)
- EWP Equatorial western Pacific ($125^\circ E-160^\circ W, 20^\circ S-20^\circ N$)
- SAFM South African summer monsoon
- NAIM Northern Australia-Indonesia summer monsoon
- SAMM South American summer monsoon
- NAFM North African summer monsoon
- SEAM Southeast Asian summer monsoon
- WNP Western North Pacific summer monsoon
- ENPM Eastern North Pacific summer monsoon
- ENAM Eastern North Atlantic summer monsoon
- WI Winter phase ($P70-P10$)
- SP Spring phase ($P16-P28$)
- SU Summer phase ($P34-P49$)
- FA Fall phase ($P52-P64$)

Fig. 15. As in Fig. 3, but for Pentad 61 (A and B) and Pentad 70 (C and D).
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OLR の赤道対称成分による熱帯における対流活動の年変化

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赤道対称な半旬平均 OLR データを用いて、赤道付近における対流活動の中心域と時期を決定し、対流活動の年変化機構について調べた。赤道アフリカや赤道アメリカの大陸上では、太陽の季節的進行に伴って春と秋に対流活動ピークをもつ半年周期が卓越する。

赤道海洋域における積雲活動は、ロスビーの変形半径内（約 15°N-15°S）で卓越し、その季節変化は主として変形半径外からの遠隔作用によって支配される。赤道西太平洋（EWP）における対流活動発期は 11 月から翌年の 3 月である。西部北太平洋の夏のモンスーン（WNPM）に伴うトラフが 11 月までに赤道に到達すると、気圧場は赤道対称となり、東西および南北方向の気圧傾度でやって収束が起こる。1 月に赤道に向かって張り出していた北太平洋高気圧の南縁で北よりの風の収束がもっとも強くなり、EWP の最盛期を迎える。冬季の EWP は赤道上における波数 1 と 2 の東西循環のエネルギー源となる。地上における低圧帯の位相は、季節とともにインド洋-西太平洋にまたがって時計回りに移動する。すなわち、75°E では春から夏に北進、10°N では夏から秋に東進、155°E では秋から冬に南下、さらに 10°S では冬から春に西進する。EWP における対流活動は、南進位相の時に活性化する。

東南アジアの夏のモンスーン（SEAM）トラフの南側では、気圧傾度に向かって吹く風が赤道域に発散をもたらすので、赤道インド洋（EIO）上の夏季の対流活動は比較的弱い。EIO における対流活動最盛期は、経度 80°E-140°E（EIO1）と 100°E-130°E（EIO2）とで大きく異なる。EIO1 における対流活動は、SEAM が終わってから盛んになり、10 月が最盛期となる。この原因はインド洋高気圧によってもたらされた赤道上の西高東低の気圧傾度が西風を加速し、いわゆる β-効果によって収束が起こるためである。一方、EIO2 ではシベリアからの高気圧の張り出しのために、12 月が対流活動最盛期となる。この冬のモンスーンは、EIO2 上の対流活動に対して二重の効果を及ぼす。まずシベリア高気圧南縁の南シナ海上では、気圧傾度に向かう北よりの風がスマトラやボルネオの対流活動中心域に南北の収束をもたらす。次にアラビア海やベンガル湾へのシベリア高気圧の張り出しは、赤道に沿う西高東低の気圧傾度を増大し、β-効果によって収束をもたらす。地上低圧帯の北進位相、すなわち春には、EIO 上の対流活動は強く、西太平洋と対照的である。

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