Evolution of Tropical Circulation Anomalies Associated with 30–60 day Oscillation of Globally Averaged Angular Momentum during Northern Summer

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Abstract

Intraseasonal variations of upper and lower-level zonal winds, outgoing longwave radiation (OLR) and globally averaged angular momentum (GAM) for northern summers of 1977–1984 are studied using lag correlation statistics. The role of surface frictional drag on atmospheric angular momentum is also investigated by studying the temporal and spatial distribution of surface wind stress in the tropics and its relationship with zonal wind anomalies.

It is shown that the 30–60 day GAM fluctuation is accompanied by zonal propagation of convection and 850 mb zonal wind anomalies in the tropical belt. On the other hand, the associated 250 mb zonal wind and surface wind stress anomalies exhibit pronounced standing oscillations. The magnitude of wind stress anomaly is strongly influenced by the climatological zonal wind in the tropics. Large wind stress anomalies are only found in the tropical central Pacific where surface easterly from the climatological Walker circulation is strong and persistent. Therefore, although the tropical lower-level wind anomalies associated with the 30–60 day oscillation propagate continually from the Indian Ocean to the eastern Pacific, large momentum exchange between the earth and atmosphere only occurs at the time when the low level wind anomalies are in phase with climatological easterlies in tropical central Pacific. From the present results, it is suggested that momentum exchange between the lower and upper troposphere may occur in regions of active convection via vertical momentum transport. The tropical central Pacific plays a key role in the linking the atmosphere and earth through angular momentum exchange in intraseasonal time scales.

1. Introduction

Because of its conserved property, angular momentum is one of the most basic measures of the general circulation of the atmosphere. Recently, many studies have shown that global atmospheric angular momentum (GAM) fluctuates with a variety of time scales in concert with the solid earth. Because of the conservation of total angular momentum, these fluctuations must involve exchange of the momentum between atmosphere and solid earth. The angular momentum change of solid earth is most commonly identified by the change of the rotation rate of the earth estimated by length of day (LOD) measurements (Langley *et al.*, 1981; Rosen and Salstein,

1983). The LOD is known to fluctuate over a wide range of time scales. Meteorologically most interesting is the strong coherent variation between LOD and atmospheric angular momentum fluctuations in periods around 50-days. It has been suggested that this coherent variation is related to the tropical 30–60 day oscillation.

There have been an abundance of recent theories of the tropical 30–60 day oscillation (*e.g.*, Lau and Peng, 1987; Miyahara, 1988; Sui and Lau, 1988 and many others). Most recent theories of the 30–60 day oscillation aim at the mechanism for zonal propagation of the asymmetric part of the motion associated with tropical convection. But the dynamical connection between the tropical 30–60 day oscillation and global angular momentum has never been made

very clear. Most of the above-mentioned linear or quasi-linear theories of the oscillations predict only weak or no zonal mean wind response to localized tropical heating. However, large changes in zonal mean wind, in particular at upper levels are necessary for changes in GAM. Recently Sardeshmukh and Hoskins (1985) showed that nonlinearity of the large scale flow plays an important role in the maintenance of the upper level zonal flow. Therefore, in order to understand the underpinning of GAM fluctuations, it may be necessary to know the mechanism of nonlinear dynamics of the large scale flow and the coupling between angular momentum and convection in the tropics. The documentation of the basic structures of zonal wind and outgoing longwave radiation (OLR) anomalies associated with the change of GAM may provide *a priori* information for further development of a nonlinear theory of intraseasonal oscillations.

In the first part of this paper (Lau *et al.*, 1989; hereafter it is referred to as LKS), the zonal mean statistics of atmospheric AM and outgoing longwave radiation in the intraseasonal time scales are examined. The main results are as follows. The globally averaged upper tropospheric angular momentum (GAM) exhibits a very sharp peak at 50-day period. The intraseasonal fluctuations of GAM are coupled with zonally averaged OLR fluctuations with three basic meridional modes, pertaining to the tropics, subtropics and the extratropics. It was found that the 50-day fluctuation of GAM is contributed mostly from the tropical mode, and subsequently, it was argued that frictional torque is likely to be the main source of GAM fluctuation at the intraseasonal time scale.

LKS further pointed out that the 50-day periodicity is pronounced in the upper-tropospheric zonal wind and in the time series of GAM, but not in the lower level zonal wind. Yet, they suggested that the major source of GAM fluctuation at the 50-day time scale is the surface frictional torque in the tropics. If the angular momentum exchange between solid earth and atmosphere is largely due to frictional drag, then how can a surface wind with little power in the 50-day period give rise to fluctuations of GAM in the same period? This apparent paradox need to be resolved. One possibility is that the horizontal distribution of surface wind stress anomaly may be quite different from the distribution of surface wind anomaly in the tropics, and consequently, the time series of the former is different from the time series of the latter when averaged over the globe. To clarify this, we compare the time evolution of the spatial structure of the surface wind stress with that of the lower-level zonal wind in the time scale of 30–60 days. We also examine if there are preferred geographical locations of surface wind stress that may be related to the 30–60 day fluctuation of

GAM. Only results for the northern summer will be presented in this paper.

2. Data and method

(a) Data

The data sets used in this study are 5-day averages of NMC analyzed 250 and 850 mb winds and outgoing longwave radiation (OLR) from NOAA satellite. The spatial resolution is 5 degree latitude and 5 degree longitude. The 8 summers from 1977 to 1984 are selected and used in the present study. Summer is defined as the period from May 21–25 to September 8–12. The OLR has missing data for 1978. To emphasize the 30–60 day fluctuations, a 31-point bandpass filter was applied to time series of all the variables. Only filtered time series were used in this study. More detailed knowledge of the filtered data and the response function of the bandpass filter can be obtained from Knutson and Weickmann (1987).

(b) Method

Previous studies (Rosen and Salstein, 1983; Krishnamurti and Gadgil, 1984) have shown that total atmospheric angular momentum has a large contribution from the momentum at the jetstream level. LKS show that the total angular momentum at 250 mb level has a pronounced peak at 50-day periodicity. Thus, the time variation of 250 mb total AM may be regarded as an useful indicator of GAM in the intraseasonal time scale, although the former can be different in detail from the latter. In the present study the globally averaged 250 mb angular momentum is calculated as a reference time series for constructing the time evolution of zonal wind and OLR associated with the 30–60 day fluctuation of GAM.

The filtered globally averaged AM (FGAM) is defined as

$$FGAM = \frac{\sum_i (u_i a \cos \phi_i) (2\pi a \cos \phi_i) (a \Delta \phi)}{4\pi a^2}$$

where u is the 30–60 day filtered zonal wind at 250 mb, a the radius of the earth, ϕ the latitude, and the subscript i indicates i -th latitude. $\Delta \phi$ is five degree. FGAM is further divided by the earth radius to elucidate the mean magnitude of zonal wind and is redefined to FGAM/ a throughout this paper.

Lag correlation coefficients of lag from -20 days to $+20$ days with 5-day interval between FGAM and zonal wind, and OLR in the tropical belt are calculated to obtain the time evolution of zonal wind and OLR patterns associated with different phases of FGAM. In computing the lag correlation coefficient, FGAM for each summer is confined to the period from May 21–25 to September 8–12, but the corresponding time series of zonal wind and OLR are lagged. For example, to obtain the -20 day lag

correlation between FGAM and OLR, the period of OLR time series used is that from May 1–5 to August 19–23.

3. 30–60 day fluctuations of GAM and OLR

FGAM at 250 mb level is calculated using the 30–60 day filtered zonal wind during the 8 summers of 1977–1984. As shown in Fig. 1, FGAM varies between about -1 ms^{-1} and 1 ms^{-1} and has 2–3 cycles each summer. Largest fluctuation occurs during 1981 summer and minor variations during 1980 summer. In this paper, the time series shown in Fig. 1 is used as the reference time series for obtaining various lag correlations.

Before obtaining the relationship between the FGAM and OLR, the distribution of OLR variability in the tropical belt is discussed. In Fig. 2 the root-mean-square (rms) of 30–60 day filtered OLR is shown. Large fluctuations are found over the Indian Ocean. The maximum amplitude is about 20 Watt m^{-2} . Relatively large fluctuations are also found in the subtropical Western Pacific and in the ITCZ region along the Eastern Pacific near $5\text{--}10^\circ\text{N}$. Small OLR fluctuations are distributed mostly in climatologically dry regions, *i.e.*, the subtropical eastern oceans in both hemispheres, Sahara, and northern edge of Australia. This rms distribution is similar to that obtained by Lau and Chan (1988), where they showed that 40–50 day variance of OLR contributes about 15–25 % of the total variance in the tropical belt. The 15–25 % contribution of variance indicates that the rms shown in Fig. 2 contributes about 40–50 % of the

Fig. 1. The time series of globally averaged angular momentum divided by the earth radius at 250 mb level as calculated using 30–60 day filtered zonal wind during the 8 summers of 1977–1984. Each summer is separated by dotted lines.

Fig. 2. Distribution of the root-mean-square (rms) of 30–60 day filtered OLR. Unit is Watt m^{-2} . Regions of the rms bigger than 14 Watt m^{-2} are shaded.

about 40–50 % of the total rms.

Figure 3 shows the distributions of lag correlation coefficients between the time series of FGAM shown in Fig. 1 and those of the 30–60 day filtered OLR at all grid points in the domain. The correlation coefficient of 95% significance level is about 0.25^* . At -20 day lag (Fig. 3a), 20 days before the time of maximum FGAM, positive correlations indicating below normal convection appear in the northern subtropical western Pacific. Relatively large convection appears in the tropical central Pacific, and the convection anomaly increases in the Indian Ocean. At -10 day lag (Fig. 3b), the convection in the Indian Ocean is much enhanced and slightly moved north-eastward. At the time of maximum FGAM (lag 0, Fig. 3c), the center of convection is moved further to the east and extended to the subtropical central Pacific, and dry anomalies locate in the subtropical central Pacific of the Southern Hemisphere and over the central America. This pattern is consistent with a local Hadley circulation over the tropical Pacific with rising (sinking) motion in the northern (southern) hemisphere subtropics. Also noted is dry condition developing over the Indian Ocean. At $+10$ day lag (Fig. 3d), the dry condition is much enhanced as it moves over to the eastern Indian Ocean. The spatial pattern of $+10$ day lag correlation is similar to that of -10 day lag, except with the sign reversed. Similarly, the lag correlation pattern at $+20$ day (Fig. 3e) is similar to the negative of the simultaneous correlation shown in Fig. 3c, indicating that the mutual coupling between OLR and FGAM occurs mainly with the time scale of about 40 days. Overall, Figs. 3a–3e are characterized by the OLR anomalies moving from the Indian Ocean to the subtropical central Pacific. It should be mentioned that the characteristics of anomaly, *e.g.*, moving character, can not be determined definitively from the correlation coefficients alone. But we also made maps of correlation coefficient multiplied by rms (not shown), and the discussions in this paper are made based on both maps.

* If we assume a de-correlation time scale of order of 10 to 15 days for the filtered time series, the total number of degree of freedom is approximately 60–90 for the entire data period. This gives a 95 % significance level between 0.25 to 0.21.

Fig. 3. Distributions of lag correlation coefficients between the time series of FGAM shown in Fig. 1 and those of the 30–60 day filtered OLR at all grid points in the domain. Lags are –20 days for (a), –10 days (b), 0 day (c), +10 days (d), and +20 days (e). Contour interval is 0.1. Negative values are represented by broken lines.

4. Time evolution of upper and lower-level zonal winds

The variability of zonal wind in time scale of 30–60 day is shown in terms of rms. The rms of zonal wind at 250 mb level shown in Fig. 4a is distributed rather uniformly in the zonal direction. The minimum appears in the tropical region with the magnitude about $2\text{--}3\text{ ms}^{-1}$ and the maximum appears in the jetstream latitudes in both hemispheres. On the other hand, the 850 mb rms (Fig. 4b) is characterized by a more regional structure, *i.e.*, the wind fluctuation is largest in the tropical western Pacific and Indian Ocean while it is relatively weak in the northern hemisphere extratropics, except in the northern

central Pacific.

We now examine the time evolution of zonal winds at 250 mb and 850 mb levels using the 30–60 day filtered zonal winds. Shown in Fig. 5 are the distributions of lag correlation coefficients between the time series of FGAM and those of 250 mb zonal winds at all grid points in the domain. It is noted that in all figures in Fig. 5, large correlations appear in the tropical region, indicating that 30–60 day fluctuations in GAM originate from the tropics. This is in agreement with the recent results of other investigators, *e.g.*, Madden (1987, 1988), Benedict and Haney (1988), and Risbey and Stone (1988). At 20 days before the time of maximum FGAM, the zonal wind anomalies (shown in Fig. 5a) are mostly negative in

Fig. 4. Distribution of the rms of zonal wind at (a) 250 mb (b) 850 mb. Contour interval is 1 ms^{-1} for (a) and 0.5 ms^{-1} for (b). Shading indicates the rms greater than 4 m s^{-1} for (a) and 2 m s^{-1} for (b).

the tropical belt except in Indonesian region. Large negative correlations appear over the eastern edges of the tropical Pacific and Atlantic Oceans. At -10 day lag (Fig. 5b), the zonal wind anomaly becomes positive in most regions of the tropics. Large positive correlations are found in the subtropical western Pacific and equatorial central Pacific. At the time of its maximum (0 day lag, Fig. 5c), FGAM is accompanied by positive zonal wind anomalies in the tropics and subtropics, with maxima over the subtropical central Pacific in the Southern Hemisphere and the eastern tropical Pacific and Atlantic. At $+10$ day lag (Fig. 5d), the westerly anomalies are much reduced in most tropical regions and the easterly anomalies become significant over Indonesian region. At $+20$ day lag (Fig. 5e), easterly anomalies prevail in the tropics. It is noticed that the 30–60 day fluctuations of 250 mb zonal wind are characterized by apparent standing oscillations occurring with two nodes located in Indonesian region and the east coast of South America.

The time evolution of 850 mb zonal wind associated with the change of FGAM is shown in Figs. 6a–e. At 20 days before the maximum FGAM (Fig. 6a), the lower-level wind is mostly easterly. Large easterly anomalies are found in the subtropical Indian Ocean of the Northern Hemisphere and the subtropical Pacific in the Southern Hemisphere. At -10 day lag (Fig. 6b), the anomalies migrate to the tropical central Pacific and westerly anomalies become evident in the western Indian Ocean. At 0 day lag (Fig. 6c), the easterly anomalies move further to the

east to northern Central American. But the westerly anomalies over the Indian Ocean are much enhanced and spread into the subtropical Indian Ocean and Indonesian region. As shown in Fig. 6d and 6e, the positive anomalies migrate continuously to the east and die out in the eastern Pacific at 20–30 days after the maximum FGAM. At $+10$ day lag, the easterly anomalies originated in the Indian Ocean propagate to the east and the above cycle is repeated.

Overall, the time evolution of lower-level wind is characterized by a pattern of eastward propagation from the Indian Ocean to South America. It is also interesting to note that 20 days before the maximum FGAM, lower-level wind anomaly is mostly easterly in the tropics, but after the maximum FGAM, low level westerly anomaly dominates the tropics. In other words, westerly angular momentum extracted from the solid earth by the surface easterly wind stress precedes or leads to westerly acceleration in the upper-troposphere, resulting in the positive peak of FGAM 20–30 days later. After the maximum FGAM, the globally averaged angular momentum decreases due to the destruction of westerly momentum at the surface.

Comparing Fig. 6 with the OLR and 250 mb zonal wind in Figs. 3 and 5, the following broad picture emerges. Generally, lower-level wind anomaly is convergent (divergent) over the region of increased (reduced) convection. Convection and lower-level wind anomalies propagate coherently to the east from the Indian Ocean to the eastern Pacific in the tropics. However, the 250 mb zonal flow exhibit an

Fig. 5. As in Fig. 3 except the lag correlation between FGAM and the 30–60 day filtered 250 mb zonal winds at all grid points in the domain.

Fig. 6. As in Fig. 5 except the lag correlation coefficient between FGAM and the 30–60 day filtered 850 mb zonal winds at all grid points in the domain.

essentially stationary feature with the center over Indonesian region shifting abruptly to the eastern Pacific. Although there is some indication of upper-level divergence (convergence) over the region of increased (reduced) convection, it is difficult to understand the stationarity of the upper-level wind anomalies in terms of the simple linear response to a moving heat source (*e.g.*, Gill, 1980; Kang, 1986). This suggests that the linear dynamics is not enough to explain the mutual coupling between 30–60 day fluctuations of convection and upper-level circulations in the tropics and that the mechanism of vertical transport of westerly momentum associated with the 30–60 day propagation of tropical convection and the nonlinearity of the large scale flow itself may be important. It is also interesting to note that the zonal winds in the subtropical region of the Southern Hemisphere has a barotropic vertical structure. This barotropic structure may be a result of Rossby wave propagated from tropical convection region to the extratropics in the winter hemisphere and may be related to the tropical-extratropical meridional mode discussed in LKS.

5. 30–60 day fluctuations of surface wind stress

The globally averaged atmospheric angular momentum can be changed only by two mechanisms: mountain torque and frictional torque. LKS and Madden (1987) suggested that intraseasonal fluctuations of global angular momentum are caused by the frictional torque in the tropics and closely related to the 30–60 day oscillation of tropical convection. Our results so far support this notion. The frictional torque acting on an atmospheric strip of finite width around a latitude circle is proportional to the surface wind stress integrated around the latitude circle. But as discussed in the introduction, low level wind and surface wind stress are not necessarily the same. It is necessary to delineate the surface wind stress distribution and its evolution in time with the change of FGAM. In this section, the anomaly of surface wind stress varying within the time scale of 30–60 days is estimated using filtered zonal wind and climatological mean wind at 850 mb level. To compute the surface wind stress properly, surface wind data may be required. Since this is not available for the present study, we used 850 mb data. For the 30–60 day time scale, the wind stress calculated from the 850 mb wind may provide only an qualitative measure of the surface momentum exchange.

The stress in zonal direction (τ_x) is usually expressed as

$$\tau_x = -C_D u |V| \quad (1)$$

where u is the zonal wind, $|V|$ is the total wind speed, and C_D is the drag coefficient. In this study

C_D is taken to be the constant value of 1.5×10^{-3} . The wind is decomposed into the climatological mean denoted by the over-bar and transient part denoted by the superscript prime $u = \bar{u} + u'$ and then $|V|$ is approximated as

$$|V| \approx |\bar{V}| \left(1 + \frac{\bar{u}u'}{|\bar{V}|^2} \right) \quad (2)$$

In the derivation of (2), it is assumed that $2\bar{u} \gg u'$ and v' , and $\bar{u} \gg \bar{v}$. Figures 7a and 7b clearly show that the magnitude of the climatological zonal wind during the summer is much larger than that of the meridional wind in the tropical belt. Substitution of (2) and $u = \bar{u} + u'$ into (1) gives, after neglecting a small nonlinear transient term,

$$\tau_x \approx -C_D \left[\bar{u}|\bar{V}| + u'|\bar{V}| + \frac{\bar{u}^2 u'}{|\bar{V}|} \right]. \quad (3)$$

The wind stress expressed as in (3) can be decomposed into the climatological and transient parts, $\tau_x = \bar{\tau}_x + \tau'_x$. If the transient component of zonal wind u' in (3) is replaced by the 30–60 day filtered zonal wind, the zonal component of surface wind stress in the time scale of 30–60 day can be approximately estimated with the following equation (positive in the sense of atmosphere gaining westerly momentum):

$$\tau'_x = -C_D \left[u'|\bar{V}| + \frac{\bar{u}^2 u'}{|\bar{V}|} \right]. \quad (4)$$

The surface wind used to calculate surface wind stress is estimated crudely from the wind at 850 mb level by simply taking 0.7 of the 850 mb wind. Madden (1987) used 0.75 same as that suggested by Wylie *et al.* (1983). However, as pointed out by Madden, the factor 0.75 produces a slight overestimate of the surface wind. The 0.7 factor is a typical value of the ratio estimated by McDonald (1938) for the equatorial area. Therefore, the estimated 30–60 day filtered surface wind stress (FSWS) is about half (exactly 0.49) of the value calculated by substituting both the climatological wind and 30–60 day wind anomaly at 850 mb level into Eq. (4).

The spatial distributions of $\bar{u}$ and $|\bar{V}|$ at 850 mb level, which are used to calculate FSWS in Eq. (4), are shown in Figs. 7a and 7c, respectively. Strong easterlies are found in the tropical Pacific and Atlantic. The easterly in the tropical Pacific is a manifestation of the persistent Walker Circulation during the summer. In other tropical regions the zonal wind is weak. The absolute magnitude of total wind shown in Fig. 7c is characterized by a strong regional structure. This structure can be inferred from Fig. 7a since $\bar{u}$ is much larger than $\bar{v}$ in the tropics, as mentioned earlier in this Section. In the tropics, the maximum of $|\bar{V}|$ is located at the eastern equatorial Pacific.

Fig. 7. 850 mb climatological means during the northern summer. (a) is the zonal wind at 850 mb level, (b) the meridional wind, and (c) the absolute magnitude of the wind. Shading in (c) indicates the magnitude greater than 6 ms^{-1} .

The time evolution of FSWS associated with changes of FGAM is illustrated in Fig. 8. The surface wind stress anomaly shown in the figure is obtained using Eq. (4), where the anomalous zonal wind (u') at each time lag is calculated by multiplying the lag correlation coefficient at each time lag shown in Fig. 6 by the rms of 850 mb zonal wind. The anomalous zonal wind thus calculated represents the 850 mb zonal wind anomaly associated with 1 ms^{-1} change of FGAM. At -20 day lag (Fig. 8a), FSWS has weak positive values in most regions of the tropics, although relatively large anomalies are found in several localized regions. At -10 day lag (Fig. 8b), large positive anomalies are found in the tropical Pacific and negative anomalies become evident in the eastern Indian Ocean. At 0 day lag (Fig. 8c), positive wind stress anomalies are weakened in the tropical eastern Pacific, but negative anomalies are extended in the Indian Ocean

and the subtropical western Pacific. In the tropics, regions of positive and negative wind stress anomalies are separated at the dateline during the time of maximum FGAM. The distribution is consistent with the low level wind response from an equatorial heat source near the dateline. At $+10$ day lag (Fig. 8d), the wind stress anomaly in the tropical Pacific is replaced by a well organized pattern of large negative value. The spatial pattern of Fig. 8d resembles a mirror image of that of wind stress at -10 day lag. At $+20$ day lag (Fig. 8e), the spatial extent of negative anomaly is decreased and more localized in the tropical eastern Pacific.

Overall, the time evolution of FSWS is characterized by alternating positive and negative wind stress anomalies of large magnitude with well organized spatial pattern in the tropical Pacific. It is noted that the time variation of FSWS in the Indian Ocean and Indonesian region is relatively weak and

Fig. 8. Distributions of surface wind stress anomaly varying in the time scale of 30–60 days. The wind stress is computed based on Eq. (4) in text. (a) and (b), respectively, are the distributions at 20 days and 10 days before the maximum of FGAM, (c) at the time of maximum FGAM, (d) and (e), respectively, at 10 days and 20 days after the maximum of FGAM. Contour interval is $3 \times 10^{-3} \text{ m}^2 \text{ s}^{-2}$.

not well organized. As shown in the previous section, the time evolution of 850 mb zonal wind anomaly is characterized by a pattern of almost continuous zonal propagation. But the surface wind stress shown above is characteristic of a standing oscillation. The reason is due to the asymmetric distribution of climatological zonal wind. The anomalous wind stress as expressed by Eq. (4) can be approximated by $\tau'_x \approx -2C_D|\bar{u}|u'$, since $|\bar{u}| \approx |\bar{V}|$. Therefore, in the Indian Ocean and Indonesian region where $\bar{u}$ is small, the wind stress should be weak although anomalous lower-level zonal wind is large, particularly in the region of zero $\bar{u}$. On the other hand, in the tropical Pacific where $\bar{u}$ is large the wind stress anomaly is much larger than that of other regions in the tropics for the same anomaly of lower-level zonal wind.

It is also noted that FGAM has a 90 degree phase difference in time with the surface wind stress anomaly averaged over the globe, as pointed out by Madden (1987). In Fig. 8, the surface wind stress anomalies at the time lags -10 and -20 days are positive in most regions of the tropics, but reversed situation is true for the anomalies at $+10$ and $+20$ day lags. These indicate that the positive angular momentum is continuously supplied by (depleted from) the tropical earth surface for about 20 days before (after) the time of maximum FGAM. Also it can be inferred from Fig. 8c that the wind stress anomaly averaged over the domain is about zero at the time of maximum FGAM.

6. Conclusion

In the present study, life cycles of atmospheric circulation features are constructed to illustrate the time evolution of upper and lower-level zonal winds and convection anomalies during a typical 30–60 day variation of globally averaged angular momentum. The spatial pattern of OLR anomaly is similar to that of a typical 30–60 day oscillation with convection anomalies moving from the Indian Ocean to the subtropical central Pacific during Northern summer (Lau and Chan, 1985). The time evolution of 850 mb zonal wind anomaly is manifest in an eastward propagating pattern with convergence (divergence) over the region of increased (reduced) convection. However, the time evolution of upper-level zonal wind anomaly is characterized by a standing oscillation occurring with two nodes located in the east coast of south America and Indonesian region. It is difficult to understand the stationary nature of the 250 mb zonal wind anomalies in terms of a simple Gill-type linear response to tropical convection.

It is also shown that there is a preferred geographical location of large surface wind stress anomaly in the tropical Pacific. The time evolution of surface wind stress anomaly associated with the variation of 30–60 day fluctuation of global angular momen-

tum is characterized by alternating positive and negative anomalies in the tropical Pacific, associated with fluctuations of the Walker circulation. The wind stress anomalies in the Indian Ocean and Indonesian region, where the climatological zonal wind is weak, is relatively small and not well organized. Also shown is that for about 20 days before (after) the peak of global angular momentum, maximum westerly angular momentum of the atmosphere is supplied by (depleted from) the earth surface in the tropical Pacific.

Results in the present study supports the basic picture of angular momentum exchange between the earth and atmosphere proposed by LKS. In addition, the following scenario of 30–60 variation of global angular momentum is suggested (Fig. 9). First, anomalous tropical convection and circulation associated with the 30–60 day oscillation originates in the Indian Ocean and propagates eastward. At this time the lower level wind anomalies are located in the Indian and western Pacific Oceans, and the momentum exchange between the earth surface and atmosphere is insignificant because of the weak climatological surface wind over those regions (Fig. 9a). However, when the convection anomalies reach the Indonesian region, low level easterly wind anomaly increases in the tropical central Pacific enhancing the climatological easterlies in the Walker circulation. During this period, large positive angular momentum is supplied from the earth surface to the atmosphere. The positive angular momentum supplied is transported to the upper-troposphere via strong rising motion in regions of active convection. This is manifest in the strong westerly wind anomalies found over the maritime continent (Fig. 9b). As the deep convection propagates into the central Pacific, the convection anomaly is reduced but the globally averaged angular momentum anomaly reaches its maximum magnitude. The center of the upper level westerly wind shifts to the equatorial eastern Pacific/South America region in a way resembling a see-saw or standing oscillation of zonal wind between the maritime continent and South America (Fig. 9c). This is followed by a reduction in GAM associated with the emergence of a region of upper level easterlies over the maritime continent and flux of westerly angular momentum from atmosphere to solid earth (Fig. 9d). The largest sink of atmospheric angular momentum occurs over the central Pacific at 20-day lag (Fig. 9e). Here the distribution of upper level zonal wind anomaly is completely reversed from those at 0-day. It is important to note that the equatorial central Pacific region appears to be the region where large momentum exchange takes places between ocean and atmosphere during all phases of the 30–60 day oscillation. The timing of large momentum exchange and the importance of tropical central Pacific in the momen-

Fig. 9. Schematics showing qualitative features of upper and lower level zonal wind, tropical convection and surface wind stress during a complete cycle of the tropical 30–60 day oscillation. The maximum zonal winds is depicted by heavy dark arrows. Direction of angular momentum exchanges between earth and atmosphere is indicated by broad hatched arrows.

tum exchange are consistent with those suggested by Madden (1987). The possible manner in which the surface wind stress sets up east-west sea-level pressure gradient to provide a net torque on the solid earth is discussed in LKS.

The standing oscillation in upper level zonal wind associated with 30–60 day GAM variation cannot be explained in terms of simple linear wind response to tropical heating. Nonlinear effects in the maintenance of upper level zonal flow and the vertical transport of angular momentum in the region of convection needs to be further documented and understood. Furthermore, the change of zonal wind due to surface friction and/or changes in meridional eddy momentum flux (zonal mean of $u'v'$) from the tropics to the extratropics may also be important. Jhun (1987) showed that the meridional tilt of eddy struc-

ture is sensitive to the magnitude of friction in the tropics. His result is consistent with those obtained by Hayashi (1970) and may support the above consideration. These are topics for future research.

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全球平均角運動量の 30–60 日振動にともなう熱帯循環の時間発展

(北半球が夏の場合)

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1977–1984 年の北半球が夏の季節について、対流圏上層下層の東西風、OLR、全球平均角運動量 (GAM)、の季節内変動間の相互関連についてラグコリレーションを使って調べた。また、大気角運動量変動に対する地表面摩擦の役割についても、熱帯地表面ストレスの時間空間変動と東西風変動の関連を探ることにより調べた。

30–60 day GAM の変動は、熱帯地域を東方に移動する対流活動と 850 mb 東西風変動を伴うことが示された。他方、250 mb 東西風や地表面ストレスの変動は滞在波的振舞いを示す事がわかった。ストレス変動の大きさは熱帯東西風の気候値に大きく依存する。大きなストレス変動は、平均的な Walker 循環に伴う強い東風が存在する熱帯中部太平洋に見られる。このことは、30–60 day 振動に伴う下層風はインド洋から東太平洋まで連続的に移動するのに対して、地球と大気間の角運動量交換は、主に下層風変動と気候値的な意味での偏東風の位相が一致する時期に熱帯中部太平洋地域で起こる事を示している。また、大気角運動量の上層下層間の交換は対流活動に伴う鉛直方向の運動量輸送によって行われている事が示唆される。熱帯中部太平洋地域は、季節内振動時間スケールの地球・大気角運動量交換に重要な役割を果たしている。