Dynamics of Super Cloud Clusters, Westerly Wind Bursts, 30–60 Day Oscillations and ENSO: An Unified View

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Abstract

In this paper we present results of several preliminary numerical experiments to study the mechanism of the observed variabilities in wind and convection associated with super cloud clusters(SCC), westerly wind bursts and 30–60 day oscillations in the western Pacific region. Using the Lau and Peng(1987) model, we show that the generation of eastward propagating precipitation pattern associated with the 30–60 day oscillation can be identified as a SCC. During its development stage, the SCC is accompanied by convective clusters moving in opposite direction(westward) to the SCC. Our results suggest that these westward propagating cloud clusters are associated with Rossby waves produced at the development stage of the 30–60 day disturbance due to mutual adjustment of the large scale flow and heating. If the boundary forcing is zonally symmetric and in the absence of other external forcings, the SCC eventually settles down to an organized eastward propagating precipitation pattern accompanied by a planetary east-west circulation cell in the equatorial zonal plane. It is found that intraseasonal oscillation in the tropical atmosphere is a multi-scale process. Three basic spatial scales are identified, i.e., synoptic scale motions associated with a cloud cluster(1000–2000 km), size of the SCC complex(~2000–4000 km) and planetary circulation scale(wavenumber 1 and 2). The first is associated with westerly wind burst, double cyclone formation and high-frequency fluctuations of order of several days. The second and third are associated with slow eastward propagation associated with the 30–60 day oscillation.

The relationship between SCC and westerly wind burst is studied by examining the detailed spatial and temporal variation of the model SCCs as they propagate over zonally varying sea surface temperature. It is found that wave-CISK heating in the lower troposphere as well as surface heating are crucial in leading to the strongly asymmetric low level westerly wind bursts similar to those observed in the real atmosphere. The implication of the present results in terms of the development of a unified theory of low frequency oscillation is also discussed.

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1. Introduction

In a recent observational study using GMS high-resolution satellite IR data, Nakazawa (1988) showed that the 30–60 day oscillations in the equatorial western Pacific region is associated with a large scale convective complex (hereafter referred to as a super cloud cluster, SCC) which is comprised of many short-period, synoptic-scale convective clusters. Each SCC has a horizontal scale of several thousand kilometers and appears to move eastward with a phase speed of about 5–10 m/s along the equator (see Fig. 1). Individual cloud cluster within a SCC fluctuates with a period of several days and propagates westward with comparable phase speed.

It has also been known for some time, that the climatological surface easterlies over the equatorial western and central Pacific often undergo short period (2–5 days) episodic fluctuations reversing to strong westerlies. These rather energetic fluctuations are known as “westerly wind bursts” (Nitta and Motoki, 1987, Luther *et al.*, 1983). The occurrence of westerly wind burst is at times associated with the formation of a symmetric cyclone pair straddling the equator (Keen, 1982, 1987). It has also been shown that the “westerly wind burst” is closely related to and may be instrumental to the onset of ENSO (Murakami and Sumathipala, 1988 and Lukas, 1987).

Independent of the above observations, many recent studies have shown that strong convective activities associated with the 30–60 day oscillations are confined to the Indian Ocean/Western Pacific region (Knutson and Weickmann, 1987, Lau and Chan, 1985, 1986a). In particular, Lau and Chan (1985) found that as the 30–60 day oscillation propagates into the equatorial western Pacific region, it seems to stall and spawn additional convection before the convection dies off as the waves propagate into the eastern Pacific. It has been pointed out that over warm water in the tropical western Pacific, the 30–60 day oscillation may act as a trigger leading to the onset of the El Niño Southern Oscillation (ENSO) (Lau and Chan, 1986b, 1988).

There are three basic features that are common to the SCC, westerly wind burst and the 30–60 day oscillation: (i) they are all collocated and have maximum amplitude over SST of 28°C or higher *i.e.*, the western Pacific and (ii) they are associated with the appearance of surface westerly wind and enhanced convection over the equatorial western and central Pacific and (iii) they all seemed to be strongly modulated by the ENSO. These observations raised a number of obvious questions. Is there a relationship between SCC and westerly wind burst? How can one dynamically explain the eastward propagation of the SCC and the westward propagating individual clusters imbedded therein? Is the propagation and or-

ganization of SCC an intrinsic part of the 30–60 day oscillation? Finally, is there a relationship between the above phenomena and the onset of ENSO and if so, is the onset effected through the action of individual occurrence or a collection of all the above? These are the questions that a viable theory of low frequency oscillations in the tropics must answer. In this paper, we show that the embryo of such theory is now emerging. Needless to say, much more work is needed to substantiate the results presented here. We hope the results in this paper will lay the groundwork and stimulate more future research in these areas.

In a recent paper, Lau and Peng (1987) (hereafter referred to as LP) showed, based on idealized numerical experiments and eigenvalue calculations, that the 30–60 day oscillations can be interpreted as unstable Kelvin-type modes destabilized by the so-called mobile wave-CISK mechanism. The basic tenet of this theory is now supported by a large number of independent theoretical studies (*e.g.* Chang and Lim, 1988., Wang, 1988, Miyahara, 1987, Takahashi, 1987 and others). The mobile wave-CISK mechanism requires a positive feedback between low level moisture convergence and latent heating. In addition, it restricts positive latent heating only in region of rising motion, and allows no corresponding negative heating in the descending part of the motion. Such restriction is equivalent to transporting perturbation energy from high to low wavenumbers and likely to contribute to the planetary scale selection of the model 30–60 day oscillations. In LP's experiments, they found some westward propagating convection initially, but once the eastward propagating 30–60 day disturbance becomes established, the westward propagating disturbances become strongly damped. Because of the idealized initial conditions used in LP, the full range of possible response to initial conditions was not explored.

In this paper we present some preliminary results of numerical experiments to study the mechanism for the observed variabilities in wind and convection in the western Pacific region. We shall show that under more general initial conditions, the generation of a 30–60 day eastward propagating precipitation pattern in the LP model which can be identified as SCC, is accompanied by convective clusters moving in opposite direction to that of the SCC itself. The results suggest that the westward propagating cloud clusters are produced at the initial stage of the 30–60 day disturbance due to mutual adjustment of the large scale flow and heating. The relationship between the SCC and westerly wind burst is studied by examining the detailed spatial and temporal variation of the 30–60 day SCCs under the influence of zonally varying sea surface temperature. Finally, a discussion of the present results in terms of the development of a unified theory of low frequency os-

Fig. 1. Time-longitude section of deep convection index based on three-hourly GMS IR data and twice-daily 850 mb wind averaged between 2.5°N–2.5°S during the period November 3 to 23, 1986.

cillation will be presented.

2. Observational background

This section is devoted to a selective discussion of some basic features of observations that are directly related to the numerical experimentation in this paper. It is not intended to be a detailed description of any of above-mentioned phenomena. For such descriptions, the readers should be referred to references cited in Section 1 and many other works available in the literature. Fig. 1 shows the time-longitude section of an index of deep convection from three-hourly GMS IR data and twice-daily 850 mb wind from the Japanese Meteorological Agency for the period of Nov. 3–23, 1986. The figure which is reproduced from Nakazawa (1988) embodies the

essential features of all the phenomena discussed above. During the above period, a strong 30–60 day oscillation was found to propagate from the Indian Ocean to the western Pacific and might have led to further warming of the western and central Pacific during the 1986–87 ENSO (Nitta and Motoki, 1987). The eastward migration of the SCC associated with the 30–60 day oscillation is evident in the movement of the centers of convection from around 100°E to 180° from the beginning to end of the period. Individual cloud clusters appear as westward propagating features in intervals of one to two days. A closer look suggests that the eastward and westward movements appeared to be closely coordinated *i.e.*, a SSC moves eastward in discrete manner due to the successive amplification or forma-

tion of new cloud clusters further east, which propagates westward and becomes damped or moved away from the equator. Strong low level westerly wind appears to accompany the convective outbursts resulting in several strong "westerly wind bursts" around Nov.11–13, 14–16, 18–23. It is also noticed that the low level wind appears to have a wavy appearance (strong meridional wind component) immediately to the west of the convection centers in the region 120–150 E. The above features will be relevant to the numerical simulations to be discussed subsequently.

3. SCC and 30–60 day oscillation simulation

a. Model description

The model used for simulation is the same as in LP, –5 vertical level spectral model with rhomboidal truncation to 15 wavenumbers. A semi-implicit time differencing scheme with Robert-Asselin type averaging is used in the integration. This provide strong damping of gravity waves in the model. Both the fully nonlinear and the quasi-linear (positive-only latent heating, no advection) version of the model are used for the following experiments. In the nonlinear case, the nonlinear terms are computed on a Gaussian grid at a resolution of approximately 7.5 degree by 4.5 degree. The vertical heating profile used has maximum heating between 500 and 700 mb (see Sui and Lau (1988)) and section 6 for detailed discussion. The atmosphere has no mean motion and the horizontal mean temperatures are given by the observed equatorial zonal mean temperatures, with fixed specific humidity. In all experiments discussed in this section, the sea surface temperature (SST) is uniform at 27°C. Initial perturbation quantities are all zero except the velocity potential or divergence at the lowest model level (900 mb) which is given by the following function

$$F(x) = A \sum_{m=1}^{15} \cos\{m(\lambda - r(m))\}$$

imposed at the grid points along the 2°S and 2°N latitude and equals zero elsewhere, where A is the amplitude, λ is the longitude and $r(m)$, for $m=1, 2, \dots, 15$ are randomly generated phases. Four experiments are carried out. In the first three, no dissipation terms are included in addition to the damping effect inherent in the numerical scheme of the model. In the fourth, linear dissipation is included. These experiments are:

Case A, quasi-linear with initial 900 mb divergence given by $F(x)$.

Case B, quasi-linear with initial 900 mb velocity potential given by $F(X)$.

Case C, same as B, except nonlinear.

Case D, same as B, except with Rayleigh friction and Newtonian cooling with an e-folding time to 10 days

b. Organization and propagation of super cloud clusters

Figs. 2 and 3 show the first 12-day time-longitude cross section of the total latent heating (in precipitation unit of mm/day) at the equator in case A and B respectively. Since the magnitude of the perturbation used is different for different cases, the actual amount of precipitation should not be taken too seriously. In the following we focus the discussion on the variation of the precipitation pattern. Case A and B corresponds respectively to initial random perturbation in the small scale (divergence) and large scale (velocity potential) field. Although the initial conditions are spectrally quite different in these two cases, the equatorial rainfall patterns that eventually developed are quite similar, particularly in the major convective regions. Initially, westward propagating disturbances appear to dominate. As will be shown later, these are Rossby waves excited by the initial large imbalance in the mass and wind field. However, eastward propagating disturbances appear to outgrow the westward propagating disturbances after about 10 days. They do so by forming new deep convection/precipitation centers eastward of the existing convective centers which simultaneously spin off westward propagating disturbances about every 2 days. The convective complex including the eastward and westward propagating disturbances will be hereafter identified as a model SCC. The transient westward propagating features are greatly weakened at the end of about 12 days. At this stage, the eastward propagating disturbance consists of a single eastward propagating SCC resembling that discussed in LP. The eastward migration speed of the SCC is about 9 ms⁻¹. This corresponds to a recurrent time around the equator of about 50 days, in agreement with that predicted from linear analysis of LP. The westward propagating disturbance are somewhat faster (16 m/s) than the average eastward speed (10–15 m/s) and are damped while moving away from the source region. The spatial scale, the propagating speed and the life-time of individual convective clusters are quite comparable to those shown in Fig. 1. Eventually (results not shown for case A and B but may refer to case C and D and LP) the westward propagating disturbances die out while the eastward disturbance grow stronger and so that the structure of the model SCC becomes less similar to that shown in Fig. 1. The above results suggest that the complex structure of the observed SCC is not part of the final equilibrium structure generated by the mobile wave-CISK mechanism of LP but arises as the atmosphere undergoes an adjustment towards that equilibrium via the interac-

tion of the mobile wave-CISK and the pre-existing, unbalanced wind field.

The effect of nonlinear advection is included in case C. Fig. 4 shows the time-longitude plot of the rainfall for this case. The growth of the eastward propagating disturbances and the organization of the SCC from westward propagating individual convective clusters is obvious. In this case, the time taken for the disturbance to grow to the mature stage is of the order of 25–30 days, much longer than required for the linear case B. The eastward propagation speed is somewhat slower at approxi-

mately $8\text{--}10\text{ ms}^{-1}$. At the end of about 30 days, a number of SCCs emerge near the longitude from 0 to 120° . As in the linear case, the stronger the eastward propagating component of the SCC, the weaker the westward propagating component. Since the same initial conditions are used as in case B, the larger variability in phase speed and the generation of secondary SCCs in this case are apparently due to nonlinear advection.

To investigate the effect of dissipation on the structure and propagation of the model SCC, we repeat case B but includes Rayleigh friction and New-

Fig. 2. Time-longitude section of precipitation associated with model super cloud cluster during the first 12 days of integration from random initial conditions in the divergence field (Case A). Maximum precipitation is about 5 mm/day.

Fig. 3. Same as in Fig. 2, except for Case B.

tonian cooling with e-folding time of 10 days. Fig. 5 shows the rainfall pattern for this case. The same general features as the other cases can be seen. Initially, only westward disturbances are present. Between 5 to 12 days, as the SCC grows, westward moving disturbances appear to emanate from "parent" precipitation centers which leapfrog eastward by generating successive centers downstream. At the end of 36 days, the three distinct, well-organized eastward propagating SCC precipitation patterns emerge. As in the previous cases, the eastward propagation becomes increasingly well developed, the less apparent the westward propagation appears. By the end of 20 days, the westward propagating distur-

bances disappear completely. Dissipation appears to damp out the westward but not the eastward propagating disturbances.

c. Horizontal velocity field

In order to unravel the detailed structure and evolution of the circulation pattern associated with the SCC and its westward propagating components, Figs. 6 a-d show the time sequence at a half day interval of the 900 mb streamlines (with the velocity vectors plotted to indicate the relative strength) associated with the westward moving cloud clusters at the developing stage of the 30–60 day SCC for Case A. We have examined the detailed spatial and

Fig. 4. Same as in Fig. 2, except for Case C.

Fig. 5. Same as in Fig. 2, except for Case D.

Fig. 6. Time sequence showing horizontal distribution of the 900 mb circulation (solid contour) and precipitation (dashed contour) pattern at 12 hour interval for selected window shown for Case A in Fig. 2.

temporal evolution of the velocity field during the four distinct intensification and eastward movement of the SCC shown in Fig. 2. They all show very similar behavior. Thus only the results for the limited domain shown in Fig. 2 are shown. In Fig. 6a, when a precipitation center is developed along the equator, the low level easterly inflow to the heat source has little meridional velocity component indicating a Kelvin wave structure. On the other hand, the cyclonic inflow on the western side of the precipitation center is characteristic of a Rossby wave. As the convection and circulation become more intense in the next 12 hours, excessive inflow from the Rossby wave creates a divergence pattern (reduced precipitation) symmetric about but off the equator to the west of the precipitation center while the Kelvin wave pulls the convergence center to the east. As a result two precipitation centers appear (Fig. 6b). The splitting is complete in the next 12 hours. After the splitting, the precipitation associated with the Kelvin wave first weakens or even dies out completely due to the interaction with the vertical motion associated with the Rossby wave, but it eventually re-intensifies. Soon a new precipitation center due to the intensification of the unstable Kelvin wave re-appears about 10 degree to east of the initial position. On the other hand, the Rossby wave which is stable with respect to mobile wave-CISK, decays and propagates westward continually. Secondary intensification of the Rossby waves many also occur as they move into new convergence areas left behind by previous transients. The signature of Rossby waves in the form of "twin cyclones" straddling the equator and moving away from the main precipitation center is very clear in Fig. 6c and d. As a result, strong wave-like features are found in the westerlies to the west of the heat source while relative steady easterlies are found on the eastern side of the SCC. This feature was noted in the discussion of Fig. 1. The above process is repeated about every 2 days, leading eventually a mature and well organized eastward propagating SCC whereby the Rossby waves are strongly damped.

d. Effect of the mobile wave-CISK on zonal scale selection

It is obvious from the previous result that the eventual organization of the heating pattern into a 30–60 day oscillation from apparently random initial conditions involves vigorous spatial re-scaling, through the generation of Rossby and Kelvin wave modes. It is seen from Fig. 2–5 that in spite of the initial presence of all spatial scales, the lowest few wavenumbers rapidly dominate the flow in less than 10 days. This is because the positive-only latent heating transports perturbation energy from high to low wavenumbers. Figs. 7a and b show the shift of the spectra of the kinetic energy normalized by

the corresponding total energy, at the 700 mb level in the vicinity of the equator for the case B and C respectively. In Case B, for the first 5 days or so, the spatial scale is between wavenumber 10 to 2 and the dominant scale is wavenumber 2 after the initial re-scaling. The only nonlinearity in this case is that due to the positive-only heating. The energy spectrum becomes quite steady even though the amplitude of the perturbation is continuously growing (not seen in Fig. 7, because of the normalized spectrum). In the nonlinear Case C, the dominant spatial scale in the first 5 days or so is wavenumbers 2 to 10 as in Case B. Subsequently, the dominant scale shifts from wavenumber 1, then 2 and again 1 at the end of the simulation and it is conceivable that it may still vacillates further. The spectral shift in wavenumber due to the positive-only heating in the mobile wave-CISK mechanism for case B and C is shown in Fig. 8a and b respectively. Here, the difference between spectra of the vertical velocity (or heating field) before and after imposing the positive-only heating condition is shown as a function of time (see also LP for details). Dashed and solid contours indicate energy loss or gain relative to wavelike heating respectively. The transfer of energy from high to low wave number as a result of the restriction of heating only in the localized region of rising motion is obvious. The largest loss occurs for short waves (wavenumber 15) and the largest gain occurs at the long waves (wavenumber zero to two) for both case B and C. The effect of nonlinear advection (case C, Fig. 8b) apparently cause more variability in the scales of the large scale motion as they evolve.

4. Westerly wind burst experiments

a. Vertical structure of westerly wind associated with SCC

It is clear from LP and other recent studies (*e.g.* Takahashi 1987) that the propagation speed and vertical structure of the 30–60 day oscillation is sensitive to the level of maximum heating in the troposphere. The vertical zonal wind distribution of the SCC associated with the 30–60 day oscillation for different vertical heating distributions is examined in this section. Three cases are considered with a CISK heating profile having

- I Maximum heating near 300 mb,
- II Maximum heating between 500 and 700 mb and
- III Heating profile as in Case II and surface evaporation and sensible heating proportional to the 900 mb wind according to a bulk aerodynamic formulation. All the sensible heating is assumed to be realized in the lowest model layer (1000–900 mb, see Sui and Lau, 1988 for details).

When the heating is concentrated in the upper troposphere (Case I, Fig. 9a), the vertical scale of the disturbance is large, the westward tilt with height extend throughout the troposphere. The westerly wind is maximum in the upper troposphere and to the east of the heat source. At the surface, the westerly (Rossby) is slightly stronger than the easterly (Kelvin) inflow towards the heat source. The eastward propagation speed for this mode is about 23 ms^{-1} (20 day oscillation). When the heating maximum is between 700 and 500 mb (Case II, Fig. 9b), the vertical scale is reduced and the westward tilt with height is limited to the lower troposphere. At the surface, the westerlies flow is stronger

than the easterlies, suggesting the increasing importance of Rossby waves. This mode propagates with an eastward phase speed of about 9 ms^{-1} (50-day oscillation). When latent and sensible heat flux is included, the vertical distribution of the SCC changes drastically (Case III, Fig. 9c). First, the vertical phase tilt is reduced and the zonal wind maximum is further away from the convective center. Second, the east-west asymmetry of the surface zonal winds become very pronounced. Strong surface westerlies are found in a shallow layer to the west of the convective center and the easterlies are considerably weaker. The eastward propagation speed is about 13 ms^{-1} (36-day oscillation). Evaporation provided

Fig. 7. a. Normalized energy spectra for 700 mb flow near at the equator as a function of time for Case B.

b. Same as in Fig. 7a, except for case C.

Fig. 8. a. Difference in energy spectra of vertical velocity before and after positive-only heating as a function of time for case B. Spectra are normalized by the instantaneous total energy.

b. Same as Fig. 8a, except for case C.

Fig. 9. Vertical cross-section of zonal wind mobile wave-CISK mode for different heating profiles. a) maximum heating near 300 mb. b) maximum heating between 500 and 700 mb and c) same as in b) but include surface heating.

additional source of moisture for deep convection through wave-CISK. More importantly, it appears sensible heating is responsible for the presence of the shallow layer of westerlies in the wake of the eastward propagating SCC. Thus, low tropospheric heating and in particular, surface heating may be important in leading to an asymmetric low level westerly wind in the model. This may have important implication on the connection between SCC and the westerly wind burst phenomenon in reality.

b. Effect of zonally asymmetric SST

The effect of asymmetric SST and surface heating on the structure and propagation of mobile wave-CISK modes have been examined in detail in Sui and Lau (1988). The following is a simplified discussion of an experiment focusing on westerly wind burst. The SCC structure shown in Fig. 9c is allowed to propagate through a zonally varying SST distribution as shown in Fig. 10 which mimics the climatological SST distribution along the equator. The meridional distribution is assumed to be Gaussian, with a e-folding width of 10° latitude. Here, the effect of the varying SST is to produce a varying amount moisture available for moist convection through evaporation as well as to provide different amount of surface heating as a function of the air-sea temperature difference and wind. The detailed evolution of the SCC vertical structure and phase speed as a function of the longitude is similar to that shown in Sui and Lau (1988) and will not be repeated here. Only features deemed to be related to SCC and westerly wind burst will be highlighted below.

Fig. 11a shows the longitude-time section of precipitation for the first 96 day of the integration. As the SCC (maximum precipitation region) propagates over the warm water, it undergoes vigorous dynamical adjustment. This is manifest in the eastward propagation of the SCC due to Kelvin waves and the westward propagation of convective ele-

ments associated with Rossby waves. The SCC disappears over the eastern Pacific, due to the inhibition of deep convection over the cold water. It revives somewhat over the warm and moist Amazon region, before disappearing over the Atlantic and finally recovered over the Indian Ocean/Western Pacific region. The cycle appears to be repeated indefinitely about every 36 days. Although the SCC appears to turn on and off at different longitude, it appear to follow more or less the same longitude-time path indicating a somewhat constant underlying phase speed. This is because the large scale zonal winds in the lower troposphere maintained both by convection and surface heating are basically continuous over the entire journey around the equator (Fig. 11b). The wavy appearance of the surface westerlies behind and the relatively steady easterlies ahead of the mobile heat source is quite apparent. It is also noticed that this east-west asymmetry is most pronounced over the warm part of the ocean between longitude 90 and 180. These features are similar to that shown in Fig. 1.

Fig. 12 shows the time series of 900 mb zonal wind at a fixed point around 160°E . The strong asymmetry and the front-like reversal between the westerly and the easterly component of the zonal wind is clearly seen. The easterly component is quite steady but the westerly component is very spiky and contains many high frequency fluctuations. This is related to the strong east-west asymmetry of the intrinsic structure of the SCC noted in Fig. 9c and the global picture described in Figs. 11a and b. As the SCC propagates eastward, an observer over the warm ocean will first experience the steady surface easterlies due to Kelvin waves. The passage of the convective center marks the abrupt shift from easterlies to westerlies. This is followed by high frequency zonal wind fluctuation in the westerly phase due to Rossby cyclonic couplet shed off by the main heat source as it propagates further to the east. The above situation is repeated in 36-day cycles.

Fig. 10. Spatial variation of SST used for the westerly wind burst experiment.

For each cycle, the westerly phase of the oscillation lasts about 20–25 days and the easterly phase about 10–15 days. Qualitatively the overall scenario is remarkably similar to the observed westerly wind burst, although the transition from westerly to the easterly phase in the model is less sharp compared with the observed wind fluctuation over the equatorial central Pacific (*e.g.* Luther *et al.*, 1983).

5. Discussions and conclusions

From the results presented, it appears that a hierarchy of spatial scales and substructures are embedded in the 30–60 day oscillations. Three basic spatial scales can be identified: the size of the SCC complex including the eastward and westward propagat-

ing convection centers and the circulation around it (~ 2000 – 4000 km, see Figs. 2–3); the size and separation associated with individual cloud clusters (~ 1000 – 2000 km) and the extent of the large scale circulation (wavenumbers zero, 1 and 2). Condensational heating in localized region of the SCC is responsible for the generation of the planetary scale motions associated with the 30–60 day oscillations.

One of the key results of this paper is that the mobile wave-CISK mechanism is responsible for two basic dynamical regimes in the tropical atmosphere. In the *adjustment regime*, there is a large imbalance between the large scale circulation and the heating fields. As a result, vigorous dynamical adjustment occurs via Kelvin and Rossby waves. The for-

Fig. 11. Time-longitude section of a) precipitation (mm/day) and b) 900 mb wind (ms^{-1}) associated with the model SCC.

Fig. 12. Time series of 900 mb zonal wind at 160°E showing the strong asymmetry between the westerly and the easterly phase of the model 30–60 day oscillation.

mer is responsible for the slow eastward migration of the SCC, and the latter accounts for the high-frequency fluctuation in surface westerlies, formation of twin-cyclones or vortices straddling the equator, and westward propagation of individual cloud clusters within the SCC. In the absence of external forcing, the dynamical fields eventually approach an *organization regime*. In this regime, the dynamical and heating fields are in approximate dynamic (non-stationary) equilibrium, dominated by Kelvin waves. Because Kelvin waves are unstable with respect to mobile wave-CISK, they are selectively am-

plified. Due to their non-dispersive nature, they can sustain the growth of the collective motion of the heat source and circulation in the form of individual eastward propagating SCC. The *organization regime* is manifested in the uniform eastward propagating, organized 3-D structure found in LP, Hayashi and Sumi (1986) and many observational studies of the 30–60 day oscillations.

The westerly wind burst experiment suggests that as the intensity of the deep convection changes due to zonally varying SST, the circulation and heating field are continuously forced out of balance and

Hierarchy structure of intraseasonal oscillations over the Tropical Pacific Ocean

Fig. 13. Schematic showing the hierarchy structure of intraseasonal oscillations over the tropical Pacific as inferred from the model simulation. Horizontal scales shown are only approximate. Large overlap may occur within the super cloud cluster substructures.

hence dominated by the adjustment regime. The structure and propagation of SCCs in the adjustment regime agree remarkably well with the observed SCCs and westerly wind burst, considering the crudeness of the present model. In reality, given that the tropical atmosphere is constantly being perturbed by a wide range of forcings from the lower and lateral boundaries, and new clusters forming and decaying within the tropics due to nonlinear scale interactions, a well adjusted circulation and heating field may not be possible in a short time. Even in the absence of external forcing, it is found from our experiments that an adjustment time scale of the order of 10–20 days is required before a well-organized SCC emerges. If this can be applied to the real atmosphere, this means that the convection complex associated with the 30–60 day oscillation initially generated over the Indian Ocean and western Pacific would have reached over the cold water of the eastern Pacific where convection is inhibited. In other words, there will not be sufficient time for the circulation and heating to adjust over the warm water region ($SST > 28^{\circ}\text{C}$). Hence, we expect that truly organized eastward propagation of SCC may occur only infrequently. This suggests that the adjustment regime as manifested in the structure and propagation of SCCs found by Nakazawa (1988) and related westerly wind fluctuation should be a common occurrence over the western Pacific.

The above discussions lead us to the important question of the origin of westerly wind bursts. The results show that the sudden occurrence of westerly wind over the open warm water region (equatorial western and central Pacific) may be due to

- (a) a direct response of the lower tropospheric wind to the passage of the eastward propagating main SCC heat source associated with the 30–60 day oscillation and/or
- (b) zonal wind fluctuations on the equatorial side of the twin cyclonic vortices emanating from the main heat source.

The former is directly related to strong convection/precipitation while the latter may or may not be accompanied by strong convection. This may be the reason why the most intense convection is often found near the eastern terminus of the westerlies and that some westerlies outbursts are not related *locally* to convection (Keen, 1987). The rapid expansion of the region of surface westerly is the result of the SCC complex extending both eastward (by Kelvin waves) and westward (by Rossby waves) during the adjustment regime. A summary of the above discussion is illustrated schematically in Fig.13.

Finally, it is noted that the results presented here is entirely consistent with the view that the 30–60 day oscillation may be instrumental in triggering the

onset of ENSO (Lau, 1985, Lau and Chan, 1986b, 1988). The most recent work of Lau and Shen (1988) suggests that a new set of unstable coupled ocean-atmosphere modes arises as a result of including the physics of the 30–60 day oscillation in the atmospheric model. This suggests that the entire hierarchy of motions depicted in Fig. 13 may be dynamically unstable when coupled to the ocean. Ongoing work on the extension of the present multi-layer atmospheric model to include ocean-atmosphere interaction indicates that the mobile wave-CISK modes are further modified by coupled interactions. Such interactions may further link the SCC, westerly wind burst and the 30–60 day oscillation as a major players in the onset of ENSO.

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超雲団、西風バースト、30-60 日振動そして ENSO の力学： ある統一した視点

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本論文では、西太平洋における超雲団 (SCC)、西風バーストそして 30-60 日振動と関連した風と対流の観測された変動の機構を解明するための初歩的な数値実験の結果が示される。Lau and Peng (1987) のモデルを用いると、30-60 日振動と関連した東進する降水パターンの生成が SCC として認められる。その発達期において SCC は SCC とは逆に西進する対流クラスターを伴っている。私たちの結果は、これらの西進するクラウドクラスターは、大規模場の流れと加熱の相互調節によって、30-60 日周期の発達期にできるロスビー波と関連していることを示唆している。もし下面境界からの強制が東西一様で、他の外力もないならば、SCC は、赤道東西面で惑星規模の東西循環セルを伴った、組織化された東進する降水パターンに次第に落ち着く。熱帯大気での季節内変動は複合スケールの過程であることがわかった。3 つの基本的な空間スケールが認められる。すなわち、クラウドクラスターと関連した総観規模運動のサイズ (波数 1 と 2)。最初のスケールは西風バースト、対低気圧の生成、そして数日のオーダーを持つ高周波変動と関連している。2 番目、3 番目のスケールは、30-60 日振動のゆっくりした東進と関連している。

SCC と西風バーストとの関係を、海面水温が東西で変化している下でのモデル SCC の空間・時間変動を詳細に調べることによって研究した。海面からの加熱と同様に対流圏下部での wave-CISK 加熱が、実際の対流で観測される強い非対称性を持った下層の西風バーストを形成する上で重要であることがわかった。低周波振動の統一的理論の発展から見た本結果の示唆する点についても議論される。

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