Effects of the Mean Walker Circulations on the Zonal Variability of Tropical CISK Waves

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ABSTRACT
An idealized zonally asymmetric mean state in the Tropics is used to investigate the effects of atmospheric mean Walker circulations and sea surface temperatures on the zonal variability of tropical waves with positive-only wave–CISK heating.

First, the effects of the zonally symmetric vertical and meridional shears of the mean zonal flow are investigated with different vertical profiles of convective heating and temperature lapse rate. With the observed vertical profile of temperature lapse rate in the Tropics, the effect of the vertical shear of the mean zonal flow on the fast waves (period about 2 weeks) associated with a maximum heating in the upper troposphere is found to be opposite to that on the slow waves (period about 4 weeks) with a maximum heating in the middle troposphere, while the vertical shear effects with an idealized constant lapse rate are quite similar. The meridional shear of the mean zonal flow acts only to reduce the effects of the vertical shear of the mean zonal flow.

Second, with the existence of atmospheric mean Walker circulations, tropical waves increase and decrease alternately when propagating around the globe. It is found that the influence of the mean Walker circulations is mainly from the effects of the vertical shear of the mean zonal flow, and the tendencies of the zonal variations for the fast waves and the slow waves are almost opposite.

Third, the amplitudes of zonal variations of the tropical waves as a consequence by the atmospheric mean Walker circulations are comparable to those arising from the zonally asymmetric SSTs in the Tropics. It is found that the combined effect of the mean state makes the fast waves propagate over the colder regions without decreasing, while the slow waves strongly anchor over the warmer regions, implying other processes might be important in sustaining the slow waves over the colder regions where convection is weaker.

1. Introduction
Planetary-scale tropical waves are of various frequencies. Generally, they can be divided into fast and slow tropical waves with periods around 1–2 weeks and 6–7 weeks, respectively. The slow waves are most significant in the tropospheric Tropics, whereas the fast waves can propagate vertically and are, in part, responsible for mean flow oscillations in the stratosphere. Zonal variability of these waves has been found in both observations and GCM simulations (e.g., Hayashi 1974; Hayashi and Golder 1980; Rui and Wang 1990). Investigations with linear models (Hayashi 1970, 1971; Lindzen 1974) have addressed the topic of tropical CISK–waves in a resting atmosphere and with convective heating that does not vary spatially nor temporally with surface properties, such as the sea surface temperature (SST). Therefore, their models cannot be used to study the variability (both in time and in longitude) of tropical waves.

The most important surface factor in affecting the atmospheric convection, which drives tropical waves, is the SST in the Tropics. In observations (e.g., Graham and Barnett 1987; Rui and Wang 1990; Salby and Henderson 1994), deep convection seems to anchor over certain regions with high SST. This is quite understandable because the air in these regions is moist since there is substantial evaporation over the warm sea surface. Numerical simulations (Sui and Lau 1989; Lau et al. 1989; Miyahara 1987), by including the SST effect on tropical convection, showed that tropical waves significantly decrease while propagating over colder regions. Other authors (e.g., Wang and Rui 1990) also suggested that the seasonal meridional movement of maximum SST regions can affect the excitation of tropical Kelvin waves, since Kelvin waves are most effectively excited when the highest SST is located at the equator. This

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seasonality of tropical convective activity was also observed by Salby and Hendon (1994). Observations (Yasunari 1979 and 1980; Krishnamurti and Subrahmanyan 1982; Lau and Chan 1986) showed that meridional movement accompanies the prevailing eastward propagating perturbations in the Tropics. Li and Wang (1994) found that with convective heating modified by SST the asymmetry (about the equator) of the SST distribution causes meridional propagation of tropical wave activity.

Another possibly important factor for the variability of transient tropical waves is the mean wind state. Zhang and Webster (1989) investigated the effect of the meridional shear of the mean zonal flow on the structures of tropical waves. They found that the meridional shear of the mean zonal flow has the most profound influence on Rossby waves, while its influence on Kelvin waves is almost negligible. Tropical convection, however, was not included in their model. The effects of vertical shear of the mean zonal flow on the structure and excitation of tropical waves have been investigated recently (Lim et al. 1990; Zhang and Geller 1994; Shen et al. 1996). Their results showed that the vertical shear of the mean zonal flow has selective effects on the excitation of tropical waves. Specifically, westward (eastward) propagating tropical waves are preferentially excited with westerly (easterly) vertical wind shear [westerly (easterly) shear means westerly (easterly) wind increasing with altitude] in both the linear wave–CISK case and the positive-only wave–CISK case (no “convective” cooling with wave-induced descending motions), except only eastward propagating waves are excited in the positive-only wave–CISK case. This implies that a seasonality of tropical waves may result from the seasonal variation of the vertical wind shear.

In the Tropics, however, the atmospheric mean Walker circulation produces significant zonal asymmetries. The mean Walker circulation in the Tropics is somewhat related to the SST distribution in the Tropics although consideration of all of the processes involved in the formation of the tropical mean state is very complicated. Thus, the effect of the mean wind state may be considered partly as another aspect of the SST effect on propagating tropical waves. The main purpose of this study is to see the effects on tropical waves that arise through the mean Walker circulation compared with the effects of the zonal distribution of SST in the Tropics. The asymmetry (about the equator) of the mean Walker circulation and SST is not investigated in this study.

Since tropical Rossby–gravity waves were not excited by wave–CISK (e.g., Shen et al. 1996) and were found to be probably related to extratropical wave activity (e.g., Hayashi and Golder 1978; Zhangvil and Yanai 1980, 1981; Yanai and Lu 1983), the variability of Rossby–gravity waves is not discussed in this study. A positive-only wave–CISK heating is used so that the tropical waves in this study are the eastward propagating Kelvin waves (or wave packets). The difference between the so-called fast and slow tropical waves in this study is uniquely determined by presumably different vertical distributions of the convective heating. Many other possible mechanisms for the origin of slow waves (Madden–Julian waves) have also been suggested. One is phase-lagged wave–CISK (e.g., Davies 1979; Cho et al. 1994), in which the convective heating lags the lower level large-scale convergence. Another is evaporation–wind feedback (Emanuel et al. 1994), which argues that convective heating above the surface is mainly determined by the surface heat exchange. Wang and Rui (1990) suggested frictional wave–CISK and emphasized the role of boundary friction in tropical wave selection in period and zonal scale. The origin of tropical waves is far from resolved at this time. Since the mean wind state, such as vertical wind shear, affects wave growth by affecting the wave phase structure (Zhang and Geller 1994; Shen et al. 1996), mean wind state effects on tropical waves might exist in all of the above theories. It should be pointed out that in our following study with wave–CISK, the slow phase speed results from the prescribed vertical distribution of CISK heating with its maximum at a relatively low altitude (about 400 mb). Tropical convective heating is important for tropical wave activity but the heating structure is very complicated and is not well understood at this time. Hence, it is possible that this vertical distribution may not represent the actual situation for the observed slow waves. Precisely speaking, this study deals with tropical waves resulting from wave–CISK heating with the prescribed vertical distributions. Some features of the resulting waves are shown in this study.

A linearized equation model with prescribed mean state (atmospheric wind state and SST), which is described in Shen et al. (1996), is used for the investigation (see appendix for model equations). In section 2, the observed tropical mean states and the idealized mean states that are used in this study are discussed. In section 3, the mean temperature lapse rate influence on the effect of vertical shear of the zonally symmetric mean zonal wind (hereafter, referred to as vertical wind shear) on wave–CISK is investigated since this has not been addressed by previous studies (e.g., Lim et al. 1990; Zhang and Geller 1994; Shen et al. 1996). Section 4 gives results on the effect of zonally asymmetric mean states on tropical waves. Also, a comparison between the zonally asymmetric mean state effect and the zonally asymmetric SST effect is done. A discussion of the representativeness of the model results to the real atmospheric situation and the possible applications of these results are given in section 5. Section 6 is a brief summary of the main points of this study.

2. Mean states in the Tropics

a. Mean states in the observation

This section discusses “data” used in this study from the European Centre for Medium-Range Weather Fore-
casts (ECMWF) analysis. Figure 1 shows the 3-yr (1986–88) averaged monthly mean zonal wind fields at both upper and lower levels in the troposphere for January (top row), April (second row), July (third row), and October (bottom row), respectively. In the lower troposphere, easterlies are dominant in the Tropics except there are westerlies over the Indian Ocean (~40°E–100°E) and western Pacific (~110°E–160°E). The zonal wind and its vertical shear are seen to vary significantly with longitude. Obviously, in the upper troposphere, strong westerlies are dominant away from the Tropics, with large meridional shears of the mean zonal flow around 15°N(S).

The 12°S–12°N averaged zonal wind was used as the tropical zonal wind. The altitude–longitude cross sections of the asymmetric portion of the tropical zonal wind with the zonally averaged zonal winds subtracted are shown in Fig. 2. There are two distinct regions of vertical wind shear that show relatively little annual shift in longitude. These two distinct vertical wind shear regions were found to be related to the mean Walker circulations (its associated vertical velocity is seen in Fig. 3).

Fig. 1. ECMWF monthly mean zonal wind (unit: m s⁻¹) fields for January, April, July, and October, respectively (upper to lower).
The magnitudes of the monthly averaged meridional winds in the Tropics (not shown) were found to be much smaller than those of the zonal winds. Also, meridional winds of small zonal scales dominate in the Tropics. This implies that the large-scale mean vertical motion should be mainly determined by the zonal wind, according to the continuity equation. The vertical velocity field will be discussed in the next section. Also, the zonal variation of temperatures (not shown) was found to be very small throughout the entire tropical atmosphere, except at the surface where variations of about 2°C were found. Although the meridional gradient of temperature is appreciable in the monthly averaged temperature field, this gradient is less significant in the equatorial regions.

b. Mean state used in experiments

In the previous section, we have seen that the details of the tropical mean state are rather complicated but, in principle, a large zonal scale Walker circulation dominates. To focus on the impact of this dominant mean circulation on propagating tropical waves, we use an idealized mean state instead of the observed one. This idealized mean state is obtained as follows: First, we compute the $12^\circ$S–$12^\circ$N averaged monthly mean vertical velocity in January at 500 mb; second, we use this mean vertical velocity to parameterize a stationary heating forcing.

$$ F = \begin{cases} M\eta(p)\cos\left(\frac{\varphi}{15^\circ}\right)W(\lambda) & \text{if } \left|\frac{\varphi}{15^\circ}\right| < 1 \\ 0 & \text{otherwise,} \end{cases} $$

where $M$ is a constant giving the intensity of the forcing. Here, $\eta(p)$ represents the vertical heating profile. In the calculation, $\eta$ is taken to be maximum at 450 mb. Here, $\varphi$ denotes the latitude, $\lambda$ the longitude, and $W$ the vertical velocity at 500 mb. Next, we put this forcing into a 16-layer linearized model without mean flow to obtain...
the solutions for the wind and temperature fields (the description of such models can be found in Lau and Peng 1987; Shen et al. 1996). Since the problem is linear, both analytic and numerical approaches can be used to obtain the solutions. These wind and temperature fields, so obtained, are used as the mean state in our investigation. There are two purposes for the application of this idealized mean state: 1) to exclude the large extratropical winds while keeping the meridional shear in the tropical mean flow; 2) to eliminate small-scale wind features and assure that the mean flow is a solution of the continuity equation.

Examining these results, we checked the magnitudes of the upward motion at different levels over the central Pacific (−180°) for the model-generated case, and found that they (not shown) are quite close to the observations (the Fourier-filtered large-scale motion). The meridional wind is not shown, but its magnitude was found to be much smaller than that of the zonal wind. This is in agreement with tropical observations. The formation of the zonally asymmetric temperature field in the Tropics and its relation to the mean flow are very complicated. Our linearized model with a prescribed heating is unable to simulate this temperature field well even though it does simulate the mean zonal wind well. However, our numerical results indicated that the model-generated temperature perturbation is very small (<0.5°C). Thus, in the following experiments, the model-simulated zonally asymmetric horizontal and vertical winds will be used as the zonally asymmetric mean wind state and the mean temperature field is replaced by its zonal mean.

3. Mean temperature lapse rate influence on the effect of zonally symmetric vertical wind shear on wave excitation

The effects of the zonally symmetric vertical wind shear on the excitation of tropical waves have been discussed, but with a constant mean temperature lapse rate, +6°C km⁻¹ in the troposphere (Lim et al. 1990; Zhang and Geller 1994; Shen et al. 1996). Before discussing cases with more complicated mean states, possible influences of the lapse rate on these previous results related to the effects of the zonally symmetric vertical wind shear on wave excitation are discussed. The effects of the meridional shear of the mean zonal flow are also discussed in this section.

The solid line in Fig. 5a shows the vertical profile for the temperature lapse rate (unit: °C km⁻¹) used in the previous work. The dotted line is the vertical distribution of the saturated adiabatic lapse rate. The saturated adiabatic lapse rate at a certain altitude is defined as the rate of temperature decrease with respect to height for a saturated parcel when this parcel is adiabatically displaced upward. Figure 5b shows the zonally averaged vertical profiles of the observed and the saturated adiabatic temperature lapse rates in the Tropics (10°S–10°N) for January. The zonal and annual variations of the vertical lapse rate profile (not shown) were found to be negligible. Obviously, CISK is possible only at altitudes where the saturated lapse rate is smaller than that for the mean temperature. The tropospheres in Figs. 5a and 5b are seen to be conditionally unstable, but the most unstable layers are at different altitudes in these two cases. In Fig. 5a, the lower troposphere is more unstable than the upper troposphere, while the upper troposphere is more unstable than the lower troposphere in Fig. 5b. The actual mechanisms for tropical convec-

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The zonally averaged vertical velocity has been subtracted out.
tion are related to mesoscale processes, and these are not well understood at the present time. In our experiments, prescribed wave–CISK heating profiles are used. However, since different vertical heating profiles imply different phase speeds, the difference between those shown in Figs. 5a and 5b may make the role of the vertical wind shear significantly different for the waves of different phase speeds in these two cases.

Figure 6a shows the vertical profiles for the heating and the temperature lapse rate used in the experiments. Heating profiles 1 and 2 are associated with fast waves with periods of about 2 weeks and slow waves with periods of about 4 weeks, respectively, in the positive-only wave–CISK case. Lapse rate profiles 1 and 10 represent the observed and the constant lapse rate cases shown in Fig. 5, respectively. Between them, 8 profiles are inserted, with equal intervals at the same altitude.

To examine wave behavior, we use the linearized equation model by Shen et al. (1996). As in Shen et al. (1996), with the given lapse rate profiles and vertical wind shears, we integrate the model long enough until a single wave or wave packet dominates the resulting picture. First, positive-only wave–CISK is used with no mean flow. Figure 6b shows that the growth rate, $d \ln A / (dt)$ ($A$ is amplitude of temperature or velocity perturbation), for the fast waves decreases with the increasing number of the lapse rate profile used, while the growth rate of the slow waves is relatively insensitive to these changes in the lapse rate profile.

Figure 7a shows the tropical ($12^\circ$S–$12^\circ$N) averaged vertical velocity (shaded area denotes upward motion) and temperature [solid (dotted) line: higher (lower) than the zonal-mean temperature at the same altitude] distributions in longitude and pressure for the fast waves with the constant lapse rate in the troposphere. It is seen that this wave packet has a backward phase tilt for the temperature field up to about 200 mb and an opposite phase tilt for both the temperature and vertical velocity fields above. It can also be seen that the wave packet obtains kinetic energy from the available potential energy in the upper troposphere, where the temperature and vertical velocity are positively correlated, while kinetic energy is lost above the tropopause. Although the pictures are slightly different (not shown) for different profiles of temperature lapse rate, the above features are the same. Figure 7b is the same as 7a but for the slow waves. It is similar to Fig. 7a, but the phase tilt transition
occurs at lower altitudes. The mechanism for the observed Madden–Julian oscillations is still unclear although many possible mechanisms have been proposed. However, it is interesting that the fast and slow waves (defined by the vertical profile of wave–CISK heating) have the observed feature that the slow waves are only present in the troposphere while the fast waves propagate upward.

Experiments have been performed to show the effects of the vertical wind shear on wave excitation with different vertical lapse rate profiles. The profiles of the vertical wind shear used in the experiments are shown in Fig. 8a in which profile 1 approximately represents the maximum easterly shear and profile 11 represents the maximum westerly vertical wind shear in the Tropics. Profiles 2–10 are inserted with equal intervals at the same altitudes. Figure 8b shows that the effect of the vertical wind shear in exciting fast waves depends on the temperature lapse rate profile. The vertical profiles of temperature lapse rate 1, 6, and 10 refer to those given in Fig. 6a. For the case with the constant lapse rate profile 10, the growth rate of the fast waves increases (decreases) with increasing easterly (westerly) vertical wind shear. This is consistent with the conclusions of Shen et al. (1996) with positive-only wave–CISK heating where constant vertical wind shears with height were used; however, the effect of the vertical wind shear on the wave growth is seen to be very sensitive to the lapse rate profile. For the observed lapse rate profile 1, the growth rate of the fast waves decreases (increases) with increasing easterly (westerly) vertical shear. This is opposite to that in the case with lapse rate profile 10. The wave growth rate for the lapse rate profile 6 case is seen to be relatively insensitive to the vertical wind shear.

It was found (not shown) that with linear wave–CISK heating, the previous conclusions about the vertical wind shear effect on fast wave growth with a constant lapse rate profile in Shen et al. (1996) still qualitatively applies in the case with the observed lapse rate profile. In the presence of westerly vertical shear, however, the preference for the westward propagating waves is less manifest with the observed lapse rate profile than is the case with the constant lapse rate profile.

Figure 9a is similar to Fig. 8b except that the mean zonal flow is confined to the tropical region with an e-folding meridional scale of 15°, which is similar to that of the equatorial westerlies and easterlies in Fig. 4. It can be seen that the confinement (to the equator) of the vertical wind shear reduces the effect of the vertical wind shear on wave excitation but this confinement does not change the relationship between the vertical

![Fig. 7. The height–longitude profiles of the tropical vertical velocity (shadow: upward motion) and temperature (solid/dotted line: higher/lower than the mean temperature at the same altitude). (a) Fast wave packet; (b) slow wave packet. The longitude positions of the ascending regions in the figure are not meaningful because the wave packets are mobile.](image)
wind shear and the wave growth. Figure 9b is similar to 9a but for the slow tropical waves. The solid line is for the case with the constant lapse rate and is very similar to the results of Lim et al. (1990). Changes in the lapse rate have little influence on the wave growth. It can be seen that the westerly vertical shear suppresses the wave growth, whereas the influence of the easterly vertical shear on wave growth is insignificant.

4. Effects of zonally asymmetric mean state on the tropical waves

a. Effect of the zonally asymmetric mean state

In this section, we use the same linearized model and methods that were used in the previous section but with the model-generated zonally asymmetric mean wind state (see Fig. 4) to investigate the effect of the mean Walker circulation on the tropical waves. The observed vertical profile of zonally averaged mean temperature in the Tropics is used. Rayleigh friction and Newtonian cooling are used with an $e$-folding scale of 5 days. Zonally asymmetric SST effects are not included in this section.

Positive-only wave–CISK heating is used. Strictly speaking, the use of positive-only heating makes the mean state and waves no longer separable; however, in the positive-only heating case, the ascending region shrinks to occupy a small zonal region, which is much smaller than the dominant scale of the mean vertical velocity. Thus, the amplitude of the zonally concentrated ascending region should be much larger than that of the mean vertical velocity, so using the positive-only heating determined by the waves alone is a reasonable approximation.

In the following experiments, the efficiency factor for the heating is chosen to make the zonally averaged growth rate be zero. Figure 10 shows the time evolution (from upper to lower) of the vertical velocity field at 500 mb (left) and temperature field at 250 mb (right) for the fast waves. It can be seen that the ascending region (solid line: upward motion) is more zonally concentrated than the warm region (with temperature higher than the zonally averaged mean temperature). The period of the wave packet is about 2 weeks. For the zonal
Fig. 10. Time evolutions (upper to lower) of the vertical velocity field at 500 mb (left) and temperature field at 250 mb (right) for the fast waves. The time interval between two successive panels is two days.

wind perturbations (the meridional wind perturbations are much smaller), planetary zonal scales (not shown) were found to be dominant. The wave structures were found to be very similar to those in the case without mean flow. However, as can be seen, the intensity of the wave packet changes with longitude. Both the vertical velocity and temperature perturbations tend to decrease over the Indian Ocean and western Pacific where
the vertical wind shear is easterly, and to increase over the eastern Pacific (~170°W–80°W) where the vertical wind shear is westerly.

The vertical profile of the wave structure at each time is not shown but was found to be very similar to the fast wave case shown in Fig. 7a. Since the upward motion has almost no tilt in the troposphere, the temperature and vertical velocity are positively correlated in the upper troposphere. The decrease (increase) of the wave packet is seen to be associated with the scale of the warmer region in the upper troposphere being less (more) zonally concentrated. This is consistent with the discussion in Shen et al. (1996) about the effect of vertical wind shear on the selective excitation of the tropical waves in the positive-only heating case.

Figures 11a and 11b show the normalized zonal variations of the amplitude of the maximum upward motion with longitude for fast and slow waves, respectively (the zonal variation here is equivalent to the time evolution for propagating waves). The most significant difference between the fast waves and the slow waves is that the fast waves diminish over the western and middle Pacific but grow when moving over the eastern Pacific, while the slow waves develop oppositely. It can be seen that the increases (decreases) of the maximum upward motion at different levels are not exactly in phase, but this is expected for two reasons. First, the vertical profiles of the amplitude for the maximum upward motion are not the same with the different “local” mean states at different longitudes. Second, it takes time for the evolving pattern to adjust to a stable pattern (with structures unchanged) subject to a “local” mean state. Thus, for the propagation of the wave packet, its wave structure is always in “adjustment.” The more in-phase relation between 500- and 850-mb curves in the slow wave case (Fig. 11b) is quite understandable since the slow waves have more time for their adjustment to the stable patterns determined by the “local” mean states. Fortunately, even for the fast wave case, the above complication does not keep us from seeing the general tendency. The above features in the longitudinal variation were also found for the amplitude of the temperature perturbation (not shown).

To see the relative importance of the mean zonal flow in the mean Walker circulation effect, experiments with $\overline{u}(x, y, z)$ only are performed. The zonal variations of the velocity and temperature were found to be very similar to those in Figs. 11a and 11b. Thus, we can conclude that the influence of the other fields (mainly the vertical velocity field) on the excitation and zonal variation of tropical waves is less important than that of the zonal flow where the vertical wind shear effect is much more important than the meridional wind shear effect, as was discussed in section 3. Experiments with the term, $w' \frac{\partial \overline{u}}{\partial z}$ ($w'$ denotes the wave-related vertical velocity), excluded and all the others present were also performed. The exclusion of this term makes the results significantly different and the range of the zonal amplitude variation much smaller.

It was also found that the influence of the mean zonal flow by the zonal advection terms, $\overline{u}(\frac{\partial F}{\partial x})$ ($F$ includes $u', v',$ and $T'$), in the case of positive-only heating is not as important as in the linear wave–CISK case. The importance of the vertical advection terms, $w' (\frac{\partial G}{\partial z})$ ($G$ includes the mean zonal wind ($\overline{u}$) and the mean temperature ($T_\infty$)), increases in the case of positive-only heating, while they are negligible in the linear wave–CISK case. The reason for this is probably the following. The zonal concentration of the ascending region makes the magnitude of the upward motion larger for the positive-only heating case than for the linear waves, with the same magnitude of the zonal wind perturbation ($u'$). Since the wave energy generation, which determines the wave growth, is concentrated in this narrow region in the positive-only heating case, the terms, $w' (\frac{\partial G}{\partial z})$, become more important than in the linear wave–CISK case with the same magnitude of the zonal wind perturbation. Because $(\frac{\partial T_\infty}{\partial z})$ in $w' (\frac{\partial T_\infty}{\partial z})$ represents the lapse rate of the mean temperature, the effect of the lapse rate in

\^2 Mean meridional and vertical velocities are taken to be zero, and the mean temperature is a function of altitude only.
the case of positive-only heating becomes more significant than in the case linear CISK case.

b. A comparison between the effects of the zonally asymmetric mean wind state and SST

As was discussed previously, a very significant factor in determining the tropical mean wind state is the quasi-stationary SST distribution in the Tropics, although several other factors are also important. An indirect effect of the SST on tropical wave behavior has already been discussed in the investigation of the effects of the zonally asymmetric mean wind state in the previous section. The direct effect of the SST takes place by modulating the intensity of the tropical moist convection. We will refer to this direct effect as the effect of the SST.

Figure 12a shows the zonal distribution of the ECMWF equatorial surface air temperature in January. The solid line is smoothed in the same manner as was done for Fig. 3a. The smoothed line represents the observed distribution reasonably well. The zonal distribution of the surface moisture mixing ratio (the solid line in Fig. 12b) was obtained by using the simple formula suggested by Wang (1988) according to the statistical relationship between the SST and the surface mixing ratio,

\[ q_s = (0.94 \times \text{SST} - 7.64) \times 10^{-3}, \]

where SST is in units of degrees Celsius. The intensity of the tropical convection is determined by many other factors, however. Graham and Barnett (1987) found that the relationship between the intensity of tropical convection and SST is nonlinear. They found that the intensity of the tropical convection is rather insensitive to the SST over warmer regions, where the SST is above 27.5°C. They also found that no deep convection occurs over cold regions. Applying this suggestion, we use the dotted line in Fig. 12b as another possible way for the SST to affect the tropical convection (hereafter, this is referred to as case 2 and the other as case 1). The range of the amplitude variation in case 2 is twice as large as that in case 1. We did not impose zero convection in the colder regions because no other energy sources were used in the present model.

Experiments with no mean flow and zonally asymmetric SST effects were conducted. The SST effects simply make the waves increase over the warmer regions and decrease over colder regions. Figure 13 shows the rates of amplitude change with respect to time (from the temperature perturbations at 250 mb for the fast wave case) as a function of longitude for SST effects by case 1 and case 2 forcings (dotted and dashed curves) and the mean wind state effect (solid curve). The effect of the mean state on the tropical wave increase and decrease is seen to be almost opposite to that of the SST. The SST effect makes the waves grow over the western and middle Pacific, while the effect of the mean state makes the waves grow over the eastern Pacific. The zonal variation of the rate of change of the wave amplitude resulting from the effect of the mean state is as large as that from the SST effect in case 2. Thus, the effect of the mean state turns out to sustain these tropical waves when they propagate away from the warmer SST regions. With the inclusion of both the SST and the mean state, it was found that the waves grow over a broader region with a maximum between those by the SST effect and the mean state effect (see Fig. 13).

For the slow waves (not shown), the SST effect is almost the same as for the fast waves. Since the effect of the mean state on the slow waves is similar to that of the SST, the effect of the mean state makes the larger
amplitudes of the slow waves be strongly restricted to the warmer SST regions. Other processes such as sensible heating in the lower troposphere (Sui and Lau 1989) might be important for the global propagation of the slow waves, particularly for the global propagation of their zonal wind perturbations. These mechanisms are not discussed in this study.

5. A discussion

a. Representativeness of the model results

Tropical waves are found to interact strongly with the zonally asymmetric mean state. Since waves with wavenumbers larger than the truncation limit \( M = 15 \) are excluded in the modeling, larger zonal-scale waves should be predominant both for the mean state and for the transient waves in order for our numerical integration, with the present truncation, to be considered reliable. For the mean state in our experiments, waves whose wavenumbers are larger than 4 were excluded. For the wave perturbations, the influence of waves with wavenumbers larger than 11 should be negligible for our numerical truncation to not distort our results. It has been found that planetary scale waves are dominant for all wave perturbation variables except for the vertical velocity. Thus, a problem might be caused by the inclusion of interaction terms with the form \( w' (\partial G/\partial z) \) \( G \) includes \( u, v, \) and \( T, \) due to the zonal concentration of \( w'. \) The magnitude of \( v \) is much smaller than that of \( u \) in the mean state and the effect by \( v \) was found to be negligible. Thus, \( w' (\partial T /\partial z) \) cannot be a problem. Neither can \( w' (\partial \bar{u} /\partial z) \) because \( \bar{T} \) is not dependent on \( x. \) For \( w' (\partial \bar{u} /\partial z) \), unlike the linear wave–CISK case, however, this term is important in determining the zonal variation of the tropical waves and is discussed/investigated in the following.

Crum and Dunkerton (1992) have shown that in the positive-only heating case, the simulated scale of the ascending region depends on the zonal resolution and the dissipation used in the numerical model. They also suggested several possible scale-dependent dissipations for the wave packet to have a noninfiniteesimal ascending region, but as Shen et al. (1996) pointed out, the present results can be considered to be relevant to the waves with wavenumbers higher than the truncation limit being severely dampened and large scales being scarcely affected. Analytic solutions (e.g., Crum and Dunkerton 1992; Wang and Xue 1992) also indicated that a vertical velocity perturbation with a certain zonally concentrated ascending region consists of all wavenumbers. So, it is clear that although the scale of the ascending region in the modeling is constrained by the numerical zonal resolution, this scale is different from the smallest wavelength in the modeling. Figure 14 shows the composition of the vertical velocities at 500 mb and 850 mb for the fast wave packet whose vertical structure is seen in Fig. 7a. Although the scale of the ascending region is small, the contribution from the small zonal-scale waves is less important than the large-scale waves. Experiments with the exclusion of wavenumbers larger than 11 in \( w' \) were conducted. The results are very similar to those previously obtained with the inclusion of wavenumbers larger than 11. Thus, the influence of the wavenumbers larger than 11 in \( w' \) is negligible for these results, implying no problem with the interaction term, \( w' (\partial \bar{u} /\partial z) \), in our numerical integrations.

b. The effect of the lapse rate on the previous results

The effects of the vertical wind shear on tropical waves has been shown to depend on the vertical profile of the mean temperature lapse rate. One possible reason is that due to the zonal concentration of the ascending region, the main region for wave energy generation is concentrated in this narrow region (see Figs. 7a and 7b). Because of this zonal concentration of the ascending region, the vertical advection terms, such as \( w' (\partial \bar{T} /\partial z) \) and \( w' (\partial \bar{u} /\partial z) \), become more important for energy generation as was discussed in section 4. This is probably why in the positive-only heating case, the vertical wind shear effect on fast waves (with wave maximum energy generation at the upper troposphere) is sensitive to the change of the vertical profile of the lapse rate (in this study, the largest change is in the upper troposphere).

![Fig. 14. The amplitudes of vertical velocity at the equator as a function of wavenumber. Solid: at 500 mb; dotted: at 850 mb.](image-url)
But there is no significant difference for the linear CISK–waves where the zonal advection terms, such as $\bar{u}(\partial T/\partial x)$ and $\bar{u}(\partial w/\partial x)$, are more important (Zhang and Geller 1994). The effect of the vertical wind shear on the slow waves has been found to be insensitive to the change of the vertical profile of the lapse rate. One possible reason for this may be that the differences between the lapse rates used in the current experiments are small in the middle troposphere, where the heating rate (or $\eta$) is the largest and the maximum energy generation occurs for the slow waves.

**c. Seasonality of tropical waves**

The principal focus of this study is the effect of the mean Walker circulation, which is mainly realized by the zonality of the mean vertical wind shear effect, on the zonal variability of propagating tropical waves. The seasonality of tropical waves mentioned here is purely due to the seasonality of the mean Walker circulation.

From Figs. 1 and 2, the main seasonal variations of the mean Walker circulation are due to its variation in intensity and its meridional movement. Generally, the Walker circulation is stronger during the solstitial seasons (January and July) than during the equinoctial seasons (April and October). The Walker circulations during the solstitial seasons have larger meridional scales than the Walker circulations during the equinoctial seasons. This implies that the zonal variability of the tropical waves during the solstitial seasons should be larger than during the equinoctial seasons due to the large meridional shear in the equinoctial seasons. Also, waves are not exactly symmetric about the equator due to the asymmetry of the mean state about the equator, but these aspects of the problem are not treated here. It should be pointed out that the tropical waves should also have interannual variations related to the interannual variations of the tropical SST and the above mean Walker circulation.

**6. Summary**

The influence of the mean atmospheric Walker circulation on tropical wave variability and the relative importance of the vertical mean wind shears in this influence have been investigated compared with the effect of zonally asymmetric SST. The fast and slow tropical waves in this study are uniquely determined by the supposedly different vertical profiles of the wave–CISK heating. Since the observed temperature lapse rate in the upper troposphere seems to be quite different from the idealized constant lapse rate used in previous studies (Lim et al. 1990; Zhang and Geller 1994; Shen et al. 1996), experiments have been carried out to examine the sensitivity of the effects of the vertical shear of the mean zonal flow to the temperature lapse rate. For the reader’s convenience, a summary of the principal conclusions from this paper and those by Shen et al. (1996) about the mean state effects on tropical waves is given in Table 1. The following is a detailed summary of this paper:

1) For fast waves with positive-only wave–CISK, where the maximum heating is assumed to be located in the upper troposphere, the effect of the zonally symmetric vertical wind shear significantly changes when the observed temperature lapse rate is used instead of the idealized constant lapse rate, which has been used in many previous studies. With the observed temperature lapse rate, easterly wind shear (easterly wind increasing with altitude) suppresses the eastward propagating fast waves and westerly wind shear preferentially excites them, while this is almost opposite with the idealized constant lapse rate. Since the vertical advection terms, $w'(\partial T/\partial z)$ and $w'(\partial u/\partial z)$, which are important in the positive-only wave–CISK case, are less important in the linear CISK case (where the zonal advection terms, $\bar{u}(\partial u/\partial z)$ and $\bar{u}(\partial T/\partial z)$, are most important), the effect of the vertical shear of the mean zonal flow on the fast waves is not as sensitive to this difference in the temperature lapse rate profile in the linear CISK case. For the slow waves, the vertical wind shear effect is not sensitive to the change of the lapse rate profile. This may be attributed to the fact that the most unstable layer occurs in the middle troposphere where the difference between the observed and the idealized lapse rates is small. The meridional shear of the mean zonal flow does not affect the relationship between the wave growth rate and the vertical shear of the mean zonal flow but reduces the magnitude of the effect of the vertical shear of the mean zonal flow.

2) With the existence of the mean atmospheric Walker circulation, tropical waves have large zonal variations when propagating globally. In general, the effect of the zonally asymmetric mean wind state makes the fast tropical waves decrease over the Eastern Hemisphere ($0^\circ$–$90^\circ$E=$180^\circ$) and increase over the Western Hemisphere ($180^\circ$–$90^\circ$W–$0^\circ$), while it makes the slow waves increase over the Eastern Hemisphere and decrease over the Western Hemisphere. It is found that the existence of the zonally asymmetric meridional and vertical velocity fields and the temperature field are not so important for the above effect. The basic features of the zonal variation are mainly due to the vertical shear of the mean zonal flow.

3) For an asymmetric distribution of SST, the propagating tropical waves grow in amplitude over the warmer regions and decrease over the colder regions. The amplitude of the zonal variation of the tropical waves resulting from the zonally asymmetric mean wind state is found to be comparable to that due to the zonal variation of the SSTs. Since the warmer tropical regions (Indian Ocean and western Pacific) and the colder tropical regions (eastern Pacific) generally have different vertical shear conditions for the tropical mean zonal flow, the fast waves are sustained by the effect of the mean wind state when moving out from the warm
regions. For the slow waves, since the SST and the mean wind state have in-phase effects on slow propagating waves, the slow wave maximum amplitudes are strongly anchored over the warmer SST regions.

It should be pointed out that only a simplified mean state is used in this study and the main purpose of this study is to see the effect of the large-scale mean Walker circulation on the zonal variability for tropical CISK waves. Only the seasonality of the zonal variability of tropical waves was briefly discussed with respect to the seasonality of the mean Walker circulation in the tropics. Since the seasonality of the SST distribution is important to the wave variability, this variability must be considered in combination with the resulting Walker circulation for a more realistic picture. It should also be pointed out that only wave–CISK was used in the current investigation. Other suggested mechanisms, such as the phase-lagged wave–CISK, the frictional wave–CISK, and the evaporation–wind feedback should be incorporated into this model to examine to what extent the results of the current study are valid with these other suggested mechanisms. For verification, observations need to be examined with great detail to see whether the model-simulated wave variability is observed.

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APPENDIX

Model Equations

Nondimensional linearized model equations on a sphere in \( \sigma \) coordinate:

\[
\begin{align*}
\frac{\partial z'}{\partial t} &= \frac{1}{1 - \mu^2} \left( \frac{\partial}{\partial \lambda} g' + \frac{\partial}{\partial \mu} f' \right) - \alpha z' \\
\frac{\partial \delta'}{\partial t} &= \frac{1}{1 - \mu^2} \left( \frac{\partial}{\partial \lambda} g' + \frac{\partial}{\partial \mu} f' \right) \\
&- \Delta f \left( \frac{\pi u'}{(1 - \mu^2)} + \phi' + T_0 \ln P' \right) - \alpha \delta' \\
\frac{\partial T'}{\partial t} &= -\left( \frac{\pi}{1 - \mu^2} \frac{\partial}{\partial \lambda} T' - \frac{u'}{(1 - \mu^2)} \frac{\partial}{\partial \lambda} T' \right) \\
&- \frac{\partial}{\partial \mu} \frac{\partial}{\partial \lambda} T' - \phi' \frac{\partial}{\partial \sigma} T' \frac{\partial}{\partial \sigma} T' \frac{\partial}{\partial \sigma} + \frac{\partial T'}{\partial \sigma} \\
&+ \frac{k T' \omega}{P} + \frac{k T' \omega'}{P} + Q' - \beta T' 
\end{align*}
\]
\[ \frac{\partial \ln P_1^*}{\partial \tau} = -\frac{\pi}{1 - \mu^2} \frac{\partial \ln P_0^*}{\partial \lambda} - \frac{\mu}{1 - \mu^2} \frac{\partial \ln P_2^*}{\partial \lambda} - \frac{\sigma}{\partial \mu} \frac{\partial \ln P_1^*}{\partial \mu} - \delta' \frac{\partial \ln P_2^*}{\partial \sigma} \]

and

\[ \frac{\partial \phi^*}{\partial \ln \sigma} = -T' \]  

\[ \theta^* = \frac{v'^* + \psi'^* - \sigma^* \frac{\partial \phi^*}{\partial \sigma} - \sigma \frac{\partial \phi^*}{\partial \sigma} - \frac{\partial \phi^*}{\partial \sigma} - \frac{\partial \phi^*}{\partial \sigma}}{\partial \mu} \]

\[ T' = (1 - \mu^2) \frac{\partial \ln P_2^*}{\partial \lambda} - \frac{\partial \ln P_1^*}{\partial \lambda} - T' \]

\[ \xi = 2\mu + \frac{1}{1 - \mu^2} \frac{\partial \phi^*}{\partial \lambda} - \frac{\partial \phi^*}{\partial \mu} \]

\[ \delta = \frac{1}{1 - \mu^2} \frac{\partial \phi^*}{\partial \lambda} + \frac{\partial \phi^*}{\partial \mu} \]

\[ \xi' = \frac{1}{1 - \mu^2} \frac{\partial \phi^*}{\partial \lambda} - \frac{\partial \phi^*}{\partial \mu} \]

\[ \delta' = \frac{1}{1 - \mu^2} \frac{\partial \phi^*}{\partial \lambda} + \frac{\partial \phi^*}{\partial \mu} \]

where

\[ F(\lambda, \mu, \sigma, \tau) = \bar{F}(\lambda, \mu, \sigma) + F'(\lambda, \mu, \sigma, \tau) \]

and

\[ \bar{F}(\lambda, \mu, \sigma) = \frac{1}{\Delta \tau} \int_0^{\Delta \tau} F(\lambda, \mu, \sigma, \tau) d\tau \]

and \( \mu = \sin(\phi) \), where \( \phi \) is the latitude, \( \lambda \) is the longitude, \( \Delta \tau \) is the time-averaging period, and \( \sigma = \rho p \), is the vertical coordinate where \( \rho \) is the surface pressure. Here \( \bar{F} \) can be \( \bar{F}(\lambda, \mu, \sigma) \) or \( \bar{F}(\sigma) \), depending on the problem. For \( T \) (temperature), we apply \( T = \bar{T} + T' + T' \), where \( \bar{T} \) is a function of height \( (\sigma) \) only. Here, \( \alpha \) and \( \beta \) are the coefficients for Rayleigh friction and Newtonian cooling, respectively. All the other variables in Eqs. (A1)–(A5) have their convective meanings [see Shen et al. (1996) for details]. The vertical velocity \( (\sigma) \) is calculated directly with the continuity equation.

Here, \( Q' \) in the temperature equation denotes the diabatic heating. Wave–CISK scheme is used to calculate this diabatic heating. Thus, only the wave-induced lower-level convergence is used to determine the position and intensity of the diabatic heating (actually, it is caused by organized small-scale moist convective activities). Since these small convections cannot be resolved with the excited waves in a linearized model, the vertical distribution of the heating is then prescribed in the model and determines the wave phase speed in this study. Here, \( \sigma = 0 \) is applied at both upper and bottom boundaries. Semi-implicit time scheme (see Shen et al. 1996) is applied for model integration.

REFERENCES


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