

Equatorial Superrotation and the Factors Controlling the Zonal-Mean Zonal Winds in the Tropical Upper Troposphere

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ABSTRACT

The response of the zonal-mean zonal winds in the tropical upper troposphere to thermal forcing in the Tropics is studied using an idealized general circulation model with 18 vertical levels and simplified atmospheric physics. The model produces a conventional general circulation, with deep easterly flow over the equator, when integrated using zonally invariant and hemispherically symmetric boundary conditions, but persistent equatorial superrotation (westerly zonal-mean flow over the equator) is obtained when steady longitudinal variations in diabatic heating are imposed at low latitudes. The superrotation is driven by horizontal eddy momentum fluxes associated with the stationary planetary wave response to the applied tropical heating. The strength of the equatorial westerlies is ultimately limited by vertical steady eddy momentum fluxes, which are downward in the tropical upper troposphere, and by the zonally averaged circulation in the meridional plane, which erodes the mean westerly shear via vertical advection.

The transition to superrotation can be prevented by specifying offsetting zonally invariant heating and cooling anomalies on either side of the equator to create a “solstitial” basic state with a single dominant Hadley cell straddling the equator. Superrotation is restricted in the solstitial climate because the strength of the mean meridional overturning is enhanced, which increases the efficiency of vertical advection, and because the cross-equatorial flow in the upper troposphere provides an easterly zonal acceleration that offsets some of the momentum flux convergence associated with tropical eddy heating. The cross-equatorial flow aloft also reduces the stationary planetary wave response in the summer hemisphere. These results suggest that hemispheric asymmetry in the mean meridional circulation is responsible for maintaining the observed mean easterly flow in the tropical upper troposphere against the westerly torques associated with tropical wave sources.

1. Introduction

One of the more exotic behaviors exhibited by simplified atmospheric general circulation models (GCMs) is the transition from a conventional general circulation, with westerly midlatitude jet streams and easterly flow in the Tropics, to a mean climate exhibiting equatorial superrotation, or westerly zonal-mean winds over the equator. Suarez and Duffy (1992), Saravanan (1993), and Woolnough (1997) all obtained radical superrotating general circulations, with intense westerly winds in the tropical upper troposphere and major changes in global climate, using idealized two-level GCMs forced with large zonal-wavenumber-2 tropical eddy heating anomalies. Equatorial superrotation has also been reported in multilevel GCMs with idealized physics (I. M. Held, Bernhard Haurwitz Memorial Lecture, 1999 American Meteorological Society Annual Meeting, available online at <http://www.gfdl.gov/~ih/>,

hereafter referred to as HE99) and in “aquaplanet” GCMs with zonal-wavenumber-1 sea surface temperature variations along the equator (Battisti and Ovens 1995; Neale 1999; Hoskins et al. 1999; Inatsu et al. 2002). Unlike the superrotating regimes in two-level GCMs, which appear abruptly when the eddy heating exceeds a certain threshold amplitude and often persist for several hundred days after the thermal forcing is removed, the circulation changes in superrotating multilevel GCM experiments are largely confined to the tropical upper troposphere and show no evidence of hysteresis.

Hide’s theorem (Hide 1969; Held and Hou 1980) demands that superrotating flows be supported by eddy angular momentum fluxes that are directed up the local angular momentum gradient. For example, the westerly phase of the quasi-biennial oscillation in the stratosphere is supported by upward fluxes of westerly momentum in vertically propagating Kelvin waves. Neale (1999) and Hoskins et al. (1999) determined that the upper-tropospheric superrotation in their aquaplanet GCM was supported by horizontal steady eddy momentum fluxes associated with the stationary wave re-

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sponse to the fixed thermal contrasts imposed along the equator. Hoskins et al. also demonstrated that steady tropical eddy heating leads to westerly mean-flow anomalies over the equator in a multilevel primitive equation model integrated using an initial value technique. In contrast, transient eddies provide most of the eddy momentum flux required to maintain the superrotation in two-level GCMs (Saravanan 1993), which explains why the unusual superrotating regimes in these models are able to persist after the tropical eddy heating is removed.

Saravanan (1993) hypothesized that the persistence of conventional general circulations in two-level models forced with weak tropical eddy heating could be attributed to a negative feedback between tropical zonal wind speed and baroclinic wave dissipation on the subtropical flanks of the midlatitude jet streams, and that this feedback could be overwhelmed by a sufficiently strong westerly acceleration at low latitudes, leading to a novel climate with an alternative momentum balance. However, Panetta et al. (1987) and HE99 have argued that the interaction of baroclinic waves with the mean flow cannot be accurately simulated with only two vertical levels, and Neale (1999) and Hoskins et al. (1999) determined that the transient eddy momentum fluxes in the superrotating climatology of their aquaplanet model were similar to those in the conventional general circulation. These results suggest that the exotic superrotating climates in two-level GCMs are simply artifacts of the severely truncated vertical resolution and that transient eddies do not actually provide a strong negative feedback on the mean zonal-flow at low latitudes. However, it remains unclear how the westerly accelerations associated with tropical wave sources are ultimately balanced, and why multilevel GCMs superrotate when forced with boundary conditions that are intended to roughly approximate annual-mean or equinoctial conditions on earth.

An understanding of the deep easterly zonal-mean flow observed over the equator has only recently begun to emerge. Lindzen and Hou (1988) noted that a single Hadley cell straddling the equator dominates the mean meridional circulation throughout much of the year, and demonstrated that an axisymmetric model with maximum rising motion located off the equator produces easterly flow in the tropical upper troposphere due to the easterly Coriolis torque experienced by parcels of air moving toward the equator. Hou (1993) and Hou and Molod (1995) subsequently determined that hemispherically asymmetric, zonally invariant heating anomalies also lead to cross-equatorial meridional flows and easterly zonal wind anomalies over the equator in three-dimensional GCMs. However, cumulus friction (Schneider and Lindzen 1977; Rosenlof et al. 1986), gravity wave drag (Lindzen 1981; Huang et al. 1999), extratropical waves (Schneider and Watterson 1984; Held and Phillips 1990; Saravanan 1993), and the strength of the Hadley circulation (Woolnough 1997;

HE99) have also been cited as potentially important factors in the tropical momentum balance.

Lee (1999) performed a cross-spectral analysis of the 200-hPa zonal and meridional winds in the National Centers for Environmental Prediction (NCEP) reanalysis and found that the zonally symmetric component of the transient momentum flux provides an easterly acceleration at low latitudes, while transient eddies, steady eddies, and the annually averaged zonal-mean flow are all associated with a convergence of westerly momentum at the equator. Since the seasonal cycle of the Hadley circulation is the main source of temporal variability in the zonally averaged tropical winds (Dima and Wallace 2003), Lee's results suggest that hemispheric asymmetry in the mean meridional circulation is responsible for maintaining the observed mean easterly flow in the tropical upper troposphere against the westerly torques associated with tropical wave sources. However, Lee cautions that the sparsity of data at low latitudes makes it difficult to diagnose the exact balance of zonal accelerations in the tropical upper troposphere, and it remains unclear how the large seasonal variations in the tropical mean zonal wind field are produced and maintained.

Lee (1999) also demonstrated that increasing the global mean sea surface temperature in a perpetual-equinox, aquaplanet GCM leads to enhanced intraseasonal eddy activity in the Tropics and westerly mean flow over the equator. GCM experiments performed using realistic boundary conditions, a full seasonal cycle, and doubled CO₂ (Boer et al. 2000; Huang et al. 2001) have also shown a moderate increase in tropical zonal wind speed and total atmospheric angular momentum. However, the strong equatorial superrotation obtained using GCMs with simplified boundary conditions suggests that the potential response of the zonally averaged flow could be much larger if the tropical wave forcing were to increase significantly in response to either anthropogenic or natural climate change (Pierrehumbert 2000). Thus, a better understanding of the sensitivity, stability, and maintenance of the zonal-mean zonal wind distribution at low latitudes would improve both reconstructions of past climates and predictions of future climate changes.

In this study, a general circulation model with 18 vertical levels, a flat lower boundary, and idealized atmospheric physics is used to study the zonally averaged response to tropical eddy heating and other types of low-latitude forcing, with an emphasis on the mean zonal winds in the tropical upper troposphere. The results of these experiments demonstrate that cross-equatorial flow is essential for preventing superrotation when wave forcing is present at low latitudes and also reveal a surprisingly important role for vertical momentum transport in the Tropics. The paper is organized as follows: the model and its control climate are described in section 2. Section 3 describes the response of the model to tropical eddy heating, zonally invariant accel-

erations, and hemispheric asymmetry, with an emphasis on the processes that control the mean zonal wind field. Section 4 contains a summary of results, a discussion of the zonal-mean zonal wind balance over the equator, and our conclusions regarding the occurrence of equatorial superrotation in models.

2. Model and control climate

a. Model description

The general circulation model used in this study consists of the dynamical core from the National Center for Atmospheric Research (NCAR) Community Climate Model, version 3.6, integrated using the simplified boundary conditions and idealized parameterizations for atmospheric physics suggested by Held and Suarez (1994, hereafter HS94). The model has T42 horizontal spectral resolution, 18 vertical (sigma) levels, and a flat (constant geopotential) lower boundary. Frictional effects near the surface are represented by a horizontally uniform linear drag (Rayleigh friction). Radiative transfer and other physical processes that affect the atmospheric temperature distribution are approximated by linear relaxation (Newtonian cooling) toward a specified zonally invariant and hemispherically symmetric reference temperature distribution. The numerical scheme also includes weak biharmonic horizontal diffusion at all levels to damp small-scale noise and harmonic diffusion in the top three levels to absorb vertically propagating wave energy. Further details of the dynamical core and the idealized physical parameterizations can be found in Kiehl et al. (1998) and HS94, respectively.

Unless stated otherwise, all of the model runs described below were integrated for a total of 730 days (2 yr) starting from a resting basic state with small thermal perturbations added to break symmetry. The statistics from the final 550 days (~18 months) of each experiment are used to characterize the mean climate associated with each set of boundary conditions and applied forcings. Longer integrations indicate that this averaging period adequately represents the mean climate. A limited number of integrations performed using different model specifications also suggest that the results presented below are relatively insensitive to the model resolution, damping parameters, and the reference temperature profile (see Kraucunas 2001).

b. Control run

Figure 1 shows the time-averaged, zonal-mean zonal winds, mean meridional circulation, and horizontal transient kinetic energy from a control run integrated using only the zonally invariant and hemispherically symmetric HS94 damping parameters and boundary conditions. The mean zonal wind field (Fig. 1a) in this control, or "C," climate exhibits well-defined westerly jet streams and surface westerlies in the midlatitudes, easterly trade winds in the subtropical boundary layer,

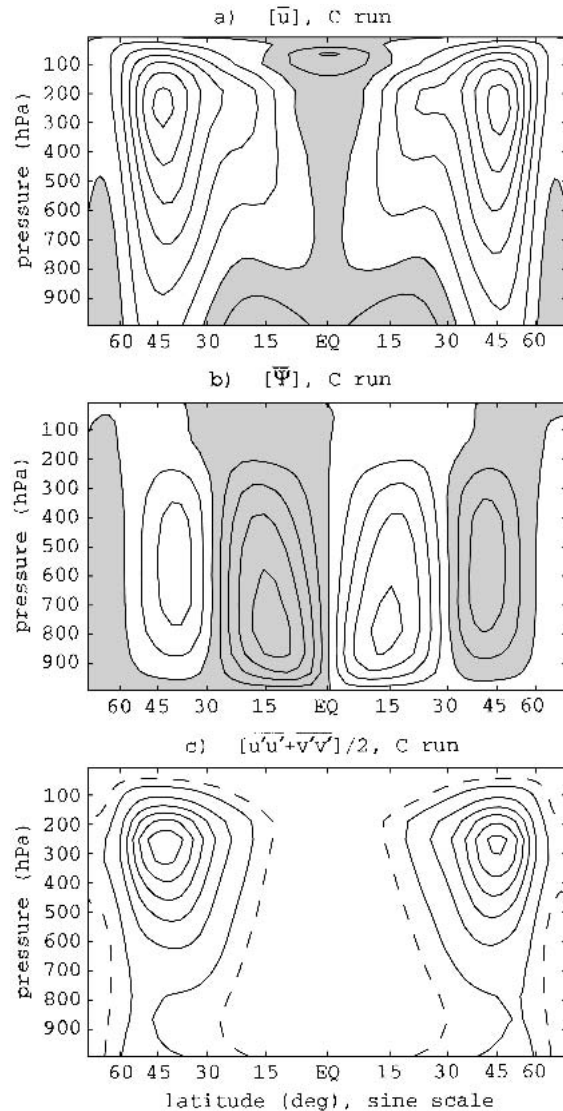


FIG. 1. The mean (a) zonal winds, (b) meridional mass circulation, and (c) transient kinetic energy in the C run. The contour intervals are (a) 5 m s^{-1} , (b) $2 \times 10^{10} \text{ kg s}^{-1}$, and (c) $50 \text{ m}^2 \text{ s}^{-2}$, with a dashed contour added at $25 \text{ m}^2 \text{ s}^{-2}$ in (c). Negative (easterly/counterclockwise) values are shaded. Note that the horizontal axis is sine of latitude and the vertical axis is linear in pressure, so equal areas on the plot represent equal amounts of atmospheric mass.

and weak easterly flow over the equator and near the poles. These features are similar to the zonal-mean flow observed in aquaplanet models forced with perpetual-equinox boundary conditions (Hess et al. 1993; Battisti and Ovens 1995; Hoskins et al. 1999), and also to the annually averaged, zonal-mean zonal wind distribution on earth. One unusual aspect of the mean zonal wind distribution in the C run is the weak westerly flow in the subtropical midtroposphere, which extends almost to the equator near the top of the boundary layer. There is also a weak easterly anomaly over the equator near

100 hPa, which has been studied by Williamson et al. (1998) and Zadra et al. (2002).

The mean meridional circulation in the C climate (Fig. 1b) resembles the observed annually averaged circulation in the meridional plane (see, e.g., Dima and Wallace 2003). However, the Hadley cells in the idealized model are somewhat weaker, and centered lower in the troposphere, than the annually averaged Hadley cells on earth, while the midlatitude Ferrell cells are slightly stronger. Previous studies (Lindzen and Hou 1988; Hou 1993) have demonstrated that the strength of the Hadley circulation increases when the maximum heating is either moved off the equator or latitudinally confined, so the weak Hadley cells in the C run are consistent with the hemispheric symmetry of the HS94 boundary conditions and the diffuse tropical heating associated with Newtonian relaxation (the influence of a stronger Hadley circulation is evaluated in section 3). The strong midlatitude Ferrell cells in the C climate, on the other hand, suggest that extratropical baroclinic waves are particularly active in the idealized model. The relatively large horizontal transient kinetic energy maxima (Fig. 1c) and the location of the jet streams deep in the midlatitudes (Fig. 1a) are also indicative of robust extratropical transient eddy activity.

c. Mean zonal wind balance

The time-averaged, zonal-mean zonal wind tendency equation can be used to diagnose how different processes affect the zonal-mean zonal wind distribution. Using overbars (primes) and brackets (asterisks) to denote temporal and longitudinal averages (deviations), respectively, the time rate of change of the zonal-mean zonal wind may be written (in advective form):

$$\begin{aligned} \frac{\partial[\bar{u}]}{\partial t} = & 2\Omega \sin\phi[\bar{v}] - \frac{[\bar{v}]}{\cos\phi} \frac{\partial}{\partial\phi} ([\bar{u}] \cos\phi) - [\bar{\omega}] \frac{\partial[\bar{u}]}{\partial p} \\ & + [\bar{F}_x] - \frac{1}{a \cos^2\phi} \frac{\partial}{\partial\phi} ([\bar{u}^* \bar{v}^*] \cos^2\phi) - \frac{\partial}{\partial p} [\bar{u}^* \bar{\omega}^*] \\ & - \frac{1}{a \cos^2\phi} \frac{\partial}{\partial\phi} ([\bar{u}' \bar{v}'] \cos^2\phi) - \frac{\partial}{\partial p} [\bar{u}' \bar{\omega}']. \quad (1) \end{aligned}$$

Here the notation is standard, hydrostatic balance is assumed, and we have neglected the small zonal accelerations associated with the vertical Coriolis term ($2\Omega \cos\phi[\bar{\omega}]/\rho g$), the metric term $[\bar{u}\bar{\omega}]/apg$, and the horizontal diffusion of momentum.

The first three terms on the right-hand side of (1) represent the zonal accelerations associated with the mean meridional circulation acting on the mean zonal wind distribution. The first two terms are often combined to yield $[\bar{v}](2\Omega \sin\phi - \partial[\bar{u}]/\partial y)$, the acceleration associated with the advection of zonal momentum by the mean meridional wind (in Cartesian form). Here, $[\bar{F}_x]$ is the mean frictional drag. The final four terms represent the zonal accelerations associated with the

horizontal and vertical fluxes of zonal angular momentum by steady eddy circulations and transients; these terms are positive (i.e., induce a westerly acceleration) when the momentum fluxes are convergent. Lee (1999) explicitly separates the transient terms into zonally asymmetric and zonally symmetric components (e.g., $[\bar{u}' \bar{v}'] = [\bar{u}^{*'} \bar{v}^{*'}] + [\bar{u}'] [\bar{v}']$), but the time-averaged, zonally symmetric transient momentum fluxes are small in all of our model integrations due to the absence of temporal variability in the boundary conditions.

The zonal-mean zonal wind tendency was negligible over the final 550 days of all of our model integrations, so the terms on the right-hand side of (1) were calculated at each resolved latitude and pressure level in order to evaluate the relative roles played by zonally symmetric, transient, and steady eddy processes in maintaining the time-averaged, zonal-mean zonal wind distribution associated with each set of boundary conditions. Derivatives were calculated using centered finite differences. The mean zonal wind tendency, the terms neglected in (1), and any computational errors were collected in a residual term that is combined with the frictional term (which is zero outside of the boundary layer) for display purposes. The vertical and horizontal transient momentum flux convergences are also combined in all of the figures presented below.

Figure 2 displays the mean zonal wind balance components for the C run. The steady eddy terms were found to be small, which is consistent with the zonally invariant boundary conditions in the control integration, and are not shown. In the boundary layer, the zonal accelerations associated with the mean meridional flow (Fig. 2a) are balanced by frictional drag (Fig. 2b). The dominant balance in the upper troposphere is between the mean meridional term (which is dominated by the mean Coriolis acceleration) and the transient eddy momentum flux convergence (Fig. 2c). These relationships are similar to those calculated for annually averaged conditions on earth (Peixoto and Oort 1992), although the midlatitude transient eddy forcing is slightly stronger in the C run, which is consistent with the robust midlatitude baroclinic wave activity in the idealized model.

The zonal accelerations near the equator are very small in the C run. Transient eddies provide a weak ($\sim 0.1 \text{ m s}^{-1} \text{ day}^{-1}$) easterly acceleration in the tropical upper troposphere due to the dissipation of equatorward-propagating extratropical transient eddies that occasionally reach low latitudes. The anomalous westerly flow near the top of the boundary layer in Fig. 1a appears to be associated with the diffuse poleward outflow in the upper branches of the Hadley cells (Fig. 1b), which induces weak westerly accelerations through the depth of the free troposphere (Fig. 2a). The resulting easterly vertical shear over the equator, combined with the upward mean vertical flow, yields a weak westerly acceleration in the tropical upper troposphere due to mean vertical advection (Fig. 2d), which balances the

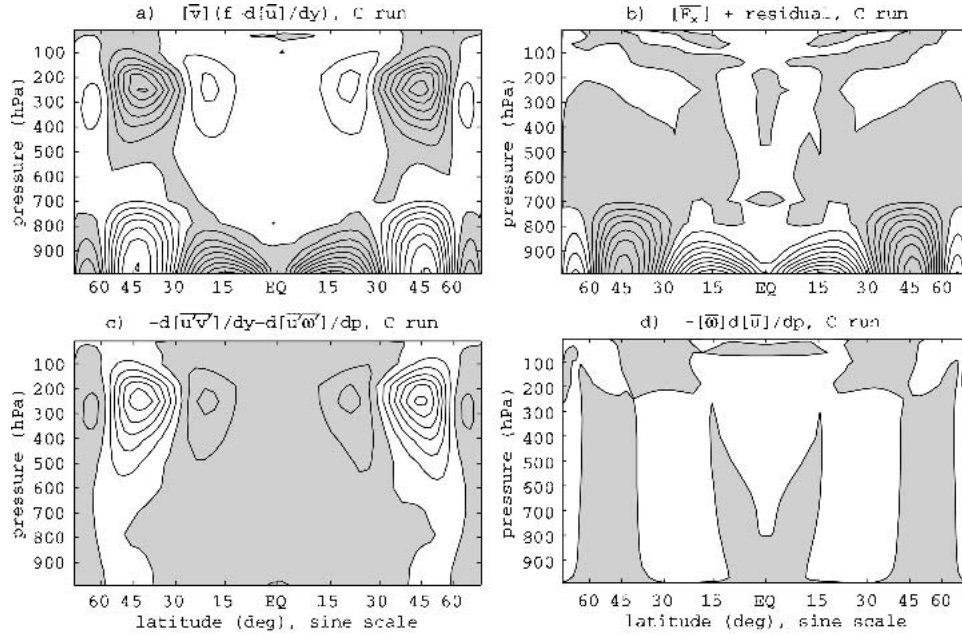


FIG. 2. The mean zonal wind balance components for the C run [see text regarding Eq. (1) for explanation]. The contour interval is $1 \text{ m s}^{-1} \text{ day}^{-1}$ and easterly zonal accelerations are shaded.

weak easterly acceleration associated with the dissipation of transients. As shown below, the mean vertical advection term becomes stronger if there is an increase in either the mean vertical shear or the mean vertical velocity over the equator.

3. Results

a. Response to tropical eddy heating

To study the atmospheric response to large-scale thermal contrasts at low latitudes, a steady, zonally asymmetric heating perturbation with the following structure was added to the thermodynamic tendency equation of the model at all time steps:

$$Q^*(\lambda, \phi, \sigma) = Q_0 \sin(n\lambda) \exp\left[-\left(\frac{\phi - \phi_0}{\Delta\phi}\right)^2\right] \cdot \sin\left(\frac{\sigma - \sigma_t}{\sigma_b}\right), \sigma_t < \sigma < \sigma_b.$$

This paper focuses on the response to a Q^* with $Q_0 = 1 \text{ K day}^{-1}$, $n = 2$, $\phi_0 = 0^\circ$, $\Delta\phi = 15^\circ$, $\sigma_t = 0.2$, and $\sigma_b = 0.8$ (Figs. 3a,b), which has a three-dimensional structure designed to roughly approximate the zonally asymmetric component of the annually averaged tropical latent heating distribution on earth (Figs. 3c,d; cf. Schumacher et al. 2004). The horizontal distribution of Q^* is also identical to the eddy heating used by Suarez and Duffy (1992) and Saravanan (1993), and similar to the mean eddy heating anomalies produced by aquaplanet

models forced with zonal SST variations (Neale 1999; D. S. Battisti 2000, personal communication).

Figure 4 shows aspects of the mean climate in the C^* run, which was integrated from rest with Q^* imposed, and Fig. 5 shows the mean zonal wind and temperature differences between C^* and C. The most striking feature of the C^* mean zonal wind field (Fig. 4a) is the uninterrupted band of westerly flow stretching across the entire tropical upper troposphere, including a superrotation of nearly 20 m s^{-1} over the equator. Outside of this region, the C^* mean zonal wind distribution is similar to the control climate, although the midlatitude jet streams are slightly weaker ($\sim 10\%$). The mean zonal wind difference between C^* and C (Fig. 5a) highlights the concentration of the response in the tropical upper troposphere, and also reveals a slight poleward migration of the midlatitude jets and a slight weakening of the subtropical trade winds. The mean meridional circulation and transient kinetic energy in C^* (Figs. 4b,c) also bear a strong resemblance to their counterparts in the C run (Figs. 1a,b), although the tropical Hadley cells and the midlatitude transient kinetic energy maxima are both slightly ($\sim 15\%$) weaker in the C^* climate and there are sharp increases in both transient kinetic energy and mean temperature (Fig. 5b) coincident with the large zonal wind increase in the tropical upper troposphere.

Additional experiments, described in Kraucunas (2001), indicate that the mean climate produced by the idealized model is independent of the initial state used, that is all runs integrated with Q^* produce the same mean climate as that shown in Fig. 4, and all runs inte-

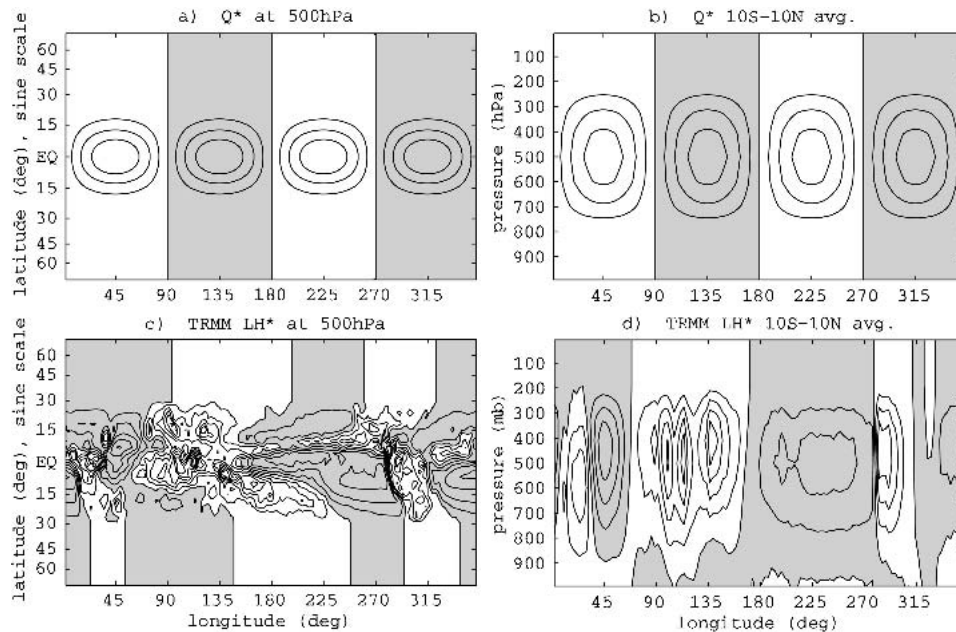


FIG. 3. The imposed tropical eddy heating perturbation, Q^* , (a) at 500 hPa and (b) averaged between 10°N and 10°S , and the annual-mean eddy latent heating distribution derived from TRMM precipitation measurements by Schumacher et al. (2004), (c) at 500 hPa, and (d) averaged between 10°N and 10°S . The contour interval is 0.25 K day^{-1} , and negative values (i.e., latent heating rates that are less than the zonal average at each latitude) are shaded.

grated with only the zonally symmetric HS94 forcing produce a mean climate similar to the C run (even when the final time step of the C* run or a stronger superrotating flow is used as an initial state). Thus, our model does not exhibit any evidence of multiple steady states under fixed boundary conditions. It was also determined that the mean zonal wind response over the equator is not very sensitive to the horizontal or vertical profile of the eddy heating, and that the superrotation strength is roughly linear with respect to the eddy heating amplitude. The weakening of the Hadley cells and the increases in transient kinetic energy and temperature in the tropical upper troposphere also appear to be consistent responses to tropical eddy heating across a broad range of eddy heating parameters.

Although the focus of this paper is on the zonally averaged response to eddy heating at low latitudes, it is useful to briefly examine the three-dimensional structure of the steady eddy response forced by Q^* , which is depicted in Fig. 6, because the momentum fluxes associated with these features play an important role in the mean zonal wind balance. The time-averaged eddy horizontal winds and geopotential height variations in the upper troposphere (Fig. 6a) indicate that regions with eddy heating exhibit warmer tropospheric temperatures, easterly zonal wind anomalies near the equator, and anticyclonic flow at low latitudes in both hemispheres; while areas with eddy cooling exhibit the opposite deviations from the zonal mean. This pattern is similar to the steady response to tropical eddy forcing

obtained by Matsuno (1966) and Gill (1980) using linear shallow-water models with resting basic states, except that the strongest circulation features in the C* climate are collocated in longitude with the maximum heating and cooling, rather than one-quarter wavelength to the west, which indicates that the rotational (i.e., Rossby wave) response is much stronger than the divergent (i.e., Kelvin wave) response in the superrotating case. An analysis of the vorticity budget in the C* run, along with experiments performed using a linearized version of the model, reveal that the intensification and eastward shift in the planetary wave response at upper levels can be attributed to both the strong westerly zonal-mean flow aloft (Kraucunas 2001; see also Philps and Gill 1987) and nonlinearity (Hendon 1986; Sardeshmukh and Hoskins 1988).

The horizontal propagation of planetary wave activity through a region in which vorticity increases with latitude is associated with a net flux of westerly zonal momentum in the opposite direction (see, e.g., Eliassen and Palm 1961). A careful examination of the eddy zonal wind vectors in Fig. 6a reveals that the steady eddies aloft in the C* run are associated with equatorward zonally averaged fluxes of westerly momentum in the upper troposphere (Fig. 6b), which suggests that the upper-level stationary planetary wave response to Q^* is responsible for driving the robust equatorial superrotation in the idealized model. A plot of the total (eddy plus zonal mean) time-averaged zonal winds and vertical velocities over the equator in C* (Fig. 6c) indicates

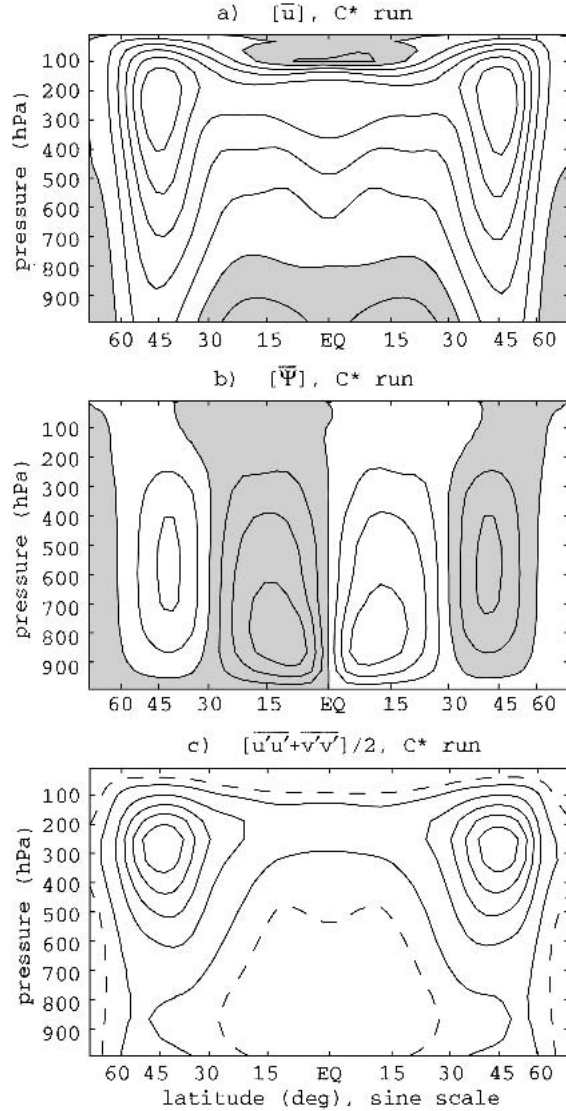


FIG. 4. The mean (a) zonal winds, (b) meridional mass circulation, and (c) transient kinetic energy in the C^* run, which was forced with the tropical eddy heating Q^* . The contour intervals are (a) 5 m s^{-1} , (b) $2 \times 10^{10} \text{ kg s}^{-1}$, and (c) $50 \text{ m}^2 \text{ s}^{-2}$, with a dashed contour added at $25 \text{ m}^2 \text{ s}^{-2}$ in (c). Negative (easterly/counterclockwise) values are shaded.

that the strongest westerly winds are centered above the regions with eddy cooling, where weak adiabatic descent dominates the vertical motion field (note that ω is positive for motion toward the surface), while regions with eddy heating are characterized by strong upward velocities and weak vertical shears. Thus, the steady eddy circulations in C^* are also associated with vertical fluxes of zonal momentum (Fig. 6d).

Figure 7 shows the difference between the mean zonal wind balance components (1) in C^* and those from the control integration, including the horizontal and vertical steady eddy momentum flux convergences (which were very weak in the C run). The largest zonal

accelerations in Fig. 7 are associated with changes in the advection of angular momentum by the mean meridional flow (Fig. 7a), boundary layer friction (Fig. 7b), and the transient momentum flux convergence (Fig. 7c). However, these accelerations are associated with the small mean zonal wind changes in the extratropics, rather than processes near the core of the equatorial superrotation. For example, the mean meridional term (Fig. 7a) exhibits easterly acceleration anomalies in the subtropical boundary layer, indicating that the slight reduction in trade wind strength noted in Fig. 5a may be attributed to the decline in the strength of the Hadley circulation in the C^* run.

The horizontal steady eddy momentum flux convergence (Fig. 7e) is the only component of the C^* mean zonal wind balance that provides a westerly acceleration in the region where the maximum zonal wind changes are observed in Fig. 5a. Hence, the equatorial superrotation in the idealized model is driven by the eddy momentum fluxes depicted in Fig. 6b. Interestingly, the vertically oriented steady eddy momentum fluxes are associated with an easterly acceleration in the upper troposphere (Fig. 7f) that closely mirrors the westerly acceleration associated with the horizontal eddy response. The vertical steady eddy momentum fluxes also induce a westerly acceleration in the midtro-

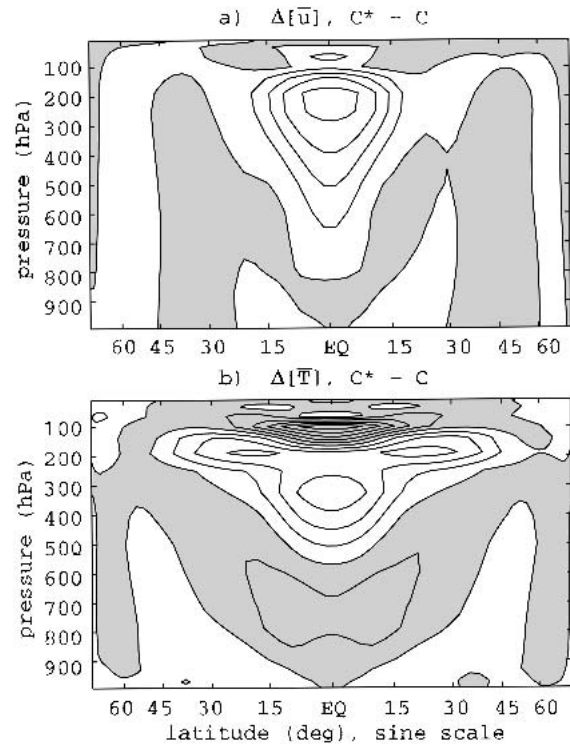


FIG. 5. The mean (a) zonal wind and (b) temperature differences between the C^* and C runs, which show the response of the control climate to the tropical eddy heating Q^* . The contour intervals are (a) 5 m s^{-1} and (b) 1 K . Negative (easterly/colder) values are shaded.

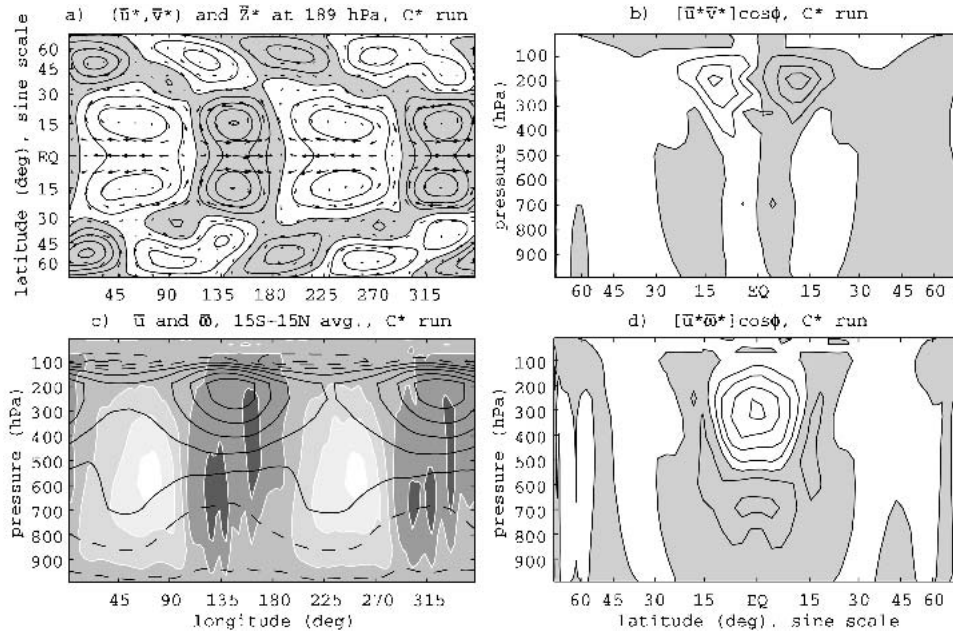


FIG. 6. Aspects of the steady eddy response in the C* run: (a) eddy geopotential height variations and horizontal eddy winds in the upper troposphere, where the contour interval is 20 m with negative height anomalies shaded, and the longest vector represents a wind speed of 12 m s^{-1} ; (b) the zonally averaged meridional flux of zonal angular momentum by steady eddies, where the contour interval is $2 \text{ m}^2 \text{ s}^{-2}$ and southward fluxes are shaded; (c) the total (eddy plus zonal mean) mean zonal winds and mean vertical pressure velocities averaged between 15°S and 15°N , where the black contour lines depict the mean zonal wind field with a contour interval of 5 m s^{-1} and easterly values are dashed, and the shaded contours indicate mean vertical pressure velocity with a contour interval of 0.01 Pa s^{-1} and the lightest (darkest) shading corresponding to upward (downward) velocities of 0.03 (0.01) Pa s^{-1} ; and (d) the zonally averaged vertical flux of zonal angular momentum by steady eddies, where the contour interval is $0.025 \text{ m Pa s}^{-2}$ and upward fluxes are shaded.

posphere, which indicates that the correlation between the vertical and zonal velocity perturbations noted in Fig. 6c is responsible for spreading the superrotation downward into the midtroposphere. The divergence of the equatorward steady eddy momentum fluxes in Fig. 7e induces an easterly acceleration in the subtropics. Thus, the net effect of the steady eddy circulations in the C* climate is to extract westerly momentum from the subtropics, especially at upper levels, and deposit it in the midtroposphere over the equator.

The westerly acceleration associated with the net steady eddy momentum flux convergence in the tropical midtroposphere is ultimately balanced by mean vertical advection (Fig. 7d), which provides an easterly acceleration over the equator in the C* climate because the upward flow in the rising branch of the Hadley circulation is collocated with a strong westerly vertical shear. The temporal evolution of the zonally averaged circulation in C* (not shown) indicates that the mean zonal winds over the equator accelerate rapidly at first, then more slowly as the vertical shear increases, until finally the mean vertical advection term offsets the steady eddy momentum flux convergence over the equator. Experiments with weaker and stronger eddy heating anomalies indicate that the horizontal steady

eddy momentum flux convergence, the mean vertical advection term, and the maximum westerly flow over the equator all increase roughly linearly with respect to the amplitude of the heating. These results demonstrate that vertical advection provides an important restoring force for controlling mean wind perturbations over the equator under hemispherically symmetric boundary conditions, although it should be noted that the mean rising motion cannot provide an easterly acceleration over the equator until a westerly mean shear has developed.

The mean zonal wind changes and steady eddy circulations in the C* climate are similar in structure to the mean responses obtained by Battisti and Ovens (1995), Neale (1999), and Hoskins et al. (1999) using aquaplanet GCMs with zonal-wavenumber-1 tropical SST anomalies imposed under otherwise zonally invariant and hemispherically symmetric boundary conditions. Hence, our model captures the fundamental dynamical response to thermal contrasts at low latitudes obtained using models with more sophisticated physical parameterizations. The eddy heating, steady eddy circulation features, and steady eddy momentum fluxes in our model are actually somewhat weaker than the mean diabatic heating anomalies and eddy statistics re-

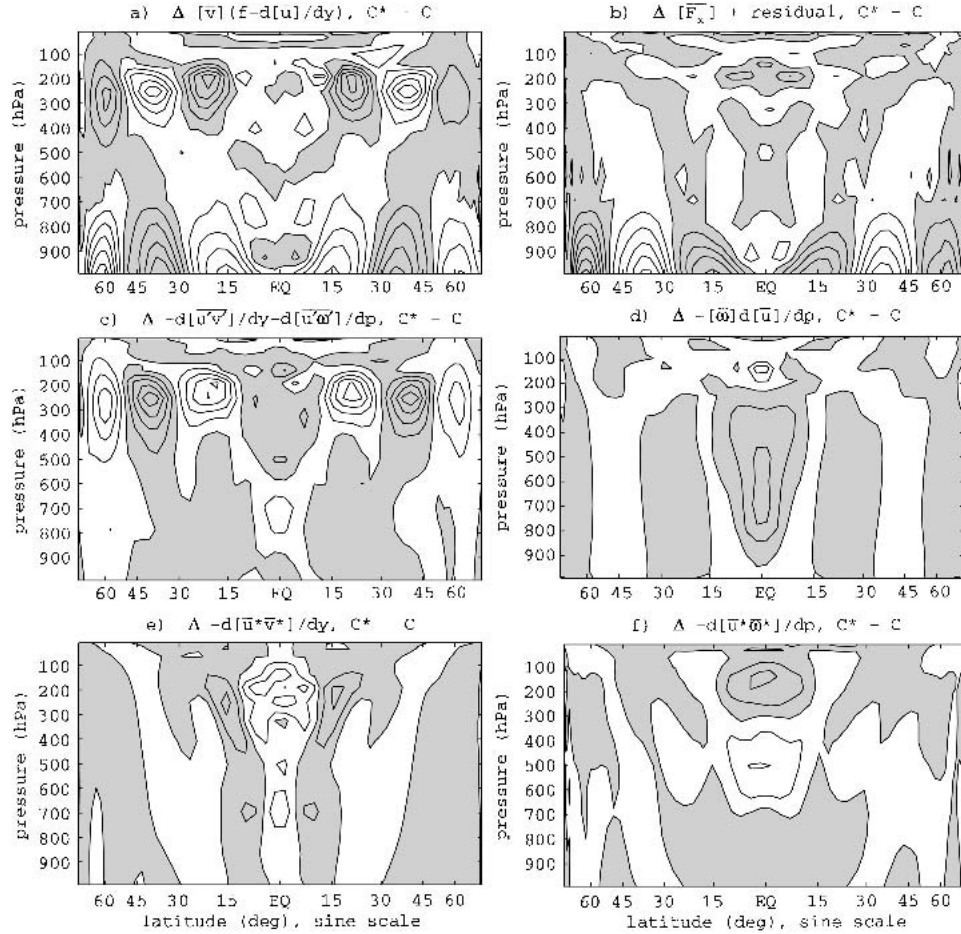


FIG. 7. The change in the mean zonal wind balance components [see Eq. (1)] between the C and C* runs. The contour interval is $0.25 \text{ m s}^{-1} \text{ day}^{-1}$ and easterly acceleration anomalies are shaded.

ported by Neale (1999). It will be demonstrated later that the relatively strong mean zonal wind response in our model (relative to the amplitude of the eddy heating and stationary wave response) may be attributed to the weaker Hadley circulation, and hence weaker mean vertical advection, produced by the HS94 boundary conditions. However, first we will confirm the relative roles played by horizontal and vertical eddy momentum fluxes in maintaining the superrotating mean flow over the equator.

b. Response to mean zonal accelerations

In this section we describe the results of model runs forced with zonally invariant zonal accelerations derived from the mean zonal wind balance components in the C* integration. These experiments will help verify and extend our conclusions regarding superrotation and the stability of the tropical zonal-mean flow, as well as allow us to evaluate the role of steady eddies in producing the mean temperature, mean meridional circulation, and transient kinetic energy changes in the C*

climate. In some of these experiments the model was forced by adding one or more of the components illustrated in Fig. 7 directly to the zonal wind tendency, while in other runs these accelerations were first multiplied by -1 to study the influence of easterly zonal torques in the Tropics. The results of these experiments, and also those described in the previous and following sections, are summarized in Table 1, which lists the imposed thermal and/or mechanical forcing for each run, along with the maximum mean zonal wind response (relative to the appropriate control run) and maximum horizontal steady eddy momentum flux convergence over the equator.

Figure 8 shows aspects of the mean climate from the “F” run, which was forced with a zonally invariant zonal acceleration equal to the net (horizontal plus vertical) steady eddy momentum flux convergence from the C* run (i.e., Figs. 7e,f). The mean zonal wind difference between F and C (Fig. 8a) is virtually identical to the mean zonal wind difference between C* and C (Fig. 5a), which confirms that the mean zonal wind response in C* can be explained by considering the fluxes of

TABLE 1. Description of model integrations. The heating and zonally invariant zonal accelerations used to force each run are listed in the second and third columns (see text for further explanation). The final two columns report the maximum horizontal steady eddy momentum flux convergence and mean zonal wind change between 10°S and 10°N, relative to the appropriate control integration.

Run	Heating	Zonal acceleration	$-d[\bar{u}^*\bar{v}^*]/dy$ maximum 10°S–10°N ($\text{m s}^{-1} \text{ day}^{-1}$)	$\Delta[\bar{u}]$ maximum 10°S–10°N (m s^{-1})
C	—	—	—	—
C*	Q^*	—	+0.67	+22
F	—	$-d[\bar{u}^*\bar{w}^*]/dp - d[\bar{u}^*\bar{v}^*]/dy$ from C*	—	+24
Fh	—	$-d[\bar{u}^*\bar{v}^*]/dy$ from C*	—	+103
Fh-	—	$+d[\bar{u}^*\bar{v}^*]/dy$ from C*	—	-21
Fh-*	Q^*	$+d[\bar{u}^*\bar{v}^*]/dy$ from C*	+0.97	+15
S	Q_s	—	—	—
S*	$Q_s + Q^*$	—	+0.60	+6
SF	Q_s	$-d[\bar{u}^*\bar{w}^*]/dp - d[\bar{u}^*\bar{v}^*]/dy$ from C*	—	+7
SFh	Q_s	$-d[\bar{u}^*\bar{v}^*]/dy$ from C*	—	+18
E	Q_e	—	—	—
E*	$Q_e + Q^*$	—	+0.86	+9
TRMM	LH	—	+0.38	—
TRMM*	LH*	—	+0.43	+13

momentum associated with the three-dimensional steady eddy response to Q^* . Advection by the mean vertical flow (not shown) remains the dominant easterly acceleration over the equator in the F run. However, the mean temperature response, mean meridional circulation, and transient kinetic energy in F (Figs. 8b,c,d) all exhibit considerably smaller departures from the C climate than the corresponding fields from the C* integration (Figs. 5b, 4b,c). We will briefly discuss these differences before returning to the mean zonal wind response.

The C* and F climates both exhibit mean temperature changes that are consistent with the demands of thermal wind balance, namely, cooling in the lower stratosphere and warming in the tropical upper troposphere and near the subtropical tropopause. However, the longitudinal collocation of the largest temperature and vertical velocity variations near the equator in the C* run (see Figs. 6a,c) gives rise to an upward steady eddy heat flux (not shown) that amplifies the warming in the upper troposphere and leads to cooling in the lower troposphere. The C* climate also includes pole-

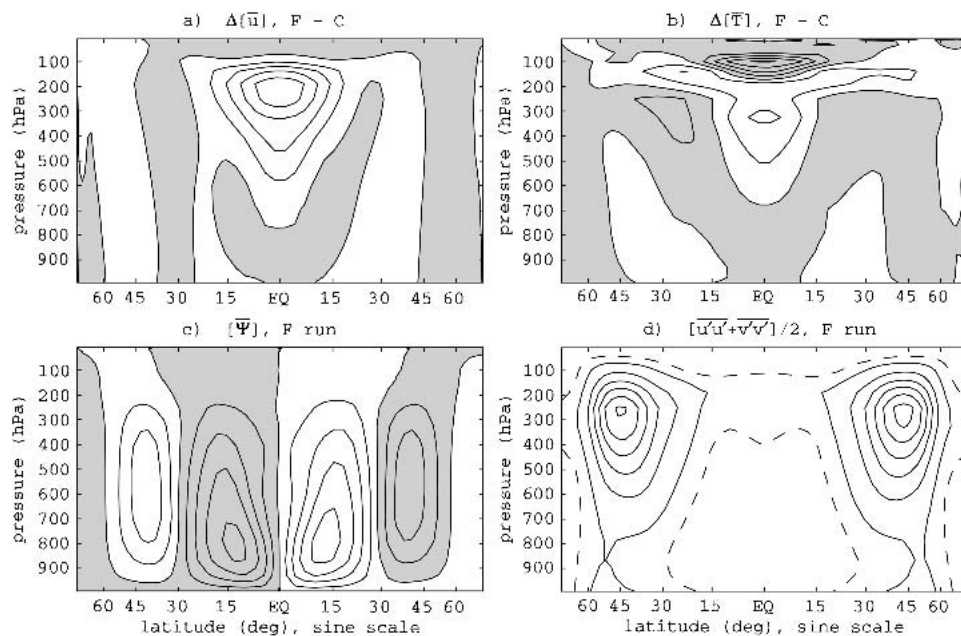


FIG. 8. The mean (a) zonal wind and (b) temperature differences between the F and C runs, and the mean (c) meridional mass circulation and (d) transient kinetic energy in the F run, which was forced with a zonally invariant zonal acceleration equal to the net (horizontal plus vertical) steady eddy momentum flux convergence from the C* run. The contour intervals are (a) 5 m s^{-1} , (b) 1 K , (c) $2 \times 10^{10} \text{ kg s}^{-1}$, and (d) $50 \text{ m}^2 \text{ s}^{-2}$, with a dashed contour added at $25 \text{ m}^2 \text{ s}^{-2}$ in (d). Negative (easterly/colder/counterclockwise) values are shaded.

ward steady eddy heat fluxes in the upper troposphere (not shown) that enhance the warming near the subtropical tropopause, offset some of the warming associated with the vertical steady eddy heat fluxes over the equator, and accomplish some of the poleward heat transport normally provided by the mean meridional circulation, which accounts for the weak Hadley cells in the C* run. Increases in static stability over the equator also reduce the mean meridional overturning required to maintain the meridional temperature gradient, which explains why the small temperature changes in the F run also lead to a slight reduction in the Hadley circulation (cf. Fig. 9c with Fig. 1b).

A cross-spectral decomposition was performed on

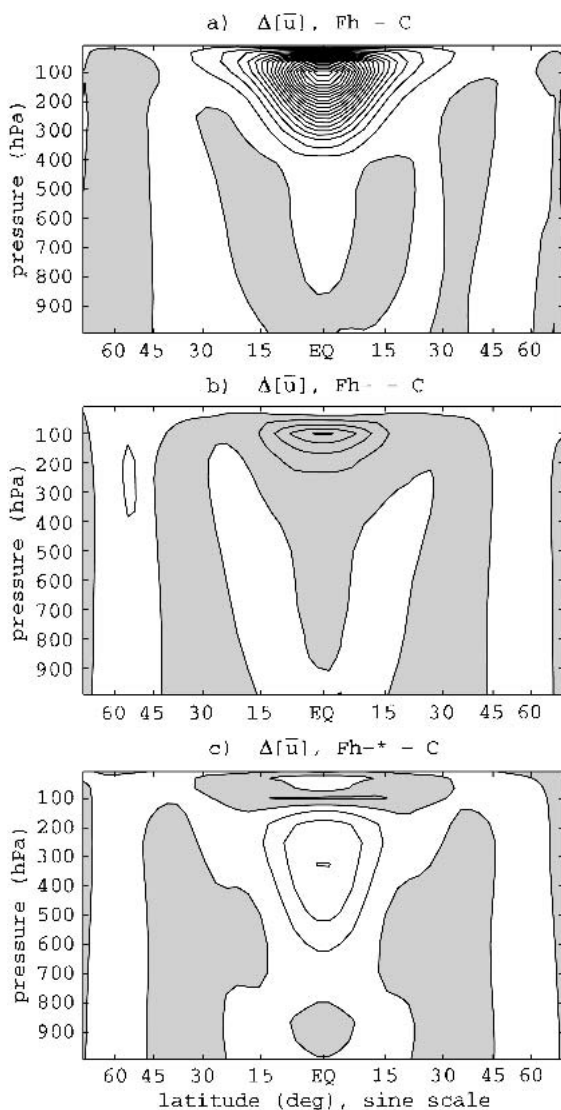


FIG. 9. The mean zonal wind change, relative to the C run, for the (a) Fh, (b) Fh-, and (c) Fh-* runs, which illustrates the model's response to easterly versus westerly zonal accelerations in the tropical upper troposphere (see text). The contour interval is 5 m s^{-1} and easterly zonal wind anomalies are shaded.

the daily zonal and meridional winds in C* and F to determine the origin of the large transient kinetic energy increases over the equator in Figs. 4c and 8d, and to determine why the C* run exhibits a larger increase than the F run. The results of this decomposition (not shown) indicate that roughly one-third of the transient kinetic energy increase over the equator in C* (Fig. 4c) is associated with variability in the zonally averaged flow (i.e., $[\bar{u}][\bar{u}]' + [\bar{v}][\bar{v}]'$), another third is concentrated at zonal-wavenumber-2 with periods longer than 10 days, and the remaining third is distributed over other zonal wavenumbers (mainly wavenumbers 3–8) at periods of 3–10 days; the F run (Fig. 8d) has low-frequency zonal-mean and high-frequency zonally asymmetric components that are similar to those in C*, but the wavenumber-2 feature is absent. The obvious interpretation of these results is that temporal variability in the stationary wave response to the Q^* tropical eddy heating accounts for the $\sim 50\%$ larger transient kinetic energy increase over the equator in C*, relative to F, and that the transition to superrotation gives rise to a small increase in the number of extratropical eddies that are able to reach the equator. Cross-spectral decompositions of the small transient eddy momentum fluxes near the equator in C, C*, and F lead to a similar conclusion. Neale (1999) and Hoskins et al. (1999) report that midlatitude transient eddies also fail to penetrate deeply enough into the Tropics to interact with the zonal-mean flow over the equator in their superrotating aquaplanet model, which suggests that the negative feedback between tropical zonal wind speed and extratropical transient eddy dissipation postulated by Saravanan (1993) does not operate in models with multiple vertical levels.

To confirm our diagnosis of the relative roles played by the horizontal and vertical steady eddy momentum fluxes in the C* climate, another zonally invariant zonal forcing experiment was performed using only the horizontal component of the C* steady eddy momentum flux convergence (Fig. 7e) to accelerate the zonal-mean zonal winds. The mean zonal wind response in this “Fh” run (Fig. 9a) exhibits an intense ($>100 \text{ m s}^{-1}$) westerly wind anomaly in the tropical upper troposphere and lower stratosphere. This result confirms that the horizontal steady eddy momentum fluxes in C* drive the transition to superrotation. It was also determined that the Fh climate makes a rapid transition to the F climate when a zonal acceleration equal to the vertical steady eddy momentum flux convergence from C* (Fig. 7f) is added to the forcing, which confirms that the primary role of the vertically oriented steady eddy momentum fluxes in C* is to limit the magnitude of the mean zonal wind response by redistributing momentum downward.

Several additional experiments were performed with zonally invariant accelerations that are easterly over the equator (see Table 1) to gain insight into the stability of the mean zonal flow with respect to easterly

versus westerly torques. The Fh- run was performed using an acceleration equal to the horizontal steady eddy momentum flux *divergence* from C*, which was obtained by multiplying the zonal acceleration depicted in Fig. 7e by -1 . This acceleration is not intended to simulate any real process in the atmosphere, although one could think of it as roughly simulating the effects of cumulus friction, gravity wave drag, or some other unresolved physical processes that provides an easterly acceleration in the tropical upper troposphere. The mean zonal wind response in the Fh- run (Fig. 9b) exhibits an easterly anomaly of 20 m s^{-1} in the tropical upper troposphere and lower stratosphere, which is considerably weaker than the response to the equivalent westerly acceleration (Fig. 9a). Advection by the mean vertical flow balances the imposed zonal acceleration in both Fh and Fh-, but the westerly (easterly) wind perturbation in Fh (Fh-) is associated with an increase (decrease) in upper-tropospheric temperatures and tropical static stability via thermal wind balance, which leads to a weaker (stronger) Hadley circulation, weaker (stronger) vertical flow in the upper troposphere, and hence a stronger (weaker) mean zonal wind response.

Since the earth's troposphere does not exhibit equatorial superrotation, despite the presence of strong zonally asymmetric heating in the Tropics (Figs. 3c,d), additional experiments were performed with both tropical eddy heating and easterly accelerations present to determine if the mean-state changes induced by fixed easterly torques could prevent superrotation. The Fh-* run, which was performed with both Q^* and the Fh-zonal acceleration applied, produced a mean zonal wind response (Fig. 9c) similar in structure, and $\sim 30\%$ weaker than, the mean zonal wind response in C* (Fig. 5a). Experiments performed using other fixed easterly accelerations produced similar responses (not shown). An analysis of the zonal momentum balance (1) in the Fh-* run indicates that the horizontal steady eddy momentum flux convergence at the equator is $\sim 50\%$ larger in Fh-* than in C* (see Table 1), while the mean vertical advection is $\sim 30\%$ weaker due to the smaller zonal wind response aloft. These results suggest that fixed easterly accelerations in the tropical upper troposphere cannot prevent the transition to superrotation because the mean flow changes induced by easterly torques lead to stronger eddy momentum fluxes.

To further investigate the influence of the zonal-mean climate on the steady eddy response, eddy heating experiments were performed using a linearized version of the model with the C, C*, and Fh-* mean climates as basic states. The results of these experiments (not shown) indicate that the structure of the stationary wave response and the magnitude of the corresponding eddy momentum fluxes are sensitive to the zonal-mean zonal wind distribution. The strongly superrotating C* mean zonal wind field yields a linear stationary wave response similar in magnitude and structure to that

shown in Fig. 6, while the Fh-* basic state and the C basic state produce linear stationary wave responses with stronger horizontal circulation features and more concentrated eddy momentum fluxes. The decrease in the amplitude of the stationary wave response as the mean zonal winds over the equator become more westerly may be attributed in part to the reduction in the meridional gradient in absolute vorticity, which decreases the amplitude of the Rossby wave source (Sardeshmukh and Hoskins 1988). The weaker eddy fluxes in the superrotating basic state, along with the prominent role played by mean vertical advection in the mean zonal wind balance for both the C* and Fh-* run, indicate that the transition to superrotation itself is responsible for limiting the magnitude of the equatorial westerlies. The sensitivity of the eddy response to the mean zonal wind distribution also suggests that the strong steady eddy momentum fluxes obtained by Neale (1999) and Hoskins et al. (1999) might be related to the robust subtropical jet streams in their aquaplanet model.

c. Influence of hemispheric asymmetry

The similarity between our results and those obtained using aquaplanet models forced with steady thermal contrasts along the equator suggests that equatorial superrotation is a ubiquitous response to tropical eddy forcing under hemispherically symmetric (or perpetual-equinox) boundary conditions. Furthermore, the results of the preceding section suggest that easterly accelerations associated with subgrid-scale processes are unlikely to be responsible for preventing superrotation on earth. In this section, we evaluate the influence of hemispheric asymmetry on the response to tropical eddy heating. Previous authors have suggested that hemispheric asymmetry in the Hadley circulation, especially the tendency for the strongest rising motion to occur off the equator, is responsible for maintaining the observed deep easterly flow over the equator (Lindzen and Hou 1988; Hou 1993; Hou and Molod 1995; Lee 1999).

Following Hou (1993), we determined that a reasonable facsimile of the mean meridional circulation on earth during solstitial months could be produced in our model by adding a steady, zonally symmetric, vertically uniform thermal forcing with the following structure:

$$Q_s(\phi) = (1 \text{ K day}^{-1}) \exp \left[- \left(\frac{\phi - 7.5^\circ}{7.5^\circ} \right)^2 \right] \\ - (0.5 \text{ K day}^{-1}) \exp \left[- \left(\frac{\phi + 15^\circ}{15^\circ} \right)^2 \right], \\ \sigma_t < \sigma < 1.$$

The latitudinal position and narrow meridional profile of the zonally invariant heating north of the equator is intended to replicate the intertropical convergence

zone in the summer hemisphere, while the weaker and broader cooling anomaly south of the equator is intended to represent the radiative cooling typically observed in the winter hemisphere during solstitial months (see, e.g., Nigam et al. 2000). Although the cooling amplitude is weaker than the heating amplitude, the broader meridional scale of the cooling ensures that the net heat added to the system is nearly zero.

The mean climate in the solstitial control, or "S," run (Fig. 10) bears a strong resemblance to the zonally averaged climate observed during solstitial months on earth (Peixoto and Oort 1992). Comparing Fig. 10b with Fig. 1b, we see that the solstitial heating anomaly causes the winter hemisphere Hadley cell to strengthen and expand across the equator, with strong rising motion centered at 7.5°N , enhanced sinking motion in the Southern Hemisphere, and strong cross-equatorial flow. The mean meridional winds (Fig. 10c), which reach 3 m s^{-1} near 200 hPa over the equator, are also stronger and more concentrated in the upper troposphere than the diffuse poleward flow produced by the HS94 forcing. The advection of zonal momentum by this cross-equatorial flow (Fig. 10d) is associated with an easterly acceleration of greater than $1 \text{ m s}^{-1} \text{ day}^{-1}$. However, the mean zonal winds in the S run (Fig. 10a) are only slightly more easterly in the tropical upper troposphere than those in the C integration, which is consistent with the weak response to easterly torques noted in the previous section. The enhanced Hadley

circulation also gives rise to a strengthening and equatorward shift in the winter hemisphere jet stream, which is accompanied by corresponding shifts in the transient kinetic energy and momentum flux convergence (not shown).

The strong hemispheric asymmetry produced by the solstitial heating pattern changes the response of the model to tropical eddy heating dramatically. The mean zonal winds in the S^* run (Fig. 11a), which was performed with both Q_s and Q^* applied, are very similar to the mean zonal wind in the solstitial control run (Fig. 10a), and the mean zonal wind difference between S^* and S (Fig. 11b) reveals only a weak westerly anomaly over the equator. Although the zonal winds in the tropical midtroposphere are westerly rather than easterly in S^* , the winds at the equatorial tropopause are easterly, and a broad region of weak ($< 5 \text{ m s}^{-1}$) winds stretches from the equator to 15°N . The mean meridional circulation, mean temperature, and transient kinetic energy in the S^* run (not shown) are also similar to their counterparts in S, which is consistent with the lack of a strong mean zonal wind response, although the steady eddy heat fluxes associated with the stationary wave response do give rise to a slight ($\sim 0.5 \text{ K}$) warming of the upper troposphere and a slight ($\sim 10\%$) reduction in the Hadley circulation. Runs with different eddy heating anomalies applied, including anomalies centered at the same latitude as the zonal-mean heating, yielded similar results.

As in C^* , the vertical fluxes of zonal momentum as-

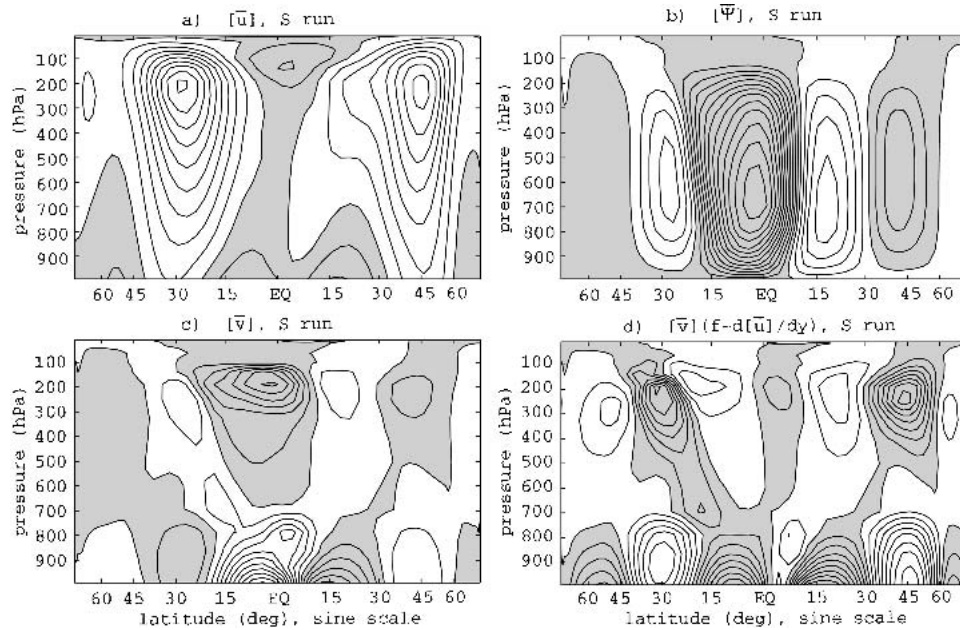


FIG. 10. The mean (a) zonal winds, (b) meridional mass circulation, (c) meridional winds, and (d) zonal wind tendency associated with advection by the mean meridional winds in the S run forced with the zonally invariant, hemispherically asymmetric heating function Q_s . The contour intervals are (a) 5 m s^{-1} , (b) $2 \times 10^{10} \text{ kg s}^{-1}$, (c) 0.5 m s^{-1} , and (d) $1 \text{ m s}^{-1} \text{ day}^{-1}$. Negative (counterclockwise/southward/easterly) values are shaded.

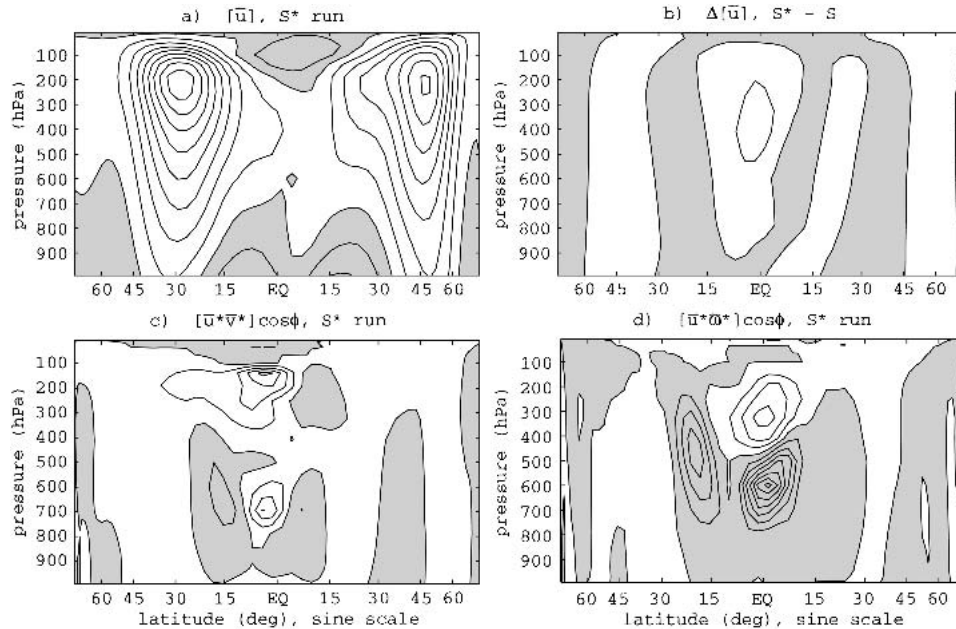


FIG. 11. The mean (a) zonal winds, (b) zonal wind change (relative to the solstitial control run), and zonally averaged (c) meridional and (d) vertical fluxes of zonal angular momentum by steady eddies in the S^* run, which was forced with both the zonally invariant solstitial heating function Q_s and the tropical eddy heating Q^* . The contour intervals are (a), (b) 5 m s^{-1} , (c) $2 \text{ m}^2 \text{ s}^{-2}$, and (d) $0.025 \text{ m Pa s}^{-2}$. Negative (easterly/southward/upward) values are shaded.

sociated with the steady eddy response in the S^* run (Fig. 11d) act to offset much of the westerly acceleration associated with the horizontally oriented steady eddy circulations. However, the horizontal steady eddy momentum fluxes in S^* (Fig. 11c) do exhibit one curious property: they emanate almost entirely from the winter hemisphere. Watterson and Schneider (1987) found that planetary waves forced at low latitudes in the presence of a strong cross-equatorial flow preferentially propagate in the direction of that flow, which tends to enhance the eddy circulation in the winter hemisphere and reduce the eddy response in the summer hemisphere. Since the meridional propagation of planetary waves is associated with a net flux of westerly momentum in the opposite direction, the strong cross-equatorial flow in the tropical upper troposphere of our model (Fig. 10c) enhances the equatorward steady eddy momentum fluxes from the winter hemisphere and sharply reduces the fluxes from the summer hemisphere, leading to the pattern in Fig. 11c. Integrations performed using a linearized version of the model confirm that the cross-equatorial meridional flow in the solstitial basic state promotes an enhanced stationary wave response and larger eddy momentum fluxes in the winter hemisphere, along with a reduced response in the summer hemisphere. Dima et al. (2005, hereafter DWK) report a similar relationship between the meridional winds and eddy momentum fluxes on earth.

Figure 12 displays selected components of the mean

zonal wind balance [see Eq. (1)] in the S^* run minus the corresponding terms from the S run (the omitted components show little or no change over the equator). The net (horizontal plus vertical) steady eddy momentum flux convergence in S^* (Fig. 12a) provides a strong westerly acceleration in the midtroposphere over the equator, along with easterly accelerations in the subtropics, which is similar to the zonal acceleration pattern associated with the steady eddy response in C^* (Figs. 7e,f). The steady eddy momentum flux convergence over the equator is offset primarily by changes in the advection of zonal momentum by the mean vertical flow (Fig. 12b), which provides an easterly acceleration in the midtroposphere because the weak easterly vertical shear that was present in S has been replaced with a weak westerly vertical shear (cf. Figs. 10a and 11a). This balance between steady eddy forcing and mean vertical advection is similar to the dominant mean zonal wind balance over the equator in C^* , except that the strong upward motion in the rising branch of the solstitial Hadley circulation allows vertical advection to balance the stationary wave forcing with only a small change in the mean zonal wind distribution. In addition, the strong cross-equatorial flow in the S^* climate allows small changes in the meridional shear to yield a substantial increase in the easterly zonal acceleration associated with advection of momentum by the mean meridional flow in the upper troposphere (Fig. 12c), which offsets both the residual westerly steady eddy acceleration and the small westerly acceleration associated with

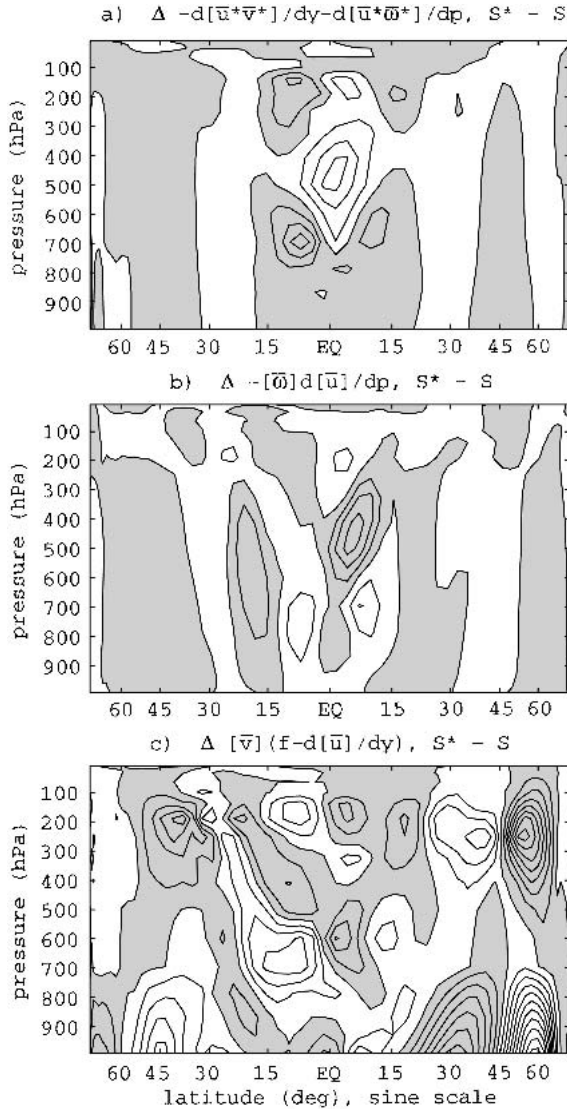


FIG. 12. The mean zonal wind tendencies associated with changes in the (a) total (horizontal plus vertical) steady eddy momentum flux convergence, (b) advection of zonal momentum by the mean meridional winds, and (c) advection of zonal momentum by the mean vertical flow between the S^* and S runs [see Eq. (1)]. The contour interval is $0.25 \text{ m s}^{-1} \text{ day}^{-1}$ and easterly zonal acceleration anomalies are shaded.

mean vertical advection above the zonal wind maximum. Hence, the weak mean zonal wind response to tropical eddy heating in the solstitial climate may be attributed to both the strength of the meridional overturning and the presence of cross-equatorial flow aloft.

Several integrations were also performed with fixed zonally invariant zonal accelerations in the solstitial climate (see Table 1). The SF run, which was forced with both the Q_s solstitial heating pattern and the F forcing described in the previous subsection (i.e., a zonally invariant zonal acceleration equal to the net steady eddy momentum flux convergence from the C^* run), pro-

duced a mean zonal wind response (not shown) nearly identical to the S^* mean zonal wind response in Fig. 11b. The SFh run, which was forced with Q_s and the horizontal component of the steady eddy momentum flux convergence from the C^* run, exhibited a westerly wind anomaly of 18 m s^{-1} , but this anomaly was concentrated in the lower stratosphere between 150 and 50 hPa, indicating that the downward flux of momentum by vertically oriented steady eddy circulations remains important for preventing superrotation above the tropopause in the solstitial climate. Analogous runs forced with zonal accelerations equal to the horizontal and net steady eddy momentum flux convergences from the S^* run yielded similar results (not shown).

d. Role of enhanced meridional overturning

The prominent role played by vertical advection in the mean zonal wind balances of the solstitial and superrotating climates suggests that the mean zonal winds in the tropical upper troposphere are simply sensitive to the strength of the mean meridional overturning. HE99 and Woolnough (1997) have also suggested that the response to tropical eddy heating may depend strongly on the strength of the mean meridional circulation. To test this hypothesis, we performed several model runs with the following “superequinox” heating anomaly imposed:

$$Q_e(\phi) = (1 \text{ K day}^{-1}) \exp\left[-\left(\frac{\phi}{7.5^\circ}\right)^2\right] - (0.3 \text{ K day}^{-1}) \left\{ \exp\left[-\left(\frac{\phi + 15^\circ}{15^\circ}\right)^2\right] + \exp\left[-\left(\frac{\phi - 15^\circ}{15^\circ}\right)^2\right] \right\}.$$

This zonally invariant heating pattern is similar to the solstitial heating perturbation (Q_s), except that it is hemispherically symmetric with heating at the equator and cooling in each hemisphere.

The Q_e heating yields a control climate with stronger Hadley cells (Fig. 13b) and subtropical jet streams (Fig. 13a) than those obtained using HS94 boundary conditions (Fig. 1). Interestingly, this “E” climate is similar to the mean state obtained when aquaplanet models are integrated with zonally symmetric, perpetual-equinox boundary conditions (Battisti and Ovens 1995; Hoskins et al. 1999), which suggests that the differences between the C run and the control climates produced by aquaplanet models arise mainly because the physical parameterizations in aquaplanet models produce stronger zonally averaged diabatic heating in the Tropics than the HS94 parameterizations. Aquaplanet models also tend to produce dual ITCZs, with heating maxima at low latitudes in both hemispheres, and weak equatorward flow in between, which could explain why their control climates exhibit stronger easterly flow over the equator than our idealized model.

The mean zonal winds in the E^* run (Fig. 13c) indicate that the response to tropical eddy heating in the

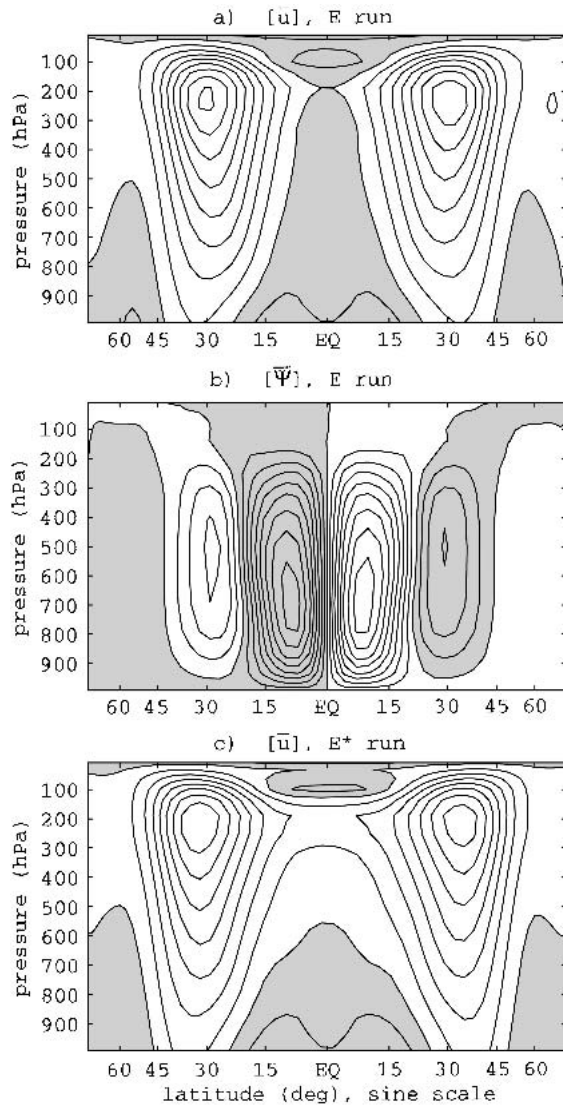


FIG. 13. The mean (a) zonal winds and (b) meridional mass circulation in the E (superequinox control) run, which was forced with the zonally invariant, hemispherically symmetric heating function Q_e , and (c) mean zonal winds in the E* run, which was forced with both Q_e and the tropical eddy heating Q^* . The contour intervals are (a) 5 m s^{-1} , (b) $2 \times 10^{10} \text{ kg s}^{-1}$, and (c) 5 m s^{-1} . Negative (counterclockwise/easterly) values are shaded.

superequinox context is similar in structure to the response obtained under HS94 boundary conditions with westerly mean flow in the upper troposphere over the equator. The mean zonal wind balance in E* (not shown) indicates that advection by the mean vertical flow remains the dominant restoring force opposing the westerly acceleration over the equator associated with the steady eddy response, but the response is much weaker in E* than in C* because the mean vertical shear does not need to be as strong to balance the steady eddy momentum flux convergence (which is only slightly stronger in E*, as indicated in Table 1).

The maximum zonal wind response also occurs higher in the troposphere in E* than in S*, which suggests that the easterly acceleration associated with cross-equatorial flow in the upper troposphere (Fig. 10d) is responsible for limiting the maximum zonal wind change near the tropopause in the solstitial climates. Additional experiments (not shown) indicate that the adding a zonally invariant zonal acceleration equal to the easterly zonal acceleration associated with cross-equatorial flow in the solstitial climate (in effect, Fig. 10d minus Fig. 2a) brings the superequinox solutions even closer to their solstitial counterparts.

e. Response to TRMM latent heating fields

To examine the relationship between the mean meridional circulation, tropical stationary wave forcing, and the upper-tropospheric mean zonal winds with a more realistic representation of tropical heating, two final experiments were performed using the annually averaged latent heating distribution calculated by Schumacher et al. (2004) from Tropical Rainfall Measuring Mission (TRMM) satellite data. The first experiment was performed using only the zonally asymmetric component of the TRMM latent heating field (Figs. 3c,d), and the second experiment was performed using the full (eddy plus zonal mean) TRMM latent heating distribution, which includes a deep zonal-mean heating centered at approximately 7°N .

The zonally asymmetric component of the TRMM heating produced a mean climate with superrotation in the tropical upper troposphere (Fig. 14a), and steady eddy momentum fluxes (not shown, but see Table 1) similar in structure and magnitude to those obtained using the idealized Q^* heating. In contrast, the full TRMM latent heating field produced a mean zonal wind field (Fig. 14c) with weak zonal flow across a broad latitudinal range. The mean meridional circulations from these two runs (Figs. 14b,d) indicate that the inclusion of the zonal-mean heating leads to a stronger and less hemispherically symmetric mean meridional circulation, particularly in the upper troposphere, and the advection of momentum by this mean meridional flow (not shown) provides an easterly acceleration that offsets the momentum flux convergence associated with the steady eddy circulations in the tropical upper troposphere. Similar results were obtained using the TRMM-derived eddy and total latent heating patterns for individual seasons. These results suggest that the reason that the earth does not experience equatorial superrotation in the troposphere is because the zonally averaged tropical heating, and hence the mean meridional circulation, almost always exhibit some amount of hemispheric asymmetry.

4. Conclusions

A general circulation model with simplified atmospheric physics and zonally invariant boundary condi-

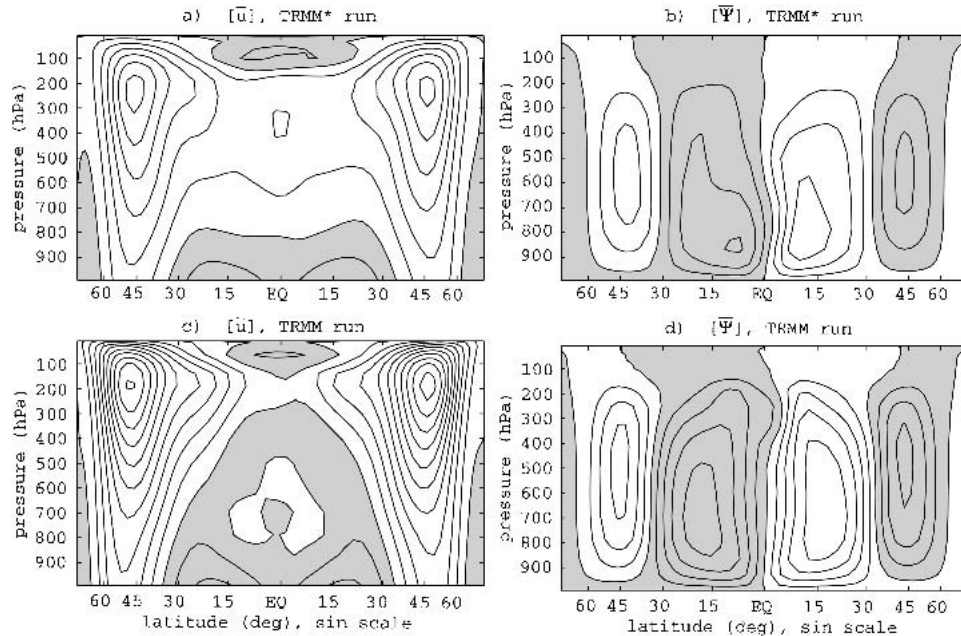


FIG. 14. The mean (a) zonal winds and (b) meridional mass circulation in the TRMM* run, which was forced with the annual-mean eddy latent heating pattern depicted in Figs. 3c,d, and the mean (c) zonal winds and (d) meridional mass circulation in the TRMM run, which was forced with the total (eddy plus zonal mean) annual-mean latent heating, as derived by Schumacher et al. (2004). The contour intervals are (a),(c) 5 m s^{-1} , and (b),(d) $2 \times 10^{10} \text{ kg s}^{-1}$. Negative (easterly/counterclockwise) values are shaded.

tions was subjected to a variety of idealized thermal and mechanical forcings to investigate the maintenance and stability of the zonal-mean zonal winds in the tropical upper troposphere. The results of these experiments indicate that 1) longitudinal variations in diabatic heating at low latitudes invariably give rise to a steady eddy response that includes a net flux of westerly momentum from the subtropics to the midtroposphere over the equator; 2) under hemispherically symmetric boundary conditions, the momentum flux convergence associated with tropical wave forcing leads to persistent equatorial superrotation (westerly zonal-mean winds over the equator); 3) the response of the zonal-mean flow to tropical eddy heating is reduced dramatically in a “solstitial” climate with a single dominant Hadley cell straddling the equator; and 4) extratropical transient eddies and specified zonally invariant easterly accelerations at low latitudes do not have a strong influence on the mean zonal winds over the equator.

The main conclusion that can be drawn from these results is that hemispheric asymmetry is crucial for maintaining the mean easterly flow in the tropical upper troposphere against the westerly accelerations associated with zonally asymmetric tropical heating. Hemispheric asymmetry in the zonal-mean heating distribution inhibits the transition to superrotation in two ways: by increasing the strength of the mean meridional overturning, which enables the zonally averaged flow in the meridional plane to more efficiently erode any mean zonal wind anomalies, and by inducing cross-

equatorial flow aloft, which provides a mean easterly acceleration in the tropical upper troposphere. Vertically oriented steady eddy momentum fluxes, which arise due to the collocation of the vertical velocity variations and zonal wind anomalies forced by the eddy heating at low latitudes, also act to limit the magnitude of the mean zonal wind response over the equator under both hemispherically symmetric and solstitial boundary conditions by redistributing momentum in the vertical. In our model, all three of these processes need to be present in order to prevent superrotation.

Another interesting aspect of our solstitial experiments is that the cross-equatorial meridional flow aloft restricts the meridional propagation of stationary waves into the summer hemisphere, so that the meridional fluxes of westerly angular momentum associated with steady eddies in the tropical band emanate predominantly from the winter hemisphere. Although this effect does not have a large influence on the mean zonal wind response in our model, the cancellation between the easterly acceleration associated with the cross-equatorial flow aloft and the westerly acceleration associated with the horizontal eddy momentum fluxes suggests that a more fundamental relationship may exist between stationary waves and the mean meridional flow in the tropical upper troposphere. DWK have recently found a similar relationship between the mean meridional flow and eddy momentum fluxes in the NCEP reanalysis. The ubiquitous cancellation between the vertical and horizontal steady eddy momentum flux

convergences in our model also suggests that vertically oriented steady eddy circulations could represent an important feedback on the zonal-mean zonal winds at low latitudes. Tropical stationary waves also appear capable of influencing the mean temperature distribution in the upper troposphere and lower stratosphere.

The tropical upper troposphere is often thought of as a region with small zonal accelerations relative to those in the extratropics, the stratosphere, or the boundary layer. The results of this study, however, combined with the observational analyses of Lee (1999) and DWK, suggest that the zonal-mean zonal winds over the equator actually reflect a balance between strong westerly accelerations associated with longitudinal variations in diabatic heating and strong easterly accelerations associated with the mean meridional circulation and vertically oriented stationary waves. On earth, these easterly accelerations are apparently strong enough to offset the westerly accelerations associated with tropical wave sources, especially during solstitial months, leading to the deep easterly flow observed over the equator. However, in multilevel GCMs with hemispherically symmetric boundary conditions, eddy forcing at low latitudes can accelerate the zonal winds in the equatorial upper troposphere until the vertical wind shear becomes large enough for vertical advection of momentum by the mean meridional circulation to balance the net eddy momentum flux convergence over the equator. The sensitivity of the mean zonal wind balance at low latitudes to both zonal and hemispheric asymmetries in diabatic heating raises the interesting possibility that the mean zonal wind distribution in the Tropics may have been radically different during past climates, and could change again in the future.

The importance of hemispheric asymmetry in regulating the mean flow over the equator also suggests that climate models that replace the seasonal cycle with annual-mean or perpetual-equinox boundary conditions may be overly sensitive to tropical wave forcing. The superrotating general circulations obtained by Battisti and Ovens (1995), Neale (1999), and Hoskins et al. (1999) using aquaplanet models with zonal-wavenumber-1 sea surface temperature anomalies at low latitudes and otherwise zonally symmetric, perpetual-equinox climates almost certainly fall into this category. In contrast, Ting and Held (1990) only obtained a small increase in tropical zonal wind speed when a large sea surface temperature dipole was imposed along the equator in an aquaplanet model with perpetual-solstice boundary conditions. However, Inatsu et al. (2002) obtained strong equatorial superrotation when a similar model was forced with zonal-wavenumber-1 sea surface temperature variations at the equator, and we have also found it difficult to prevent superrotation in an aquaplanet version of CCM3.6 forced with eddy sea surface temperature variations at low latitudes under perpetual-solstice boundary condi-

tions. Additional work is clearly needed to evaluate the relationship between the zonally averaged climate and the dynamical response to eddy forcing in this more complicated setting.

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