Hemispherical Asymmetry of Tropical Precipitation in ECHAM5/MPI-OM during El Niño and under Global Warming

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ABSTRACT

Similarities and differences between El Niño and global warming are examined in hemispherical and zonal tropical precipitation changes of the ECHAM5/Max Planck Institute Ocean Model (MPI-OM) simulations. Similarities include hemispherical asymmetry of tropical precipitation changes. This precipitation asymmetry varies with season. In the boreal summer and autumn (winter and spring), positive precipitation anomalies are found over the Northern (Southern) Hemisphere and negative precipitation anomalies are found over the Southern (Northern) Hemisphere. This precipitation asymmetry in both the El Niño and global warming cases is associated with the seasonal migration of the Hadley circulation; however, their causes are different. In El Niño, a meridional moisture gradient between convective and subsidence regions is the fundamental basis for inducing the asymmetry. Over the ascending branch of the Hadley circulation, convection is enhanced by less effective static stability. Over the margins of the ascending branch, convection is suppressed by the import of dry air from the descending branch. In global warming, low-level moisture is enhanced significantly due to warmer tropospheric temperatures. This enhances vertical moisture transport over the ascending branch of the Hadley circulation, so convection is strengthened. Over the descending branch, the mean Hadley circulation tends to transport relatively drier air downward, so convection is reduced.

1. Introduction

Climate changes associated with anthropogenic forcings, such as greenhouse gases, are becoming influential on human life. The most pronounced feature is the warming of the earth's surface and the troposphere, almost everywhere. This global warming is found not only in climate model simulations (e.g., Boer et al. 2000; Dai et al. 2001b; Delworth and Knutson 2000; Hansen et al. 1995; Houghton et al. 2001; Meehl et al. 2007; Lucarini and Russell 2002; Manabe et al. 1991; Meehl et al. 2005, 2006; Mitchell et al. 2000; Stott et al. 2006; Teng et al. 2006; Washington et al. 2000; Watterson et al. 1999; Yonetani and Gordon 2001) but also in observations (e.g., Houghton et al. 2001; Trenberth et al. 2007; Jones and Moberg 2003; Mann et al. 1998; Mears et al. 2003). Climate model simulations also show substantial tropical precipitation changes at the end of the twenty-first century (e.g., Allen and Ingram 2002; Boer et al. 2000; Dai et al. 2001a; Douville et al. 2002; Meehl et al. 2007, 2000, 2003; Roeckner et al. 1999; Williams et al. 2001). However, the spatial distribution of the precipitation changes shows poor agreement among climate model simulations on a regional basis (e.g., Allen and Ingram 2002; Houghton et al. 2001; Meehl et al. 2007; Neelin et al. 2003, hereafter NCS03, 2006).

To understand global warming impacts on regional tropical precipitation, Chou and Neelin (2004, hereafter CN04) proposed two dominant mechanisms: the upped-ante and the rich-get-richer (anomalous gross moist stability, $M'$) mechanisms. In awarmer climate, the atmospheric boundary layer (ABL) moisture over convective regions increases under convective quasi

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equilibrium (QE). This so-called QE mediation does not occur over nonconvective regions. This creates a horizontal gradient of low-level moisture anomalies between convective and nonconvective regions, which is fundamental for those two mechanisms of regional tropical precipitation changes. In the upped-ante mechanism, dry inflow from nonconvective regions to convective regions reduces precipitation over margins of convective regions. In the rich-get-richer mechanism, the enhanced low-level moisture over convective regions decreases the effective static stability $M$, so low-level convergence and the associated precipitation are enhanced over regions with strong climatological convergence and precipitation. These mechanisms have also been found in a more comprehensive coupled atmosphere–ocean model (Chou et al. 2006).

El Niño events can be used as an analog to study possible mechanisms that lead us to understand how global warming affects tropical precipitation. An El Niño–like sea surface temperature (SST) anomaly pattern is often found in global warming simulations (Meehl et al. 2007; Meehl and Washington 1996; Teng et al. 2006). Thus, mechanisms that induce tropical precipitation changes in El Niño may also change tropical precipitation in global warming. On the other hand, El Niño also creates large-scale warming over the entire tropical troposphere, which is similar to global warming, but with a weaker amplitude. Therefore, mechanisms for the El Niño–induced tropical precipitation changes under this warming condition can also be found in the global warming case (NCS03). In this study, we also focus on a comparison between the El Niño and global warming cases but from another point of view.

Under tropical tropospheric warming during El Niño, Chou and Lo (2007, hereafter CL07) examined observations for an asymmetric response in zonally averaged tropical precipitation changes. From boreal winter to spring, when El Niño is at its maximum strength, tropical precipitation over the Southern Hemisphere enhances, while tropical precipitation over the Northern Hemisphere reduces. Such asymmetry of tropical precipitation anomalies is associated with a seasonal migration of the Hadley circulation and the distribution of the ABL moisture anomalies between the ascending and descending branches of the Hadley circulation. This distribution of the ABL moisture anomalies between the ascending and descending branches should also be expected in the global warming case. Thus, the asymmetry of the tropical precipitation changes discussed in CL07 could also be found in the global warming case.

In this study, we explore the similarities and differences between global warming and El Niño under the influence of large-scale tropospheric warming in the ECHAM5/Max Planck Institute Ocean Model (MPI-OM) coupled general circulation model (CGCM) simulations, which are from the World Climate Research Program’s (WCRP) Coupled Model Intercomparison Project Phase 3 (CMIP3) multimodel dataset for the Fourth Assessment Report (AR4) of the Intergovernmental Panel on Climate Change (IPCC). We first briefly describe the ECHAM5/MPI-OM CGCM and discuss the method used in this analysis. We then derive moisture and moist static energy (MSE) equations in section 3. In section 4, we analyze El Niño simulated by this CGCM and discuss effects that are associated with the asymmetry in El Niño. We next analyze the global warming case and discuss the associated mechanisms. Comparisons between El Niño and global warming are discussed in section 6, followed by conclusions.

2. Model and method

a. Model

To examine the hemispherical asymmetry of tropical precipitation occurring in El Niño and global warming, simulations from ECHAM5/MPI-OM are analyzed. The ECHAM5 is the latest version of the ECHAM atmospheric model. This model uses spherical harmonics with standard triangular truncation at wavenumber 63 (T63), which is equivalent to a spatial resolution of $1.875^\circ \times 1.875^\circ$, with 31 vertical sigma levels that extend up to 10 hPa. The main changes from the ECHAM4 model are in the radiation and cloud schemes and in the coupling between land surface and atmosphere. An implicit scheme is used for coupling land surface and atmosphere (Schulz et al. 2001). The heat transfer in soil is also calculated by an implicit scheme. For the presence of snow, the top of the snow layer is considered as the top of the soil model. A new set of land surface data, including vegetation ratio, leaf area index, forest ratio, and background albedo, has been derived from a global 1-km-resolution dataset (Hagemann 2002). The MPI-OM uses a generalized orthogonal curvilinear C grid and includes an improved scheme for slope convection and a Gent–McWilliams style eddy-induced mixing parameterization (Gent and McWilliams 1990). An increased horizontal resolution is employed between 5°S and 5°N, with a grid spacing of 0.5° in the meridional and 2.5° in the zonal. The model has 23 vertical levels, with 10 over the upper 300 m. Flux adjustment is not included in the ECHAM5/MPI-OM simulations (Jungclaus et al. 2006). More de-
tails of these developments can be found in Roeckner et al. (2003) for the atmospheric component and in Marsland et al. (2003) for the oceanic component.

b. Method

The simulations of the twentieth-century climate in couple models (20C3M) experiment for the period of 1860–2000 and the IPCC scenario A2 for the twenty-first century are used. We consider only one realization in this study. To understand the characteristics of the hemispherical asymmetric responses in tropical precipitation during El Niño and global warming, these two different temporal variabilities are separated. We first remove the seasonal cycle, based on the seasonal slice of 1961 to 1990, for each of the grid points. We then use a linear regression to obtain the linear trend in 1960–2010 for each month separately. We note that the linear trend component calculated from the period of 1960–2100 may not be appropriate because the trend peaks up after 2040. However, the results in this study should not be altered by choice of the studying period. To be consistent with the analysis in IPCC AR4 (Meehl et al. 2007), which used the period in the twentieth century as the current climate for comparison, we thus choose 1960–2100 as our studying period. The fast Fourier transform (FFT) was used to get the interannual variability between 2 and 10 yr. The time series of the 1-yr running mean of global precipitation anomalies from 1960 to 2100 is used as an example. Figure 1a is the original data without the annual cycle, so it includes not only the trend component but also other climate variabilities, such as interannual and interdecadal variabilities. The part of the interannual variability is shown in Fig. 1b and the trend component with an increased rate of 0.014 mm day$^{-1}$ per decade is shown in Fig. 1c. To obtain a composite of the El Niño events in the ECHAM5/MPI-OM simulations, the Niño-3.4 index, which is the interannual component of the SST anomalies averaged over 5°S–5°N, 170°–120°W, is used. The warm event (El Niño) is defined as the December–February (DJF) Niño-3.4 index greater than 1.6°C (one standard deviation). On the other hand, the cold event (La Niña) is defined as the DJF Niño-3.4 index less than −1.6°C. Twenty-nine El Niño events and 23 La Niña events are found during 1960–2100. The changes induced by global warming are defined as differences of the trend component between the future climate (2071–2100) and the current climate (1961–90).

3. Moisture and moist static energy budgets

In the tropics, precipitation is mainly controlled by convective processes. Thus, a vertically integrated moisture budget is used to examine precipitation changes, and vertically integrated MSE budget, on the other hand, is used to examine anomalous vertical motion associated with convection. For time or ensemble averages, the time derivative term is negligible, so changes in the vertically integrated moisture equation can be written as

$$P' \approx -(\langle \omega \partial_q \rangle - \langle \mathbf{q} \mathbf{v} \mathbf{q} \rangle - \langle \mathbf{v} \cdot \nabla \mathbf{q} \rangle) + E',$$

(1)

where the overbar denotes climatology in the El Niño case or the current climate in the global warming case, and the prime represents the departure from the climatology or the current climate. The precipitation $P$ is in energy units (W m$^{-2}$), which, divided by 28, is millimeters per day. Here, $E$ is evaporation, $\omega$ is pressure velocity, $\mathbf{v}$ is horizontal velocity, and the specific humidity $q$ is in energy units by absorbing the latent heat per unit mass $L$. The vertical integral is indicated with the use of angle brackets and denotes a mass integration through the troposphere with $p_T$ as the depth of the troposphere:

$$\langle X \rangle = \frac{1}{g} \int_{p_s}^{p_T} X \, dp,$$

(2)

where $g$ is gravity and $p_s$ is surface pressure.

Following the moisture Eq. (1), changes in the vertically integrated MSE equation can also be written in a similar form:

$$\langle \omega' \partial_t \mathbf{R} \rangle \approx -\langle \mathbf{q} \mathbf{h}' \rangle - \langle \mathbf{v} \cdot \nabla (q + T) \rangle' + F_{\text{net}}',$$

(3)

where $T$ is atmospheric temperature in energy unit that absorbs the heat capacity at constant pressure $C_p$, and the MSE is $h = q + s$. The dry static energy is $s = T + \phi$, with $\phi$ being the geopotential. The net energy into the atmospheric column is

$$F_{\text{net}} = F_t - F_s.$$  

(4)

The net heat flux at the top of the atmosphere (TOA) is

$$F_t = S_t^\downarrow - S_t^\uparrow - R_t^\downarrow$$  

(5)

and the net heat flux at the surface is

$$F_s = S_s^\downarrow - S_s^\uparrow + R_s^\downarrow - R_s^\uparrow - E - H.$$  

(6)

Subscripts $s$ and $t$ on the solar ($S^\downarrow$ and $S^\uparrow$) and longwave ($R^\downarrow$ and $R^\uparrow$) radiative terms denote surface and model top, and $H$ is sensible heat flux. Positive $F_t$ and $F_s$ indicate downward heat fluxes. The nonlinear terms $\langle \omega \partial_t \mathbf{q} \rangle$ and $\langle \mathbf{q} \mathbf{v} \mathbf{q} \rangle$ are neglected in both moisture and MSE equations since they are small and do not change the result of our diagnosis. Note that monthly data are used here, so transients are not included in the nonlinear terms.
As discussed in the previous studies (Chou et al. 2006; CL07), accurately computing the terms associated with $\omega \partial_p h$ ($\omega \partial_p s$ more precisely) in Eq. (3) is technically difficult due to the strong dependence of these terms on convection depth. Ideally, $\omega$ should be very small near the top of convection and becomes zero above the convection top if tropical convection dominates. Thus, the contribution of $\omega \partial_p h$ (or $\omega \partial_p s$) is also small near the convection top and becomes zero above the convection top even though $\partial_p h$ is large at high altitudes. In other
words, convection height should not be important here. However, \( \omega \) is not zero above the convection top in most CGCM simulations because of the influence of other processes besides tropical convection. This leads to huge errors in calculating \( \langle \omega i \rangle(h) \) (or \( \langle \omega p \rangle(s) \)) without defining convection height. On the other hand, \( \omega p \) is insensitive to convection height since little moisture is found at high altitudes, that is, \( \partial_p \approx 0 \). Thus, we only analyze the contribution from the moisture part, such as \( \langle \omega p \rangle(h) \) and \( \langle \omega p \rangle(q) \) in Eq. (3) and neglect the terms associated with \( \langle \omega i \rangle(s) \). The terms \( \langle \omega i \rangle(h) \) and \( \langle \omega i \rangle(q) \) integrating from 1000 to 30 hPa are shown here as a reference, so the cloud-top effect (Yu et al. 1998) is not included in this calculation. By neglecting the contributions of \( \langle \omega i \rangle(s) \) and \( \langle \omega i \rangle(q) \), the MSE budget is not closed, and the associated uncertainties may create some caveats. For instance, the term \( \langle \omega i \rangle(h) \) is a small difference between two large terms \( \langle \omega p \rangle(h) \) and \( \langle \omega p \rangle(q) \) in the MSE budget could be modified when adding the contribution of \( \langle \omega p \rangle(s) \). Another point regards the anomalous vertical velocity in the MSE budget. Vertical motion in the tropics may not all be related to convection, so the anomalous vertical velocity obtained from the MSE budget could be inconsistent with \( \omega \) found in the CGCM simulations. Here we focus only on some key issues associated with convection and the corresponding circulation, which dominate tropical climate, so the vertically integrated moisture and MSE budgets are still useful here, even without counting the effect of \( \langle \omega p \rangle(s) \). Overall, the analysis of the vertically integrated moisture budget is more reliable than the MSE budget because the moisture budget does not depend on convection depth.

4. The El Niño case

a. Temporal and spatial variations

An El Niño event often occurs over a period of two years and peaks at the end of the first year (year 0) or the beginning of the second year (year 1), so Fig. 2 shows the two-year evolution of the El Niño composite of hemispherical averages over the tropics for departures from climatology (1961–90). The maximum anomalies of SST and tropospheric temperature in the ECHAM5/MPI-OM simulations are much stronger than the observations (Jungclaus et al. 2006), but the ENSO variability is relatively realistic among 20 IPCC AR4 models (van Oldenborgh et al. 2005). The corresponding precipitation anomalies are also reasonable in the ECHAM5/MPI-OM simulations, especially the horseshoe pattern (Dai 2006). It implies that the ENSO variability is not affected by the feature of a double intertropical convergence zone (ITCZ) found in the ECHAM5/MPI-OM simulations (Jungclaus et al. 2006). Both SST and tropospheric temperature anomalies are symmetric with respect to the equator with a slightly stronger amplitude for the Southern Hemisphere anomalies at their respective peak phase. This symmetric feature is similar to observations found in CL07. The maximum SST anomalies occur in January of year 1, while the maximum tropospheric temperature anomalies peak around April–June of year 1, a one-and-a-half-season lag to the maximum SST anomalies. This lagged relationship between SST and tropospheric temperature anomalies has been discussed by several studies (e.g., Kumar and Hoerling 2003; Sobel et al. 2002; Su et al. 2005). The tropical precipitation anomalies (Fig. 2c), on the other hand, show a clear asymmetry between the Northern and Southern Hemispheres, which varies with seasons. In the first half of year 1 around January–May, the tropical precipitation averaged over the Southern Hemisphere increases, but the precipitation averaged over the Northern Hemisphere decreases. The maximum asymmetry occurs around February–April of year 1, a one-season lag to the maximum SST anomalies, which is consistent with the observations discussed in CL07. In the second half of year 1 around June–December, however, the asymmetry reverses: the Southern Hemispherical tropical precipitation decreases, while the Northern Hemispherical tropical precipitation increases. The asymmetry in the second half of the year occurs during both year 0 and year 1. In the observation (CL07), the asymmetry in the second half of year is not as clear as in the ECHAM5/MPI-OM simulations. This might be because both SST and tropospheric temperature anomalies exist longer than the observations. Thus, no clear temporal asymmetry, discussed in CL07, is found in the ECHAM5/MPI-OM simulations. Note that the tropical mean precipitation shows little change during El Niño.

Further examining those anomalies associated with El Niño, the zonal averages of the anomalies are shown in Fig. 3. Both surface and tropospheric temperature anomalies are roughly symmetric to the equator but expand a little farther poleward in the Southern Hemisphere than in the Northern Hemisphere. Note that surface temperature includes both land surface temperature and SST. The meridional gradient of the tropospheric temperature anomalies (Fig. 3b) is smoother than that of the surface temperature. The tropospheric temperature anomalies also persist longer than the surface temperature anomalies (Figs. 3a,b). The maximum surface temperature anomalies are found around February of year 1, while the maximum tropospheric temperature anomalies occur around June of year 1. This
lagged relationship between the surface and tropospheric temperature anomalies is consistent with Figs. 2a and 2b. The zonally averaged tropical precipitation anomalies tend to vary with the seasonal movement of the tropical convection zone (thick line), especially for the positive precipitation anomalies (Fig. 3c). The main positive precipitation anomalies move to the Southern Hemisphere in the boreal winter and spring, but they...
are over the Northern Hemisphere in the boreal summer and autumn. Negative precipitation anomalies are found both north and south of the positive anomalies, with a much stronger amplitude over the Northern Hemisphere. The seasonal variation of the zonal precipitation anomalies implies an asymmetric pattern of the tropical precipitation anomalies on two sides of the equator during El Niño. The maximum asymmetry
occurs in the boreal spring of year 1, with negative anomalies over the Northern Hemisphere and positive anomalies over the Southern Hemisphere. The zonal moisture anomalies also show similar seasonal variation, but the positive anomalies are meridionally wider than the precipitation anomalies and the negative anomalies are much weaker (Fig. 3d). All four variables shown in Fig. 3 are similar to the observations (Fig. 2 in CL07).

To examine the spatial patterns of the anomalies associated with El Niño, the seasons with maximum amplitudes are used: DJF for surface temperature, March–May (MAM) for tropospheric temperature, and February–April (FMA) for precipitation. The positive SST anomalies are found over the equatorial Pacific, with maximum SST anomalies occurring just east of the date line (Fig. 4a). This SST anomaly pattern is similar to observed El Niño SST anomalies. Warm surface temperature anomalies are also found over South America, which might be due to a more downward solar radiation associated with less clouds and precipitation (Fig. 4c). The warm tropospheric temperature anomalies (850–200 hPa) are found over the entire tropics and the pattern of these tropospheric temperature anomalies is also similar to observations (Wallace et al. 1998; Su et al. 2003). The corresponding precipitation anomalies are found mainly over the tropical Pacific. The positive precipitation anomalies are along the equator and the maximum anomalies occur near the date line, a little more westward than the maximum SST anomalies. Negative precipitation anomalies are also found both north and south of the positive precipitation anomalies and over the warm pool region, a well-known horseshoe pattern (Ropelewski and Halpert 1987). Relatively weak negative precipitation anomalies occur over South America, which are associated with less convective clouds and may induce the warm surface temperature anomalies over this area. Overall, El Niño is well simulated by the ECHAM5/MPI-OM CGCM.

b. Diagnosis

To understand effects that cause the asymmetry of the tropical precipitation changes in El Niño, we examine the zonal averages of those terms in moisture and MSE equations in FMA of year 1 when the asymmetry of the tropical precipitation changes is at its maximum strength. Figure 5 shows the zonal averages of the anomalies in the tropics (30°S–30°N). Main positive precipitation anomalies are found between 10°S and 7°N, with maximum precipitation anomalies near 5°S (Fig. 5a). Thus, the positive precipitation anomalies are over the equatorial region where the SST anomalies are
the warmest, but the maximum precipitation anomalies shift southward to the Southern Hemisphere. Two major negative precipitation anomalies are also found over 7°–23°N and 10°–30°S, with minima at 10°N and 15°S, respectively. The amplitude of the minimum precipitation anomalies over the Northern Hemisphere is comparable to the maximum precipitation anomalies, but the amplitude of the minimum anomalies over the Southern Hemisphere is much weaker. The corresponding tropospheric moisture changes also show maximum anomalies over the Southern Hemisphere. Because of the influence of the maximum warm SST anomalies at the equator, the maximum moisture anomalies are closer to the equator than the maximum precipitation anomalies. Negative moisture anomalies are found over the Northern Hemisphere, with minimum anomalies around 14°N. No negative moisture anomalies are found over the Southern Hemisphere.
where negative precipitation anomalies occur (Figs. 5a,b), so the meridional gradient of the tropospheric moisture anomalies is stronger over the Northern Hemisphere than over the Southern Hemisphere.

In the vertically integrated moisture budget, $-\langle \omega \partial_p \tilde{q} \rangle$, which is associated with anomalous vertical velocity, is the most dominant term on the rhs of Eq. (1), contributing more than 70% of zonal precipitation anomalies. The pattern of $-\langle \omega \partial_p \tilde{q} \rangle$ is also very similar to the distribution of zonal precipitation anomalies, further evidence that $-\langle \omega \partial_p \tilde{q} \rangle$ dominates tropical precipitation changes. The term $-\langle \tilde{w} \partial_p q' \rangle$, the direct moisture effect (CN04), contributes about 10% of the positive precipitation anomalies over the Southern Hemisphere but has very little contribution to negative precipitation anomalies. Positive $-\langle \tilde{w} \partial_p q' \rangle$ expands a little more southward, into south of 10°S, where negative precipitation anomalies are found. The horizontal moisture transport $-\langle \mathbf{v} \cdot \nabla q \rangle'$ coincides well with the precipitation anomalies and affects both positive and negative precipitation anomalies by about 10%. Evaporation anomalies are roughly opposite to the distribution of $-\langle \mathbf{v} \cdot \nabla q \rangle'$, so a cancellation between these two effects can be expected. The anomalous evaporation has a maximum around 10°N where the precipitation anomalies are at a minimum. This implies that local evaporation does not cause the precipitation anomalies.

As discussed above, $-\langle \omega \partial_p \tilde{q} \rangle$ is the most dominant term in the moisture budget for tropical precipitation anomalies. Here, $-\langle \omega' \partial_p \tilde{q} \rangle$ is associated with anomalous vertical motion that is related to convection. Figure 6 shows zonally averaged pressure velocity anomalies in FMA of year 1. Maximum anomalous upward motion is found just south of the equator and north of the maximum mean upward motion where the ascending motion of the mean Hadley circulation dominates. Maximum anomalous downward motion, on the other hand, is found around 10°N, the northern margin of the ascending branch of the mean Hadley circulation. This anomalous downward motion reduces tropospheric moisture over the Northern Hemisphere. Weak anomalous downward motion is also found at the southern margin of this Hadley ascending branch. Thus, the Hadley circulation is enhanced but becomes narrower, with both maximum ascending and descending branches shifting toward the equator. This distribution of the anomalous vertical motion is consistent with $-\langle \omega' \partial_p \tilde{q} \rangle$ shown in Fig. 5c, proving that $-\langle \omega' \partial_p \tilde{q} \rangle$ is a good indicator for vertical velocity anomalies.

To examine effects that are associated with this anomalous vertical motion, we analyzed the MSE budget. As discussed in the previous section, we only analyzed the effects of $\langle \tilde{w} \partial_p q' \rangle$ and ignored terms associated with $\langle \tilde{w} \partial_p s \rangle$. However, we did calculate terms related to $\langle \tilde{w} \partial_p h \rangle$ by using a constant depth of convection at 30 hPa, so the cloud-top effect (Yu et al. 1998) was not included in the calculation of $\langle \tilde{w} \partial_p h \rangle$. Thus, terms related to $\langle \tilde{w} \partial_p h \rangle$ shown in Figs. 5c and 5d are just for reference. The effect of $-\langle \tilde{w} \partial_p q' \rangle$, a part of $-\langle \tilde{w} \partial_p h \rangle$, that is associated with the rich-get-richer mechanism (CN04; Chou et al. 2006), is positive over the Southern Hemisphere but negative and small over the Northern Hemisphere. Thus, $-\langle \tilde{w} \partial_p q' \rangle$ has an uneven impact on vertical velocity between the Northern and Southern Hemispheres, creating an asymmetry between two hemispheres. Note that $-\langle \tilde{w} \partial_p q' \rangle$ extends a little more southward and opposes the anomalous downward motion in the Southern Hemisphere. The horizontal MSE transport $-\langle \mathbf{v} \cdot \nabla (q + T) \rangle'$ follows the distribution of the zonally averaged $-\langle \tilde{w} \partial_p \tilde{q} \rangle$ relatively well. Over the Southern Hemisphere, $-\langle \mathbf{v} \cdot \nabla (q + T) \rangle'$ has a maximum at 7°S and a minimum around 17°S, with a similar amplitude compared to other terms on the rhs of the MSE budget Eq. (3). Over the Northern Hemisphere, on the other hand, $-\langle \mathbf{v} \cdot \nabla (q + T) \rangle'$ is the dominant term on the rhs of Eq. (3), with a minimum at 10°N. Its
amplitude is much greater than the negative anomalies in the Southern Hemisphere (around 17°S). The term \( -\langle \mathbf{v} \cdot \nabla (q + T) \rangle \) is dominated by \( -\langle \mathbf{v} \cdot \nabla q \rangle \) (Fig. 5c) because \( -\langle \mathbf{v} \cdot \nabla T \rangle \) is relatively small in the tropics. Therefore, instead of analyzing \( -\langle \mathbf{v} \cdot \nabla (q + T) \rangle \), we further analyzed three components of \( -\langle \mathbf{v} \cdot \nabla q \rangle \), which are \( -\langle \mathbf{v} \cdot \nabla q \rangle \), \( -\langle \mathbf{v} \cdot \nabla q \rangle \), and the nonlinear term \( -\langle \mathbf{v} \cdot \nabla q \rangle - \langle \mathbf{v} \cdot \nabla q \rangle \). The term \( -\langle \mathbf{v} \cdot \nabla q \rangle \) has the largest contribution (Fig. 7), while the other two terms \( -\langle \mathbf{v} \cdot \nabla q \rangle \) and \( -\langle \mathbf{v} \cdot \nabla q \rangle - \langle \mathbf{v} \cdot \nabla q \rangle \) tend to cancel each other out. The effect of \( -\langle \mathbf{v} \cdot \nabla q \rangle \) has been referred to as the upward-ante mechanism (NCS03; CN04; Chou et al. 2006). Two meridional inflows at the lower troposphere transport dry air from the descending regions to convective regions, so negative \( -\langle \mathbf{v} \cdot \nabla q \rangle \) is found over the northern and southern margins of convective regions. Negative \( -\langle \mathbf{v} \cdot \nabla q \rangle \) is larger over the Northern Hemisphere than over the Southern Hemisphere because of the stronger inflow associated with the Hadley circulation (Fig. 6) and the larger meridional gradient of the moisture anomalies in the Northern Hemisphere (Fig. 5b). Because the maximum moisture anomalies lean more toward the equator than the maximum mean convergence (Figs. 5c,d), positive \( -\langle \mathbf{v} \cdot \nabla q \rangle \) is found around 7°S. The last term on the rhs of Eq. (3), the net heat flux into the atmosphere \( F_{\text{net}} \), is similar to the net surface heat flux \( F_{s} \) because \( F_{s} \) is relatively small (not shown). Thus, \( F_{\text{net}} \) is similar to the distribution of the SST anomalies, which is roughly symmetric to the equator, with a maximum at the equator. The asymmetry of the tropical precipitation anomalies cannot be directly associated with the symmetric \( F_{\text{net}} \). However, the positive \( F_{\text{net}} \) is important because \( F_{\text{net}} \) initiates the whole process that leads to the tropical precipitation asymmetry (CL07). Overall, the positive precipitation anomalies near the equator are associated with \( F_{\text{net}} \), while the effects of \( -\langle \mathbf{v} \cdot \nabla q \rangle \) and \( -\langle \mathbf{v} \cdot \nabla q \rangle \) shift the maximum precipitation anomaly southward to the Southern Hemisphere and \( -\langle \mathbf{v} \cdot \nabla q \rangle \) reduces tropical precipitation over margins of the Hadley ascending area. This creates the asymmetry of the tropical precipitation anomalies between the Northern and Southern Hemispheres. Both effects of \( -\langle \mathbf{w} \partial_{t} q \rangle \) and \( -\langle \mathbf{v} \cdot \nabla q \rangle \) are associated with the mean Hadley circulation.

5. The global warming case

a. Temporal and spatial variations

Figure 8 shows tropical averages of surface temperature, tropospheric temperature (850–200 hPa), and precipitation changes for the trend component between 2071–2100 and 1961–90. The 2-yr period, repeating one more year, is for comparison to the El Niño case, which is also shown in a 2-yr period—its growing to decaying phases. The surface temperature anomalies are roughly symmetric to the equator and its seasonal variation is very weak (Fig. 8a). The Northern Hemispherical average is slightly warmer than the Southern Hemispherical average. This might be a result of more land over the Northern Hemisphere than the Southern Hemisphere. The tropospheric temperature anomalies are also symmetric to the equator and the seasonal variation is weak too. The difference in the tropospheric temperature anomalies between the Northern and Southern Hemispheres is even smaller than that of the surface temperature anomalies. Note that the amplitude of the surface temperature anomalies (≈3°C) is smaller than that of the tropospheric temperature anomalies (≈4.5°C). Unlike the surface and tropospheric temperature anomalies, the tropical precipitation anomalies show strong asymmetry between the Northern and Southern Hemispheres. The mean precipitation anomalies averaged over the entire tropics is 0.2 mm day\(^{-1}\). The Northern Hemispherical average reaches its maximum 0.6 mm day\(^{-1}\) in the boreal summer [July–September (JAS)] and decreases to its minimum −0.2 mm day\(^{-1}\) in the boreal winter [January–March (JFM)], and vice versa for the Southern Hemispherical average. In other words, the hemispherical averages are a departure from the tropical average by approximately ±0.4 mm day\(^{-1}\). This asymmetry of the hemispherically averaged tropical precipitation anomalies has a distinct seasonal cycle, which is consistent with the movement of tropical convection zones, while the tropical average does not show any such seasonal variation.

We further examine the zonal averages of the anomalies induced by global warming, the results of which are shown in Fig. 9. Both surface and tropospheric temperature anomalies are positive over the en-
tire tropics. The zonal surface temperature anomalies over the Southern Hemisphere are relatively weaker and have a stronger meridional gradient than the anomalies over the Northern Hemisphere. The tropospheric temperature anomalies, on the other hand, are roughly similar over both hemispheres, with maximum anomalies occurring around 10°S in June. The surface and tropospheric temperature anomalies do not show a clear meridional movement. However, the precipitation and tropospheric moisture anomalies have a very ap-
parent meridional movement. The positive precipitation anomalies follow the seasonal movement of mean convection. The corresponding negative precipitation anomalies are found outside the positive precipitation anomaly area. In JAS, the positive precipitation anomalies are over the Northern Hemisphere and most negative precipitation anomalies are over the Southern Hemisphere. In JFM, on the other hand, the positive anomaly area.
anomalies are over the Southern Hemisphere and most negative anomalies are over the Northern Hemisphere. The positive precipitation anomalies, with a maximum amplitude of around 1.6 mm day$^{-1}$, are much stronger than the negative anomalies, with a maximum amplitude of around $-0.4$ mm day$^{-1}$. The tropospheric moisture anomalies have a similar meridional movement to the precipitation anomalies, but their values are all positive over the entire tropics ($30^\circ$S–$30^\circ$N). Its amplitude, with a maximum amplitude $1.6$ g kg$^{-1}$, is much greater than the positive anomalies in the El Niño case (Fig. 3d), which has a maximum at $0.6$ g kg$^{-1}$.

What do the results shown in Figs. 8 and 9 imply when considering trends for hemispherical averages, especially for the tropical precipitation anomalies? The hemispherical asymmetry of the tropical precipitation anomalies implies that different trends of the tropical precipitation changes exist between the Northern and Southern Hemispheres and also between different seasons, but that such different trends are not found in both temperature anomalies. In the tropics, the trends of the surface and tropospheric temperature anomalies are similar between the Northern and Southern Hemispheres and between JAS and JFM (Fig. 10). The Northern Hemispherical surface temperature anomalies increase slightly more quickly than those of the Southern Hemisphere, while the tropospheric temperature anomalies are roughly the same for both hemispherical averages. Moreover, these temperature trends do not show any clear seasonal variation: both trends in JAS and JFM are similar. These are consistent with the results shown in Fig. 8. On the contrary, the precipita-
tion anomaly trend is quite different between hemispheres and seasons. In JAS, the Northern Hemispheric average in the tropics increases significantly, while the Southern Hemispheric average decreases only slightly. In JFM, the trends over the Northern and Southern Hemispheres are reversed. Thus, Figs. 10e and 10f imply that the hemispherical average of tropical precipitation becomes larger in a wet (summer) season and smaller in a dry (winter) season under global warming. In other words, the range of tropical precipitation between wet and dry seasons becomes larger in a warmer climate. The rate of the decreasing trend is slower than the rate of the increasing trend, because of an increasing trend of the tropically averaged precipitation. The results shown in Fig. 10 also imply that the JAS and JFM precipitation difference between the Northern and Southern Hemispheres is widened when the atmosphere becomes warmer. A stronger seasonal variation, as well as a widened hemispheric precipitation difference, is also found in other CGCM simulations (Chou et al. 2007).

Figure 11 shows the spatial distribution of the surface temperature, tropospheric temperature, and precipitation anomalies in JAS and JFM. The tropospheric temperature anomalies are roughly symmetric to the equator in both seasons, with the maximum anomalies occurring at lower latitudes. The surface temperature anomalies are warmer over land than over ocean. A weak El Niño–like SST anomaly pattern over the eastern Pacific is found in both seasons, with stronger amplitudes in JFM. A Rossby wave response to those El Niño–like SST anomalies is found in the JFM tropospheric temperature anomalies but is unclear in the JAS anomalies. Most precipitation changes occur near the equator (Figs. 11e,f). The amplitude of the positive precipitation anomalies is larger than that of the negative anomalies. Most positive tropical precipitation anomalies are over the Northern Hemisphere in JAS and over the Southern Hemisphere in JFM. This is consistent with the previous analysis shown in Figs. 8c and 9c. The typical El Niño–like response in precipitation anomalies, such as positive anomalies over the equatorial central Pacific and a horseshoe pattern of negative precipitation anomalies, is not clear.

b. Diagnosis

Positive zonal precipitation anomalies are found over the Northern Hemisphere in JAS and over the Southern Hemisphere in JFM (Figs. 12a, 13a), with maxima around 10°N and 10°S, respectively. Relatively weak but meridionally broad negative precipitation anomalies are found on the other side of the equator, the Southern Hemisphere in JAS and the Northern Hemisphere in JFM. No clear minimum precipitation anomalies are found in both JAS and JFM. The zonal moisture anomalies also only have maxima around 10°N and 10°S, without clear minima. The meridional gradients on both sides of the maximum moisture anomalies are similar in magnitude. Unlike the moisture anomalies in the El Niño case (Fig. 5b), all moisture anomalies in JAS and JFM are positive. Analyzing the vertically integrated moisture budget, \( -\langle \omega_p q' \rangle \) is the dominant effect contributing to the zonal tropical precipitation anomalies, more than 60% at respective maximum precipitation anomalies in JAS and JFM. Their meridional distribution is roughly similar to the corresponding precipitation anomalies, especially the positive anomalies. This effect of \( -\langle \omega_p q' \rangle \) is associated with the direct moisture effect (CN04). Because of the increase of low-level moisture due to global warming (CN04; Held and Soden 2006), the ascending motion of the mean Hadley circulation transports more moisture vertically and increases tropical precipitation over the ascending branch of the Hadley circulation. On the other hand, the descending motion transports relatively drier air downward and reduces precipitation over the descending branch. Thus, positive \( -\langle \omega_p q' \rangle \) is over the Northern Hemisphere and negative \( -\langle \omega_p q' \rangle \) is over the Southern Hemisphere in JAS, and vice versa in JFM. Note that the positive \( -\langle \omega_p q' \rangle \) is greater than negative \( -\langle \omega_p q \rangle \) in amplitude. The effect of \( -\langle \omega_p q \rangle \), which is associated with anomalous vertical motion, becomes secondary and contributes only about 30% of the maximum precipitation anomalies. The maximum \( -\langle \omega_p q \rangle \) is over the same hemisphere as the maximum precipitation anomalies, but the meridional distribution of \( -\langle \omega_p q \rangle \) is not similar to the precipitation anomaly distribution shown in Figs. 12a and 13a. To the north and south of the positive maximum \( -\langle \omega_p q \rangle \), negative anomalies are found. The positive \( -\langle \omega_p q \rangle \) is just slightly larger than the negative anomalies in amplitude. Small negative \( -\langle \omega_p q \rangle \) is also found over the winter hemisphere, 15°–25°S in JAS and 15°–25°N in JFM. On the hemispheric scale, \( -\langle \omega_p q \rangle \) does not show a clear contribution to the asymmetry of tropical precipitation anomalies because of a cancellation between positive and negative anomalies. The horizontal moisture transport \( -\langle v \cdot \nabla q' \rangle \) shows small positive anomalies over the areas with maximum precipitation anomalies, 5°N in JAS and 5°S in JFM, but relatively large negative anomalies to the north and south of the maximum \( -\langle v \cdot \nabla q' \rangle \), especially in JFM. Positive evaporation anomalies are over the entire tropics, with an amplitude of less than 10 W m\(^{-2}\). The zonal evaporation anomalies do not vary with latitude too much.
FIG. 11. As in Fig. 4, but for the global warming case: (a), (c), (e) JAS and (b), (d), (f) JFM. In (c) and (d), the scale of the y axis on the left is for the solid line and the one on the right is for the dashed line.
Even though the effect of $-\langle \omega' \partial_z f \rangle$ is secondary, the association of vertical velocity changes with the Hadley circulation is still important. Figure 14 shows anomalous vertical motion in JAS and JFM. We first examine vertical velocity anomalies within $10^\circ N$–$10^\circ S$. In JAS, anomalous upward motion is over the Northern Hemisphere, and anomalous downward motion is over the Southern Hemisphere. In JFM, the vertical velocity anomalies reverse: anomalous downward motion over the Northern Hemisphere and anomalous upward motion over the Southern Hemisphere. These vertical velocity anomalies imply anomalous circulations on the pressure–latitude domain. These anomalous circulations in JAS and JFM are in the same direction as the respective mean Hadley circulation, but only on the inner side (near the equator) of the Hadley circulation. The anomalous vertical motion within $10^\circ S$–$10^\circ N$ shows two local maxima: below 800 hPa and above 300 hPa. At the lower troposphere, the anomalous vertical motion implies a cyclonic circulation in JAS and an anticyclonic circulation in JFM on the pressure–latitude domain. A similar anomalous circulation is also found at the upper troposphere, implying a vertically deepened Hadley circulation associated with an increase of tropical convection height. On the outer side of the Hadley circulation in $10^\circ$–$25^\circ N$ and $10^\circ$–$25^\circ S$, anom-
lous vertical motion, which opposes the mean Hadley circulation, is also found. In JAS, anomalous upward motion is in 10°–25°S and anomalous downward motion is in 10°–25°N. In JFM, anomalous downward motion is in 10°–25°S and anomalous upward motion is in 10°–25°N. Overall, Fig. 14 indicates a shrinking Hadley circulation under global warming. This meridional distribution of anomalous vertical motion is consistent with the anomalous vertical motion indicated by $\langle \omega' \partial_p \bar{q} \rangle$ in Figs. 12c and 13c. Thus, $\langle \omega' \partial_p \bar{q} \rangle$ can be used to represent vertical velocity anomalies in the global warming case. These complicated vertical motion changes imply a potential difficulty in determining the strength of the Hadley circulation in the future climate.

To understand how vertical motion associated with convection changes in a warmer climate, the MSE budget is examined. Because of neglecting the effect of $\langle \omega \partial_p \mathcal{L} \rangle$, the MSE budget analysis discussed here can only give us a rough and qualitative idea about effects that influence the vertical motion. However, we did calculate terms related to $\langle \omega \partial_p \mathcal{F} \rangle$, with a constant depth of convection (Figs. 12c,d, 13c,d). Because those terms related to $\langle \omega \partial_p \mathcal{F} \rangle$ do not include the cloud-top effect, we only use them as a reference. The maximum $\langle \omega \partial_p q' \rangle$ is in the same hemisphere as the maximum
This implies that $-(\omega \partial_p q')$, a part of the rich-get-richer mechanism (CN04), should have some contribution to the anomalous upward motion. On the opposite side of the equator, $-(\omega \partial_p q')$ tends to be small and negative, so the effect of $-(\omega \partial_p q')$ is negligible. The distribution of $-(v \cdot \nabla (T + q)')$ is similar to $-(v \cdot \nabla q')$ but a little stronger. This indicates that the effect of $-(v \cdot \nabla q')$ is relatively weak compared with $-(v \cdot \nabla q')$, so $-(v \cdot \nabla q')$ can be used to represent the effect of the horizontal MSE transport $-(v \cdot \nabla (T + q)')$. Compared with the three components of $-(v \cdot \nabla q')$, $-(v \cdot \nabla q')$ is the dominant effect (Fig. 15), which is associated with the upped-ante mechanism (CN04; NCS03). The low-level meridional winds transport relatively dry air from the descending regions to the margins of the ascending area. This dry advection suppresses convection over these margins. The effect of $-(v \cdot \nabla q')$ also produces relatively small positive anomalies over areas with positive anomalous ascending motion, especially in JFM. The term $F_{\text{net}}$ in the El Niño case is also more concentrated near the equator, with the maximum more than 10 W m$^{-2}$, but the global warming $F_{\text{net}}$ is broader and weaker, with the maximum around 5 W m$^{-2}$. The global warming SST anomalies do have an El Niño–like component, but with relatively weaker amplitude, compared not only with the El Niño SST anomalies but also with their corresponding global component. Even though the El Niño–like component in SST anomalies is weak, an El Niño–induced tropospheric temperature anomaly pat-
tern is clearly found in JFM, with a relatively smooth warm tropospheric temperature anomaly pattern in the background (Fig. 11b). The tropospheric temperature anomalies in the El Niño case have a much broader pattern than the associated SST anomalies, but they are still weaker and spatially smaller than those anomalies in the global warming case. Thus, the tropical precipitation anomalies in the El Niño case are induced by the forcings mainly in the tropics, while the tropical precipitation anomalies in the global warming case could be induced by the forcings not only from the tropics but also from higher latitudes.

The asymmetry of tropical precipitation anomalies is found in both El Niño and global warming cases. Compared with the precipitation anomalies in the boreal winter (JFM), when the precipitation asymmetry is at maximum in both El Niño and global warming cases, the amplitudes of positive and negative anomalies are more comparable in the El Niño case than in the global warming case. This is because the precipitation changes in the global warming case have an increasing trend for the entire tropical average (30°S–30°N). Unlike the hemispherical average, this tropically averaged change does not vary with season (Figs. 10c,f). Because the tropical-mean precipitation is enhanced under global warming, for departures from this tropical average of precipitation anomalies, the magnitudes of hemispherical averages for positive and negative precipitation anomalies become more similar (Chou et al. 2007). In a zonal average, both positive and negative precipitation

Fig. 15. As in Fig. 7, but for the global warming case: (a) JAS and (b) JFM.
anomalies in the global warming case expand broader meridionally, especially for negative anomalies, while the precipitation anomalies in the El Niño case are more concentrated near the equator because of the maximum SST anomalies at the equator. For the negative precipitation anomalies, the El Niño case has a minimum over the Northern Hemisphere in JFM, but the global warming case does not have a clear minimum over the same hemisphere. The tropospheric moisture anomalies induced by global warming are all positive and their amplitude is much stronger than those anomalies associated with El Niño. The El Niño case has a stronger meridional gradient of the moisture anomalies on the Northern Hemisphere than the Southern Hemisphere in JFM. This leads to an uneven distribution of \(-\langle v \cdot \nabla q \rangle\)' in the El Niño case. This uneven distribution of \(-\langle v \cdot \nabla q \rangle\)' is not clear in JFM of the global warming case.

The main effects that are associated with the precipitation asymmetry for El Niño and global warming are different. In the El Niño case, the dynamical feedback associated with anomalous vertical motion via \(-\langle \omega_\cdot \partial_q \rangle\) dominates. In the global warming case, however, the direct moisture effect \(-\langle \omega_\cdot \partial_q \rangle\), which is directly linked to the mean Hadley circulation, dominates. During El Niño, the warm SST anomalies enhance convection and create weak tropospheric temperature anomalies. The weak tropospheric temperature anomalies enhance the low-level moisture, causing the precipitation anomalies to become asymmetric because of the dynamical feedback of \(-\langle \omega_\cdot \partial_q \rangle\). Because the positive moisture anomalies are small, the direct moisture effect via \(-\langle \omega_\cdot \partial_q \rangle\) in the moisture budget is also relatively small. Under global warming, on the other hand, the tropospheric warming is much stronger and has a broader spatial scale, so the moisture increases are much more than in the El Niño case. Thus, zonal tropical precipitation changes under global warming is mainly associated with the direct moisture effect \(-\langle \omega_\cdot \partial_q \rangle\).

As discussed in the previous section, \(-\langle \omega_\cdot \partial_q \rangle\) can also affect tropical precipitation via the dynamical feedback associated with anomalous vertical motion. The term \(-\langle \omega_\cdot \partial_q \rangle\) is larger in the global warming case than in the El Niño case, so stronger anomalous vertical motion should be expected under global warming if there is no other effect. However, the anomalous vertical motion in global warming (Fig. 14) actually has only half the amplitude of the anomalous vertical motion in El Niño (Fig. 6). This implies that other effects, such as \(-\langle \omega_\cdot \partial_q \rangle\), must cancel the effect of \(-\langle \omega_\cdot \partial_q \rangle\) on vertical motion. In this study, we neglected the contribution of \(-\langle \omega_\cdot \partial_q \rangle\) in our MSE budget analysis because \(-\langle \omega_\cdot \partial_q \rangle\) strongly depends on convection height. Even though we did calculate \(-\langle \omega_\cdot h \rangle\) as a reference, we still omitted the cloud-top effect (Yu et al. 1998) in this calculation. Thus, \(-\langle \omega_\cdot h \rangle\) shown in Figs. 5, 12, and 13 cannot give us too much insight related to the effect associated with the changes of convection height. Under global warming, tropopause tends to be higher (Holzer and Boer 2001; Santer et al. 2003; International Ad Hoc Detection and Attribution Group 2005), so tropical convection could be deepened. Thus, \(-\langle \omega_\cdot s \rangle\) becomes larger and can compensate the stronger effect of \(-\langle \omega_\cdot q \rangle\) in the global warming case. Note that \(-\langle \omega_\cdot h \rangle\) in the El Niño case (Fig. 5d) does show the similar amplitude to those in the global warming case (Figs. 12d, 13d). More detailed analysis should be done before making a definite conclusion regarding the influence of \(-\langle \omega_\cdot q \rangle\). Besides the amplitude, the vertical and meridional distributions of vertical velocity anomalies are also different between these two cases. The global warming case has clear maximum vertical velocity anomalies at the upper troposphere, while the El Niño case does not have such local maxima. The anomalous upward motion found between 10° and 25°N in JFM for the global warming case (Fig. 14b) is not clear for the El Niño case. This implies that possible extratropical forcings may have an influence on the tropical precipitation. Overall, the current MSE budget analysis can only give us a qualitative idea about those effects that are associated with anomalous vertical motion, especially for the effect associated with \(-\langle \omega_\cdot q \rangle\). Thus, caution must be taken when using \(-\langle \omega_\cdot q \rangle\) to explain the effect of \(-\langle \omega_\cdot h \rangle\).

In the previous studies (CN04; Chou et al. 2006), the “regional” tropical precipitation changes under global warming were dominated by the dynamical feedback of \(-\langle \omega_\cdot q \rangle\) and \(-\langle v \cdot \nabla q \rangle\) via \(-\langle \omega_\cdot \partial_q \rangle\), which are associated with the rich-get-richer mechanism and the upped-ante mechanism, respectively. This study, on the other hand, shows that the direct moisture effect \(-\langle \omega_\cdot q \rangle\) is more important to zonal tropical precipitation changes via the Hadley circulation. This has also been discussed by Held and Soden (2006). In Chou et al. (2006), we focused on the regional precipitation changes over convective regions. In this study, the focus is on both convective regions and subsidence regions that are associated with the Hadley circulation. Over convective regions, which are discussed by both studies, most \(-\langle \omega_\cdot \partial_q \rangle\) are positive over convective centers but negative over convective margins. Thus, the amplitude of \(-\langle \omega_\cdot \partial_q \rangle\) averaged over convective regions becomes smaller because of the cancellation between convective centers and convective margins. On the other hand, \(-\langle \omega_\cdot \partial_q \rangle\) is positive everywhere in convective regions,
where $\overline{\omega}$ is negative, that is, upward motion. Therefore, the zonal average of $-\langle \overline{\omega} q' \rangle$ becomes more dominant than the zonal average of $-\langle \omega' \partial_y q' \rangle$. The direct moisture effect involves the mean Hadley circulation and zonal moisture changes, which are usually well simulated by most climate models, so the direct moisture effect should be more robust among climate model simulations. On the other hand, the dynamical feedback associated with anomalous vertical motion can be very different between climate models (Tan et al. 2007, manuscript submitted to J. Climate). Thus, partitioning between the dynamical feedback via $-\langle \omega' \partial_y q' \rangle$ and the direct moisture effect $-\langle \overline{\omega} q' \rangle$ may be model dependent. Nevertheless, the direct moisture effect does tend to become more important to tropical precipitation changes on the zonally averaged basis.

7. Conclusions

In this study, we aim to understand how zonal tropical precipitation and the associated circulation react to a uniform forcing, such as global warming and the El Niño–induced tropical warming. The ECHAM5/MPI-OM CGCM simulations under the IPCC scenario A2 are used. In both El Niño and global warming cases, an asymmetry of tropical precipitation changes between the Northern and Southern Hemispheres is found. This asymmetry varies with season, negative anomalies over the Northern Hemisphere and positive anomalies over the Southern Hemisphere in the first half of the year and vice versa in the second half of the year. The precipitation asymmetry is slightly different between the El Niño and global warming cases. In El Niño, the amplitudes of positive and negative hemispherically averaged precipitation anomalies are similar, with the maximum around $\pm 0.2$ mm day$^{-1}$. The tropical average of the precipitation anomalies is near zero. In global warming, the amplitude of positive anomalies is larger than that of the negative anomalies, with the maximum at $+0.6$ and $-0.2$ mm day$^{-1}$, respectively. This uneven amplitude between positive and negative anomalies is due to a nonzero tropical average of the precipitation anomalies, which is $0.2$ mm day$^{-1}$. This hemispherical asymmetry in both cases is consistent with the meridional movement of zonal precipitation anomalies, which is associated with the seasonal migration of the Hadley circulation.

The seasonal variation of the precipitation asymmetry in global warming suggests a different view for detecting global warming signals in precipitation changes. First, the tropical precipitation tends to increase in the summer hemisphere and decrease in the winter hemisphere. The summer hemisphere is associated with a wet season and the winter hemisphere is associated with a dry season for most places in the world. Thus, the wet season becomes wetter and the dry season becomes drier. In other words, the seasonal precipitation range, a difference between wet and dry seasons, is widened in a warmer climate. Second, the precipitation difference between the Northern and Southern Hemispheres is enhanced. This enhancement of the hemispherical difference in tropical precipitation is not necessarily associated with the change of the Hadley circulation.

Dominant effects associated with the tropical precipitation asymmetry are different between the El Niño and global warming cases. In the El Niño case, warm SST anomalies over the equatorial eastern Pacific initiate the whole process. The asymmetry of the tropical precipitation anomalies is associated with the dynamical feedback via anomalous vertical motion. This dynamical feedback includes the effect associated with vertical moisture transport by mean circulation and the effect of horizontal moisture advection. These two major effects are associated with the distribution of low-level moisture anomalies and the mean Hadley circulation. Warmer SST and troposphere enhance low-level moisture and create a meridional moisture gradient from the descending branch to the ascending branch of the Hadley circulation. Over the ascending branch, convection is enhanced because of less effective static stability ($M$), so the Hadley circulation is strengthened, following which the meridional moisture gradient is further enhanced. On the background of the meridional moisture gradient, the meridional winds at the lower troposphere associated with the Hadley circulation transport dry air from the descending area to the ascending area, which suppresses convection over the margins of the ascending area (Fig. 7). Besides the effect associated with less effective static stability, positive horizontal moisture advection mainly associated with $-\mathbf{v} \cdot \nabla q'$ is also found over the ascending area and enhances convection over this area (Fig. 7). Unlike El Niño, the dominant effect for the precipitation asymmetry in global warming is much more direct. Low-level moisture is enhanced in a warmer climate, so the ascending motion of the mean Hadley circulation can transport more water vapor vertically and enhances tropical precipitation over the ascending branch of the Hadley circulation. Over the descending branch, the downward motion associated with the Hadley circulation transports relatively drier air, so the precipitation is suppressed. This effect over both ascending and descending branches is termed as the direct moisture effect.

The dynamical feedbacks that dominate in the El Niño case are also found in the global warming case.
because the meridional moisture gradient also occurs in the global warming case under QE mediation (CN04). However, the anomalous vertical motion associated with those effects is smaller than in the El Niño case. Moreover, the contribution of the anomalous vertical motion to the precipitation asymmetry in the global warming case is weaker than in the El Niño case because of cancellation in the zonally averaged vertical moisture transport that is associated with the anomalous vertical motion. The zonally averaged anomalous vertical motion in the global warming case is much more complicated than in the El Niño case. This implies that other effects should also affect the vertical motion in the tropics. Further examination of mechanisms that induce such complicated anomalous circulation is subject to ongoing work.

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