Depth of convection and the weakening of tropical circulation in global warming

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Abstract

Anthropogenic forcings, such as greenhouse gases and aerosols, are starting to show their influence on the climate, as evidenced by a global warming trend observed in the past century. The weakening of tropical circulation, a consequence of global warming, has also been found in observations and in 21st-century climate model simulations. It is a common belief that this weakening of tropical circulation is associated with the fact that global-mean precipitation increases more slowly than water vapor. Here we propose a new mechanism for this robust change, which is determined by atmospheric stability associated with depth of convection. Convection tends to extend higher in a warmer climate, due to an uplifting of the tropopause. The higher the convection, the more stable the atmosphere. This leads to a weakening of tropical circulation.

1. Introduction

Under global warming, the global-mean precipitation of general circulation model (GCM) simulations tends to increase at a rate of around $2\% \text{ K}^{-1}$, while the atmospheric water vapor increase roughly follows the Clausius-Clapeyron expansion of saturated water vapor with a constant relative humidity, which is around $7.5\% \text{ K}^{-1}$ (Allen and Ingram 2002; Held and Soden 2006; Stephens and Ellis 2008; Trenberth et al. 2003; Vecchi and Soden 2007; Wentz et al., 2007). Based on a simplified equation, $P = M_c q_s$, where P is precipitation, M_c is convective mass flux and q_s is a typical boundary layer mixing ratio, the fractional change of global-mean precipitation is roughly equal to the sum of the changes in convective mass flux and the boundary layer mixing ratio, i.e., $P'/\bar{P} = M'_c/\bar{M}_c + q'_s/\bar{q}_s$, where $(\bar{})$ is climatology and $()'$ is the change. Since the water vapor increases much faster than the global-mean precipitation, tropical circulation should be weakened in the future warmer climate (Held and Soden 2006; Knutson and Manabe 1995; Vecchi and Soden 2007), i.e., $M'_c/\bar{M}_c < 0$. The weakening of tropical circulation is very robust among the 21st-century GCM simulations and occurs most in a longitudinal direction (Held and Soden 2006; Vecchi and Soden 2007). A weakening of the Walker circulation has already been observed (Vecchi et al. 2006).

What kind of mechanisms could cause this weakening of tropical circulation? Two possible mechanisms have been proposed by Chou et al. (2009, hereafter CHOU09): the upped-ante mechanism and the effect of deepened convection. The upped-ante mechanism is associated with a dry advection from subsidence regions to convective regions. This mechanism usually occurs over margins of convective regions. The effect of deepened convection, on the other hand, increases the gross moist stability and stabilizes the atmosphere. In this study, we would like to further examine the effect of deepened convection, which is commonly found under global warming. The model and data used here is briefly described in section 2. A vertically-integrated water vapor budget used to determine possible contributions to precipitation changes is discussed in section 3.

The effect of deepened convection has been examined by using an intermediate climate system in section 4, followed by discussion and summary.

2. The model and data

a. The quasi-equilibrium tropical circulation model

To examine the effect of atmospheric stability associated with convection depth on tropical circulation, a coupled ocean-atmosphere-land model of intermediate complexity (Neelin and Zeng 2000; Zeng et al. 2000) with prescribed divergence of ocean heat transport (Q-flux) is used. Based on the analytical solutions derived from the Betts-Miller moist convective adjustment scheme (Betts and Miller 1993), typical vertical structures of temperature, moisture and winds for deep convection are used as leading basis functions for a Galerkin expansion (Neelin and Yu 1994; Yu and Neelin 1994). The atmospheric model constrains the flow by quasi-equilibrium thermodynamic closures and is referred to as QTCM1 (quasi-equilibrium tropical circulation model with a single vertical structure of temperature and moisture for deep convection). Because the basis functions are based on vertical structures associated with convective regions, these regions are expected to be well represented and similar to a general circulation model (GCM) with the Betts-Miller moist convective adjustment scheme. Far from convective regions, QTCM1 is a highly truncated Galerkin representation equivalent to a two-layer model. A cloud-radiation scheme (Chou and Neelin 1996; Zeng et al. 2000), simplified from the full radiation schemes (Fu and Liou 1993; Harshvardhan et al. 1987), is included. In this scheme, the deep and cirrocumulus/cirrostratus cloud fraction is estimated by an empirical parameterization (Chou and Neelin 1999).

An intermediate land-surface model (Zeng et al. 2000) is used to simulate the interaction between the atmosphere and land-surface. This model simulates processes such as evapotranspiration and surface hydrology in a single land surface layer for calculating energy and water

budgets. Soil moisture is balanced by precipitation, evaporation, surface run-off and ground run-off. This model does not include snow and sea-ice feedback, so the surface warming at higher latitudes is weaker than in most coupled GCM simulations. A slab mixed-layer ocean model with a fixed mixed-layer depth of 50m is used. The Q-flux is obtained from a prescribed seasonal sea surface temperature (SST) run. In this study, a revised QTCM1 version 2.3 is used. The main improvement in the model physics is the inclusion of a simple ABL which assumes a steady state, vertically homogeneous mixed layer with fixed height (Stevens et al. 2002).

b. Vertical structure of vertical velocity

Under quasi-equilibrium convective closures, a typical vertical profile of deep convection is prescribed in QTCM1, so vertical velocity can be written as

$$\omega(x, y, p, t) = -\Omega(p) \nabla \cdot \mathbf{v}_1(x, y, t), \quad (1)$$

where ω is vertical velocity and $\nabla \cdot \mathbf{v}_1$ is divergence induced by baroclinic winds. The vertical profile of vertical velocity $\Omega(p)$ is independent of time and horizontal position, but depends on convection depth (Neelin and Yu 1994; Yu et al. 1998).

With the prescribed vertical profile of vertical velocity, the vertically integrated vertical advection of moisture and dry static energy can be written as

$$\langle \omega \partial_p q \rangle = -M_q \nabla \cdot \mathbf{v}_1 \quad \text{and} \quad (2)$$

$$\langle \omega \partial_p s \rangle = M_s \nabla \cdot \mathbf{v}_1, \quad (3)$$

where q is specific humidity and the dry static energy is $s = T + \phi$, with ϕ being the geopotential. Both moisture q and temperature T are in energy units by absorbing the latent heat per unit mass L and the heat capacity at constant pressure C_p , respectively. $\langle \rangle$ denotes a mass integration through the entire troposphere. M_s is dry static stability:

$$M_s = -\langle \Omega \partial_p s \rangle, \quad (4)$$

and M_q is gross moisture stratification:

$$M_q = \langle \Omega \partial_p q \rangle. \quad (5)$$

Thus,

$$\langle \omega \partial_p h \rangle = M \nabla \cdot \mathbf{v}_1, \quad (6)$$

where moist static energy (MSE) is $h = s + q$ and the gross moist stability is $M = M_s - M_q$ (Yu et al. 1998).

c. Experiment design

A set of experiments are designed to study the effect of convection depth. In the control experiment, the current climate with the normal CO_2 concentration, i.e., $1 \times \text{CO}_2$, and seasonal Q-flux, obtained from a fixed SST run with observed seasonal climatology, is used. In the standard global warming experiment (M_s+0), the same Q-flux and a doubled CO_2 concentration, i.e., $2 \times \text{CO}_2$, are used. The convection height in these two experiments is fixed at 150 hPa. The differences between these two experiments are induced by the increase of the anthropogenic greenhouse gas, CO_2 . Because convection height is determined beforehand, such as in (1), and the associated effect is pre-calculated in QTCM1, so convection depth can be much more easily changed than most coupled GCMs. In this study, four experiments with different depths of convection are conducted with the same setting as that of the standard global warming experiment (M_s+0), i.e., $2 \times \text{CO}_2$ and the same Q-flux. Changes in convection height affect $\langle \omega \partial_p s \rangle$, i.e., M_s in QTCM1 (see (3)), but have much less of an impact on $\langle \omega \partial_p q \rangle$, due to little water vapor at upper troposphere. In other words, changes in M_s are roughly equivalent to changes in convection height. Thus, we varied M_s

in those four experiments, instead of directly changing convection height. The convection heights for the experiment M_s-1 , M_s+1 , M_s+2 and M_s+3 are fixed roughly at 155 hPa, 145 hPa, 141 hPa and 137 hPa, respectively, so one (M_s-1) is shallower than in the standard global warming experiment (M_s+0) and three (M_s+1 , M_s+2 and M_s+3) are deeper than in the standard run. All experiments are 50-yr runs, but averaged only over the last 40 years. Figure 1 shows precipitation anomalies for the experiments M_s-1 , M_s+0 and M_s+1 minus the control experiment. The patterns are similar, but the magnitudes of precipitation anomalies tend to be smaller as convection becomes deeper. The precipitation anomalies are further reduced in the other two experiments, M_s+2 and M_s+3 , but with similar spatial patterns (not shown). Overall, all five experiments show reasonable changes induced by global warming.

d. Data

The 21st-century GCM simulations from the World Climate Research Programme's (WCRP's) Coupled Model Intercomparison Project phase 3 (CMIP3) multi-model dataset (Table 1) in the A1B scenario are used here. As shown later in the budget analysis (Table 2), the moisture budget is slightly imbalanced in some models, e.g., GFDL_CCM2.1, JP_CCSR3.2H and FR_IPSL_CM4.1. Causes for this imbalance might be associated with the interpolation of model variables from the original vertical coordinate to the standard pressure coordinate. This could be larger over regions with variations in topography. Thus, saving flux variables from the original coordinate is important in the budget analysis. Another possible cause is related to the transient component. In our calculation, monthly data are used, so the transient terms are excluded. We also notice that the availability of model variables is not consistent in some model. At some grid points, for instance, the horizontal velocity u and v exist, while the moisture q is missing. This creates an inconsistency between the flux terms, such as $-\nabla \cdot \mathbf{v}q$, and the advection terms, such as $-\mathbf{v} \nabla \cdot q$. Overall, the moisture budget in most models is still balanced quite nicely (Fig. 2).

3. Global water vapor budget

To more precisely estimate the change of precipitation, a column-integrated water vapor budget is used, which can be written as

$$P - E \approx -\langle \nabla \cdot \mathbf{v}q \rangle, \quad (7)$$

where pressure velocity ω is assumed to be zero at the surface p_s and at tropopause p_T . For simplicity and consistency, the precipitation (latent heat release) P is in energy units (Wm^{-2}) which, divided by 28, becomes mm day^{-1} . E is evaporation (surface latent heat flux) and $\mathbf{v}$ is horizontal velocity. The anomalous column-integrated moisture budget can be further written in an advection form:

$$\begin{aligned} P' - E' &\approx -\langle \nabla \cdot \mathbf{v}q \rangle' \\ &\approx -\langle \omega \partial_p q \rangle' - \langle \mathbf{v} \cdot \nabla q \rangle' \\ &\approx -\langle \bar{\omega} \partial_p q' \rangle - \langle \omega' \partial_p \bar{q} \rangle - \langle \mathbf{v} \cdot \nabla q \rangle'. \end{aligned} \quad (8)$$

The nonlinear term $-\langle \omega' \partial_p q' \rangle$ is neglected here. In a warmer climate, based on (7) and (8), the fractional change of $P - E$ should be equal to the change in the column-integrated convergence of moisture flux:

$$\frac{P' - E'}{\bar{P} - \bar{E}} = \frac{-\langle \bar{\omega} \partial_p q' \rangle}{-\langle \nabla \cdot \bar{\mathbf{v}} \bar{q} \rangle} + \frac{-\langle \omega' \partial_p \bar{q} \rangle}{-\langle \nabla \cdot \bar{\mathbf{v}} \bar{q} \rangle} + \frac{-\langle \mathbf{v} \cdot \nabla q \rangle'}{-\langle \nabla \cdot \bar{\mathbf{v}} \bar{q} \rangle} + \text{residual}_q. \quad (9)$$

The first term on the r.h.s. of (9) $-\langle \bar{\omega} \partial_p q' \rangle$ is a thermodynamic component, which is mainly associated with the water vapor change. The second term on the r.h.s. $-\langle \omega' \partial_p \bar{q} \rangle$ is a dynamic component, which is related to the change in tropical circulation (CHOU09; Emori and Brown 2005; Held and Soden 2006).

Most precipitation occurs over convective regions, which are roughly defined with $\omega < 0$ (upward motion) at 500 hPa, e.g., indicated by thick dashed curves in Fig. 1, so we focus on the moisture budget (9) averaged over convective regions. Figure 2a shows a rough balance between the fractional changes of $P - E$ and $-\langle \nabla \cdot \mathbf{v}q \rangle$, so the moisture in convective regions is conserved in all climate models. Since the horizontal moisture advection and the residual term (the third and fourth terms on the r.h.s. of (9)) are relatively small for most climate model simulations, compared to the thermodynamic and dynamic components (Table 2), they are usually neglected in the first-order approximation. Thus, (9) could become similar to the simplified precipitation equation discussed in section 1, i.e., $P'/\bar{P} = M'_c/\bar{M}_c + q'_s/\bar{q}_s$, if the fractional changes of precipitation and evaporation averaged over convective regions were the same. For global averages, the fractional changes of precipitation and evaporation must be equal due to the balance of the global moisture budget (Held and Soden 2006; Lambert and Webb 2008; Vecchi and Soden 2007). Averaged over convective regions, on the other hand, it is not guaranteed that the precipitation change is equal to the evaporation change. In fact, the evaporation change is usually smaller than the precipitation change (Table 2), making the fractional change of $P - E$ larger than that of precipitation in most climate model simulations (Fig. 2b). Unlike the global precipitation change, which is controlled by the global energy balance with its fractional change being much smaller than the Clausius-Clapeyron thermal scaling (Held and Soden 2006; Lambert and Webb 2008; Vecchi and Soden 2007), $P - E$ averaged over convective regions does not depend on the global energy balance, so the fractional change of $P - E$ could be either larger or smaller than the Clausius-Clapeyron thermal scaling. For the climate model simulations used in this study, however, the fractional change of $P - E$ is still consistently smaller than the Clausius-Clapeyron thermal scaling (Fig. 2c), even though some of those changes in $P - E$ are much larger than the fractional change of precipitation (Fig. 2b).

Considering the moisture convergence terms, i.e., the right hand side of (9), the fractional change of the thermodynamic component is smaller than the value indicated by the Clausius-

Clapeyron thermal relation, but is still larger than the fractional change of $P - E$ (Table 2). Thus, the fractional change of the dynamic component must be negative, as shown in Table 2, and tropical circulation should be weakened in the future warmer climate. In the 16 climate models (Table 1) that we analyzed here, all show a weakening of tropical circulation, which is indicated by pressure velocity at 500 hPa (Fig. 2d). Since the change of $P - E$ is not controlled by the global energy balance, the negative fractional change of the dynamic component should not be directly associated with less efficient radiative cooling via precipitation change, as indicated in previous studies (Held and Soden 2006; Vecchi and Soden 2007). In other words, tropical circulation could be strengthened in a warmer climate even though the fractional change of precipitation is much less than the fractional change of water vapor. For instance, the fractional change of $P - E$ could be larger than the Clausius-Clapeyron thermal scaling if the change of evaporation were much less than that of precipitation, or even negative.

4. Impacts of deepened convection on the strength of tropical circulation

If tropical circulation cannot be determined by different fractional changes between the global-mean precipitation and water vapor, what kind of mechanisms control the change of tropical circulation? In previous studies (Chou and Neelin 2004, hereafter CN04; Chou et al. 2006; CHOU09), several mechanisms for regional tropical precipitation change have been proposed through the further analysis of the MSE budget, which is

$$\langle \nabla \cdot \mathbf{v}h \rangle \approx F^{net}. \quad (10)$$

The net energy flux into the atmosphere is $F^{net} = F_t - F_s$. The net heat flux at the top of the atmosphere (TOA) is $F_t = S_t^\downarrow - S_t^\uparrow - R_t^\uparrow$ and the net heat flux at the surface is $F_s = S_s^\downarrow - S_s^\uparrow + R_s^\downarrow - R_s^\uparrow - E - H$. Subscripts s and t on the solar ($S^\downarrow$ and $S^\uparrow$) and longwave ($R^\uparrow$ and $R^\downarrow$) radiative terms denote surface and model top, and H is sensible heat flux. Positive F_t and F_s indicate downward heat fluxes. The anomalous MSE budget can then be written as

$$\langle \omega \partial_p h \rangle' + \langle \mathbf{v} \cdot \nabla (T + q) \rangle' \approx F^{net'} \quad (11)$$

or

$$\langle \omega' \partial_p \bar{h} \rangle \approx -\langle \bar{\omega} \partial_p h' \rangle - \langle \mathbf{v} \cdot \nabla (T + q) \rangle' + F^{net'}. \quad (12)$$

Thus, the fractional changes of the column-integrated MSE budget can be written as

$$\frac{\langle \omega' \partial_p \bar{h} \rangle}{\langle \omega \partial_p h \rangle} = \frac{-\langle \bar{\omega} \partial_p h' \rangle}{\langle \omega \partial_p h \rangle} + \frac{-\langle \mathbf{v} \cdot \nabla (q + T) \rangle'}{\langle \omega \partial_p h \rangle} + \frac{F^{net'}}{\langle \omega \partial_p h \rangle} + residual_h. \quad (13)$$

In (13), the change in the vertical velocity that is associated with tropical circulation $\langle \omega' \partial_p \bar{h} \rangle$ is mainly determined by changes in the term associated with atmospheric stability $-\langle \bar{\omega} \partial_p h' \rangle$, the horizontal MSE advection $-\langle \mathbf{v} \cdot \nabla (q + T) \rangle'$ and the net energy flux into the troposphere $F^{net'}$. In a warmer climate, the tropospheric vapor water increases and mainly concentrates at lower troposphere, so the atmosphere becomes more unstable. On the other hand, the vertical profile of the warming induced by latent heating should follow the moist adiabatic process; thus the atmospheric temperature increases more at higher troposphere than at lower troposphere (Karl et al. 2006; Trenberth et al. 2007). This amplification of tropospheric warming can stabilize the atmosphere. Although these two effects tend to cancel each other out, moisture contributes slightly more than temperature (CHOU09). In other words, the atmosphere becomes a little more unstable in a warmer climate when considering only changes in temperature and moisture. This effect has been termed the “rich-get-richer” mechanism (CN04; CHOU09). In a warmer climate, tropical convection also tends to extend higher (Fig. 3) due to the uplifting of the tropopause (Holzer and Boer 2001; Santer et al. 2003; International Ad Hoc Detection and Attribution Group 2005; Lorenz and DeWeaver 2007), which stabilizes the atmosphere because of the increase in its effective static stability (CHOU09). The changes of convection depth are around 2~3%, i.e., deepening for all 16 GCM simulations (Table 2). Convection height here is roughly estimated by

assuming a similar MSE at the base and top of convection, so it represents the maximum level of convection (e.g., Yu et al. 1998). Besides the changes associated with atmospheric stability, other effects, such as the MSE advection, also affect tropical circulation. In convective regions, the MSE advection is usually negative since the inward flow at lower troposphere brings relatively dry air from subsidence regions to convective regions; this process is termed the upped-ante mechanism (CN04; CHOU09). This effect often occurs over margins of convective regions. The change in the net energy input into the atmosphere is relatively small since changes in evaporation and sensible heat tend to balance the radiative cooling when averaging over a big enough domain (Lambert and Webb 2008; Vecchi and Soden 2007). Overall, tropical circulation is strengthened by the rich-get-richer mechanism, but weakened by the upped-ante mechanism and the effect of convection depth.

According to the discussion above, one of the robust mechanisms that can reduce tropical circulation is associated with depth of convection. To examine this effect associated with the change of convection depth, an intermediate climate model coupled with a mixed-layer ocean (Neelin and Zeng 2000; Zeng et al. 2000) is used. This model (QTCM1) has been used in studying mechanisms of regional tropical precipitation changes under global warming (Neelin et al. 2003; CN04; Chou et al. 2007).

In a set of experiments, convection height is gradually increased from 155 hPa~137 hPa, which are equivalent to -1.2%~3.3% of changes in convection depth (Table 3), similar to those in the GCM simulations (Table 2), so the atmosphere becomes more stable. The fractional change of the global-mean water vapor varies between $7.0\%K^{-1}$ and $8.2\%K^{-1}$, which roughly follows the Clausius-Clapeyron scaling. The change of the global-mean precipitation, on the other hand, ranges from $2.5\%K^{-1}$ to $1.5\%K^{-1}$, which is far below the corresponding changes in water vapor (Fig. 4). According to the argument discussed in previous studies (Held and Soden 2006; Vecchi and Soden 2007), tropical circulation must weaken in all experiments because the increase in precipitation is slower than in water vapor. On the contrary, tropical circulation is not always

weakened in these experiments. In Fig. 4a, the fractional change of vertical velocity ω at 500 hPa, which can be used to roughly estimate the strength of tropical circulation (Vecchi and Soden 2007), is positive when convection is shallower, and becomes negative as convection becomes deeper. In other words, tropical circulation is not necessarily weakened even when the fractional change of the global-mean precipitation is consistently less than that of the global-mean water vapor. The vertical profile of changes in vertical velocity is shown in Fig. 5. The vertical profiles with weakened tropical circulation (M_s+2 and M_s+3) are similar to those in Fig. 3. It implies that the changes in vertical velocity shown in Fig. 3 are associated with the deepening of tropical convection. In the MSE budget shown in Table 4, $-\langle \bar{\omega} \partial_p h' \rangle$, which is associated with the opposite sign of the change in the effective static stability, is gradually reduced as convection becomes deeper, so the atmosphere becomes more stable. The horizontal MSE advection, which is another mechanism that can suppress tropical circulation (CN04; CHOU09), is negative in all experiments, but does not vary too much among these experiments. The anomalous net energy input into the atmosphere $F^{net'}$ increases as convection deepens, so $F^{net'}$ tends to enhance, not suppress, tropical circulation. Thus, the gradual weakening of tropical circulation in these experiments is mainly due to the change in atmospheric stability that is associated with depth of convection.

On the r.h.s. of the moisture budget (9), since the fractional change of water vapor does not vary too much, the thermodynamic component is roughly constant among these QTCM1 experiments, ranging from 5.3% to 6.6% (Table 3). The dynamic component, on the other hand, varies greatly (from 4.2% to -4.6%) due to the variation in the strength of tropical circulation. Both the horizontal moisture advection and the residual term are relatively small. Thus, the fractional change in the convergence of moisture flux becomes smaller as the atmosphere becomes more stable (Table 3). On the l.h.s. of (9), the precipitation averaged over convective regions remains relatively unchanged as the atmosphere becomes more stable, while the evaporation becomes larger (Fig. 4). This leads to a smaller change in $P - E$ (Table 3). In other words, the weaker (stronger) convergence of moisture flux associated with the weakened (enhanced) tropical circu-

lation compensates for the increase (decrease) of the evaporation over convective regions, so the precipitation change does not vary too much with the change in convection depth. Overall, the fractional change of $P - E$ averaged over convective regions varies with convection depth, but not fractional changes in either the global-mean precipitation or water vapor, which are controlled by different mechanisms.

5. Discussion and summary

As global warming becomes dominant, many climate changes are starting to show. One of the well-known climate changes is a weakening of tropical circulation, which could affect rainfall amount and intensity both in regional and global scales. Previous studies (Soden and Held 2006; Vecchi and Soden 2007) show that the atmospheric water vapor increases faster than precipitation, so the associated tropical circulation must weaken. However, we demonstrated here that this condition cannot guarantee the weakening of tropical circulation.

We provided a possible mechanism, which is associated with depth of convection, for inducing the weakening of tropical circulation under global warming. Conventionally, vertical motion associated with tropical circulation is determined by atmospheric stability. Atmospheric stability is usually affected by the vertical profile of atmospheric temperature and moisture (Karl et al. 2006; Trenberth et al. 2007), which tends to cancel each other out if the atmosphere is mainly controlled by convection, such as in the Tropics (CHOU09). A previous analysis (Yu et al. 1998) shows that atmospheric stability can also be affected by depth of convection. Under global warming, a deepening of convection is consistently found in the CMIP3 dataset. This leads to a weakening of tropical circulation in the CMIP3 global warming simulations. The change in atmospheric stability also modifies the evaporation gradient between convective and subsidence regions, which compensates for the change in the convergence of moisture flux to maintain a similar change in global-mean precipitation. Overall, by examining mechanisms for the weakening of tropical circulation, this study offers an alternative way to detect global warming impacts on tropical cir-

ulation.

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REFERENCES

- Allen, M. R., and W. J. Ingram, 2002: Constraints on future changes in climate and the hydrologic cycle. *Nature*, **419**, 224-232.
- Betts, A. K., and M. J. Miller, 1993: The Betts-Miller scheme, Chapter 9 in *The Representation of Cumulus Convection in Numerical Models of the Atmosphere*. (Eds. K. A. Emanuel and D. J. Raymond.) *Amer. Meteor. Soc., Meteor. Mon.*, **24**, No. 46, 107-121.
- Chou, C., and J. D. Neelin, 1996: Linearization of a longwave radiation scheme for intermediate tropical atmospheric models. *J. Geophys. Res.*, **101**, 15129-15145.
- Chou, C., and J. D. Neelin, 1999: Cirrus detrainment-temperature feedback. *Geophys. Res. Lett.*, **26**, 1295-1298.
- Chou, C., and J. D. Neelin, 2004: Mechanisms of global warming impacts on regional tropical precipitation. *J. Climate*, **17**, 2688-2701.
- Chou, C., J. D. Neelin, J.-Y. Tu, and C.-T. Chen, 2006: Regional tropical precipitation change mechanisms in ECHAM4/OPYC3 under global warming. *J. Climate*, **19**, 4207-4223.
- Chou, C., J.-Y. Tu, and P.-H. Tan, 2007: Asymmetry of tropical precipitation change under global warming. *Geophys. Res. Lett.*, **34**, L17708, doi:10.1029/2007GL030327.
- Chou, C., J. D. Neelin, C.-A. Chen, and J.-Y. Tu, 2009: Evaluating the “rich-get-richer” mechanism in tropical precipitation change under global warming. *J. Climate*, **22**, 1982-2005.
- Emori, S., and S. J. Brown, 2005: Dynamic and thermodynamic changes in mean and extreme precipitation under changed climate. *Geophys. Res. Lett.*, **32**, L17706, doi:10.1029/2005GL023272.
- Fu, Q., and K. N. Liou, 1993: Parameterization of the radiative properties of cirrus clouds. *J. Atmos. Sci.*, **50**, 2008-2025.

- Harshvardhan, R. Davies, D. A. Randall, and T. G. A. Corsetti, 1987: fast radiation parameterization for general circulation models. *J. Geophys. Res.*, **92**, 1009-1016.
- Held, I. M., and B. J. Soden, 2006: Robust responses of the hydrological cycle to global warming. *J. Climate*, **19**, 5686-5699.
- Holzer, M., G. J. Boer, 2001: Simulated changes in atmospheric transport climate. *J. Climate*, **14**, 4398-420.
- International Ad Hoc Detection and Attribution Group, 2005: Detection and attributing external influences on the climate system: a review of recent advances. *J. Climate*, **18**, 1291-01314.
- Karl, T. R., S. J. Hassol, C. D. Miller, and W. L. Murray (eds), 2006: *Temperature Trends in the Lower Atmosphere: Steps for Understanding and Reconciling Differences*. A report by the Climate Change Science Program and Subcommittee on Global Change Research, Washington, DC.
- Knutson, T. R., and S. Manabe, 1995: Time-mean response over the tropical Pacific to increased CO₂ in a coupled ocean-atmosphere model. *J. Climate*, **8**, 2181-2199.
- Lambert, F. H., and M. J. Webb, 2008: Dependency of global mean precipitation on surface temperature. *Geophys. Res. Lett.*, **35**, L16706, doi:10.1029/2008GL03438.
- Lorenz, D. J., and E. T. DeWeaver, 2007: Tropopause height and zonal wind response to global warming in the IPCC scenario integrations. *J. Geophys. Res.*, **112**, D10119, doi:10.1029/2006JD008087.
- Neelin, J. D., and J.-Y. Yu, 1994: Modes of tropical variability under convective adjustment and the Madden-Julian Oscillation. Part I: Analytical theory. *J. Atmos. Sci.*, **51**, 1876-1894.
- Neelin, J. D., and N. Zeng, 2000: A quasi-equilibrium tropical circulation model-formulation. *J. Atmos. Sci.*, **57**, 1741-1766.
- Neelin, J. D., C. Chou, and H. Su, 2003: Tropical drought regions in global warming and El

- Niño teleconnections. *Geophys. Res. Lett.*, **30**, 2275, doi:10.1029/2003GL018625.
- Santer, B. D., and coauthors, 2003: Contributions of anthropogenic and natural forcing to recent tropopause height changes. *Science*, **301**, 479-483.
- Stephens, G. L., and T. D. Ellis, 2008: Controls of global-mean precipitation increases in global warming GCM experiments. *J. Climate*, **21**, 6141-6155.
- Stevens, B., J. Duan, J. C. McWilliams, M. Münnich, and J. D. Neelin, 2002: Entrainment, Rayleigh friction, and boundary layer winds over the tropical Pacific. *J. Climate*, **15**, 30-44.
- Trenberth, K. E., A. Dai, R. M. Rasmussen, and D. B. Parsons, 2003: The changing character of precipitation. *Bull. Am. Meteorol. Soc.*, **84**, 1205-1217.
- Trenberth, K. E., and coauthors, 2007: in *Climate Change 2007, The Physical Science Basis* Chapter 3, eds Solomon S. et al. (Cambridge Univ. Press, Cambridge) pp235-336.
- Vecchi, G. A., B. J. Soden, A. T. Wittenberg, I. M. Held, A. Leetmaa, and M. J. Harrison, 2006: Weakening of tropical Pacific atmospheric circulation due to anthropogenic forcing. *Nature*, **441**, 73-76.
- Vecchi, G. A., and B. J. Soden, 2007: Global warming and the weakening of the tropical circulation. *J. Climate*, **20**, 4316-4340.
- Wentz, F. J., L. Ricciardulli, K. Hilburn, and C. Mears, 2007: How much more rain will global warming bring? *Science*, **317**, 233-235.
- Yu, J.-Y., and J. D. Neelin, 1994: Modes of tropical variability under convective adjustment and the Madden-Julian Oscillation. Part II: Numerical results. *J. Atmos. Sci.*, **51**, 1895-1914.
- Yu, J.-Y., C. Chou, and J. D. Neelin, 1998: Estimating the gross moist stability of the tropical atmosphere. *J. Atmos. Sci.*, **55**, 1354-1372.
- Zeng, N., J. D. Neelin, and C. Chou, 2000: A quasi-equilibrium tropical circulation model-implementation and simulation. *J. Atmos. Sci.*, **57**, 1767-1796.

TABLE CAPTIONS

Table 1: A list of the 16 coupled atmosphere-ocean climate model simulations in the A1B scenario from the CMIP3 archive.

Table 2: Fractional changes of the moisture budgets for 16 CMIP3 models on the A1B emission scenario between the last and first 20 years of the 21st century. Rows 1-3 are averaged over the entire globe and the rest are averaged only over convective (ascending) regions. The values are normalized by the global-mean surface temperature change.

Table 3: Fractional changes of the QTCM moisture budgets for different convection depths. From left to right, convection becomes deeper. Rows 1-3 are averaged over the entire globe and the rest are averaged only over convective (ascending) regions, indicated by thick dashed curves in Fig. 1. The values are normalized by the global-mean surface temperature change.

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FIGURE CAPTIONS

Figure 1: Precipitation differences between experiments with $1\times\text{CO}_2$ and $2\times\text{CO}_2$ concentration for different convection heights: (a) M_s-1 , (b) M_s+0 and (c) M_s+1 . The thick dashed curves indicate a convective margin, i.e., $\omega = 0$ at 500 hPa. The unit is mm day^{-1} .

Figure 2: Scattering plots for fractional changes of hydrological variables and the strength of tropical circulation averaged over convective regions between the last and first 20 years of the 21st century for each of the climate models listed in Table 1. (a) $P - E$ vs. the column-integrated convergence of the moisture flux $-\langle \nabla \cdot \mathbf{v}q \rangle$, (b) $P - E$ vs. precipitation, (c) surface temperature vs. $P - E$, and (d) $P - E$ vs. the strength of tropical circulation (vertical velocity at 500 hPa).

Figure 3: The anomalous pressure velocity between the last and first 20 years of the 21st century. The anomalies are averaged over convective regions for the 16 CMIP3 models listed in Table 1. Positive values indicate downward motion (reduced) and negative values indicate upward motion (enhanced).

Figure 4: Fractional changes of precipitation and water vapor, and tropical circulation in the QTCM1 experiments associated with different convection depths. The numbers are normalized by the change of global-mean surface temperature. Depth of convection becomes shallower from left to right. (a) Pressure velocity at 500 hPa and (b) evaporation averaged over convective regions (indicated by thick curves in Fig. 1), and global-mean (c) precipitation and (d) column-integrated water vapor.

Figure 5: The anomalous pressure velocity averaged over convective regions (indicated in Fig. 1) between experiments with $1\times\text{CO}_2$ and $2\times\text{CO}_2$ concentrations for different convection heights: M_s-1 , M_s+0 , M_s+1 , M_s+2 and M_s+3 .

TABLE 1: A list of the 16 coupled atmosphere-ocean climate model simulations in the A1B scenario from the CMIP3 archive.

Model	Description
CA_CCCMA3.1_T47	Canada, Canadian Centre for Climate Modelling and Analysis, CGCM3.1 Model (T47)
CA_CCCMA3.1_T63	Canada, Canadian Centre for Climate Modelling and Analysis CGCM3.1 Model (T63)
CN_IAP	China, LASG, Institute of Atmospheric Physics, FGOALS1.0_g Model
FR_CNRM_CM3	France, Centre National de Research Meteorologiques, CM3 Model
FR_IPSL_CM4.1	France, IPSL/LMD/LSCE, CM4 V1 Model
GFDL_CCM2.0	USA, NOAA Geophysical Fluid Dynamics Laboratory, CM2.0 Model
GFDL_CCM2.1	USA, NOAA Geophysical Fluid Dynamics Laboratory, CM2.1 Model
JP_CCSR3.2H	Japan, CCSR/NIES/FRCGC, MIROC Model V3.2, high resolution
JP_CCSR3.2M	Japan, CCSR/NIES/FRCGC, MIROC Model V3.2, medium resolution
JP_MRI2.3	Japan, Meteorological Research Institute (MRI), CM2.3.2a
MPI_ECHAM5	Germany, Max Planck Institute for Meteorology, ECHAM5/MPI OM
NCAR_CCSM3	USA, NCAR Community Climate System Model, CCSM3.0
NCAR_PCM1	USA, NCAR Parallel Climate Model (Version 1)
Russia_INMCM3	Russia, Institute for Numerical Mathematics, INMCM3.0 Model
UKMO_HadCM3	UK, Hadley Center for Climate Prediction, Met Office, HadCM3 Model
UKMO_HadGEM1	UK, Hadley Centre for Climate Prediction, Met Office, HadGEM1 Model

Table 2: Fractional changes of the moisture budgets for 16 CMIP3 models on the A1B emission scenario between the last and first 20 years of the 21st century. Rows 1-3 are averaged over the entire globe and the rest are averaged only over convective (ascending) regions. The values are normalized by the global-mean surface temperature change.

term	CA_CCCMA3.1_T47	CA_CCCMA3.1_T63	CN_IAP	FR_CNRM_CM3	FR_IPSL_CM4.1	GFDL_CCM2.0	GFDL_CCM2.1	JP_CCSR3.2H	JP_CCSR3.2M	JP_MR2.3	MPI_ECHAM5	NCAR_CCSM3	NCAR_PCM1	Russia_INMCM3	UKMO_HadCM3	UKMO_HadGEM1
global $P' / \bar{P}$	0.0196	0.0189	0.0131	0.0155	0.0221	0.0087	0.0060	0.0156	0.0163	0.0239	0.0201	0.0242	0.0168	0.0175	0.0099	0.0107
global $E' / \bar{E}$	0.0195	0.0188	0.0172	0.0153	0.0216	0.0085	0.0057	0.0156	0.0164	0.0244	0.0200	0.0242	0.0168	0.0177	0.0099	0.0106
global $q' / \bar{q}$	0.0846	0.0806	0.0705	0.0763	0.0849	0.0809	0.0862	0.0780	0.0776	0.0817	0.0836	0.0705	0.0672	0.0661	0.0715	0.0729
$\omega' / \bar{\omega}$ at 500hPa	-0.0231	-0.0218	-0.0226	-0.0242	-0.0179	-0.0282	-0.0299	-0.0225	-0.0252	-0.0258	-0.0232	-0.0210	-0.0164	-0.0232	-0.0169	-0.0226
$P'E' / \bar{P}\bar{E}$	0.0332	0.0287	0.0104	0.0407	0.0492	0.0254	0.0373	0.0340	0.0386	0.0346	0.0516	0.0321	0.0315	0.0336	0.0341	0.0278
$-\langle \nabla \cdot \mathbf{v}q \rangle' / -\langle \nabla \cdot \mathbf{v}q \rangle$	0.0285	0.0254	0.0099	0.0362	0.0346	0.0157	0.0250	0.0226	0.0251	0.0283	0.0494	0.0284	0.0257	0.0355	0.0318	0.0227
$-\langle \mathbf{v} \cdot \nabla q \rangle' / -\langle \mathbf{v} \cdot \nabla q \rangle$	0.0017	-0.0003	0.0006	-0.0061	0.0020	-0.0059	-0.0070	-0.0065	-0.0096	0.0023	-0.0048	-0.0045	-0.0051	-0.0068	-0.0076	-0.0064
$-\langle \omega \partial_p q \rangle' / -\langle \nabla \cdot \mathbf{v}q \rangle$	0.0268	0.0258	0.0093	0.0423	0.0322	0.0232	0.0320	0.0291	0.0347	0.0261	0.0541	0.0329	0.0305	0.0423	0.0450	0.0288
$-\langle \bar{\omega} \partial_p q' \rangle' / -\langle \nabla \cdot \mathbf{v}q \rangle$	0.0596	0.0549	0.0480	0.0513	0.0484	0.0498	0.0659	0.0576	0.0592	0.0567	0.0481	0.0539	0.0497	0.0500	0.0398	0.0479
$-\langle \omega' \partial_p \bar{q} \rangle' / -\langle \nabla \cdot \mathbf{v}q \rangle$	-0.0392	-0.0370	-0.0377	-0.0268	-0.0267	-0.0460	-0.0287	-0.0399	-0.0359	-0.0409	-0.0167	-0.0330	-0.0270	-0.0227	-0.0300	-0.0373
$P' / \bar{P}$	0.0234	0.0209	0.0149	0.0178	0.0258	0.0117	0.0101	0.0190	0.0204	0.0267	0.0269	0.0298	0.0220	0.0215	0.0154	0.0151
$E' / \bar{E}$	0.0174	0.0162	0.0172	0.0051	0.0098	0.0008	-0.0096	0.0082	0.0076	0.0213	0.0079	0.0277	0.0133	0.0133	0.0013	0.0055
$\mathbf{z}'_T / \bar{\mathbf{z}}_T$	0.0270	0.0257	0.0266	0.0295	0.0287	0.0270	0.0329	0.0299	0.0336	0.0249	0.0259	0.0285	0.0309	0.0255	0.0241	0.0251

TABLE 3: Fractional changes of the QTCM moisture and MSE budgets for different M_s , which indicates the contribution of vertical temperature profile. Rows 1-3 are averaged over the entire globe (80°S-80°N) and the rest are averaged only over convective (ascending) regions. The values are normalized by the global surface temperature change.

term	M_s-1	M_s+0	M_s+1	M_s+2	M_s+3
global $P'/\bar{P}$	0.0253	0.0219	0.0200	0.0176	0.0148
global $E'/\bar{E}$	0.0245	0.0217	0.0200	0.0181	0.0156
global $q'/\bar{q}$	0.0695	0.0724	0.0761	0.0790	0.0821
$\omega'/\bar{\omega}$ at 500 hPa	0.0290	0.0166	0.0039	-0.0093	-0.0219
$P' - E'/\bar{P} - \bar{E}$	0.0906	0.0695	0.0476	0.0284	0.0088
$-\langle \nabla \cdot \mathbf{v}q \rangle' / -\langle \nabla \cdot \mathbf{v}\bar{q} \rangle$	0.0906	0.0695	0.0478	0.0285	0.0089
$-\langle \mathbf{v} \cdot \nabla q \rangle' / -\langle \nabla \cdot \mathbf{v}\bar{q} \rangle$	-0.0111	-0.0063	-0.0039	-0.0003	0.0018
$-\langle \omega \partial_p q \rangle' / -\langle \nabla \cdot \mathbf{v}\bar{q} \rangle$	0.1017	0.0758	0.0517	0.0288	0.0221
$-\langle \bar{\omega} \partial_p q' \rangle / -\langle \nabla \cdot \mathbf{v}\bar{q} \rangle$	0.0531	0.0572	0.0621	0.0638	0.0662
$-\langle \omega' \partial_p \bar{q} \rangle / -\langle \nabla \cdot \mathbf{v}\bar{q} \rangle$	0.0418	0.0183	-0.0032	-0.0249	-0.0459
$P'/\bar{P}$	0.0376	0.0305	0.0252	0.0196	0.0153
$E'/\bar{E}$	-0.0099	-0.0044	0.0050	0.0117	0.0211
$z'_T/\bar{z}_T$	-0.012	0.00	0.0124	0.0226	0.0331

TABLE 4: Fractional changes of the QTCM moist static energy budgets for different M_s , which represents different convection depths. From left to right, convection becomes deeper. The values are averaged over convective (ascending) regions and normalized by the global-mean surface temperature change.

term	M_s-1	M_s+0	M_s+1	M_s+2	M_s+3
$\langle \omega' \partial_p \bar{h} \rangle / \langle \overline{\omega \partial_p h} \rangle$	0.0375	0.0149	-0.0067	-0.0303	-0.0501
$-\langle \bar{\omega} \partial_p h' \rangle / \langle \overline{\omega \partial_p h} \rangle$	0.1259	0.0819	0.0368	-0.0030	-0.0464
$-\langle \mathbf{v} \cdot \nabla (q + T) \rangle' / \langle \overline{\omega \partial_p h} \rangle$	-0.0401	-0.0412	-0.0562	-0.0542	-0.0648
$F^{net} / \langle \overline{\omega \partial_p h} \rangle$	-0.0011	0.0092	0.0340	0.0453	0.0691

FIGURE 1: Precipitation differences between experiments with 1×CO₂ and 2×CO₂ concentration for different convection heights: (a) M_s-1 , (b) M_s+0 and (c) M_s+1 . The thick dashed curves indicate a convective margin, i.e., $\omega = 0$ at 500 hPa. The unit is mm day⁻¹.

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