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Weakening of tropical free tropospheric temperature gradients with global warming

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Abstract:	The weak temperature gradients in the tropical free troposphere due to the vanishing Coriolis force near the equator lead to a strong dynamical coupling over the entire tropics. Using theory and a suite of targeted model experiments, we show that the weak temperature gradients further weaken under global warming. We show that the temperature gradient is set by the circulation strength, with a weaker circulation being associated with weaker gradients. Thus, the known scaling difference between atmospheric radiative cooling and static stability that leads to a slow-down of the circulation under warming also leads to a weakening of the temperature gradients in the tropical free troposphere. The impact from the weakening circulation on the weakening of temperature gradients is shown to dominate over the impact of masked CO$_2$ forcing and the El-Nino like tropical Pacific warming pattern in model projections. Key to the result is the non-linear zonal momentum advection term. Using the well-known Matsuno-Gill model with correct scaling of heating and static stability may give the correct sign of the response in the temperature gradients, but incorrect scaling, due to the linear momentum damping in that model. The robust scaling of the magnitude of the tropical quasi-stationary structure with temperature opens possibilities for theoretical advances on questions of societal relevance, ranging from changes in tropical cloudiness to heat stress under climate change.
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Dear *Journal of the Atmospheric Sciences* Editor:

Enclosed is a manuscript entitled "Weakening of tropical free tropospheric temperature gradients with global warming" that we respectfully submit for consideration for publication in *Journal of the Atmospheric Sciences*. The manuscript is coauthored by Heng Quan, Yi Zhang and Stephan Fueglistaler. Heng Quan will be the corresponding author. The work we present is original and has never been submitted for publication elsewhere. Thank you in advance for your time and consideration.

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Weakening of tropical free tropospheric temperature gradients with global warming

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ABSTRACT: The weak temperature gradients in the tropical free troposphere due to the vanishing Coriolis force near the equator lead to a strong dynamical coupling over the entire tropics. Using theory and a suite of targeted model experiments, we show that the weak temperature gradients further weaken under global warming. We show that the temperature gradient is set by the circulation strength, with a weaker circulation being associated with weaker gradients. Thus, the known scaling difference between atmospheric radiative cooling and static stability that leads to a slow-down of the circulation under warming also leads to a weakening of the temperature gradients in the tropical free troposphere. The impact from the weakening circulation on the weakening of temperature gradients is shown to dominate over the impact of masked CO_2 forcing and the El-Nino like tropical Pacific warming pattern in model projections. Key to the result is the non-linear zonal momentum advection term. Using the well-known Matsuno-Gill model with correct scaling of heating and static stability may give the correct sign of the response in the temperature gradients, but incorrect scaling, due to the linear momentum damping in that model. The robust scaling of the magnitude of the tropical quasi-stationary structure with temperature opens possibilities for theoretical advances on questions of societal relevance, ranging from changes in tropical cloudiness to heat stress under climate change.

1. Introduction

Due to the small Coriolis force at low latitudes, the tropical free troposphere cannot sustain horizontal temperature gradients as large as at higher latitudes (Charney 1963). Any strong horizontal buoyancy or temperature gradients produced by deep convection would be quickly homogenized by gravity waves (Bretherton and Smolarkiewicz 1989). Consequently, on climate time scales horizontal pressure and temperature gradients can be assumed to be small, and the ”weak temperature gradient (WTG)” approximation allows to simplify the equations governing the atmospheric dynamics (Sobel and Bretherton 2000; Sobel et al. 2001).

The WTG approximation may be used to parameterize tropical planetary-scale circulation in column models (SCMs) and cloud resolving models (CRMs). For example, Sobel and Bretherton (2000) proposed to parameterize the vertical velocity in SCMs in a way that represents the dominant large-scale balance between diabatic heating and vertical advection of potential temperature (Sobel et al. 2001). This approach can be generalized to CRMs (Raymond and Zeng 2005; Sessions et al. 2010; Wang and Sobel 2011; Daleu et al. 2012; Warren et al. 2020); whereby an alternative approach is the “damped gravity wave” method (Kuang 2008; Blossey et al. 2009; Romps 2012; Edman and Romps 2014).

Together with the convective quasi-equilibrium (QE) approximation (i.e. moist convection maintains the vertical temperature profile close to a moist adiabat (Arakawa and Schubert 1974; Emanuel et al. 1994), the QE-WTG framework is the foundation to understand many aspects of tropical climate and changes therein for example due to global warming. In the QE-WTG framework, the tropical troposphere can be seen as consisting of a boundary layer with a substantial temperature gradient and a relatively homogeneous free troposphere whose temperature is determined by the subcloud moist static energy (MSE) in the regions of deep convection where subcloud MSE maximizes (e.g. Emanuel et al. (1994)). This framework has been used to explain the amplified warming over land (Byrne and O’Gorman 2018), an apparent super-moist adiabatic amplification in the tropical temperature trend profile (Flannaghan and Fueglistaler 2014), the trend of tropical heat extremes (Byrne 2021; Zhang et al. 2021), the enhanced precipitation contrast between wet and dry regions with warming (Neelin et al. 2003; Chou and Neelin 2004; Zhang and Fueglistaler 2019), and the SST pattern effect and its impact on climate sensitivity (Ceppi and Gregory 2017; Fueglistaler 2019; Fueglistaler and Silvers 2021). Thus, the magnitude of the tropical free tropospheric temperature gradient is of paramount importance for climate, and in the following we address the question how global warming will affect the tropical free tropospheric temperature gradients; specifically, whether the ”weak temperature gradient” will get weaker or stronger.

The paper is organized as follows. Section 2 provides a brief introduction to the relevant theory and mechanisms. Section 3 describes the numerical models and experiments used in this study. Section 4 discusses the results from model simulations with coupled Atmosphere-Ocean General Circulation Models (GCMs), and Atmospheric GCM simulations with prescribed sea surface temperatures. The simulations show a robust weakening of the temperature gradients independent of the question to what extent global warming results in an El-Nino like warming pattern in the tropics. Similarly, the simulations show that the masked CO_2 forcing is not a major contributor. The weakening of the temperature gradients must result from the slow-down of the atmospheric circulation under global warming, and the zonal momentum equation is used to derive a scaling between temperature gradient and circulation strength. Section 5 shows that idealized mock Walker cell simulations with a CRM follow the theoretical scaling. Finally, Section 6 summarizes the results and conclusions, and discusses implications.

2. Theory

a. Background

The relation between temperature gradients, pressure gradients and the momentum budget is discussed in Charney (1963). However, the impact of the fundamental slow-down of the atmospheric circulation ((Held and Soden 2006)) due to the different scaling of atmospheric radiative cooling and static stability with temperature (the latter being set by the boundary layer specific humidity, which scales approximately like Clausius-Clapeyron) on the quasi-stationary tropical structure of the circulation, pressure and temperature remains incompletely understood.

Because of the quasi-stationary geographic structure of atmospheric latent heating in the tropics, the tropics show a pronounced quasi-stationary wave structure in the troposphere, whereby temperature gradients maximise in the upper troposphere (warm anomlies in the regions of deep convection) and around the tropical tropopause (cold anomalies over the deep convective regions), with geopotential gradients maximizing in-between (e.g. Fueglistaler (2019)). The model proposed by Gill (1980) provides an elegant approach to understand the tropical tropospheric quasi-stationary structure as the consequence of steady equatorial Rossby and Kelvin waves emanating from the localized heating in the regions of deep convection. The "Gill model" is widely regarded as the basis for any discussion of the large-scale structure of the tropical atmosphere, and would seem the natural starting point for the problem of interest here. However, in order to arrive at an analytical solution, the Gill model represents dissipative processes as linear momentum and diabatic damping (their equations 2.6 - 2.9). The magnitude of the momentum damping coefficient is very important as it sets the length scale of the solution, but the term is physically poorly justified and operates largely as a "tuning" parameter. Our analysis below emphasizes the importance of the momentum balance for the temperature gradient, and the Gill model may not be able to provide the insights necessary to understand the relation between circulation strength and temperature gradient. In passing we note that a superficial look at the Gill solution may suggest an increase in the stationary wave amplitude since the latent heating term Q (precipitation) increases under global warming. The change in static stability with global warming, however, must also be considered, which is - slightly less obvious - encoded in the gravity wave phase speed $c = \frac{NH}{\pi}$ where stratification N is determined by static stability. Because a larger static stability decreases the stationary wave amplitude - which fights against the increase of the latent heating term Q - in a warmer climate, it is not obvious whether the stationary wave amplitude (hence free-tropospheric temperature gradients) will be larger or smaller in a warmer climate just from the Gill model.

b. Expected scaling based on the equatorial zonal momentum balance

The zonal momentum equation at a certain height $z = H$ in the free-troposphere sufficiently far away from the surface is

$$\frac{\partial u}{\partial t} + \mathbf{v} \cdot \nabla u - f v = -\frac{1}{\rho}\frac{\partial p}{\partial x}, \tag{1}$$

where u and v are zonal and meridional velocities, $\mathbf{v}$ is the three-dimensional velocity vector, p is pressure, and ρ is air density. The quasi steady-state zonal pressure gradient force in the equatorial upper troposphere is primarily balanced by the zonal advection of zonal momentum (Bao et al. 2022)

$$u\frac{\partial u}{\partial x} \approx -\frac{1}{\rho}\frac{\partial p}{\partial x}. \tag{2}$$

In the following, we will demonstrate how the left side of Eq. (2) corresponds to the strength of the overturning circulation W, while the right side is associated with horizontal (virtual) temperature gradients (δT_v) in the free troposphere, linking δT_v directly to the strength of the atmospheric circulation.

The left side of Eq. (2) corresponds to the overturning circulation strength through mass conservation

$$\frac{\partial u}{\partial x} + \frac{\partial w}{\partial z} = 0, \tag{3}$$

which results in

$$\frac{U}{L} \sim \frac{W}{H}, \tag{4}$$

where L is on the scale of 10^4 km, the width of the Equatorial Pacific basin. In addition, the scale of the zonal variation of u, which we denote as δU in the following, is similar to u itself, i.e., $\delta U \sim U$. Therefore, the left side of Eq. (2) scales as

$$\frac{W^2 L}{H^2}. \tag{5}$$

The right side of Eq. (2) corresponds to zonal temperature gradients via the hydrostatic balance and the ideal gas law:

$$\ln \frac{p}{p_s} = -\frac{g}{R_d} \int_0^H \frac{\mathrm{d}z}{T_v}, \tag{6}$$

where T_v is the height-dependent virtual temperature, and p is the pressure at height $z = H$. In order to arrive at a scaling, we approximate T_v as constant with height, which is typically valid when other variables vary more rapidly with height. This simplifies to the hypsometric equation:

$$\ln \frac{p}{p_s} = -\frac{gH}{R_d T_v}. \tag{7}$$

Taking the zonal derivative of Eq. (7) and ignoring the zonal variation in p_s, we get

$$\frac{\partial \ln p}{\partial x} = -\frac{gH}{R_d} \frac{\partial}{\partial x} \left(\frac{1}{T_v} \right), \tag{8}$$

leading to the scaling

$$\frac{\delta p}{p} \sim \frac{gH\delta T_v}{R_d T_v^2}. \tag{9}$$

Combining this with the ideal gas law, the right side of Eq. (2) scales as

$$\frac{gH\delta T_v}{LT_v}. \tag{10}$$

We now equate the scalings in Eq. (5) and Eq. (10) and arrive at the following scaling:

$$\frac{\delta T_v}{T_v} \sim \frac{W^2 L^2}{gH^3}. \tag{11}$$

This equation links the temperature gradient $\frac{\delta T_v}{T_v}$ to the overturning circulation strength W, horizontal length scale L, gravitational acceleration g, and the height of the troposphere H.

c. The importance of the large-scale aggregation of deep convection

The quasi-stationary structure of the tropical atmosphere reflects the large-scale distribution of deep convective heating. Any change in the geographic distribution of deep convection projects on the quasi-stationary structure and hence also on the temperature gradients in the free troposphere. The typical El-Nino like warming pattern over the tropical Pacific leads to an eastward expansion of deep convection, and thus to a weakening of the Walker cell and upper tropospheric temperature gradients over the Pacific. We address the question to what extent the surface warming pattern affects the temperature gradients in the free troposphere with targeted GCM experiments with prescribed SSTs.

3. Methods

a. General circulation model (GCM) simulations

We use the Geophysical Fluid Dynamics Laboratory (GFDL) Forecast-oriented Low Ocean Resolution version of CM2.5 (CM2.5-FLOR) (Vecchi et al. 2014) and its atmosphere model AM2.5 (Delworth et al. 2012) to conduct patterned and uniform SST warming simulations. The atmosphere and land components of CM2.5-FLOR uses a horizontal resolution of $0.25° \times 0.25°$ and 32 vertical levels, and the ocean and sea ice components use lower resolution. The greenhouse gas concentrations except CO_2 and aerosol emissions correspond to the conditions of the year 2000. We run the following experiments:

1. CM2.5-FLOR idealized CO_2 increase simulation. The CO_2 concentration starts from the observed value at the year 2000 and increases by 1% per year for 140 years (a quadrupling by the year 140). Both CO_2 concentration and SST are changing in this experiment with the coupled Atmosphere Ocean GCM. Averaged fields computed from the first and last 10 years are referred to as "present climate" and "warmer climate", respectively, and their difference is the response to the CO_2 forcing.

2. Atmospheric GCM simulations with AM2.5 with prescribed SSTs from the coupled GCM experiment (1) as oceanic boundary condition. The prescribed SSTs are the mean annual cycles of the first and last 10 years of the coupled experiment (1). Both experiments are integrated for 40 years to ensure equilibration, and the last 10 years of both experiments are averaged to obtain the Atmospheric GCM "present climate" and "warmer climate" states. The CO_2 concentration is fixed at the value of the year 2000. These atmospheric GCM climate states allow direct comparison with coupled GCM simulations in (1) and isolate the responses to SST warming under fixed CO_2 forcing, which helps us quantify the masked CO_2 forcing effect (details below).

3. In order to quantify the importance of the geographic structure of the SST increase in the coupled GCM simulations, the atmospheric GCM is run with prescribed SSTs from the "present climate" with a uniform increase corresponding to the global mean SST increase (approximately +3 K). The CO_2 concentration is fixed at the value of the year 2000. Comparison of the results with the true (patterned) warming of experiment (2) allows to quantify the impact of the SST warming pattern on the tropical free tropospheric temperature gradients.

b. Cloud resolving model (CRM) simulations

We use the System for Atmospheric Modeling (SAM, Khairoutdinov and Randall (2003)) version 6.11.5 cloud resolving model (CRM) for 2-D (longitude/height) mock Walker simulations without rotation similar to Kuang (2012) and Wofsy and Kuang (2012). The model is nonhydrostatic, uses bulk microphycics and a simple Smagorinsky-type scheme for subgrid turbulence, and computes the surface sensible heat, latent heat and momentum fluxes based on the Monin–Obukhov similarity theory. The vertical grid has 64 levels, starting at 25 m and extending up to 27 km, and the vertical grid spacing increases from 50 m at the lowest levels to roughly 1 km at the top of the domain. The model has a rigid lid at the top with a wave-absorbing layer occupying the upper third of the domain to prevent the reflection of gravity waves. The domain width along the x-direction is $L = 10,240$ km with a 2-km horizontal resolution, and solid wall boundary conditions are employed at the two edges.

The SSTs are prescribed and linearly decrease by 8 K from the left boundary ($x = 0$) to the right boundary ($x = 10240$ km), mimicking the east-west SST gradient across the equatorial Pacific. Similar to previous studies (Kuang 2012; Wofsy and Kuang 2012), we use prescribed uniform radiative cooling rates Q_{rad} throughout the troposphere (where the temperature is warmer than 207.5 K), and use a Newtonian relaxation towards 200 K in the stratosphere (Pauluis and Garner 2006). The prescribed radiative cooling allows experiments to disentangle the effects of atmospheric humidity on latent heat and static stability, and on the atmospheric radiative cooling. We run two sets of simulations:

1. Simulations with fixed $Q_{\mathrm{rad}} = -1.7$ K/day and a domain average SST ranging from 294 K to 303 K with an increment of +1.5 K (i.e. 7 simulations to cover the range).

2. Simulations with the domain average SST fixed at 300 K and the radiative cooling Q_{rad} ranging from -2.9 K/day to -0.9 K/day with an increment of 0.2 K/day (i.e. 11 simulations to cover the range).

All simulations are run for 150 days and reach equilibrium after approximately 50 days. All our results below show averaged fields computed from the last 50 days of hourly model output.

4. GCM results

Throughout this section, we analyse the virtual temperature $T_v = T(1+0.61q)$ (i.e. including the effect of water vapor mixing ratio on density) at the 500 hPa pressure level. We focus on zonal temperature gradients close to equator where the Coriolis force is smallest, and the WTG approximation is most appropriate. The meridional gradients in the GCM simulations are discussed only to the extent necessary for the purpose of this paper. In order to avoid the additional complications due to off-equatorial latent heating particularly pronounced during the South Asian monsoon, we focus the following discussion on the results for the month of January; results for the annual mean fields are similar (see Supplement, Figure S2).

a. GCM simulations show weaker temperature gradients in a warmer climate

Figure 1(a) shows that the canonical structure (see also Bao et al. (2022)) of the 500 hPa virtual temperature of the present-climate, with spatial variability of order several Kelvin and temperature maxima over the regions of deep convection (for example, the equatorial Western Pacific warm pool and the Amazon). The global warming simulation retains the geographic structure of the present climate, but the warming structure Figure 1(b,d) reveals an anticorrelation to the anomaly structure of the base state: Regions that are warmer than the average in the base state experience less than average warming, and *vice versa*. Correspondingly, the width (Fig. 1c; quantified in terms of standard deviation) of the frequency distribution of the anomalies from the mean decreases in the ”warmer climate” compared to the ”present climate” simulation.

In the following, we test three possible mechanisms that could explain the decrease in the free tropospheric temperature gradient associated with global warming: Masked CO_2 forcing, changes in the geographic distribution of deep convection due to an El-Nino like SST warming pattern, and the weakening of the tropical diabatic circulation.

Masked CO_2 forcing. The first hypothesis is the masked CO_2 forcing. As pointed out by Merlis (2015), although the CO_2 concentration increase is homogeneous over the globe in the global warming simulations, the radiative forcing of CO_2 is not. In the convective regions such as the Western Pacific warm pool, the CO_2 radiative forcing is reduced, or “masked”, compared to subsidence regions, by deep-convective clouds and abundant water vapor (see also Zhang and Huang (2014)). Hence, one may hypothesize that this could induce larger free tropospheric warming in the subsiding regions (consistent with the warming pattern visible in Fig. 1(b)). Note that this mechanism is based on the impact on atmospheric radiative cooling, and not inhomogeneous radiative forcing of the surface energy balance. In order to test this hypothesis, We conduct a mechanism-denial model simulation, in which we force AM2.5 (the atmosphere model of CM2.5) with the SST increase from the CM2.5-FLOR CO_2 increase simulation but the CO_2 concentration is fixed at the value of the year 2000 (methods section). This simulation yields, compared to the

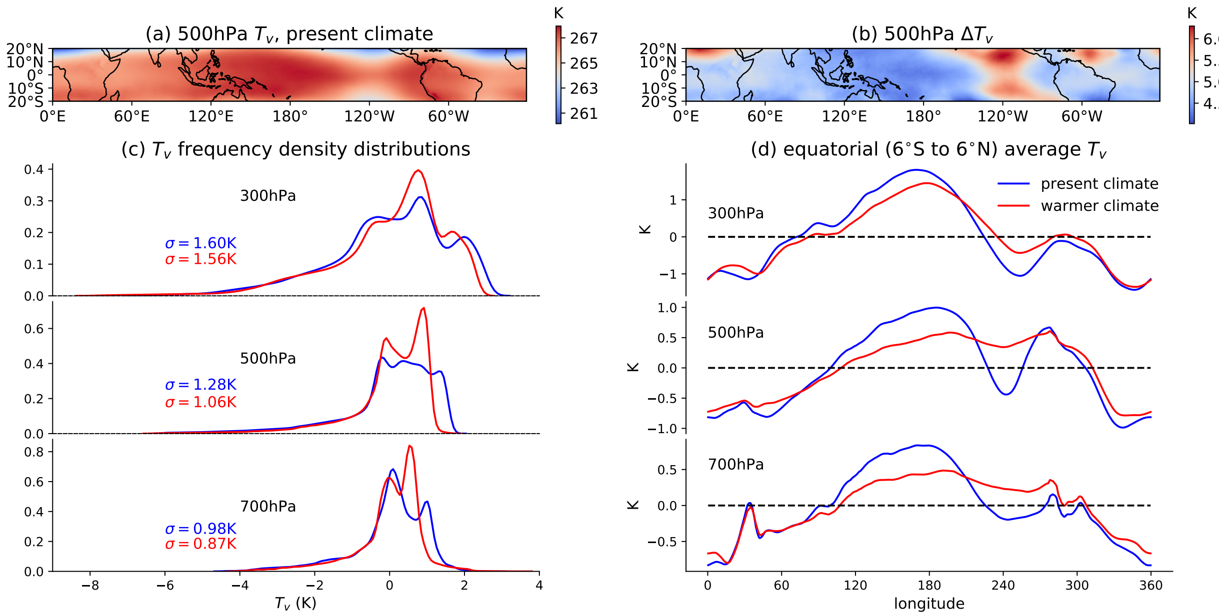

Fig. 1. Weaker temperature gradients in a warmer climate. (a) January climatological mean 500 hPa virtual temperature (T_v) in present climate (year 1-10 of the CM2.5-FLOR idealized CO_2 increase simulation). (b) Response of January climatological mean 500 hPa virtual temperature (ΔT_v) under global warming, calculated as the difference between warmer climate (year 131-140) and present climate (year 1-10). (c) The frequency distributions of T_v anomalies from tropical (20°N - 20°S) mean in present climate and warmer climate at 300 hPa, 500 hPa and 700 hPa levels, with their standard deviations σ listed. (d) The zonal profiles of T_v anomalies close to equator (meridional average between 6°N to 6°S) in present climate and warmer climate at 300 hPa, 500 hPa and 700 hPa levels.

reference simulation with the ”masked CO_2 forcing” (Fig. 1), a similar or even stronger weakening of the temperature gradients in both the frequency distribution (Fig. 2(a)) and the equatorial meridional mean (Fig. 2(b)).

El-Nino like SST warming. Coupled GCM simulations yield an El-Nino like warming pattern over the tropical Pacific in the future (Dong et al. 2019). That is, the cold eastern tropical Pacific is warming more than the warm western tropical Pacific. This leads to an eastward expansion of deep convection and a weakening of the Walker cell over the tropical Pacific. Observed SST trends in recent decades do not show this warming pattern, and there is debate to what extent Walker cell strength trends are due to the weakening of the diabatic atmospheric circulation (Vecchi et al. 2006) or related to patterned SST warming, and what may cause the difference in the warming

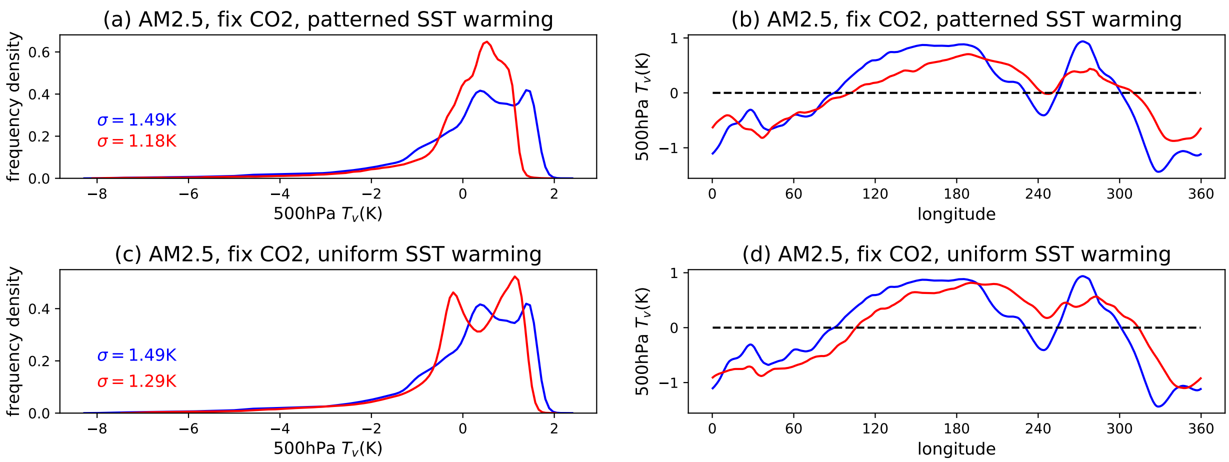

FIG. 2. Robust temperature gradient weakening across scenarios. (a) and (b) Same as figure 1 (c) and (d) but for AM2.5 forced by the patterned SST warming from CM2.5-FLOR idealized CO_2 increase simulation and with fixed CO_2 concentration. (c) and (d) Same as figure 1 (c) and (d) but for AM2.5 forced by uniform SST warming resulting the same global mean SST increase as the CM2.5-FLOR idealized CO_2 increase simulation and with fixed CO_2 concentration. See methods section for simulation details.

pattern between coupled GCMs and observations (e.g. Po-Chedley et al. (2021)). The impact of the El-Nino like warming pattern in coupled GCMs on the Walker cell - and hence also on the free tropospheric temperature structure (see also Kamae et al. (2015)) - is undisputed, and the question of interest here is whether this effect dominates, or just contributes, to the weakening of the temperature gradients shown in Figure 1.

In order to quantify the impact of the El-Nino like warming pattern, we conduct a second mechanism-denial experiment in which we force AM2.5 with a uniform SST increase corresponding to the global mean SST warming in the CM2.5-FLOR simulation. The results of this simulation are compared to the simulation with the patterned SST change; both simulations use the same CO_2 concentration (at the value of the year 2000). This ”uniform warming” simulation results in temperature gradient weakening in both frequency distribution (Fig. 2(c)) and equatorial zonal profile (Fig. 2(d)), that is smaller, but of comparable magnitude (in terms of reduction of standard deviation), to the ”patterned warming” simulation. Thus, the patterned SST warming trend amplifies the weakening of the temperature gradients, but is not the dominant reason of weaker temperature gradients in a warmer climate: The temperature gradients also decrease substantially under uniform warming.

Slow-down of the circulation. The weaker scaling of radiative cooling compared to the scaling of the static stability under global warming implies a a slow-down of the circulation (Held and Soden 2006; Vecchi and Soden 2007). This slow-down weakens the zonal momentum advection in the equatorial free troposphere, and a corresponding weakening of the pressure gradient as required by the zonal momentum balance - which is equivalent to a weaker temperature gradient (Fig. 1d). Having shown that "masked CO_2" forcing does not lead to, and the patterned warming contributes but is not the dominant reason for, weaker temperature gradients, we discuss the "circulation slow-down" mechanism in detail in the next section.

b. Weaker temperature gradients attributed to weaker circulation

To facilitate the analysis of GCM output on pressure levels, we shift to the pressure coordinate and we focus on geopotential $\Phi = gz$, as Φ and T_v are closely related if we rewrite the hydrostatic balance (equation 6) in pressure coordinate as

$$\int_{p_s}^{p} -R_d T_v \mathrm{d}\ln p = \int_0^z g\mathrm{d}z'. \tag{12}$$

That is, the 300 hPa geopotential height z and its response under global warming, Δz, have almost identical spatial pattern compared to 500 hPa virtual temperature (see Supplement, Figure S1(b,d)). The temperature gradient weakening in Fig. 1(c)(d) is also reflected in the pressure gradient weakening in Fig. S1(f)(h). Therefore, we demonstrate that the circulation slow-down decreases the temperature gradients by showing that the circulation slow-down leads to weaker pressure gradients due to the steady-state zonal momentum balance:

$$\frac{\partial \Phi}{\partial x} = -\vec{v}\cdot\nabla u + fv + r, \tag{13}$$

where the four terms represent pressure gradient force, momentum advection, Coriolis force and the residual term. As before, we focus on the near-equatorial zonal structure, and show the 10-year January averages of the present and warmer climate coupled GCM simulations. Not surprisingly, when close to equator, the dominant balance is between the pressure gradient force and the momentum advection term (i.e. $\frac{\partial \Phi}{\partial x} = -\vec{v}\cdot\nabla u$) in both present climate (first 10 years) and

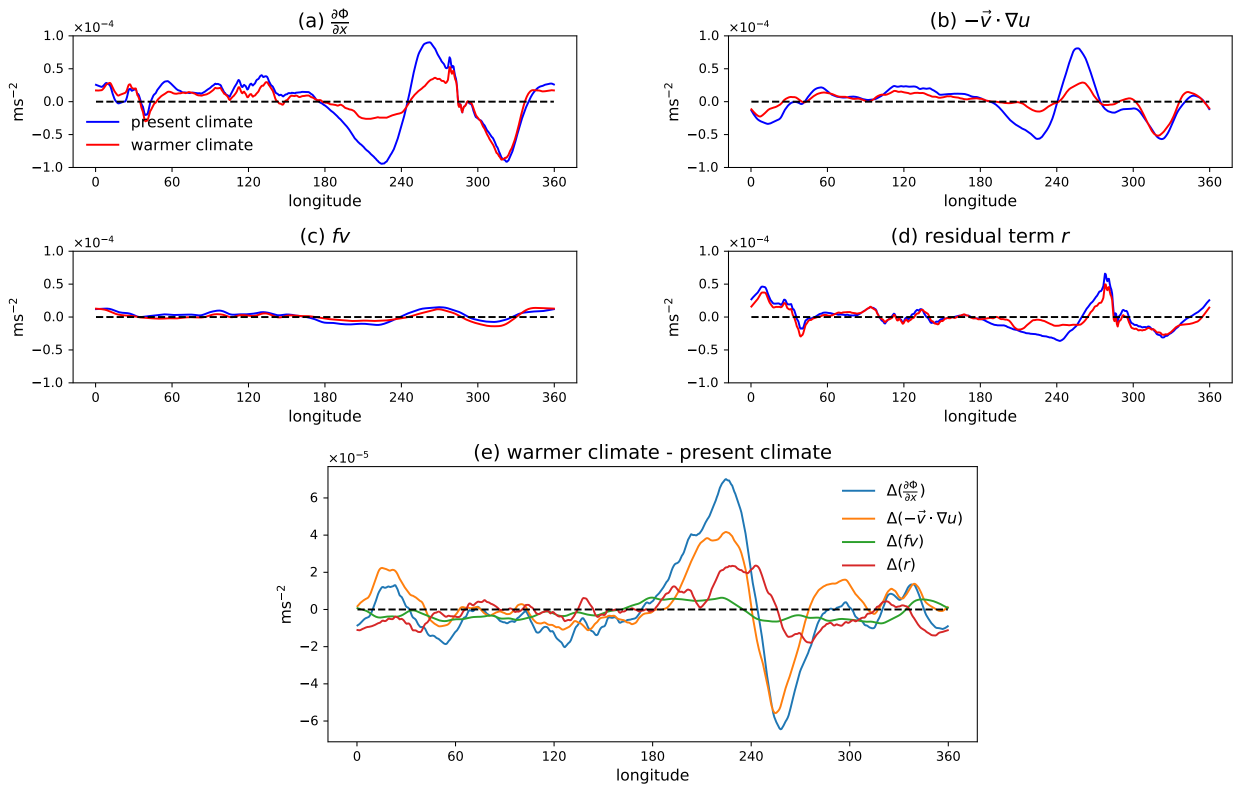

FIG. 3. 300hPa zonal momentum budgets close to equator (meridional average between 6°N to 6°S) for present (year 1-10) and warmer (year 131-140) climate in the CM2.5-FLOR idealized CO_2 increase simulation. (a) (minus) Pressure gradient force $\frac{\partial\Phi}{\partial x}$. (b) Zonal momentum advection $-\vec{v}\cdot\nabla u$. (c) Coriolis force fv. (d) The residual term r. (e) Responses (difference between warmer climate and present climate) of four terms to global warming. All terms are January averages in 10 years.

warmer climate (last 10 years) (figure 3(a)(b)), while the Coriolis force and the residual term are relatively small (figure 3(c)(d)).

Following global warming, the pressure gradient force and the momentum advection term become weaker in the equatorial Pacific (Figure 3(a)(b)). The response of the pressure gradient $\Delta\left(\frac{\partial\Phi}{\partial x}\right)$ is almost equal to the response of the momentum advection $\Delta\left(-\vec{v}\cdot\nabla u\right)$ due to the weaker circulation (Figure 3(e)), with a correlation over all longitudes of 0.91. Consistent with Bao et al. (2022), we find the reduction of the momentum advection $\Delta\left(-\vec{v}\cdot\nabla u\right)$ is dominated by $\Delta\left(-u\frac{\partial u}{\partial x}\right)$ (Supplement, Figure S4). Therefore, we attribute the weaker 300 hPa zonal pressure gradient as well as the weaker 500 hPa zonal temperature gradient close to equator to weaker momentum advection, which is a consequence of the weaker circulation (primarily weaker Walker circulation in equatorial Pacific) in a warmer climate.

Before analysing the relation between circulation strength and temperature gradients more quantitatively (Section 5), we briefly comment on the weakening of the meridional temperature and pressure gradients, which is particularly prominent over the subtropical Eastern Pacific and North Africa (Fig. 1(b)). In these regions, the Coriolis force is no longer negligible, and the reduction of the pressure gradient force $\Delta\left(-\frac{\partial\Phi}{\partial y}\right)$ in response to global warming is balanced by the reduction of the Coriolis force $\Delta(-fu)$ (figure S3), itself a consequence of weaker westerly wind. Future work may focus on this result, and its relation to the discussion of the response of the subtropical jet to global warming (Rivière 2011; Woollings et al. 2023).

5. Theoretical scalings and CRM results

In the following, we seek theoretical understanding using a simple model, aligned with the hierarchical approach (Held 2005). As mentioned before, the linear Matsuno-Gill model would be an obvious starting point due to its ability to reproduce the spatial pattern of 500 hPa ΔT_v. By converting the predicted change of convective heating (i.e. precipitation) to the forcing Q in the Gill model thermal equation (equation 2.8 in Gill (1980)), the tropical free troposphere temperature gradients are weaker in a warmer climate [1] (Keil et al. (2023), their Figure 5). However, the linear Matsuno-Gill model cannot give the correct explanation. According to the zonal momentum equation (equation 2.6 in Gill (1980)) in the Matsuno-Gill model, one would attribute a weaker pressure gradient $-\frac{\partial p}{\partial x}$ along the equator to a weaker momentum damping that is *linearly* proportional to the zonal wind, i.e. $-\epsilon \cdot u$. Below, we show that numerical model simulations do not follow the linear scaling inherent in the Matsuno-Gill model, but follow the (quadratic) scaling derived in Section 2.

Hence, we turn to a 2-dimensional (longitude/height) numerical model simulation in order to evaluate the theoretical scaling (Section 2, eq. 11) based on the zonal momentum equation with the dominant term $-\vec{v} \cdot \nabla u$. In these 2-D mock Walker cell simulations deep convection gradually becomes weaker away from the warm end, and is absent in the colder part of the domain.

[1] Note: Keil et al. (2023) did not adjust gravity wave phase speed (i.e. stratification) in the Gill model, which increases under global warming. Therefore, their results cannot be regarded as a "global warming" calculation.

The numerical experiments employ uniform radiative cooling rates Q_{rad} throughout the troposphere and linear SST profiles as shown in Fig. 4(a). This configuration is similar to the real-world equatorial Pacific and forces the majority of deep convection (and precipitation) to develop in the leftmost (warmest) 20% of the domain (Fig. 4(a)), resulting in a mock Walker circulation (Fig. 4(b)). The circulation strength is controlled by variation of the radiative cooling rate. In the limit where the steady-state thermodynamic energy equation is dominated by a balance of radiative cooling and vertical motion (typical in subsidence regions above cold SSTs), we can relate the strength of the vertical motion required by the scaling to the prescribed radiative cooling as

$$w = \frac{Q_{\mathrm{rad}}}{S} = \frac{Q_{\mathrm{rad}}}{\frac{\partial T}{\partial z} + \frac{g}{c_{\mathrm{p}}}}, \tag{14}$$

where $Q_{\mathrm{rad}} < 0$ is the radiative cooling rate in Ks^{-1} and $S = \frac{\partial T}{\partial z} + \frac{g}{c_{\mathrm{p}}}$ is the dry stability in Km^{-1}.

Equation 14 shows that, in addition to variations of the radiative cooling rate, the circulation strength can also be modified by variations in the static stability. As mentioned before, under global warming both parameters change, and the net slow down results from the static stability scaling being larger than the radiative cooling scaling with warming. Hence, we run two sets of experiments:

1. Variation of the domain average SST from 294 K to 303 K with a step of 1.5 K., while fixing the radiative cooling rate Q_{rad} at -1.7 K/day. Higher SSTs lead to higher subcloud specific humidity, which increases the static stability S. Hence, at fixed radiative cooling the simulation with higher SSTs is expected to show a weaker circulation (see also Appendix A, Figure A1).

2. Variation of the radiative cooling rate Q_{rad} from -2.9 K/day to -0.9 K/day with a step of 0.2 K/day, while fixing the domain average SST at 300 K to fix stability S. At fixed SSTs and hence fixed S, the simulation with a smaller radiative cooling rate is expected to show a weaker circulation (see also Appendix A, Figure A2).

From the hydrostatic equation 6, we define an effective tropospheric virtual temperature T_v^* to represent the free troposphere temperature as:

$$T_v^* = -\frac{gH}{R_d \ln \frac{p}{p_s}}, \tag{15}$$

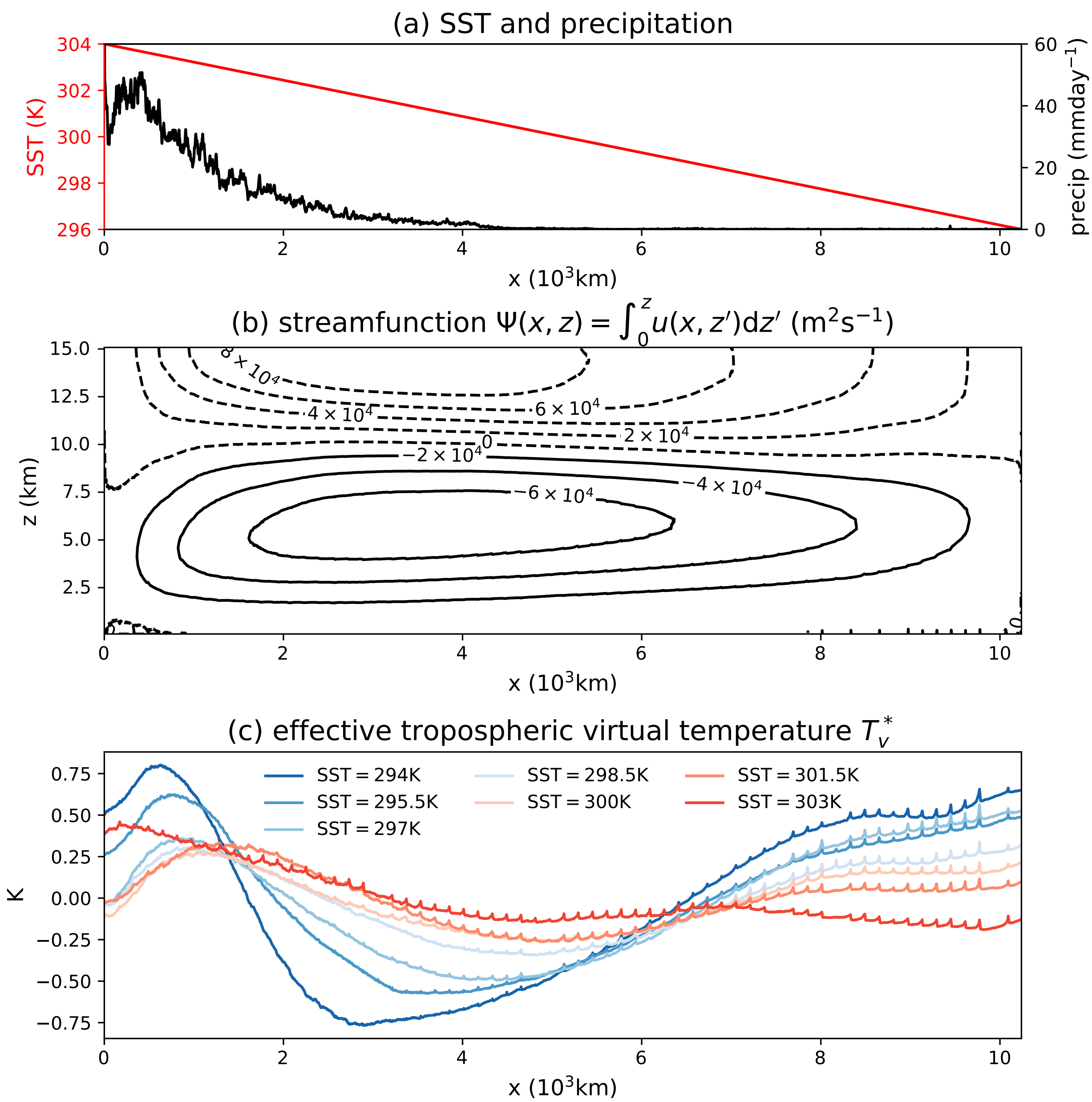

FIG. 4. (a) SST and precipitation profiles and (b) the overturning streamfunction for the simulation with $Q_{\rm rad} = -2.1$ K/day and mean SST of 300 K. (c) The effective tropospheric virtual temperature T_v^* (details in text) anomalies (from averages along x direction) in SAM simulations with fixed radiative cooling rate $Q_{\rm rad} = -1.7$ K/day but different mean SSTs.

which is the virtual temperature of an isothermal atmosphere with the same pressure-height relation at pressure p (in the upper troposphere) as the (not isothermal) atmosphere in the numerical model simulation.

For the simulations with varying radiative cooling but equal SSTs, the equivalent virtual temperature is evaluated for $H = 10\,\mathrm{km}$. For the simulations with varying SSTs but equal radiative cooling, the height level is determined as the level where the (horizontal) average temperature $\langle T \rangle = 220\,\mathrm{K}$. See Appendix A for motivation and further details. Actual virtual temperature at different levels from the $z = 5\,\mathrm{km}$ level to the $z = 10\,\mathrm{km}$ level show similar results as the effective tropospheric virtual temperature T_v^* (Supplement, Figure S6). Hence, the effective virtual temperature is a convenient and accurate parameter.

Figure 4(c) shows that, as expected, the simulations with higher SSTs have a weaker circulation, and a weaker temperature gradient in the free troposphere (i.e. a flatter T_v^* profile). Similarly (not shown), the simulations with equal SSTs but varying radiative cooling rate show a weaker circulation and a flatter T_v^* profile for the simulations with smaller radiative cooling. Qualitatively, the 2-D CRM results are consistent with the GCM results and confirm that a weaker circulation is associated with weaker temperature gradients in free troposphere. Thus, the idealized mock Walker cell simulations can be used to quantitatively evaluate the scaling between circulation strength and temperature gradients.

The horizontal temperature gradient is expected to scale with the square of subsiding velocity (W^2) all else held fixed (Equation 11). Replacing W in equation 11 with the peak subsiding velocity w_{peak} in the subsiding branch (details in appendix A) gives

$$\langle | \frac{\partial T_v^*}{\partial x} | \rangle \propto w_{\mathrm{peak}}^2, \tag{16}$$

where $\langle | \frac{\partial T_v^*}{\partial x} | \rangle$ is the average (absolute) horizontal T_v^* (eq. 15) gradient along the x-direction. The proportionality coefficient in equation 16 is given by equation B8 in the Appendix B.

Figure 5(a) shows that the theoretical relation in equation 16 (curves) is consistent with the model results (dots), both when we change Q_{rad} with fixed SST (blue) and change SST with fixed Q_{rad} (red). Consistent with equation 14, Figure 5(b) shows that $\langle | \frac{\partial T_v^*}{\partial x} | \rangle \propto Q_{\mathrm{rad}}^2$ when varying Q_{rad} with fixed SST, and Figure 5(c) shows that $\langle | \frac{\partial T_v^*}{\partial x} | \rangle \propto S^{-2}$ when varying SSTs with fixed Q_{rad}. Therefore, the mock Walker cell numerical model simulations quantitatively confirm that the weaker circulation causes weaker temperature gradients in tropical free troposphere, and that the temperature gradient scales with the *square* of circulation strength (i.e. w_{peak}^2). Using the linear Matsuno-Gill model (e.g. Keil et al. (2023)) to analyse the relation between circulation and temperature gradients, and

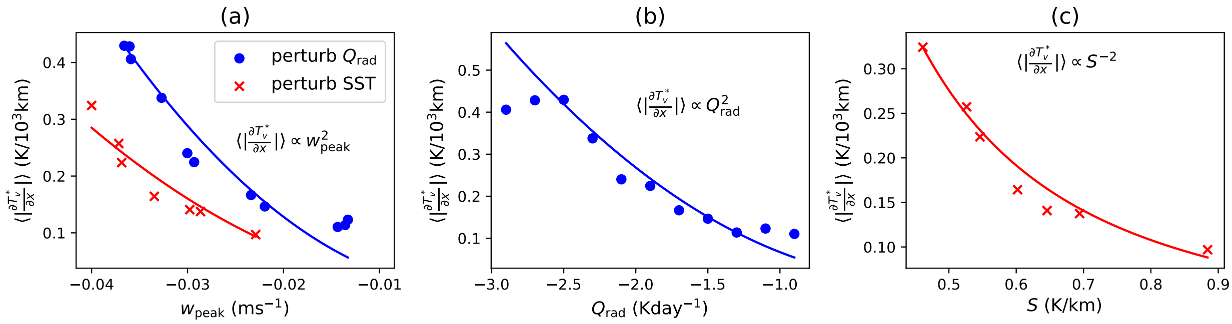

FIG. 5. Weaker circulation causes weaker temperature gradients in CRM simulations. (a) The average (absolute) horizontal T_v^* (defined in equation 15) gradient along x direction, $\langle|\frac{\partial T_v^*}{\partial x}|\rangle$, as a function of the peak subsiding velocity $w_{\rm peak}$ in the subsiding branch (details in appendix A) when we change $Q_{\rm rad}$ under fixed SST (blue) and change SST under fixed $Q_{\rm rad}$ (red). The dots are model results, and the curves are parabolas going through the origin (equation B8 in appendix B). (b) Same as (a) but showing $\langle|\frac{\partial T_v^*}{\partial x}|\rangle$ as a function of $Q_{\rm rad}$ when we change $Q_{\rm rad}$ under fixed SST. (c) Same as (a) but showing $\langle|\frac{\partial T_v^*}{\partial x}|\rangle$ as a function of stability S when we change SST under fixed $Q_{\rm rad}$.

how it may change under global warming, gives the incorrect result that the temperature (pressure) gradient scales *linearly* with circulation strength (i.e. U or W) because the pressure gradient force $-\frac{1}{\rho}\frac{\partial p}{\partial x}$ is balanced by the linear momentum damping term $-\alpha u$ ((equation 2.6 in Gill (1980)).

Finally, we note that the $\langle|\frac{\partial T_v^*}{\partial x}|\rangle \propto w_{\rm peak}^2$ theory seems to hold only for a certain range of the circulation strength. Figure 5(b) shows that when the radiative cooling is smaller (in magnitude) than -1.5 K/day or larger (in magnitude) than -2.5 K/day, the scaling is no longer accurate. Inspection of the circulation structure (Supplement, Figure S7) shows that the simulations show a regime shift, changing from a single cell to a double cell when $Q_{\rm rad}$ becomes smaller (in magnitude) than -1.5 K/day. This regime shift has also been noted in similar situations by Lutsko and Cronin (2023). The scaling and its evaluation at a height of ≈ 10 km applies to the situation of a single overturning cell with 10 km being robustly in the upper troposphere (see Appendix B). The cause for the regime shift, and adapting the scaling and its evaluation to multiple cells, are beyond the scope here, but are important questions for future research.

6. Conclusions and outlook

Due to the smallness of the Coriolis parameter at low latitudes, the tropical free troposphere cannot maintain temperature gradients as large as at higher latitudes (Charney 1963). Using theory and a hierarchy of models, we demonstrate that horizontal temperature gradients in the tropical free troposphere will be even weaker in a warmer climate. This is because the magnitude of the temperature gradients scale with the circulation strength, and the circulation strength decreases with global warming (Held and Soden 2006; Vecchi and Soden 2007). The weaker circulation corresponds to weaker horizontal momentum advection, which then causes weaker horizontal pressure gradients in the tropical free troposphere as required by the steady-state zonal momentum equation. Due to hydrostatic balance, the weaker pressure gradients correspond to weaker temperature gradients. This theoretical expectations is confirmed by the GCM simulations and quantitatively verified by the CRM mock Walker circulation simulations shown in this paper.

Linear models, such as the Matsuno-Gill model forced by predicted change of convective heating (i.e. precipitation) and stratification also yield a weakening of the temperature gradient in the free troposphere due to the weakening of the circulation. However, the scaling would not be correct as the Matsuno-Gill model employs a linear momentum damping, whereas in reality the zonal momentum equation is quadratic. The 2-D mock Walker circulation model simulations shown here confirm the quadratic scaling.

Our results have implications for both idealized modeling and theoretical understanding of the tropical atmosphere. Parameterizations of large-scale dynamics in SCMs and CRMs need to correctly represent the weakening of temperature gradient in simulations of global warming. Our results highlight the importance of the non-linear momentum advection term for the understanding of changes in atmospheric dynamics with global warming and climate change in general. Direct implications of the weakening of the temperature gradient in tropical free troposphere include processes that depend on the static stability of the lower troposphere, including heat extremes (Sherwood and Huber 2010; Zhang et al. 2021) and low cloud amount (Klein and Hartmann 1993; Ceppi and Gregory 2017; Fueglistaler 2019).

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lation results. Computations in this research are done on the Tiger cluster and Stellar cluster of Princeton University.

Data availability statement. The model outputs and data used in this study are available upon request.

APPENDIX A

Two ways to vary the strength of the mock Walker circulation

The circulation strength (i.e. subsidence velocity) may be modified (eqn. 14) by either changing SSTs (to change static stability S) or by changing the radiative cooling rate Q_{rad}.

When increasing SSTs under fixed Q_{rad}, the subsiding velocity w indeed becomes smaller (Figure A1a). This is because the stability $S = \frac{\partial T}{\partial z} + \frac{g}{c_{\mathrm{p}}}$ in the free troposphere (vertically averaged between the $\langle T \rangle = 260\,\mathrm{K}$ level and the $\langle T \rangle = 220\,\mathrm{K}$ level) becomes larger in Fig. A1(b). The theoretical relation $w_{\mathrm{peak}} = \frac{Q_{\mathrm{rad}}}{S}$ (line) is consistent with the model results (dots) in Fig. A1(c). Note that in Figure A1(a)(b) we follow Jeevanjee (2022) and use temperature as the vertical coordinate. This is because the circulation structure (i.e. w in figure A1(a)) remains approximately fixed in temperature coordinates, so this choice simplifies our analysis when warmer SST deepens the troposphere.

Decreasing Q_{rad} under fixed SST, the subsiding velocity w becomes smaller (Figure A2a). In this experiment, the stability $S = \frac{\partial T}{\partial z} + \frac{g}{c_{\mathrm{p}}}$ near the peak subsiding velocity in Fig. A2(b) has a constant value $S_0 \approx 0.8\,\mathrm{K/km}$ for different radiative cooling rates. The theoretical relation $w_{\mathrm{peak}} = \frac{Q_{\mathrm{rad}}}{S_0}$ (line) is consistent with model results (dots) in Fig. A2(c) when Q_{rad} is between $-2.5\,\mathrm{K/day}$ and $-1.5\,\mathrm{K/day}$. When Q_{rad} is smaller (in magnitude) than $-1.5\,\mathrm{K/day}$ or larger (in magnitude) than $-2.5\,\mathrm{K/day}$, however, the subsiding velocity departs from the expected scaling with the radiative cooling rate. Inspection of the mass stream function shows that there is a regime shift (figure S7) where equation (14) is no longer accurate.

APPENDIX B

The theoretical relation between temperature gradient and subsiding velocity

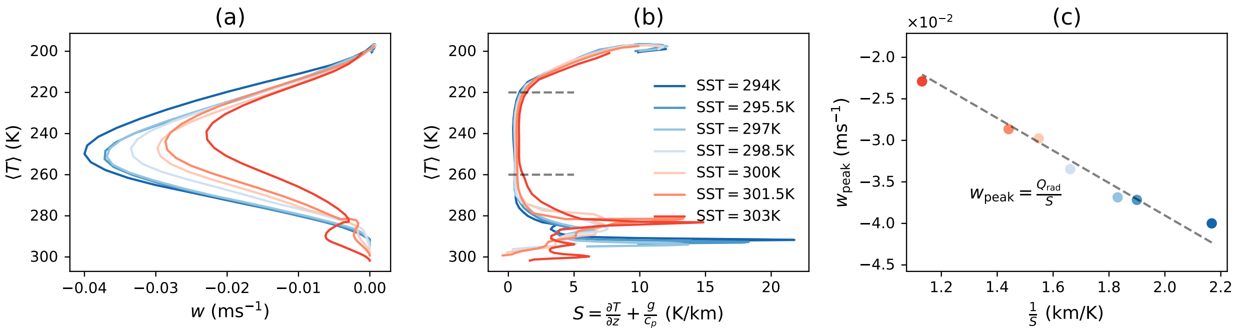

FIG. A1. Tropospheric vertical profiles of (a) vertical velocity w and (b) stability $S = \frac{\partial T}{\partial z} + \frac{g}{c_p}$ in the subsiding branch (i.e. average along x direction above the 10% coldest SSTs) of SAM simulations with the same radiative cooling rate $Q_{\text{rad}} = -1.7$ K/day but different mean SSTs. Notice here we use temperature as the vertical coordinate. (c) The peak subsiding velocity w_{peak} as a function of $\frac{1}{S}$, where S is averaged between the two dotted lines in (b) - the $\langle T \rangle = 260$ K level and the $\langle T \rangle = 220$ K level. The dotted line in (c) shows their predicted relation $w_{\text{peak}} = \frac{Q_{\text{rad}}}{S}$ based on theory (details in text).

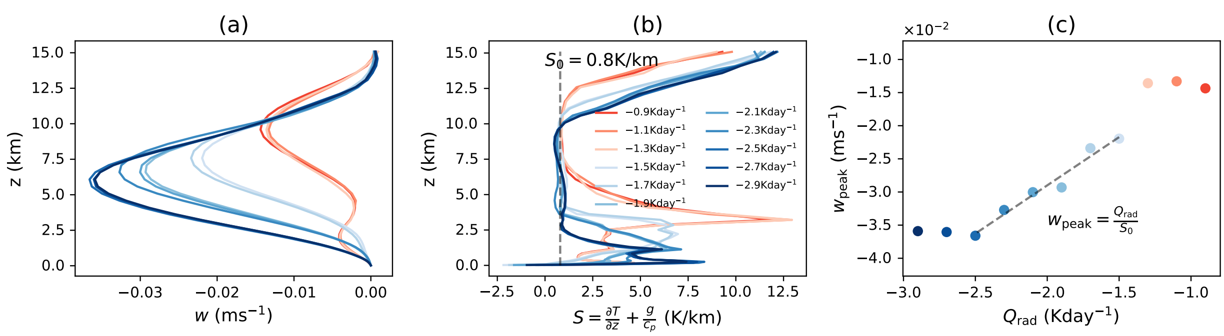

FIG. A2. Tropospheric vertical profiles of (a) vertical velocity w and (b) stability $S = \frac{\partial T}{\partial z} + \frac{g}{c_p}$ in the subsiding branch (i.e. average along x direction above the 20% coldest SSTs) of SAM simulations with the same mean SST of 300 K but different radiative cooling rates Q_{rad}. (c) The peak subsiding velocity w_{peak} as a function of the radiative cooling rate Q_{rad}. The dotted line shows their predicted relation $w_{\text{peak}} = \frac{Q_{\text{rad}}}{S_0}$ based on theory (details in text), where $S_0 \approx 0.8$ K/km is the constant stability in mid-troposphere for different radiative cooling rates as marked in (b).

First, we derive how u scales with w from the continuity equation. We assume that the subsiding vertical velocity w close to the cold end of the domain $x = L$ (Fig.A1(a) and Fig.A2(a)) has the highly idealized form of

$$w = w_{\text{peak}} \sin \frac{\pi z}{H}, \tag{B1}$$

where $H = 10$km when we change $Q_{\rm rad}$ under fixed SST and H is the level where $\langle T \rangle = 220\,$K when we change SST under fixed $Q_{\rm rad}$. We also assume the horizontal velocity u at the $z = H$ level (figure not shown) has the highly idealized form of

$$u = u_{\rm peak} \sin \frac{\pi x}{L}, \tag{B2}$$

where the domain width $L = 10240\,$km. Evaluating the continuity equation 3 at $(x, z) = (L, H)$ gives

$$u_{\rm peak} = -\frac{L}{H} w_{\rm peak}. \tag{B3}$$

The functional forms of this idealization are applicable only to cells with length scales L and H; the regime shift to double cells violates this idealization.

Second, we derive how pressure gradient scales with w from the momentum equation. Hourly CRM output enables us to decompose the momentum advection terms into stationary terms and eddy terms, and rewrite the momentum equation (13) as (Yang et al. 2013)

$$-\overline{\frac{1}{\rho}\frac{\partial p}{\partial x}} + \left(-\overline{u}\frac{\partial \overline{u}}{\partial x}\right) + \left(-\overline{w}\frac{\partial \overline{u}}{\partial z}\right) + \left(-\frac{\partial \overline{u'u'}}{\partial x}\right) + \left(-\frac{\partial \overline{u'w'}}{\partial z}\right) + \overline{r} = 0, \tag{B4}$$

where $\overline{()}$ is the time average. The six terms in this equation are plotted in the Supplement Figure S5 (a) to (f) for the simulation with $Q_{\rm rad} = -2.1\,$K/day and mean SST of 300 K. At the $z = H$ level in free troposphere, the dominant balance is between the pressure gradient force and the stationary horizontal advection of horizontal momentum (figure S5(g)), so we assume

$$-\overline{\frac{1}{\rho}\frac{\partial p}{\partial x}} + \left(-\overline{u}\frac{\partial \overline{u}}{\partial x}\right) = 0. \tag{B5}$$

Utilizing equation (B2) for u and neglecting density variation, the pressure gradient has the form

$$\frac{\partial p}{\partial x} = -\rho u_{\rm peak}^2 \frac{\pi}{2L} \sin \frac{2\pi x}{L}, \tag{B6}$$

so that the average (absolute) horizontal pressure gradient along the x-direction is

$$\langle|\frac{\partial p}{\partial x}|\rangle = \frac{\rho}{L}u_{\text{peak}}^2 = \frac{\rho L}{H^2}w_{\text{peak}}^2. \tag{B7}$$

Finally, we derive the temperature gradient scaling with w from the hydrostatic equation. We compute the x-derivative of the effective tropospheric virtual temperature T_v^* defined in equation 15. Neglecting the horizontal variation of p_s (we would get an additional constant term because p_s gradient is tied to SST gradient and approximately fixed), we can relate $\langle|\frac{\partial T_v^*}{\partial x}|\rangle$ to $\langle|\frac{\partial p}{\partial x}|\rangle$ by

$$\langle|\frac{\partial T_v^*}{\partial x}|\rangle = \frac{gH}{R_d}\frac{1}{p\cdot(\ln p/p_s)^2}\langle|\frac{\partial p}{\partial x}|\rangle = \frac{g}{R_d}\frac{1}{p\cdot(\ln p/p_s)^2}\frac{\rho L}{H}w_{\text{peak}}^2. \tag{B8}$$

This equation tells us the coefficient in equation 16. Using $w_{\text{peak}} = \frac{Q_{\text{rad}}}{S}$, we can also get the coefficients for $\langle|\frac{\partial T_v^*}{\partial x}|\rangle \propto Q_{\text{rad}}^2$ and $\langle|\frac{\partial T_v^*}{\partial x}|\rangle \propto S^{-2}$. Figure 5 shows that the theoretical relations derived here (curves) are consistent with model results (dots).

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