A Single-Column Simulation-Based Decomposition of the Tropical Upper-Tropospheric Warming

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ABSTRACT: An atmospheric global climate model (GCM) and its associated single-column model are used to study the tropical upper-tropospheric warming and elucidate how different processes drive this warming. In this modeling framework, on average the direct radiative process accounts for 13% of the total warming. The radiation increases the atmospheric lapse rate and triggers more convection, which further produces 74% of the total warming. The remaining 13% is attributable to the circulation adjustment. The relative importance of these processes differs in different regions. In the deep tropics, the radiative–convective adjustment produces the most significant warming and accounts for almost 100% of the total warming. In the subtropics, the radiative–convective adjustment accounts for 73% of the total warming and the circulation adjustment plays a more important role than in the deep tropics, especially at the levels above 200 hPa. When the lateral boundary conditions (i.e., the temperature and water vapor advections) are held fixed in single-column simulations, the tropospheric relative humidity significantly increases in the radiative–convective adjustment in response to the surface warming. This result, in contrast to the relative humidity conservation behavior in the GCM, highlights the importance of circulation adjustment in maintaining the constant relative humidity. The tropical upper-tropospheric warming in both the full GCM and the single-column simulations is found to be less strong than the warming predicted by reference moist adiabats. This evidences that the sub-moist-adiabatic warming occurs even without the dilution effect of the large-scale circulation adjustment.

KEYWORDS: Climate change; Climate models; Tropical variability

1. Introduction

Tropical upper-tropospheric warming (TUTW) is a prominent feature observed in both observational datasets (Fu et al. 2011; Po-Chedley and Fu 2012; Po-Chedley et al. 2015; Santer et al. 2017; Suárez-Gutiérrez et al. 2017; Tuel 2019) and model projections of the twenty-first-century climate change (Bony et al. 2006; Collins et al. 2013; Santer et al. 2005; Flannaghan et al. 2014). The warming is horizontally uniform in the tropics and reaches its maximum at around 200 hPa (Fig. 1a).

TUTW has many influences on other processes in climate change. The warming amplification increases the emission of longwave flux, constituting a negative radiative feedback, which contributes to the slower surface warming rate at the equator than at the Arctic (Pithan and Mauritsen 2014). As the relative humidity holds approximately constant in climate change, the positive water vapor feedback is enhanced by TUTW following the Clausius–Clapeyron relationship (Bony et al. 2006; Held and Soden 2006). Concerning the dynamics, the TUTW increases the temperature gradient between the tropics and high latitudes in the upper troposphere, shifting the jets and storm tracks poleward (Yin 2005) and strengthening the Brewer–Dobson circulation (BDC; Garcia and Randel 2008).

The causes of the TUTW have been traced to deep convections and lapse-rate change (Knutson and Manabe 1995; Flannaghan et al. 2014; Fueglistaler et al. 2015; Andrews and Webb 2018; Tuel 2019). Being part of the complex tropical climate system, deep convection inevitably interacts with the radiative processes (Li et al. 2019) and large-scale circulations

(Charney 1963). Based on the previous findings, the mechanism for TUTW can be hypothesized as follows. We can think of the tropics as two boxes: convecting box and nonconvecting box. In the convecting box, deep convection tightly couples the surface with the upper troposphere. As sea surface temperature (SST) warms, deep convection is enhanced with more latent heat release which efficiently increases the temperature in the upper troposphere. In the nonconvecting box, the SST perturbation is largely confined within the boundary layer and lower troposphere; the warming in the upper troposphere is thus small. Because of the small Coriolis force in the tropics, the temperature gradient produced by the radiative–convective adjustment cannot be maintained in the free atmosphere. In consequence, the large-scale circulation transports the heat from the convecting box to the nonconvecting box, resulting in a uniform TUTW.

The aims of this paper are 1) to validate the above hypothesis in the climate model and 2) to elucidate how different processes (i.e., radiation, convection, and large-scale circulation) drive the TUTW. Specifically, we will address the questions of how much of the TUTW could be explained by the local radiative–convective adjustment alone and how the large-scale circulation modifies the pattern of the radiative–convective adjustment and contributes to the TUTW.

To accomplish our goals, we use a GCM together with its single-column model (SCM) to separate the effects of different processes on the TUTW in a set of simulation experiments, following a similar method applied by Wang and Huang (2020) to analyze the atmospheric adjustment to the CO₂ forcing. In this work, we use prescribed SST as the forcing to generate the TUTW. As detailed in section 2, we determine the atmospheric warming due to radiative, radiative–convective, and circulation

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FIG. 1. Temperature changes in response to the warming SST. (a) Overall temperature change, obtained from Exp. CAM_S4 and CAM_S1. (b) Temperature change driven by the radiative process, obtained from Exp. PORT_S4 and CAM_S1. (c) Temperature change driven by the convective process, obtained from Exp. SCM_S4C1 and PORT_S4. (d) Temperature change driven by the circulation adjustment, obtained from Exp. SCM_S4C4 and SCM_S4C1. (e) Temperature change driven by the radiative-convective adjustment, obtained from Exp. SCM_S4C1 and CAM_S1. The black boxes, ranging from 30°S to 30°N and 100 to 350 hPa, indicate the region where tropical upper-tropospheric warming is quantified. Units: K.

TABLE 1. Overview of experiments. SST1 is generated from the CCSM4 piControl experiment in CMIP5. SST4 is generated from CCSM4 abrupt4×CO₂ experiment in CMIP5. LST1 is the land surface temperature archived from the CAM_S1 experiment. LST4 is the land surface temperature archived from the CAM_S4 experiment.

Experiment	Atmospheric model	Sea surface temperature	Land surface temperature	Radiation	Convection	Circulation
CAM_S1	CAM5	SST1	Active	Active	Active	Active
CAM_S4	CAM5	SST4	Active	Active	Active	Active
PORT_S4	PORT	SST4	LST4	Active	Prescribed-CAM_S1	Prescribed-CAM_S1
SCM_S1C1	SCAM5	SST1	LST1	Active	Active	Prescribed-CAM_S1
SCM_S4C1	SCAM5	SST4	LST4	Active	Active	Prescribed-CAM_S1
SCM_S4C4	SCAM5	SST4	LST4	Active	Active	Prescribed-CAM_S4

adjustments by forcing the atmospheric column with component temperature tendencies obtained from the full GCM simulations. First, we suppress the convective and circulation adjustments by fixing the tendencies due to the convection and the circulation and thus allow only the radiative process to adjust the atmospheric temperature to respond to the SST warming. The radiative adjustment is found to increase convective available potential energy (CAPE) in the atmosphere. Then, convective adjustment is allowed to join the radiative adjustment in the single-column simulation, which is found to consume the CAPE to restore the local radiative–convective equilibrium. The effect of the circulation adjustment is determined by comparing the radiative–convective adjustment simulated by the SCM with the full GCM simulation. We note that the term “adjustment” here refers to the atmospheric response to the prescribed SST warming, not to be confused with the rapid adjustment associated with the CO₂ radiative forcing.

2. Methods and model descriptions

In this paper, we aim to elucidate the mechanisms controlling TUTW and focus on the effects of atmospheric processes, including radiation, convection, and circulation. This justifies the use of the prescribed-SST approach in our simulation experiments. Because TUTW is driven by the warming SST rather than the direct CO₂ radiative effect, we fix the CO₂ concentration to 1138.8 ppmv, corresponding to the 4 × CO₂ condition often used in global warming simulations. The climate model used to simulate the full atmospheric response with all the processes involved is CAM5 (Neale et al. 2012). It is configured on the Eulerian spectral grids with a T42 wavenumber truncation (64 grids in latitude and 128 grids in longitude). There are 30 vertical levels with the model top at around 3 hPa. The choice of physical parameterizations includes the RRTMG radiation scheme (Mlawer et al. 1997), the University of Washington shallow convection scheme (Park and Bretherton 2009), and the Zhang–McFarlane deep convection scheme (Zhang and McFarlane 1995).

The sea surface temperatures and sea ice concentrations (SICs) prescribed in CAM5 runs are obtained from CCSM4 (Community Earth System Model version 4) simulations in CMIP5 dataset (Taylor et al. 2012). The control SST (SST1 in Table 1) is monthly mean data averaged over the last 30 years (out of 1300 years) in the piControl experiment; the perturbed

SST (SST4 in Table 1) is monthly mean data averaged over the last 10 years (out of 150 years) in the abrupt4×CO₂ experiment.

The CAM5 experiments, as well as the radiative adjustment experiment and the radiative–convective adjustment experiment explained below, are summarized in Table 1. The overall adjustment is defined as the difference between experiment (hereinafter Exp.) CAM_S4 and CAM_S1; the adjustment driven by the radiative process is defined as the difference between Exp. PORT_S4 and CAM_S1; the adjustment driven by the convective process is defined as the difference between Exp. SCM_S4C1 and PORT_S4; and the adjustment driven by the circulation change is defined as the difference between Exp. SCM_S4C4 (equivalent to CAM_S4) and SCM_S4C1.

a. Radiative adjustment

We use the Parallel Offline Radiative Transfer model (PORT; Conley et al. 2013) to simulate the atmospheric temperature change in radiative adjustment. PORT is a special configuration of CAM, in which only radiation modules are called. The original PORT distributed with CAM is designed to be compatible with CAM4 and the radiation scheme was an obsolete radiation model CAMRT. In this work, we replace CAMRT with RRTMG to make the PORT consistent with CAM5. It is validated that PORT can accurately reproduce the heating rates and the radiative fluxes output from CAM5.

The temperature change at a specific time step in radiative adjustment is computed as

$$\left(\frac{dT}{dt}\right)_i = H(T_i + \Delta T_i, TS4) - H(T_i, TS1). \quad (1)$$

The term ΔT_i is the temperature change driven by the radiative process at the i th integration; $H(T_i, TS1)$ is the radiative heating rate at atmospheric temperature (T_i) and surface temperature (TS1). This heating rate, which describes the control climate, is output from experiment CAM_S1 (Table 1). The term $H(T_i + \Delta T_i, TS4)$ is the radiative heating rate calculated by PORT. The physical interpretation of the radiative adjustment is that the atmosphere adjusts its temperature (ΔT) to compensate for the flux imbalance perturbed by the surface temperature change (from TS1 to TS4). The radiative temperature adjustment at the next time step is obtained by adding the temperature change driven by the radiative heating rate difference:

$$\Delta T_{i+1} = \Delta T_i + \left(\frac{dT}{dt} \right)_i \times \Delta t. \quad (2)$$

The time step in PORT, Δt , is set to be 1 h. The boundary conditions of surface temperature (TS4) are output from experiment CAM_S4. The surface temperature here consists of both the SST and the land surface temperature (LST). The background atmospheric temperatures (T_i) are archived from experiment CAM_S1. Both TS4 and T_i are output at an hourly frequency. Other variables influencing the radiative transfer calculation, such as the water vapor contents and the cloud fractions, are output together with T_i from experiment CAM_S1. PORT reads in the archived data and inputs the data to the radiation module at each time step. The simulation reaches the equilibrium after a half-year integration, characterized by a near-zero heating rate difference [Eq. (1)]. We spin up the PORT model for 1 year and run it for another year for the analysis.

b. Radiative–convective adjustment

The single-column atmosphere model (SCAM; Gettelman et al. 2019) is used to compute the temperature change driven by the radiative–convective process. SCAM contains identical physical parameterizations to the standard CAM but simplifies the dynamical forcing (Gettelman et al. 2019). Most of the moist processes are incorporated in SCAM, including deep convection, shallow convection, microphysics, and macrophysics. The water vapor tendencies and temperature tendencies associated with large-scale circulation are prescribed in lateral boundary conditions. The lateral boundary conditions also include the prescribed cloud advection and aerosol advection.

To reproduce the identical results of CAM5 from SCAM, we output the advective tendencies from experiment CAM_S1 at each time step (20 min). SCAM reads the tendencies as the lateral boundary condition and performs the column atmospheric physical calculation. We have verified the identical results between CAM5 and SCAM. So experiments SCM_S1C1 and SCM_S4C4 can reproduce the same results as the experiments CAM_S1 and CAM_S4. To reduce the demand for the storage of the offline tendency data, we subsample the tendency data at reduced 32×32 (latitude $\times$ longitude) grids out of the original 64×128 grids. SCAM is run only at the subsampled grids. We have validated that most of the zonal mean features concerned here agree well with those calculated on the original grids.

The surface temperatures used to force the SCAM perturbation run, archived from experiment CAM_S4, are the same as those in the radiative adjustment experiment PORT_S4. Given that the time scale of the radiative–convective equilibrium is hundreds of days (Cronin and Emanuel 2013), the SCAM is spun up for 5 years and run for another 5 years for the analysis.

c. Moist adiabat

As an air parcel is lifted from the lowest atmospheric level, it follows the dry adiabatic process first until it reaches the lifted condensation level (LCL), and then follows the moist adiabatic process. The initial condition of the rising parcel is set by the

annual mean 2-m atmospheric temperature, 2-m relative humidity, and surface pressure.

The dry adiabatic lapse rate below the LCL is

$$\Gamma_d = \frac{g}{c_{pd}} \quad (3)$$

and the moist adiabatic lapse rate above the LCL is

$$\Gamma_m = \Gamma_d \frac{1 + \frac{L_v r_v}{RT}}{1 + \frac{L_v^2 r_v}{c_{pd} R_v T^2}}, \quad (4)$$

where c_{pd} is the specific heat capacity, L_v is the latent heat of vaporization, r_v is the water vapor mixing ratio, R is the specific gas constant of dry air, R_v is the specific gas constant of water vapor, and T is temperature. The above moist-adiabatic lapse rate assumes all condensates precipitate out immediately and $r_v \ll 1$. It does not include the latent heat of the fusion either. Miyawaki et al. (2020) showed that these two assumptions do not significantly change the results.

3. Results

a. Temperature

Temperature changes driven by different processes are shown in Fig. 1. Figure 1a depicts the overall zonal mean temperature change simulated by standard CAM runs (Exp. CAM_S4 – CAM_S1). The TUTW extends from 400 hPa up to the tropopause around 100 hPa in the vertical direction and from 30°S to 30°N in the latitudinal direction, with the maximum centered around 200 hPa. The largest warming is about 7.5 K, within the range of results shown in the CMIP5 model simulations (Huang et al. 2016).

The radiative impact on temperature change is largely confined to the boundary layer (Fig. 1b; Exp. PORT_S4 – CAM_S1). The two maximum centers are over the Arctic and the oceans around the Antarctic, where the sea ice melts most. The radiative contribution to the free troposphere is uniform with values mostly less than 1 K. The atmospheric lapse-rate changes in the western Pacific, averaged over 10°S–10°N, 120°–160°E, are shown in Fig. 2. The lapse rate (absolute value) increases throughout the whole troposphere, especially in lower levels. According to the classic view of convective adjustment described in Manabe and Wetherald (1967), convection will adjust the excessive atmospheric lapse rate back to the moist-adiabatic lapse rate by moving energy from the lower layer to the upper layer. Although the direct temperature contribution of radiation to the TUTW is not significant, radiation is an important trigger for the convective and the circulation adjustments, as is described in more detail below.

The temperature change driven by convection discloses a triangle-shaped structure (Fig. 1c; Exp. SCM_S4C1 – Exp. PORT_S4); the convective warming extends from the near-surface to the upper troposphere, with a noticeably higher reach up to 70 hPa and larger magnitude in the deep tropics (15°S–15°N). The convection decreases the atmospheric lapse

FIG. 2. Lapse-rate change driven by different processes, averaged over the western Pacific, ranging from 10°S to 10°N and from 120° to 160°E. (a) Lapse-rate change driven by the radiation, obtained from Exp. PORT_S4 and CAM_S1. (b) Lapse-rate change driven by the convection, obtained from Exp. SCM_S4C1 and PORT_S4. (c) Lapse-rate change driven by circulation adjustment, obtained from Exp. SCM_S4C4 and SCM_S4C1. The lapse-rate change plotted here is the change in its absolute value.

rate from the near surface up to 150 hPa (Fig. 2b), manifesting an offsetting effect to the radiation. Figure 1e shows the radiative–convective adjustment simulated by the single-column model (Exp. SCM_S4C1 – SCM_S1C1). The pattern noticeably differs from the overall warming pattern (Fig. 1e vs Fig. 1a). The warming driven by radiative–convective adjustment is narrower in the upper troposphere and does not extend to the subtropics. In the subtropics above 200 hPa, the convective adjustment does not show noticeable warming, in sheer contrast to the warming amplification in the total temperature change (Fig. 1a). This result is consistent with our hypothesis (see the introduction) that without circulation adjustment, the deep tropics where deep convection occurs more frequently has higher TUTW than the subtropics.

To understand how different processes coordinate to arrive at a radiative–convective equilibrium state, we plot the heating rate decompositions in Fig. 3. The standard heating rates output from CAM and SCAM are the longwave radiative heating rate, the shortwave radiative heating rate, the heating rate of turbulence, the heating rate of moist processes, and the heating rate of the dynamical core (for CAM only). In Fig. 3, the radiative heating rate is the summation of the longwave and the shortwave radiative heating rates; the convective heating rate is the summation of the turbulent heating rate and moist processes heating rate; the circulation heating rate is the heating rate of the dynamical core. When SST warms, the convective heating is significantly enhanced in the upper troposphere in the deep tropics (Fig. 3b). The convective warming is mostly contributed by the enhanced latent heat release. In the subtropics, heating rate change shows a similar pattern to that in the deep tropics. The convective heating is also enhanced, but with a smaller magnitude than that in the deep tropics (Fig. 3e).

The temperature adjustment driven by the circulation change (Exp. SCM_S4C4 – SCM_S4C1) is shown in Fig. 1d. The temperature adjustment shows a complementary pattern to that driven by the convection (Fig. 1c). There are two warming centers in the southern and northern subtropical upper troposphere. The warming extends from 400 hPa up to the stratosphere.

Circulation change increases the lapse-rate magnitude in the lower troposphere and decreases the lapse rate in the upper troposphere (Fig. 2c) in the western Pacific. The magnitude of the lapse-rate change is smaller compared to those driven by the radiation (Fig. 2a) and the convection (Fig. 2b). The lapse-rate changes have an opposite effect in triggering the convection; the lower troposphere favors more convection but the upper troposphere suppresses the convections.

The warmings in the stratosphere and the troposphere arise from two different mechanisms. In the troposphere, the warming is subject to the weak temperature gradient (WTG) constraint (Charney 1963). Owing to the weak Coriolis force, gravity waves (Nicholls et al. 1991) can rapidly transport the energy from the deep tropics to the subtropics, resulting in a horizontally uniform warming pattern in the whole tropics. In the stratosphere, the warming resembles the bullhorn-shaped pattern noted by Huang et al. (2016), which is caused by the acceleration of Brewer–Dobson circulation. The speeding up of the Brewer–Dobson circulation is a robust feature in global warming simulations (Butchart 2014). The acceleration is due to the meridionally steepened temperature gradient in the subtropical upper troposphere and lower stratosphere (Garcia and Randel 2008). The steepened temperature gradient is evident in Fig. 1a, where the tropical upper troposphere warms and the extratropical lowermost stratosphere cools. This steepened temperature gradient can also be observed in the convective adjustment (Fig. 1c), which indicates that the acceleration of large-scale Brewer–Dobson circulation can be predicted from the radiative–convective adjustment.

The heating rate balance shows the convective heating in the upper troposphere is further enhanced by the circulation adjustment in the whole tropics (Figs. 3c,f). In the deep tropics, the circulation enhancement of convective heating (Fig. 3c) exceeds the radiation enhancement (Fig. 3b). In the subtropics, although the enhancement of convective heating is narrower and confined below 200 hPa, the magnitude of circulation enhancement (Fig. 3f) is comparable to the radiation enhancement (Fig. 3e). The enhanced convective heating suggests the

FIG. 3. Heating-rate decompositions. In the deep tropics (15°S – 15°N), (a) climatological heating rates in Exp. CAM_S1, (b) heating-rate changes driven by radiative–convective adjustment, obtained from Exp. SCM_S4C1 and CAM_S1, and (c) heating-rate changes driven by circulation adjustment, obtained from Exp. SCM_S4C4 and SCM_S4C1. (d)–(f) As in (a)–(c), but for the subtropics (15° – 30°N/S).

circulation change facilitates convection to release more latent heat and amplify the TUTW.

To give a summary, we quantify the TUTW driven by different processes. We calculate the mean temperature change in the whole tropics ranging from 350 to 100 hPa and from 30°S to 30°N (as marked by the black box in Fig. 1). To further analyze the warming due to different mechanisms and in different regions, we divide the region into the deep tropics (15°S – 15°N) and subtropics (15° – 30°N/S), as well as the upper layer (100–200 hPa) and lower layer (200–350 hPa). The fractional contributions of each process to the overall warming are listed in Table 2. The average TUTW is about 6.2 K, nearly twice the 3.5 K warming at the surface. Among the overall warming in the whole tropics, 13% is contributed by the direct radiative effect. The radiative–convective adjustment produces 87% (13% + 74%) of the total warming (see boldface in Table 2). The rest of 13% is contributed by the large-scale circulation change. Although convection plays a dominant role in the formation of TUTW, radiation triggers the convective response and the circulation changes reinforce it. When divided into different regions, the radiative–convective adjustment can reproduce all of the upper-tropospheric warming in the deep tropics. The contribution from the circulation is negligible regarding the mean value, which we note results from compensating effects between different regions (Fig. 1d). In the

subtropics, the contribution from circulation is significant, especially in the upper layer (100–200 hPa). The circulation effect accounts for 57% of the total warming there (see boldface in Table 2).

b. Humidity

Along with the tropical warming, the water vapor concentration increases throughout the troposphere, with a maximum in the tropical upper troposphere (Fig. 4a). The radiative–convective adjustment predicts a more intense moistening in the tropics (Fig. 4b), which is offset by the circulation effect (Fig. 4c). The circulation effect manifests an enhanced poleward water vapor transport. The middle row of Fig. 4 shows the relative humidity change. One robust feature among GCM projections is the approximately conserved relative humidity during global warming (Fig. 4d; Held and Soden 2006; Pierrehumbert et al. 2007). However, the relative humidity driven by the radiative–convective adjustment is found to significantly increase in the tropical upper troposphere. The increase is generally more than 20%, with a maximum exceeding 35% in the subtropics at around 150 hPa (Fig. 4e). The relative humidity increase is suppressed after the circulation adjustment is included (Fig. 4f). The cloud change shows similar patterns as the relative humidity change. The overall cloud change discloses an upward shift in high cloud (Fig. 4g; Zelinka

TABLE 2. The percentage of contributions from different processes [radiative (rad), convective (conv), and circulation (circ)] to the overall warming. The tropics range from 30°S to 30°N; the deep tropics range from 15°S to 15°N; and the subtropics range from 30° to 15°S and from 15° to 30°N. The boldface values represent those discussed in section 3a.

	Tropics			Deep tropics			Subtropics		
	Rad	Conv	Circ	Rad	Conv	Circ	Rad	Conv	Circ
100–350 hPa	13%	74%	13%	13%	87%	0%	13%	60%	27%
100–200 hPa	15%	56%	29%	15%	83%	2%	16%	27%	57%
200–350 hPa	12%	83%	5%	12%	89%	–1%	11%	77%	12%

and Hartmann 2010; Bony et al. 2016). Consistent with the relative humidity, the radiative–convective adjustment and circulation adjustment induce respectively significant but compensating effects in the high cloud with maxima occurring at around 150 hPa in the subtropics (Figs. 4h,i).

The significant increase in relative humidity and high cloud is not observed in conventional radiative–convective equilibrium simulations. For example, cloud-resolving models simulate a nearly constant relative humidity with various SST perturbations (Romps 2014). GCMs in radiative–convective equilibrium configurations obtain an upward shift of the high cloud, which is similar to Fig. 4g (Bony et al. 2016). The increase of relative humidity in our simulations is due to the fixed atmospheric temperature advection and water vapor advection in the single-column model runs. As a result, the subtropical upper troposphere is not warmed as much (Fig. 1c vs Fig. 1a) and the water vapor is stacked in the tropics (Fig. 4b). Both of these factors contribute to the relative humidity increase. We use Eq. (5) to separate their contributions:

$$\begin{aligned} \text{RH}(Q_{\text{scm}}, T_{\text{scm}}) - \text{RH}(Q_4, T_4) = & [\text{RH}(Q_{\text{scm}}, T_4) - \text{RH}(Q_4, T_4)] \\ & + [\text{RH}(Q_4, T_{\text{scm}}) \\ & - \text{RH}(Q_4, T_4)] + \text{Residual}. \end{aligned} \quad (5)$$

Here RH is the relative humidity, Q is the specific humidity, and T is the atmospheric temperature. The subscript “scm” indicates the variables are from the Exp. SCM_S4C1, and the subscript 4 indicates the variables are from the Exp. CAM_S4. We use monthly mean temperature and specific humidity to calculate the relative humidity in Eq. (5) and the results are shown in Fig. 5. The first term (Fig. 5b) resembles the pattern of circulation-induced specific humidity change (Fig. 4c) and the second term resembles the pattern of circulation-induced temperature change (Fig. 1d). Thus, in deep tropics, the relative humidity increase primarily arises from the accumulation of water vapor. In the subtropics, the water vapor accumulation dominates the lower part of the upper troposphere and the temperature change dominates the upper part of the upper troposphere.

c. Moist adiabat

The moist-adiabatic process is recognized to be the key process that governs the vertical temperature structure in the tropics (Holloway and Neelin 2007). However, the moist adiabat quantitatively overpredicts the TUTW in GCMs. The

overprediction is argued to be caused by the entrainment and the large-scale circulation (Miyawaki et al. 2020). As the large-scale circulation is prescribed in the single-column model, it is of interest to quantify how the radiative–convective adjustment and the circulation adjustment contribute to the overprediction separately. Our following discussion is based on the western Pacific where deep convection occurs most frequently and influences the whole tropics (Flannaghan et al. 2014; Andrews and Webb 2018). The model-simulated temperature profiles are averaged over 10°S–10°N, 120°–160°E. We calculate the moist adiabat on each grid based on the climatological mean 2-m temperature and relative humidity from model simulations using the method described in section 2c and average the moist adiabats within the same region.

Figure 6a shows the GCM-simulated climatological atmospheric profile in unperturbed climate (Exp. CAM_S1) and the corresponding moist adiabat. The GCM-simulated atmospheric temperature is lower than the moist adiabat because of the entrainment of dry air (Romps 2016; Po-Chedley et al. 2019). The temperature differences between simulated profiles and their corresponding moist adiabats are shown in Fig. 6b. In all experiments, the moist adiabats overestimate the model-simulated temperatures. At 200 hPa, in control climate (Exp. CAM_S1), the moist adiabat overpredicts the GCM profile by 4.3 K (blue line). In the single-column model, the overprediction increases to 7.5 K (orange line, Exp. SCM_S4C1). When the circulation effect is included, the overprediction amounts to 6.2 K (red line, Exp. CAM_S4). The result indicates the temperature difference between the GCM simulation and the moist adiabat becomes larger as the SST warms. This is reminiscent of the increase of CAPE as SST warms (Romps 2016).

Figure 6c shows the TUTWs predicted by the GCM, the single-column model, and the moist adiabat. The moist adiabat predicts a 9.8 K warming at 200 hPa (red line), which is noticeably greater than the GCM-simulated 7.9 K (green line) and the single-column model simulated 7.4 K. In GCM, the overprediction of TUTW by the moist adiabat is 1.9 K; in the single-column model, the overprediction increases to 2.4 K. This result suggests the overestimation of the moist adiabat in the western Pacific arises from the radiative–convective adjustment, even when there is no large-scale circulation adjustment. One plausible reason for explaining the overprediction in single-column simulation might be the entrainment. Miyawaki et al. (2020) showed the overprediction in the upper troposphere could be decreased by 10.4% when the entrainment rate is decreased.

FIG. 4. Water vapor fractional changes $[(Q_2 - Q_1)/Q_1]$; Q is specific humidity, relative humidity changes, and cloud fraction changes in response to the warming SST. (a) Overall water vapor change, obtained from Exp. CAM_S4 and CAM_S1. (b) Water vapor change driven by the radiative-convective adjustment, obtained from Exp. SCM_S4C1 and CAM_S1. (c) Water vapor change driven by the circulation adjustment, obtained from Exp. SCM_S4C4 and SCM_S4C1. (d)–(f) As in (a)–(c), but for relative humidity. (g)–(i) As in (a)–(c), but for cloud fraction. Units: %.

FIG. 5. (a) Relative humidity difference between the single-column model simulation (Exp. SCM_S4C1) and the full model simulation (Exp. CAM_S4). The relative humidity is calculated based on the monthly mean specific humidity and monthly mean atmospheric temperature. (b) As in (a), but computed by replacing the atmospheric temperature in single-column model simulation with the temperature in the full model simulation. (c) As in (a), but computed by replacing the specific humidity in single-column model simulation with the specific humidity in the full model simulation. (d) Residual computed by subtracting (b) and (c) from (a). Units: %.

4. Conclusions and discussion

We use the radiative transfer model PORT, single-column model SCAM, and standard GCM CAM to investigate the mechanisms of the tropical upper-tropospheric warming and elucidate how different processes drive the tropical upper-tropospheric warming.

We prescribe SST perturbations as forcing to reproduce the tropical upper-tropospheric warming. The radiative process warms the whole troposphere (Fig. 1b). The warming effect is largely confined to the boundary layer. Although the direct contribution of the radiative process to the tropical upper-tropospheric warming is uniform and less than 1 K, the radiative adjustment significantly increases the lapse rate (absolute

value) in the lower troposphere (Fig. 2a), which triggers more convection (Figs. 3b,e). The convective warming extends from the surface to the tropopause with a larger magnitude in the deep tropics (Fig. 1c). The convective warming is narrow in the upper troposphere and does not extend to the subtropics. The convection decreases the lapse rate in convecting regions (Fig. 2b), manifesting an opposite effect to the radiation. The subtropical upper troposphere is warmed by the circulation adjustment (Fig. 1d). The temperature change driven by circulation adjustment manifests two evident warming centers in the subtropical upper troposphere, complementary to the convective adjustment, due to the weak temperature gradient constraint in the tropics. The circulation change also feeds back to the convection and enhances the latent heat release in the

FIG. 6. Temperature profiles averaged over the western Pacific (10°S – 10°N , 120° – 160°E). (a) GCM-simulated atmospheric temperature and the moist adiabat. The moist adiabat is computed based on the 2-m temperature and humidity. The data for both the GCM-simulated profile and moist adiabat are from Exp. CAM_S1. Blue line: GCM profile; red line: moist adiabat. (b) Temperature differences between moist adiabats and model-simulated temperatures. The moist adiabats are computed from their corresponding model experiments. Blue line is from Exp. CAM_S1; orange line is from Exp. SCM_S4C1; red line is from Exp. CAM_S4. (c) Tropical upper-tropospheric warmings predicted by the full GCM, the single-column model, and the moist adiabat. Red line: moist-adiabat-predicted warming, obtained from Exp. CAM_S4 and CAM_S1; orange line: SCAM-simulated warming, obtained from Exp. SCM_S4C1 and CAM_S1; blue line: CAM-simulated warming, obtained from Exp. CAM_S4 and CAM_S1.

upper troposphere in both the deep tropics and subtropics (Figs. 3c,f). The circulation slightly increases the lapse rate in the lower troposphere and decreases the lapse rate in the upper troposphere in the convecting regions (Fig. 2c).

We quantify the tropical upper-tropospheric warming by averaging temperature from 100 to 350 hPa and from 30°S to 30°N . The average warming is about 6 K. The largest contribution to the overall warming is from the convection (74%), followed by the radiation (13%) and the circulation (13%) (Table 2). In the deep tropics (15°S – 15°N), the radiative–convective adjustment could reproduce almost all of the warming (100%). The contribution from the circulation is negligible (around 0%) regarding the regional mean values. In subtropics, the contribution from the circulation becomes important, especially for the upper layer (100–200 hPa), where the circulation contributes 57% to the total warming.

The moist adiabat overpredicts the GCM-simulated and the single-column model-simulated TUTW. The moist adiabat predicts a 9.8 K warming at 200 hPa in the western Pacific; CAM predicts a 7.9 K warming and SCAM predicts a 7.4 K warming at the same location. It is interesting to note that the single-column simulation does not agree better with the moist-adiabatic prediction. This result suggests that the radiative–convective adjustment leads to subadiabatic warming in the western Pacific, even when there is no dilution by the circulation.

This study analyzes the atmospheric warming in a prescribed-SST framework, which is justified in that we focus on the atmospheric processes. One caveat of this method is that the prescription of SST suppresses the feedback from the ocean–atmosphere interactions. Although the prescribed-SST experiment well reproduces the tropical upper-tropospheric warming simulated by the atmosphere–ocean coupled GCMs, the underlying physics may not be the same. It would be worthwhile to include an active ocean model to verify the

conclusions here in future works. Another caveat is that the convection is not explicitly resolved in climate models. The parameterizations of convection differ significantly from model to model. The entrainment coefficient is one of the most sensitive variables causing uncertainty in climate models (Knight et al. 2007). Our quantification is only based on one climate model. The sensitivity of our result to different parameterization schemes and different entrainment coefficients needs to be further explored.

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Data availability statement. The CESM codes can be downloaded from National Center for Atmospheric Research (NCAR) website (<http://www.cesm.ucar.edu/models/cesm1.2/>).

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