



## RESEARCH LETTER

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## Key Points:

- Moist adiabatic adjustment overpredicts the tropical upper tropospheric temperature response to increased CO<sub>2</sub> across CMIP5 climate models
- The overprediction is nonzero after accounting for the presence of a climatological large-scale circulation and the direct effect of CO<sub>2</sub>
- GFDL AM2.1 aquaplanet simulations show that overprediction decreases with decreasing climatological entrainment rate

## Supporting Information:

- Supporting Information S1

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## Quantifying Key Mechanisms That Contribute to the Deviation of the Tropical Warming Profile From a Moist Adiabat

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**Abstract** Climate models project tropical warming is amplified aloft relative to the surface in response to increased CO<sub>2</sub>. Here we show moist adiabatic adjustment overpredicts the multimodel mean 300 hPa temperature response by 16.6–25.3% across the CMIP5 model hierarchy. We show three mechanisms influence overprediction: climatological large-scale circulation, direct effect of increased CO<sub>2</sub>, and convective entrainment. Accounting for the presence of a climatological large-scale circulation and the direct effect of CO<sub>2</sub> reduces the CMIP5 multimodel mean overprediction by 0.7–7.2% and 2.8–3.9%, respectively, but does not eliminate it. To quantify the influence of entrainment, we vary the Tokioka parameter in aquaplanet simulations. When entrainment is decreased by decreasing the Tokioka parameter from 0.1 to 0, overprediction decreases by 9.6% and 10.4% with and without a large-scale circulation, respectively. The sensitivity of overprediction to climatological entrainment rate in the aquaplanet mostly follows the predictions of zero-buoyancy bulk-plume and spectral-plume models.

**Plain Language Summary** Climate models project tropical warming is amplified in the upper troposphere in response to increased CO<sub>2</sub>. This is expected based on a simple model that considers the latent heat released from condensation of water vapor in a rising plume (moist adiabat). The vertical profile of warming has important implications for the strength of convective storms, the subtropical climate, and the climate sensitivity. Here we compare the moist adiabatic prediction of the response to increased CO<sub>2</sub> across a hierarchy of climate models. We find that the moist adiabatic temperature profile overpredicts the tropical warming aloft in response to increased CO<sub>2</sub>. We quantify the influence of three mechanisms that are missing in the simple model of a rising plume: (1) the presence of a climatological large-scale circulation, (2) the radiative effect of increased CO<sub>2</sub>, and (3) the mixing of dry environmental air into the rising plume. Accounting for the first two mechanisms reduces the overprediction but does not eliminate it. In idealized aquaplanet simulations, we find that weaker mixing leads to smaller overprediction, in agreement with the expectation that the mixing of dry air dilutes rising plumes and decreases the latent heat released from condensation.

### 1. Introduction

One of the earliest general circulation model (GCM) predictions of the response to increased atmospheric CO<sub>2</sub> is amplified warming aloft in the tropics (Manabe & Wetherald, 1975; Manabe & Stouffer, 1980). This prediction has since been confirmed qualitatively by observations (Flannaghan et al., 2014; Santer et al., 1996; Thorne et al., 2011), state-of-the-art models such as coupled Atmosphere-Ocean GCMs (AOGCMs, Santer et al., 2005; Vallis et al., 2015), and high-resolution cloud resolving models (CRMs, Lau et al., 1993; Romps, 2011). The tropical temperature response has important implications for the global climate, as it sets the (1) static stability in the tropics, which influences the strength of deep convection (Singh & O’Gorman, 2013; Seeley & Romps, 2015) and cloud feedbacks (C. Zhou, M. D. Zelinka, et al., 2016); (2) meridional temperature gradient, which influences the position of the subtropical jet and storm tracks (Shaw et al., 2016); and (3) lapse rate feedback in the tropics, which exerts a strong influence on the global climate sensitivity owing to the large contribution of the tropics to the global mean (Popke et al., 2013; Po-Chedley et al., 2018).

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Amplified tropical upper tropospheric warming in response to increased CO<sub>2</sub> is predicted following moist adiabatic adjustment and the Clausius-Clapeyron relation (Held, 1993). In particular, moist adiabatic adjustment predicts a 9 K warming aloft in response to a 4 K surface warming assuming a tropical surface temperature of 298 K and a fixed surface relative humidity of 80%. Thus, the moist adiabat predicts the vertical structure of warming in response to surface warming (indirect effect of increased CO<sub>2</sub>) and humidity change in regions of deep convection. It does not consider several important additional effects, such as the response outside regions of climatological deep convection, the radiative response that arises in the absence of significant surface warming (direct effect of increased CO<sub>2</sub>), and the influence of convective processes such as entrainment.

The moist adiabat is a model of a rising plume and does not account for the presence of a climatological large-scale circulation with regions of ascent and descent. Andrews and Webb (2018) show that localized SST warming in the tropical western Pacific (region of climatological ascent) in HadGEM2-A results in amplified warming aloft whereas warming in the eastern Pacific (region of climatological descent) does not. Previous authors show that the tropospheric temperature response is more strongly linked to the surface temperature response when it is filtered or weighted by climatological precipitation, which is a proxy for deep convection (Flannaghan et al., 2014; Fueglistaler et al., 2015; Sobel et al., 2002).

While the direct effect of CO<sub>2</sub> does not affect sea surface temperatures, previous work shows that it leads to a nearly uniform warming in the free troposphere (He & Soden, 2015; Wang & Huang, 2020). In the absence of surface warming and humidity change, the moist adiabat would predict no warming aloft and thus underpredict the temperature response to the direct effect of CO<sub>2</sub>. However, changes in near-surface air temperature and relative humidity due to the direct effect of CO<sub>2</sub> can also impact overprediction. For example, an increase in surface relative humidity would increase the warming aloft.

We expect convective entrainment will lead to a weaker temperature response aloft than predicted by the moist adiabat because an entraining plume releases less latent heat aloft. Thus, in the presence of climatological entrainment, the moist adiabat should overpredict the temperature response aloft. Previous work shows that zero-buoyancy bulk-plume and spectral plume models, which include the effect of climatological convective entrainment, successfully capture the increase of convective available potential energy (CAPE) to warming obtained in CRMs (Singh & O’Gorman, 2013; Seeley & Romps, 2015; Zhou & Xie, 2019). Increasing CAPE with warming is consistent with the overprediction of upper tropospheric warming by the moist adiabatic response since CAPE quantifies the deviation of the temperature profile from a moist adiabat. Furthermore, Po-Chedley et al. (2019) show the moist adiabatic response to multimodel mean surface warming in the RCP8.5 scenario of Coupled Model Intercomparison Project Phase 5 (CMIP5) overpredicts the temperature response aloft in individual models. They show that including the effect of entrainment via the bulk plume model of Romps (2016) better captures the RCP8.5 temperature response. However, the influence of climatological entrainment on the temperature response to increased CO<sub>2</sub> in GCMs has not been examined in detail.

Here we quantify the moist adiabatic temperature response to increased CO<sub>2</sub> and show it overpredicts the modeled temperature response across the CMIP5 model hierarchy. We quantify the importance of three mechanisms for the overprediction of the moist adiabat: (1) the presence of a climatological large-scale circulation, (2) the direct effect of CO<sub>2</sub>, and (3) convective entrainment. We quantify the importance of (1) and (2) using the CMIP5 model hierarchy and (3) by varying the parameterized entrainment rate in idealized aquaplanet simulations.

## 2. Methods

### 2.1. CMIP5 Models

We examine the effect of the large-scale circulation and the direct effect of CO<sub>2</sub> on the tropical temperature response to increased CO<sub>2</sub> across the CMIP5 model hierarchy (Taylor et al., 2012). The CMIP5 models used in this study are listed in Table S1 in the supporting information. To set the baseline from which the contributions of the large-scale circulation and the direct effect of CO<sub>2</sub> are quantified, we consider the total response (labeled as “T”) to increased CO<sub>2</sub> as follows.

At the complex end of the CMIP5 model hierarchy, we define the total response in AOGCMs as the response to quadrupling CO<sub>2</sub>, which is quantified by

$$\text{abrupt4xCO}_2 - \text{piControl}. \quad (1)$$

We average the last 30 years of the 150-year simulation to study the near-equilibrium response in AOGCMs. In the mid-range of complexity of the CMIP5 model hierarchy, we define the total response in atmospheric GCMs (AGCMs), which prescribe the SST based on observations (Gates, 1992), as the sum of the indirect (increased SST with fixed CO<sub>2</sub>) and direct (fixed SSTs with quadrupled CO<sub>2</sub>) effects of increased CO<sub>2</sub>. The indirect effect is quantified using two different CMIP5 simulations: (1) the response to patterned SST warming (amipFuture–amip) and (2) the response to uniform 4 K warming (amip4K–amip). Thus, the total AGCM response to patterned warming, hereafter AGCMp, is quantified by

$$(\text{amipFuture} - \text{amip}) + (\text{amip4xCO2} - \text{amip}), \quad (2)$$

and the total AGCM response to uniform warming, hereafter AGCMu, is quantified by

$$(\text{amip4K} - \text{amip}) + (\text{amip4xCO2} - \text{amip}). \quad (3)$$

We average over the entire 30 years of each AGCM simulation. Finally, at the simple end of the CMIP5 model hierarchy, we quantify the total response in aquaplanet (AQUA) simulations as the sum of the indirect effect of uniform 4 K warming and the direct effect of quadrupled CO<sub>2</sub>, that is,

$$(\text{aqua4K} - \text{aquaControl}) + (\text{aqua4xCO2} - \text{aquaControl}). \quad (4)$$

We average over the last 5 years of each AQUA simulation.

We remove the impact of the climatological large-scale circulation outside regions of deep convection by averaging the response only over regions where ascent at 500 hPa exceeds –35 hPa/day, that is,  $\omega < -35$  hPa/day. Following Sherwood et al. (2014), this corresponds to the 75th percentile value in the multimodel mean climatology of the AOGCM (piControl) and AGCM (amip) simulations. The response after removing regions outside of large-scale climatological ascent is labeled as “T-L”.

Finally, we remove the impact of the direct effect of CO<sub>2</sub> by subtracting the direct effect from the total response over regions of deep convection. Given the constraints of the CMIP5 archive, we can only remove the direct effect from the AGCM and AQUA simulations (see Equations 2–4). The response after removing regions outside of large-scale climatological ascent and the direct CO<sub>2</sub> effect is labeled as “T-L-D”.

## 2.2. GFDL AM2.1 Aquaplanet GCM

In order to understand the importance of convective entrainment for the tropical tropospheric temperature response to surface warming, we configure the GFDL AM2.1 aquaplanet GCM (hereafter GFDL) with the Relaxed Arakawa-Schubert (RAS) convection scheme (Moorthi & Suarez, 1992). In the RAS scheme, the Tokioka parameter ( $\alpha$ ) controls the minimum entrainment rate ( $\epsilon_{\min}$ ) as follows:

$$\epsilon_{\min} = \frac{\alpha}{D}, \quad (5)$$

where  $D$  is the depth of the planetary boundary layer. This constraint only affects plumes that detrain above 500 hPa. Thus, the Tokioka parameter controls the entrainment rate of deep convection only. In previous studies,  $\alpha$  was varied to study the influence of convective entrainment on the Madden-Julian Oscillation (Tokioka et al., 1988), the Intertropical Convergence Zone (Kang et al., 2008), the El Niño-Southern Oscillation (Ham et al., 2013; Jang et al., 2013; Kim et al., 2011), and tropical clouds and precipitation (Lin et al., 2013). The default climatological value is  $\alpha = 0.025$  in GFDL. To investigate the role of entrainment on the tropical temperature response, we perturb  $\alpha$  from its default climatological value as follows:  $\alpha = 0, 0.00625, 0.0125, 0.05, \text{ and } 0.1$ . Increasing the Tokioka parameter beyond 0.1 does not further increase the entrainment rate. Thus, the range of bulk entrainment rates obtained here represent nearly the full extent of the entrainment rate regime that can be studied by perturbing the Tokioka parameter in GFDL.

We vary the Tokioka parameter in two configurations of the GFDL model: (1) the standard aquaplanet, hereafter GFDLaqua, configured with the same SST profile used in the CMIP5 AQUA simulations (specifically the Qobs SST profile as defined in Neale & Hoskins, 2000), and (2) rotating radiative-convective equilibrium (RCE) configured with a spatially uniform SST of 300 K, hereafter GFDLrce. The RCE configuration allows us to test the robustness of our results in the absence of a large-scale circulation, which is a common idealized model configuration for the tropics (Wing et al., 2018). For both configurations we investigate the response to a uniform SST warming of 4 K (GFDLaqua4 K and GFDLrce4K) with fixed  $\alpha$ . Following Tan et al. (2019), the GFDL aquaplanet uses RRTMG radiation and does not include the radiative effects of ozone and clouds.

### 2.2.1. Zero-Buoyancy Bulk-Plume and Spectral-Plume Models

We compare the relationship between overprediction of the moist adiabat and climatological entrainment in GFDLrce and GFDLaqua to zero-buoyancy bulk-plume and spectral plume models. The zero-buoyancy bulk-plume model is a simple 1-D model that includes the effect of a single entraining plume in RCE. To compare the predictions of the zero-buoyancy bulk-plume models to GFDL, we diagnose the bulk plume fractional entrainment rate  $\epsilon$  using the bulk plume continuity equation (see equation 7 in Romps, 2014):

$$\epsilon = \frac{1}{M} \left( \frac{\partial M}{\partial z} + d \right), \quad (6)$$

where  $z$  is height,  $M$  is the convective mass flux ( $\text{kg m}^{-2} \text{s}^{-1}$ ), and  $d$  is the detrainment mass flux per unit height ( $\text{kg m}^{-3} \text{s}^{-1}$ ).  $M$  and  $d$  are directly output from the RAS convection scheme. We vertically average  $\epsilon$  over pressure from 850–300 hPa to quantify the mean strength of entrainment in the free troposphere. Here we compare the GFDL output to three different bulk plume models. For Singh and O’Gorman (2013), hereafter SO13, we set the relative humidity to 85% to fit the relationship between overprediction and entrainment in GFDLrce and 80% to fit GFDLaqua. We vary  $\hat{\epsilon}$  (where  $\epsilon = \hat{\epsilon}/z$ ) to study how the strength of entrainment influences overprediction. For Romps (2014), hereafter R14, we set the ratio of gross evaporation to gross condensation  $\alpha = 0.80$  to fit GFDLrce and  $\alpha = 0.75$  to fit GFDLaqua and vary  $\epsilon$  directly, which is constant with height. Finally, for Romps (2016), hereafter R16,  $\epsilon = a\gamma/\text{PE}$  where  $\gamma$  is the fractional lapse rate of saturation specific humidity,  $a$  is a constant, and PE is precipitation efficiency. We set PE = 1 for both GFDLrce and GFDLaqua and vary  $\epsilon$  by varying  $a$ . Alternatively, when  $\epsilon$  is varied directly, the R16 model behaves nearly identically to R14.

The spectral plume model of Zhou and Xie (2019), hereafter ZX19, assumes a one-to-one relationship between the entrainment rate of a plume and its level of neutral buoyancy ( $z_d$ ), that is,  $\epsilon[z_d]$ .  $\epsilon[z_d]$  depends on the tropopause height  $z_t$  and two fitting parameters  $\epsilon_0$  and  $k$ . As  $\epsilon[z_d]$  is not directly output from the RAS scheme, we infer  $\epsilon[z_d]$  in GFDL such that the following criterion is satisfied:

$$\bar{h}^*(z_d) = h_{\epsilon[z_d]}(z_d), \quad (7)$$

where  $\bar{h}^*$  is the saturation moist static energy (MSE) in GFDL and  $h_{\epsilon[z_d]}$  is the MSE of a plume with entrainment rate  $\epsilon[z_d]$  according to ZX19. We convert  $\epsilon[z_d]$  to pressure coordinates (i.e.,  $\epsilon[p_d]$ ) and average the entrainment rates of plumes that detrain between 850–300 hPa to quantify the mean strength of the spectral entrainment rate in the free troposphere. We fit the GFDLrce climatology by setting  $z_t = 14.61$  km,  $k = 1.00$ , and  $\epsilon_0 = 0.33 \text{ km}^{-1}$  in ZX19. We fit the GFDLaqua climatology by setting  $z_t = 16.02$  km,  $k = 0.60$ , and  $\epsilon_0 = 0.20 \text{ km}^{-1}$ . To mimic the effect of varying the Tokioka parameter in the ZX19 model, we vary the mean  $\epsilon[z_d]$  by varying  $k$  while holding  $z_t$  and  $\epsilon_0$  fixed, which produces the best fit to the GFDL results (see Figure S1).

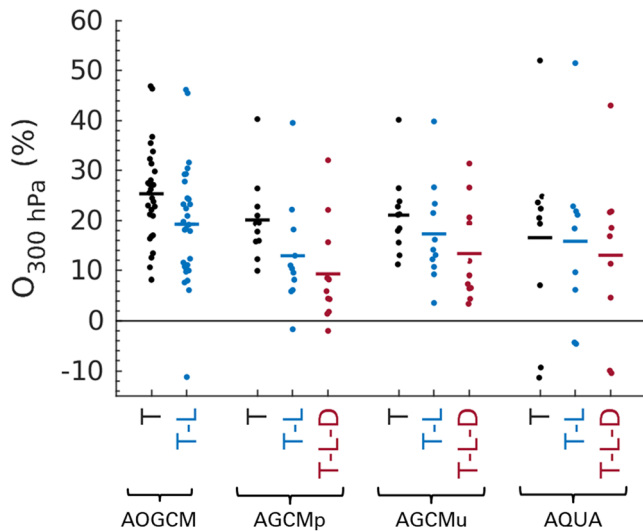
### 2.3. Calculating the Moist Adiabat and Its Overprediction

We calculate the moist adiabatic temperature at each latitude and longitude by setting the initial condition of the rising parcel as the annual mean 2 m temperature, relative humidity, and surface pressure. For models where the 2 m fields are not available, we interpolate the three-dimensional temperature and humidity fields to the surface pressure. Where the surface pressure is greater than the lowest pressure level of the vertical grid (1,000 hPa), we linearly extrapolate from the 1,000 hPa value. Similar results are obtained if the moist adiabat is computed using 950 hPa instead of 2 m fields (see Tables S2 and S3).

We integrate the dry adiabatic lapse rate  $\Gamma_d = \frac{g}{c_{pd}}$  up to the lifted condensation level (LCL). During this dry ascent, we assume that the water vapor mixing ratio is conserved. Above the LCL, we calculate temperature by integrating the moist-adiabatic lapse rate  $\Gamma_m$  following the definition in the American Meteorological Society (AMS) glossary (AMS, 2020).

$$\Gamma_m = \Gamma_d \frac{1 + \frac{L_v r_v}{RT}}{1 + \frac{L_v^2 r_v}{c_{pd} R_v T^2}}, \quad (8)$$

where  $L_v$  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,  $T$  is temperature, and  $c_{pd}$  is the isobaric specific heat



**Figure 1.** Overprediction across the CMIP5 model hierarchy. For each model configuration, black dots denote overprediction for the total response to increased CO<sub>2</sub> (labeled T) of individual models, and the black horizontal line is the mean. Overprediction averaged only over regions of climatological deep convection (where  $\omega < -35$  hPa/day at 500 hPa and labeled T-L) is shown in blue. Overprediction averaged over regions of deep convection and without the direct CO<sub>2</sub> effect is shown in red (labeled T-L-D).

symbols in Figure 1). The multimodel mean overprediction of the total response is comparable across the model hierarchy, ranging from 25.3% for AOGCM to 20.1%, 21.1%, and 16.6% for AGCMp, AGCMu, and AQUA, respectively. The overprediction is largest in the upper troposphere (Figure S2).

In what follows, we focus on quantifying the impact of the following mechanisms on overprediction: (1) the presence of a climatological large-scale circulation, (2) direct effect of CO<sub>2</sub>, and (3) convective entrainment.

### 3.2. Presence of a Climatological Large-Scale Circulation

Overprediction is smaller in regions of climatological deep convection such as the western Pacific warm pool (see region inside red contour in Figures 2a–2c). Conversely, overprediction is large over the eastern Pacific, which is characterized by climatological descent (see region outside red contour in Figure 2a–2c). Overprediction over the eastern Pacific is smaller in AGCMu compared to AGCMp, suggesting that enhanced future surface warming in the eastern Pacific contributes to overprediction. Overprediction is zonally uniform in AQUA (Figure 2d) and nearly meridionally uniform as most of 10°N/S is a region of climatological deep convection. Similar results are obtained when regions of climatological deep convection are defined using a precipitation threshold of 8 mm/day (see Figure S3).

When averaged only over regions of deep convection, multimodel mean overprediction decreases by 6.1%, 7.2%, and 3.8% for AOGCM, AGCMp, and AGCMu, respectively (compare blue to black horizontal lines in Figure 1 and see Figure S4). This decrease is statistically significant at the 5% level when comparing changes among individual models (Table S4). The multimodel mean overprediction over regions of deep convection for AQUA decreases by 0.7% and is not statistically significant because the 10°N/S region is dominated by deep convection (see Figure 2d). Clearly, the climatological large-scale circulation has an influence on the tropical temperature response to increased CO<sub>2</sub>, but accounting for this does not eliminate overprediction.

### 3.3. Direct Effect of CO<sub>2</sub>

When the response to the direct effect of CO<sub>2</sub> is removed, the multimodel mean overprediction over regions of deep convection further decreases by 3.6%, 3.9%, and 2.8% for AGCMp, AGCMu, and AQUA, respectively (compare red to blue horizontal lines in Figure 1 and see Figure S5). A *t* test shows that this decrease is statistically significant at the 5% level for all three model configurations (Table S5). As shown in previous work, the temperature response to the direct effect of CO<sub>2</sub> is vertically uniform in the free troposphere (compare vertical structure of black and orange lines in Figure S6) and hence does not follow a moist adiabat. The

capacity of dry air. This moist adiabat is a simplified form of a moist pseudoadiabat where it is assumed that all condensates precipitate out immediately and  $r_v \ll 1$ . Furthermore, we do not consider the effect of freezing (latent heat of fusion). Calculating the moist adiabat using alternative definitions such as the pseudoadiabatic and reversible adiabatic lapse rates, and including the ice phase, does not significantly change our results (see Tables S2 and S3).

We quantify the overprediction  $O_p$  of the moist adiabatic response at a pressure level  $p$  as follows:

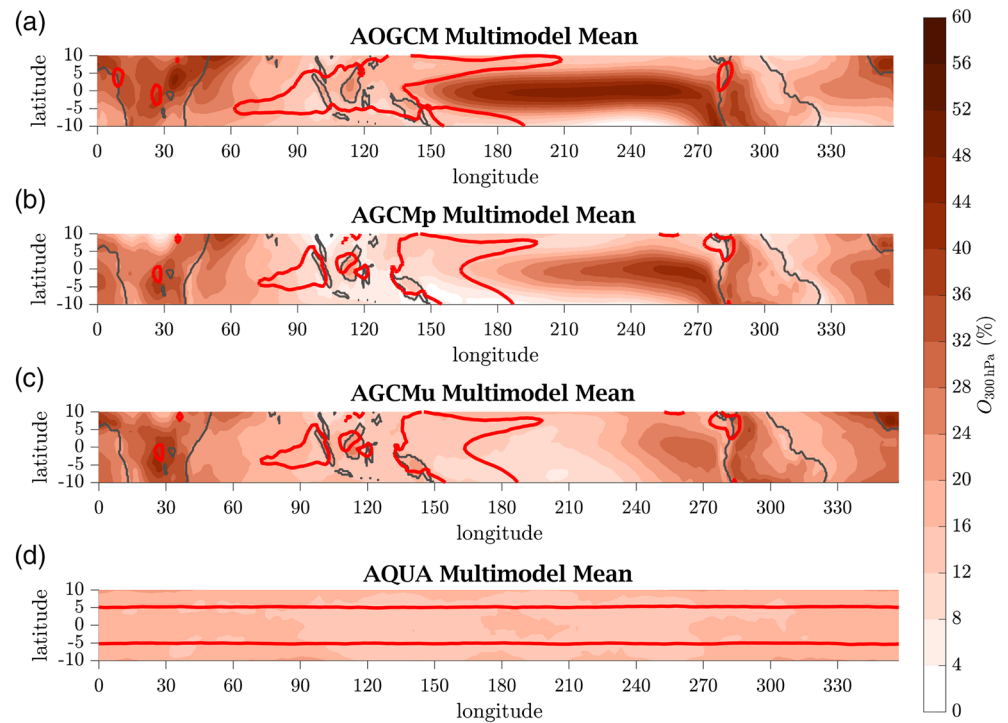
$$O_p = \frac{\Delta T_m(p) - \Delta T(p)}{\Delta T(p_s)}, \quad (9)$$

where  $\Delta$  denotes the difference between the warmer and control climates,  $T(p)$  is the GCM temperature at pressure  $p$ ,  $T_m(p)$  is the moist adiabatic temperature at pressure  $p$ , and  $T(p_s)$  is the surface temperature. We evaluate overprediction at 300 hPa following Fueglistaler et al. (2015). The tropical-mean overprediction is obtained by horizontally averaging between 10° S and 10° N.

## 3. Results

### 3.1. Overprediction Across the CMIP5 Model Hierarchy

When considering the total response to increased CO<sub>2</sub> (indirect plus direct effects), moist adiabatic warming systematically overpredicts the upper tropospheric warming across the CMIP5 model hierarchy (black



**Figure 2.** (a) Spatial structure of the overprediction of the moist adiabat at 300 hPa in response to warming for the AOGCM multimodel mean. The red contour denotes the boundary of the multimodel mean climatological deep convection (where  $\omega < -35$  hPa/day at 500 hPa). (b–d) The same for AGCMp, AGCMu, and AQUA, respectively.

overprediction associated with the direct  $\text{CO}_2$  effect is driven by a small but significant warming of 2 m air temperature and is further amplified by an increase in the 2 m relative humidity (compare dashed to solid orange lines in Figure S6).

### 3.4. Convective Entrainment

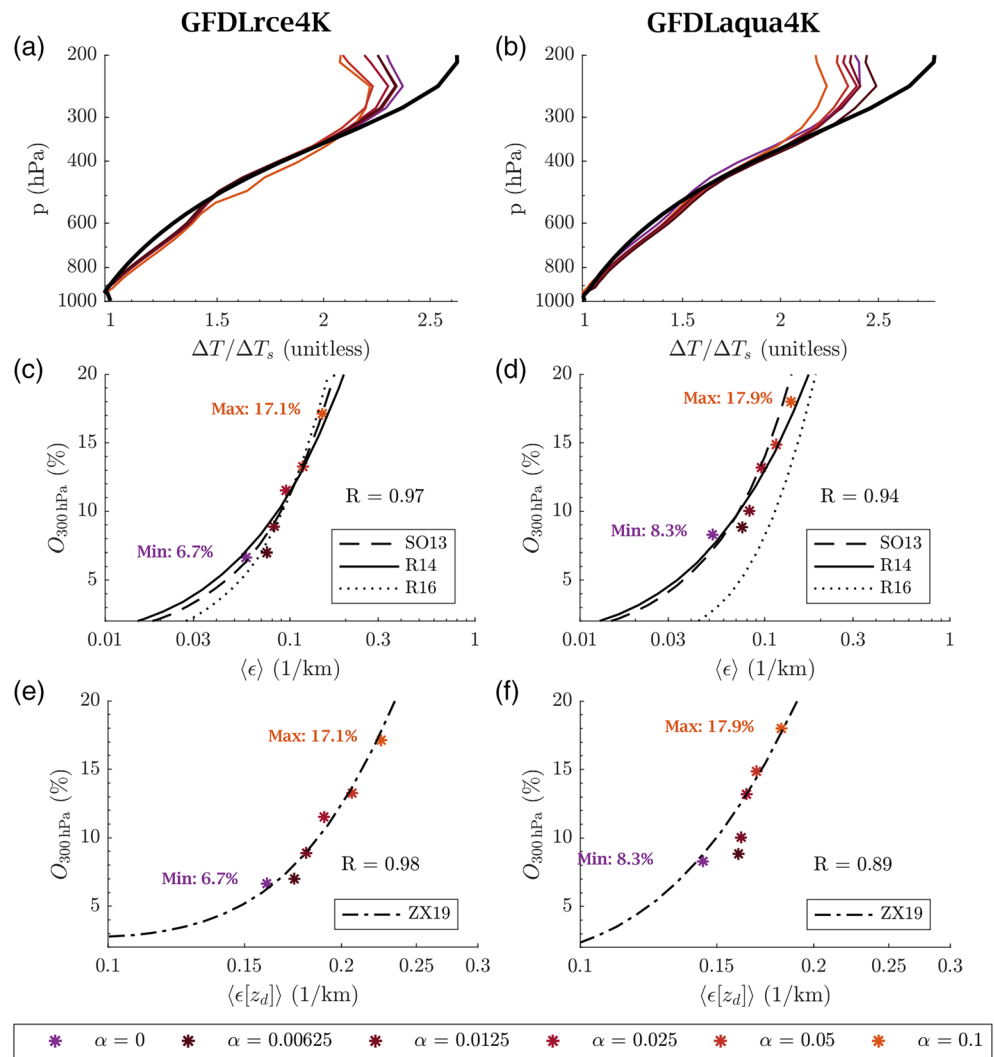
With the default Tokioka parameter ( $\alpha = 0.025$ ), the moist adiabat overpredicts the T-L-D response in GFDLrce4 K and GFDLaqua4 K by 11.6% and 13.2%, respectively (see Figure S7, Table S3). The magnitude of overprediction in GFDL is similar to that of the CMIP5 aqua4 K multimodel mean.

When the Tokioka parameter is decreased, warming is enhanced aloft and approaches the moist adiabatic response in both GFDLrce (Figure 3a) and GFDLaqua (Figure 3b) configurations. When the Tokioka parameter is decreased from 0.1 to 0, overprediction decreases from 17.1% to 6.7% in GFDLrce and 17.9% to 8.3% in GFDLaqua. Overprediction in the GFDL model is significantly correlated with the logarithm of the diagnosed bulk plume (Figures 3c and 3d) and spectral plume fractional entrainment rates (Figures 3e and 3f). The bulk plume models mostly capture this relationship for GFDLrce (see lines in Figure 3c). SO13 and R14 also capture this relationship for GFDLaqua (see lines in Figure 3d), but R16 does not. The fit of the R16 model cannot be improved by tuning the PE parameter, which is already set to its maximum value of 1. The ZX19 spectral plume model captures the relationship for both GFDLrce (see line in Figure 3e) and GFDLaqua (see line in Figure 3f).

## 4. Summary and Discussion

### 4.1. Summary

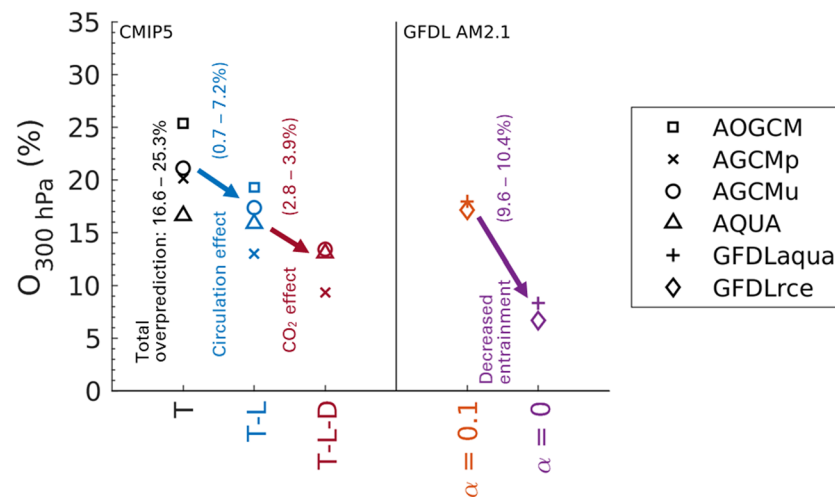
Here we investigate the accuracy of the moist adiabatic prediction of the tropical upper tropospheric temperature response to increased  $\text{CO}_2$ . We found that the moist adiabat overpredicts the multimodel mean tropical upper tropospheric warming at 300 hPa to increased  $\text{CO}_2$  by 16.6–25.3% across the CMIP5 model hierarchy (black symbols in Figure 4). We quantified the importance of three mechanisms, not included in the moist adiabat theory, to the overprediction: (1) the presence of a large-scale circulation, (2) the direct effect of  $\text{CO}_2$ , and (3) convective entrainment. The presence of a large-scale circulation and the direct effect



**Figure 3.** Temperature response in the GFDL aquaplanet when varying the Tokioka parameter  $\alpha$  for the (a) RCE (GFDLrce4K) and (b) aquaplanet (GFDLaqua4K) configurations. The moist adiabatic response is shown as a thick black line for reference. Overprediction of the moist adiabat decreases with decreasing strength of the climatological vertically averaged bulk-plume entrainment ( $\epsilon$ ) for (c) GFDLrce4K and (d) GFDLaqua4K. (e and f) Similar except the x axis shows the climatological spectral entrainment rates averaged within the free troposphere,  $\langle \epsilon[z_d] \rangle$ . The relationship between overprediction and entrainment predicted by zero-buoyancy bulk-plume models from Singh and O’Gorman (2013) (labeled SO13), Romps (2014) (labeled R14), and Romps (2016) (labeled R16), and the spectral-plume model from Zhou and Xie (2019) (labeled ZX19) are shown as black lines in panels c–f.

of  $\text{CO}_2$  were quantified using the CMIP5 archive. The importance of climatological convective entrainment was quantified by varying the Tokioka parameter in idealized aquaplanet simulations. Our conclusions are as follows:

1. The climatological large-scale circulation has a significant impact on overprediction. Overprediction is largest outside tropical regions of climatological deep convection defined by  $\omega > -35$  hPa/day. Overprediction is smaller but nonzero in tropical regions of deep convection defined by  $\omega < -35$  hPa/day. The contribution of the large-scale circulation to overprediction ranges from 0.7–7.2% (difference between black and blue symbols in Figure 4).
2. The direct effect of increased  $\text{CO}_2$ , which impacts the surface relative humidity and tropospheric temperature response through changes in radiative energy balance, tropical circulation, and precipitation, contributes to overprediction. The contribution of the direct  $\text{CO}_2$  effect to overprediction ranges from 2.8–3.9% (difference between blue and red symbols in Figure 4).



**Figure 4.** Summary of the contributions of the large-scale circulation, direct CO<sub>2</sub> effect, and convective entrainment to the overprediction of the moist adiabat response to increased CO<sub>2</sub>. The circulation effect is obtained from the difference in the CMIP5 multimodel mean overprediction averaged over the tropics ( $\pm 10^\circ$ ) and averaged only over regions of deep convection, with the range indicating results across the model hierarchy. The CO<sub>2</sub> effect is quantified by removing the direct effect of CO<sub>2</sub> over regions of deep convection while retaining the indirect effect of CO<sub>2</sub> (SST increase). The range of the entrainment effect is obtained from decreasing the Tokioka parameter ( $\alpha$ ) from 0.1 to 0 in GFDLaqua and GFDLrce.

- Parameterized convective entrainment contributes significantly to overprediction in the GFDL aquaplanet model configured with various Tokioka parameters. As the Tokioka parameter is decreased from 0.1 to 0, overprediction decreases by 10.4% for GFDLrce and 9.6% for GFDLaqua (difference between orange and purple symbols in Figure 4). Overprediction is significantly correlated with the logarithm of the climatological entrainment rate in the GFDL model. The relationship between overprediction and the climatological entrainment rate in the GFDL model mostly follows the prediction of zero-buoyancy bulk-plume models of Singh and O’Gorman (2013), Romps (2014), and Romps (2016), and the spectral plume model of Zhou and Xie (2019).

#### 4.2. Discussion

We showed that climatological convective entrainment contributes significantly to the overprediction of the moist adiabatic response to warming. Our results are in agreement with Tripathi (2014) and Po-Chedley et al. (2019), who previously argued that convective entrainment affects the overprediction of the moist adiabat in response to LGM and RCP8.5 climate changes.

In our study, perturbing the Tokioka parameter in an aquaplanet model by an order of magnitude did not capture the full intermodel spread of overprediction in the AQUA CMIP5 model response. Some possible reasons that our experiment failed to capture the full spread of overprediction include the following: (1) The RAS convection scheme is not used by all CMIP5 aquaplanet models, and other convection schemes may show greater sensitivity to entrainment, (2) the entrainment response to warming (rather than the climatological entrainment) may influence overprediction, and (3) convective processes other than entrainment may influence overprediction. The importance of (1) may be addressed by running experiments using a different convection scheme. The importance of (2) may be quantified by prescribing different entrainment rates in a warmer climate. Prescribing different Tokioka parameters in the control and warm climates of the GFDL aquaplanet leads to a large range of overprediction ( $-40.4\%$ – $73.5\%$ ; see Figure S8). However, parameterized entrainment must be compared to more direct measures of entrainment such as those diagnosed from CRM simulations (Romps, 2010). Future work could also explore (3) by quantifying the influence of precipitation efficiency and cloud radiative effects. Fully understanding the relationship between entrainment and overprediction using theory and CRMs is an important area of future work.

This work highlights that while moist adiabatic adjustment provides a useful qualitative understanding of the tropical temperature response to increased CO<sub>2</sub>, it has limitations as a quantitative theory. Incorporating

the mechanisms identified here is important for improving the accuracy of the prediction. A full understanding of tropical lapse rate changes is critical to provide confidence in tropical climate predictions of the response to increased CO<sub>2</sub>.

## Data Availability Statement

Data supporting the conclusions are available through Knowledge@UChicago (<https://doi.org/10.6082/uchicago.2204>).

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