Strengthening of convective mass fluxes despite weakening of the mean circulation in super-parameterized warming simulations

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Abstract

Previous work has established that warming is associated with an increase in dry static stability, a weakening of the tropical circulation, and a decrease in the lower-tropospheric convective mass flux. Using both idealized and realistic global warming simulations with super-parameterized convection, we find that the weakening of the tropical circulation can occur at the same time as a strengthening of the middle- and upper-tropospheric convective mass flux. Our analysis shows that this strengthening results from changes in the stratiform heating and "environmental" vertical motion that occur in the spaces between the convective clouds.

Understanding the response of tropical ascent to warming using an energy balance framework

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Key Points:

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- The ascending area narrows as the heating there increases relative to the cooling
 in the sinking air
- The circulation weakens with warming because radiative cooling increases more
 slowly than stability
- Enhanced vertical moisture gradients and stratiform heating support decreased
 ascent area

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12 Abstract

Previous work has established that warming is associated with an increase in dry 13 static stability, a weakening of the tropical circulation, and a decrease in the convective 14 mass flux. Using a set of idealized simulations with specified surface warming and super-15 parameterized convection, we find support for these previous conclusions. We use an en-16 ergy and mass balance framework to develop a simple diagnostic that links the fractional 17 area covered by the region of upward motion to the strength of the mean circulation. We 18 demonstrate that the diagnostic works well for our idealized simulations, and use it to 19 understand how changes in tropical ascent area and the strength of the mean circula-20 tion relate to changes in heating in the ascending and descending regions. We show that 21 the decrease in the strength of the mean circulation can be explained by the relatively 22 slow rate at which atmospheric radiative cooling intensifies with warming. In our sim-23 ulations, decreases in tropical ascent area are balanced by increases in non-radiative heat-24 ing in convective regions. Consistent with previous work, we find a warming-induced de-25 crease in the mean convective mass flux. However, when we condition by the sign of the 26 mean vertical motion, the warming-induced changes in the convective mass flux are non-27 monontonic and opposite between the ascending and descending regions. 28

²⁹ Plain Language Summary

The circulation of the atmosphere is expected to weaken in a future warmer climate. Despite an expected increase in precipitation, studies show that the average strength of stormy updrafts (as measured by the average speed of the updrafts multiplied by their area) will also decrease. We use simulations with realistic representations of storms to test these ideas, and derive an equation that may help us better understand how the strength of the circulation is linked to changes in its energy balance.

36 1 Introduction

The mean tropical circulation is closely coupled to convection. Mean rising motion occurs in relatively warm and moist regions, in association with active deep cumulus convection. This upward motion is balanced by slow, radiatively driven subsidence that occurs in drier regions, where cumulus convection is suppressed. For reasons discussed below, both the tropical mean circulation and the low-level convective mass flux are expected to weaken in a future, warmer climate (Betts & Ridgway, 1989; Chou & Chen, 2010; Knutson & Manabe, 1995; Held & Soden, 2006; Schneider et al., 2010; Seager et al., 2010; Vecchi & Soden, 2007). The mean circulation is the net vertical mass flux over an area large enough to contain many convective clouds, and is the sum of the mass fluxes of narrow convective updrafts and downdrafts and the vertical mass flux of their broad, quiescent environment.

Betts and Ridgway (1989) were the first to suggest that the tropical mean circulation would weaken with warming. Using a simple model of the tropical boundary layer, they found that the subsidence required for thermodynamic equilibrium weakened as sea surface temperatures warmed. Their conclusions were supported by the results of Knutson and Manabe (1995), who found that a global circulation model (GCM) simulated a weakening of the mean circulation with warming.

The response of the mean circulation can be understood in terms of the area-averaged dry static energy budget:

$$\frac{\partial \overline{s}}{\partial t} = -\overline{\mathbf{v}_h} \cdot \nabla \overline{s} - \overline{w} \frac{\partial \overline{s}}{\partial z} + \overline{Q_R} + \overline{Q_c}.$$
(1)

Here s is dry static energy, \mathbf{v}_h is the horizontal wind vector, w is the vertical velocity, Q_R is the radiative heating rate, Q_c is the non-radiative heating rate due to cloud processes and turbulence, and an overbar represents an area average. In a dry-statically stable atmosphere, $\partial s/\partial z > 0$. In the tropics, and for time scales longer than a few hours or a day at most, equation (1) can be approximated by the "weak temperature gradient" (WTG) balance (Charney, 1963; Sobel & Bretherton, 2000; Sobel et al., 2001):

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$$\overline{w}\frac{\partial \overline{s}}{\partial z} = \overline{Q_R} + \overline{Q_c}.$$
(2)

In the absence of non-radiative heating (i.e., for $\overline{Q_c} = 0$), (2) reduces to a balance between radiative cooling ($\overline{Q_R} < 0$) and the warming due to downward advection of dry static energy (i.e., $\overline{w} < 0$). Many studies have shown that in a warming climate the tropical static stability increases, roughly following the more stable moist adiabat associated with warmer surface temperatures. The fractional change in the static stability is larger than the fractional change in the radiative cooling rate, so that (2) implies slower sub-

⁷⁰ sidence in non-convective regions.

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Held and Soden (2006) proposed that the globally averaged convective mass flux
will also decrease with warming. This is the vertical mass flux associated with convective updrafts (and downdrafts). The mass flux associated with convective updrafts is given
by

$$M_u = \rho \sigma_u w_u, \tag{3}$$

where ρ is the density of the air, σ_u is the fractional area covered by the convective up-

drafts, and w_u is the vertical velocity of the convective updrafts. A similar formula gives

 M_d , the mass flux associated with convective downdrafts. Held and Soden (2006) expressed

⁷⁹ the globally averaged moisture budget in the form

$$P = (M_u)_B q_B. (4)$$

Here P is the surface precipitation rate, $(M_u)_B$ is the lower-tropospheric value of M_u , 81 and q_B is the lower-tropospheric water vapor mixing ratio. Changes in the globally av-82 eraged latent heat release associated with precipitation are mainly balanced by changes 83 in atmospheric radiative cooling (e.g., Riehl & Malkus, 1958). Allen and Ingram (2002) 84 argued that, as the climate warms, the fractional increase in P will be much smaller than 85 the fractional change of q_B , which increases following Clausius-Clapeyron scaling. Based 86 on this idea, Held and Soden (2006) concluded from (4) that M_B must decrease with warm-87 ing. Vecchi and Soden (2007) provided support for this conclusion, based on an anal-88 ysis of results from a suite of GCMs. 89

In summary, energy balance suggests that the subsidence *velocity* will weaken with warming, and moisture balance suggests that the lower-tropospheric convective mass flux will also weaken with warming. Caution is needed before concluding that these changes imply a weakening of the circulation as measured by the total upward (or downward) mass flux. For example, slower subsidence over a broader region can give the same downward mass flux. In fact, an expansion of subsiding regions and a contraction of ascending regions have been observed in the historical record and produced in simulations of warming (e.g., Byrne & Schneider, 2016a, 2016b; Byrne et al., 2018; Hu & Fu, 2007; Lau
& Kim, 2015; Lu et al., 2007; Su et al., 2017, 2019; Wodzicki & Rapp, 2016).

Additionally, although the mass flux of the large-scale mean circulation is expected 99 to be upward when and where the mass flux of deep convective clouds is upward, there 100 are important differences between the two (Arakawa & Schubert, 1974; Betts, 1998). The 101 mean mass flux includes partially cancelling contributions from much stronger local con-102 vective updrafts and downdrafts, as well as vertical motions in the broad environment 103 between the convective drafts. For example, Schneider et al. (2010) estimated that the 104 rate at which mass ascends in convective updrafts may be up to five times larger than 105 the mean upward motion, simply because of compensating downward motions in the same 106 region. A more detailed discussion is given later. 107

For the reasons outlined above, the current paradigm of warming-induced weakening of both the tropical mean circulation and the tropical convective mass flux needs further study, in part because a weakening of the tropical circulation and/or convective mass flux may contribute to a weakening of teleconnections between the topics and middle latitudes (Bui & Maloney, 2018, 2019; Wolding et al., 2017). In this paper we use idealized simulations of global radiative-convective equilibrium (RCE) to investigate warminginduced changes to the mean circulation and convective mass flux.

115 2 Methods

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2.1 Model

Simulations of the future climate are sensitive to convective parameterizations (e.g., 117 Maher et al., 2018). Models participating in the Coupled Model Intercomparison Project 118 Phase 5 vary widely in their projections of the strength of the mean tropical circulation 119 (Byrne et al., 2018), mostly due to differences in convective parameterizations between 120 the models (Schiro et al., 2019). The resulting uncertainties can be avoided by using global 121 convection-resolving models (CRMs; e.g., Stevens et al., 2019), but for now the high com-122 putational cost of such models limits their use in global climate change simulations. Su-123 perparameterization offers an intermediate option, in which the conventional parame-124 terizations of cloud and boundary-layer processes are replaced by a simplified CRM em-125 bedded within each GCM grid cell (Grabowski, 2001; Khairoutdinov & Randall, 2001; 126 Khairoutdinov et al., 2005; Randall et al., 2003). The CRM explicitly simulates the cloud-127

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scale dynamics by solving the equation of motion. Cloud microphysics, radiative trans-128 fer, and turbulent mixing are parameterized on the CRM's fine grid. Despite the two-129 dimensionality of the embedded CRMs, which is needed to reduce the high computational 130 cost of simulations, superparameterization enables more realistic simulations of a num-131 ber of convectively coupled global processes (summarized by Randall et al., 2016). Key 132 for the present study is that super-parameterization makes it possible to directly diag-133 nose changes in the convective mass flux and heating associated with cloud systems that 134 are explicitly simulated by the CRM. 135

We will present results from a superparameterized (SP) version of the Community Atmosphere Model 4 (CAM4), using the finite-volume dynamical core and a 0.9°×1.25° horizontal grid. Each GCM grid column hosts an embedded two-dimensional CRM, which has a horizontal grid spacing of 4 km with 32 columns. The CRMs share the bottom 24 of CAM4's 26 layers. Microphysics are computed by the CRM using a single-moment microphysics scheme, described in detail by Khairoutdinov and Randall (2003). For more details on SPCAM4 see Kooperman et al. (2016).

Following the experimental design of the RCE model intercomparison project (RCEMIP; 143 Wing et al., 2018), we simulated RCE using uniform solar insolation and uniform sea sur-144 face temperatures (SSTs) on a non-rotating planet. Using SSTs of 295 K, 300 K, and 145 305 K, the model was run for three simulated years. These simulations are among those 146 analyzed by Wing et al. (submitted). Our results are based on analysis of hourly and 147 daily mean output for a 15-day extension at the end of year 3. Data were saved on the 148 native CAM4 $0.9^{\circ} \times 1.25^{\circ}$ horizontal grid, and for 26 hybrid-sigma model levels, with 149 hybrid sigma vertical coordinate σ_h . 150

2.2 Diagnostics

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For both the RCE and earth simulations, we make use of a diagnostic quantity computed by the CRM and saved on the GCM grid. This is the non-radiative temperature tendency due to the embedded CRM (model variable name "SPDT"), horizontally averaged over the CRM's grid.

The model also provides convective mass flux diagnostics, which are computed using vertical velocities on the CRM's grid. The mass fluxes are computed on each CRM time step, then time-averaged for output at a specified interval. Following (3), the con-

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vective mass flux is the sum of the air density times the vertical velocity times the fractional area over CRM grid cells for which the average of the vertical velocities at the model level top and bottom is greater than (for updrafts) or less than (for downdrafts) a specified vertical velocity threshold. CRM cells with weaker vertical velocities do not contribute to the average. The specified threshold vertical velocity is used to distinguish convection from non-convective processes, such as turbulence and gravity waves.

The threshold is of course somewhat arbitrary. To explore the sensitivity to the thresh-165 old, and also to obtain some information about the range of convective intensity, we com-166 puted convective mass fluxes using hourly vertical velocities in the same way as described 167 above, but with three different vertical-velocity thresholds $(0.5 \text{ m s}^{-1}, 1 \text{ m s}^{-1}, \text{ and } 2$ 168 m s^{-1}). A test comparing the global mean convective mass fluxes computed using hourly 169 versus instantaneous vertical velocities, both using the 2 m s⁻¹ threshold, showed good 170 agreement. In the plots shown later, we use the lowest threshold, 0.5 m s^{-1} , to diagnose 171 the total convective mass flux. We categorize mass fluxes between 0.5 m s^{-1} and 1 m172 s^{-1} as "weak" convection, between 1 m s^{-1} and 2 m s^{-1} as "moderate" convection, and 173 greater than 2 m s⁻¹ as "intense" convection. 174

- 175 **3 Results**
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3.1 Mean circulation

The global-mean precipitation rate increases from 2.6 mm day⁻¹ in the 295 K RCE simulation to 3.2 mm day⁻¹ (a 4% K⁻¹ increase) and 3.8 mm day⁻¹ (3.5% K⁻¹) in the 300 K and 305 simulations, respectively. As expected for a boundary layer that is warming but maintaining a roughly constant relative humidity, the simulated increase in the water vapor mixing ratio of the lowest model level is about 6.8 % K⁻¹.

Figure 1 shows the domain-mean dry static stability profiles for these simulations. The dry static stability increases with warming, particularly at upper-levels, as expected for a nearly moist-adiabatic lapse rate adjusting to warmer surface temperatures.

As discussed earlier, a weakening of the mean subsidence velocity is an expected response to warming. However, this does not directly imply a weakening of the mean circulation as measured by the rate that mass circulates. In order to say something about potential changes to the mean circulation with warming, it is necessary to take into account any changes in the fractional areas covered by mean ascent and descent.



Figure 1. Mean static stability profiles for simulations. The vertical axis is the hybrid-sigma vertical coordinate, σ_h . We have divided by c_p , the specific heat of air at constant pressure, so the units are K km⁻¹.

We define regions of ascent and descent at each level based on their daily mean vertical velocity. Mass balance requires that the mass fluxes of the ascending and descending regions are equal. This can be written as

$$\overline{M_{up}} = -\overline{M_{dn}} \equiv \overline{M},\tag{5}$$

193 or as

$$\overline{w_{up}}\rho_{up}\alpha_{up} = -\overline{w_{dn}}\rho_{dn}\alpha_{dn}.$$
(6)

In these equations, the subscripts "up" and "dn" refer to regions of ascent ($\overline{w} > 0$) and descent ($\overline{w} < 0$), respectively. In (5), we take advantage of the fact that $\overline{M_{up}} = -\overline{M_{dn}}$ to define the symbol \overline{M} (no subscript), which can be called the mass flux of the "overturning circulation." For our simulations of global RCE without rotation, we can take advantage of the weak horizontal temperature gradients and apply equation (2) to substitute for the vertical velocities in equation (5) in terms of the area-averaged heating rates. Assuming that $\rho_{up} \approx \rho_{dn}$, and making the substitution $\alpha_{dn} = 1 - \alpha_{up}$, we find that

$$\alpha_{up} \approx \frac{1}{1 - \gamma \frac{\overline{Q_{up}}}{\overline{Q_{dn}}}},$$

(7)

²⁰³ where we define

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$$\gamma \equiv \frac{(\partial \overline{s}/\partial z)_{dn}}{(\partial \overline{s}/\partial z)_{up}},\tag{8}$$

and \overline{Q} is the mean total heating (i.e., $\overline{Q_R} + \overline{Q_c}$). Neglecting the effects of differences in 205 the dry static stability between the ascending and descending regions (i.e., for $\gamma \approx 1$), 206 equation (7) says that the fractional area covered by mean ascent decreases as the ra-207 tio of mean total heating between the ascending and descending regions increases. The 208 ratio Q_{up}/Q_{dn} is always ≤ 0 . In fact, we can use equation (7) to show that for, $\gamma =$ 209 1, the area-averaged heating rate (across both ascending and descending regions) is zero. 210 When the mean (absolute) heating rates of the two regions are equal, $\overline{Q_{up}} = \overline{Q_{dn}}$ and 211 (7) gives $\alpha_{up} = 0.5$. Larger heating rates in the ascending region, relative to the cool-212 ing rates in the descending region, imply stronger rising vertical velocities in the ascend-213 ing region than sinking vertical velocities in the descending region (equation 2). The de-214 scending region must therefore cover a larger fractional area than the ascending region. 215 This is what we see in our simulations. Figure 2a shows the difference between α_{up} cal-216 culated using equation (7) and the true α_{up} . Throughout the mid-troposphere, roughly 217 between levels $\sigma_h = 0.7$ to $\sigma_h = 0.3$, the approximate α_{up} given by equation (7) is nearly 218 equal to the true α_{up} , shown in Figure 2b. That is, between these levels, the error of the 219 approximation is small. The approximation slightly overestimates α_{up} between σ_h 220 0.9 and $\sigma_h = 0.7$, and is very poor in the boundary layer and stratosphere. At low lev-221 els, while some of the disagreement between the approximation and α_{up} may be due to 222 differences in density between the humid ascending and dry descending regions, it is more 223 likely driven by the weaker applicability of the WTG approximation at those levels. 224

Figure 2c shows the percentile of the column relative humidity (CRH; column precipitable water divided by the precipitable water of a saturated column with the same temperature profile) where these regions tend to be located. Regions of ascent in the up-



Figure 2. a, The difference between the approximate fractional ascent area, α_{up} , given by equation (7) and the true α_{up} . b, α_{up} . c, The column relative humidity (CRH) percentile that separates regions of mean ascent (right of the line) from regions of mean descent (left of the line). d, The mean vertical velocity, \overline{w} , for the ascending (solid line) and descending (dashed line) regions. e, The true (solid) and approximated, using equation (9) (dashed) mass flux of the mean circulation, \overline{M} .

per troposphere have CRH values larger than about 0.75, while ascending air can occur
in the lower troposphere even for CRH values less than 0.5.

- Figure 2d shows the mean vertical velocity in the ascending and descending regions. Mean rising motions are stronger than sinking motions throughout most of the troposphere. Again, ignoring the small differences in static stability between the ascending and descending regions, 2d together with equation (2) implies that the heating in the ascending region is stronger than the cooling in the descending region. This is reflected in the values of α_{up} , which are less than 0.5 at almost all levels.
- In addition to a weakening of the mean subsidence velocity, we find that the fractional area covered by the ascending region decreases with SST. This decrease is stronger in the upper troposphere, above $\sigma_h = 0.5$. At those same levels the fractional area covered by subsidence increases with SST. Figure 2e shows that the overturning mass flux (which is equal and opposite between the ascending and descending regions) weakens with warming throughout most of the troposphere (between about $\sigma_h = 0.9$ and $\sigma_h = 0.25$). This is consistent with previous work, and reflects the fact that both the fractional area



Figure 3. a, At $\sigma_h = 0.51$, the percent change in the strength of the mean circulation, \overline{M} , that results from prescribed fractional changes in $\overline{Q_{up}}$ and $\overline{Q_{dn}}$ of the 295 K simulation, given the change in the dry static stability between the 295 K and 300 K simulations. The dashed line shows the percent change in $\overline{Q_{dn}}$ between the 295 K and 300 K simulations. b, The simulated percent change in $\overline{Q_{dn}}$ between the 295 K and 300 K simulations. b, The simulated and 305 K simulations (brown). c, The simulated percent change in $\overline{Q_{up}}$ (solid), and the percent change in $\overline{Q_{up}}$ needed to strengthen \overline{M} (dashed).

of the ascending region and the mean vertical velocity in the ascending region decreasewith increasing SST.

Using our approximation for α_{up} with equation (2), we can approximate the strength of the overturning circulation by

$$\overline{M} \approx \frac{\rho}{(\partial \overline{s}/\partial z)_{up}} \left(\frac{\overline{Q}_{up}}{\overline{Q}_{dn}} - \overline{Q}_{up}} \right). \tag{9}$$

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The approximate values are shown as the dashed line in Figure 2e. At most levels, the 248 approximation accurately diagnoses the vertical profile of \overline{M} and its decrease with SST, 249 although there are some larger errors in the upper troposphere (between $\sigma_h = 0.6$ and 250 $\sigma_h = 0.3$). Equation (9) directly shows how the strength of the mean circulation is in-251 versely proportional to the dry static stability in the ascending region. For $\gamma \approx 1$, this 252 means that the strength of the mean circulation is inversely proportional to the mean 253 tropical dry static stability. Equation (9) shows that increases in \overline{M} may be possible given 254 larger fractional increases in the term, $\frac{\overline{Q_{up}} \overline{Q_{dn}}}{\overline{Q_{dn}} - \overline{Q_{up}}}$, than in dry static stability. 255

We can use equation (9) to explore the magnitude of changes in $\overline{Q_{up}}$ and $\overline{Q_{dn}}$ re-256 quired to strengthen \overline{M} . Figure 3a shows, for $\sigma_h = 0.51$, the percent change in \overline{M} that 257 results from prescribed fractional changes in the $\overline{Q_{up}}$ and $\overline{Q_{dn}}$ of the 295 K simulation, 258 given the change in the dry static stability between the 295 K and 300 K simulations. 259 In the descending region, a positive change indicates an intensification of the cooling. 260 We include lines showing the zero percent change in heating. At their intersection, we 261 see that between the 295 K and 300 K simulations, \overline{M} would weaken due to the increased 262 dry static stability alone. The horizontal dashed line shows the simulated percent change 263 of $\overline{Q_{dn}}$ between 295 K and 300 K. Given the simulated strengthening of cooling in the 264 descending region ($\approx +4\%$ between the 295 K and 300 K simulations) at this level, we 265 would begin to see an increase in the strength of the circulation for around a doubling 266 of the heating in the ascending region. 267

Figure 3b shows the simulated percent change in $\overline{Q_{dn}}$ between the 295 K and 300 268 K simulations (orange line) and between the 295 K and 305 K simulations (brown line) 269 at each model level. Panel c of the same figure shows the simulated change in Q_{up} (solid 270 lines) and the change in $\overline{Q_{up}}$ that would be required to get a strengthening of \overline{M} at that 271 level, given the simulated changes in $\overline{Q_{dn}}$ and dry static stability (dashed line). Through-272 out the mid-troposphere (between $\sigma_h = 0.8$ and about $\sigma_h = 0.4$), the simulated in-273 crease in $\overline{Q_{up}}$ is smaller than the change in $\overline{Q_{up}}$ needed to strength the mean circulation. 274 Between the 295 K and 300 K simulations, and for σ_h in the range 0.95 to 0.8, we see 275 the opposite: the increase in $\overline{Q_{up}}$ exceeds that needed to strengthen \overline{M} . This is consis-276 tent with the pattern of \overline{M} change that is simulated by the numerical model (see the solid 277 curves in Figure 2e). 278

In the quiescent descending region, mid-tropospheric changes in cooling with surface warming are dominated by small increases in radiative cooling. As the surface warms, the fractional rate of increase in radiative cooling is smaller than that of the dry static stability. Increases in the strength of the mean circulation would be possible only with very large and implausible increases in heating in the ascending region. This is another way of understanding why the overturning circulation weakens with warming.

We can use equation (7) to gain insight into the physical processes involved with the decrease of α_{up} with increasing SST. Figure 4a shows γ and $-\overline{Q_{up}}/\overline{Q_{dn}}$. Throughout most of the troposphere ($\sigma = 0.8$ through $\sigma = 0.4$), $\gamma \approx 1$ and is not sensitive to

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Figure 4. a, Minus the ratio of total heating between the ascending and descending regions (solid), and γ from equation (7). b, Total heating for the ascending (solid) and descending (dashed) regions. c, For the ascending region only, non-radiative (solid), longwave radiative (dashed), and shortwave radiative (dot-dashed) heating. d, For the ascending region only, heating associated with the convective mass flux $(\frac{1}{c_p} \frac{M_c}{\rho} \frac{\partial \overline{s}}{\partial z};$ solid) and stratiform heating $(\widehat{Q_c}; \text{ dashed})$.

the SST. From Figure 4a, we see that increases in $-\overline{Q_{up}}/\overline{Q_{dn}}$ are associated with the 288 decrease in α_{up} with SST. Figure 4b shows $\overline{Q_{up}}$ and $\overline{Q_{dn}}$ for each SST. In the descend-289 ing region, cooling intensifies as the SST increases, throughout most of the column. This 290 is due to increased longwave emission to space from a warmer, wetter atmosphere (e.g., 291 Pendergrass & Hartmann, 2014). With no change in $\overline{Q_{up}}$, the stronger radiative cool-292 ing rate in the descending region would lead to an *increase* in α_{up} . But what we actu-293 ally see is a decrease in α_{up} , because the mean heating rate in the ascending region is 294 also intensifying, and at a faster rate than the cooling over the descending region (i.e., 295 $-\overline{Q_{up}}/\overline{Q_{dn}}$ increases). This provides support for the suggestion of Schiro et al. (2019) 296 that changes to convection in the ascending region are closely linked to changes in as-297 cent area with warming. 298

Figure 4c shows the contributions to $\overline{Q_{up}}$ from both non-radiative processes ($\overline{Q_c}$), and radiation. The radiative portion is further subdivided into its longwave $\overline{Q_{Rl}}$ and shortwave $\overline{Q_{Rs}}$ parts. Longwave cooling intensifies in the ascending region (just as it does in the descending region), but there is very little change in shortwave heating of the ascending region. The main message of panel c is that the increases in $\overline{Q_{up}}$ with SST in the ascending region are due to increases in non-radiative heating. For a GCM grid cell, the mass flux of the mean circulation, \overline{M} , can be written as the sum of the net convective mass flux M_c and the "environmental" mass flux \widetilde{M} (Arakawa & Schubert, 1974):

$$\overline{M} = M_c + M. \tag{10}$$

The environmental mass flux is associated with weak vertical motion in the broad regions between the convective updrafts and downdrafts. It is typically but not always down-

ward. Eq. (10) shows that when convection is not active the mean mass flux is equal to

the environmental mass flux; without convection, the domain is "all environment."

For averages over areas large enough so that the fractional area occupied by convective updrafts is $\ll 1$ (such as a GCM grid cell), $\overline{Q_c}$ may be written as

$$\overline{Q_c} = \frac{M_c}{\rho} \frac{\partial \overline{s}}{\partial z} + L\widetilde{C} + \overline{D}(s_c - \overline{s}) + \overline{Q_{turb}},\tag{11}$$

where $\frac{M_c}{\rho} \frac{\partial \overline{s}}{\partial z}$ is the warming due to the net convective mass flux, L is the latent heat of condensation and/or freezing, \tilde{C} is the environmental condensation rate, \overline{D} is the detrainment mass flux, and $\overline{Q_{turb}}$ is the dry static energy transport due to turbulence (Arakawa & Schubert, 1974). We can isolate the (non-radiative) cloud and turbulent heating apart from (i.e., not including) the contribution from $\frac{M_c}{\rho} \frac{\partial \overline{s}}{\partial z}$ using Equation (11) as

$$\widehat{Q_c} = \overline{Q_c} - \frac{M_c}{\rho} \frac{\partial \overline{s}}{\partial z},\tag{12}$$

³²¹ where we define

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$$\widehat{Q_c} \equiv L\widetilde{C} + \overline{D}(s_c - \overline{s}) + \overline{Q_{turb}}.$$
(13)

We use (12) to diagnose $\widehat{Q_c}$ from the model output. Eq.(13) shows that contributions to $\widehat{Q_c}$ come from environmental (non-convective) condensation $L\widetilde{C}$, convective detrainment of dry static energy $D(s_c - \overline{s})$, and turbulent transport of dry static energy, but for simplicity we refer to $\widehat{Q_c}$ as the "stratiform heating." A large fraction of $\widehat{Q_c}$ comes from phase changes that occur outside of convective updrafts. Houze (1977) used radar data with other observations to show that such stratiform condensation is a major component of the heating in tropical convective systems. We expect $L\tilde{C} > 0$ where there is environmental rising motion in near-saturated environments (e.g., in stratiform anvil clouds), and $L\tilde{C} < 0$ where condensed water/ice is evaporating/melting as it interacts with an unsaturated portion of the environment.

Figure 4d shows the non-radiative heating, $\overline{Q_{c,up}}$, in the ascending region. The solid 332 curves show $\frac{M_c}{\rho} \frac{\partial \overline{s}}{\partial z}$ and the dashed curves show $\widehat{Q_c}$. The increase in $\widehat{Q_c}$ with SST is roughly 333 monotonic, except between the 295 K and 300 K simulations where a small decrease be-334 tween $\sigma_h = 0.7$ and $\sigma_h = 0.55$ may be due to the deepening of the troposphere. In-335 terestingly, below $\sigma = 0.4$ the change in $\frac{M_c}{\rho} \frac{\partial \overline{s}}{\partial z}$ with SST is not monotonic. For the 336 295 K and 300 K simulations, heating associated with M_c is larger than $\widehat{Q_c}$ below $\sigma_h =$ 337 0.6. This is reversed in the 305 K simulation, for which $\widehat{Q_c}$ larger than $\frac{M_c}{\rho} \frac{\partial \overline{s}}{\partial z}$ through-338 out the troposphere. Above $\sigma_h = 0.5$ for all simulations, and throughout the troposphere 339 for the 305 K simulation, most of the increase in the total $\overline{Q_c}$ of ascending regions comes 340 from an increase in $\widehat{Q_c}$ with SST. Between the 295 K and 300 K simulations, the increase 341 in $\frac{M_c}{\rho} \frac{\partial \overline{s}}{\partial z}$ roughly equals that of \widehat{Q}_c . 342

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3.2 Stratiform heating

What accounts for the large increase of \widehat{Q}_c with warming in the ascending regions? 344 In the upper troposphere, the increase is directly related to the decrease in the fractional 345 area of the ascending region with warming. Through its contribution to the increase in 346 $\overline{Q_{up}}$, it plays a role in determining the response of the mean circulation to warming. That 347 is, increases in $\widehat{Q_c}$ in the ascending region with SST keep the mean circulation from weak-348 ening as much as it would given no change in heating in the ascending region. In places 349 where \widetilde{M} is upwards we can write environmental latent heating, $L\widetilde{C}$, in terms of the ad-350 vection of environmental moisture by M as, 351

$$L\widetilde{C} \approx -L\frac{\widetilde{M}}{\rho}\frac{\partial\widetilde{q}}{\partial z}, \quad (\widetilde{M} > 0),$$
 (14)

where \tilde{q} is the water vapor mixing ratio of the environment. We can use equation (14) to estimate the fraction of $\widehat{Q_c}$ due to environmental condensation. The dashed lines in Figure 5 are the heating from $-L\frac{\widetilde{M}}{\rho}\frac{\partial \overline{q}}{\partial z}$ in ascending regions only. We compute this term for grid cells where \widetilde{M} is upwards, and then multiply the result by the fraction of the



Figure 5. For ascending regions: a, Contributions to $\widehat{Q_c}$ from environmental condensation, estimated as $-L\frac{\widetilde{M}}{\rho}\frac{\partial \overline{q}}{\partial z}$, and the remainder $\widehat{Q_c} - \left(-L\frac{\widetilde{M}}{\rho}\frac{\partial \overline{q}}{\partial z}\right)$. b, Negative vertical water vapor mixing ratio gradient. c, Mean upwards \widetilde{M} .

ascending region where $\widetilde{M} > 0$. We use the grid cell mean water vapor mixing ratio, \overline{q} , rather than the environmental \widetilde{q} , and account for the ice phase using the temperaturedependent partitioning between the liquid and ice phases used in the model. The solid lines are the remainder when environmental condensation is subtracted from \widehat{Q}_c .

In ascending regions, heating due to environmental condensation increases with SST, 360 particularly between the 300 K and 305 K simulations. Figure 5b shows that most of this 361 increase is from the stronger vertical moisture gradient, which increases following the in-362 crease in the slope of the saturation mixing ratio curve with temperature. Between the 363 300 K and 305 K simulations, panel c shows that below $\sigma_h = 0.6$ and above $\sigma_h = 0.4$ 364 there is an additional contribution from a stronger upward \widetilde{M} (computed as the aver-365 age \widetilde{M} where $\widetilde{M} > 0$ multiplied by the fraction of the ascending region where $\widetilde{M} > 0$ 366 0).367

We would like to point out that environmental air moving *upwards* is interesting and contrary to how it is typically conceptualized. The environmental mass flux, as defined presently, is sensitive to our choice of the vertical velocity threshold we use to define convection, because we calculate \widetilde{M} using equation (10). It is possible that we have underestimated the convective mass flux; if so, this would favor upward values of \widetilde{M} . However, rising air in the environment is physically interpretable: Positive values of \widetilde{M} in the middle and upper troposphere favor the formation of stratiform anvil clouds, which

are well-understood to be important for the heating associated with mesoscale convec-

376 tive systems.

377

378

Equations
$$(10)$$
 and (11) allow us to rewrite Equation (2) as

$$\frac{M}{\rho}\frac{\partial\overline{s}}{\partial z} = \overline{Q_R} + \widehat{Q_c}.$$
(15)

Writing WTG balance in terms of \widetilde{M} , as in equation (15), allows us to diagnose the rel-379 evant terms in the heating balance important for the environmental mass flux of a large 380 area, regardless of whether or not the area contains convection (Arakawa & Schubert, 381 1974; Chikira, 2014). For grid cells containing active convection, equation (15) says that 382 the dry static energy advection by the environmental mass flux, $\frac{M}{\rho} \frac{\partial \overline{s}}{\partial z} = \frac{1}{\rho} (\overline{M} - M_c) \frac{\partial \overline{s}}{\partial z}$, 383 balances the combination of stratiform heating and radiative cooling. For grid cells that 384 don't contain active convection, $M_c = 0$, $\widetilde{M} = \overline{M}$, $\widehat{Q_c}$ is typically negligible, and equa-385 tion (15) reduces to a balance between radiative cooling and advection of dry static en-386 ergy by the mean vertical motion. Equation (15) shows us that the environmental mass 387 flux of an area is upwards when the net heating $\overline{Q_R} + \widehat{Q_c}$ is positive. Part of the heat-388 ing that the upwards \widetilde{M} balances is due to \widetilde{M} being upwards. That is, heating from en-389 vironmental condensation associated with upwards M, in part balances the advective 390 cooling due to the upwards \widetilde{M} . This is a way to understand why \widetilde{M} is upward. 391

In ascending regions, where an upwards \widetilde{M} contributes to \widehat{Q}_c , we can subtract the 392 environmental condensation term, approximated using equation (14), from \widehat{Q}_c to roughly 393 estimate the magnitude of heating from detrainment and turbulent fluxes. Figure 5a shows 394 that this heating is large in the upper troposphere. Some of this it may be from a co-395 variance between \widetilde{M} and $\partial \widetilde{q} / \partial z$ on time scales faster than a day; these would be missed 396 in our calculation of $L\widetilde{C}$ using daily mean values. The apparent importance of detrain-397 ment and turbulence in the heating budget in ascending regions requires further inves-398 tigation. This is an important question because the increase of $\widehat{Q_c}$ with warming, which 399 is largely contributed to by an increase in this residual heating term in the upper tro-400

⁴⁰¹ posphere, is directly linked to the decrease of α_{up} with warming, and hence how the mean ⁴⁰² circulation responds to warming.

403

3.3 Response of the convective mass flux to warming

In our simulations of global RCE, the decrease of α_{up} with SST is directly related 404 to fact that the net non-radiative heating in ascending regions intensifies faster than the 405 radiative cooling in descending regions. In the upper troposphere, an increase in $\widehat{Q_c}$ ex-406 plains most of the increase in the total $\overline{Q_c}$ in ascending regions, between the 300 K and 407 305 K simulations. Much of the increase is due to the intensification of the vertical mois-408 ture gradient with warming. Between the 295 K and 300 K simulations, there is an ad-409 ditional contribution to the increase in $\overline{Q_c}$ from $\frac{M_c}{\rho} \frac{\partial \overline{s}}{\partial z}$. Interestingly, this is not the case between the 300 K and 305 K simulations, where $\frac{M_c}{\rho} \frac{\partial \overline{s}}{\partial z}$ in the ascending region *decreases* 410 411 throughout the lower- and mid-troposphere. We are thus motivated to more closely in-412 vestigate how the heating associated with the convective mass flux changes with increas-413 ing SST. 414

The non-monontonic changes in $\frac{M_c}{\rho} \frac{\partial \overline{s}}{\partial z}$ in ascending regions are due to non-monotonic changes in M_c with warming rather than changes in $\frac{\partial \overline{s}}{\partial z}$, which increases monotonically with SST. For the remainder of this section, we will thus focus on changes of the convective mass flux with SST.

We will extend this analysis to the entire domain, rather than just focusing on the ascending region alone, to get a more complete understanding of how M_c changes with SST. Figure 6a shows the vertical profiles of the global mean M_c . Consistent with previous work suggesting a decrease in M_c with warming (Held & Soden, 2006; Vecchi & Soden, 2007), we find that M_c weakens throughout the troposphere.

The decrease in the global mean M_c with SST is monotonic, with the largest de-424 crease occurring between the 300 K and 305 K simulations between about $\sigma_h = 0.9$ and 425 $\sigma_h = 0.4$. When we separately consider ascending and descending regions, we find that 426 the change of M_c with SST is not monotonic with SST and opposite between the regions. 427 Figure 6e shows the total M_c in the ascending and descending regions. Below $\sigma_h = 0.35$, 428 we see a very small increase (decrease) in M_c in ascending (descending) regions between 429 the 295 K and 300 K simulations. The change is opposite between the 300 K and 305 430 K simulations, with M_c decreasing (increasing) in ascending (descending) regions. While 431



Figure 6. (Top row) Global mean profiles of simulated M_c for different intensity bins. a, Total convective mass flux. b, Convective mass flux from weak convection; c, moderate convection; d, intense convection. (Bottom row) Mean convective mass flux for the ascending (solid) and descending (dashed) regions: e, total; f, weak convection; g, moderate convection; h, intense convection.

the combination of the changes produces a monotonic decrease in the global mean M_c 432 with SST, the region-specific patterns of change have unique effects on the heating bud-433 gets of the ascending and descending regions (and hence α_{up}). For example, the decrease 434 of M_c in descending regions and increase in ascending regions that we observe between 435 the 295 K and 300 K simulations, neglecting how heating from other processes changes, 436 increases the ratio $\overline{Q_{up}}/\overline{Q_{dn}}$, and hence contributes to a decrease in α_{up} . Conversely, the 437 increase of M_c in descending regions and decrease in ascending regions that we observe 438 between the 300 K and 305 K simulations is related to the opposite change in α_{up} . 439



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vection. Below $\sigma_h = 0.7$ there is very little change in M_c from intense convection between the 295 K and 300 K simulations, and a decrease between the 300 K and 305 K simulations. Conversely, above $\sigma_h = 0.7$, we see an increase in M_c from intense convection between the 300 K and 305 K simulation.

Panels f-h of Figure 6 show M_c from weak, moderate, and intense convection in the 450 ascending and descending regions. The magnitude of M_c from weak convection is com-451 parable between the regions, except between $\sigma_h = 0.6$ and $\sigma_h = 0.3$ where M_c in the 452 ascending region is negative and about zero in the descending region. As in the pattern 453 of change of the total M_c in ascending and descending regions with SST, the change of 454 M_c from intense and moderate convection is non-monotonic and opposite between the 455 regions. We see from panel g that the decrease of M_c from moderate convection below 456 $\sigma_h = 0.5$ in the global mean profile comes from a decrease in M_c in descending regions 457 between the 295 K and 300 K simulations, with little change over the ascending region; 458 and a large decrease between the 300 K and 305 K simulations over the ascending re-459 gion, while M_c in the descending region strengthens. For M_c from intense convection, 460 the global mean increase that we observe above $\sigma_h = 0.65$ comes not from an inten-461 sification of M_c over ascending regions but from an intensification in *descending* regions 462 (which is amplified in the global mean profile because of an increase in α_{dn}). 463

Changes to the mean convective mass flux can be due to changes in the frequency
 of convective updrafts, the vertical velocity of convective updrafts, and/or updraft frac tional area.

In a warmer climate, updraft speeds are expected to intensify, with the largest in-467 creases occurring for the most intense updrafts (Singh & O'Gorman, 2015). This, in part, 468 explains a similar shift with warming in both modeled and observed precipitation (e.g., 469 Chou et al., 2012; Del Genio et al., 2007; Fischer & Knutti, 2016; Lau et al., 2013; Pen-470 dergrass & Hartmann, 2014; Sun et al., 2007). Figure 7a,d show the log_{10} distribution 471 of vertical velocities in ascending and descending regions, respectively. Frequencies are 472 calculated as the average fraction of GCM grid cells with at least one CRM grid column 473 containing an hourly vertical velocity that exceeds a threshold value. These frequencies 474 thus include information about the number of GCM grid cells in the ascending/descending 475 region containing an updraft at a certain speed (areal coverage) and the temporal fre-476

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Figure 7. For the 295 K simulation, the frequency of hourly mean vertical velocities on the CRM grid exceeding threshold values for the a, ascending and d, descending regions. b, Difference in the frequency between the 300 K and 295 K simulations for the ascending region and e, descending region. c, Difference in the frequency between the 305 K and 295 K simulations for the ascending region and f, descending region.

quency of the updraft speed. Strong vertical velocities are more common in the ascend-ing than in the descending region.

Figure 7b,c show the differences in the frequency distributions of vertical veloci-479 ties between the 295 K and warmer simulations in the ascending region. Note that the 480 differences in Figure 7b,c,e,f are also in log_{10} space, and therefore should be interpreted 481 as ratios rather than arithmetic differences. Changes between the distributions can be 482 influenced by changes to the fraction of GCM grid cells within each region containing 483 updrafts at a certain speed and/or by changes to the temporal frequency of those up-484 drafts. Between the 295 K and 300 K simulations, we see an increase in the frequency 485 of all updraft speeds at all levels, except at speeds below 2 m s⁻¹ between $\sigma_h = 0.5$ and 486 $\sigma_h = 0.35$. This is consistent with the intensification of M_c that we observe in the as-487 cending region between these two simulations. Between the 295 K and 305 K simulation, 488 we now see a decrease in the frequency of all updrafts (blue colors between threshold ver-489 tical velocities of 0 and about 2 m s^{-1}), with an intensification of the frequency of the 490 strongest updrafts (> about 2 m s⁻¹ above $\sigma_h = 0.7$ and > about 4 m s⁻¹ between $\sigma_h =$ 491 0.9 and $\sigma_h = 0.7$). Again, this is consistent with the weakening of M_c that we observe 492 in ascending regions between these simulations. 493

Figure 7e,f show the differences in the frequency distributions of vertical velocities 494 between the 295 K and warmer simulations in the descending region. Here we see a de-495 crease in the frequency of all updrafts between the 295 K and warmer simulations, ex-496 cept between $\sigma_h = 0.8$ and $\sigma_h = 0.6$ between 295 K and 305 K. The patterns of change, 497 however, are quite different between panels e and f. Between the 295 K and 300 K sim-498 ulations, updrafts decrease in frequency at almost all speeds, except in the upper tro-499 posphere for $w > 3 \text{ m s}^{-1}$. Between the 295 K and 305 K simulations, however, we see 500 an increase in the frequency of all updrafts stronger than about 2 m $\rm s^{-1}$ throughout the 501 column. 502

Generally, we observe an increase in the frequency of intense vertical velocities, which agrees with previous work showing that the vertical velocity distribution becomes more positively skewed with warming (Pendergrass & Gerber, 2016). This is directly related to an increase in the mean convective available potential energy (CAPE). Figure 8 shows the mean undilute CAPE for parcels lifted from the lowest model level as a function of CRH. The maximum CAPE shifts towards wetter columns in the warmer simulations.

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Figure 8. Simulated mean convective available potential energy (CAPE) for a parcel based in the lowest model level as a function of column relative humidity.

There is also an increase in the magnitude of the maximum CAPE by about 500 J kg⁻¹ with each 5 K increase in SST.

Figure 9a,d show the average updraft fractional area in ascending and descending regions for the 295 K simulation. This is calculated as the average fraction of the CRM grid with vertical velocities exceeding a threshold value, neglecting GCM grid cells that have no updrafts at or exceeding the threshold intensity. Above about 1 m s^{-1} , the fraction of the CRM domain within each GCM grid cell that is ascending is larger in the ascending than in the descending region. As expected, strong updrafts occupy a smaller fraction of the area in both regions.

Figure 9b,c show the differences in the updraft fractional area distributions between the 295 K and warmer simulations in the ascending region. The patterns of change are similar to those for the velocity distributions. Between the 295 K and 300 K simulations, the fractional area occupied by updrafts generally increases throughout the column (except for large vertical velocities and at the upper levels). Between the 295 K and 305 K simulations, weak updrafts become much (> 10%) smaller while strong updrafts (> 1 m s⁻¹) become larger.

Figures 9e,f show the differences in the updraft fractional area distributions between the 295 K and warmer simulations in the descending region. Changes between the 295 K and 300 K simulations are small. Interestingly, between the 295 K and 305 K simulations, we see a large increase in updraft fractional area for all intensities throughout most of the column.

In summary, despite a monotonic decrease in the global mean M_c , we find non-monotonic changes when we subdivide by the ascending and descending regions. In general, con-

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Figure 9. For the 295 K simulation, the average updraft fractional area, calculated as the average fraction of the CRM grid with vertical velocities exceeding threshold values for the a, ascending and d, descending regions. b, Difference in the average updraft fractional area between the 300 K and 295 K simulations for the ascending region and e, descending region. c, Difference in the average updraft fractional area between the 305 K and 295 K simulations for the ascending region and e, descending region. c, Difference in the average updraft fractional area between the 305 K and 295 K simulations for the ascending region and e, descending region. c, Difference in the average updraft fractional area between the 305 K and 295 K simulations for the ascending region and f, descending region. White contours in b,c,e,f are drawn every ± 0.10 starting at 0.



Figure 10. Frequency distribution, in average daily percent global area covered, of column relative humidities for bins 0.02 wide.

vective updrafts become less common with increased SST, with increases in the size and frequency of the strongest updrafts. However, when looking at changes in just the ascending and descending regions, we find variations that are not consistent with SST and region. A particularly interesting result is the simultaneous decrease in M_c in ascending regions and increase in descending regions between the 295 K and 305 K simulations. That is, the ascending and descending regions are becoming more similar to each other in the 305 K simulation.

539

3.4 Humidity dependence

Up until this point, we have discussed changes to the circulation using averages of 540 quantities taken over very large areas, i.e., either the entire domain, or the ascending/descending 541 regions. In this section, we will discuss the variations to the mean circulation, convec-542 tive mass flux, and heating terms with SST in more detail by looking at these terms in 543 CRH space. In the real world, tropical temperature and pressure are close to uniform. 544 Moisture and convective morphology, however, are not. Regions of active deep convec-545 tion and mean ascent are generally located in the most humid parts of the tropics. Re-546 gions of mean descent are conversely located in drier regions. Analyzing changes to the 547 circulation in humidity space thus allows us to assess regional changes. In our global RCE 548 simulations, humid regions move around in space and time rather than being linked to 549 geographically fixed SST patterns, the presence of land and topography, or rotation, as 550 they are on earth. Figure 10 shows the average daily distribution in percent of global 551 area covered by various CRHs on the GCM grid. Values at the extremes of the distri-552 bution are rare, but become more common in the warmer SST simulations. This is seen 553 as a widening and flattening of the CRH distribution, with values drier than 0.5 and wet-554 ter than 0.8 both becoming more common with warming, at the expense of the moder-555

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Figure 11. Mean mass flux, \overline{M} (top row), and total heating \overline{Q} (bottom row), binned by column relative humidity (CRH) percentile. a, \overline{M} for the 295 K simulation. b, Difference of \overline{M} between the 300 K and 295 K simulations. c, Difference of \overline{M} between the 305 K and 295 K simulations. d-f, As in a-c but for \overline{Q} . White contours in a (d) are drawn every 0.01 kg/m²/s (4 K/day). Grey contour line in all panels are the zero contour of \overline{M} binned by CRH percentile for the a,d 295 K; b,e 300 K; and c,f 305 K simulations.

ate CRHs. An increase in the frequency of dry columns may be associated with an increased state of organization, which dries the atmosphere (Wing et al., submitted).

Figure 11a shows \overline{M} for the 295 K simulation binned by CRH percentile. We bin 558 by percentile here because of the shift in the distribution of CRH, and in order to cap-559 ture differences between the distributions while retaining all of the data. In this figure, 560 the curve separating regions of descent and ascent slopes from drier columns at low lev-561 els to wetter columns at the upper levels. Figures 11b-c show the difference of \overline{M} binned 562 by CRH percentile between the 300 K and 295 K simulations, and between the 305 K 563 and 295 K simulations, respectively. As we showed in Figure 2e, \overline{M} weakens with increas-564 ing SST between $\sigma_h = 0.8$ and $\sigma_h = 0.3$. Figure 11b-c shows that the intensification 565 of \overline{M} at low levels is occurring for most values of the CRH. 566

The largest decreases in \overline{M} below $\sigma_h = 0.3$ are in columns more humid than the 80th percentile of the CRH. Figures 11d-f show $\overline{Q} = \overline{Q_c} + \overline{Q_R}$ binned by CRH for the 295 K simulation, and its difference from the warmer simulations. Between $\sigma_h = 0.5$ and $\sigma_h = 0.3$, \overline{Q} in this area is increasing with SST, with little change in \overline{Q} below $\sigma_h =$ 0.5. We conclude that the decreases in \overline{M} that occur for columns more humid than the 80th percentile of CRH and between $\sigma_h = 0.8$ and $\sigma_h = 0.3$ are driven by increases in $\frac{\partial \overline{s}}{\partial z}$. In the part of the column where \overline{Q} is intensifying but \overline{M} is weakening, the increases in dry static stability are outpacing those of \overline{Q} .

In the previous section we showed that the global mean convective mass flux in our global RCE simulations decreases with increasing SST due to a general decrease in the frequency of weak and moderate updrafts. M_c associated with intense convection increases in the warmer SST simulations mainly because of increases in the frequency and areal coverage of strong updrafts in descending regions. We now analyze changes in M_c and related quantities in CRH space to paint a more detailed picture of what is happening in our simulations.

Figure 12a shows the total M_c binned by CRH percentile for the 295 K simulation. 582 The black contour shows the CRH that separates regions of descent (to the left of the 583 line) from regions of ascent (to the right of the line). M_c is generally much larger in the 584 the ascending region than in the descending region, and the strongest convective mass 585 fluxes are located in the wettest columns. For columns within the 30th to 70th CRH per-586 centiles, M_c has two maxima. The lower tropospheric maximum is associated with shal-587 low cumulus convection, while the upper-level maximum may be associated with weak 588 convection in stratiform clouds (Houze, 1977). Figures 12b-c are similar but for the 300 589 K and 305 K simulations, respectively. A general weakening of M_c with warming is ap-590 parent. Figures 13a-c show the same information that is presented in Figures 12a-c, but 591 panels b and c now show the difference between the warmer simulations and the 295 K 592 simulation. The largest changes are generally in the wettest columns, above the 90th per-593 centile of the CRH. 594

Figures 12d-l and 13d-l show M_c for weak, moderate, and intense convection binned by CRH percentile for each simulation and the differences in intensity-specific M_c between the 295 K and warmer simulations. Weak convection is shallow. It is the main contributor to the total convective mass flux at low levels for the driest columns, and near the tropopause for columns between the 30th and 50th percentiles of CRH. There is a secondary peak in M_c from weak convection over the the wettest columns (> 95th percentile). This may be large-scale saturated ascent, rather than weak cumulus convection.

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Figure 12. M_c binned by column relative humidity (CRH) percentile from (top row, a-c) all convection; (second row, d-f) weak convection; (third row, g-i) moderate convection; and (bottom row, j-l) intense convection. Left column (a,d,g,j) shows values for the 295 K simulation; center column (b,e,h,k) shows values for the 300 K simulation; and right column (c,f,i,l) 305 K simulation. Black contour line in each panel is the zero contour line of the mean mass flux, \overline{M} , binned by CRH for the corresponding simulation.



Figure 13. As in Figure 12 but the center and right columns now show the difference between the 300 K and 295 K simulations (center column) and 305 K and 295 K simulation (right column). Black contour line in each panel is the zero contour line of the mean mass flux, \overline{M} , binned by CRH for the (left column) 295 K, (center column) 300 K, and (right column) 305 K simulations.

Moderate convection extends higher than weak convection and has its strongest mass fluxes in the wettest columns. The convective mass flux of intense convection is located higher in the troposphere because large values of vertical velocity are more common there (see Figure 7a). It is located in columns where CAPE is large, but in wetter columns than where CAPE is largest (see Figure 8). This is because CAPE maximizes in columns with CRHs around 70%, where dry air entrainment can decelerate cumulus updrafts (e.g. Romps & Kuang, 2010).

In the previous section we showed that the total M_c decreases monotonically with 609 SST, despite competing and non-monotonic trends between the ascending and descend-610 ing regions. Between the 295 K and 300 K simulations, we find an intensification of M_c 611 in ascending regions despite a weakening of the total M_c . This is mainly due to an in-612 crease of M_c from moderate and intense convection in ascending regions. Figures 13b,e,h,k 613 show that this intensification is occurring in the wettest columns (CRH > 90th percentile). 614 The weakening of the global mean M_c between these two simulations is driven by a de-615 crease of M_c from convection of all intensities in the descending region, and is amplified 616 by the increase in α_{dn} . Figures 13b,e,h show that this weakening is spread across the de-617 scending region. 618

Between the 295 K and 305 K simulations, there is a large decrease in M_c , primarily from a weakening of M_c from moderate convection in the ascending region (Figures 6g and 13i). This is despite an intensification of M_c from intense convection in the descending region. Looking at Figures 6h, 12l, and 13l, this increase in M_c in the descending region is small, and occurs for columns wetter than the 20th CRH percentile.

Neelin et al. (2003) and Chou and Neelin (2004) discuss mechanisms responsible 624 for increases in precipitation at the centers of convective areas with simultaneous decreases 625 at their edges. For the wettest regions, there is sufficient moisture to trigger deep con-626 vection. In warmer climates, more precipitation is produced in these regions because the 627 air contains more water vapor. At the edges of the precipitation centers, precipitation 628 is inhibited by dry air input resulting from stronger low level moisture gradients. We see 629 this pattern of change in M_c , which is closely related to precipitation, in some of the dif-630 ferences between our simulations. Figure 13h,k,l, which show the change in M_c between 631 the 295 K and 300 K simulations for the moderate and intense convection categories and 632 between the 300 K and 305 K simulations for the intense category, each have a pattern 633

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of intensifying convective mass fluxes in the wettest columns and a decrease at the margins. This pattern is most evident in panel l. However, we do not observe this pattern in the difference of the overall M_c between the simulations (Figure 13b,c), and observe the reverse pattern for the change in M_c associated with weak convection (Figure 13e,f).

638

4 Discussion and Conclusions

The warming-induced weakening of the sinking in clear-sky regions can be under-639 stood by considering the tropical clear-sky energy balance. The atmosphere is contin-640 uously losing energy radiatively, and this energy sink is balanced by adiabatic descent 641 and warming. For a given radiative cooling rate, the strength of the sinking motion is 642 dictated by the mean tropical static stability profile, which is nearly constant through-643 out the real-world tropics due to the inability of the tropical atmosphere to support strong 644 pressure gradients. The tropical static stability profile roughly follows a moist adiabat, 645 and will likely become more stable with surface warming in addition to becoming more 646 stable in direct response to increased CO_2 (Merlis, 2015). Increased static stability makes 647 sinking motions in clear-sky regions more efficient, so that for a given amount of radia-648 tive cooling slower subsidence is required to maintain energy balance. 649

This argument has been used to explain why the tropical circulation appears to weaken with warming. However, the rate that mass circulates through the mean flow also depends on the fractional area covered by the ascending and descending regions, and the mean vertical velocity in the ascending regions.

We have used the weak temperature gradient approximation, which is globally applicable in our idealized numerical simulations, to develop a simple diagnostic that links the fractional area covered by the ascending region (α_{up}) to the strength of the mean circulation. We show that it is a good approximation in simulations of global radiativeconvective equilibrium (RCE) using a model with superparameterized convection. We use it to demonstrate how large increases in non-radiative heating in the ascending region are associated with the decrease in α_{up} with increasing sea surface temperature (SST).

We find that the increase in heating in the ascending region, $\overline{Q_{up}}$, is primarily due to an increase in what we have euphemistically called "stratiform heating," which includes heating due to environmental (non-convective) condensation, detrainment, and turbulence. Most of this increase in the lower troposphere, and about half in the upper tro-

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posphere, can be explained by increases in the vertical moisture gradient with warming,
which make latent heating via environmental condensation more efficient for a constant
environmental mass flux in the spaces between the convective updrafts and downdrafts.
An investigation of the changes to stratiform heating, and the relationship (if any) between stratiform heating and the heating associated with the convective mass flux, would
be an interesting subject for future work.

Byrne and Schneider (2016a) recently developed a theory aimed at explaining the 671 width of the intertropical convergence zone (ITCZ) using an energetics framework. Byrne 672 and Schneider (2016b) use this theory to analyze the observed and modeled contraction 673 of the ITCZ with warming, and determine that it is primarily the result of increased at-674 mospheric moisture, which steepens the meridional moist static energy gradient, and en-675 hances ITCZ cooling through advective and eddy heat transfer. While we similarly find 676 that increased atmospheric moisture is related to the contraction of the ascending area, 677 our analysis shows that it is associated with increased heating in the ascending area. There 678 are important differences between the framework proposed by Byrne and Schneider (2016a) 679 and that outlined here. Our simple approximation for the fractional area of the ascend-680 ing region is made possible by assuming weak temperature gradients, which allows us 681 to neglect horizontal energy transport entirely. This is a good assumption for the trop-682 ics and for our global RCE simulations. On earth, the Hadley circulation extends into 683 the subtropics where stronger horizontal temperature gradients and rotation give rise 684 to larger horizontal heat transports. Additionally, our framework applies to the total as-685 cending area, rather than the area of the zonal mean tropical circulation. Nevertheless, 686 our framework may useful for understanding warming-induced changes in the zonally asym-687 metric Walker circulation. 688

Using our diagnostic for α_{up} , we obtained an expression for the strength of the mean 689 overturning circulation, \overline{M} . It is inversely proportional to the mean tropical dry static 690 stability, and proportional to the magnitude of the product of the mean heating rates 691 over the ascending and descending regions. Increases in \overline{M} with surface warming are pos-692 sible given sufficient intensification of the heating in ascending regions and/or cooling 693 in descending regions. The rate of increase in atmospheric radiative cooling with warm-694 ing is small, however. We have shown that implausibly large increases in heating in the 695 ascending regions would be needed to strengthen \overline{M} , given the rate that dry static sta-696

⁶⁹⁷ bility strengthens with warming. This is why the strength of the mean circulation de-⁶⁹⁸ creases with warming.

An important question is whether the framework that we have proposed for relat-699 ing ascending area and the strength of the mean circulation to regional heating rates is 700 too simple to be used to understand the real world. Popp and Bony (2019) noticed a re-701 lationship between the degree to which ITCZ convection is zonally clustered and the strength 702 and width of the ITCZ's circulation. They attribute the observed and modeled link be-703 tween increased clustering and a weakening/widening of the Hadley circulation to de-704 creases in the net heating of the ITCZ, which is in part due to increased radiative cool-705 ing from a larger equatorial non-convective region. Their results suggest that the frame-706 work we have proposed may be useful for understanding observed variability in the trop-707 ical circulation. Other studies have attributed model spread in the simulation of ITCZ 708 width to differences in the modeled response of clouds and radiation to warming (Su et 709 al., 2019), much of which appears to be related to diversity in convective parameteriza-710 tions (Schiro et al., 2019). Our results emphasize the importance of the net heating in 711 the ascending, convective region for the tropical ascent area. We pose that a better un-712 derstanding of how non-radiative heating over ascending regions responds to warming 713 will help build confidence in future projections of tropical ascent area change. 714

We have also used our simulations of global RCE with superparameterized convection to investigate changes to the convective mass flux, M_c , with warming. Consistent with Held and Soden (2006), we find decreases in M_c with increased SST. Generally, these decreases are due to less frequent convection, despite intensification of the strongest updraft speeds. Our use of a super-parameterized model with an earth-like domain size makes this result particularly relevant to the real world. Previous work has investigated changes to M_c using simulations with parameterized convection (e.g., Vecchi & Soden, 2007).

⁷²² When we investigate the warming-induced changes of M_c in more detail, we find ⁷²³ that the largest changes tend to be located in the wettest columns (column relative hu-⁷²⁴ midity percentile > 0.9), where M_c is already large. However, the fractional area cov-⁷²⁵ ered by these columns is very small (see Figure 10). The change in M_c in the much broader ⁷²⁶ dry descending region, despite its very weak magnitude, plays an important role in ex-⁷²⁷ plaining why the global mean total M_c decreases monotonically with warming. These ⁷²⁸ changes are reinforced by the increase in α_{dn} with SST. For reasons we have not deter-

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mined, the change in M_c in the ascending and descending regions is non-monotonic and opposite between the regions.

A limitation of our study is the simple bulk microphysics scheme employed by the 731 model. Previous work has shown that in simulations of RCE, single-moment microphysics 732 parameterizations underestimate low clouds when compared to two-moment schemes, 733 which produce more realistic cloud fraction profiles (Igel et al., 2015). Simulated trop-734 ical ascent area has been demonstrated to be sensitive to cloud physics parameters (Schiro 735 et al., 2019). Given the importance of clouds in the atmosphere's radiative heating bud-736 get, our results may be sensitive to the microphysics parameterization. The sensitivity 737 of our results to the microphysics parameterization is left for future work. 738

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