Variations in tropical cyclone frequency response to solar and CO$_2$ forcing in aquaplanet simulations

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Abstract The response of global tropical cyclone (TC) frequency to solar and carbon dioxide radiative forcing is examined in TC-permitting aquaplanet general circulation model simulations. With an energetically consistent slab ocean lower boundary condition, the simulations show a larger response to positive radiative forcing from increased carbon dioxide than a solar constant increase with a comparable global-mean radiative forcing. Prescribed sea surface temperature (SST) simulations reveal that both the direct response to radiative forcing (radiative forcing with unchanged SST) and the patterned-SST response vary between forcing agents. The forcing-agent dependence of the patterned-SST response of TC frequency can be accounted for by the variation in simulated intertropical convergence zone shifts. The forcing-agent dependence of the direct response of TC frequency to radiative forcing can be accounted for by the variation in direct circulation changes and in normalized moist static energy deficit changes. That the direct TC response differs across forcing agents suggests that solar radiation manipulation geoengineering schemes will not return TC frequency to that of an unperturbed climate.

1. Introduction

The response of tropical cyclone (TC) activity to climate change is an aspect of the physical climate response to anthropogenic forcing with significant societal implications. There are some projections that are well constrained by physical reasoning (e.g., increase in TC precipitation with warming from the increase in water vapor concentration) and others that are less certain, such as the change in the global TC frequency [Knutson et al., 2010]. Much recent research has made use of TC-permitting general circulation models (GCMs) that are typically run with prescribed sea surface temperature (SST) boundary conditions with perturbation SSTs taken from lower-resolution coupled climate simulations [Bengtsson et al., 2007; Zhao et al., 2009; Murakami et al., 2012] or with idealized uniform SST warming perturbations [Held and Zhao, 2011; Zhao et al., 2013]. Generally, global-mean TC frequency decreases in climate change simulations, though Emanuel [2013] and Merlis et al. [2013a] found increased global frequency. A potentially important source of uncertainty in attributing past TC changes and projecting future TC changes comes from competing influences of different radiative forcing agents, such as sulfate aerosols compared to carbon dioxide. The research presented here aims to build understanding of radiative forcing dependence of TC response in idealized TC-permitting GCM simulations.

We use a combination of slab ocean and prescribed SST boundary conditions to carefully examine differences in TC frequency changes between solar constant and carbon dioxide (CO$_2$) forcing (section 2). The response of TCs to climate change can be decomposed into direct and temperature-dependent changes [Held and Zhao, 2011; Zhao et al., 2013]. The “direct response” to radiative forcing agents, which has also been called the “fast response” or “troposphere adjustment,” is widely discussed in the radiative forcing-feedback decomposition of the top-of-atmosphere (TOA) net radiation budget [e.g., Zelinka et al., 2013; Sherwood et al., 2015]. In addition to TOA radiation changes, the troposphere’s direct, temperature-independent response to increased CO$_2$ has been shown to reduce the overturning of the mean tropical circulation [Bony et al., 2013; Merlis, 2015] and this, in turn, may reduce TC genesis [Held and Zhao, 2011; Zhao et al., 2013]. An operational definition of direct responses is provided by atmospheric GCM simulations with perturbed radiative forcing and unchanged prescribed SST—this explicitly suppresses temperature-dependent changes. However, there are analogous changes in coupled ocean-atmosphere simulations with abruptly increased forcing for times before the temperature has warmed substantially [e.g., Bony et al., 2013].
We can formalize the decomposition of direct and temperature-dependent responses in the global number of TCs \( \langle N \rangle \) with the following equation for CO\(_2\)-forced changes:

\[
\frac{d\langle N \rangle}{d\text{CO}_2} \Delta \text{CO}_2 \approx \frac{d\langle N \rangle}{dT_s} \frac{\partial T_s}{\partial \text{CO}_2} \Delta \text{CO}_2 + \frac{d\langle N \rangle}{d\text{CO}_2} \Delta \text{CO}_2. \tag{1}
\]

with surface temperature \( T_s \) and the variable CO\(_2\) denoting either the carbon dioxide concentration or the radiative forcing of perturbed CO\(_2\). With the latter choice, the temperature-dependent TC response is determined by the combination of the sensitivity of TC frequency to temperature and the climate sensitivity, the change in surface temperature for a given radiative forcing. Held and Zhao [2011] found \( \partial T_s / \partial \text{CO}_2 < 0 \) in uniform SST perturbation simulations with unchanged CO\(_2\) concentration and \( \partial \langle N \rangle / \partial \text{CO}_2 < 0 \) in simulations with unchanged SST and increased CO\(_2\) concentration in TC-permitting GCM simulations [see also the results of the TC-permitting GCM intercomparison, Zhao et al., 2013].

One can consider an analogous equation to (1) for solar radiative forcing and we present the results of this decomposition for solar and CO\(_2\) forcing. This is motivated, in part, by solar radiation manipulation (also known as solar radiation management) geoengineering schemes. In the context of solar radiation geoengineering, the surface warming from a positive radiative forcing of an increase in CO\(_2\) is envisioned to be canceled by a surface cooling from a decrease in solar constant \( S_0 \). (In practice, stratospheric sulfate aerosols or another means would be used to reflect solar radiation and perturbed solar constant simulations are a convenient idealization.) This combination would return the temperature to that of an unperturbed climate state \( (\Delta T_s \to 0) \) and temperature-dependent responses would be eliminated, but the direct responses to the two forcing agents would remain:

\[
\frac{d\langle N \rangle}{d\text{SRM}} \approx \frac{d\langle N \rangle}{d\text{CO}_2} \Delta \text{CO}_2 + \frac{d\langle N \rangle}{dS_0} \Delta S_0. \tag{2}
\]

with the combination of \( \Delta \text{CO}_2 \) and \( \Delta S_0 \) chosen to minimize the surface temperature change in the solar radiation manipulation, SRM, scheme. The extent to which the direct response to CO\(_2\) and solar forcing differ then determines how close the geoengineered climate is to that of the unperturbed climate. We note that this framework of competing direct responses in solar radiation manipulation schemes can naturally be considered for climate variables beyond global TC frequency.

We focus on the possible forcing agent dependence of TC responses [Emanuel and Sobel, 2013; Sobel et al., 2016]. In addition to the solar radiation manipulation geoengineering motivation, there is interest in anthropogenic aerosols [Mann and Emanuel, 2006; Dunstone et al., 2013] for interpreting historically observed TC changes of the late 20th and early 21st century [Sobel et al., 2016]. Here we examine solar constant changes rather than anthropogenic aerosol changes, as we expect the results to be less model dependent. However, the solar constant forcing is clearly a distinct perturbation. The hemispheric structure of its radiative forcing, for example, is modest compared to that of sulfate aerosol radiative forcing. This will, therefore, provoke distinct meridional shifts in the intertropical convergence zone (ITCZ) position [e.g., Yoshimori and Broccoli, 2008], which can be viewed as the atmosphere’s energetic response to interhemispheric asymmetries in radiative forcing, radiative feedbacks, or surface fluxes [Kang et al., 2008; Yoshimori and Broccoli, 2008; Bischoff and Schneider, 2014]. We highlight this difference because of the importance of meridional ITCZ shifts in determining the TC frequency response to climate changes [Merlis et al., 2013a].

The results presented here build on the work of Merlis et al. [2013a], which included both solar and carbon dioxide (CO\(_2\)) forcing perturbations in slab ocean aquaplanet GCM simulations. Both forcing agents featured a poleward shift in ITCZ position for positive radiative forcing and simulated a concomitant increase in TC frequency. The magnitude of the forcing differed between the solar and CO\(_2\) simulations, and this precluded isolating forcing agent differences in TC response. We return to this question here.

To date, only Emanuel and Sobel [2013] have discussed physical mechanisms for the radiative forcing agent dependence of environmental TC variables. They found differences between solar and CO\(_2\) forcing in single column model simulations. Here, we assess both changes in explicitly simulated TCs (section 3) and environmental variables (section 4). Emanuel and Sobel [2013] found increased tropical cyclone potential intensity in energetically consistent, interactive surface temperature simulations with positive radiative forcing and larger increases for solar constant forcing than for CO\(_2\) forcing. While they did not examine the direct
response to radiative forcing, these can be inferred from differences between their SST-only perturbation simulations and their interactive surface temperature simulations with perturbed radiative forcing. According to this linearity assumption, the direct CO₂ changes were larger than the direct S₀ changes and both tended to offset the temperature-dependent changes in variables such as mean precipitation and tropical cyclone potential intensity. We find this competition between direct and temperature-dependent changes in many aspects of the simulation results in TC-permitting GCM simulations here.

The preceding discussion of the decomposition of changes into direct and temperature-dependent changes (1) and solar radiation manipulation geoengineering schemes (2) did not distinguish between global-mean surface temperature changes and spatially varying surface temperature changes. This is another means by which solar and CO₂ forcing may differ. In what follows, we show larger simulated changes with CO₂ forcing than solar forcing that are consistent with greater meridional shifts in the general circulation (ITCZ latitude, in particular). Simulating these shifts is precluded in column modeling approaches and one of our central findings is that temperature-dependent responses that depart from the global or tropical-mean are important in determining variations in the TC response across radiative forcing agents.

2. General Circulation Model

We use the Geophysical Fluid Dynamics Laboratory’s High-Resolution Atmospheric Model (HiRAM) at C180 resolution, which has a cubed sphere dynamical core with approximately 50-km horizontal resolution and 32 vertical levels. The convection parameterization is based on a single entraining plume [Bretherton et al., 2004] that results in a substantial portion of precipitation in the tropics occurring through the large-scale condensation scheme rather than through the convection scheme. HiRAM’s simulated TC frequency depends non-monotonically on the convection scheme’s entrainment parameter for reasons discussed by Zhao et al. [2012]. In comprehensive configurations with the observed SST prescribed, HiRAM simulates interannual variability and long-term trends of TC frequency well [Zhao et al., 2009; Shaevitz et al., 2014]. At this resolution, the full intensity distribution of TCs is not simulated (maximum surface windspeed \( \approx 40 \) m s\(^{-1}\)). The TC tracking algorithm follows previous HiRAM publications [Zhao et al., 2009; Merlis et al., 2013a].

We perform HiRAM simulations with an aquaplanet (water covered) lower boundary condition with either a slab ocean or prescribed-SST boundary condition. This idealization aids in physical interpretation and is computationally advantageous, as TCs form at all longitudes. For example, Figure 1 displays a snapshot of the surface wind speed and precipitation. The eastern part of the domain has TCs that have propagated north from their genesis region on the poleward flank of the ITCZ and the western part of the domain has TCs forming that are near the mean ITCZ latitude.

For climate change simulations, there is sensitivity to the HiRAM configuration. With comprehensive boundary conditions (e.g., Earth’s continents and observed SST for the control simulation), HiRAM has decreasing global...
TC frequency with warming [Zhao et al., 2009; Held and Zhao, 2011]. In Earth-like aquaplanet simulations with slab ocean boundary conditions, HiRAM simulates an increase in global TC frequency with warming that is associated with poleward ITCZ shifts [Merlis et al., 2013a]. The difference between the simulated TC frequency response to climate change in comprehensive and Earth-like aquaplanet HiRAM configurations may result from the relatively large, though not implausible, ITCZ shifts of \( \approx +0.6 \) lat per kelvin of tropical warming in the idealized configuration. Finally, in aquaplanet simulations with uniform thermal forcing (SST and insolation the same at all latitudes), HiRAM simulates a reduction in global TC frequency with warming for both spherical and \( f \)-plane (constant Coriolis parameter) geometries [Zhou et al., 2014; Merlis et al., 2016].

### 2.1. Slab Ocean: Control Simulation

In slab ocean simulations, the surface boundary condition is a 20 m slab of water that interactively determines the surface temperature, consistent with the surface radiative and turbulent fluxes and the convergence of a prescribed ocean heat transport. All aspects of the simulation boundary conditions and forcing are symmetric between the hemispheres, except for a prescribed northward ocean heat transport that follows the functional form of Kang et al. [2008]. The convergence of the ocean heat transport is zero equatorward of 40° latitude and is composed of a negative half sine function in the southern hemisphere extratropics and a positive half sine function in the northern hemisphere extratropics. The hemispheric asymmetry in the lower boundary condition can be thought of as an idealization of the Atlantic ocean’s cross-equatorial heat transport, though the magnitude of the prescribed cross-equatorial heat transport (2.3 PW) is substantially larger than that of Earth’s climate. The simulations are integrated for 10 years and averages over the last 5 years, when they are equilibrated, are shown. The control CO2 concentration is 300 ppmv and the control solar constant \( S_0 \) is 1400 W m\(^{-2}\). The control simulation is identical to that of Merlis et al. [2013a], and more detailed descriptions of the model configuration can be found in Merlis et al. [2013a] and Ballinger et al. [2015].

The convergence of the prescribed northward ocean heat transport warms the northern hemisphere extratropics and cools the southern hemisphere extratropics, resulting in a hemispherically asymmetric climate with a warmer northern hemisphere tropics (Figure 2). The northward ocean heat transport provokes a southward cross-equatorial atmospheric energy transport from the northern hemisphere to the southern hemisphere. Consistent with the energetically direct, cross-equatorial Hadley cell (Figure 3), the ITCZ and maximum in precipitation are located in the northern hemisphere (near 8° latitude in the control simulation, Figure 2). Figure 2 shows the normalized moist static energy (MSE, \( h = c_p T + gz + Lq \)) deficit defined by

\[
\chi = \frac{h_b - h_m}{h_1 - h_b},
\]

with saturation surface moist static energy \( h_1 \), boundary layer moist static energy \( h_b \) (evaluated at the lowest atmospheric model level, 35 m altitude), and midtroposphere moist static energy \( h_m \).
This variable $\chi$ appears in quasi-equilibrium theories for the convective mass flux and its changes have been interpreted as a measure of perturbations to the timescale with which the free troposphere can moisten and transition to the saturated, TC state [Emanuel et al., 2008]. Note that subsequent publications [e.g., Emanuel and Sobel, 2013] replaced $h_0$ with the saturation moist static energy of the free troposphere in (3), though this definition of $\chi$ will differ if convective quasi-equilibrium does not hold [Zhou, 2015]. Low values of $\chi$ are conducive to TC genesis and there is a weak local minimum near the ITCZ (Figure 2), where the free-tropospheric relative humidity is high.

The TC genesis frequency simulated in the control simulation is shown in Figure 4. The peak in genesis occurs on the poleward side of the ITCZ near 10° latitude. There are 364 TCs per year globally in this simulation. This is more than in Earth observations, but the asymmetric climate state, which is akin to a perpetual late summer environment, and the absence of continents both substantially raise the number in the idealized simulations. We include all TCs to better accumulate statistics and note where there are differences in what follows. Merlis et al. [2013a], in contrast, showed the frequency of hurricane-strength TCs. The number of TCs is sensitive to a variety of aspects of the mean climate, as the broader range of aquaplanet simulations with prescribed SST presented in Ballinger et al. [2015] shows. Our subsequent analysis is focused on radiatively forced changes with the hope that the TC changes and their relationship to environmental parameters are not sensitively dependent on the control climate state.

2.2. Prescribed SST: Control Simulation

The time- and zonal-mean SST is prescribed from the equilibrium phase of the control slab ocean simulation. This follows Ballinger et al. [2015], though they used the time-mean SST that included weak deviations...
from the zonal mean. Weak deviations from the zonal-mean SST do not noticeably affect the simulated TC frequency in the control climate. The prescribed-SST simulations are integrated for 10 years and all years are included in the averages presented.

The mean climate of the prescribed-SST control simulation is quite similar to the slab ocean counterpart (Figure 2, solid vs. dashed). The differences in precipitation and evaporation are everywhere less than 1 mm day$^{-1}$ and 0.5 mm day$^{-1}$, respectively. The differences in normalized MSE deficit $\chi$ are also small ($\approx 5\%$ in the region of interest, Figure 2).

The control prescribed-SST simulation has a similar distribution of TC genesis in latitude compared to the corresponding slab ocean simulation (black solid vs. black dashed in Figure 4). There are 397 TCs per year globally in the control prescribed-SST simulation, which is about 10% greater than the corresponding slab ocean simulation. That the global number is higher is consistent with the expectation that ocean cooling (here exclusively from the turbulent surface fluxes, a subset of the true ocean coupling) under TCs limits their intensification. The slab ocean depth of 20 m may exaggerate this discrepancy relative to larger slab depths, and other factors may also contribute to the difference in control climate TC frequency. For the perturbation simulations, we focus on the fractional changes rather than the changes in absolute number to limit the influence of the bias in the control climate $N$.

### 2.3. Radiative Forcing and Perturbation Simulations

We perform perturbation simulations where either the CO$_2$ concentration is quadrupled, 4×CO$_2$, or the solar constant $S_0$ is increased from 1400 W m$^{-2}$ to 1450 W m$^{-2}$, a 3.5% increase, for slab ocean simulations. The direct response to radiative forcing is obtained from prescribed-SST simulations with the SST from the control simulation. These also define the troposphere-adjusted radiative forcing [Hansen et al., 2005] in which the direct changes in tropospheric temperature, water vapor, and cloud fields are allowed to alter the radiative forcing estimate [see also Sherwood et al., 2015]. The global-mean radiative forcing is 7.6 W m$^{-2}$ for 4×CO$_2$ and is 7.9 W m$^{-2}$ for the 3.5% increase in solar constant.

![Figure 4. Number of tropical cyclone genesis events per year per degree latitude for (a) slab ocean simulations (Slab), (b) prescribed SST simulations with perturbed radiative forcing and unchanged SST (Direct), (c) prescribed SST simulations with perturbed SST and unchanged radiative forcing parameters (SST Only), and (d) prescribed SST simulations with perturbations to both radiative forcing and prescribed SST (Both). The control simulation is shown in black and perturbation simulations with increased solar constant in red and increased CO$_2$ in blue.](image-url)
We isolate the temperature-dependent changes by performing simulations with the spatially varying perturbation SST prescribed from the perturbation $S_0$ or CO$_2$ slab ocean simulations. These “SST Only” simulations use the control values for $S_0$ and CO$_2$ for the GCM’s radiative transfer calculations.

Finally, we perform prescribed-SST simulations with both the perturbation SST and radiative forcing. These prescribed-SST simulations with “Both” perturbations allow us to examine linearity within the prescribed-SST simulation framework (do the direct and temperature-dependent responses sum to the response found in “Both”?) and allow us to examine whether prescribed-SST simulations can reproduce the changes simulated with the slab ocean boundary condition. If not, this suggests high-resolution coupled ocean-atmosphere GCM simulations [e.g., Murakami et al., 2015] have advantages over global “downscaling” simulations where atmospheric GCMs simulations are performed with prescribed SSTs taken from lower-resolution coupled simulations. In total, we present the results of three slab ocean simulations and seven prescribed-SST simulations.

We have performed select simulations with a reduced solar constant, as would be necessary for solar radiation manipulation (2). The simulated response to the decrease in radiative forcing is quite similar to the opposite of the response to positive radiative forcing with the same forcing agent [see also, Merlis et al., 2013a].

3. TC Frequency Changes

We begin with an overview of the simulated TC frequency changes. Figure 4 shows the time- and zonal-mean genesis frequency and Figure 5 shows the percent change in global TC number $\langle N \rangle$ relative to the corresponding slab ocean or prescribed-SST control simulation. We first present the slab ocean simulations and then present the prescribed-SST simulations used to decompose the TC response into direct and temperature-dependent components (following equation (1) and the analogous equation for solar forcing).

3.1. Slab Ocean TC Frequency Changes

The simulated global number of tropical cyclones increases for both solar (red) and CO$_2$ (blue) radiative forcing (Figure 5). This result is consistent with Merlis et al. [2013a]. The region of genesis shifts poleward with warming (Figure 4a), following the poleward shift in the atmospheric general circulation (Figures 3 and 6d).

The simulations presented in Merlis et al. [2013a] had a larger magnitude solar constant perturbation (a 100 W m$^{-2}$ increase there vs. a 50 W m$^{-2}$ increase here) and did not document a difference between forcing agents. Here, with comparable global-mean radiative forcing, it is clear that there are differences: there is a larger increase in TC number under CO$_2$ forcing (15%) than under solar forcing (9%). These changes are about a factor of four lower than the changes in hurricane-strength TCs [Merlis et al., 2013a], indicating changes in the intensity distribution [see also Ballinger et al., 2015, Figure 3].

The northern hemisphere tropics (between the equator and 30° N) warms by 4:0 K and 4:1 K for CO$_2$ and solar forcing, respectively. The tropical warming is comparable to the global-mean warming, though there is somewhat more CO$_2$-forced global warming (4.4 K) than solar-forced warming (4.1 K). The variation in tropical-mean warming is consistent with the slightly higher solar radiative forcing, but this cannot account for the variation in TC frequency response. If the change in $\langle N \rangle$ was rescaled by the global- or tropical-mean warming, the variation in $\langle N \rangle$ would only be slightly altered. Therefore, the variation in global TC frequency response between forcing agents may be associated with
either differences in the spatially varying temperature-dependent response (as can be operationally defined by differences in SST change pattern, but may more fundamentally be related to differences in spatial variations in perturbation energy budgets) or differences in the direct responses. We use prescribed-SST simulations to address the contribution of each of these two means of radiative-forcing dependent response next.

### 3.2. Direct TC Frequency Changes

The genesis region does not change in the direct response to increased radiative forcing (SST unchanged from control) and the peak of TC genesis remains near 10° latitude (Figure 4b). Changes in genesis frequency are subtle in Figure 4b; however, there are systematic changes for 4×CO₂ (Figure 5). The direct response of TC frequency to an increase in CO₂ with unchanged SST is a reduction (Figure 5), as previously described by Yoshimura and Sugi [2005] and Held and Zhao [2011]. The aquaplanet simulations here have a reduction of 6% for the 4×CO₂ simulation (Figure 5). This is smaller than the 10% reduction for 2×CO₂ in

![Figure 6. Time- and zonal-mean change in (first row) SST, (second row) precipitation, (third row) evaporation, and (fourth row) normalized moist static energy deficit (dimensionless). (a, d, g, j) Slab ocean (dashed lines) and prescribed-SST simulations with both radiative forcing and SST changes (solid lines) are shown in the left column. (b, e, h, k) Prescribed-SST simulations with unchanged SST and perturbed radiative forcing are shown in the middle column and (c, f, i, l) prescribed-SST simulations with perturbed SST and unchanged radiative forcing parameters are shown in the right column. Solar-forced changes are shown in red and CO₂-forced changes are shown in blue. The latitude of maximum precipitation in the control simulation is indicated with a gray vertical line.](image)
comprehensive boundary condition simulations with the same GCM [Held and Zhao, 2011]. This possibly results from a lower sensitivity of TC genesis to weakening ascent when the control circulation is strong, as is the case here because a representation of tropical ocean heat transport is not included and the atmosphere, therefore, transports more energy and mass to balance the energy budget [Held, 2001; Levine and Schneider, 2011].

The direct response of global TC frequency to an increase in \( S_0 \) is near zero in these simulations (Figure 5). There are, however, changes in the intensity distribution for both direct forcing simulations that shifts the peak to weaker intensity. This is consistent with decreasing potential intensity, though other factors may also contribute to alter the intensity distribution. Given that the slab ocean 4×CO₂ simulation has a larger increase in \( \langle N \rangle \) than the simulation with a +3.5%\( S_0 \) perturbation, the CO₂ direct response reduction does not account for the radiative forcing agent dependence of the TC changes found in the slab ocean (Figure 5).

3.3. SST-Only TC Frequency Changes

The temperature-dependent TC frequency response is an increase in the global mean and a poleward shift in the region of genesis (Figs. 4 and 5). These changes are broadly similar to the changes in the slab ocean simulations, though there are quantitative differences. The temperature-dependent response of \( \langle N \rangle \) is +12 % for CO₂ and +6% for solar forcing. The 6% larger increase in \( \langle N \rangle \) for CO₂ compared to solar SST-only simulations is similar to the 7% larger increase in \( \langle N \rangle \) between CO₂-and solar-forced slab ocean simulations, but the direct response tends to offset this difference.

The northern hemisphere tropical-mean warming is similar (differing by \( \sim0.1 \) K); however, there are subtle differences in the pattern of the SST perturbations between solar and CO₂ forced simulations (Figure 6a). The differences simulated here (\( \sim0.2 \) K) are smaller in magnitude than those in comprehensive coupled simulations forced by single forcing agents [e.g., aerosol-only and greenhouse gas-only simulations analyzed in Xie et al., 2013]. This suggests (i) a limitation of uniform warming perturbation simulations if one is interested in radiative forcing agent dependence and (ii) an exceedingly precise knowledge of SST changes is needed if one wants to use these changes to interpret TC changes.

3.4. Both (Perturbed SST and Radiation) TC Frequency Changes

Figure 5 shows that the prescribed SST simulations with both perturbed SST and radiative forcing are similar to the sum of the individual responses of the two prescribed-SST simulations for CO₂, though there is evidence of non-additivity for the solar forcing. In these simulations, the global TC frequency increases by 7% under CO₂ forcing and increases by 4% for solar forcing. Relative to perturbation SST-only simulations, “Both” simulations include the direct response that reduces the difference between the forcing agents. The combined response of \( \langle N \rangle \) in prescribed SST simulations is then about half as large as the slab ocean response, which naturally includes both SST and radiative forcing changes.

The weaker magnitude of the response of the global-mean TC frequency in prescribed-SST simulations with both perturbed SST and radiative forcing compared to the slab ocean simulations suggests that time-fluctuations in the SST may need to be prescribed to reproduce the full response of simulations with interactive SST in prescribed-SST simulations. The SST fluctuations themselves may or may not be the ultimate cause of the differences. One can envision time-variations in the large-scale circulation or thermodynamic fields are suppressed when a time-independent SST is prescribed and it is the variations in these other fields, rather than the SST itself, that is key for TC genesis.

4. Tropical Environment Changes

We present several environmental variables in the aquaplanet HiRAM simulations. Our choice of variables is motivated by (i) those that have been argued to be important for global TC frequency changes in response to radiative forcing and the accompanying warming and (ii) those used in tropical cyclone genesis indices [e.g., Emanuel and Nolan, 2004; Tippett et al., 2011]. We do not explicitly evaluate TC genesis indices here, as Camargo et al. [2014] has shown that they have difficulty capturing the direct response to increased CO₂.

Held and Zhao [2011] argued that decreases in mean ascending mid-tropospheric vertical velocity in genesis regions, as a proxy for the level of convective activity, account for reduced global TC frequency for both
direct and temperature-dependent responses to CO₂ [see also Zhao and Held, 2012]. Changes in ascending vertical velocity were largely successful in capturing the simulated global TC frequency changes in the TC-permitting model intercomparison described in Zhao et al. [2013]. Here, we show changes in the Eulerian-mean stream function from which changes in mean vertical velocity can be inferred (Figure 3).

Emanuel et al. [2008] argued that the normalized moist entropy deficit, defined analogously to (3) but with equivalent potential temperature, increases in warming scenarios and this tends to suppress genesis. They also showed its increase accounted for a reduction in genesis in TC downscaling simulations of future climate projections. The normalized MSE or moist entropy deficit increases under warming because the numerator increases (in proportion to the saturation deficit \(\sim 7\% \text{K}^{-1}\), if relative humidity is unchanged) more rapidly than the denominator (in proportion to the energetically constrained surface flux change \(\sim 2\% \text{K}^{-1}\)). The increase in \(\chi\) can be interpreted as an increase in the timescale for the subsaturated mean state of the free troposphere to transition to the saturated free troposphere state of a TC via surface evaporation. Climate change simulations robustly feature increases in the normalized MSE deficit, though the subsequent TC downscaling simulations of Emanuel [2013] had an increase in global TC frequency. This variable \(\chi\) has also been incorporated in genesis indices [Emanuel, 2010].

Last, Merlis et al. [2013a] showed a sensitive dependence of TC genesis on ITCZ latitude. The ITCZ can shift poleward from changes in the cross-equatorial ocean heat transport or as a result of internal atmospheric feedbacks under increased radiative forcing. Warmed climate states with unchanged ITCZ position (brought about by a combination of increased radiative forcing and decreased ocean heat transport in aquaplanet slab ocean simulations) had a decrease in global TC frequency, consistent with comprehensive boundary condition simulations under uniform warming [Held and Zhao, 2011]. A poleward shift in the ITCZ increases the relevant planetary vorticity component of the absolute vorticity, a standard variable in genesis indices.

Tropical cyclone potential intensity (PI) also commonly appears in TC genesis indices. The slab ocean and perturbed SST simulations have increases in PI and the direct response to both forcing agents is a decrease in PI. The magnitude of these PI changes has some radiative forcing agent dependence, with larger CO₂-forced than solar-forced changes for the direct response decrease. The temperature-dependent increases found in the other simulation types (slab, SST only, and “Both”) have weak radiative forcing agent dependence (\(\sim 1\%\) compared to an increase of \(\sim 10\%\)), with typically slightly larger solar-forced increases in the TC genesis regions. We focus on evaporation changes because they contribute to PI changes, which depend on the air–sea enthalpy disequilibrium [Emanuel, 1987; Wing et al., 2015], and their changes shed light on how the denominator of the normalized MSE deficit \(\chi\) changes.

Additional environmental factors that commonly appear in TC genesis indices include the vertical wind shear and relative humidity of the free troposphere [Emanuel and Nolan, 2004]. We do not show these variables, as these are favorable for TC genesis near the ITCZ and their changes do not appear to be the dominant factors in accounting for the changes in TC genesis.

In what follows, the simulated changes in the mean SST, precipitation, evaporation, normalized MSE deficit, and Hadley circulation are presented for the slab ocean simulations, followed by the prescribed-SST simulations used to decompose the changes into direct and temperature-dependent responses (Figure 6).

### 4.1. Slab Ocean Tropical Environment Changes

The slab ocean simulations have a patterned SST change with greater warming poleward of the control simulation ITCZ than equatorward of it (Figure 6a). There is a sharper gradient in the SST change in the CO₂-forced simulation than in the solar-forced simulation. The change in precipitation is a dipole with a decrease equatorward of and near the control simulation ITCZ latitude (8° latitude) and increased precipitation poleward of this (Figure 6d). The Eulerian-mean stream function has substantial positive anomalies near the zero stream function boundary between the cross-equatorial and summer hemisphere cells, indicating a poleward shift in this boundary and in the location of maximum ascent (Figure 3). The poleward shift of the precipitation maximum is larger in the CO₂-forced simulation than in the solar-forced simulation: maximum in P shifts 3.0° latitude poleward for CO₂ forcing and 2.5° latitude poleward for solar forcing. The difference in ITCZ shifts between the two forcing agents is significant at the 1% level according to a Student’s t test using either the time series of annual or monthly mean ITCZ latitude. In addition to poleward shifts, the stream function changes imply a reduced magnitude of ascent in the genesis region.
In the energetic framework for ITCZ latitude \( \phi_i \), the ratio of the vertically integrated, northward MSE flux at the equator \( \frac{\langle \mathbf{v} \mathbf{n} \rangle}{\langle \mathbf{v} \mathbf{n} \rangle} \) and the net energy input to the atmospheric column determine the ITCZ latitude [Kang et al., 2008; Yoshimori and Broccoli, 2008; Bischoff and Schneider, 2014]. For equilibrated climates with slab ocean boundary conditions, the net energy input to the atmospheric column is simply the TOA net radiation \( N_{TOA} \) and the ITCZ latitude is given by the following expression:

\[
\phi_i \approx - \frac{1}{a} \frac{\langle \mathbf{v} \mathbf{n} \rangle}{N_{TOA}},
\]

with Earth radius \( a \) [see Bischoff and Schneider, 2014 for derivation]. An examination of the cross-equatorial atmospheric energy transport reveals that the CO2-forced and solar-forced simulations have comparable changes in the cross-equatorial atmospheric energy transport \( \frac{\langle \mathbf{v} \mathbf{n} \rangle}{\langle \mathbf{v} \mathbf{n} \rangle} \). This implies the two forcing agents have comparable interhemispheric changes in the temperature-dependent component of the perturbation TOA energy budget—the interhemispheric radiative feedbacks that determine the numerator in equation (4). The solar-forced simulations, then, have weaker ITCZ shift as a result of the larger magnitude change in tropical-mean \( N_{TOA} \) (the denominator in equation (4)) compared to CO2-forced changes. This, in turn, is a consequence of the spatial pattern of the radiative forcing. For the same global-mean forcing, the solar constant forcing is larger in low latitudes, consistent with the distribution of the insolation, than it is for CO2 forcing. The larger tropical-mean solar forcing could in principle be offset by a differing tropical-mean radiative feedback, but that is not critical in these simulations or likely in general given the similarity of TOA response to different forcing agents [e.g., Merlis et al., 2014]. This line of reasoning relates the weaker ITCZ shift to an aspect of the radiative forcing, rather than the more numerous factors that can alter the spatial structure of the SST changes.

Evaporation increases in the tropical mean (Figure 6g), as expected [O’Gorman et al., 2012]. There is latitudinal structure in the changes that is associated with the ITCZ shift because this is where the minimum in evaporation occurs (Figure 2). There is a bigger increase in evaporation for solar forcing than CO2 forcing, consistent with energetic arguments for changes in the hydrological cycle [O’Gorman et al., 2012; Emanuel and Sobel, 2013].

The overall environmental response is consistent with the discussion of Merlis et al. [2013a]: factors that tend to reduce TC genesis under warming, like a reduction in mean ascent or an increase in the normalized MSE deficit, are overwhelmed by the poleward ITCZ shift that tends to increase genesis. Here, the variation in simulated TC frequency changes between the forcing agents can be accounted for by the variation in the ITCZ shifts: CO2-forced simulations have a larger poleward shift than solar-forced simulations. This straightforward description neglects the variations in the direct response of TC frequency and the tropical environment, which we discuss next.

### 4.2. Direct Tropical Environment Changes

The direct changes in the mean climate are smaller than those simulated with slab ocean boundary condition or perturbed SST simulations (shown next). Note that the range of vertical axis values of the middle column of plots in Figure 6 is smaller than the other columns. The SST change is identically zero in these simulations (Figure 6b); however, there are changes in the hydrological cycle. The direct response to 4× CO2 is a decrease in precipitation near the ITCZ (Figure 6e) that is consistent with the weakening cross-equatorial Hadley circulation (the stream function changes are negative, shown in blue colored contours, and coincide with the positive stream function climatology, in solid contours, of the cross-equatorial cell, Figure 3). This weakening of the tropical overturning circulation is robustly simulated in comprehensive GCMs [Bony et al., 2013] and arises from the spatially patterned CO2 radiative forcing that reduces the required atmospheric energy transport [Merlis, 2015]. The spatially patterned CO2 radiative forcing in the tropics results from the climatological distribution of clouds and water vapor—the additional CO2 has less of an impact on the TOA radiation in the presence of these other longwave absorbers [Zhang and Huang, 2014; Huang et al., 2016]. The direct response in the Hadley circulation and precipitation are weaker for solar forcing than CO2 forcing (Figs. 3 and 6e), though there is also a slight precipitation decrease and weakening of the stream function maximum for solar forcing. We suggest this results from more spatially uniform solar radiative forcing that arises from both low and high clouds influencing the planetary albedo, in contrast to
the spatial pattern of tropical CO$_2$ forcing that is sensitive to high clouds, which re-emit from cold temperatures, but not low clouds, which re-emit from warm temperatures.

The direct response of evaporation to radiative forcing is a reduction for both solar and CO$_2$ forcing [Bala et al., 2010 and Figure 6h here]. The tropical-mean evaporation response can be understood from atmospheric energy balance considerations: the latent heating in the atmosphere (equal to the surface latent heat flux and proportional to evaporation in the mean) balances net radiative flux changes, where both surface and TOA fluxes alter the atmospheric energy balance. Atmospheric absorption of solar radiation—primarily by water vapor—leads to an increase in solar absorption within the atmosphere when the solar constant is increased. This, in turn, is balanced by a decrease in latent heating. The direct radiative response to increased CO$_2$ is comprised of a TOA flux decrease that is larger than the increase in the net longwave flux to the surface. This combined decrease in the longwave radiative loss by the atmosphere is balanced by a decrease in latent heating [O’Gorman et al., 2012]. These energetic arguments constrain the global-mean response of the hydrological cycle, though they offer useful guidance for the tropics as well, provided the energy transport between the tropics and extratropics do not change dramatically.

The direct response of the normalized MSE deficit is an increase, consistent with the reduction in evaporation, that is larger for CO$_2$ than for solar forcing (Figure 6k). The variation in the direct response of the normalized MSE deficit between solar and CO$_2$ forcing arises from the varying response in evaporation (proportional to $\gamma$’s denominator) and also a reduction in free-troposphere relative humidity in response to $4\times$CO$_2$ (proportional to $\gamma$’s numerator) that is absent from the solar-forced simulation. Direct CO$_2$-forced changes in relative humidity have been previously documented in GCM simulations [Colman and McAvaney, 2011]. In the region of interest here, this reduction in tropospheric relative humidity appears to be a natural consequence of the reduction in time-mean ascent (Figure 3).

The reduction in global TC frequency in the CO$_2$ forcing simulation is consistent with the weakened ascent and the increased normalized MSE deficit. In contrast, these changes are weaker for solar forcing and there is not an appreciable change in global TC frequency (Figure 5).

4.3. SST-Only Tropical Environment Changes

In prescribed-SST simulations with perturbed SSTs taken from the $4\times$CO$_2$ or solar-forced slab ocean simulations, the simulated mean climate response bears substantial similarities to the slab ocean mean climate response. Precipitation, evaporation, and the normalized MSE deficit all have similar pattern of temperature-dependent changes as the slab ocean simulations (Figures 6f, 6i, and 6l compared to Figures 6d, 6g, and 6j). The magnitude of the changes in the SST-only simulations is slightly larger than the slab ocean simulations, consistent with partly offsetting direct responses. The temperature-dependent changes in precipitation (Figure 6f) and the Eulerian-mean stream function (Figure 3) are the dominant changes in the total response, as simulated with slab ocean or prescribed-SST simulations with both SST and radiative forcing perturbed (described next). The sensitivity of TC genesis to the poleward shift in ITCZ position in these SST-only simulations suggests this is the dominant temperature-dependent genesis response (Figure 4c).

The normalized MSE deficit increases in response to the SST perturbation (Figure 6l) and TC genesis-weighted ascending vertical velocity decreases. Both of these changes in isolation would tend to decrease genesis, though it increases in these simulations.

While we use these prescribed, perturbation-SST simulations to determine the temperature-dependent changes, we do not advocate that the patterned SST is the ultimate cause of the variation in ITCZ shifts (see the preceding discussion of energetic arguments for ITCZ shifts). Kang and Held [2012] have shown the limitations of interpreting changes in SST as being the underlying cause of ITCZ shifts and Merlis et al. [2013b] have shown counter-intuitive relationships between changes in surface temperature gradients and tropical circulations that can be accounted for by analyzing circulation energetics.

4.4. Both (Perturbed SST and Radiation) Tropical Environment Changes

We examine the simulated changes in the tropical environment in prescribed-SST simulations with both SST and radiative forcing parameters perturbed with an eye toward addressing whether these simulations have different TC frequency response than the corresponding slab ocean boundary condition simulations (Figure 5) because of differences in the simulated changes in the mean tropical environment. Figure 6
shows that there are only modest differences in the mean climate changes that are similar under the different boundary conditions (dashed vs. solid lines in Figures 6d, 6g, and 6j).

By definition, the SST is the same as the slab ocean simulations (Figure 6a). Precipitation changes both show a poleward shift and the CO2-forced shift is larger than that from solar forcing—both of which are similar to slab ocean simulations (dashed vs. solid in Figure 6d). The evaporation change is similar to that of the corresponding slab ocean simulation and has comparable variations between solar and CO2 forcing (Figure 6g). Likewise, the normalized MSE deficit increases more for CO2 than solar forcing and these changes follow the slab simulations closely (Figure 6j).

From the simulated mean climate, it would be difficult to anticipate the more muted change global TC frequency in these prescribed-SST simulations compared to the slab ocean simulations (Figure 5). This potentially poses a difficulty for TC genesis indices: even if the mean and perturbation slab ocean simulations were used to optimally construct an index [e.g., Tippett et al., 2011; Camargo et al., 2014], it would not necessarily capture the result that the similar environmental change simulated here has a weaker TC frequency response in the prescribed-SST simulations. Either time-variations in the mean circulation and thermodynamic environment or air–sea coupling influences on climate changes (via transient disturbances such as convectively coupled waves) are potential candidate explanations for the discrepancy in TC genesis between boundary conditions.

5. Conclusions

We have analyzed TC-permitting simulations of solar and CO2 forcing with aquaplanet boundary conditions. The global TC genesis frequency has a weaker response to solar forcing than to CO2 forcing. The simulations have comparable global-mean radiative forcing and comparable tropical-mean surface warming (Figure 6a). This leaves two candidates for the varying TC response: (i) direct responses and (ii) temperature-dependent responses associated with patterned warming. We assessed these using prescribed SST simulations with (i) unchanged SST and increased radiative forcing for the direct response and (ii) patterned-perturbation SST simulations with unchanged CO2 and solar constant for the temperature-dependent responses.

Both the direct and temperature-dependent responses varied across the forcing agents. The direct CO2 response is a decrease in TC frequency, as has been found in comprehensive boundary condition simulations [Yoshimura and Sugi, 2005; Held and Zhao, 2011], while there is little direct solar forcing response in TC frequency. This is consistent with (i) a larger direct weakening of the Hadley circulation for CO2 and therefore weakened ascent in the genesis region and (ii) a larger CO2-forced increase in the normalized MSE deficit \( \chi \) that results from radiative forcing agent dependence of the evaporation changes and free-troposphere relative humidity. Both the mean ascent and \( \chi \) have been argued to be important in determining genesis changes [Emanuel et al., 2008; Held and Zhao, 2011].

The temperature-dependent response is a larger increase for CO2 than for solar forcing in these simulations. This is associated with a larger poleward shift in the ITCZ, and a sensitive dependence of TC genesis on ITCZ position in this idealized simulation configuration [Merlis et al., 2013a]. The magnitude of the temperature-dependent response dominates the direct response. The total response to a combined change in SST and radiative forcing (Both) in prescribed-SST simulations is more muted than the simulated changes with slab ocean boundary conditions. This is plausibly the result of suppressed changes in low frequency variability of the tropical environment or higher frequency transient disturbances with prescribed-SST boundary conditions.

The varying direct response between solar and CO2 forcing, both in TC genesis and in the overturning circulation, is one way in which solar radiation management will not return the climate to its unperturbed state, even if the surface temperature changes are eliminated. This potentially has important consequences for the regional hydrological cycle that warrant further examination.

Our simulations and analysis of the variations between solar constant and CO2 radiative forcing should provide a useful benchmark for analyses of the physical mechanisms underlying the TC response to anthropogenic aerosols. The decomposition of changes into direct and spatially patterned temperature-dependent responses provides guidance on sources of forcing-agent dependence. We note that the geographic structure of anthropogenic aerosol forcing is weighted less heavily toward the tropics than solar constant
perturbations, so the direct responses to the tropical environment and TCs may be even weaker. In contrast, the seasonality of solar and aerosol forcing is an important factor that the simulations with time-independent forcing presented here do not include. The greater summer-season radiative forcing of either anthropogenic aerosol forcing or comprehensive simulations with perturbed solar constant forcing mediates convergence zone changes [Merlis, 2016] and may be one reason why TC-season PI is more sensitive to aerosol forcing than greenhouse gas forcing in climate simulations of the last century [Sobel et al., 2016]. A thorough examination of the seasonal cycle of tropical climate changes across radiative forcing agents would complement the time-independent climate states analyzed here.

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