Some Counterintuitive Dependencies of Tropical Cyclone Frequency on Parameters in a GCM

MING ZHAO
GFDL/UCAR, Princeton, New Jersey
ISAAC M. HELD AND SHIAN-JIANN LIN
NOAA/GFDL, Princeton, New Jersey

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ABSTRACT

High-resolution global climate models (GCMs) have been increasingly utilized for simulations of the global number and distribution of tropical cyclones (TCs), and how they might change with changing climate. In contrast, there is a lack of published studies on the sensitivity of TC genesis to parameterized processes in these GCMs. The uncertainties in these formulations might be an important source of uncertainty in the future projections of TC statistics.

This study investigates the sensitivity of the global number of TCs in present-day simulations using the Geophysical Fluid Dynamics Laboratory High Resolution Atmospheric Model (GFDL HIRAM) to alterations in physical parameterizations. Two parameters are identified to be important in TC genesis frequency in this model: the horizontal cumulus mixing rate, which controls the entrainment into convective cores within the convection parameterization, and the strength of the damping of the divergent component of the horizontal flow. The simulated global number of TCs exhibits nonintuitive response to incremental changes of both parameters. As the cumulus mixing rate increases, the model produces nonmonotonic response in global TC frequency with an initial sharp increase and then a decrease. However, storm mean intensity rises monotonically with the mixing rate. As the strength of the divergence damping increases, the model produces a continuous increase of global number of TCs and hurricanes with little change in storm mean intensity. Mechanisms for explaining these nonintuitive responses are discussed.

1. Introduction

About 90 tropical cyclones (TCs) develop each year around the globe. At present there is little understanding of what controls this number. Neither is there a consensus on how it will change with changing climate, including that associated with the anthropogenic greenhouse gas warming. Much of the difficulty is due to the large range of spatial and temporal scales associated with TC genesis. Recently, high-resolution global climate models (GCMs) have been increasingly utilized to attack these problems even though current resolutions (typically 20–100 km) are still too coarse to adequately resolve aspects of individual storm structures and intensities (e.g., Sugi et al. 2002; McDonald et al. 2005; Yoshimura et al. 2006; Oouchi et al. 2006; Bengtsson et al. 2007b; Gualdi et al. 2008; LaRow et al. 2008; Zhao et al. 2009; Sugi et al. 2009; Held and Zhao 2011). Areas of agreement and disagreement are reviewed by Knutson et al. (2010).

There is some evidence that the simulated TC statistics may significantly depend on the details in the formulations of physics and dynamics in these models (e.g., LaRow et al. 2008). For example, the simulated present-day global annual number of TCs is about 105 in Zhao et al. (2009) and roughly 78 in Oouchi et al. (2006). Further, in contrast to an overproduction of western Pacific storms in Zhao et al. (2009) and Bengtsson et al. (2007a), Oouchi et al. (2006) reported a large underestimate of the western Pacific storms in their model. It is not clear what causes these differences and how they might impact the simulated TC response to warming. A direct comparison of these results is often complicated by the fact that the differences among the various studies often involve...
simultaneous changes of multiple elements including models, forcings, and TC detecting and tracking algorithms. Therefore, it would be useful to pursue a systematic study of TC sensitivity under a more controlled condition in order to focus on specific aspects of the physics in a model. As one step in this direction, we investigate the sensitivity of the global number of TCs in the context of present-day simulations using a TC-permitting AGCM with altered physical parameterizations.

It has long been noted that GCM-simulated tropical mean climate as well as transient activity depend sensitively on the parameterization of moist convection (e.g., Slingo et al. 1994; Zhang and McFarlane 1995; Anderson et al. 2004; Lin et al. 2006; Benjamin et al. 2008). In particular, an inhibition of the parameterized deep convection through enhanced lateral mixing into convective cores generally leads to a colder and drier troposphere as well as increased tropical transient activity, both of which could have profound impact on TC genesis. Therefore, it would be interesting to study the response of TC statistics to an alteration of this cumulus mixing rate.

Explicit numerical diffusion is typically required in GCMs to suppress grid-scale noise and enhance model stability in integration. With our finite-volume formulation of the transport processes, vorticity, potential temperature, and all tracers are transported consistently by the same monotonicity-preserving scheme (Lin and Rood 1997; Lin 2004). Therefore, there is no need to add additional diffusion terms to any of these prognostic variables. However, the divergent component of the horizontal flow remains undamped. To control the grid-scale noise, we have utilized a scale-selective biharmonic form of hyper-diffusion to the divergence component. We refer to it here as the divergence damping.

The impact of divergence damping occurs primarily at points where there exists concentrated diabatic heating associated with resolved-scale moist convection (i.e., resolved-scale condensation and induced vertical motion). Resolved-scale convection can become more frequent as a GCM moves toward finer spatial resolution and/or as its parameterized deep convection is more inhibited. Since resolved-scale convective velocities and induced subsidence at neighboring grids tend to be exaggerated in a GCM of relatively coarse spatial resolution with the hydrostatic assumption, the divergence damping helps to alleviate the flow distortion by smoothing out localized upward and downward velocities. In the tropical atmosphere, convective heating associated with cumulus clouds is distributed horizontally quickly through subgrid-scale internal gravity waves (Bretherton and Smolarkiewicz 1989) and the troposphere rarely gets fully saturated with overturning motion over a typical scale of current GCM grids. As the spatial resolution of a GCM continuously improves, more convection tends to be resolved, yet a hydrostatic model will increasingly overestimate resolved-scale convective velocities (Pauluis and Garner 2006) and their suppression effect on neighboring grid points. Therefore, the divergence damping can be thought of as an important part of convection representation in a hydrostatic model that significantly relies on resolved-scale convection for tropical convective transport. This representation can potentially affect convection organization and therefore TC genesis in a GCM.

While the cumulus mixing rate has long been recognized to be important in GCM simulations, the impact of divergence damping or related horizontal mixing schemes is usually not considered as important except for the control of small-scale noise. It is a common practice for GCM modelers to use these parameters to optimize their simulation results. In this paper, we investigate how both parameters can affect GCM simulated TC statistics. Section 2 briefly describes the model and the physics perturbation experiments. Section 3 presents the results. Some possible connections between the global number of TCs and the dynamic and thermodynamic properties of the atmospheric environment are further explored in sections 4–6. Finally, section 7 provides a summary and discussion.

2. The model and perturbation experiments

The model used for this study is a newer version (below referred to as HIRAM2.2) of the Geophysical Fluid Dynamics Laboratory (GFDL) High Resolution Atmospheric Model (HIRAM) utilized in Zhao et al. (2009), Zhao and Held (2010), and Held and Zhao (2011) for studies of global hurricane climatology, variability, and change with global warming. The main difference is that the HIRAM2.2 incorporates a new land model (LM3) to be described in a future paper (P. C. D. Milly 2012, personal communication). The atmospheric dynamical core of the model was also updated to improve efficiency and stability. As a result of these changes, there are minor retunings of the atmospheric parameters in the cloud and surface boundary layer parameterizations necessary to achieve the top-of-atmosphere (TOA) radiative balance. This model is also the version of HIRAM used for the GFDL participation in the Fifth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC AR5) high-resolution time-slice simulations, and will be documented further in upcoming papers related to those simulations.

We first pursued the control simulation by prescribing the climatological sea surface temperature (seasonally varying with no interannual variability) using time-averaged (1981–2005) Hadley Centre Global Sea Ice and Sea Surface Temperature (HadISST) data (Rayner et al.
2003). We have also verified that the new model produces very similar characteristics of TC statistics compared to those in the earlier version of the model in Zhao et al. (2009). For the physics perturbation experiments, we conducted identical runs as the control experiment except with separate incremental changes in the two parameters.

The first set of experiments modifies the nondimensional parameter $e_0$ that controls the cumulus lateral mixing rate. The model uses a modified version of the convection scheme of Bretherton et al. (2004), which assumes a single bulk entraining and detraining plume. The cumulus mixing rate used in the model is $e_0/H$, where $H$ is the convective depth determined from the previous time step for each grid column. The mixing generates inhomogeneous mixtures and the entrainment/detrainment profiles are determined based on a parcel buoyancy sorting algorithm. For details about the formulation of cumulus mixing, entrainment and detrainment, readers are referred to Bretherton et al. (2004) [note that $e_0$ here is the same as $e_0$ in their Eq. (18)].

We perturb $e_0$ from the default value 10 in Zhao et al. (2009) to 6, 8, 12, and 14 (below referred to as E6, E8, E12, and E14). As we mentioned in the introduction, an increase of $e_0$ leads to enhanced cumulus lateral mixing and entrainment rate that reduces the buoyancy of cumulus clouds and results in shallower convective depth. The overall effect is a reduction of the parameterized deep convective activity.

The second set of experiments modifies the divergence damping parameter $d_0$. The divergence damping term is added to the vector invariant form of the momentum equation in the following form:

$$\frac{\partial \mathbf{V}}{\partial t} = -\left(\frac{d_0 A_0}{\Delta t}\right) \mathbf{V}(\nabla^2 D) + \text{other terms}, \quad (1)$$

where $\mathbf{V}$ denotes the vector of horizontal wind, $D$ is the divergence of horizontal wind, $\Delta t$ is the dynamical time step, $A_0$ is the smallest value of the finite-volume cell area [the corner cell of the cubed sphere; $A_0 = 1.652 \times 10^9 \text{ m}^2$ for the current (C180) resolution], and $d_0$ is a nondimensional constant coefficient. See Lin and Rood (1997) for description of the rest of the numerics. Taking the curl of Eq. (1) shows that the vorticity field is not directly affected by the damping term. Applying the divergence operator to Eq. (1) yields a $\nabla^4$ form of a hyperdiffusion for the divergence field with $d_0$ setting the strength of the diffusion. We vary $d_0$ from its default value of 0.16 to 0, 0.06, 0.12, and 0.18 (below referred to as D0, D6, D12, and D18). Further increasing $d_0$ leads to numerical instability of the model due to the explicit formulation of this damping term, which prevents us from assessing the impact of this parameter over a larger range.

Both the control and the physics perturbation experiments are 20 yr long. The storm detection and tracking algorithm is the same as that described in Zhao et al. (2009) except that the surface wind speed now uses 10-m winds instead of the lowest model level winds (35 m) in Zhao et al. (2009). The purpose of this modification is to generalize the storm detection and tracking algorithm so that it does not depend on the vertical discretization of the model. Comparison of the wind speed statistics between 10 m and the model lowest level (35 m) show a roughly 10% reduction over ocean. Walsh et al. (2007) recommend reduction of wind speed criteria for TCs to account for the effect of the coarse spatial resolution on storm structure. A 10% reduction is in the range recommended for a 50-km model used here. We have reduced the maximum surface wind speed criteria in the storm tracking algorithm by roughly 10% from 17 to 15.3 m s$^{-1}$ for tropical storms and from 33 to 29.5 m s$^{-1}$ for hurricanes/typhoons, both to be consistent with this recommendation and to make the result on TC and hurricane frequency be consistent with those in other papers in this series. None of our conclusions is sensitive to this correction.

3. Results

Figure 1a shows the variation of annual global number of TCs and hurricanes with respect to the cumulus mixing rate parameter $e_0$ from the control and the $e_0$ perturbation experiments. As $e_0$ increases from 6 to 10 (default), there is roughly a doubling of the global number of both TCs and hurricanes. However, the fractional increase of TCs (102%) is smaller than that of hurricanes (137%) so that the hurricane–TC ratio climbs steadily as $e_0$ increases. The rise of global number of TCs is larger as $e_0$ increases from 6 to 8 than from 8 to 10. Further increment of $e_0$ to 12 results in small changes. The additional advancement of $e_0$ to 14 leads to a sharp reduction of both TCs (−20%) and hurricanes (−16%). This nonmonotonic response of global TC/hurricane frequency with cumulus mixing rate suggests the existence of competing processes affecting the TC frequency, which may respond differently to $e_0$ and together determine the transition point from an increase to a reduction. Interestingly, the default value of $e_0$ in HIRAM, which was chosen primarily to tune the model’s global TOA radiative balance (Zhao et al. 2009), happens to be close to this transition point and generates the largest number of TCs. Figure 1a also shows that in contrast to the nonmonotonic behavior of total TC/hurricane frequency response, the ratio of hurricane to TC frequency increases roughly linearly with $e_0$, indicating a systematic intensification of storms as $e_0$ increases.

Figure 1b displays the variation of annual global number of TCs and hurricanes with respect to the divergence
damping parameter $d_0$. When the divergence damping is turned off by setting $d_0$ to zero, there are only about 65 TCs and 32 hurricanes per year globally. The application of the divergence damping leads to a systematic growth of both TCs and hurricanes with the default strength of damping ($d_0 = 0.16$) yielding roughly 40% more TCs and hurricanes. As $d_0$ increases progressively from 0 to 0.18, the rise of both TCs and hurricanes are approximately linear with the hurricane–TC ratio staying relatively unchanged. Therefore, there is a large impact of divergence damping on total number of storms with little effect on storm intensity distribution. The increase of TC frequency with the strength of divergence damping is also not intuitive since the divergence damping is designed to suppress grid-scale noise/disturbances, the existence of which might be thought of as conducive for TC genesis.

To confirm that the TC intensity responses to $e_0$ and $d_0$ are indeed as suggested by the hurricane–TC ratio in Fig. 1, we have computed the global mean TC intensity by averaging the maximum intensity of each storm (as measured by 10-m winds) over all TCs from the entire global tropical ocean. The mean intensity rises fairly linearly with $e_0$ with a roughly 10% increase as $e_0$ varies from 6 to 14. In contrast, the mean intensity exhibits little change ($\sim 1\%$) with $d_0$ for the range of values that we explored here. We emphasize that HIRAM at the current 50-km resolution cannot realistically simulate the full range of intensity distribution and therefore the response of the mean intensity should not be interpreted quantitatively at face value (Zhao et al. 2009; Zhao and Held 2010).

### 4. Relation to mean vertical motion

Zhao and Held (2012) explored six atmospheric parameters to understand the simulated hurricane frequency response to global sea surface temperature warming patterns. They include four commonly used TC genesis parameters (e.g., Camargo and Sobel 2005; Camargo et al. 2007a, b; Emanuel et al. 2008; Emanuel 2008, 2010), namely the potential intensity (PI), 600-hPa relative humidity ($RH_{600}$), the magnitude of vertical shear of vector wind between 200 and 850 hPa $S$, and 850-hPa absolute vorticity $\eta_{850}$. In addition, they also examined two less commonly used parameters: the vertical shear of zonal wind between 200 and 850 hPa $S_z$ and the 500-hPa vertical pressure velocity $\omega_{500}$. Zhao and Held (2012) found that while all parameters exhibit strong correlation to storm frequency response in some basins, $\omega_{500}$ is skillful in all basins. Held and Zhao (2011) demonstrate that $\omega_{500}$ can also be used to understand the global mean reductions of TC frequency in their idealized climate change experiments in which total greenhouse gas effects are broken down into the effect of increasing SST with fixed CO$_2$ and the effect of increasing CO$_2$ with fixed SSTs. Therefore, we focus on $\omega_{500}$ to help us understand the response of global TC frequency in these physics perturbation experiments. We have also examined the other five parameters and an additional parameter $x_m$ as defined in Eq. (3) in Emanuel et al. (2008), but none of them is able to explain the simulated global TC frequency responses. For example, for $x_m$, the changes in this parameter are small (at most $\pm 5\%$) and do not explain the nonmonotonic variation of $N$ with $e_0$.

Following Zhao and Held (2012), we first compute monthly TC genesis frequency over each $4 \times 5$ (latitude–longitude) grid box and obtain a climatological TC genesis function $G(x, y, m)$ ($x$ is longitude, $y$ is latitude,
and \( m = 1, 12 \) from the control simulation. We then weight local monthly mean fields \( \bar{\omega}_{500}(x, y, m) \) at each grid box \((x, y)\) by \( G(x, y, m) \) to obtain an annual mean—that is, 
\[
\overline{\omega_{500}}_{x,y} = \sum_{m} \left( \bar{\omega}_{500}(x, y, m)G(x, y, m) / \sum_{m} G(x, y, m) \right).
\]

The genesis weighting over the 12 months provides a simple and objective way of defining the index at the most relevant times of year. We then compute the aggregated index by a spatial average over all grid boxes where a TC genesis occurs in the control simulation. No genesis weighting is pursued for the spatial average. For both the control and the perturbation experiments, \( G(x, y, m) \) is held to be the same as the control simulation so that changing \( G(x, y, m) \) would not impact the genesis-weighted indices. We refer to this genesis weighted index as simply \( \overline{\omega_{500}} \) in the following.

The index \( \overline{\omega_{500}} \) is a measure of the midlevel atmospheric total convective mass flux with a negative value indicating a net upward mass flux. Figure 2 shows a scatterplot of the fractional changes of annual global TC count \( N \) versus fractional changes in \( \overline{\omega_{500}} \). The simulation produces a negative value of \( \overline{\omega_{500}} \) \((-26 \text{ hPa day}^{-1})\) since TC genesis occurs on average over regions of general convective activity. Note that Fig. 2, a positive \( \Delta \overline{\omega_{500}} \overline{\omega_{500}} \) corresponds to a negative \( \overline{\omega_{500}} \overline{\omega_{500}} \) and therefore an increase of upward convective mass flux.

For both perturbation experiments, the scatterplots in Fig. 2 exhibit a high correlation (>0.9) between the fractional change in \( N \) and that in \( \overline{\omega_{500}} \). However, the linear regression yields very different slopes. In the case of \( e_0 \)-perturbation experiments, the slope is roughly 1.4, indicating an increase of \( N \) with decreasing \( \overline{\omega_{500}} \) (or increasing convective mass flux) at a rate faster than linear but slower than quadratic. The suggested functional form is \( N = a_0 \overline{\omega_{500}}^{-b} \) with \( b = 1.4 \). In contrast, the linear regression for the \( d_0 \)-perturbation experiments leads to a much steeper slope (\( b \approx 6 \)). Since the variability in TC frequency may not only be regulated by but also contribute to the variability of \( \overline{\omega_{500}} \) through TC-induced mass flux, a possible explanation of the large difference in the regression slopes is that the variation of \( \overline{\omega_{500}} \) is primarily the cause for changes in \( N \) in the \( e_0 \)-perturbation experiments while the variation of \( \overline{\omega_{500}} \) in the \( d_0 \)-perturbation experiments is largely caused by the variation in \( N \), for which other mechanisms affected by \( d_0 \) must be responsible. As we will discuss below, other aspects of the responses in large-scale dynamic and thermodynamic environment support this explanation.

If we accept that the \( d_0 \)-perturbation experiments provide a rough estimate of the effect of TCs on \( \overline{\omega_{500}} \), then, globally, a 1% change of \( N \) would lead to roughly a 0.15% change in \( \overline{\omega_{500}} \), which is small but not negligible. This component of change in \( \overline{\omega_{500}} \) may be written as \( \delta \overline{\omega_{500}}^{\delta_0} = 0.15 \delta N / N \). Removing this effect in the \( e_0 \)-perturbation experiments \( [i.e., \delta N / N = 1.4 (\delta \overline{\omega_{500}}^{\delta_0} / \overline{\omega_{500}})] \) would bring the \( N \overline{\omega_{500}} \) relationship close to linear \( (\delta N / N \approx 1.15 \delta \overline{\omega_{500}}^{\delta_0} / \overline{\omega_{500}}) \).

We now provide additional evidence that \( e_0 \) can strongly alter the strength of tropical atmospheric convective overturning motion and therefore the TC genesis index \( \overline{\omega_{500}} \) whereas \( d_0 \) has little direct impact on \( \overline{\omega_{500}} \) except perhaps through changes in TCs themselves. Figures 3b and 3c show a comparison for the response of the probability density function (PDF) \( P_\omega \) of monthly mean 500-hPa vertical pressure velocity index \( \overline{\omega_{500}} \) to different experiments; the legend shows the correlation coefficient \( r \) and regression slope \( s \). The dotted (one-to-one) line shows the slope for a linear relationship.

Fig. 2. Scatterplots of the fractional changes (perturbation minus control) in annual global TC count \( \Delta N / N \) versus fractional changes in 500-hPa vertical pressure velocity index \( \Delta \overline{\omega_{500}} / \overline{\omega_{500}} \). Numbers 1, 2, 3, and 4 denote respectively the E6, E8, E12, and E14 experiments; 5, 6, 7, 8 denote respectively the D0, D6, D12, and D18 experiments. Dashed lines show linear regression for each set of the perturbation experiments; the legend shows the correlation coefficient \( r \) and regression slope \( s \). The dotted (one-to-one) line shows the slope for a linear relationship.
regions of upward motion if the total upward mass flux over the tropical ocean stays the same among the $e_0$-perturbation runs. It turns out the total upward mass flux also increases modestly as $e_0$ increases from 6 to 12 and stays roughly unchanged from $e_0 = 12$ to $e_0 = 14$ (Fig. 3e). This further contributes to the strengthening of the mean upward vertical motion with increasing $e_0$.

In general, the broadening of $P_v$ with increasing $e_0$ is due to a gradual transition of tropical deep convection from the (weak) parameterized to more intense resolved-scale convection. All other things being equal, the intensification of the tropical atmospheric convective overturning motion would suggest an enhancement of TC genesis frequency (Fig. 3a). This is indeed what the model simulated as $e_0$ increases from 6 up to 10. However, as $e_0$ increases from 10 to 14, the TC genesis index ceases to decrease and begins to increase (become less negative) even though the mean $\omega$ averaged over the ascent regions in the entire tropical ocean and for all months continues to decrease (become more negative). This indicates a more complicated (non-uniform) response of the seasonal and spatial distribution of $\omega$ to $e_0$. Moreover, as we will discuss below, there is evidence that other processes may act against the TC genesis in this model as $e_0$ increases.

In contrast to the $e_0$-perturbation experiments, Fig. 3c shows very little changes in $P_v$ in the $d_0$-perturbation experiments, indicating that the strength of the divergence damping hardly affects the seasonal-scale tropical
convective overturning motion. This leads to our hypothesis that the variation in TC genesis index in the $d_0$ perturbation experiments (Fig. 2) is most likely due to the variation of TC frequency itself.

Although $P_\omega$ and $\Delta P_\omega$ over the strong ascent regions ($\omega < -50 \text{ hPa day}^{-1}$) are relatively small compared to that in the weak ascent regions, they constitute a large fraction of the total upward mass flux and its changes in the perturbation experiments, since $\omega$ is large in these regions. To illustrate this, Figs. 3a–c except with $P_\omega$ and $\Delta P_\omega$ multiplied by $od_\omega$ ($od_\omega = 15 \text{ hPa day}^{-1}$, the bin size used) so as to display the distribution of the total upward mass flux and its changes within different $\omega$ regimes. The integration of $\omega P_\omega od_\omega$ curve over the negative $\omega$ regimes in Fig. 3d gives the total upward mass flux. Similarly, the changes in total upward mass flux from the perturbation experiments can be found by integrations over the negative $\omega$ regimes in Figs. 3e and 3f. Figure 3e shows that as $e_0$ increases, the increases of upward mass flux summed over the strong ascent regions are generally larger than the reductions summed over the weak ascent regions despite the fact that the augmentation of strong ascent area is much smaller than the diminishment of the weak ascent area (Fig. 3b). Therefore, the shift of moist convection from weak/parameterized to more intense/resolved-scale regimes also strengthens the overall tropical upward motion. In contrast, Fig. 3f shows again that there are little differences among the $d_0$-perturbation experiments.

The shift of deep moist convection from the parameterized to the resolved-scale regime in the $e_0$-perturbation experiments can also be clearly seen from the response in the fraction of large-scale precipitation $f_0$, which is defined as the fraction of total tropical ($30^\circ$S–$30^\circ$N) precipitation resulted from the large-scale cloud module instead of the parameterized convection module. Figure 4 shows that $f_0$ increases rather linearly with $e_0$ from 15% in E6 to 60% in E14. Nearly all of the increase in large-scale precipitation compensates the reduction in the precipitation from the parameterized convection so that the total tropical precipitation increases only modestly ($<3\%$). In contrast, the alteration of $d_0$ has little impact on $f_0$. The variation of total tropical precipitation among the $d_0$-perturbation experiment is very small ($<0.4\%$). Since the $d_0$-perturbation experiments produce large variations in global TC frequency and the TC-associated precipitation comes primarily from the large-scale cloud module in this model, the small response of $f_0$ to $d_0$ also indicates that TC contribution to global tropical precipitation is small in this model. This is not inconsistent with recent estimates based on satellite observations (e.g., Jiang and Zipser 2010).

To summarize, Figs. 2–4 indicate that the response of global TC count in the $e_0$-perturbation experiments can be understood through changes in seasonal-scale convective mass flux as measured by the $\overline{\omega_{500}}$ index, whose variation is primarily driven by changes in the strength of convection. However, the correlation between TC frequency response and the variation of $\overline{\omega_{500}}$ in the $d_0$-perturbation experiments is most likely due to the effect of TCs on $\overline{\omega_{500}}$. Two remaining questions are the following: What causes the rise of global TC frequency with the increase of $d_0$? Why does the global TC count drop as $e_0$ increases further beyond some large values? Next, we explore another aspect of the simulation—the noisiness, or lack of spatial and temporal coherence of the convection—which we believe may be the key to answering these questions.

5. Small-scale noise

For this analysis we used 4 times daily high-frequency output of instantaneous 850-hPa vorticity, also used for TC detection and tracking. In retrospect, it might have been useful to also save the high-frequency $\omega$ and/or horizontal divergence for this analysis. But we do not have this available from these simulations. We first conducted a 1D Fourier transform for each longitudinal distribution of 850-hPa vorticity and then computed the variance spectra averaged over all tropical ($30^\circ$S–$30^\circ$N) latitude bands from the 20-yr model output. Figure 5a displays the variance spectra from the control and the $d_0$-perturbation experiments. The variance spectra roughly follow a $-1$ power law for wavenumbers between 10 and 100 (wavelength 4000–400 km), consistent

FIG. 4. Variation in the fraction of large-scale precipitation $f_0$ in the tropics with respect to (bottom x axis) the cumulus mixing parameter and (top x axis) the divergence damping parameter. The fraction of large-scale precipitation is defined as the fraction of total tropical precipitation that comes from the large-scale cloud module instead of the parameterized convection module.
with the variance being dominated by 2D geostrophic turbulence over these spatial scales. At higher wavenumbers the slopes become steeper and are clearly impacted by $d_0$.

To display more clearly the sensitivity of the variance spectra to $d_0$, Fig. 5b shows fractional changes in variance at each wavenumber from the $d_0$-perturbation and the control experiments. As $d_0$ increases there are systematic reductions of variance for wavenumbers greater than 60. This suppression of small-scale noise is expected from the formulation of the divergence damping. However, the diminishment of the smaller-scale noise also appears to induce an enhancement of the variance at larger scales (wavenumbers smaller than 60). Since the variance spectrum declines rapidly with wavenumber in the control simulation (Fig. 5a), the absolute increase in variance at larger scales is significant. The result suggests a picture in which the suppression of small-scale noises encourages the development of disturbances at larger scales, which may provide a more coherent environment for TC genesis. Explanation for the suggested scale interaction may be that small-scale random convection tends to compete for moisture and energy, resulting in a difficulty for larger-scale disturbances to emerge and grow. According to this speculation, the suppression of small-scale weak disturbances is helpful for the organization of convection, allowing moisture and energy to be effectively fed into larger-scale disturbances for TC genesis.

The alteration of $e_0$ also affects the noisiness of convection. In particular, as $e_0$ increases, there is systematic growth in the spectral power with roughly uniform fractional changes across all wavenumbers. Figure 6 depicts the variation of global TC count from the $e_0$- and $d_0$-perturbation experiments in a two-dimensional space measuring both the strength and the noisiness of convection. The strength of convection is estimated by $f_l$ as shown in Fig. 4. The noisiness of convection $n_c$ is estimated by the spectral power averaged for wavenumbers greater than 60 (see Fig. 5). Both $f_l$ and $n_c$ are expressed in fractional changes from the control simulation. Contours of constant TC frequency at 80 and 100 yr$^{-1}$ are manually added.

**Fig. 5.** (a) Variance spectra of 850-hPa vorticity from the control and the $d_0$-perturbation experiments. Variance spectra are computed using fast Fourier transform (FFT) for each latitude using 6-hourly model output. The spectra are then averaged over all tropical (30$^\circ$S–30$^\circ$N) latitude bands for the 20-yr simulation period. (b) As in (a), but for fractional changes in variance at each wavenumber.
that TC frequency can be thought of as a function of $f_1$ and $n_c$, contours of constant TC frequency at 80 and 100 yr$^{-1}$ are added manually in Fig. 6.

As $e_0$ increases incrementally, both $f_1$ and $n_c$ rise and tend to act against each other in the global TC genesis frequency. At low values of $f_1$, the isolines for TC frequency tend to be vertical (parallel to lines of constant $n_c$), indicating a dominant effect of $f_1$ on the global TC count with little impact from $n_c$. However, at high values of $f_1$, the isolines curve horizontally (parallel to lines of constant $f_1$), suggesting that $n_c$ takes control of the TC genesis frequency when $f_1$ is sufficiently large. As $e_0$ increases, one can imagine that the line connecting the $e_0$ simulations travels across the isolines twice, resulting in a nonmonotonic response of the global TC count to $e_0$. In contrast, the $d_0$ perturbation experiments produce a simple monotonic decrease of global TC count with increasing $n_c$ since the variation in $f_1$ is small among this set of experiments.

6. Mean tropical profiles

Finally, Fig. 7 compares the response in vertical profiles of the parameterized convective mass flux, temperature, and humidity averaged over all tropical ocean areas.

Fig. 7. Differences (perturbation minus control) in vertical profiles of (a) the parameterized convective mass flux $\Delta M_c$, (b) temperature $\Delta T$, and (c) relative humidity ($\Delta RH$) between each of the $e_0$-perturbation experiments and the control experiment. (d)–(f) As in (a)–(c), respectively, but for the $d_0$-perturbation experiments.
areas for all seasons between the $e_0$ and $d_0$-perturbation experiments. Figure 7a shows that as $e_0$ decreases from its default value 10 to 8 and 6, there are systematic increases (decreases) of the parameterized convective mass flux above (below) 700 hPa. Since the parameterized deep and shallow plumes can strongly interact each other over time through impact on the boundary layer and the lower-tropospheric thermodynamic properties, the change of entrainment rate parameter influences populations of both deep and shallow plumes and therefore the parameterized mass flux at all levels. The changes in parameterized convective mass flux in Fig. 7a must then be largely balanced by changes in the resolved-scale (explicit) convection since the total tropical convective mass flux is strongly constrained by net tropospheric radiative cooling, which stays relatively unchanged among these prescribed SST experiments.

The response of resolved-scale convection tends to be slave to and to compensate changes in the parameterized convection. For example, as $e_0$ increases, the reduction of parameterized deep convection would initially lead to much colder free troposphere and moister boundary layer, which promotes resolved-scale convection. The resolved-scale convection heats the free troposphere through condensation of large-scale water vapor. This compensation means that the equilibrium atmosphere does not cool as much as directly suggested by the changes in parameterized deep convection. This negative feedback makes the net response of temperature and humidity profiles more complicated and may be sensitive to the parameterization of large-scale cloudiness, microphysics, and convection–cloud–radiation interactions.

Despite these complications, some of the response of temperature and humidity profiles may still be qualitatively understood by changes in the parameterized convection. The increase of parameterized deep convective mass flux generally leads to a warmer free troposphere over most of the convective depth (below $\sim$250 hPa). Near the top of the convection, the increase in the parameterized deep cumulus clouds generates more penetrative mixing (Bretherton et al. 2004) and entrainment of cloud-top air. Further evaporation of cloud condensate and associated downdrafts lead to a cooling effect near the top (above $\sim$250 hPa; see Fig. 7b). Figure 7c shows that the increase of parameterized deep convection generally gives rise to an atmosphere that is moister near the top (through additional cumulus detrainment) and drier in the lower part of the convective depth. Figures 7a–c also demonstrate that the opposite is true as $e_0$ increases from its default value 10 to 12 and 14. Overall, the response of the tropical mean sounding to an increase of $e_0$ can be generally characterized as a systematic increase of atmospheric convective instability.

In contrast to the $e_0$-perturbation experiments, Figs. 7d–f display very small responses for each of the fields from the $d_0$-perturbation experiments. The only significant response may be in the temperature field for the $d_0 = 0$ experiment, which produces a warming of the troposphere by roughly 0.1–0.3 K. Careful examination of Fig. 7e reveals a small but systematic cooling effect with the strength of the divergence damping that tends to quickly saturate as $d_0$ increases beyond 0.06. This cooling effect may be due to the suppression of some disorganized convection, allowing additional large-scale precipitation and heating, as may be seen in Fig. 4. In general, the fact that there are very small changes in the tropical atmospheric thermodynamic properties in the $d_0$-perturbation experiments provides further evidence that the small variations in $\sigma_{500}$ (Fig. 2) among the $d_0$-perturbation experiments is most likely a result of the variation in TCs.

7. Discussion

At current resolutions (20–100 km), the uncertainties in the details of a model’s formulations of physics and dynamics may be an important source of uncertainties in simulated TC statistics and possibly their future changes. We have explored the sensitivity of global number of TCs in the context of present-day simulations using the GFDL HIRAM with altered physics parameterizations related to moist convection. We found two parameters/processes that are important in controlling the global TC count in this model. They are the cumulus mixing rate and the strength of damping in the divergent component of the horizontal flow. The simulated global number of TCs exhibits nonintuitive response to incremental changes of both parameters.

As the cumulus mixing rate parameter $e_0$ increases, the model produces an initial sharp rise in global number of TCs/hurricanes, but then reaches a maximum and decreases quickly as $e_0$ attains larger values. The response of the global TC count is well correlated with an atmospheric TC genesis index $\sigma_{500}$, which measures the seasonal-scale midtropospheric total convective mass flux averaged over all TC development regions. While the parameterized convective mass flux is directly controlled by $e_0$, the response of total convective mass flux to $e_0$ is complicated and involves intricate feedbacks from resolved-scale convection, cloudiness and convection–cloud–radiation interactions. These make it difficult to quantify the response of an aggregated index $\sigma_{500}$ over the TC development regions and seasons. Nevertheless, the analysis suggests that the following processes may qualitatively explain the simulated TC response in this model.

As $e_0$ increases, the enhanced cumulus mixing and entrainment reduces in-cloud buoyancy, resulting in...
decline of parameterized deep convective mass flux. This generally cools and dries much of the upper troposphere and moistens the boundary layer and lower troposphere, leading to growth of resolved-scale convection. Compared to the parameterized convection, the resolved-scale convection provide more intense latent heat release and vertical velocities, which tend to enhance surface heat fluxes, making the convection more favorable for cyclogenesis. The gradual shift of tropical convection from parameterized to resolved-scale convection leads to an intensification of tropical convective overturning motion, which can be measured by examination of the 500-hPa vertical velocity PDF. All other things being equal, this shift in the strength of convection would lead to an increase of global TC frequency with $e_0$, which is what the model simulated as $e_0$ increases up to 10. However, the flattening and unexpected drop of the global TC count as $e_0$ advances from 10 to 12 and 14 suggests that other processes may set in to prevent TC genesis in this model.

One explanation for the nonintuitive response of TC frequency at large $e_0$ is guided by the analysis for the $d_0$-perturbation suite, which perturbs the strength of the divergence damping. As the divergence damping $d_0$ increases, the model produces a monotonic increase of global TC count, which is also not intuitive since the damped small-scale noises/disturbances might be thought of as favorable for TC genesis. A detailed examination of the seasonal and large-scale dynamic and thermodynamic properties of the tropical atmosphere reveals little difference among the $d_0$-perturbation experiments. Therefore, the decrease in the noisiness of convection appears to be the only possible cause of the simulated differences in TC frequency for the $d_0$-perturbation experiments. The explanation may be that the excessive amount of disorganized small-scale convection tends to compete for moisture and energy, making it hard for coherent large-scale disturbances to develop. The effect of the noisiness of convection on TC genesis can also be used to understand the drop of TC frequency at large $e_0$ in the $e_0$-perturbation experiments, since the rise of $e_0$ not only enhances the strength of convection but also increases the noisiness of convection. These effects tend to counteract each other in their impact on TC genesis, as shown in Fig. 6.

The results in Fig. 6 also suggest that the dependence of the global TC frequency on the strength (measured by the amount of resolved-scale convection) and the noisiness of convection is nonlinear and regime dependent. At small $e_0$, the parameterized convection dominates tropical convective activity and convection is generally weaker and more persistent (less intermittent). In this kind of regime, TC frequency is highly sensitive to the strength of convection (i.e., the amount of resolved-scale convection) and not sensitive to the noisiness of convection. At larger $e_0$, resolved-scale convection starts to dominate tropical convective transport. Resolved-scale convection generally has much larger vertical velocity and a much stronger suppression effect to surrounding atmosphere through induced transient warming and drying. Excessive amounts of the resolved-scale convection make it difficult to organize, leading to an overall noisiness in the flow. In this kind of regime, the noisiness of the convection takes control of the TC genesis frequency and produces a detrimental effect to global TC genesis despite the continuing rise in the strength of convection. Therefore, the shift from parameterized to resolved-scale convection with increasing $e_0$ is also accompanied by a gradual transition of TC genesis from a strength-controlled regime to a noisiness-controlled regime.

The flow noisiness introduced by resolved-scale convection may be an issue in a high-resolution hydrostatic model whose parameterized moist convection is so inhibited that the total tropical convective transport must depend significantly on resolved-scale convection. The problem may be alleviated by the application of the divergence damping or equivalent. In this regard, the divergence damping may be considered part of convection parameterization, which can impact convection organization in this kind of model.

Our explanations for the parameter dependence in this model’s TCs remain tentative. Parameter studies with other models of various resolutions will be needed to better test our understanding of the statistics of TC genesis in global models.

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