A Numerical Study of the Global Formation of Tropical Cyclones

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Abstract This study examines the large-scale factors that govern global tropical cyclone (TC) formation and an upper bound on the annual number of TCs. Using idealized simulations for an aquaplanet tropical channel, it is shown that the tropical atmosphere has a maximum capacity in generating TCs, even under ideal environmental conditions. Regardless of how favorable the tropical environment is, the total number of TCs generated in the tropical channel possesses a consistent cap across experiments. Analyses of daily TC genesis events reveal further that global TC formation is intermittent throughout the year in a series of episodes at a roughly 2-week frequency, with a cap of 8–10 genesis events per day. Examination of different large-scale environmental factors shows that 600-hPa moisture content, 850-hPa absolute vorticity, and vertical wind shear are the most critical factors for this global episodic TC formation. Specifically, both the 850-hPa absolute vorticity and the 600-hPa moisture are relatively higher at the onset of TC formation episodes. Once TCs form and move to poleward, the total moisture content and the absolute vorticity in the main genesis region subside, thus reducing large-scale instability and producing an unfavorable environment for TCs to form. It takes ~2 weeks for the tropical atmosphere to remoisten and rebuild the large-scale instability associated with the Intertropical Convergence Zone before a new TC formation episode can occur. These results offer new insight into the processes that control the upper bound on the global number of TCs in the range of 80–100 annually.

Plain Language Summary Understanding the mechanisms behind the upper bound on the global tropical cyclone (TC) number is important yet still elusive. Using TC simulations on an aquaplanet tropical channel, this study shows that the tropical atmosphere possesses a maximum capacity in producing TCs globally, regardless of how ideal the environmental conditions are. In particular, global TC formation does not occur throughout the year but in a series of episodes with a frequency of about 2 weeks, with ~8–10 TCs per episode. Examination of the TC genesis index shows that midlevel moisture content and 850-hPa absolute vorticity associated with the ITCZ instability are key factors governing the episodic global TC formation.

1. Introduction

The long-term variability of global tropical cyclone (TC) frequency is an important yet open question. Historical records of TC activities show that the total annual number of TCs over all ocean basins varies in the range of 80–100 TCs (Frank & Young, 2007; Gray, 1968, 1975; Lander & Guard, 1998; Ramsay, 2017), with no definite trend either globally or in each hemisphere (Figure 1a). While being long documented, there is currently no extant theory that explains why the Earth’s tropical atmosphere can produce a specific range of 80–100 TCs instead of an arbitrary number of TCs annually. Using historical data, Hoogewind et al. (2020) recently sought to estimate the maximum potential genesis and found that the atmosphere could plausibly fit many more storms each year than actually occur. However, the total number of the TCs is still bounded in their estimation, even under the most favorable conditions.

From the energetic standpoint, the existence of such an upper bound on the number of TCs is consistent with the fact that only a certain amount of energy from the tropical region can be transported to higher latitudes to maintain the energy balance and stability of the Earth’s climate system. Due to the specific distribution of solar radiation on the Earth’s surface, too many or too few TCs would result in an unstable configuration of the large-scale circulation that would not be otherwise maintained. Thus, it is anticipated that the global
annual TC number must be upper-bounded. Despite such reasonable expectation of the global TC number from the energetic argument, no complete dynamical explanation or current theories could explain what sets the global annual TC count. In fact, many observational and modeling studies have provided different insight into causes of variability in this total TC number, yet they have not directly addressed the number itself nor its upper bound (e.g., Bengtsson et al., 1982; Caron et al., 2011; Emanuel & Nolan, 2004; Manganello et al., 2012; Murakami & Sugi, 2010; Oouchi et al., 2006; Ramsay, 2017; Zhao et al., 2009; Walsh et al., 2015).

Numerous studies of the environmental parameters known to be important for TC activity have found a list of large-scale conditions required for TC formation (e.g., Bister & Emanuel, 1997; Gray, 1968; Molinari et al., 2000; Nolan et al., 2007; Ritchie & Holland, 1997; Riehl & Malkus, 1958; Simpson et al., 1997; Walsh et al., 2020; Yanai, 1964; Zhang & Bao, 1996). Several common conditions include (i) relatively warm sea surface temperature (SST); (ii) a preexisting cyclonic disturbance at the lower troposphere; (iii) weak vertical wind shear; (iv) a moist lower-to-middle troposphere; or (v) existence of a tropical upper tropospheric trough. Under these favorable conditions, a TC may emerge and subsequently move to higher latitude under the influence of beta-drift and southeasterly steering flows on the southwestern edge of subtropical highs. For practical applications, these environmental factors are often combined into different indices such as a Genesis Potential Index (GPI; see, e.g., Nolan et al., 2007; Camargo & Wing, 2016) or a ventilation index (e.g., Tang & Emanuel, 2012), to quantify the environmental favorability for TC genesis and understand how variability in these environmental conditions causes changes in TC genesis.

The above large-scale environmental conditions are however only a part of what governs variability in TC count, because the climatology of TC formation is also known to strongly depend on the characteristics of each ocean basin. In fact, various field experiments and modeling studies showed that the formation of TCs varies widely among basins and inherits strong regional characteristics, such that no single set of criteria can be applied. For example, TC formation in the North Atlantic basin often shows a strong connection to active tropical waves associated with African Jet (Avila & Pasch, 1992; DeMaria, 1996; Dieng et al., 2017;
Landsea, 1993; Molinari et al., 2000; Pasch et al., 1998). In the Northwestern Pacific basin, previous studies showed that tropical cyclogenesis is closely tied to variability in Intertropical Convergence Zone (ITCZ) and monsoon activities (Gray, 1968, 1982; Ferreira & Schubert, 1997; Harr et al., 1996; Mark & J., 1993; Ritchie & Holland, 1997; Yanai, 1964). For the Northeastern Pacific basin, the multiscale interaction of terrain and tropical waves seems to be the main mechanism in generating tropical disturbances that could potentially evolve into TCs (Halverson et al., 2007; Kieu & Zhang, 2008, 2009a, 2009b; Molinari et al., 1997; Wang & Magnusdottir, 2006; Zehnder et al., 1999). These studies indicate that the large-scale dynamics of how precursor disturbances form is important for understanding the global TC count, as also suggested from the recent analyses by Hoogewind et al. (2020).

To eliminate the regional asymmetric forcings related to the land-surface or other orographic effects, global TC simulations on an aquaplanet have been extensively conducted to study the characteristics of the global TC distribution (e.g., Chavas & Reed, 2019; Chavas et al., 2017; Merlis et al., 2013; Rauscher et al., 2013; Reed & Jablonowski, 2011; Walsh et al., 2020). For example, Merlis et al. (2013) demonstrated that the TC frequency would increase due to radiatively forced warming associated with the poleward shifting of the ITCZ. In the absence of variability in thermodynamic forcing, Chavas and Reed (2019) demonstrated a dynamical dependence of the latitudinal distribution of genesis on the Coriolis parameter, as well as the minimum distance of genesis from the equator. Likewise, Fedorov et al. (2018) demonstrated that TC count decreases rapidly with increasing meridional sea surface temperature gradient using planetary-scale beta-plane experiments. In contrast, the aquaplanet simulations by Li et al. (2013) showed that specific intraseasonal variability might not have significant impacts on global TC formation. Instead, TCs may form spontaneously owing to the internal physics and dynamics in an idealized aquaplanet environment. This behavior appears to be consistent with a recent study by Patricola et al. (2018), which found that filtering out African Easterly waves does not substantially alter the number of storms in the Atlantic basin.

To better understand the dynamical mechanisms underlying global TC formation, (Wang et al., 2019, hereinafter W19) recently presented a theoretical study for the large-scale dynamics of TC formation, based on the ITCZ breakdown model. Using dynamical transition theory, W19 found that the number of tropical disturbances that could potentially grow into TCs on any given day possesses an upper bound, independent of thermodynamic favorability. Specifically, the maximum number of daily TC disturbances must be less than ~12 in the current Earth's climate due to the inherent dynamics of the tropical atmosphere. Such an upper bound on the number of TCs that the tropical atmosphere can produce is consistent with previous modeling studies of the ITCZ breakdown in Ferreira and Schubert (1997) and observations. For example, Figure 1b displays a satellite snapshot of a rare event of nine simultaneous storms in the tropical belt in September 2017. In this regard, W19's result aligns with recent observational analyses in the maximum number of TCs by Hoogewind et al. (2020) and the linkage between storm count and precursor disturbance count in current and future climate states in Vecchi et al. (2019).

The finding of W19 that the tropical atmosphere has a limited capacity to produce TC precursors offers a key new building block toward understanding controls on global TC count. What remains missing, though, is how frequently TCs emerge in the atmosphere. Consider, for example, a situation in which 10 TCs could appear every single day over an entire year. This implies that the total annual number of TCs that the Earth can produce is equal to ~3,500 storms, which is far too high. However, a scenario in which 10 TCs emerge every 15 days would cut down the total number of TCs to ~250 storms per year, which seems to be much more reasonable for the Earth-like aquaplanet atmosphere. In this regard, knowing the frequency of TC formation is required to explain the current bound of ~80–100 TCs as observed annually in the real atmosphere.

In this study, we aim to address how frequently global TC formation can occur in the tropical atmosphere using idealized simulations. We further aim to evaluate which atmospheric large-scale factors are most important in setting this frequency of global TC formation. Answering these questions, together with the upper bound on the number of TCs that the tropical atmosphere is likely to produce at a given time as proposed in W19, will provide important insight into the fundamental question of why the current Earth's atmosphere can support only 80–100 TCs annually.

Our paper is constructed as follows. Section 2 provides the details of experiment design and the TC tracking algorithm used in this study. The results from various experiments and related analyses are presented in section 3, and the summary of key findings is provided in section 4.
2. Experiment Description

2.1. Model Description
To understand the frequency of global TC formation beyond the analytical ITCZ breakdown model, idealized simulations with a full-physics model on an aquaplanet are conducted in this study. The Weather Research and Forecasting (WRF) model (Version 3.9 Skamarock et al., 2008) is employed in an aquaplanet channel between 45° S to 45° N. All experiments are set up with a single domain at a homogeneous resolution of 27 km, with 31 vertical levels and model top at 18 km. The model domain consists of 1,479 × 400 grid points in the west-east and the north-south directions, thus covering the entire symmetric tropical and midlatitude regions from 45° S to 45° N. For our TC formation study, this domain suffices to capture all stages of TC development including early formation, poleward movement as well as the final dissipation stage at higher latitudes over colder SST, thus allowing for detailed examination of TC formation mechanisms at the global scale in an idealized, zonally symmetric setting.

For the model physical representations, all experiments are conducted with a common suite of physical parameterizations including (i) a modified version of the Kain and Fritsch (1990) cumulus parameterization scheme in which deep convection and a broad range of shallow convection are both parameterized; (ii) the Yonsei University planetary boundary layer (PBL) parameterization with the Monin-Obukhov surface layer scheme; (iii) the Rapid Radiative Transfer Model (RRTM) scheme for both longwave and shortwave radiation that include the full seasonal cycle of shortwave radiation (Mlawer et al., 1997); and (iv) the Lin et al. (1983) cloud microphysics scheme. Along with these physical parameterization schemes, the Eulerian dynamical core based on the Runge-Kutta finite difference scheme at a time step of 30 s is employed for all simulations. Given the specific domain configuration that covers the entire tropical belt, the model boundary condition is set to be periodic in the zonal direction and open in the meridional direction. Note that the choice of open north-south boundary condition is because there exists no periodicity or prescribed boundary input for the idealized aquaplanet setting. Moreover, the asymmetry in solar incoming radiation between Northern and Southern Hemispheric summers prevents the complete symmetry between the north and south lateral boundaries. Thus, the open boundary is most physically appropriate for our simulations in this study.

To allow for the interaction of the tropical atmosphere with the ocean surface, a one-dimensional mixed layer ocean coupling scheme is used. This ocean coupling option, which is available in the WRF model (V3.9), is based on the study by Pollard et al. (1972) (hereafter referred to as the OMLM-Pollard scheme) that allows for surface flux exchange without ocean dynamics. Physically, this OMLM-Pollard scheme is a bulk mixed-layer model that simulates the ocean mixed layer in an integral sense. This model requires a given initial mixed-layer depth and a deep-layer lapse rate in the thermocline at each model grid point (Wang & Duan, 2012). Given its simplicity and efficiency in long-term climate modeling (e.g., Davis et al., 2008; Kim & Hong, 2010; Wang & Duan, 2012), this scheme is adopted herein for our global TC formation simulations.

It should be noted that the choice of a fixed homogeneous 27-km resolution in this study is a compromise between the computational requirement, the length of simulations, the efficiency of diagnostic analyses, and our current available data storage. Previous modeling studies of tropical cyclogenesis showed that this horizontal resolution suffices to produce model vortices up to at least Category 3 (e.g., Bengtsson et al., 1982; Broccoli & Manabe, 1990; Chavas et al., 2017; Vecchial et al., 2019; Wehner et al., 2010; Zhao et al., 2009). In principle, a higher resolution is preferred to obtain better simulations of TC inner core structures (i.e., the grid spacing should be finer than 10 km to have full eye dynamics; see, e.g., Wehner et al., 2010; Zhao et al., 2009). Due to the large amount of the model output at a homogeneous solution and the long 2-year model integration with multiple sensitivity experiments, this 27-km resolution is therefore adopted here with a caveat that the fine inner-core structure of mature TCs may not be fully resolved. Such details are also not the focus of our work as in recent works using global climate model aquaplanet simulations at similar resolution of ~25 km (Chavas & Reed, 2019; Chavas et al., 2017; Fedorov et al., 2018).

2.2. Experiment Design
To set up a reference for global TC formation, a control (CTL) experiment with a 5-year simulation is first carried out, starting from a tropical sounding (Jordan, 1958). The first year of the model simulation is considered as a spin-up integration to allow WRF to develop a tropical atmosphere consistent with the given boundary
Figure 2. (a) The latitudinal profile of SST used to initialize the WRF model and (b) the horizontal distribution of geopotential perturbation (shaded, unit m² s⁻²) and wind perturbations associated with tropical Kelvin waves at z = 1 km that are added from the second year of the CTL experiment at a prescribed interval. Solid/dotted contours denote the relative vorticity with contour interval 10⁻⁶ s⁻¹.

Figure 2a: The latitudinal profile of SST used to initialize the WRF model.

Figure 2b: The horizontal distribution of geopotential perturbation (shaded, unit m² s⁻²) and wind perturbations associated with tropical Kelvin waves at z = 1 km. Solid/dotted contours denote the relative vorticity with contour interval 10⁻⁶ s⁻¹.

conditions and model physics. For this CTL experiment, the initial surface boundary is constrained by an ocean surface with initial SST distribution given by the following the zonally averaged profile:

\[ SST = T_0 e^{-\left(\frac{y - y_0}{R}\right)^2}, \]

where \( R = 25^\circ \) is the half width, \( y_0 \) is the latitude of symmetry that is set equal to 0 at the equator, and \( T_0 = 30.5^\circ \) C is the maximum SST at \( y_0 \). This profile has maximum SST at the equator and quickly decreases with latitudes as seen Figure 2a. While there are various idealized SST profiles proposed in previous studies (e.g., Neale & Hoskins, 2000), all these profiles share an important property that SST deceases at higher latitudes. This property is important to drive the tropical atmosphere toward radiative-convective equilibrium as discussed in Neale and Hoskins (2000), with an easterly flow and a low-pressure belt in the equatorial region and a higher-pressure belt with westerly flows in the extratropical region.

With the first year of model integration used as a spin-up, we implemented an external forcing starting from the second year of model integration that constantly excites the tropical region to produce a large number of disturbances and promote TC genesis. Unlike the realistic Earth’s surface for which the asymmetric distribution of landmass and topography is a source of atmospheric disturbances, the aquaplanet setting lacks such a source of perturbations related to the topography-atmosphere interaction. Given our purpose of examining the maximum number of TCs that the tropical atmosphere can produce, an external stirring mechanism is, therefore, needed to create the most favorable conditions for TC formation.

An approach to effectively stir the tropical region for this purpose is to add equatorial waves to the model state at a given interval. Among several dominant tropical waves in the equatorial regions that could affect TC formation (e.g., Frank & Roundy, 2006; Kiladis et al., 2009; Schreck, 2015; Schreck & Molinari, 2011; Schreck et al., 2012; Wheeler, 2002; Wheeler et al., 2000), tropical Kelvin waves are one of the disturbance modes that help promote TC genesis. For example, semi-Lagrangian analyses by Ventrice and Thorncroft (2013) or Schreck (2015) suggested that Kelvin waves can interact with other systems and amplify preexisting precursors that later develop into TCs. An important characteristic of tropical Kelvin waves is their persistence in providing a favorable vorticity-rich environment for TC embryos to develop (Figure 2b), which could account for TC-related disturbances in the tropical region.

Given the potential roles of Kelvin waves in TC formation, we designed an external source of perturbations based on the tropical Kelvin wave mode as follows (e.g., Holton, 2004):

\[ u(x, y) = U_0 e^{-\frac{\gamma^2}{R^2}}, \]

\[ v(x, y) = 0, \]

\[ \phi(x, y) = c U_0 e^{-\frac{\gamma^2}{R^2}}, \]

where \( \beta \) represents the variation of the Coriolis parameter in the meridional direction, \( \gamma \) is the scaled distance from the equator, \( c \) is the wave propagation speed, and \( U_0 \) denotes the wave amplitude. To ensure
the balanced requirement among these perturbations, the temperature perturbation is directly derived from the geopotential perturbation given by Equation 3, while the moisture is left unchanged. We stress that this external source of Kelvin perturbations is independent of any internal tropical wave modes that exist in the model. In this regard, our external Kelvin wave source is merely to examine if constantly stirring the tropical region can help enhance tropical cyclogenesis, rather quantifying the relative contributions of different tropical wave modes to TC formation as discussed in the previous studies.

By default, these Kelvin-wave perturbations have a wavelength of about 3,800 km (corresponding to zonal Wave Number 11) with a vertical structure given by a normal mode function (Wheeler, 2002). Such a vertical structure provides a simple relationship between the vertical wavelength of a normal mode in a constant stratification atmosphere and its equivalent depth, and it is applied simultaneously to both the wind and mass fields. Note that the use of specific zonal Wave Number 11 for the tropical Kelvin wave perturbations in this study follows from previous studies by Ferreira and Schubert (1997), and W19, which showed that the most unstable mode in the tropical region is upper bounded by zonal Wave Number 12 (i.e., all possible unstable modes must have wave number $m < 12$ as derived from W19’s results).

In all experiments, the Kelvin wave amplitude $U_0$ is set equal to 1 m s$^{-1}$ with a phase speed $c = 10$ m s$^{-1}$. These perturbations are added every 10 days over the entire domain for the CTL experiment (Figure 2b). We note that this 10-day interval is not arbitrary, but it is roughly the time required for the Kelvin waves to travel one wave length, given the wave speed in the range of 5–10 m s$^{-1}$ and wave length of $\sim 3,800$ km as designed. Moreover, the amplitude of the Kelvin wave perturbation has to be sufficiently small such that the addition of these perturbations on the model state would not affect the model numerical stability. As such, the wave amplitude of 1 m s$^{-1}$ is kept fixed in all experiments in this study.

With the above CTL configuration, a number of sensitivity experiments are then conducted to examine the robustness of the CTL experiment in simulating global TC formation. In the first sensitivity experiment (K5), the tropical Kelvin waves are added at an interval of 5 days instead of 10 days as in the CTL experiment. Such a shorter time window of adding Kelvin wave perturbations is to examine how TC formation changes with different triggering frequencies as compared to the CTL experiment. To further isolate the effect of the Kelvin wave excitation, another experiment (NOK) is conducted in which no Kelvin wave perturbations are added during the entire simulation. The relative difference among the CTL, K5, and NOK experiments will reveal how critical the tropical perturbations are in driving global TC formation.

In the second set of sensitivity experiments, the model resolution is reduced from 27 km in the CTL experiment to 54 km (E54) and 81 km (E81) to test the robustness of the CTL conclusions to horizontal resolution. Due to the large requirement of computational resource and storage for homogeneous idealized experiments, only two sensitivity experiments with coarser resolutions of 54 and 81 km are presented herein. For a coarse resolution of $> 100$ km, TCs are very poorly resolved, and genesis is rare, while much higher resolution ($< 10$ km) is not feasible with our current computational resources. Except for these changes in the model resolution, the same procedures and model physics in all sensitivity experiments are applied as in the CTL experiment.

In the final sets of sensitivity experiments, we test a range of ocean mixed layer depths and relaxation timescales below as part of our analysis. Specifically, the ocean mixed layer (OML) depth is varied to 10 m (M10) and 30 m (M30) with the relaxation time kept fixed at $\tau = 30$ days. Likewise, two additional experiments are also conducted with a fixed OML $= 1$ m, but with the relaxation time is equal to 5 days (T5) and 50 days (T50). Here, the ocean relaxation time is defined as the timescale required for the ocean model temperature and OML depth to relax back to their original values. These sensitivity experiments are needed, because the oceanic responses to the seasonal variability depend strongly on ocean mixed layer depth and the relaxation timescale. That is, a thicker OML would result in a smaller seasonal change. Similarly, a longer relaxation time would allow the ocean to exhibit stronger seasonal and diurnal variations due to external solar and atmospheric forcing instead of quickly adjust back to the mean state. Given the sensitivity of TC formation on surface temperature conditions, it is expected that the global climatology of TC formation should display some dependence on the OML and ocean relaxation time. Detailed descriptions of all experiment configurations are given in Table 1. Unlike the CTL experiment that was integrated for up 5 years to obtain sufficiently long data for more in-depth spectral analyses, all sensitivity experiments were run for 2 years due to the limit in our computational resource.
Table 1
A List of Experiments Conducted in This Study

| Experiment | External KW frequency | Resolution | OML | Ocean coupling relaxation time |
|------------|-----------------------|------------|-----|-------------------------------|
| CTL        | 10 days               | 27 km      | 1 m | $\tau = 30$ days              |
| K5         | 5 days                | 27 km      | 1 m | $\tau = 30$ days              |
| NOK        | no external Kelvin wave | 27 km    | 1 m | $\tau = 30$ days              |
| E54        | 10 days               | 54 km      | 1 m | $\tau = 30$ days              |
| E81        | 10 days               | 81 km      | 1 m | $\tau = 30$ days              |
| M10        | 10 days               | 27 km      | 10 m| $\tau = 30$ days              |
| M50        | 10 days               | 27 km      | 50 m| $\tau = 30$ days              |
| T5         | 10 days               | 27 km      | 1 m | $\tau = 5$ days               |
| T50        | 10 days               | 27 km      | 1 m | $\tau = 50$ days              |

2.3. TC tracking Algorithm

Given model output from the WRF simulations, an important step is to have an efficient algorithm that can track all model TCs detected over the entire domain. There are various versions of TC tracking schemes designed for numerical model outputs, which are based on different TC criteria such as the maximum wind speed at different levels, relative vorticity, warm core magnitude, TC locations, or duration (e.g., Broccoli & Manabe, 1990; Bengtsson et al., 1982; Camargo, 2013; Camargo & Zebiak, 2002; Horn et al., 2014; Oouchi et al., 2006; Strachan et al., 2013; Tsutsui & Kasahara, 1996; Ullrich & Zarzycki, 2017; Walsh & Katzfey, 2000; Walsh et al., 2007, 2015; Zarzycki & Ullrich, 2017; Zhao et al., 2009). A common feature in all TC-tracking schemes is that the number of TCs detected from these schemes is sensitive to different model configurations and horizontal resolutions. For example, Walsh et al. (2007) noticed that using a criterion of the maximum wind speed of 17.5 m s$^{-1}$ for 10-m wind can be adapted for horizontal resolutions smaller than about 10 km, while for a 30-km spaced grid, a wind speed threshold of about 17.0 m s$^{-1}$ is applicable, and for a very coarse resolution of about 125 km, a 10-m wind speed of roughly 14.5 m s$^{-1}$ should be chosen. Thus, the threshold of the maximum wind speed for detecting a model vortex seems to increase inversely to the model resolution. Likewise, one could simplify the tracking process by focusing on few key parameters as found in (Chavas & Reed, 2019; Chavas et al., 2017; Ullrich & Zarzycki, 2017), which use only the surface pressure and warm core thresholds to avoid the dependence of the surface wind speed on the model resolution.

Because of such a sensitivity of the TC tracking algorithm to specific models and configuration, we present here an algorithm to track TCs specifically for our 27-km resolution model output, which is based on several basic criteria as follows:

- A preconditioning is applied over the entire domain to search for potential TC center locations. Specifically, the minimum surface pressure is checked at all model grid points to see if there is any value below 1,004 hPa. If so, a patch of a size $20^\circ \times 20^\circ$ around that grid point is marked such that the search will skip this whole patch to reduce redundant computational time, thus speeding up the search of vortex centers.
- Once the potential TC center patches are marked, the location of the maximum relative vorticity center will be found within each patch, and the relative distance between the surface pressure minimum location and the vorticity maximum location in each patch will be computed. Any distance smaller than 150 km will be marked as a next candidate for a TC center for further check in the next step. Otherwise, the patch will be discarded. This check is based on the physical condition that the circulation center and the minimum central pressure of a typical TC must be sufficiently close to one another.
- After the consistency between the maximum vorticity center and the minimum surface pressure is confirmed, the 10-m wind speed maximum is searched within a region of $3^\circ \times 3^\circ$ square around the center of the minimum surface pressure. A criterion of the maximum 10-m wind greater 17.5 m s$^{-1}$ at a distance of 300 km from the minimum surface pressure is also applied to make sure that the circulation structure could indeed represent a typical 34-kt radius of TCs.
- Next the thermodynamic structure of a TC by computing temperature anomaly at 400 hPa for the TC centers that pass the three previous steps. Here, the temperature anomaly is calculated by subtracting the temperature averaged within an area of $18^\circ \times 18^\circ$ around any TC center. A possible TC center that has the temperature anomaly greater than 3K will be finally tagged as the storm center and recorded.
Table 2
Minimum Threshold Detection Criteria for Tropical Cyclones With 27, 54, and 81 km Resolutions

| Model resolution | Latitude range | Gale-force wind (m s⁻¹) | Minimum pressure (hPa) | Storm outer size (km) | Distance between $P_{\text{min}}$ and vorticity centers (deg) | Temperature anomaly at 400 hPa (K) | Duration (days) |
|------------------|----------------|-------------------------|------------------------|----------------------|---------------------------------------------------------------|----------------------------------|----------------|
| 27 km            | −37° S to 37° N| 17.5                    | 1,004                  | 300                  | 5                                                             | 3                                | 3              |
| 54 km            | −37° S to 37° N| 14                      | 1,008                  | 200                  | 5                                                             | 3                                | 3              |
| 81 km            | −37° S to 37° N| 12                      | 1,010                  | 100                  | 5                                                             | 3                                | 3              |

Finally, the life cycle of each detected storm is checked to ensure that any detected TC has to possess a lifetime of at least 3 days to avoid vortices with too short lifetime.

For all experiments, this tracking program is applied to 6-hourly model output such that a complete history of TC center locations is recorded. Sensitivity analyses for this tracking algorithm, which are needed for detecting TCs from model gridded output as emphasized in Ullrich and Zarzycki (2017) and Zarzycki and Ullrich (2017), are presented in Supporting Information. Detailed TC-tracking criteria for the 27-km control run as well as the criteria for the two sensitivity experiments at 54- and 81-km resolutions are given in the Table 2, which has a similar range as those used in previous studies (e.g., Camargo, 2013; Oouchi et al., 2006; Walsh et al., 2007; Zarzycki & Ullrich, 2017).

3. Results
3.1. Global TC Distribution

We begin with a broad picture of the TC climatology developed in the CTL configuration. Figure 3a displays all TC tracks during the second year of the WRF model simulation. As expected, most TCs form in the tropical region between 5° and 10° latitude and then move poleward and westward under the large-scale easterly flows and eventually drift to higher latitudes due to the effect of beta drift. Of interest is that the total number of TCs detected during the entire year is bounded at approximately 350 in the aquaplanet setting, regardless of how frequent the tropical atmosphere is perturbed by the tropical Kelvin waves or not. As shown in Figures 3b and 3c, adding Kelvin wave perturbations every 5 days (K5) or 10 days or no adding Kelvin wave perturbations at all (NOK) produces a very similar TC climatology in terms of the total TC number as well as genesis locations and large-scale track patterns. This is a noteworthy result, because the aquaplanet tropical atmosphere and the SST condition are, by design, very favorable for TC development. Despite the ideal thermodynamic environment and constant stirring of the tropical region, the tropical atmosphere produces a similar number of TCs. In this regard, the consistency of the TC count among different experiments suggests that the maximum potential genesis of the tropical atmosphere must be governed by some internal dynamical or energetic constraints rather than specific triggering mechanisms.

In terms of seasonality of the global TC count, Figure 4 shows the number of TCs for the summer and winter seasons separately in each hemisphere. The Northern Hemisphere TC summer season is defined to be an interval from days 118–300 in the second year of the CTL simulation, and the remaining days are for the Southern Hemisphere TC season summer (see Figure 4). As expected, TCs form almost exclusively in the warm seasons, although here annual genesis is more symmetric about the equator than that found in the real world. Each hemisphere’s warm season produces ~170 TCs as shown in Figures 4a–4c, whether or not the external wave...
Figure 4. (a) Seasonal distribution of the locations of TC formation in the Northern Hemisphere summer (red) and the Southern Hemisphere summer (blue) for the CTL experiment, (b–c) Hovmöller diagrams of the sea surface temperature and daily precipitation during the second year of the CTL experiment, (d–f) similar to (a)–(c) but for the K5 experiment, and (g–i) for the NOK experiment.

Perturbations are added. Notice that the total TC count for each hemispheric summer is close but not exactly the same due to more solar incoming in the Southern Hemispheric summer. Thus, the model develops some large-scale asymmetry with slightly more TCs in the Southern Hemisphere, consistent with such difference in solar incoming between the two summers. As shown by the corresponding SST and precipitation distributions, the strong seasonal signal of TC counts is mostly related to the migration of the ITCZ between the two hemispheres. These results align with the findings of Merlis et al. (2013) and explain the distinct distribution of the total TC count between the two seasons in all simulations shown in Figure 4.

Unlike the real atmosphere for which TC counts between the Northern and Southern Hemispheres are drastically different, one notices in Figure 4 that the storm count is symmetric between the hemispheres in all experiments. This difference is likely due to the fact that the observed regional TC climatology is determined not only by SST distribution but also by a number of other large-scale conditions such as the abundance of tropical disturbances or vertical wind shear (Gray, 1968, 1982). Moreover, the topography of the Northern Hemisphere alters global and regional circulation patterns, including the monsoon, which further affects the ITCZ as well as the large-scale environment (e.g., Moore & Philander, 1997; Pike, 1971; Philander et al., 1996, 1996; Takahashi & Battisti, 2007; Xie & Philander, 1994), ultimately favoring genesis in the Northern Hemisphere. In our aquaplanet configuration, there is no such land-sea interaction or land coverage asymmetry, thus resulting in more homogeneous TC formation in both hemispheres.

Given the importance of the ocean surface temperature to the migration of the ITCZ and related seasonality of global TC formation, Figure 5 shows the global TC counts and related SST response for our experiments using different ocean mixed layer depths and relaxation times. Despite a similar qualitative pattern of the seasonal response in all experiments, the magnitude and the meridional migration of the ITCZ are substantially different among these experiments. Specifically, increasing the ocean mixed layer depth leads to smaller ITCZ migration (M10 Figures 5a and 5b and M30 Figures 5c and 5d). As the OML becomes sufficiently deep (>30 m), the ITCZ becomes more strongly constrained to remain near the equator as shown in
Figure 5d, and TC count decreases. In fact, for OML > 50 m (not shown), the ITCZ is largely confined within 10° S to 10° N, and fewer TCs (<100 TCs) form. In contrast, OML = 10 m gives the same number of storms as OML = 1 m in the CTL experiment, indicating that the annual count is little affected for a sufficiently shallow OML.

Decreasing the ocean temperature relaxation time plays a similar role to increasing the ocean mixed-layer depth. This is because a shorter relaxation time acts as a stronger forcing back toward the climatological-mean and thus damps the seasonal cycle. As shown in Figures 5e–5h, a shorter relaxation time results in smaller amplitude meridional ITCZ movement. In fact, for relaxation time less 5 days, the seasonal migration of the ITCZ is mostly between 10° S to 10° N, and the corresponding SST difference is also significantly less responsive due to the rapid adjustment of the mixed layer toward the reference state. This constraint on ITCZ migration to remain near the equator translates to less favorable conditions for
barotropic instability to develop as discussed in W19. As such, fewer tropical disturbances can be produced, which accounts for an overall smaller number of TCs.

The consistent symmetry in TC count across all of these experiments apparently confirm that the asymmetry in TC climatology between the seasons and between the Northern and Southern Hemispheres in the real atmosphere can be attributed to heterogeneity in the land-sea distribution. Given the consistent upper bounds in TC formation irrespective of adding Kelvin wave perturbations or not, it is natural to ask a question of what are the large-scale factors that govern such a cap on global TC formation in this favorable and symmetric environment. We tackle this question next.

3.2. TC Episodic Formation

To examine the frequency of TC development at the global scale, Figure 6a shows a time series of the number of daily TC genesis events detected from the model output during the second year of the CTL simulation. Here, the number of daily TC genesis events is defined as the number of new TCs detected within the main genesis domain between $[20^\circ S$ and $20^\circ N]$ (cf. Figure 2). Such a limit in searching for new TC genesis events within the $20^\circ S$ to $20^\circ N$ region is consistent with the seasonal migration of the ITCZ shown in Figure 5, which is our main focus here. This choice also avoids the interference of storms previously formed in the tropical region and subsequently moving to higher latitudes, rather than the newly formed systems.

As can be seen from the time series of the daily genesis events, global genesis is indeed not continuous but takes place in a series of episodes that are superimposed on the seasonal cycle, even under ideal zonally symmetric thermodynamic forcing. Spectral analysis of this TC genesis time series (Figure 6b) first shows a dominant subseasonal spectral peak emerging (see blue line in Figure 6b) at a period of $\sim$14 days (2 weeks), which corresponds to the largest variability in the TC number during the summer months. Along with this dominant frequency, one notices several other frequencies in the range of 5–10, 17–20, and 25–30 days, which are likely related to different tropical wave modes reported in, for example, Chen and Sui (2010), Kiladis et al. (2009), and Molinari et al. (2007).
Among these frequencies, we note that the dominant 2-week frequency corresponds to the largest variation of the TC daily genesis events (Figure 6a). Furthermore, the maximum number of new TC genesis events for any given day is apparently capped ($\leq 10$), which is in line with W19 and Ferreira and Schubert (1997) (cf. Figure 6a). Taken together, a 2-week frequency of TC formation with a cap in the number of TCS produced in each cycle suggests that the total number of TCS that the tropical atmosphere can generate must be limited per year ($\sim 350$ in our experiments herein). If one assumes that the Earth is half covered by land, then the total number of TCS is expected to be cut by half, which is closer to the observed range 80–100 TCS in the real atmosphere. Of course, this number is modified further in the real world by variations in SST and environmental favorability. However, here we have arrived at a reasonable order of magnitude estimate that is consistent in all experiments.

From the scaling perspective, the upper limit in the daily number of TC genesis events shown in Figure 6 is also consistent with the typical TC horizontal scale discussed in Hoogewind et al. (2020). Assume, for example, an inter-TC spacing of 3,000 km diameter, then the largest number of TCS that the atmosphere can generate per day is roughly equal to the number of TCS that can be arranged around the Earth’s circumference, which is about $\sim 14$ TCS for any given time. This scale estimation is well captured from both satellite images and model simulations (cf. Figure 1b). As an illustration, Figure 7 shows a snapshot of the model output when 11 tropical disturbances emerge at the beginning of a TC formation episode as captured in the CTL experiment, along with a similar snapshot at the end of this episode about 2 weeks later. The upper bound and the contrast in the number of TCS counted at the beginning and at the end of this TC formation outbreak are well captured in these snapshots, thus highlighting the unique episodic nature of the global TC formation.

It should be mentioned that the 2-week cycle of TC formation as obtained in the CTL experiment is significantly longer than the typical life cycle of a real TC ($\sim 7–10$ days). Physically, the longer lifetime of TCs in the CTL experiment could be linked to our idealized setting that allows a perfect environment for TCs to develop and maintain their properties, so long as the TCs stay over a relatively warm ocean surface. In the real atmosphere, the TC life cycle is generally shorter, as most TCs either make a landfall or move to higher latitudes where they dissipate over colder water. Regardless of this difference in the TC lifetime, the noticeable contrast in the TC number at the beginning and at the end of a TC episode is apparent and indicates some fundamental changes in the large-scale environment for TC development that we wish to explore.

While W19 can help explain the upper bound on TC formation in a given episode, we also need to understand why these episodes peak at the 2-week frequency and which environmental factors contribute to this behavior. To tackle this question, we analyze the TC genesis potential index (GPI) proposed by Emanuel and Nolan (2004) and Nolan et al. (2007).

\[
GPI = 10^6 \eta^2 \left( \frac{\chi}{50} \right)^3 \left( \frac{\text{VPI}}{70} \right)^3 (1 + 0.1 \text{Vshear})^{-2},
\]

where $\text{Vshear}$ denotes the total vertical wind shear, $\text{VPI}$ denotes Emanuel’s theoretical maximum potential intensity, $\eta$ is the low-level absolute vorticity at 850 hPa, and $\chi$ is midlevel relative humidity at 600 hPa. Here, the vertical wind shear is computed for 850 and 200 hPa levels, while the midlevel relative humidity is defined at 600 hPa level following Camargo and Wing (2016). While GPI has been used frequently to study climatological variability (e.g., Emanuel, 2005; Camargo & Wing, 2016; Zhao & Held, 2012), the GPI formulation given by Equation 5 has yet to be applied to study global TC formation on shorter timescales.
alternative formulation by Emanuel (2005) uses the midlevel entropy deficit in lieu of the relative humidity, which has a stronger theoretical basis but is ultimately similar for our purposes (see the supporting information).

Figure 8 shows the time series of the GPI index and the daily TC genesis number. Overall, GPI displays a similar variation with the TC genesis frequency, especially at the seasonal time scale. Within each hemispheric summer season, GPI also shows an episodic variability similar to the number of TC genesis events, with higher GPI for the day with more TC genesis events. Note also that once a genesis outbreak has occurred, there is no more room for additional events for a few days, and the corresponding GPI decreases gradually. This suggests that GPI can be in fact useful to examine the variability of global TC formation beyond its original design for monthly or longer climatology (Emanuel & Nolan, 2004). Of course, high GPI is only a necessary indication for tropical cyclogenesis to take place and by no means sufficient. However, the similar behaviors of GPI and TC genesis event time series as shown in Figure 8 suggest that GPI does contain some important information of the large-scale conditions that control TC genesis variability on these timescales.

To examine how variations in GPI may help understand this episodic behavior, Figure 9 compares the composite zonal-mean GPI at the onset and the end of the TC formation 2-week cycle for each hemisphere summer. Here, the composite GPI at the onset of TC genesis episodes is computed by taking a zonal average of GPI at the beginning of all episodes. Note that an onset event is defined to be a local maximum in daily TC genesis count that is at least 2 weeks apart from other peaks during each hemisphere summer season. This yields a total of nine episodes for the Northern Hemisphere summer (Days 118–300) and eight for the Southern Hemisphere summer (the remaining days) for our composite as shown in Figure 6a. Similarly, the composite GPI at the end of TC genesis episodes is an average of GPI over all the days with a minimum number of TC genesis events in each hemisphere summer.

One notices in Figure 9 that the composite GPI generally peaks in the region between 8–12° in both hemispheres, which is consistent with the main location of TC development associated with the ITCZ shown in Figures 3 and 4. Moving poleward, GPI decreases because potential intensity decreases over the cold SSTs and vertical wind shear increases from thermal wind balance. Moving equatorward, the GPI also decreases rapidly as absolute vorticity decreases, which explains why TCs are not allowed to form too close to the equator (Chavas & Reed, 2019).

Of relevance to global TC formation frequency is that GPI at the onset of a TC formation episode is significantly higher than that at the end of the episode (highlighted as a dark blue area in Figure 9). The signal is robust in both hemispheres over a range of experiments, with an overall GPI about 25% higher at the onset within 8–12° than that at the end of the episode. These results indicate that there is intrinsic variability in the large-scale environmental favorability for TC genesis associated with this episodic behavior.

To examine which factors within GPI contribute the most to the total variations in GPI shown in Figure 9, it is necessary to separate each factor in Equation 5. Let $G_1$ and $G_2$ be the composite GPI at the onset and at the end of all TC formation episodes. Take a natural logarithm of $G_1$ and $G_2$ and subtract from each other, we obtain the following relationship:

$$\frac{\Delta \text{GPI}}{\text{GPI}} = \frac{G_1 - G_2}{G_2} = \frac{3}{2} \frac{\Delta \eta}{\eta} + 3 \frac{\Delta \chi}{\chi} + 3 \frac{\Delta \text{VPI}}{\text{VPI}} - 0.2 \frac{\Delta \text{Vshear}}{1 + 0.1 \text{Vshear}}. \quad (6)$$
Figure 9. The composite of the zonally-averaged GPI at the onset and end of TC formation episodes in the Northern Hemisphere (right panels) and the Southern Hemisphere (left panels) obtained from (a and b) CTL, (c and d) K5, (e and f) NOK, (g and h) M10, and (i–k) M50 experiment. The red/blue solid lines denote the composite GPI at the onset/end of the TC genesis episode, which is averaged over all episodes in the second year of the CTL simulation. The thin lines denote GPI for each episode, and the shaded region highlights the GPI difference between the start and the end of the TC genesis episode. The blue boxes highlight the latitude range where the majority of TC genesis takes place in each hemisphere.
Figure 10. The relative contribution of each individual factor in the GPI index including zonally averaged potential intensity $V_{JI}$ (purple), 600-hPa relative humidity (cyan), absolute vorticity (red), and vertical wind shear between 850- to 200-hPa levels (blue) for CTL experiment in the Southern Hemisphere (left panels) and the Northern Hemisphere (right panels) for (a and b) CTL, (c and d) K5, (e and f) NOK, (g and h) M10, and (i–k) M50 experiment.

where the bar denotes the average over both the zonal direction and all episodes and the notation $\Delta$ denotes the difference of each GPI factor between the onset and the end of all episodes. Here, we have assumed that each factor $\Delta \bar{\eta}$, $\Delta \bar{\chi}$, $\Delta \bar{VPI}$, and $\Delta \bar{Vshear}$ are sufficiently less than 1 within the main TC genesis region. This linear framework allows one to evaluate which factor contributes the most to the difference in GPI as expected, so long as the relative difference of each factor on the right-hand side of Equation 6 is sufficiently small as can be seen in Figure 10. While these four factors are relatively independent in the sense that they...
Figure 11. Similar to Figure 6 but for (a and b) the 54-km (E54) resolution experiment and (c and d) the 81-km (E81) resolution experiment.

represent different large-scale conditions, it should be noted that they are ultimately related to each other via the internal modes of variability in any climate system and hence do not truly vary independently. Our main purpose here is, however, to separate their relative change between the onset and the end of TC episodes. Thus, the interrelation among these factors at the climate equilibrium will not be examined herein.

Among the four factors shown in Figure 10, one notices that the 600-hPa relative humidity, the 850-hPa absolute vorticity, and the wind shear each contribute significantly to the overall change in GPI between the onset and the end of TC genesis episodes, while the potential intensity factor plays the least significant role. Further examination of the three main factors in Figure 10 shows that the 600-hPa relative humidity and the absolute vorticity are consistently higher while the vertical wind shear is relatively weaker at the onset of all episodes in the region where GPI peaks. Similar results are also obtained for the Southern Hemisphere with the exception of wind shear, which has a smaller contribution as compared to that in the Northern Hemisphere.

Note that in both hemispheres the potential intensity factor does not contribute much, because SSTs are slow to evolve and thus tend not to vary too strongly over such a short timescale. Meanwhile, the higher absolute vorticity in this fixed latitude band reflects an increase in low-level vorticity within the ITCZ leading up to cyclogenesis onset. Likewise, higher midlevel relative humidity indicates a moister free troposphere in the vicinity of the ITCZ. These factors, in addition to a reduction in vertical wind shear, all act together to enhance the favorability for TC genesis. Additional weaker wind shear can also help protect a newly developed system such that tropical disturbances can be organized and subsequently grow, albeit its roles are not as consistent in both hemispheres as the other factors.

As a string of TC disturbances emerge and evolve into TCs under such favorable conditions, the moisture content in the region between 8° and 12° gradually decreases with time. Similarly, the large-scale vorticity within the ITCZ region from 8–12° is higher at the onset of a TC formation episode, yet it starts to decrease with time until the end of the episode. Such a pattern of TC formation favorability emerging and subsiding in an episodic manner indicates that TCs not only consume deep organized convection and dry out the environment around the storms while transporting moisture within the storm poleward out of the deep tropics but also act as a pathway to remove the instability related to the ITCZ breakdown as found in W19. From this perspective, the 2-week cycle can be considered as a time scale for the combined effects of remoistening the tropical atmosphere and restoring the ITCZ flow field toward a critical state before instability can recur and a new episode of TCs can form. Formulation of a closed-form theory for this dynamical evolution would be a valuable avenue of future work.

3.3. Model Resolution Sensitivity

Because of the dependence of TC formation to the model horizontal resolution, a final set of sensitivity experiments is examined to verify whether the results obtained in the CTL experiment with the 27-km resolution are robust. Except for the coarser resolution, we note again that all other model configurations and initialization are kept unchanged in these resolution sensitivity experiments to isolate the effects of resolution.
In addition, the TC tracking algorithm must be modified to properly capture TCs at the lower resolution as given in Table 2.

As expected from the coarse-resolution experiments (E54 and E81), both the number of TCs and the strengths of TCs in these sensitivity experiments are significantly less than CTL, yielding 151 and 137 TCs, respectively, as compared to ∼350 TCs from the CTL experiment. Along with fewer TCs, the average TC intensity is also significantly lower with the maximum 10-m wind of 29, and 25 m s$^{-1}$ for the 54-, and 81-km resolutions, respectively (not shown). Despite this difference in the global TC counts, the episodic nature of TC formation is still however well maintained. Figure 11 shows the time series of TC genesis events for the E54 and E81 experiments, similar to that shown in the CTL experiment. There again is a dominant seasonal spectral peak associated with the ITCZ migration (the black spectrum in Figure 11), as well as a shorter frequency peak ($\approx$20 days) for both the 54- and the 81-km resolution (the red spectrum in Figure 11).

While the 2-week TC formation frequency is somewhat less apparent for the coarse resolutions than that in the CTL experiment, which is likely due to TC detection at coarse resolutions, the episodic nature of TC formation is evident among all experiments. Despite the range of the dominant spectral peaks is shifted, we note that these spectral peaks are still mostly concentrated in the range of 15–25 days. This is an important property of the tropical atmosphere where it takes some time for favorable conditions to reach a critical threshold before TC formation can occur.

To verify the difference in the large-scale conditions between the onset and the end of the TC formation episodes, Figures 12–13 show a composite analysis of GPI for 7 TC formation episodes obtained from the E54 and E81 experiments, respectively, along with the percentage contribution of each large-scale factor. One can see the same characteristics of the GPI distribution in the tropical region similar to the 27-km resolution experiment, with the peaks of GPI in the 8–12° region in both hemispheres. Likewise, the composite GPI difference between the onset and the end of the TC formation episode is still apparent in the region 8–12°, similarly to what obtained in the CTL experiments.

Separating each individual GPI factor for these coarse-resolution experiments confirms that the larger absolute vorticity, the higher midlevel moisture, and the weaker vertical wind shear are the dominant factors distinguishing the onset and the end of each TC formation episode. Unlike the 27-km experiments, PI has a slightly stronger role for the E54 experiment, albeit it is still relatively smaller compared to absolute vorticity and midlevel moisture. Taken together, our experiments suggest that global TC formation is not a
random and continuous process, but it instead proceeds in episodic manner at frequency of about 2 weeks in a simplified world with zonally symmetric forcing.

It should be noted that the above intermittent behavior of global TC formation is robust not only to the model resolution but also to different TC tracking algorithms. As discussed in section 2, tracking a TC formation and its subsequent motion from a numerical simulation depends also on various criteria used to define a TC in the model world beyond the model resolution. This tracking sensitivity issue has been examined thoroughly in recent studies by (Horn et al., 2014; Ullrich & Zarzycki, 2017; Zarzycki & Ullrich, 2017), which proposed a tracking sensitivity analysis approach for evaluating the robustness of results in TC genesis research. Our sensitivity analyses of tracking algorithm following their approach do capture similar episodic development of TC formation as well as the dominant role of midlevel moisture and absolute vorticity as shown above (see the supporting information). In this regard, the tracking sensitivity analyses also support the general validity of the main results obtained in this study.

4. Conclusion

This study examined two central issues related to global TC genesis: (1) an upper bound on the total number of TCs that the tropical atmosphere can annually produce given ideal thermodynamic conditions and (2) the intrinsic temporal variability of TC formation. Using the WRF model with an idealized aquaplanet setting, a range of simulations were carried out to investigate the limiting factors to TC formation in a tropical channel between 45° S and 45° N. By constantly perturbing the tropical atmosphere with equatorial Kelvin wave perturbations at different intervals to mimic a triggering mechanism for tropical cyclogenesis, it was found that the tropical atmosphere indeed possesses an upper bound in producing TCs. Specifically, there are about 350 TCs detected in an aquaplanet environment, regardless of how frequently external wave perturbations are added to the model or if they are excluded. This upper bound on the total number of TCs indicates a maximum capacity of the tropical atmosphere to produce TCs, which can be observed in a favorable idealized environment.

In addition to an upper bound on the number of TCs that can form in one year, there is no clear difference in TC count between the two hemispheres in our aquaplanet simulations. Unlike the real atmosphere that has a distinct difference between Southern and Northern Hemispheres, the same number of TCs was observed in the two summer hemispheres for the idealized aquaplanet tropical channel. This is true whether or not the external perturbation excitation is applied, and it is linked to the lack of landmass and ocean current effects in the aquaplanet simulation, which prevents the asymmetry of the ITCZ between the two hemispheres where TC disturbances are hosted.

Further analyses of the temporal variability on tropical cyclogenesis frequency revealed that global TC formation does not occur continuously throughout the year. Instead, TC formation takes place in episodes at a frequency of about 2 weeks. Furthermore, the maximum number of the daily TC genesis events is not arbitrary but capped by a limit of approximately 10 genesis events at any given day. Such an upper bound in the daily TC genesis events for each episode is consistent with a recent theoretical estimation of the TC disturbances in the ITCZ breakdown model by Wang et al. (2019), which showed that the maximum number of TC disturbances for any given day cannot be larger than 12 due to the internal constraint of the tropical dynamics.

The cap on the maximum number of daily genesis events together with the episodic nature of global TC formation at a 2-week frequency found in this study offers an explanation for an upper bound on the annual global TC number. Indeed, assuming a cap of 14–15 TCs per episode and a frequency of 2 weeks between episodes as obtained from our simulations, it can be seen that the total number of TCs that the tropical atmosphere can support is ~350 TCs per year for an aquaplanet. This number will be lower if one further takes into account the Earth’s surface area, which is about half covered by land in the Northern Hemisphere. While the absolute number of TCs from a model simulation varies with different model resolutions and parameters, the episodic nature of the global TC formation and the upper bound on daily genesis events are consistent across all sensitivity experiments, irrespective of the tracking algorithm or model configurations, suggesting that it is a robust feature of a simplified tropical region.

Analyses of the zonally-averaged TC genesis potential index (GPI) showed significantly higher GPI at the onset of a TC formation episode within the main developing region 8°–12° in both hemispheres as compared
to the end of the episode. Additional analyses in which each environmental factor was isolated showed that changes in GPI were dominated by changes in 600-hPa moisture content and 850-hPa absolute vorticity, both of which are relatively larger at the onset of TC formation outbreaks. Once TCs form and move to higher latitudes, the tropical troposphere moisture and the low-level relative vorticity within 8–12° region decreases, thus producing a temporarily less favorable environment for new TC formation toward the end of TC formation episodes. It takes approximately 2 weeks for the tropical atmosphere to remoisten and restore larger absolute vorticity in the vicinity of the ITCZ before a new episode of TC formation can take place.

While model simulations could help address the fundamental questions in global TC formation, several caveats in this study must be acknowledged. First, the absolute number of TCs detected from a model simulation depends on various subjective factors such as model resolution, TC physics, or tracking criteria. Hence, the specific number of ~350 found in our aquaplanet configuration should be taken with caution. Despite this inconclusive number of total TCs, we wish to note that our primary focus in this study is more on the intrinsic genesis variability, which is found to be robust across experiments and provides insight into the maximum potential genesis of the tropical atmosphere in an ideal thermodynamic state. Likewise, all of our analyses emphasize on the relative difference in the large-scale factors among different sensitivity experiments under the same aquaplanet environment, rather than focusing on a specific TC count in all experiments. In this regard, any exact number of TCs that can be formed in the real tropical atmosphere at the global scale is still an open question.

Second, the interpretation of our analyses is limited to some degree by computational constraints. Specifically, the statistical significance of the differences in large-scale factors between the onset and the end of TC episodic development could not be conclusively assessed. Similarly, the spectral analyses are also inconclusive, thus rendering some uncertainties in determining the episodic frequency of global TC formation that we wish to quantify. This caveat was partially addressed in this work by using different sensitivity experiments such as coarser-resolution experiments or different ocean mixed layer depth to help verify the signal obtained from the control experiment.

Finally, the use of a single type of tropical Kelvin waves as an external perturbation source to create favorable conditions for TC formation cannot fully capture the entire spectrum of perturbations in the tropical region. Ideally, one should examine as many different wave types as possible for more conclusive results, which would be a fruitful avenue of future work. Due to a large number of simulations required for the external perturbation design in this study, we limited our work to the tropical Kelvin wave type perturbation. The fact that all experiments with or without external Kelvin wave perturbations could capture almost the same behavior of global TC formation indicates at least that this type of external wave does not seem to help produce more TCs. Thus, this result advocates the limited capability of tropical atmosphere in forming TCs, though additional work is needed to further verify these conclusions. In this regard, more experiments will be needed to increase the robustness and to capture other large-scale signals that can shed additional light into the episodic nature of global TC formation.

Data Availability Statement

The tropical input sounding data required to initialize all WRF idealized simulations in this study can be found in Jordan (1958). All other model parameters and configurations that are needed to reproduce our simulations are given in section 2. TC detection data output presented in this study is available online (at https://doi.org/10.13140/RG.2.2.14244.30080).

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