Dynamical Aquaplanet Experiments with Uniform Thermal Forcing: System Dynamics and Implications for Tropical Cyclone Genesis and Size

DANIEL R. CHAVAS

Department of Earth, Atmospheric, and Planetary Sciences, Purdue University, West Lafayette, Indiana

KEVIN A. REED

School of Marine and Atmospheric Sciences, Stony Brook University, State University of New York, Stony Brook, New York

(Manuscript received 1 January 2019, in final form 25 April 2019)

ABSTRACT

Existing hypotheses for the dynamical dependence of tropical cyclone genesis and size on latitude depend principally on the Coriolis parameter f . These hypotheses are tested via dynamical aquaplanet experiments with uniform thermal forcing in which planetary rotation rate and planetary radius are varied relative to Earth values; the control simulation is also compared to a present-day Earth simulation. Storm genesis rate collapses to a quasi-universal dependence on f that attains its maximum at the critical latitude, where the inverse- f scale and Rhines scale are equal. Minimum genesis distance from the equator is set by the equatorial Rhines (or deformation) scale and not by a minimum value of f . Outer storm size qualitatively follows the smaller of the two length scales, including a slow increase with latitude equatorward of 45° in the control simulation, similar to the Earth simulation. The latitude of peak size scales with the critical latitude for varying planetary radius but not rotation rate, possibly because of the dependence of genesis specifically on f . The latitudes of peak size and peak packing density scale closely together. Results suggest that temporal effects and interstorm interaction may be significant for size dynamics. More generally, the critical latitude separates two regimes: 1) a mixed wave–cyclone equatorial belt, where wave effects are strong and the Rhines scale likely limits storm size, and 2) a cyclone-filled polar cap, where wave effects are weak and cyclones persist. The large-planet limit predicts a world nearly covered with long-lived storms, equivalent to the f plane. Overall, spherical geometry is likely important for understanding tropical cyclone genesis and size on Earthlike planets.

1. Introduction

Tropical cyclone genesis and size are known to vary with latitude on Earth, though the underlying physics of this variability remains poorly understood. Prevailing hypotheses for these quantities depend principally on the local value of the Coriolis parameter f .

First, storm genesis rate increases empirically with increasing absolute vorticity, as captured by various metrics of genesis potential (Emanuel and Nolan 2004; Camargo et al. 2014). For relatively quiescent flow with

weak relative vorticity, this result reduces to a dependence on f . Similarly, a forced poleward shift of the ITCZ in idealized aquaplanet simulations has been shown to dramatically increase the genesis rate (Merlis et al. 2013). Moreover, it is well known that storm genesis in nature rarely occurs within $\sim 5^\circ$ latitude of the equator (Gray 1968). The prevailing theoretical argument for this behavior is the requirement of sufficiently large magnitude of ambient absolute vorticity to supply angular momentum to the system (Emanuel 2003; Anthes 1982). Implicitly, then, a plausible hypothesis is that genesis rate and minimum genesis latitude both depend fundamentally on f , neither of which has yet been tested experimentally. Testing physical hypotheses for genesis is difficult using observations or simulations of Earth, though, as midlatitude dynamics associated with large-scale baroclinicity and the jet stream create a

Supplemental information related to this paper is available at the Journals Online website: <https://doi.org/10.1175/JAS-D-19-0001.s1>.

Corresponding author: Daniel R. Chavas, drchavas@gmail.com

hostile thermodynamic environment that significantly depresses storm activity moving poleward out of the tropics (Tang and Emanuel 2012).

Second, outer storm size is predicted by theory to scale inversely with f . This inverse- f scaling has been demonstrated in idealized rotating radiative–convective equilibrium (RCE) simulation experiments on an f -plane in axisymmetry (Chavas and Emanuel 2014) and 3D geometry (Khairoutdinov and Emanuel 2013; Zhou et al. 2014; Merlis et al. 2016; Zhou et al. 2017). In contrast, storm size in observations tends to remain constant or increase with latitude (Kossin et al. 2007; Knaff et al. 2014), with perhaps a slight decrease toward higher latitudes (Chan and Chan 2015; Chavas et al. 2016; Schenkel et al. 2018). Alternative explanations for this observed behavior have been proposed to be related to both internal storm factors, such as inertial stability (Smith et al. 2011; Chan and Chan 2014) and storm age (Kossin et al. 2007), as well as environmental factors such as synoptic-scale variations in ambient angular momentum (Chan and Chan 2013) and the increasing probability of extratropical transition (Hart and Evans 2001), which tends to induce storm expansion (Hart et al. 2006). However, given that an inverse- f scaling decreases very rapidly moving poleward at low latitudes, such factors appear unlikely to explain the large discrepancy between observations and existing theory. Instead, a novel hypothesis is required. Perhaps the simplest such hypothesis is that storm size in nature depends in some way on the spherical geometry of a rotating planet.

The focus of this work is to test existing and novel hypotheses relevant to tropical cyclone genesis and size in spherical geometry. Given the complexity of the real Earth, an ideal experimental laboratory is a simplified Earthlike rotating rocky planet in the absence of large-scale environmental baroclinicity created by spatial heterogeneity in thermodynamic forcing, including solar insolation and land. Such a system has been analyzed in general circulation model (GCM) experiments in an aquaplanet configuration under uniform thermal forcing (Shi and Bretherton 2014; Merlis et al. 2016), which might also be referred to as “spherical rotating radiative–convective equilibrium” in the context of its f -plane counterpart. The dominant large-scale circulations are tropical cyclones that form principally at low latitudes—as is found in nature—but may propagate all the way to the poles. This experimental design eliminates large-scale baroclinicity in the climate system while retaining the essential dynamical variability of a rotating, spherical Earthlike planet. It offers significant benefits for studying both the internal dynamics of the tropical cyclone and its spatiotemporal variability, as

global model simulations generate large numbers of storms that emerge naturally within an equilibrated climate system (Chavas et al. 2017). The end result is a clean experimental testing ground for fundamental dynamical controls on tropical cyclone variability.

Our principal research questions are as follows:

- 1) How do storm size and genesis vary with latitude in a world where tropical cyclones are allowed to propagate all the way to the poles?
- 2) Is there a fundamental dynamical dependence of genesis rate on f ?
- 3) What sets the minimum genesis distance from the equator?
- 4) What sets storm size as a function of latitude, and how does this compare with nature?
- 5) Can we understand the qualitative dynamical behavior of this idealized system theoretically?

To answer these questions, we perform dynamical experiments on an aquaplanet with uniform thermal forcing in which we vary each of the two dominant planetary dynamical parameters—planetary rotation rate and planetary radius—relative to their Earth values. Additionally, we propose a hypothesis for the general behavior of this system based on its two dominant dynamical length scales and apply its outcomes in our analysis. Overall, this work serves as the dynamical analog to Merlis et al. (2016), which analyzed the dependence of storm genesis on planetary thermodynamic forcing given by the sea surface temperature. Here we extend Merlis et al. (2016) in three key directions: 1) the dependence on planetary dynamical forcing, 2) analysis of storm size in addition to genesis, and 3) direct comparison to an Earthlike historical climate simulation.

The experimental design and analysis methodology are described in section 2. Theoretical background is presented in section 3. Results are presented for genesis (section 4) and size (section 5); for each, we first characterize its latitudinal variation and then test relevant hypotheses. Conclusions and discussion are provided in section 6.

2. Experimental methodology

a. Experimental model: Community Atmosphere Model, version 5.3

The Community Atmosphere Model, version 5.3 (CAM5), is used for the simulations performed for this work. CAM5, described in detail in Neale et al. (2012), is a comprehensive global atmosphere model that is the atmospheric component of the Community Earth System Model implemented for the Coupled Model Intercomparison Project, phase 5 (CMIP5; Taylor et al. 2012).

The main modification to CAM5 for this study is the use of a high-resolution horizontal grid spacing of ~ 25 km required for tropical cyclone-permitting scales in CAM5 (Reed and Jablonowski 2011; Wehner et al. 2014) compared to the standard CMIP5 grid spacing of ~ 100 km. Furthermore, the spectral element (SE) dynamical core option (Taylor and Fournier 2010; Dennis et al. 2012) in CAM5 is adopted as it utilizes a cubed-sphere grid that allows for quasi-uniform grid spacings throughout the global domain, which is ideal for studying tropical cyclones in our idealized experimental setup (Reed et al. 2012; Zarzycki et al. 2014; Reed and Chavas 2015). CAM5 has been shown to reproduce reasonable climatologies of tropical cyclone genesis and track (globally and regionally) in realistic decadal Atmospheric Model Intercomparison Project (AMIP; Gates et al. 1999) simulations (Zarzycki and Jablonowski 2014; Reed et al. 2015a; Bacmeister et al. 2018).

b. Experimental setup: Aquaplanet with uniform thermal forcing

We employ the same globally uniform thermal forcing aquaplanet model setup as Chavas et al. (2017), following the method used in Merlis et al. (2016). This setup developed out of nonrotating radiative–convective equilibrium experiments (Popke et al. 2013; Reed et al. 2015b; Arnold and Randall 2015) and has also been examined on a sphere with uniform Coriolis parameter (Reed and Chavas 2015). The sea surface temperature is forced to be 29°C everywhere with horizontally uniform, diurnally varying insolation set to produce a mean insolation of approximately 340 W m^{-2} similar to the observed global mean. We use this setup for experiments varying planetary rotation rate Ω and planetary radius a , as described in section 2d below.

c. Storm tracking

Storm tracking for all experiments is performed using the same algorithm and detection criteria as in Chavas et al. (2017). The open-source TempestExtremes tracking algorithm (Ullrich and Zarzycki 2017) detects candidate storms at 6-hourly intervals by searching for minima in surface pressure (taken to be the storm center) on the native cubed-sphere grid that are associated with a closed contour of 4 hPa within a distance of 556 km, that is, five great-circle degrees for an Earth-sized planet. Candidate storms are connected in time by searching within a distance of 556 km at the next time increment for another candidate storm to generate a track. For a storm track to be included in the analysis it must exist for at least four time increments (with a gap of

TABLE 1. List of aquaplanet experiments. Earth values are $\Omega_E = 7.292 \times 10^{-5}\text{ s}^{-1}$ and $a_E = 6371\text{ km}$.

Name	Ω	a	Resolution
CTRL	Ω_E	a_E	ne120 (25 km)
$2\Omega_E$	$2\Omega_E$	a_E	ne120 (25 km)
$0.5\Omega_E$	$0.5\Omega_E$	a_E	ne120 (25 km)
$0.25\Omega_E$	$0.25\Omega_E$	a_E	ne120 (25 km)
$0.5a_E$	Ω_E	$0.5a_E$	ne60 (25 km)
$2a_E$	Ω_E	$2a_E$	ne240 (25 km)

24 h between increments allowed). For our real-Earth historical simulation (described below), we use a separate storm tracker that additionally searches for an upper-level warm core as described in Zhao et al. (2009). Genesis is defined as the first point in the track. For all experiments, genesis events where the maximum near-surface azimuthal-mean azimuthal wind exceeds 20 m s^{-1} are discarded, as these are associated with storms at high latitudes where interstorm interaction is strong and a preexisting storm may be falsely identified as a new track by the track stitcher.

d. Experiments

A summary of our experiments are provided in Table 1. We define as our control experiment (CTRL) an aquaplanet simulation with uniform thermal forcing in which the planetary rotation rate and planetary radius are set to the standard Earth values following the Aquaplanet Experiment (APE; <http://climate.ncas.ac.uk/apec/design.html>) protocols; that is, $\Omega_E = 7.292 \times 10^{-5}\text{ s}^{-1}$ and $a_E = 6371\text{ km}$. From CTRL, two sets of experiments are performed:

- 1) Varying planetary rotation rate: $0.25\Omega_E$, $0.5\Omega_E$, $2\Omega_E$
- 2) Varying planetary radius: $0.5a_E$, $2a_E$

For varying a , the model resolution, including grid points and diffusion, is adjusted such that the true physical grid spacing is held constant (25 km) across all simulations. This choice minimizes the potential for resolution dependencies across simulations. Additionally, Reed and Chavas (2015) found minimal sensitivity in the qualitative behavior of the simulated RCE state in uniformly rotating global simulations using the same model. Each simulation is run for 2 years, and the first 6 months of data are discarded for spinup (the system equilibrates after approximately 2 months); the remaining 18 months yield a large number of cyclones sufficient for our analysis. We do not run a corresponding $0.25a_E$ experiment because the surface area of one hemisphere becomes comparable to the characteristic area of an individual storm.

FIG. 1. Snapshots of wind speed at the lowest model level for each experiment at simulation day 365: (a) $2\Omega_E$, (b) CTRL, (c) $0.5\Omega_E$, (d) $0.25\Omega_E$, (e) $2a_E$, and (f) $0.5a_E$. Black contour indicates 12 m s^{-1} .

Snapshots of near-surface wind speed for each experiment are displayed in Fig. 1, and maps of time-mean storm count density are displayed in Fig. 2. The atmosphere is dominated by tropical cyclones, which typically form at lower latitudes and subsequently propagate poleward and westward under the influence of beta drift (Chan 2005), eventually moving toward the poles where they may interact with other storms and eventually merge or dissipate. Moreover, in the absence of horizontal heterogeneity in boundary forcing (e.g., land) in these experiments, the spatial distribution of storm activity exhibits strong zonal and interhemispheric symmetry. This symmetry is retained as either Ω or a is varied. Thus, we focus our subsequent analysis of various storm quantities to be a function of absolute latitude, with both hemispheres combined.

Finally, to compare our idealized experiments with an Earthlike climate state, we also analyze an AMIP-style historical simulation (i.e., following Atmospheric Model

Intercomparison protocols; Gates et al. 1999) over the period 1979–2012; this exact setup was examined in previous work (Reed et al. 2015a; Bacmeister et al. 2018). Note that this AMIP simulation and an earlier version of CTRL were both employed in Chavas et al. (2017). The first year of the AMIP simulation (1979) is discarded.

3. Theoretical background

We next propose a hypothesis, first derived in Theiss (2004) in the context of quasigeostrophic (QG) ocean turbulence and applied to ocean observations by Eden (2007), for the behavior of our idealized aquaplanet atmosphere. On such a planet, which lacks externally forced horizontal thermodynamic variability, one expects the behavior of the system to be governed principally by relevant governing dynamical parameters. Specifically, two key dynamical length scales exist for this system.

FIG. 2. Spatial distribution of instantaneous storm-count density for each experiment. Data are binned into $5^\circ \times 5^\circ$ latitude–longitude boxes.

The first length scale is an inverse- f scale given by

$$L_f = \frac{U_f}{f}, \quad (1)$$

where $f = 2\Omega \sin\phi$ is the Coriolis parameter, ϕ is absolute latitude, and U_f is a velocity scale. The standard definition of this length scale is the Rossby deformation radius, representing the adjustment of an unbalanced continuously stratified fluid to rotation, for which this velocity is the gravity wave phase speed NH , where N is the Brunt–Väisälä frequency and H is the fluid depth. However, for the tropical cyclone the relevant velocity is given by the tropical cyclone potential intensity v_p (Emanuel 1986), which is a velocity scale derived strictly from local thermodynamic environmental parameters; the quantity v_p/f represents the “natural” tropical cyclone length scale (Emanuel 1995). This distinction has been demonstrated explicitly in tests of the length scales v_p/f and NH/f in axisymmetric tropical cyclone experiments (Chavas and Emanuel 2014). Thus, for generality we henceforth refer to this length scale using the term “inverse- f .”

The second length scale is an inverse- β scale, commonly referred to as the Rhines scale (Rhines 1975), given by

$$L_\beta = \frac{\pi}{2} \sqrt{\frac{U_\beta}{\beta}}, \quad (2)$$

where $\beta = df/dy = (2\Omega/a) \cos\phi$ is the meridional gradient of f , and U_β is a velocity scale. At low latitudes this quantity may also represent the equatorial deformation radius, which takes the same mathematical form. Here we include the factor $\pi/2$ in Eq. (2) to translate the Rhines scale from an eddy wavelength (with a factor of 2π) to a vortex radius, which in principal represents one-quarter of a wavelength. We note, though, that the inclusion of a scaling factor involving π varies across studies [e.g., Theiss (2004) does not include it]. The Rhines scale is associated with the nonlinear interaction of 2D turbulence with Rossby waves (Rhines 1975). This scale emerges directly from scale analysis of the quasigeostrophic vorticity equation on a β plane (Vallis 2017, p. 446), and it marks the transition from turbulence-dominated flow for length scales much smaller than L_β , for which the nonlinear advection term dominates, to Rossby wave-dominated flow for length scales much larger L_β , for which the β term dominates and the Rossby wave times are shorter than the eddy turnover times (Vallis and Maltrud 1993). Hence, the velocity scale in Eq. (2) is typically defined as a characteristic eddy velocity at the energy containing scales in the ambient flow.

Prior analyses have applied the Rhines scale to understand the dynamics of the jet stream and storm track, jet spacing on giant planets, and the scale of extratropical eddies (Frierson 2005; Frierson et al. 2006; Chemke and Kaspi 2015; O’Gorman and Schneider 2008) and thus define U_β using an RMS velocity at the latitude of maximum eddy kinetic energy (i.e., in the vicinity of the jet) or similar quantities. However, our model setup lacks the large-scale external baroclinic forcing for midlatitude jets.¹ Moreover, our eddies of interest are the isolated tropical cyclones themselves rather than ambient waves. Notably, a tropical cyclone may readily exist in the absence of a planetary vorticity gradient (e.g., Tang and Emanuel 2012), and its energetics are generally not fundamentally altered by its presence (Peng et al. 1999); this is perhaps an important distinction from prior work analyzing quasigeostrophic eddies generated from Rossby waves, whose existence depends on β . The tropical cyclone is more appropriately considered as an isolated vortex embedded within a flow with nonzero β .

Extensive fluid mechanics research has analyzed the dynamics of an isolated vortex on a β plane. The interaction of the vortex with its environment is known to induce translational motion (Llewellyn Smith 1997; Sutyrin and Flierl 1994), including for tropical cyclones (Chan and Williams 1987; Holland 1983; Smith et al. 1995). The dynamics of this motion is intimately tied to the radiation of Rossby waves by the vortex (Flór and Eames 2002; Sutyrin and Morel 1997; Reznik 2010; Zhang and Afanasyev 2015). Wave radiation transfers energy from vortex to environment and causes vortex decay (Flierl and Haines 1994; Sutyrin et al. 1994; Smith et al. 1995), which acts principally to limit the size of the vortex (McDonald 1998; Flór and Eames 2002; Lam and Dritschel 2001). Moreover, the dynamics and propagation of a vortex is more wavelike at larger size (Flór and Eames 2002), indicative of the wave–vortex transition associated with the Rhines scale. For the tropical cyclone, wind speed varies sharply with radius (i.e., length scale) within the storm, as does the circulation depth, and thus it is not obvious which velocity scale within the tropical cyclone is most relevant. The rapidly rotating inner core does not feel β as its rotational time scales are very fast (Lam and Dritschel 2001) and the flow is in approximate cyclostrophic balance (Holland 1980). Hence, this velocity scale seems most appropriately defined as a characteristic

¹ A weak easterly upper-level jet does emerge, similar to Merlis et al. (2016), because of a weak warming feedback from the cyclones to the mean state at high latitudes; this feedback also reduces v_p [see Fig. S1 and Cronin and Chavas (2019)].

flow velocity for the broad outer circulation of the cyclone.

For simplicity and analytical tractability, we set the velocity scales to be constants. We set $U_f = 70 \text{ m s}^{-1}$, which is the mean value of v_p at higher latitudes that is nearly constant across our simulations (see Fig. S1 in the online supplemental material) using the method of Bister and Emanuel (1998). We set $U_\beta = 10 \text{ m s}^{-1}$, which is a reasonable characteristic flow speed for the outer storm circulation. We note that the radial structure of the outer circulation takes a characteristic form that is relatively stable in time (Chavas et al. 2015) and covaries minimally with variations in inner-core intensity (Weatherford and Gray 1988; Chavas and Lin 2016); hence U_β would not be expected to scale with an inner-core velocity scale such as v_p . The qualitative results presented here are not sensitive to this value of U_β , with similar outcomes for a value of 5 m s^{-1} . Thus, 10 m s^{-1} should be considered reasonable; the definition of an optimal/correct precise value requires an in-depth study and accompanying theory, particularly given the inherent uncertainty in scaling constants. These velocity scales are otherwise expected to remain relatively constant in space and time given the uniform thermal forcing of the system. As described above, each of these length scales carry various caveats and assumptions in defining the precise magnitudes of the respective velocity scales, as well as uncertainty regarding scaling constants; we do not seek to resolve these issues here and instead opt to explore what we can explain using the simplest possible approach.

The dominant dynamical nondimensional parameter in the system is given by the ratio of these two length scales, that is, L_f/L_β . This ratio may be written as

$$\frac{L_f}{L_\beta} = \sqrt{\frac{U_f^2 \beta}{U_\beta^* f^2}} = \left(\frac{U_f^2}{2U_\beta^*} \right)^{1/2} \left(\frac{\cos \phi}{\sin^2 \phi} \right)^{1/2} (\Omega a)^{-1/2}, \quad (3)$$

where we define $U_\beta^* = (\pi/2)^2 U_\beta$ to absorb the π factor. Thus, this ratio depends on the planetary velocity-scale Ωa , which has been shown to be intrinsic to the primary dynamical nondimensional parameter in the primitive equations (Frierson 2005; Koll and Abbot 2015). These prior studies used the Buckingham Pi theorem to define their version of the parameter as the ratio of an inverse- Ω length scale (akin to a latitude-independent deformation radius) to the planetary radius. While both length scales are natural choices on dimensional grounds, they lack a direct connection to the dynamics of the atmosphere itself, particularly for the planetary radius. Moreover, these choices lack any dependence on latitude, which cannot be deduced solely

FIG. 3. Inverse- f length scale L_f (black solid), Rhines scale L_β (black dashed), and their ratio (blue) as a function of latitude for $\Omega = \Omega_E$ and $a = a_E$, with $U_f = 70 \text{ m s}^{-1}$ and $U_\beta = 10 \text{ m s}^{-1}$. Critical latitude ϕ_c [Eq. (7)] is highlighted (red).

from Buckingham Pi since such factors are themselves nondimensional. In our system, this parameter emerges as a ratio of two physical length scales amenable to interpretation. The resulting nondimensional parameter [Eq. (3)] yields an additional nondimensional factor that depends on latitude—it decreases monotonically moving poleward from infinity at the equator to zero at the pole, as shown in Fig. 3 for Ω_E and a_E .

As derived in Theiss (2004), equating these two length scales yields a single critical latitude ϕ_c that demarcates a transition between two dynamical regimes in which the smaller of the two length scales is the dominant one (Fig. 3): 1) L_β is dominant equatorward of ϕ_c (where $L_f/L_\beta > 1$), and 2) L_f is dominant poleward of ϕ_c (where $L_f/L_\beta < 1$). Setting $L_\beta^2 = L_f^2$ and substituting $\sin^2 \phi = 1 - \cos^2 \phi$ yields

$$\frac{U_f^2}{U_\beta^*} \cos \phi = 2\Omega a (1 - \cos^2 \phi). \quad (4)$$

Setting $x = \cos \phi$ gives an equation that is quadratic in x given by

$$x^2 + \frac{1}{\alpha} x - 1 = 0, \quad (5)$$

where

$$\alpha = \frac{2\Omega a}{U_f} \left(\frac{U_\beta^*}{U_f} \right) \quad (6)$$

represents the latitude-independent component of the governing dynamical nondimensional parameter given by Eq. (3). The physical solution of Eq. (5) for ϕ_c is

FIG. 4. (a) Dependence of critical latitude ϕ_c [Eq. (7)] on velocity scale Ωa . (b) Joint dependence of critical Coriolis parameter f_c [blue; Eq. (8)] and ϕ_c (red) on (a, Ω) . $U_f = 70 \text{ m s}^{-1}$ and $U_\beta = 10 \text{ m s}^{-1}$. Earth values are highlighted (marker).

$$\phi_c = \cos^{-1} \left[\frac{1}{2\alpha} (\sqrt{1 + 4\alpha^2} - 1) \right], \quad (7)$$

which is also marked in Fig. 3.

The dependence of ϕ_c on Ωa is displayed in Fig. 4a. Theoretically, ϕ_c separates two regimes: 1) an equatorial belt ($L_\beta < L_f$), where tropical cyclones strongly feel the Rhines scale and size is limited by Rossby wave radiation, and 2) a polar cap ($L_f < L_\beta$), where Rhines-scale effects are weak and cyclones may fill the domain with minimal wave effects.

Finally, we define the critical Coriolis as the value of f at ϕ_c , given by

$$f_c = 2\Omega \sin \phi_c. \quad (8)$$

The joint dependences of ϕ_c and f_c on (a, Ω) are displayed in Fig. 4b. While ϕ_c decreases monotonically with increasing Ω and a (Fig. 4a), f_c decreases with increasing a but increases with increasing Ω . Thus, f_c introduces an additional dependence specifically on Ω , thereby breaking the symmetry between Ω and a in the single velocity scale Ωa . The significance of this quantity will become apparent in the analysis below.

We will test the predictions of this hypothesis for explaining the behavior of the system across our experiments. We emphasize that here we focus on the interplay between the two proposed length scales and the extent to which they can explain the system behavior. We do not explicitly analyze the underlying physics nor derive a closed-form theory from first principles, which requires deeper analysis that is beyond the scope of this manuscript. However, detailed discussion of the physical implications of our results and its relevance to existing turbulence research is provided in section 6.

4. Results: Genesis

a. Quantitative description

Storm count and genesis statistics across all aquaplanet experiments, including AMIP, are displayed in Fig. 5, which follows the aesthetics of Merlis et al. (2016), their Fig. 2). Statistics include instantaneous storm count density N and annual genesis rate G as a function of absolute latitude, as well as global instantaneous storm count $\langle N \rangle$ and global annual genesis count $\langle G \rangle$. Both $\langle N \rangle$ and $\langle G \rangle$ are normalized to Earth's surface area ($A_E = 4\pi a_E^2$) to account for variability in planetary surface area associated with varying a . The value of $\langle N \rangle$ represents the average number of storms per unit Earth's surface area at any given moment in time, and $\langle G \rangle$ includes all genesis points equatorward of the local midlatitude minimum, which occurs in the range of 40° – 70° (Figs. 5c,f,i), to minimize significant uncertainties in tracker-identified genesis events in the high-latitude region where storms interact strongly.

1) CTRL SIMULATION AND COMPARISON WITH AMIP

We first discuss the CTRL simulation results and compare them to the AMIP historical simulation to place results in the context of a present-day Earthlike climate state.

CTRL yields a global annual genesis count of 537 yr^{-1} (Fig. 5a), which is significantly larger than AMIP (71 yr^{-1}) as well as the $\sim 90 \text{ yr}^{-1}$ in the historical record. In principle the real-Earth number should be inflated to account for land area and further account for the effects of the seasonal cycle, but we do not do this here, as this will not affect the conclusion. CTRL storm count density increases monotonically from equator to pole (Fig. 5b), with the sharpest increase in count density in midlatitudes

FIG. 5. Count and genesis statistics for (a)–(c) CTRL and AMIP, (d)–(f) varying Ω , and (g)–(i) varying a . (top) Global instantaneous count density $\langle N \rangle$ (stars) and global annual genesis density $\langle G \rangle$ (squares) per unit Earth's surface area; symbol size scales with planetary radius. (middle) Zonal-mean instantaneous count density N vs latitude. (bottom) Zonal-mean annual genesis density G vs latitude; markers denote maximum and minimum genesis points, and solid curves are used for calculation of $\langle G \rangle$. Data are binned into 5°-latitude intervals moving poleward from the equator, with both hemispheres combined. Control curves are highlighted (black outline).

at approximately 50°. A similar behavior also appears in Merlis et al. (2016, their Fig. 2b). CTRL genesis density increases monotonically from the equator to 30° and then decreases monotonically back toward near zero by 50° (Fig. 5c), similar to Merlis et al. (2016, their 301-K simulation) though with peak genesis shifted slightly poleward and with a slightly smaller magnitude (3.1 here vs approximately 4 in their study). The magnitude of peak genesis density is substantially larger for CTRL than AMIP and occurs much farther poleward than AMIP. Thus, the much larger total genesis count in CTRL depends principally on the wider poleward extent of genesis in our aquaplanet simulation. Clearly, in contrast to AMIP, storms in CTRL are capable of propagating toward the poles largely unimpeded, as the thermodynamic environment is uniformly favorable for their persistence by design.

2) AQUAPLANET EXPERIMENTS: VARYING ROTATION RATE AND PLANETARY RADIUS

For our aquaplanet system, as Ω is increased, global storm count increases rapidly, though slightly sublinearly, while global annual genesis count increases rapidly and slightly superlinearly (Fig. 5d). Count density increases monotonically at all latitudes (Fig. 5e). Genesis density also increases monotonically at all latitudes (Fig. 5f), with the minor exception of at 47.5° where genesis density itself is relatively small. The latitude of peak genesis density $\phi_{G,\max}$ shifts equatorward with increasing Ω .

As a is increased, global storm count per unit Earth's surface area varies weakly and nonmonotonically (Fig. 5g), with the largest value occurring for CTRL. In contrast, global annual genesis count per unit Earth's surface area decreases rapidly. Together this indicates

FIG. 6. (a) Zonal-mean annual genesis density vs f across all experiments (colored lines) and peak genesis density ($f_{G,\max}$, $G_{\max}$) (markers); curve fit to ($f_{G,\max}$, $G_{\max}$) for linear fit (black dashed) and power-law fit (gray dashed), with equations provided. (b) As in (a), but with f and G normalized by $f_{G,\max}$ and $G_{\max}$, respectively. (c) Comparison of simulated $f_{G,\max}$ vs theoretical f_c given by Eq. (8), with 1-to-1 line (black solid); crosses denote varying Ω and filled circles denote varying a (CTRL depicted with both), with circle size scaling with a .

longer-lived storms on average. Count density increases slightly and monotonically for $\phi < 45^\circ$ but decreases sharply and monotonically for $\phi > 65^\circ$ (Fig. 5h). The meridional distribution of genesis density, including $\phi_{G,\max}$, contracts equatorward (Fig. 5i), and the magnitude of peak genesis density steadily decreases.

b. Theoretical analysis

1) GENESIS RATE VERSUS LATITUDE

We now test the hypothesis that genesis rate depends fundamentally on f . Figure 6a maps genesis density versus latitude across all aquaplanet experiments (i.e., Figs. 5f,i) into f space. Genesis density curves approximately collapse to a single universal increasing function of f moving poleward from the equator up to some peak value of f , denoted $f_{G,\max}$. A linear fit to the data for ($f_{G,\max}$, $G_{\max}$) yields a constant rate of $0.72 (1000 \text{ km})^{-2} \text{ yr}^{-1} (10^{-5} \text{ s}^{-1})^{-1}$. Slight positive curvature is evident; indeed a zero-intercept power-law fit ($G = cf^\gamma$) performs slightly better, with $c = 0.16$ and exponent $\gamma = 1.57$, which is remarkably close to the $3/2$ -power-law dependence on η employed in the genesis potential index of Emanuel and Nolan (2004). Both fits are shown in Fig. 6a.

Notably, the profiles of G versus f take on similar triangular shapes, indicating rapid increase to a peak and then rapid decrease with increasing f . This suggests that perhaps these curves may be normalized by their respective maximum values, $f_{G,\max}$ and $G_{\max}$, as shown in Fig. 6b. Indeed, the curves do approximately collapse, particularly for varied a . For $f \leq f_{G,\max}$, the consistent quasi-linear increase in G noted in Fig. 6a is evident. For $f > f_{G,\max}$, G decreases with f more rapidly for smaller Ω and more slowly for larger Ω , suggesting an additional dependence on Ω not captured in the normalization.

Finally, the simplest hypothesis for what governs $f_{G,\max}$ is the critical Coriolis parameter f_c [Eq. (8)]. Figure 6b

displays a comparison of $f_{G,\max}$ and f_c ; indeed, the simulated values closely match the theoretical prediction. There is a slight upward curvature in the relationship; for $U_\beta = 5 \text{ m s}^{-1}$ this curvature disappears, though the relationship shifts rightward to be slightly offset from the one-to-one line such that f_c exceeds $f_{G,\max}$ by a constant of approximately 10^{-5} s^{-1} . Thus, genesis rate depends principally on f , though its meridional extent is set by the constraints of spherical geometry as manifest by f_c . Physically, poleward of the critical latitude, the Rhines scale becomes large and wave dynamics become increasingly weak, thereby favoring long-lived cyclones that fill the domain and thus reduce the available space for new genesis events to occur. Alternatively, at the vortex scale, the alignment of the natural tropical cyclone length scale and the Rhines scale might somehow be optimal for genesis. Notably, the role of f_c for genesis breaks the symmetry of varying Ω and a given by our hypothesis (Fig. 4b): increasing Ω reduces $\phi_{G,\max}$ (Fig. 5f) but increases $f_{G,\max}$ (Fig. 6a), whereas the two decrease in concert for increasing a (Figs. 5i and 6a). We will return to the potential significance of this distinction in section 5 below.

2) MINIMUM GENESIS DISTANCE FROM EQUATOR

We next analyze the minimum genesis latitude ϕ_{G0} . In the absence of significant relative vorticity, the hypothesis that genesis requires sufficiently large absolute vorticity suggests that this latitude is set by a minimum threshold value of f . Estimating ϕ_{G0} precisely is difficult via the binning methodology of the previous subsection. Instead, we calculate contours of minimum storm center absolute latitude as a function of longitude across all simulations (Fig. 7a), with both hemispheres combined together. We define ϕ_{G0} as the median of each contour, which increases for smaller Ω or a . The existence of a dependence of ϕ_{G0} on a indicates that f cannot be the explanatory variable and, further, an inverse- f scaling may be excluded.

FIG. 7. (a) Minimum storm-center absolute latitude (contours) and median latitude (marker on y axis) across all experiments. Data are binned into 5°-longitude intervals across both hemispheres. (b) Comparison of meridional distance from equator to median latitude L_{G0} against theoretical prediction given by equatorial Rhines scale $L_{\beta, EQ}$ given by Eq. (9); aesthetics are as in Fig. 6c.

We thus propose an alternative hypothesis: a minimum distance required to fit the majority of the incipient storm circulation on one side of the equator. A reasonable hypothesis for a governing length scale is the equatorial Rhines scale; that is,

$$L_{\beta, EQ} = \sqrt{\frac{U_{\beta}^*}{2\Omega/a}}. \quad (9)$$

This length scale is simply Eq. (2) evaluated at $\phi = 0$. Note that we cannot distinguish this scale from a traditional equatorial deformation radius, whose velocity scale differs only by a constant factor; both length scales represent viable bounds on storm size near the equator. Figure 7b compares $L_{\beta, EQ}$ against the meridional distance from the equator to ϕ_{G0} given by

$$L_{G0} = 2\pi a \left(\frac{\phi_{G0}}{360^\circ} \right). \quad (10)$$

Indeed, L_{G0} scales closely with $L_{\beta, EQ}$ across all aquaplanet experiments. Equating Eqs. (9) and (10) yields

$$\frac{\phi_{G0}}{90^\circ} = \sqrt{\frac{U_{\beta}}{2\Omega a}}. \quad (11)$$

Thus, ϕ_{G0} indeed increases for smaller Ω or a as was found in Fig. 7a. Notably, this quantity also depends solely on the velocity scale Ωa .

In AMIP, the median minimum latitude is found closer to the equator than would be predicted by this length scale (Fig. 7b). The AMIP minimum latitude curve is consistently equatorward of CTRL, particularly in the Indian Ocean and Maritime Continent (longitudes of 50°–120°). This difference is likely due to significant positive relative vorticity anomalies associated with large-scale atmospheric troughs (e.g., Yang and Wang 2018), an effect that is minimized in our aquaplanet setup.

Overall, these results indicate that the incipient storm circulation must largely fit within a region of like-signed absolute vorticity.

5. Results: Size

a. Quantitative description

We next examine storm size as a function of latitude, displayed in Fig. 8. We focus on the size of the overall storm cyclonic circulation, ideally given by the outer radius of vanishing wind r_0 (Chavas et al. 2015). To minimize noise, we analyze the radius of 2 m s^{-1} r_2 , as there are occasions where the wind profile smoothly approaches zero but then exhibits significant variability at very small wind speeds prior to attaining zero, perhaps because of natural background variability or proximity to adjacent storms.

1) CTRL SIMULATION AND COMPARISON WITH AMIP

In CTRL, median storm size in the lowest latitude bin ($[0^\circ, 5^\circ]$) is 1345 km (Fig. 8a), which is close to the minimum distance from the equator of 1266 km (Fig. 7b). Moving poleward from the equator, size first decreases to a minimum of 887 km at 17.5° before gradually increasing up to a peak of 1208 km at 47.5°. AMIP exhibits quantitatively similar behavior, with slightly larger storms between 10° and 30° such that size remains nearly constant within 10°–45°. Poleward of 47.5°, storm size decreases monotonically in CTRL, in contrast to AMIP where storm size increases rapidly, likely because of the role of extratropical transition associated with jet stream interaction. Thus, this experiment suggests that background environmental variability, including extratropical transition, is likely *not* fundamental to the variation of storm size with latitude found in nature within $\phi \in [10^\circ, 50^\circ]$. Why size first

FIG. 8. Zonal-median storm size r_2 (thick) and interquartile range (thin) vs latitude for (a) CTRL and AMIP, (b) varying Ω , and (c) varying a . Markers denote latitude of peak size. For $0.25\Omega_E$, r_2 reaches a maximum value of 4031 km at 82.5° .

decreases with latitude at very low latitudes is not clear; we speculate that this may be a transient adjustment period following genesis, though deeper analysis is warranted.

Our analysis moving forward focuses strictly on the variation of median size with latitude. However, storm size varies substantially within a given latitude bin at all latitudes in the CTRL simulation (Fig. 8a), as it does in AMIP and in nature (e.g., Merrill 1984; Chan and Chan 2015; Chavas et al. 2016). Detailed analysis of individual storms and interstorm variability will be examined in a future manuscript.

2) AQUAPLANET EXPERIMENTS: VARYING ROTATION RATE AND PLANETARY RADIUS

Size decreases monotonically with increasing Ω at all latitudes (Fig. 8b). For $0.5\Omega_E$ and $2\Omega_E$, the latitude of peak size $\phi_{r_2,\max}$ remains constant across rotation rates. For the slowest rotation rate ($0.25\Omega_E$), size does not attain a maximum at an intermediate latitude but rather continues to increase toward the pole. This very low rotation simulation produces very few storms at any given time (Fig. 1), which may have significant unknown implications for size dynamics, particularly at high latitudes where storm diameter becomes comparable to the length of a latitude circle and thus only one storm is permitted on geometric grounds alone.

Size increases monotonically and rapidly with increasing a at low latitudes (Fig. 8c), while at higher latitudes storm size varies nonmonotonically, with size remaining approximately constant between CTRL and $2a_E$ but increasing for $0.5a_E$. Perhaps more relevant, $\phi_{r_2,\max}$ shifts rapidly equatorward, from 62.5° for $0.5a_E$ to 27.5° for $2a_E$. Geometric constraints may become significant near the poles in the $0.5a_E$ simulation given that $\phi_{r_2,\max}$ shifts poleward and the surface area of the planet is substantially reduced. This suggests that the finding of constant size in the polar cap for CTRL and $2a_E$, for which the polar cap regime occupies a much larger range of latitudes, may be more credible.

b. Theoretical analysis

Following from the background of section 3, the simplest hypothesis is that storm size will follow the smaller of the two governing length scales, that is,

$$L_1 = \min[L_\beta, L_f], \quad (12)$$

and thus size should increase moving poleward from the equator up to the critical latitude ϕ_c and decrease thereafter. Indeed, the qualitative behavior of size in CTRL (Fig. 8a)—increasing at low latitudes and decreasing at high latitudes—compares well with this theoretical prediction. Moreover, theory predicts that size should scale with $\Omega^{-1/2}$ and $a^{1/2}$ in the equatorial belt and should scale with Ω^{-1} and be constant with a in the polar cap, which is also qualitatively apparent across our experiments (Figs. 8b,c).

We test this hypothesis quantitatively against simulated median size in Fig. 9. Equatorward of $\phi_{r_2,\max}$, broad variations in size for varying Ω (Fig. 9a) and a (Fig. 9b) are captured by the analytical prediction of Eq. (12), including increasing for smaller Ω and larger a at low latitudes and decreasing with Ω at high latitudes. At low latitudes, size approximately scales with $\Omega^{-1/2}$ and $a^{1/2}$, though size increases more rapidly with latitude than is predicted by L_β . Poleward of $\phi_{r_2,\max}$, size decreases much more slowly than Ω^{-1} , though it does remain approximately constant between CTRL and $2a_E$.

Clearly in Figs. 9a and 9b there are significant differences between the latitudes of peak size $\phi_{r_2,\max}$ and the critical latitude ϕ_c [Eq. (7)]. Direct comparison ϕ_c and $\phi_{r_2,\max}$ is displayed in Fig. 9c. For varying a , $\phi_{r_2,\max}$ scales reasonably well with ϕ_c , albeit with substantial offset such that $\phi_{r_2,\max} > \phi_c$; for $U_\beta = 5 \text{ m s}^{-1}$, ϕ_c is larger and the offset is partially reduced. This offset may indicate a lag in adjustment of storm size poleward of the transition latitude. In contrast, for varying Ω , $\phi_{r_2,\max}$ remains constant while theory predicts it should decrease with increasing Ω ; we return to this discrepancy below.

FIG. 9. (a),(b) Comparison of zonal-median storm size shown in Figs. 8b and 8c against theoretical predictions for the simple length scale (L_1 ; gray with dots), the length-scale fit to $\phi_{r_2,max}$ (L_2 ; black solid), and the weighted- L_2 length scale (L_3 ; black dotted). See text for details. (c) Comparison of simulated $\phi_{r_2,max}$ vs theoretical ϕ_c given by Eq. (7); aesthetics are as in Fig. 6c.

Curiously, though size poleward of $\phi_{r_2,max}$ does not appear to scale with Ω^{-1} across experiments, it does decrease with latitude as would be expected for an f^{-1} scaling. This suggests an alternative approach in which we take the transition latitude as given and test the theory accordingly; that is,

$$L_2 = \begin{cases} L_\beta, & \text{if } \phi \leq \phi_{r_2,max} \\ \frac{c}{f}, & \phi > \phi_{r_2,max} \end{cases}, \quad (13)$$

where c is simply the constant required to match the Rhines scaling at $\phi_{r_2,max}$. The prediction of Eq. (13) is displayed in Figs. 9a and 9b. This approach yields a reasonably good fit across all simulations, suggesting that storms may indeed feel the f^{-1} scaling poleward of $\phi_{r_2,max}$.

Finally, we explore one additional avenue to improve L_2 : a dimensionally consistent combination of the two L_2 scales; that is,

$$L_3 = \begin{cases} (L_{2,\beta})^{1+\alpha}(L_{2,f})^{-\alpha}, & \text{if } \phi \leq \phi_{r_2,max} \\ (L_{2,f})^{1+\alpha}(L_{2,\beta})^{-\alpha}, & \phi > \phi_{r_2,max} \end{cases}, \quad (14)$$

where α is a constant. This effectively modifies L_2 in each regime by an additional nondimensional factor associated with the ratio of the two L_2 scales. The prediction for $\alpha = 0.15$ is displayed in Fig. 9a,b. These length scales better represent the latitudinal variation of size within experiments, and also significantly improves the representation of the $0.25\Omega_E$ -size simulation. This final step admittedly is more of a fitting exercise that masks real physical processes, such as lagged responses of storm size, but at a minimum it may provide a basis for deeper analysis in future work. Note that none of these theories capture the poleward decrease in size at very low latitudes near the equator.

Why does $\phi_{r_2,max}$ scale with ϕ_c for variable a but not variable Ω ? As noted earlier, the dependence of peak

genesis rate specifically on f_c introduces a deviation from the theoretical dependence on Ωa for varying Ω but not a . Thus, for varying a , $\phi_{G,max}$ and ϕ_c neatly shift in concert (Fig. 5i), whereas when varying Ω , their relationship is transformed via f_c . The result is a departure from the simple scaling of size with ϕ_c and may perhaps also explain the deviations in genesis for $f > f_c$. If true, this suggests a role for internal feedbacks between genesis and size, though it is not obvious how to account for such complexities within the theoretical framework presented here.

The potential for interactions between genesis and size point toward an alternative hypothesis that we briefly explore here: the qualitative state of the system—that is, cyclone sparse versus cyclone packed—may be important for size dynamics. Indeed, storm behavior has been found to differ on an f plane with a single storm as compared to several storms (Zhou et al. 2014), suggesting that storm interaction may have significant effects. Storm count and size can be combined to estimate packing density ρ_{count} as a function of latitude across our experiments, given by

$$\rho_{count} = \frac{N(\pi r_2^2)}{A}, \quad (15)$$

where N is instantaneous storm count density (Figs. 5e,h), r_2 is zonal-median storm size (Figs. 8b,c), and A is the surface area of a given 5° latitude band (multiplied by 2 to account for both hemispheres). Note that this simple definition may yield packing density values that exceed unity in the presence of a small number of large storms whose diameter is much larger than the meridional bin width. Packing density as a function of latitude is shown in Figs. 10a and 10b. Moving poleward from the equator, packing density increases to a maximum value at some latitude $\phi_{\rho,max}$ and then decreases gradually toward the pole while retaining relatively high values. The decrease toward the pole is likely partially

FIG. 10. Zonal-mean packing density ρ_{count} vs latitude for (a) varying Ω and (b) varying a ; markers denote latitude of peak ρ . For $0.25\Omega_E$, ρ_{count} reaches a maximum value of 3.1 at 82.5° . (c) Comparison of $\phi_{p,\text{max}}$ and $\phi_{r_2,\text{max}}$ across all aquaplanet experiments; a small offset is included to avoid overlapping symbols. Marker aesthetics are as in Fig. 6c.

an artifact of tracker difficulties for weak and/or merging storms. Curiously, $\phi_{p,\text{max}}$ and $\phi_{r_2,\text{max}}$ are tightly correlated across all experiments, including variable Ω (Fig. 10c). One simple hypothesis is that the transition to the L_f regime is accelerated by entering a densely packed regime in which interstorm interaction is strong. The nature of interstorm interactions and its relevance to size dynamics are currently unknown, though; deeper analysis of this internal feedback lies beyond the scope of this work.

c. Relationship to the f plane

An additional, useful thought experiment is to consider our analytic predictions in the limit of an infinitely large planet ($a \rightarrow \infty$). In this limit, L_β is infinite, and is thus irrelevant, and the densely packed polar cap expands equatorward to cover the majority of the planet. Such behavior is readily visible in the transition from smaller to larger planet size (Figs. 1f,b,e and 10c); for the large planet, a sizable fraction of the planetary surface qualitatively resembles an f -plane simulation in which the domain is fully packed with storms. Moreover, on the f plane (and constant- f sphere; Reed and Chavas 2015) no proper genesis/lysis regions exist, as storms form and meander for long time. In our simulations, as a is increased, genesis count decreases rapidly relative to storm count (Fig. 5g), indicative of longer-lived storms, and genesis is increasingly confined to near the equator (Fig. 5i). Thus, our results appear consistent with the existing bed of f -plane research: f -plane-type dynamics may be generalized to the sphere for the polar cap regime where the Rhines scale is significantly larger than the inverse- f scale.

6. Conclusions and discussion

Here we employ aquaplanet experiments under uniform thermal forcing and variable global dynamical forcing, namely variations in planetary rotation rate and

planetary radius relative to Earth values, to test hypotheses regarding tropical cyclone genesis and size. Such atmospheres are dominated by tropical cyclones that form at low latitudes and propagate poleward, as is found in nature, yet are uninhibited from traveling to high latitudes and whose statistical properties are symmetric both zonally and hemispherically. Furthermore, we propose a hypothesis that the behavior of this system depends principally on the ratio of an inverse- f scale to the Rhines scale, whose intrinsic fundamental velocity is given by Ωa . This hypothesis predicts a critical latitude separating an equatorial belt where wave-cyclone interactions are strong and a cyclone-dominant polar cap where wave effects are weak and cyclones may freely evolve.

A schematic of our results is shown in Fig. 11. We summarize our findings in the context of the five research questions presented in the introduction:

- 1) In our control aquaplanet simulation: moving poleward from the equator, storm genesis rate rapidly increases from zero to a maximum and then rapidly decreases back to near zero prior to reaching the pole. Outer storm size decreases at very low latitudes, gradually increases to a maximum near 45° and then gradually decreases to the pole; the behavior of storm size below 45° mirrors that found in an Earth-like simulation despite the absence of land or jet interactions, including extratropical transition.
- 2) Genesis rate increases quasi linearly with f from near the equator to a maximum at the critical value of f and decreases back to zero thereafter. Genesis versus f , each normalized by their values at the critical value of f , collapse to an approximate universal dependence across experiments, with some deviation poleward of peak genesis for varying rotation rate. Genesis rates decrease poleward of the critical latitude where long-lived cyclones increasingly fill the domain.

FIG. 11. Summary schematic of results. See text for details.

- 3) The minimum genesis distance from the equator scales closely with the equatorial Rhines/deformation scale. This result suggests that, in the absence of large-scale relative vorticity, genesis requires that the incipient circulation largely fit on one side of the equator.
- 4) Outer storm size qualitatively follows the smaller of the two fundamental length scales: in the low-latitude regime, size scales reasonably well with the Rhines scale, indicating that the Rhines scale likely limits storm size; in the high-latitude regime, size varies with latitude following an inverse- f scaling relative to the transition latitude. The latitude of peak size is shifted significantly poleward of the critical latitude, suggesting that temporal effects may be significant. The critical latitude scales with the latitude of peak size for varying planetary radius though not for planetary rotation rate, the latter likely because of the dependence of peak genesis rate specifically on f , which breaks the system dependence on the combined quantity Ωa for variable Ω . The latitudes of peak size and peak packing density are closely correlated, suggesting interstorm interactions may be important for size dynamics.
- 5) Overall, our simulations produce equilibrium states characterized by a sparsely packed equatorial belt and a densely packed polar cap in line with the proposed hypothesis. As with size, the transition latitude scales with the critical latitude for varying planetary radius but not planetary rotation rate.
- 6) The large-planet limit predicts a planet nearly covered with long-lived storms, dynamically consistent with existing research for tropical cyclone worlds on an f plane.

What is the relationship between our results and quasigeostrophic turbulence theory? Curiously, the role

of the Rhines scale in limiting the size of isolated vortices, such as tropical cyclones, below their “natural” inverse- f length scale at low latitudes contrasts with its role in QG turbulence theory, where it acts as the cutoff for the upscale cascade of energy input at the deformation scale at high latitudes (Held and Larichev 1996; Jansen and Ferrari 2012; Chemke and Kaspi 2015, 2016; Chemke et al. 2016). Notably, QG turbulence research typically focuses on a dry fluid forced internally either barotropically (vorticity stirring) or baroclinically (baroclinically unstable shear profile), in contrast to the thermal forcing from surface heat fluxes in the study of radiative–convective equilibrium with or without rotation. The latter physics are a necessary condition for the existence of tropical cyclones (Emanuel 1986; Cronin and Chavas 2019) and thus such phenomena may simply not be permitted within traditional QG turbulence frameworks in the first place. Nonetheless, QG turbulence presumably still plays a role in setting the background eddy noise of our simulations. Thus, it seems plausible that these cyclones may form initially from turbulent eddies and, as a result, the genesis and initial characteristics of an individual cyclone may yet be intimately tied to background eddy energetics. Once mature, though, cyclone energetics may follow the traditional constant- f theory that has been well validated for the real Earth. Furthermore, our results appear to qualitatively align with that of Theiss (2004), which examined vortices generated by quasigeostrophic turbulence in a single-layer shallow-water fluid. This outcome suggests that a tropical cyclone in the presence of β behaves qualitatively like a simple barotropic vortex, as has been found for understanding tropical cyclone motion (Chan and Williams 1987). Finally, we note that waves induced by background turbulence might also modulate the large-scale statistical behavior of tropical

cyclones; indeed equatorial modes such as the Madden–Julian oscillation do exist in this simulation setup (Pritchard and Yang 2016; Arnold and Randall 2015) and are known to affect tropical cyclone activity on Earth (Schreck et al. 2012; Klotzbach and Oliver 2015; Camargo et al. 2007). Ultimately, a detailed accounting of background eddy energetics may yet yield deeper understanding of the role of turbulence in this system.

Otherwise, this analysis yields several key unanswered questions. First, what sets the meridional rate of increase (and decrease) of genesis rate with f ? This “natural” background genesis rate on a thermodynamically ideal planet for tropical cyclones (infinite ocean heat source, near-zero environmental wind shear) currently lacks any physical explanation, and appears to be strongly temperature dependent (Merlis et al. 2016); it could also vary across models. Similarly, why genesis should follow the critical latitude is not straightforward: at the system scale, increasing storm density in the polar cap regime may impose direct spatial constraints on genesis; at the vortex scale, the alignment of the Rhines scale and the natural tropical cyclone scale could perhaps be optimal for genesis. The latter might depend directly on the energetics of the background turbulent eddies from which cyclones emerge, and indeed past work has identified a similar nonmonotonic meridional variation in QG eddy kinetic energy injection rates in atmospheric reanalysis data (Chemke et al. 2016), albeit with a peak at relatively high latitudes. Understanding this background genesis rate may be an essential building block toward a theory for global genesis on Earth. Second, why does genesis more cleanly follow the theoretical prediction as compared to size? We speculate that genesis is a clearly defined event that occurs on fast time scales [$O(1)$ day; Emanuel 2011], whereas size may evolve slowly over the storm life cycle (e.g., Chavas and Emanuel 2014; Schenkel et al. 2018) and thus induces lags within the system. This time-scale distinction may similarly explain why genesis itself appears not directly modulated by the wave effects associated with the Rhines scale. Third, do the spatial constraints of spherical geometry modify storm behavior? Our results suggest that the effects of interstorm interaction may be significant, a process that is presumably enhanced at high latitudes by the reduction in surface area with latitude on a spherical planet. Fourth, what is the detailed dynamical response of a tropical cyclone vortex to the wave dynamics underlying the Rhines scale? Past work has focused on simplified barotropic vortices, whereas the tropical cyclone is baroclinic and conforms to a specific radial wind structure (Chavas et al. 2015). Finally, what sets the large variance in size at a given latitude?

Storm size varies markedly between storms in our simulations as it does in nature, suggesting that our simulations may be useful for understanding the behavior of individual storms as well, a topic that will be explored in future work.

Beyond these larger questions, we highlight a few additional aspects of our work that warrant further research. First, there is uncertainty in precisely defining the velocity scales, particularly for the Rhines scale; here we have chosen a simple and practical route but have no doubt that more detailed analyses could alter these definitions. Second, experiments extending beyond our Earth-centric range would be valuable tests of system behavior, particularly toward higher rotation rates and larger planets capable of sustaining a large number of storms; both require exponentially greater computer power to adequately resolve smaller storms (for the former) or to simulate a larger surface area at constant resolution (for the latter). Third, similar experiments on a β plane (e.g., Fedorov et al. 2019), where β is fixed, may help to isolate intrinsic temporal variability and would remove the spatial constraints imposed by spherical geometry. Fourth, our genesis dependence results may have direct relevance to the relationship between ITCZ latitude and genesis rate found in Merlis et al. (2013). Finally, our work cannot explain the substantial zonal variability of storm genesis and size in nature (Chavas et al. 2016), which likely depends on factors not accounted for in our zonally homogeneous world.

Overall, our analysis suggests that this thermodynamically homogeneous world offers a unique experimental testing ground for the behavior of tropical cyclones on a rotating planet in general, and whose results may provide a foundation for understanding their behavior and properties on Earth.

Acknowledgments. The authors thank Morgan O’Neill, Tim Merlis, and one anonymous reviewer for their detailed feedback, and thank Joe Harindra, Malte Jansen, Paul O’Gorman, Daniel Koll, Kerry Emanuel, Tim Cronin, and Tiffany Shaw for vibrant discussions that improved this manuscript. We would like to acknowledge high-performance computing support from Cheyenne (<https://doi.org/10.5065/D6RX99HX>) provided by NCAR’s Computational and Information Systems Laboratory, sponsored by the National Science Foundation, for all of the new simulations performed for this work. Access to the AMIP model output was provided by Julio Bacmeister, Susan Bates, and Nan Rosenbloom (NCAR). Reed was supported by the U.S. Department of Energy Office of Science Grants DE-SC0016994 and DE-SC0016605.

REFERENCES

- Anthes, R., 1982: *Tropical Cyclones: Their Evolution, Structure and Effects*. Meteor. Monogr., No. 41, Amer. Meteor. Soc., 208 pp.
- Arnold, N. P., and D. A. Randall, 2015: Global-scale convective aggregation: Implications for the Madden–Julian oscillation. *J. Adv. Model. Earth Syst.*, **7**, 1499–1518, <https://doi.org/10.1002/2015MS000498>.
- Bacmeister, J. T., K. A. Reed, C. Hannay, P. Lawrence, S. Bates, J. E. Truesdale, N. Rosenbloom, and M. Levy, 2018: Projected changes in tropical cyclone activity under future warming scenarios using a high-resolution climate model. *Climatic Change*, **146**, 547–560, <https://doi.org/10.1007/s10584-016-1750-x>.
- Bister, M., and K. A. Emanuel, 1998: Dissipative heating and hurricane intensity. *Meteor. Atmos. Phys.*, **65**, 233–240, <https://doi.org/10.1007/BF01030791>.
- Camargo, S. J., K. A. Emanuel, and A. H. Sobel, 2007: Use of a genesis potential index to diagnose ENSO effects on tropical cyclone genesis. *J. Climate*, **20**, 4819–4834, <https://doi.org/10.1175/JCLI4282.1>.
- , M. K. Tippett, A. H. Sobel, G. A. Vecchi, and M. Zhao, 2014: Testing the performance of tropical cyclone genesis indices in future climates using the HiRAM model. *J. Climate*, **27**, 9171–9196, <https://doi.org/10.1175/JCLI-D-13-00505.1>.
- Chan, J. C. L., 2005: The physics of tropical cyclone motion. *Annu. Rev. Fluid Mech.*, **37**, 99–128, <https://doi.org/10.1146/annurev.fluid.37.061903.175702>.
- , and R. Williams, 1987: Analytical and numerical studies of the beta-effect in tropical cyclone motion. Part I: Zero mean flow. *J. Atmos. Sci.*, **44**, 1257–1265, [https://doi.org/10.1175/1520-0469\(1987\)044<1257:AANSOT>2.0.CO;2](https://doi.org/10.1175/1520-0469(1987)044<1257:AANSOT>2.0.CO;2).
- Chan, K. T. F., and J. C. L. Chan, 2013: Angular momentum transports and synoptic flow patterns associated with tropical cyclone size change. *Mon. Wea. Rev.*, **141**, 3985–4007, <https://doi.org/10.1175/MWR-D-12-00204.1>.
- , and —, 2014: Impacts of initial vortex size and planetary vorticity on tropical cyclone size. *Quart. J. Roy. Meteor. Soc.*, **140**, 2235–2248, <https://doi.org/10.1002/qj.2292>.
- , and —, 2015: Global climatology of tropical cyclone size as inferred from QuikSCAT data. *Int. J. Climatol.*, **35**, 4843–4848, <https://doi.org/10.1002/joc.4307>.
- Chavas, D. R., and K. Emanuel, 2014: Equilibrium tropical cyclone size in an idealized state of axisymmetric radiative–convective equilibrium. *J. Atmos. Sci.*, **71**, 1663–1680, <https://doi.org/10.1175/JAS-D-13-0155.1>.
- , and N. Lin, 2016: A model for the complete radial structure of the tropical cyclone wind field. Part II: Wind field variability. *J. Atmos. Sci.*, **73**, 3093–3113, <https://doi.org/10.1175/JAS-D-15-0185.1>.
- , —, and K. Emanuel, 2015: A model for the complete radial structure of the tropical cyclone wind field. Part I: Comparison with observed structure. *J. Atmos. Sci.*, **72**, 3647–3662, <https://doi.org/10.1175/JAS-D-15-0014.1>.
- , —, W. Dong, and Y. Lin, 2016: Observed tropical cyclone size revisited. *J. Climate*, **29**, 2923–2939, <https://doi.org/10.1175/JCLI-D-15-0731.1>.
- , K. A. Reed, and J. A. Knaff, 2017: Physical understanding of the tropical cyclone wind–pressure relationship. *Nat. Commun.*, **8**, 1360, <https://doi.org/10.1038/s41467-017-01546-9>.
- Chemke, R., and Y. Kaspi, 2015: The latitudinal dependence of atmospheric jet scales and macroturbulent energy cascades. *J. Atmos. Sci.*, **72**, 3891–3907, <https://doi.org/10.1175/JAS-D-15-0007.1>.
- , and —, 2016: The latitudinal dependence of the oceanic barotropic eddy kinetic energy and macroturbulence energy transport. *Geophys. Res. Lett.*, **43**, 2723–2731, <https://doi.org/10.1002/2016GL067847>.
- , T. Dror, and Y. Kaspi, 2016: Barotropic kinetic energy and enstrophy transfers in the atmosphere. *Geophys. Res. Lett.*, **43**, 7725–7734, <https://doi.org/10.1002/2016GL070350>.
- Cronin, T. W., and D. R. Chavas, 2019: Dry and semidry tropical cyclones. *J. Atmos. Sci.*, **76**, 2193–2212, <https://doi.org/10.1175/JAS-D-18-0357.1>.
- Dennis, J., and Coauthors, 2012: CAM-SE: A scalable spectral element dynamical core for the Community Atmosphere Model. *Int. J. High Perform. Comput. Appl.*, **26**, 74–89, <https://doi.org/10.1177/1094342011428142>.
- Eden, C., 2007: Eddy length scales in the North Atlantic Ocean. *J. Geophys. Res.*, **112**, C06004, <https://doi.org/10.1029/2006JC003901>.
- Emanuel, K. A., 1986: An air–sea interaction theory for tropical cyclones. Part I: Steady-state maintenance. *J. Atmos. Sci.*, **43**, 585–605, [https://doi.org/10.1175/1520-0469\(1986\)043<0585:AASITF>2.0.CO;2](https://doi.org/10.1175/1520-0469(1986)043<0585:AASITF>2.0.CO;2).
- , 1995: The behavior of a simple hurricane model using a convective scheme based on subcloud-layer entropy equilibrium. *J. Atmos. Sci.*, **52**, 3960–3968, [https://doi.org/10.1175/1520-0469\(1995\)052<3960:TBOASH>2.0.CO;2](https://doi.org/10.1175/1520-0469(1995)052<3960:TBOASH>2.0.CO;2).
- , 2003: Tropical cyclones. *Annu. Rev. Earth Planet. Sci.*, **31**, 75–104, <https://doi.org/10.1146/annurev.earth.31.100901.141259>.
- , 2011: Global warming effects on U.S. hurricane damage. *Wea. Climate Soc.*, **3**, 261–268, <https://doi.org/10.1175/WCAS-D-11-00007.1>.
- , and D. S. Nolan, 2004: Tropical cyclone activity and the global climate system. *26th Conf. on Hurricanes and Tropical Meteorology*, Miami, FL, Amer. Meteor. Soc., 10A.2, https://ams.confex.com/ams/26HURR/techprogram/paper_75463.htm.
- Fedorov, A. V., L. Muir, W. R. Boos, and J. Studholme, 2019: Tropical cyclogenesis in warm climates simulated by a cloud-system resolving model. *Climate Dyn.*, **52**, 107–127, <https://doi.org/10.1007/s00382-018-4134-2>.
- Flierl, G., and K. Haines, 1994: The decay of modons due to Rossby wave radiation. *Phys. Fluids*, **6**, 3487–3497, <https://doi.org/10.1063/1.868405>.
- Flór, J.-B., and I. Eames, 2002: Dynamics of monopolar vortices on a topographic beta-plane. *J. Fluid Mech.*, **456**, 353–376, <https://doi.org/10.1017/S0022112001007728>.
- Frierson, D. M., 2005: Studies of the general circulation of the atmosphere with a simplified moist general circulation model. Ph.D. thesis, Princeton University, 218 pp.
- , I. M. Held, and P. Zurita-Gotor, 2006: A gray-radiation aquaplanet moist GCM. Part I: Static stability and eddy scale. *J. Atmos. Sci.*, **63**, 2548–2566, <https://doi.org/10.1175/JAS3753.1>.
- Gates, W. L., and Coauthors, 1999: An overview of the results of the Atmospheric Model Intercomparison Project (AMIP I). *Bull. Amer. Meteor. Soc.*, **80**, 29–55, [https://doi.org/10.1175/1520-0477\(1999\)080<0029:AOOTRO>2.0.CO;2](https://doi.org/10.1175/1520-0477(1999)080<0029:AOOTRO>2.0.CO;2).
- Gray, W. M., 1968: Global view of the origin of tropical disturbances and storms. *Mon. Wea. Rev.*, **96**, 669–700, [https://doi.org/10.1175/1520-0493\(1968\)096<0669:GVOTOO>2.0.CO;2](https://doi.org/10.1175/1520-0493(1968)096<0669:GVOTOO>2.0.CO;2).
- Hart, R. E., and J. L. Evans, 2001: A climatology of the extratropical transition of Atlantic tropical cyclones. *J. Climate*, **14**, 546–564, [https://doi.org/10.1175/1520-0442\(2001\)014<0546:ACOTET>2.0.CO;2](https://doi.org/10.1175/1520-0442(2001)014<0546:ACOTET>2.0.CO;2).
- , —, and C. Evans, 2006: Synoptic composites of the extratropical transition life cycle of North Atlantic tropical cyclones: Factors determining posttransition evolution. *Mon. Wea. Rev.*, **134**, 553–578, <https://doi.org/10.1175/MWR3082.1>.

- Held, I. M., and V. D. Larichev, 1996: A scaling theory for horizontally homogeneous, baroclinically unstable flow on a beta plane. *J. Atmos. Sci.*, **53**, 946–952, [https://doi.org/10.1175/1520-0469\(1996\)053<0946:ASTFHH>2.0.CO;2](https://doi.org/10.1175/1520-0469(1996)053<0946:ASTFHH>2.0.CO;2).
- Holland, G. J., 1980: An analytic model of the wind and pressure profiles in hurricanes. *Mon. Wea. Rev.*, **108**, 1212–1218, [https://doi.org/10.1175/1520-0493\(1980\)108<1212:AAMOTW>2.0.CO;2](https://doi.org/10.1175/1520-0493(1980)108<1212:AAMOTW>2.0.CO;2).
- , 1983: Tropical cyclone motion: Environmental interaction plus a beta effect. *J. Atmos. Sci.*, **40**, 328–342, [https://doi.org/10.1175/1520-0469\(1983\)040<0328:TCMEIP>2.0.CO;2](https://doi.org/10.1175/1520-0469(1983)040<0328:TCMEIP>2.0.CO;2).
- Jansen, M., and R. Ferrari, 2012: Macroturbulent equilibration in a thermally forced primitive equation system. *J. Atmos. Sci.*, **69**, 695–713, <https://doi.org/10.1175/JAS-D-11-041.1>.
- Khairoutdinov, M., and K. Emanuel, 2013: Rotating radiative-convective equilibrium simulated by a cloud-resolving model. *J. Adv. Model. Earth Syst.*, **5**, 816–825, <https://doi.org/10.1002/2013MS000253>.
- Klotzbach, P. J., and E. C. Oliver, 2015: Modulation of Atlantic basin tropical cyclone activity by the Madden–Julian oscillation (MJO) from 1905 to 2011. *J. Climate*, **28**, 204–217, <https://doi.org/10.1175/JCLI-D-14-00509.1>.
- Knaiff, J. A., S. P. Longmore, and D. A. Molnar, 2014: An objective satellite-based tropical cyclone size climatology. *J. Climate*, **27**, 455–476, <https://doi.org/10.1175/JCLI-D-13-00096.1>; Corrigendum, **28**, 8648–8651, <https://doi.org/10.1175/JCLI-D-15-0610.1>.
- Koll, D. D., and D. S. Abbot, 2015: Deciphering thermal phase curves of dry, tidally locked terrestrial planets. *Astrophys. J.*, **802**, 21, <https://doi.org/10.1088/0004-637X/802/1/21>.
- Kossin, J. P., J. A. Knaiff, H. I. Berger, D. C. Herndon, T. A. Cram, C. S. Velden, R. J. Murnane, and J. D. Hawkins, 2007: Estimating hurricane wind structure in the absence of aircraft reconnaissance. *Wea. Forecasting*, **22**, 89–101, <https://doi.org/10.1175/WAF985.1>.
- Lam, J. S.-L., and D. G. Dritschel, 2001: On the beta-drift of an initially circular vortex patch. *J. Fluid Mech.*, **436**, 107–129, <https://doi.org/10.1017/S0022112001003974>.
- Llewellyn Smith, S. G. L., 1997: The motion of a non-isolated vortex on the beta-plane. *J. Fluid Mech.*, **346**, 149–179, <https://doi.org/10.1017/S0022112097006290>.
- McDonald, N. R., 1998: The decay of cyclonic eddies by Rossby wave radiation. *J. Fluid Mech.*, **361**, 237–252, <https://doi.org/10.1017/S0022112098008696>.
- Merlis, T. M., M. Zhao, and I. M. Held, 2013: The sensitivity of hurricane frequency to ITCZ changes and radiatively forced warming in aquaplanet simulations. *Geophys. Res. Lett.*, **40**, 4109–4114, <https://doi.org/10.1002/grl.50680>.
- , W. Zhou, I. M. Held, and M. Zhao, 2016: Surface temperature dependence of tropical cyclone-permitting simulations in a spherical model with uniform thermal forcing. *Geophys. Res. Lett.*, **43**, 2859–2865, <https://doi.org/10.1002/2016GL067730>.
- Merrill, R. T., 1984: A comparison of large and small tropical cyclones. *Mon. Wea. Rev.*, **112**, 1408–1418, [https://doi.org/10.1175/1520-0493\(1984\)112<1408:ACOLAS>2.0.CO;2](https://doi.org/10.1175/1520-0493(1984)112<1408:ACOLAS>2.0.CO;2).
- Neale, R. B., and Coauthors, 2012: Description of the NCAR Community Atmosphere Model 1 (CAM 5.0). NCAR Tech. Note NCAR/TN-486+STR, 274 pp., www.cesm.ucar.edu/models/cesm1.0/cam/docs/description/cam5_desc.pdf.
- O’Gorman, P. A., and T. Schneider, 2008: Weather-layer dynamics of baroclinic eddies and multiple jets in an idealized general circulation model. *J. Atmos. Sci.*, **65**, 524–535, <https://doi.org/10.1175/2007JAS2280.1>.
- Peng, M. S., B.-F. Jeng, and R. Williams, 1999: A numerical study on tropical cyclone intensification. Part I: Beta effect and mean flow effect. *J. Atmos. Sci.*, **56**, 1404–1423, [https://doi.org/10.1175/1520-0469\(1999\)056<1404:ANSOTC>2.0.CO;2](https://doi.org/10.1175/1520-0469(1999)056<1404:ANSOTC>2.0.CO;2).
- Popke, D., B. Stevens, and A. Voigt, 2013: Climate and climate change in a radiative-convective equilibrium version of ECHAM6. *J. Adv. Model. Earth Syst.*, **5**, 1–14, <https://doi.org/10.1029/2012MS000191>.
- Pritchard, M. S., and D. Yang, 2016: Response of the superparameterized Madden–Julian oscillation to extreme climate and basic-state variation challenges a moisture mode view. *J. Climate*, **29**, 4995–5008, <https://doi.org/10.1175/JCLI-D-15-0790.1>.
- Reed, K. A., and C. Jablonowski, 2011: Assessing the uncertainty of tropical cyclone simulations in NCAR’s Community Atmosphere Model. *J. Adv. Model. Earth Syst.*, **3**, M08002, <https://doi.org/10.1029/2011MS000076>.
- , and D. R. Chavas, 2015: Uniformly rotating global radiative-convective equilibrium in the Community Atmosphere Model, version 5. *J. Adv. Model. Earth Syst.*, **7**, 1938–1955, <https://doi.org/10.1002/2015MS000519>.
- , C. Jablonowski, and M. A. Taylor, 2012: Tropical cyclones in the spectral element configuration of the Community Atmosphere Model. *Atmos. Sci. Lett.*, **13**, 303–310, <https://doi.org/10.1002/asl.399>.
- , J. T. Bacmeister, N. A. Rosenbloom, M. F. Wehner, S. C. Bates, P. H. Lauritzen, J. E. Truesdale, and C. Hannay, 2015a: Impact of the dynamical core on the direct simulation of tropical cyclones in a high-resolution global model. *Geophys. Res. Lett.*, **42**, 3603–3608, <https://doi.org/10.1002/2015GL063974>.
- , B. Medeiros, J. T. Bacmeister, and P. H. Lauritzen, 2015b: Global radiative–convective equilibrium in the Community Atmosphere Model 5. *J. Atmos. Sci.*, **72**, 2183–2197, <https://doi.org/10.1175/JAS-D-14-0268.1>.
- Reznik, G., 2010: Dynamics of localized vortices on the beta plane. *Izv. Atmos. Ocean. Phys.*, **46**, 784–797, <https://doi.org/10.1134/S0001433810060095>.
- Rhines, P. B., 1975: Waves and turbulence on a beta-plane. *J. Fluid Mech.*, **69**, 417–443, <https://doi.org/10.1017/S0022112075001504>.
- Schenkel, B. A., N. Lin, D. Chavas, G. A. Vecchi, M. Oppenheimer, and A. Brammer, 2018: Lifetime evolution of outer tropical cyclone size and structure as diagnosed from reanalysis and climate model data. *J. Climate*, **31**, 7985–8004, <https://doi.org/10.1175/JCLI-D-17-0630.1>.
- Schreck, C. J., III, J. Molinari, and A. Aiyyer, 2012: A global view of equatorial waves and tropical cyclogenesis. *Mon. Wea. Rev.*, **140**, 774–788, <https://doi.org/10.1175/MWR-D-11-00110.1>.
- Shi, X., and C. S. Bretherton, 2014: Large-scale character of an atmosphere in rotating radiative-convective equilibrium. *J. Adv. Model. Earth Syst.*, **6**, 616–629, <https://doi.org/10.1002/2014MS000342>.
- Smith, R. K., H. C. Weber, and A. Kraus, 1995: On the symmetric circulation of a moving hurricane. *Quart. J. Roy. Meteor. Soc.*, **121**, 945–952, <https://doi.org/10.1002/qj.49712152412>.
- , C. W. Schmidt, and M. T. Montgomery, 2011: An investigation of rotational influences on tropical-cyclone size and intensity. *Quart. J. Roy. Meteor. Soc.*, **137**, 1841–1855, <https://doi.org/10.1002/qj.862>.
- Sutyrin, G. G., and G. R. Flierl, 1994: Intense vortex motion on the beta plane: Development of the beta gyres. *J. Atmos. Sci.*, **51**, 773–790, [https://doi.org/10.1175/1520-0469\(1994\)051<0773:IVMOTB>2.0.CO;2](https://doi.org/10.1175/1520-0469(1994)051<0773:IVMOTB>2.0.CO;2).
- , and Y. G. Morel, 1997: Intense vortex motion in a stratified fluid on the beta-plane: An analytical theory and its validation. *J. Fluid Mech.*, **336**, 203–220, <https://doi.org/10.1017/S0022112096004685>.

- , J. S. Hesthaven, J.-P. Lynov, and J. J. Rasmussen, 1994: Dynamical properties of vortical structures on the beta-plane. *J. Fluid Mech.*, **268**, 103–131, <https://doi.org/10.1017/S002211209400128X>.
- Tang, B., and K. Emanuel, 2012: A ventilation index for tropical cyclones. *Bull. Amer. Meteor. Soc.*, **93**, 1901–1912, <https://doi.org/10.1175/BAMS-D-11-00165.1>.
- Taylor, K. E., R. J. Stouffer, and G. A. Meehl, 2012: An overview of CMIP5 and the experiment design. *Bull. Amer. Meteor. Soc.*, **93**, 485–498, <https://doi.org/10.1175/BAMS-D-11-00094.1>.
- Taylor, M. A., and A. Fournier, 2010: A compatible and conservative spectral element method on unstructured grids. *J. Comput. Phys.*, **229**, 5879–5895, <https://doi.org/10.1016/j.jcp.2010.04.008>.
- Theiss, J., 2004: Equatorward energy cascade, critical latitude, and the predominance of cyclonic vortices in geostrophic turbulence. *J. Phys. Oceanogr.*, **34**, 1663–1678, [https://doi.org/10.1175/1520-0485\(2004\)034<1663:EECLA>2.0.CO;2](https://doi.org/10.1175/1520-0485(2004)034<1663:EECLA>2.0.CO;2).
- Ullrich, P. A., and C. M. Zarzycki, 2017: TempestExtremes: A framework for scale-insensitive pointwise feature tracking on unstructured grids. *Geosci. Model Dev.*, **10**, 1069–1090, <https://doi.org/10.5194/gmd-10-1069-2017>.
- Vallis, G. K., 2017: *Atmospheric and Oceanic Fluid Dynamics*. Cambridge University Press, 946 pp.
- , and M. E. Maltrud, 1993: Generation of mean flows and jets on a beta plane and over topography. *J. Phys. Oceanogr.*, **23**, 1346–1362, [https://doi.org/10.1175/1520-0485\(1993\)023<1346:GOMFAJ>2.0.CO;2](https://doi.org/10.1175/1520-0485(1993)023<1346:GOMFAJ>2.0.CO;2).
- Weatherford, C., and W. Gray, 1988: Typhoon structure as revealed by aircraft reconnaissance. Part I: Data analysis and climatology. *Mon. Wea. Rev.*, **116**, 1032–1043, [https://doi.org/10.1175/1520-0493\(1988\)116<1032:TSARBA>2.0.CO;2](https://doi.org/10.1175/1520-0493(1988)116<1032:TSARBA>2.0.CO;2).
- Wehner, M. F., and Coauthors, 2014: The effect of horizontal resolution on simulation quality in the Community Atmospheric Model, CAM5.1. *J. Adv. Model. Earth Syst.*, **6**, 980–997, <https://doi.org/10.1002/2013MS000276>.
- Yang, H., and B. Wang, 2018: Multiscale processes in the genesis of a near-equatorial tropical cyclone during the Dynamics of the MJO experiment: Results from partial lateral forcing experiments. *J. Geophys. Res. Atmos.*, **123**, 5020–5037, <https://doi.org/10.1029/2017JD027650>.
- Zarzycki, C. M., and C. Jablonowski, 2014: A multidecadal simulation of Atlantic tropical cyclones using a variable-resolution global atmospheric general circulation model. *J. Adv. Model. Earth Syst.*, **6**, 805–828, <https://doi.org/10.1002/2014MS000352>.
- , M. N. Levy, C. Jablonowski, J. R. Overfelt, M. A. Taylor, and P. A. Ullrich, 2014: Aquaplanet experiments using CAM's variable-resolution dynamical core. *J. Climate*, **27**, 5481–5503, <https://doi.org/10.1175/JCLI-D-14-00004.1>.
- Zhang, Y., and Y. Afanasyev, 2015: Rossby wave radiation by an eddy on a beta-plane: Experiments with laboratory altimetry. *Phys. Fluids*, **27**, 076604, <https://doi.org/10.1063/1.4926979>.
- Zhao, M., I. M. Held, S.-J. Lin, and G. A. Vecchi, 2009: Simulations of global hurricane climatology, interannual variability, and response to global warming using a 50-km resolution GCM. *J. Climate*, **22**, 6653–6678, <https://doi.org/10.1175/2009JCLI3049.1>.
- Zhou, W., I. M. Held, and S. T. Garner, 2014: Parameter study of tropical cyclones in rotating radiative–convective equilibrium with column physics and resolution of a 25-km GCM. *J. Atmos. Sci.*, **71**, 1058–1069, <https://doi.org/10.1175/JAS-D-13-0190.1>.
- , —, and —, 2017: Tropical cyclones in rotating radiative–convective equilibrium with coupled SST. *J. Atmos. Sci.*, **74**, 879–892, <https://doi.org/10.1175/JAS-D-16-0195.1>.