The Formation of Moist Vortices and Tropical Cyclones in Idealized Simulations

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ABSTRACT

The upscale aggregation of convection is used to understand the emergence of rotating, coherent midtropospheric structures and the subsequent process of tropical cyclone formation. The Cloud Model, version 1 (CM1), is integrated on an $f$ plane with uniform sea surface temperature (SST) and prescribed uniform background flow. Deep convection is maintained by surface fluxes from an ocean with uniform surface temperature. Convection begins to organize simultaneously into moist and dry midtropospheric patches after 10 days. After 20 days, the patches begin to rotate on relatively small scales. Moist cyclonic vortices merge, eventually forming a single dominant vortex that subsequently forms a tropical cyclone on a realistic time scale of about 5 days. Radiation that interacts with clouds and water vapor aids in forming coherent rotating structures. Using the path to genesis provided by the aggregated solution, the relationship between thermodynamic changes within the vortex and changes in the character of convection prior to genesis is explored. Consistent with previous studies, the approach to saturation within the midtropospheric vortex accelerates the genesis process. A novel result is that, prior to genesis, downdrafts become widespread and somewhat stronger. The increased downdraft mass flux leads to stronger and larger surface cold pools. Shear–cold pool dynamics promote the organization of lower-tropospheric updrafts that spin up the surface vortex. It is inferred that the observed inconsistency between convective intensity and thermodynamic stabilization prior to genesis results from sampling limitations of the observations wherein the important cold pool gradients are unresolved.

1. Introduction

The ability of deep, moist convection to organize into scales tens, hundreds, or even thousands of times larger than a single convection cell is arguably one of the most fascinating, and perhaps still most poorly understood, behaviors in the tropical atmosphere. This organization is often manifested as convectively coupled synoptic-scale waves (Wheeler and Kiladis 1999). However, idealized models and observations have demonstrated the propensity of convection to self-organize in the complete absence of waves. Some organization of tropical convection occurs in moderately (or strongly) sheared environments in the form of quasi-linear clusters and squall lines (Houze 2004). Other organization occurs in environments with weak shear and aligns along the shear vector (Barnes and Sieckman 1984). Arguably the most coherent form of organization is a tropical cyclone, and these are clearly favored in environments of minimal vertical shear.

During the past two decades, there has been extensive investigation of the mechanisms by which convection self-organizes in the absence of shear (Held et al. 1993; Tompkins 2001; Bretherton et al. 2005; Tobin et al. 2012; Wing and Emanuel 2014). Most modeling studies have used convection-permitting models run with a grid spacing of 2–4 km and integrated over long time scales, generally 30–100 days. Without rotation, radiation–water vapor and radiation–cloud feedbacks operate in which radiative cooling drives subsidence (Muller and Held 2012; Emanuel et al. 2013). Most modeling studies have used convection-permitting models run with a grid spacing of 2–4 km and integrated over long time scales, generally 30–100 days. Without rotation, radiation–water vapor and radiation–cloud feedbacks operate in which radiative cooling drives subsidence (Muller and Held 2012; Emanuel et al. 2013). The subsidence, in turn, produces drying, which then accentuates the troposphere’s ability to cool, forcing more subsidence. This feedback transforms an initially statistically steady field of convection into what amounts to a stable attractor, a localized region of extreme moisture and convective activity hundreds of kilometers across surrounded by dry air with a near absence of convection.

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Aggregation has also been shown to produce one or more tropical cyclones when rotation is added (Bretherton et al. 2005; Nolan et al. 2007). Simulations of so-called “spontaneous genesis” have the attractive property that a minimum of parameters is needed to specify the state in which a tropical cyclone results. Even in simulations of spontaneous genesis, a coherent vortex in the midtroposphere precedes genesis by several days. We propose that such simulations offer a framework to understand aspects of the underlying processes that lead to an extreme example of the organization of convection: that is, a tropical cyclone (TC).

The present paper explores the aggregation process in an idealized model to understand the origin of rotating coherent structures that precede genesis and to help resolve a paradox that has emerged concerning the sequence of events that result in a tropical cyclone. It is widely accepted that tropical cyclone formation is preceded by a moist, mesoscale vortex in the midtroposphere (McBride and Zehr 1981; Simpson et al. 1997; Bister and Emanuel 1997; Dunkerton et al. 2009; Raymond et al. 2011). By moist, we mean that the relative humidity over a deep layer is at least 80% and perhaps closer to 90% (Nolan 2007). The thermal state prior to genesis features enhanced stratification, with significantly reduced buoyancy within the boundary layer (Davis and Ahijevych 2013; Komaromi 2013; Zawislak and Zipser 2014a). A portion of this stability enhancement follows from gradient thermal wind balance given a midtropospheric vortex (Raymond 2012). Calculations from observations reveal a near absence of buoyancy for parcels lifted reversibly in this environment. How, then, does the vigorous convection associated with genesis manage to occur? Furthermore, why is the moist midtropospheric vortex such a preferred structure across different ocean basins with different manifestations of tropical waves?

Idealized simulations of convective organization allow us to address the above questions in a way that differs from traditional studies of genesis, in which a vortex is imposed in the initial condition. However, we make no claim that these simulations portray the full sequence of events that produces a tropical cyclone in the real atmosphere. By focusing solely on the upscale organization of convection, we explicitly exclude the important role of tropical waves and their influence on convective organization. This allows us to see the intrinsic organizing properties of convection that operate even in the presence of large-scale structures. We envision that both synoptic-scale control and upscale organization from deep convection are important for producing a strong, moist, midtropospheric mesoscale vortex that is the hallmark of the precursor to genesis. Once the pre-TC vortex forms in the idealized model, we believe that the processes simulated henceforth are relevant to understanding genesis in the real atmosphere and, in particular, to answering the conundrum about the thermal structure and intensity of convection. Based on these simulations, we suggest that the paradox arises partly from limitations of observational sampling. Downdrafts from convection produce cold pools that, in the presence of moderate vertical shear, produce strong lower-tropospheric updrafts that preferentially spin up vorticity at low levels. All the action occurs in the highly transient gradient regions at the edges of cold pools that are largely unresolved by dropsondes and too shallow (and close to the surface) to be captured by remote sensing.

We should caution the reader that this paper is not intended to be a complete parameter study of convective aggregation in a rotating atmosphere. Furthermore, the fact that aggregation is relatively unpredictable in terms of its timing, or even whether it will occur at all, is not explored. Rather, we focus on the underlying mechanisms at work that produce organized structures: first, midtropospheric moist vortices (section 3), and then the comparatively rapid genesis process itself (section 4). In the conclusions (section 5), we attempt to reconcile some apparently contradictory views of the genesis process.

2. Methodology

The present study makes extensive use of the Cloud Model, version 1 [CM1, release 17 (R17)], an idealized cloud model described in Bryan and Fritsch (2002) and further in Bryan and Morrison (2012). The model is nonhydrostatic, is fully compressible, and is integrated on an $f$ plane with doubly periodic lateral boundary conditions. A uniform horizontal grid spacing of 3 km is used, along with 66 vertical levels. The lowest level is at 50 m above the surface. The lowest 10 levels are spaced 100 m apart; then the vertical grid is stretched uniformly to 500 m at 6 km. Above 6 km, levels are spaced 500 m apart to the top of the model at 25 km. The horizontal domain is 960 km $\times$ 960 km.

The CM1 is integrated with a sixth-order Runge–Kutta advection scheme using sixth-order diffusion with a nondimensional coefficient of 0.04 (Wicker and Skamarock 2002). The physical processes used are the Morrison double-moment microphysical scheme (Bryan and Morrison 2012; unaltered from its configuration in CM1 R17), the Yonsei University boundary layer scheme (Hong et al. 2006), and the Goddard shortwave and longwave radiation scheme (Chou and Suarez 1999). Rayleigh damping is applied above 15 km with an $e$-folding time of 5 min.
The model atmosphere is initialized with random perturbations with amplitude 0.25 K and has a uniform ocean beneath with a constant sea surface temperature (SST), here chosen to be 301 K. The initial sounding is taken from the radiative-convective equilibrium solution of Rotunno and Emanuel (1987). We experimented with a moistened initial sounding versus the original Rotunno and Emanuel sounding, but the initial details of the sounding are forgotten as the model establishes its own equilibrium state within the first 10–20 days. Two types of wind profiles are used: uniform flow and piecewise linear, where the wind at 700 m AGL is specified and returns to zero at a height of 3 km. The 700-m wind consired herein is westerly at 1 m s\(^{-1}\). The Coriolis parameter is set to 5 \times 10^{-5} \text{s}^{-1}, and the Coriolis force acts only on perturbations to the prescribed mean flow. A diurnal cycle is imposed for latitude 20°N (consistent with the Coriolis parameter), longitude 120°E, and a fixed date of 15 August.

Previous studies have pointed out that whether a given model produces aggregation depends on numerous subtleties (Muller and Held 2012). Schecter and Dunkerton (2009) noted that, at coarse resolution, complete bifurcations are possible for different, but spectrally identical, random initial perturbations. Owing to the sensitivities described in previous studies, it is not the purpose of the present study to perform a complete sensitivity analysis or predictability study of the behavior of aggregation in CM1.

3. Moisture patches and vortices

This section investigates the origin of aggregation in the simulation and how mesoscale rotation develops that eventually leads to a single midtropospheric mesoscale vortex. We focus on the base-state wind profile with an imposed 1 m s\(^{-1}\) surface wind that returns to zero at 3-km altitude. It turns out that the uniform flow also aggregates but requires an additional 20 days to do so. Both simulations achieve very similar end states featuring a tropical cyclone.

As shown in Fig. 1, organization of convection progresses slowly, but persistently, until a single large, moist, midtropospheric cyclonic potential vorticity anomaly results around day 50. From this point, tropical cyclone formation proceeds on a time scale of several days. By day 10 (Fig. 1a), one can already see evidence of organization of the convection. While only water vapor is shown, the imprint of ongoing or recent convective updrafts is clearly indicated by local maxima exceeding 3 g kg\(^{-1}\) at this level (level 29, or about 6 km). Small dry patches are evident, near which there is a discernable absence of updraft signatures in the water vapor field. The propensity of convection to cluster in moist patches is already evident.

By day 20 (Fig. 1b), there has been an overall drying within the domain as the atmosphere adjusts closer to an equilibrium state. The dry and moist patches are slightly larger, and a few contours of Rossby–Ertel potential vorticity (PV) appear, indicating anomalies of amplitude less than 0.25 PV units (PVU: 1 PVU = 10^{-8} \text{K kg}^{-1} \text{m}^2 \text{s}^{-1}) By day 30 (Fig. 1c), the patches are still larger, and we now see a clear association of positive PV anomalies with moist regions and negative anomalies with dry regions. Moist and dry regions have also taken on elongated structures.

By day 40 (Fig. 1d), the amplitude of dry and moist patches has grown. One moist patch has developed into a coherent vortex. While there is some indication that the dry patch has grown, overall there is not a clear upscale growth relative to day 30. By day 50 (Fig. 1e), there is a larger cyclonic vortex and a drier dry patch that now is almost completely devoid of convection. At this time, the merger of three cyclonic vortices is underway.

By day 60 (Fig. 1f), a tropical cyclone has formed. Simultaneously, the drying outside the tropical cyclone has become extreme. Convection is almost entirely suppressed outside the now-dominant vortex. Unlike simulations in the absence of rotation, the formation of dry and moist patches in the present simulation is simultaneous. Without rotation, Bretherton et al. (2005) and Wing and Emanuel (2014) showed that dry patches emerge first. We verified this behavior in CM1 for the particular set of physical parameterizations described in section 2 by setting the Coriolis parameter to zero.

It is well known that aggregation in quasi-equilibrium simulations depends on longwave radiational cooling associated with water vapor and cloud variations. To verify this in the present context, we ran a companion simulation identical to the control simulation but with radiation computed by Newtonian relaxation: that is, effectively a uniform cooling in time that has no knowledge of the inhomogeneities of cloud or water vapor. With Newtonian cooling, moist and dry patches still form, but there is no aggregation into a single cluster through 80 days, and no tropical cyclone results (Fig. 2). The lack of aggregation agrees with the results of Bretherton et al. (2005), who found that horizontally uniform radiation (or surface fluxes of heat and moisture) prevented aggregation. While 80 days might not be long enough to determine if aggregation will occur, there appears to be little progress toward such a state in the final 50 days of integration. A snapshot of water vapor of PV at 6 km at day 50 (Fig. 3) reveals moist and dry patches with some preference for the moist patches to be associated with positive PV anomalies, but the
relationship between PV and water vapor anomalies is less clear than in the control simulation.

To quantify the structure of moist and dry patches, we first smooth the water vapor field at 6-km altitude by replacing each grid point with a $21 \times 21$ point average centered on that point. Then we find maxima (minima) of smoothed water vapor that satisfy two conditions. First, the average water vapor mixing ratio must be 1.2 (0.8) times the domain average computed at the same level. Second, maxima (minima) must be separated from nearby maxima (minima) by at least 30 km. The analysis is applied to both the control and constant-radiation simulations.
between days 25 and 50. This period is relatively steady from the perspective of convection statistics (maximum updrafts and downdrafts), and the number of moist and dry water vapor patches present in each 6-hourly field is typically about five. Less than half of the full model domain is assigned to a patch at any particular time.

While it is well known that deep convection is favored in moist regions, the propensity of strong updrafts to concentrate in moist patches is remarkable (Fig. 4). Here, we have binned vertical velocity at 6 km by integer values of ln(w/0.01), with \( w \) in meters per second, and we have summed the magnitude of vertical velocity within each bin and divided by the number of patches, moist or dry. Thus, the ordinate represents the contribution from each bin to the vertical velocity averaged over all patches. The net vertical motion, averaged over moist and dry patches, is obtained by the sum of all positive bins minus the sum of all negative bins. For moist patches, the net upward motion is \( 4.36 - 3.43 = 0.93 \text{ cm s}^{-1} \), whereas, for dry patches, the net vertical motion is \( -0.16 \text{ cm s}^{-1} \). This rate of subsidence produces a downward displacement of \( 138 \text{ m day}^{-1} \). The resulting adiabatic warming of \( 1.36 \text{ °C} \) is consistent with the computed midtropospheric radiative cooling of about \( 1.2 - 1.3 \text{ °C day}^{-1} \) (see below).

The distributions shown in Fig. 4 reveal that deep convection is strongly favored in moist patches and suppressed in dry patches. These results are consistent with Held et al. (1993), James and Markowski (2010), and Kilroy and Smith (2013). The difference in net upward motion is attributed to strong updrafts in moist patches. The last two bins, representing the strongest
updrafts, collectively contribute more than 1 cm s$^{-1}$ of upward motion to the patch-mean value. This is roughly the difference of net upward motion between moist and dry patches.

Composite vertical profiles of vorticity and divergence within moist and dry patches clearly show that moist patches rotate cyclonically, and dry patches anticyclonically, with a maximum (or minimum) vorticity near 6 km (Fig. 5). Also apparent is the reversal of the vorticity around 10 km for both moist and dry patches. The negative relative vorticity atop moist patches results from the divergent outflow from convective towers. Convergence in the midtroposphere is rather weak when examined with 6-hourly output and does not explain the vorticity. The source of vorticity will be clarified below. We also note that the average midtropospheric vorticity in moist patches in the control simulation is nearly 4 times greater than the vorticity in moist patches in the simulation with Newtonian cooling.

Initially, divergence dominates vorticity within the updraft. However, as the updraft wanes and becomes a weak downdraft, a period of convergence begins that spins up vorticity. Once created, the vorticity far outlasts the divergent motions. This result was also reported by Fang and Zhang (2011), and Wissmeier and Smith (2011). The growth of snow at this level, barely above the freezing level, coincides with the switch from updraft to weak downdraft and signifies horizontal convergence within a small stratiform region that follows the dissipation of the updraft in the midtroposphere. It is here that the vorticity is generated between continued weak ascent in the upper troposphere and the downdraft fueled by the melting of snow and evaporation of rain below.
The thesis we advance based on the foregoing analysis is that the formation of coherent moist vortices is driven by two factors. First is the positive feedback, wherein moist regions favor convection and convection further moistens where it occurs. Second, the vortical remnants produced by the convection accumulate where updrafts are more frequent. Eventually, the rotation is strong enough that coherent vortices emerge, as evidenced by the formation of closed PV contours. These structures help isolate the moisture dynamically from its surroundings (Dunkerton et al. 2009).

There are other factors. Examination of Fig. 1 shows clear evidence of vortex merger (for instance, around day 50). Merger contributes to the process of aggregation into a single dominant vortex. We should point out, however, that rotation actually delays the aggregation of convection overall in this particular model configuration. A simulation identical to the control, but with $f = 0$, consolidated into a single convective cluster in only 40 days. One possibility is that the moisture-isolating effect of coherent vortices inhibits the consolidation of convection. As is clear from Fig. 1, the simulation with rotation spends tens of days in a state where numerous small vortices become progressively stronger but remain relatively inert.

Finally, we have not considered explicitly the effect of radiation on the organizational structure. As seen in Fig. 5, the cyclonic vorticity in the simulation with simple radiation is weaker than in the simulation with full radiation. One possibility is that radiation enhances cyclonic PV anomalies directly. Given the diabatic heating from the radiative tendency, expressed as $\dot{q}$, and using the relation $\dot{q} = (1/\rho)\boldsymbol{\eta} \cdot \nabla \theta$, where the overdot denotes the Lagrangian derivative, $\rho$ is the density, and $\boldsymbol{\eta}$ is the absolute vorticity vector, we compute the Lagrangian rate of change of PV within moist patches. However, this turns out to be either small or negative in the middle troposphere. The radiative heating profiles in moist and dry patches (Fig. 7) indicate why. There is little vertical gradient of heating in the middle troposphere. Horizontal gradients of heating and their coupling with horizontal vorticity are also negligible in this context. The primary vertical gradient of heating is in the upper troposphere, and, indeed, this gradient may explain the positive PV (and vorticity) anomaly in the upper troposphere above dry patches. In moist patches, this positive PV tendency is overwhelmed by the upper-tropospheric divergence associated with convection.

The effect of radiation on rotation within patches is somewhat indirect. Apparent in Fig. 7 is a difference in heating rate between moist and dry patches. The trapping of longwave radiation by clouds (and enhanced water vapor) in moist patches reduces the rate of cooling. Because moist and dry patches are typically in close proximity, we consider the difference in heating to represent a horizontal gradient. A horizontal gradient of heating, if maintained, is consistent with a vertical, secondary circulation that can be described by the Sawyer–Eliassen (S–E) equation (Sawyer 1956; Eliassen 1962). If we idealize the situation as a moist patch of radius 30 km surrounded by a dry ring, we can use the S–E equation to compute the upward motion of the secondary circulation.

Using the S–E equation as formulated in Hendricks et al. (2004) and assuming a Rankine vortex radial profile with the vorticity as shown in Fig. 5 (thick line) within the region of solid body rotation and the Brunt–Väisälä frequency given by the domain-average profile of potential temperature, we obtain the vertical motion profile shown in Fig. 7. Here, the radial gradient of heating is assumed uniform between $r = 15$ km
and $r = 45 \text{ km}$ and specified by the difference in heating rate between moist and dry profiles shown in Fig. 7. The heating gradient below 2 km is set to zero. The upward motion in the middle troposphere is a modest $0.1–0.2 \text{ cm s}^{-1}$. Nonetheless, this circulation can be maintained for many days, with a vertical displacement of more than 100 m per day, and is strong enough to cancel the effects of radiative cooling. Bretherton et al. (2005) and Jeevanjee and Romps (2013) inferred qualitatively similar vertical circulations, although theirs were inherently unbalanced. We hypothesize that the continued moistening through the secondary circulation that arises from differential heating is sufficient to make moist patches more moist (relative to dry patches) and further bias the location of convection toward moist patches. This is an indirect effect that accentuates the preference of convection for moist patches and further induces cyclonic vorticity production from convective updrafts (and their attendant stratiform precipitation). This enhancement is absent in the simulation with Newtonian cooling.

4. Genesis and the role of downdrafts

In this section, we focus on the genesis process within the final 6.5 days of the control simulation (days 53.5–60). At the end of this period, a TC exists within the domain with wind speeds approaching hurricane strength (Fig. 8a). This 6.5-day period represents the time during which a midtropospheric vortex, the precursor to the TC, can be continuously tracked. The track of the vortex is defined by the maximum vorticity at 6 km averaged within a box $150 \text{ km} \times 150 \text{ km}$. A time series of hourly values of average vorticity (Fig. 8b) shows that the vortex steadily intensifies at 6 km, then suddenly begins intensification at 2 km around 96 h, or about 2 days prior to genesis. The exact timing of genesis, defined operationally as a sustained 10-m wind exceeding $17 \text{ m s}^{-1}$, is somewhat uncertain. Herein we denote 144 h as the time of genesis based on (i) the persistence of maximum 10-m wind speed above $17 \text{ m s}^{-1}$; (ii) a transition in the organization of upward vertical velocity; and (iii) the beginning of erosion of the cold core vortex structure. These last two factors will be discussed later. While the full simulation requires 60 days to produce a TC, the genesis process, defined as the time over which the midtropospheric vortex transitions into a tropical cyclone, occurs within a realistic period of not more than 2–3 days.

As noted in numerous previous studies, the approach to saturation is an important step on the path to TC formation (e.g., Nolan 2007). The time–height depiction of relative humidity (Fig. 9) indicates that the vortex moistens continuously during the first 4 days of the 6.5-day period. By 74 h, the entire column from the surface to 7-km altitude attains a relative humidity with respect to water at or above 80%.

Within the subdomain that defines the midtropospheric vortex ($150 \text{ km} \times 150 \text{ km}$), 74 h marks the approximate time at which a notable increase in the vertical mass flux occurs at 6 km (Fig. 10a). While the maximum vertical velocity at 6 km shows no clear trend during this period (not shown), the increase in the number of updrafts at 6 km (grid points where $w \geq 0.5 \text{ m s}^{-1}$) indicates that convection becomes more organized (Fig. 10b). In section 3, moistening was found to
spatially congregate updrafts, and this appears to be happening here as well. Downdrafts at 6 km are weaker and far less frequent than updrafts (not shown).

The mass flux at 2 km does not become generally positive until roughly 108 h, more than one day after it becomes positive at 6 km (Fig. 10a) and about 1.5 days before genesis. Unlike at 6 km, both the strength and the number of updrafts increase at 2 km (Figs. 10b, c). Updraft maxima increase from a typical value of 4 m s\(^{-1}\) prior to hour 48 to roughly 6–7 m s\(^{-1}\) (Fig. 10c) after hour 108. More significant is the increase in the number of updrafts at 2 km, defined as grid points with vertical velocity exceeding 0.5 m s\(^{-1}\) (Fig. 10b). The number of updrafts increases by roughly fivefold from an average of roughly 25 prior to 72 h to about 150 near the time of genesis (144 h). Thus, while convection becomes stronger at 2 km, it is the area covered by updrafts that dominates the increase of the area-mean upward mass flux before genesis. This result echoes that of recent studies (e.g., Zawislak and Zipser 2014b) that suggest it is the area coverage of convection, rather than the intensity of convection, that affects the genesis of tropical cyclones.

Downdrafts at 2 km, defined to be greater than 0.5 m s\(^{-1}\), behave somewhat similarly to updrafts at that level. While downdrafts become somewhat stronger prior to genesis (Fig. 10c), the number of downdrafts increases markedly. Figure 10b shows the ratio of the number of updrafts to downdrafts at 2 km. Given that the number of updrafts increases markedly with time and the fact that the ratio becomes nearly constant at 1.5, one can infer that the number of downdrafts increases markedly with time as well. However, updrafts systematically outnumber downdrafts as genesis approaches. This is consistent with the increase of the vertical mass flux at 2 km prior to genesis.

An investigation of thermodynamic variables following the midtropospheric vortex provides some context for the statistics of vertical velocity. Deviations of potential temperature from the domain average (Fig. 11a) show persistent warming above the midtropospheric vortex at 9-km altitude and cooling below at 2 km. These trends mainly arise from thermal wind balance associated with the intensification of the midtropospheric vortex.
vortex. The thermal wind relation is $I(\partial v/\partial p) = - (\partial \theta/\partial r)$, where $I$ is the inertial frequency of the vortex, $\pi$ is the Exner function, $v$ is the tangential wind, and $\theta$ is the potential temperature anomaly. Approximating with finite differences and using $\delta \pi = 0.2 \theta$, (representing the Exner function difference between 500 and 200 hPa), $\delta r = 10^5$ m; and using parameters valid at 96 h, $\delta v = 8$ m s$^{-1}$ (from model output), and $I = 2 \times 10^{-4}$ s$^{-1}$, we obtain $\delta \theta = 0.8$ K. This compares favorably to the 1-K temperature anomaly computed from the model at 96 h (Fig. 11a). Further support for the balanced nature of tropical midtropospheric vortices appears in Raymond (2012).

However, Fig. 11a also indicates a cold anomaly at the surface that oscillates and becomes stronger prior to genesis. The air–sea temperature difference for most of the early simulation and over the entire domain is roughly 1.5°C. This is consistent with the time series through 72 h. A weak diurnal cycle is evident despite the constant SST. The lowest model level cools slightly at night, and warms slightly during the day (hour 0 = 1200 UTC, or approximately sunset at 120°E, the longitude arbitrarily chosen for the model domain). After 72 h, a downward trend is apparent, with the air–sea temperature difference averaging about −2.5°C between 96 and 120 h. The air–sea temperature difference recovers after genesis occurs as surface winds (and enthalpy fluxes, not shown) increase. The spatial structure of the near-surface cooling will be explored further below.

Consistent with the low-level cooling and uppertropospheric warming, the convective available potential energy (CAPE) also decreases (Fig. 11b). Here, we have computed CAPE by reversibly lifting parcels from

![Fig. 9. Vortex-following time–height cross section of area-mean relative humidity with respect to water saturation. The area is a 150 km × 150 km box centered on the vortex at 6 km. Time zero corresponds to day 53.5 of the control simulation. The vertical dashed line at hour 74 indicates when a relative humidity of 0.8 has been reached throughout the lower and middle troposphere.]
FIG. 10. Time series of (a) vertical mass flux at 6-km (black) and 2-km (red) altitude integrated over a 150 km × 150 km box centered on the vortex at 6 km; (b) number of updrafts exceeding 0.5 m s\(^{-1}\) at 2 km (red) and 6 km (black) within a 150 km × 150 km box centered on the 6-km vortex and ratio of updrafts to downdrafts (each exceeding 0.5 m s\(^{-1}\); blue line). The updraft–downdraft ratio has been computed using a moving 10-h summation of updrafts and downdrafts. The ratio has been multiplied by 100; and (c) maximum and minimum vertical velocity (m s\(^{-1}\)) at 2 km within the vortex-centered 150 km × 150 km box.
50-m altitude, subject to an entrainment rate of 0.05 km$^{-1}$. While this value of entrainment might be smaller than occurs in reality (Romps and Kuang 2010), deep convection on a 3-km grid entrains less than it does when turbulence is resolved (Bryan and Morrison 2012). The statistics displayed are computed by lifting a parcel from every point within a 150 km$^2$ domain centered on the vortex. The median CAPE is only about 100 J kg$^{-1}$ and decreases to about 60 J kg$^{-1}$ by 136 h. The maximum CAPE drops by about the same proportion. Near the time of genesis, as the air–sea disequilibrium decreases, CAPE values recover slightly, but the maximum CAPE is still well below what it was earlier in the simulation. Of note is that, despite the decrease in CAPE prior to genesis, the buoyancy at 3 km shows little change. The maintenance of lower-tropospheric buoyancy arises from cooling of the layer to which parcels are lifted and is a consequence of thermal wind balance given an intensifying midtropospheric vortex.

Thermodynamic stabilization prior to genesis is well documented from observations (Bister and Emanuel 1997; Raymond et al. 2011; Davis and Ahijevych 2013; Zawislak and Zipser 2014a; Gjorgjievska and Raymond 2014). Yet, as stabilization maximizes, the simulated area-averaged upward mass flux becomes consistently positive at 2 km. The upward mass flux increases foremost because of the dramatic increase in the number of updrafts and secondarily to the increase of lower-tropospheric updraft strength.

The increase of downdraft mass flux (inferred from Fig. 10b) also coincides with the development of stronger and more extensive cold pools at the surface (Fig. 12). At hour 60, prior to the deep-layer moistening within the vortex, updrafts are widely separated. Cold
pools, while locally strong, are also small in horizontal extent and widely separated (Fig. 12a). A snapshot at 108 h, after the deep-layer humidity within the vortex exceeds 80% (Fig. 9) and the midtropospheric mass flux increases (Fig. 10b), shows more spatially extensive cold pools, with low-level updrafts on the flanks of the cold pools (Fig. 12b). Note that the horizontal scale of the cold pools at this time is still small compared to the broad negative temperature anomaly at 2 km that is associated with the midtropospheric vortex (Fig. 12c). By 132 h (roughly 12 h prior to genesis), the updrafts have organized along the downshear flank of the cold pool (Fig. 12d).

A cross section through the cold pool of Fig. 12d reveals that it is a shallow feature perhaps 500 m deep (Fig. 13). This represents cold pools at other times. Nonetheless, the updraft is preferentially organized on the downshear side in accord with the general concept of...
Rotunno et al. (1988). In this situation, the cold pools and vertical shear are effective for increasing the low-level upward mass flux by helping to organize the low-level updrafts.

The preference for updrafts to occur on the downshear flank of cold pools is representative of the day leading up to genesis. In Fig. 14 is shown a spatial plot of the count of updrafts at 2 km exceeding $0.5 \text{ m s}^{-1}$ during the period 122–145 h. While the time-averaged temperature anomaly reveals a rather broad negative perturbation from the environment, the updrafts cluster on the downshear gradient of this temperature anomaly. Viewed as snapshots (Fig. 12), sharp gradients in surface temperature are consistently present, but the gradients move with time and are rapidly eroded by surface fluxes (Nolan and McGauley 2012). Convective cells are instantaneously able to tap boundary layer air that is relatively unstable (Montgomery et al. 2006; Creighton et al. 2013). But these convective structures feature relatively strong low-level updrafts owing to the reduced CAPE, maintenance of lower-tropospheric buoyancy, and dynamic lifting induced at the downshear edge of cold pools. Furthermore, the shear–cold pool relationship produces lower-tropospheric updrafts that are (i) more spatially extensive and (ii) moderately stronger approaching the time of genesis, compared with earlier times, such that the sign of the vertical mass flux becomes positive over the scale of the vortex. Once this happens, vorticity in the boundary layer rapidly increases.

The vertical shear in the lower troposphere arises primarily from the structure of the pregenesis vortex itself. Initially the shear is consistent with a cold core vortex: cyclonic winds increase with height in the lower troposphere (Fig. 12b). However, the increase of lower-tropospheric circulation is highly asymmetric, meaning that the lower-tropospheric circulation spins up initially away from the center of the midtropospheric vortex. Note that Fig. 12b shows the strongest convection occurs away from the center of the midtropospheric vortex. This convection, with organized lower-tropospheric updrafts, forms a lower-tropospheric cyclonic circulation. As this low-level vortex becomes stronger, the tilt between this low-level vortex and the midtropospheric vortex influences the vertical shear (Fig. 12d). We suggest that self-induced vertical shear resulting from vortex tilt could invigorate low-level updrafts (Davis and Bosart 2006).

To investigate the robustness of the foregoing scenario, a different simulation was integrated with a uniform
wind of 1 m s$^{-1}$ through the depth of the model domain. Genesis occurred at roughly day 78.5 (by the criteria described at the beginning of this section). A snapshot at day 78, analogous to Fig. 12d, reveals that the vortex tilt is greater than 100 km (Fig. 15). In both simulations, the tilt is nonsteady. Precession of the mid- and lower-tropospheric vortices occurs as the vertical tilt decreases (not shown). From Fig. 15, as in Fig. 12d, it is apparent that the lower-tropospheric vertical shear is strongest between the surface and midtropospheric vortex centers. Surface cold pools dominate this same region, and updrafts are seen primarily on the downshear side of the cold pools, consistent with Fig. 12d. It is evident from comparing the two simulations that the presence or absence of an imposed shear of 1 m s$^{-1}$ does not alter the basic mechanism of development. The formation of a tilted vortex structure may be a preferred mechanism to increase vortex-induced vertical shear beneath the center of the midtropospheric vortex, thereby enhancing low-level updrafts in this area. In some ways, this is similar to the favorability of weak shear suggested by Nolan and McGauley (2012), but with the advantage here that an environmental shear is not imposed on the vortex. Rather, shear results from the vortex structure itself and therefore does not hinder the subsequent process of vortex alignment despite the fact that the shear can be locally 5–10 m s$^{-1}$ over the lowest 3 km.

As the time of genesis approaches, the lower-tropospheric updrafts show evidence of rotation (Fig. 16). Before roughly 120 h, there is no systematic enhancement of vorticity within the updrafts compared to the background relative vorticity of the vortex at 2 km. At 120 h, there is a dramatic and sustained increase in updraft vorticity over the background value. A lower bound of the contribution of updraft vorticity to the circulation is obtained by the enhancement relative to the background (factor of about 3 between 120 and 144 h) times the fractional area of the updrafts (4%), which yields a 12% contribution to the area-mean vorticity. However, as demonstrated in section 3, vorticity long outlasts the updraft in the midtroposphere. We expect the same to be true at 2 km; hence, vorticity enhanced in updrafts will be incorporated into the mesoscale cyclonic circulation over time and remain there.

5. Conclusions

We have explored the process of aggregation of deep, moist convection in a simulation of an idealized rotating atmosphere above a uniform, constant ocean. Previous studies have shown that, in the absence of rotation, convection will aggregate through a radiation–water vapor or radiation–cloud feedback. Indeed, in our simulations, convection achieves full aggregation into a single convective complex only with radiation that interacts with cloud and water vapor. However, our study has presented a slightly different picture in which moist and dry patches emerge simultaneously. The propensity for convection to occur in midtropospheric moist patches and to avoid dry patches, combined with the ability of convection to moisten its immediate surroundings, provides a positive feedback in addition to what radiation introduces. We have also shown how the small stratiform regions associated with isolated cumulonimbi generate small amounts of vorticity in the midtroposphere that far outlast the updrafts.

After moist patches emerge, they begin to rotate. First, the rotation occurs on the scale of the patch, perhaps 30–40 km across. Patches display closed potential vorticity contours that we hypothesize further localizes the moisture and may slow the overall aggregation into a single cluster. Vortex merger then becomes an important factor in the upscale growth to a single cluster. It is not until a single large cluster forms that rapid drying occurs throughout the remainder of the model domain.

We investigated the role of radiation and found that, because of the enhanced cloudiness and water vapor within moist patches, radiative cooling is reduced when compared to dry patches. This occurs partly in response to daytime absorption of shortwave radiation in addition
to the reduction of longwave cooling at night. Adjacent moist and dry patches give rise to a weak secondary circulation with rising motion through the center of the moist patch. The modest ascent rate of 1–2 mm s\(^{-1}\) approximately cancels the subsidence induced by radiative cooling and thus maintains the enhanced water vapor within moist patches. By extension, enhanced downward motion in the dry patches augments the drying there. This quasi-balanced secondary circulation is present only with both interactive radiation and background rotation.

Once a dominant vortex emerges in the midtroposphere, the genesis process begins. The advantage of this particular simulation design is that the precursor vortex self-organizes, and there are no parameters specified a priori that determine its structure. Nonetheless, our results are broadly consistent with previous studies that have prescribed a vortex. Moistening of the lower and middle troposphere through the positive feedbacks between water vapor and convection, radiation and water vapor, and convection and relative vorticity appear crucial for intensifying and moistening the midtropospheric vortex. Once the troposphere through 7 km exceeds 80% relative humidity, the character of convection begins to change. With greater organization of updrafts comes more organization of downdrafts, despite the high relative humidity. The downdrafts generate relatively extensive cold pools that feature negative temperature anomalies of about 2\(^\circ\)–3\(^\circ\)C. Vertical shear that arises because of the variation of the tangential circulation of the vortex with height, and vortex tilt, couples with the cold pools to produce organized updrafts in the lower troposphere. The updrafts also become relatively strong at lower altitudes because of the enhanced stratification within the vortex (Raymond and Sessions 2007).

While the updrafts in the lower troposphere are stronger within the day or two before genesis, the primary effect of the cold pools is to promote organized updraft patterns that cover dramatically more area. These updrafts contribute the majority of the total lower-tropospheric, upward mass flux within the vortex. At this stage, the updrafts also exhibit rotation that far exceeds the background value of the vortex. In accord with the study of individual updrafts that we present, the vorticity far outlasts the updrafts themselves and contributes to the increase of low-level vorticity. The combined effect of slightly stronger, numerous, and rotating updrafts is a rapid increase in the surface circulation with the enhancement of surface winds that define the emergence of a tropical cyclone.

The results presented herein address a conundrum that has emerged from observational studies of pre-genesis disturbances. Such disturbances feature warming in the upper troposphere, cooling in the lower troposphere, and enhanced water vapor through a deep layer. However, this thermodynamic state features small buoyancy for parcels lifted reversibly from the boundary layer. Yet deep convection is observed to be part of the genesis process. How are these views consistent? We posit that the dynamics of cold pools and vertical shear explains the apparent paradox. Strong and organized low-level updrafts are generated on the downshear flanks of cold pools within the vortex. The gradients of temperature in the boundary layer are highly transient. Hence dropsondes, on which most observational studies are based, are distributed too far apart in space and time to resolve such structures. Observations indicate the

![FIG. 16. Time series following the vortex of average vorticity at 2 km coincident with updrafts (black) and area-mean vorticity at 2 km (red).](image-url)
presence of a broad cold dome when, in fact, the cold pools are subvortex scale and highly transient.

Although the low-level updrafts are indeed stronger prior to genesis, the relative absence of CAPE prior to genesis implies that the updrafts at higher levels will not be more intense. In our simulation, there is no evidence that the maximum updraft (typically in the upper troposphere) increases prior to genesis. We speculate that the lack of increase in updraft strength may contribute to the rather ambiguous results concerning the presence of lightning activity before genesis. It is also likely that the strength and organization of low-level updrafts is enhanced by the moist conditions present within the vortex. Parsons (1992) noted that high relative humidity (and low-level shear) promoted strong low-level updrafts in extratropical narrow cold-frontal rainbands with minimal CAPE in the environment. Furthermore, studies by James and Markowski (2010) and Kilroy and Smith (2013) show that the effect of dry air is to inhibit updrafts rather than to enhance downdrafts. Updrafts and downdrafts together define the intensity of convection and are expected to be more numerous and stronger when the relative humidity is greater through the lower and middle troposphere. Our results, combined with previous studies, clearly show that there is no need to eliminate downdrafts in order to produce a tropical cyclone.

We must emphasize that the highly idealized simulations conducted herein ignore the synoptic-scale precursors to genesis that are well established. However, the convection-scale processes described here, and their upscale progression, should be operating in the real atmosphere even if guided by the vorticity and vertical motion patterns within tropical waves. The isolation of convective processes offers a way to reconcile seemingly contradictory genesis theories. From the simulations, we have seen that the thermodynamic stabilization within a midtropical vortex does indeed induce a bottom-heavy mass flux profile (Raymond et al. 2011), but the mass flux comprises relatively strong (and rotating) low-level updrafts similar to those that appear in the so-called “vortical hot tower” paradigm (Montgomery et al. 2006; Montgomery and Smith 2014). Furthermore, the tendency of convection to cluster where the humidity of the midtroposphere is highest and further moisten where convection occurs may augment the vertical circulation diagnosed by Wang (2012) and help explain why convection systematically occurs near the center of recirculation within critical layers of waves (Dunkerton et al. 2009).

Our results have confirmed the fundamental coupling between convection, water vapor, and vorticity that produces moist vortices localized to the middle troposphere. On this basis, it would seem that vertical shear and horizontal deformation in the midtroposphere, but perhaps not in the lower troposphere, would be most disruptive to the vertical and horizontal coherence of such vortices and hence to tropical cyclone formation. It also points to the subtleties of cloud and radiative processes that may need to be modeled in order to produce realistic climatological distributions of precursor vortices, and hence tropical cyclone genesis, and the difficulties of doing so with parameterized convection.

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