SPONTANEOUS TROPICAL CYCLOGENESIS IN A CLOUD RESOLVING NUMERICAL MODEL

BY PAUL M. HOFFMAN

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[Signature redacted]

Author

Department Earth Atmospheric and Planetary Sciences

[Signature redacted]

Certified by

Kerry Emanuel, Thesis Advisor

[Signature redacted]

Accepted by

Samuel Bowring, Chair, Committee on Undergraduate Program

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Abstract

In this thesis, spontaneous tropical cyclogenesis occurring in a cloud-resolving numerical model is studied. The model environment is one of radiative convective-equilibrium on an f-plane with doubly periodic boundary conditions and constant sea surface temperature. While a variety of initial conditions may exhibit spontaneous tropical cyclogenesis, this study focuses on one. Using assumptions of axisymmetry for the growing disturbance and focusing on the large scale processes, fields were created for a number of thermodynamic variables along constant height surfaces and as azimuthal means plotted against height.

The tropical cyclone is hypothesized to develop in three steps. First, convective aggregation creates regions of high moist static energy, and regions of cold dry air. Importantly, a deep moist column is created which provides a perfect environment the developing storm. In the second step, mid-level cyclone intensification, a mid-level cold core cyclone develops in the deep moist region, and benefits from moist static energy and potential vorticity fluxes from the upper troposphere. Exhibiting anticyclonic convergent flow, the upper troposphere is an unlikely source for the mid-level disturbance, while convective downdrafts and divergent surface flow hinder energy transport from the ocean to the growing system. In fact, a cold surface anticyclone exists near the center for much of the second step. It is not until potential vorticity anomalies advect down to the surface that the final step, low-level cyclone intensification, creates a classical hurricane structure. Potential vorticity advection stimulates cyclonic flow at the surface, extinguishing the surface anticyclone, and thereby linking the mid-level disturbance to the oceanic energy source. While like some cold core cyclones previously studied, the anticyclone as an energy source is unique to this spontaneous case.
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1 Introduction

Bister and Emanuel begin their 1997 report on hurricane Guillermo and the Tropical Experiment in Mexico (TEXMEX) by stating, “It has been known for more than 50 years that tropical cyclones do not form spontaneously.” When studying difficult phenomena it is often easiest to start with negative statements, things that the phenomena are not. Positive statements on tropical cyclone genesis are more difficult to make, and a general genesis theory remains elusive. The literature describes many approaches, including gauges of hurricane likelihood (Gray’s index, genesis parameters or potential intensity), case studies, and numerical models. While certainly the least tied to reality, numerical models have been used to help advance understanding. In many instances, such as described in the literature discussed below, models reaffirm processes seen in nature. Yet in other research, and in the case of this thesis, numerical models provide an unexpected result not seen in case studies of actual storms. After considering their applicability in other studies, these numerical model results may prove very useful, especially when they seem to defy 50 year old knowledge and exhibit a case of spontaneous genesis.

One perspective on cyclogenesis focuses on the thermodynamic processes and state of the atmosphere in which tropical cyclones develop. Forming and growing largely over warm ocean waters, the cores of tropical cyclone are marked by high humidity and temperatures, but usually little widespread conditional instability. In fact, Rotunno and Emanuel (1987, hereafter RE) create and test a cloud resolving model based on moist convective neutrality. Although the authors make some alterations for the purposes of the model (creating a condition they call “model neutrality”), the model generally maintains constant saturation equivalent potential temperature ($\Theta_e$) along angular momentum surfaces—indicating neutrality to moist convection.
This approach is in contrast, as the authors assert, to much of the previous literature that focused on instability, specifically Conditional Instability of the Second Kind (CISK). Focus on CISK overemphasized the role played by cumulus convection, where more attention should be given to processes on longer time scales and influencing the larger vortex. By assuming convective neutrality, along with much more standard assumptions like hydrostatic and gradient wind balance, hurricanes would become thermodynamically reversible, and the entire system could be thought of as a Carnot engine. In this framework, cumulus clouds are “simply agents which transfer heat from the oceanic source to upper atmospheric sink on time scales much shorter than the developing vortex.” Conditional instability does occur in the model, but it is quickly eliminated through convection; the process that is more important to tropical cyclone genesis is the uptake of heat and moisture from the ocean, and the thermodynamic conditions in the environment that can impede development.

In numerical model experiments, a finite amplitude disturbance with maximum cyclonic flow in the mid troposphere was introduced into the convectively neutral domain in a control case, with varying initial conditions to test sensitivity. Tropical cyclogenesis occurred as the vortex grew through advection of high $\Theta_v$ to the interior of the cyclone. Strong surface winds from the vortex increased heat and moisture flux from the ocean, which created local conditional instability, cumulus convection, and precipitation. Downdrafts and evaporative cooling from the cumulus clouds began to stifle widespread convective growth as cool dry air was introduced to the boundary layer. However, fluxes from the ocean continued to increase $\Theta_v$ in the boundary layer and aloft, so that eventually the downdrafts themselves had high enough $\Theta_v$ that convection was no longer slowed. Once the negative effects of downdrafts were overcome, positive feedback between surface enthalpy fluxes and strengthening winds (which in turn
increased the energy flux) led to the intensification of the tropical cyclone. The catalyzing disturbance (always with its own cyclonic flow) was found to have a lower limit of intensity, and an upper limit of size. For disturbances that were too weak, uptake from the ocean was too slow, convection was weak and was extinguished quickly by downdrafts. If genesis ever did occur, it was on time scales too long to be meaningful. Also, if the catalyst was too large, too much energy was expended to accelerate the outflow at the top of the storm, and no development occurred.

Bister and Emanuel (1997) summarize RE by first addressing the damping effect of convective downdrafts. Interestingly, it is not the convection itself that is most important for tropical cyclone genesis. Instead, the test for genesis seems to be overcoming the stabilizing downdrafts. Bister and Emanuel hypothesize, based largely on results from RE that “the elevation of $\Theta_e$ in the middle troposphere just above a near-surface vorticity maximum is a necessary and perhaps sufficient condition for tropical cyclone genesis.” Using a number of radar, satellite and in situ observations from meteorological air craft, the authors observed hurricane Guillermo developing out of a mesoscale convective system (MCS) originating over Honduras. The MCS displayed a vortex in the midlevels and a cold core near the surface—markedly different conditions from those created in RE’s simulation. As a further dissimilarity to RE, it was found that while $\Theta_e$ does increase in the middle troposphere before rapid intensification, the increase is moderate and a full day before the strong intensification. MCSs occur frequently in tropical ocean environments, and distinctions between those that create tropical cyclones and those that do not aren’t clearly understood. In the case of hurricane Guillermo, genesis occurred in a large region of stratiform precipitation. Radiative heating at the cloud top in the upper troposphere, and cooling in the mid-troposphere (likely from rain evaporation) created a
vortex. Relative humidity increased in the lower troposphere due to the rain water evaporation and downdrafts, creating a column of very moist air. While stifling convection for a time, the downdrafts advected vorticity downwards to the surface. Convection redeveloped in the saturated column, thereby further increasing vorticity below the region of maximum heating aloft and a warm core grew in the previously cold boundary layer. This mechanism is summarized in Figure 1.

In addition to their earlier hypothesis based on RE, Bister and Emanuel provided a second conjecture summarizing the proposed mechanism: evaporative cooling in a region of stratiform precipitation can produce a low level cold core vortex which in turn initiates tropical cyclogenesis. The authors employed RE's model to test the new hypothesis through control and sensitivity experiments. Results very similar to the proposed mechanism for hurricane Guillermo were observed in the control case and some of the sensitivity experiments showed that the numerical simulations were most sensitive to rain intensity and duration, with greater values of either resulting in faster cyclogenesis and stronger tropical cyclones.

It was found that downwards advection of vorticity is critically important to cyclogenesis. In the early stages of the convective downdraft, and even as
the midlevel vortex was created, flow in the boundary layer was divergent and anticyclonic. To explain the initial lower level anticyclone and its eventual change, the authors provide an interpretation based on potential vorticity (PV):

\[
P V = \frac{\omega_a \cdot \nabla \lambda}{\rho}
\]  

(1)

where \(\omega_a\) is the absolute vorticity, \(\rho\) is the density and \(\lambda\) is some conserved quantity, \(\Theta_e\) in this case. Consider a thought experiment pictured in Figure 2. An area of diabatic cooling is marked dark gray, in a larger cyclonic flow, and with constant absolute vorticity throughout the domain as shown by the arrow. Part a shows PV before any vertical motion and as a result of the cooled area. As downdrafts move through the cool region and the associated gradients, PV anomalies are created, as indicated in part b. Parcels above the area of cooling pass through the top gradient and gain PV, until the entire dark gray box has a positive PV anomaly. As parcels originally in the cool box pass through the lower gradient, they acquire the negative PV anomaly, creating divergent anticyclonic flow at the surface. Only once the cooled area descends to the surface with the downdrafts, will the negative PV anomaly be extinguished, and the area of cooling strengthens the vorticity of the whole system through the positive PV anomaly aloft.

In the case of hurricane Guillermo, the boundary layer exhibited a cold core with low \(\Theta_e\) and divergent anticyclonic flow near the surface. Much like the thought experiment, advection of the vortex to the surface created cyclonic flow, which helped concentrate heat and moisture fluxes from the ocean, and \(\Theta_e\) increased near the surface. Existing in a saturated column, and with a cold environment near the core, the local barriers to convection were greatly reduced and the warm core cyclone grew rapidly. Surface enthalpy fluxes and intensifying flow provided positive feedback, and since the cyclone was growing in an already
moist environment (thanks to the stratiform precipitation), evaporative cooling and downdrafts were less intense. While the cold core was not considered in RE, the strong dependence on oceanic fluxes and growing surface flow existed in both sets of simulations.

A second case study explored cold core cyclogenesis in the Mediterranean using the framework of potential intensity—a thermodynamic parameter dependent on the thermodynamic disequilibrium between the ocean and atmosphere, and the temperature difference between the hurricane’s thermodynamic source (ocean) and sink (upper troposphere). Emanuel (2005) studied the small hurricane that occurred in mid January 1995. Cyclogenesis took place directly beneath an upper-level cut-off cyclone and slightly to the west of a large surface trough. The upper-level low lifted, cooled, and thereby moistened the air below, creating a cold moist core below an existing vorticity maximum. Emanuel employed the same cloud resolving model from RE to simulate this case. The main differences between this application and RE were the sea surface temperature (SST) of 22°C (instead of 24°C), the increased Coriolis parameter due to the higher latitude, and the existence in the initial state of the cold-core, upper-level cyclone. While baroclinic instability may have played a role in the early formation of the two
troughs, and though initial soundings showed considerable convective instability, these effects were ignored in the axisymmetric and moist convective neutrality assumptions. An upper level cold low was introduced, with maximum winds of 20 m s\(^{-1}\) centered at a radius of 300 km and 10 km above sea level, decaying linearly to zero at the surface. Thermal wind balance was enforced. Small perturbations were introduced to the surface fluxes to initiate convection, and in just three model days rapid intensification took place.

In the studies mentioned above, cloud-resolving numerical models provided important insight into real world hurricanes. But also important to understanding tropical cyclogenesis, is exploring the results of computer simulations taking place in idealized environments. Nolan et al. (2007) applied a new convective model to examine the role of external parameters in controlling radiative-convective equilibrium. The specifics of the model will be discussed in section 2, but the results are related here. The authors began with the working hypothesis based on a previously defined genesis potential index (GP) based largely on potential intensity theory. They hypothesized that an increase of any term comprising the GP would allow a decrease of the strength of the catalyzing disturbance needed to initiate tropical cyclogenesis. The authors—as supported by results from RE—further conjectured that there would exist a lower limit of disturbance intensity, below which no hurricane would form. Most interestingly, the authors observed cyclogenesis in cases of very weak initial disturbances as well as in their complete absence, forcing them to reject their hypothesis. As in the RE case, genesis progressed more slowly than cases where the initial disturbance was stronger, but it did occur. It is suspected that a simulation with wind shear could change these results, but this test was not included in the paper.

It is this case of spontaneous tropical cyclogenesis in the Nolan et al. model that is the focus of this study, because it is a seeming contradiction to fifty year
old knowledge. Idealized environments provide a perfect space in which to study and experiment with this unexpected case, and it has great potential to improve understanding of tropical cyclogenesis. Section 2 outlines the parameters of the model, some past results, and initial conditions for this spontaneous case. Section 3 provides the results of the simulations, while theory and analysis of this case are given in section 4. Section 5 provides a summary and discussion of future work.

2 Methods

The spontaneous genesis data used in this thesis were obtained from Eric D. Rappin and are the results of simulations done in conjunction with the Nolan et al. (2007) paper. The model is an adaptation of the Weather Research and Forecast Model (WRF), set on an f-plane with doubly periodic boundary conditions. Horizontal resolution is approximately 4 km on an 800 × 800 km domain; the vertical dimension is split into 40 height/pressure levels at a resolution of 24 mb, therefore there is higher resolution at lower heights with levels becoming more spread aloft. The tropopause is arbitrarily chosen to occur at 15 km, and the stratosphere above that is driven toward an isothermal state using a relaxation scheme. While the model was run at very high temporal resolution for some applications, the state was saved less frequently for the current dataset and results in this thesis have 4 hour temporal resolution. The initial state is always one of Radiative Convective Equilibrium (RCE)—an atmospheric steady state in which loss of energy to space through longwave radiation is balanced by energy flux from the surface, with convection and radiation redistributing energy upward. As the authors point out, the tropical atmosphere is seldom in such a state, due to large-scale circulations like Hadley or Walker cells. Yet, citing Bretherton et al. (2005), they maintain that an atmospheric model in
RCE is an agreed upon and sensible equilibrium condition to use as a starting point in investigations of tropical disturbances.

RCE was first obtained on a small 200×200 km domain with a hydrostatically balanced atmosphere. Initial conditions included a temperature that was 2.5 K less than the SST, and declined at a constant lapse rate; relative humidity was set at 50% below 12 km and 0% above. To initiate convection, small random temperature perturbations were introduced at every grid point below 3 km. As other studies have corroborated, the model atmosphere takes about 50 model days to settle into a statistically steady RCE. Nolan et al. ran small domain simulations to 90 days and averaged the last 30 to determine RCE for the small subsection, which was then used for the entire domain. Importantly, especially in terms of RCE, it should also be noted that the model does not include short wave radiation. The authors point out that most effects of shortwave radiation are accounted for in other ways. Constant SST is used in place of radiative surface heating, and the tropopause is created through the upper level relaxation procedure instead of stratospheric heating. Unaccounted for is shortwave radiative heating in the troposphere, yet the savings in computation resources was judged to be worth the exclusion. In consequence, the diurnal cycle, latitudinal variations, declination of the sun, and effects of cloud albedo needed not be calculated.

RCE simulations without a mean wind were found to have unrealistically high values of potential intensity, though in a doubly periodic f-plane, inclusion of mean wind or wind shear is impossible without larger scale forcing through temperature or pressure gradients. To solve these problems, a mean westerly surface wind was introduced through a constant wind term in the calculation of surface evaporation rates:

$$|V^*| = |\vec{v} + u_m\hat{t}|$$

(2)
where $V^*$ is used to calculate evaporation, $\bar{v}$ is the 10 m wind and $u_m$ is the westerly wind. In the spontaneous genesis case (and in most other experiments by Nolan et al.), the constant wind term was 3 m s$^{-1}$. The mean surface wind broke the horizontal symmetry, and no mean wind shear was included. Bister and Emanuel, and other studies cited by Nolan et al., make it clear that wind shear is an important factor in tropical cyclogenesis, yet the authors assert that genesis usually occurs in regions of no or very low wind shear. In the Guillermo case, wind shear would have disrupted the downward advection of vorticity and likely killed the developing storm. With this justification, the exclusion of vertical shear is not seen as a significant limitation.

In this thesis, data from an atmosphere in RCE with the Coriolis parameter at 30° latitude, SST of 25° C, and mean surface wind of 3 m s$^{-1}$, are used to study the resulting spontaneous tropical cyclogenesis. Nolan et al. cite a similar case in their paper: $f=f(20°)$, SST=27.5° C and a mean wind of 3 m s$^{-1}$. Although the two sets of initial conditions are two elements in what might be a long list of viable possibilities, the study did not attempt to define the entire parameter space, but instead focused on this single case. Likewise, this thesis only examines the one case outlined above. Model days 6–12 were examined in detail. Days 1–5 were determined to be well before the first states of cyclogenesis, and those days after 12 reflected the mature hurricane. The model was only run until day 16, as the simulation space eventually became corrupted by the hurricane through its own strong convection and extent throughout the domain. Variables of interest included pressure, moist static energy, saturation moist static energy, temperature, specific humidity, and wind velocities. These variables were interpreted in time series of horizontal slices (along constant $z$ levels) and in radius vs. height fields where the variables were averaged around radius circles. In order to create radius vs. height plots of development, the
center of the tropical cyclone was determined from 3 km wind fields at every time step. By day 12, the tropical cyclone is well established and the center is easy to find. Moving back in time, the fields are less clear and the extrapolation from the more developed hurricane path was used to help determine position. Radial plots were made for each time step and include virtual temperature, azimuthal and radial winds, relative humidity, moist static energy and saturation moist static energy. Perturbation plots were also made in which the field at each level is compared to the horizontal domain average. In the case of saturation moist static energy, levels below 900 mb, the sub-cloud region, display normal moist static energy because there is likely no saturated air at those levels, and the saturated moist static energy would not make sense.

3 Results

As Nolan et al. outline, spontaneous tropical cyclogenesis occurs in two steps, convective aggregation and cyclone intensification. However, in this case, cyclone intensification can be divided further into mid-tropospheric and low-tropospheric steps. Although it is reasonable to suppose that convection in an environment of radiative-convective equilibrium would be random, locally short lived, and homogeneously distributed throughout the domain, convection often aggregates in some areas and is completely absent in others. In the project data, it was found that aggregation occurred on day 6 and intensified steadily into the tropical cyclone, as seen in fields of moist static energy (h) and saturation moist static energy (h*), quantities roughly equivalent to $\Theta_c$ and which apply to unsaturated and saturated air respectively:

\begin{align*}
    h &= c_p T + gz + L_v q \\
    h^* &= c_p + gz + L_v q^*
\end{align*}

(3)  
(4)
where $T$ is the temperature, $c_p$ is the gas constant at constant pressure, $g$ is acceleration due to gravity, $z$ is height, $L_v$ is latent heat of vaporization, $q$ is the specific humidity, and $q^*$ is the saturation specific humidity. Moist static energy and $\Theta_e$ are conserved in moist convection as parcels move through the atmosphere in adiabatic processes. Therefore, areas of high energy have come to that state through energy transport, and not as a result of air motion. Figure 3 shows moist static energy fields at 1.5, 3.05 and 5.91 km while Figure 4 gives the corresponding humidity fields and Figure 5 shows temperature. (When reading these and the following figures, keep the doubly periodic boundary condition in mind; i.e. a feature that extends to the left (top) edge of any figure will continue seamlessly on the right (bottom) edge. Also, all figures are shifted using the doubly periodic condition so that the calculated storm center is always in the center of the plot.)

The mechanics of aggregation are first clear in the 6 km plots, where low moist static energy areas are visible as early as day 7 in the upper right and lower corners of the domain. What begins as a small band of elevated energy grows steadily throughout the simulation. Examination at the lower levels shows random areas of high moist static energy through day 7 with areas of low moist static energy becoming more pronounced at the end of day 8, as the widespread convective activity becomes organized. Temperature fields reveal general warming in time, with one notable exception at the 6 km level, and stronger warming near the developing cyclone center at all levels. With the domain becoming generally warmer, it is the humidity term in Equation 3 that accounts for most of the moist static energy gradient seen in Figure 3. Specific humidity fields show drying in the areas of low moist static energy, and moistening in the areas of continued convection. Note from the studies cited earlier that the first two terms on the right hand side of Equation 3 do not effect convection in the same...
way. While low moist static energy over a warm ocean is a primed environment for convection, a warm area with low humidity is less conducive than a cooler region with high humidity. This observation is born out in the correlation of humidity and moist static energy fields. Figure 6 shows the evolution of the pressure field, which is strongly related to the rotational component of the horizontal velocity. At almost all levels, the low deepens most intensely on days 11 and 12, with slow change in the days before. The change in these fields occurs more suddenly than the corresponding changes in energy, temperature or humidity.

The sudden drop in pressure does correspond to an improved organization in azimuthal velocity, seen in Figure 7. As the evolution of azimuthal winds make clear, the tropical cyclone grows out of a circulation first concentrated between 2 and 4 km aloft. Starting on day 8, velocity begins a steady increase and on day 9 the circulation begins to move towards the surface. Through much of day 9, there is zero or even anticyclonic flow near the surface below areas of strong flow aloft; this is especially true in the area closest to the storm center. It is not until day 10 that the strongest flow has reached the surface, and days 11 and 12 show much greater organization and higher velocity. Almost as intense as the cyclonic flow in the midlevels, anticyclonic flow also develops between 10 and 12 km aloft. As radial velocities in Figure 8 show (note positive radial velocity is away from the center), the upper level anticyclone also exhibits convergent flow, while at most other layers in the model atmosphere, velocity is weak but radially outward. However, it should be noted that further examination of the wind fields at high altitude do not exhibit the same organization and axial symmetry seen at lower altitudes. Although the mean flow is anticyclonic and convergent, perhaps due to midlevel low pressure, winds are not well organized into a coherent upper level anticyclone, and areas of convergence at one section.
are sometimes balanced in the azimuthal average by areas of divergence in other sections (similarly for cyclonic and anticyclonic areas).

In contrast to the gradients of azimuthal and radial velocity, relative humidity is more constant in space and time. Figure 9 shows the obvious high humidity at low altitude associated with oceanic moisture flux, but also moist conditions near the cyclonic center as early as day 7. Regions of very moist air extend up to 6 or 8 km in altitude, with a significant moisture gradient developing in the middle and high troposphere towards the core. The driest region of the developing cyclone is usually in the middle troposphere at large radius. Relative humidity near the core remains high until day 10 when drying occurs in the midlevels and humidity increases below. By day 12, moisture is concentrated almost exclusively in the core of the tropical cyclone or at the surface, and drier conditions extend throughout the rest of the domain.

Figure 10 shows the radius vs. height field for moist static energy, and Figure 11 the saturation moist static energy. Notice the band of low moist static energy air in the middle troposphere, and the region of increased moist static energy at the top, suggesting high level clouds, and perhaps a region of precipitation in mid levels. To make horizontal gradients in this field clearer, perturbation moist static energy (deviation of the variable from the mean on a given height surface) is shown in Figure 12. The perturbation shows a midlevel warm disturbance with lower relative static energy near the surface and aloft, and is associated with a weak low pressure system at midlevels. Keep in mind that the midlevels where the warm perturbation exists is an area of low moist static energy—the moist static energy field shows that even though there is a strong warm perturbation, the mid level warm disturbance is not a maximum of moist static energy; it is a slightly higher energy area in a cold dry region. Much like the middle levels seen in hurricane Guillermo and TExMEX. Nearer to
the surface (where actual moist static energy is very high), the perturbation moist static energy is negative near the center of the system. It is not until day 11 and 12 that a high moist static energy perturbation exists at the surface. In the evolution of moist static energy, the area of maximum energy perturbation broadens and intensifies until there is high relative moist static energy from 1 to 4 km altitude, and after some fluctuations, a warm core replaces the negative moist static energy perturbation at the surface.

The radius vs. height plots discussed above assume axisymmetry. However, the assumption does not hold for vertical velocity (not shown). Cells with strong updrafts and downdrafts exist at many radii. This shows, as outlined in the introduction, that although the focus of the analysis is larger dynamics and longer time scales, the strongest vertical motions are unevenly spaced and on small short scales, associated as they are with individual convective events.

Recall the three development steps which opened this section: aggregation, mid-level cyclone intensification, and low-level cyclone intensification. The data show aggregation in days 6 and 7, transitioning to mid-level intensification on days 8–10, and lower-level cyclone intensification on days 11 and 12. The three steps summarize the varying driving forces in the growing disturbance.

4 Analysis

In many ways the tropical cyclogenesis described in the previous section is a typical cold core cyclone as described by Bister and Emanuel, and in some other ways like the case presented by RE. Most obviously, the surface cold core disturbance is seen in the plots of perturbation and complete moist static energy and verified by virtual temperature (not shown), however the cold core can also be seen in azimuthal velocity. Through thermal wind balance, horizontal gradients of temperature must be associated with vertical gradients of horizontal
geostrophic wind. Admittedly, the flow is surely not in geostrophic balance everywhere, and gradient or even cyclostrophic balance may be more appropriate at some radii, but the only way for a cold core near the surface to persist in the manner it does, is through thermal wind balance. Also, the generally weak winds at the surface make the geostrophic assumption more plausible, as these winds need balance only the geostrophic component of the stronger midlevel rotation. The wind field shows cyclonic vertical shear of the horizontal wind, consistent with a cold core cyclone and the associated temperature gradient.

In this case, as convection aggregated out of random convective cells, I hypothesize that, like the hurricane Guillermo case, cooling from downdrafts and evaporation created a cold core near the surface, while air with high moist static energy from convection remained aloft or advected down from the upper troposphere. Subsequently, evaporation of falling precipitation, like that discussed for hurricane Guillermo, intensified the midlevel disturbance. This mechanism created a strong midlevel mesocyclone with a cold, humid core, that became a hurricane only when surface fluxes connected it to the oceanic energy source.

The model includes a longwave radiative scheme, but, as discussed in section 2, no calculation of shortwave heating in the troposphere. Therefore, the top of the troposphere may be losing energy to space, and the cloud top heating discussed with hurricane Guillermo is absent; despite this, the evolution of moist static energy as seen in Figure 10 suggest that energy is advected down from the higher troposphere into the lower energy mid levels. Perturbation moist static energy illustrates the effect of this advection, as the warm disturbance grows steadily (note continued drying at higher radius in the mid levels may also amplify this result). In both cases, downward transport of moist static energy increases the temperature and downward moist static energy gradient, which is associated with a positive PV anomaly at the mid-level, and increasing
anticyclonic flow above. This energy gradient above the mesocyclone is more pronounced in fields of saturation moist static energy, suggesting that the PV anomaly is higher for saturated air. Therefore, the upper levels provide an engine for the midlevel warm disturbance. With slightly different geometry, this downward advection of potential vorticity is very similar to the thought experiment given in figure 2, and the interpretation of hurricane Guillermo given by Bister and Emanuel. Negative PV, and divergent flow in the surface cold core, sit under a growing positive PV anomaly above the mesocyclone until downward advection extinguishes the cold core and the surface undergoes rapid cyclonic intensification.

Azimuthal and radial velocities support this hypothesis. For much of the tropical cyclogenesis, flow is radially outward in the lower atmosphere and towards the center only above 6 km. While azimuthal velocity does intensify and eventually reach the surface at large radius, and while this is the only source of heat and moisture for the entire model, the radial flow is away from the center at low levels. It seems that little of the strengthening warm core in the middle troposphere is due to ocean flux. Until day 11, and the low-level cyclone intensification step, the radial inflow at the top of the troposphere is as strong as the maximum azimuthal velocity. Relative humidity fields (Figure 9) show that moisture remains high near the core, through the level of the warm disturbance and in some cases much higher. The condition of a deep moist column emphasized in Bister and Emanuel, is clearly evident in this simulation through day 10. The transition from mid- to low-level cyclone intensification seems to correspond to drying above the mesocyclone, and concentration of the moisture into the newly created surface cyclone.

In the case of hurricane Guillermo, maximum winds existed in the mid levels, in a structure much like the cold-core disturbance observed here, and strength-
ening of the flow came from the atmosphere-ocean boundary—oceanic $\Theta_e$ flux and the associated positive feedback of intensifying flow. While a PV anomaly existed, it was driven initially only by evaporative cooling and convective downdrafts. In the simulated spontaneous case, maximum flow is near the top of the troposphere, where there exists a reservoir of moist static energy and potential vorticity. While fairly persistent until day 11, the intensity of the surface cold core fluctuates. Seen mostly in the perturbation moist static energy plot, times like day 8, 12:00 show almost no cold core and surface cyclonic azimuthal flow very close to that at the midlevels. By 20:00 of the same day the cold core is firmly reestablished. The surface warm core exhibits another period of growth early in day 10, with one final oscillation to cold core at 12:00, and rapid growth after that. It is on day 11, when the warm core is firmly established, that the azimuthal velocity becomes most organized throughout the troposphere and the central pressure drops quickly. In doing so, the anticyclone at the upper levels is pushed out to higher radii, and the classical hurricane flow develops.

Unlike the case of Mediterranean cyclogenesis, dynamics from a single cut-off low did not catalyze tropical cyclogenesis. In neither the Mediterranean nor the Guillermo case, was there an analogous structure to the upper level anticyclonic flow. All three cases support the claim that a moist column and depression in $\Theta_e$ in the midlevels are necessary to tropical cyclone formation, yet all three provide slightly different catalyzing events: ocean fluxes, upper level lows, and upper level anticyclones.

The upper level anticyclone does present some questions and doubts. The upper troposphere—especially without shortwave radiative heating—is seldom the source of energy for developing disturbances. In fact, using convective neutrality to view a tropical cyclone as a reversible thermodynamic system implies that the upper troposphere is the heat sink. It should be noted that once the
hurricane is developed, this dynamic is observed, but the thermodynamic reversal from development to mature system is questionable. The conditions that brought about the state of the model at day 6 is beyond the scope of this paper, but it is curious that the midlevels would have such low moist static energy, and the upper levels would be so high, due, as it appears, to humidity differences. Also, while there is an anticyclone with radial inflow as seen in the azimuthal mean plots, upper level winds are heterogeneous and include many small scale variations and eddies.

Interestingly, Rotunno and Emanuel’s study, used as a basis for much of the research cited here, and an inspiration for the current numerical model, incorporates an initial disturbance that excludes these dynamics. The initial paper describes simulations initiated by a cyclonic disturbance that extended through most of the troposphere, with a wind maximum about where the mid level cyclone was observed and decaying to zero above. Other papers focusing on specific hurricane events use similar catalyzing wind profiles. The upper level anticyclone seems a unique feature of this spontaneous case.

5 Summary

I have provided a three step theory of spontaneous tropical cyclogenesis as observed in a cloud-resolving numerical model in radiative-convective equilibrium. Convective aggregation, the first step, is an important result of the initial conditions, and created the catalyzing concentration and organization of convection that eventually led to cyclogenesis. After convective aggregation, a cold core cyclone formed in a moist column of air with maximum winds in the mid-troposphere; this is the second step, mid-level intensification. The cold core cyclone extended to the surface with anti-cyclonic surface flow at low radius, and radially outward flow throughout the low altitudes. Thereby oceanic enthalpy
flux to the developing disturbance was minimal. The mid-level disturbance grew, instead, through moist static energy and potential vorticity flux from the upper troposphere. An upper level anticyclone with radial inflow concentrated moisture and facilitated the moist static energy flux. Figures 10 and 12 show the results of this flux and the growing midlevel disturbance. The midlevel cyclone had strong flow for much of its existence, and in a mechanism much like Figure 2, a positive vorticity anomaly was created by the $\Theta_e$ gradient above the mesocyclone and was advected downwards to intensify the mid-level disturbance. As the potential vorticity anomaly reached the surface, the final step of tropical cyclogenesis occurred. The low-level cyclone intensification linked the oceanic heat source to the mid-level disturbance, and on day 11 the tropical cyclone underwent rapid intensification with a deep and sudden pressure drop, development of an eye and strongly organized cyclonic flow. The upper level anticyclone moved to higher radius and the disturbance looked like a classical hurricane.

The mechanism driving the genesis, the cold core cyclone and upper level anticyclone is similar to previous research as a case of potential vorticity advection. However, the anticyclone as an important source of moist static energy is, to the author’s knowledge, unique to this spontaneous genesis case. Further research should be done to understand the complexity of the upper level flow, and the conditions that facilitated it. The sensitivity of the mechanism to variations in model parameters, and general exploration of the variable space for spontaneous tropical cyclogenesis would also be interesting areas for future research. It is not clear what application this thesis might have to real tropical cyclones, and the author knows of no tropical cyclone that developed in the way outlined here, but perhaps some aspects of the theory can be used to understand actual tropical cyclogenesis events.
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6 References


Figure 3: Timeseries of moist static energy fields at constant height.
Figure 4: Timeseries of specific humidity fields at constant height.
Figure 5: Timeseries of temperature fields at constant height.
Figure 6: Time series of pressure perturbation (deviation from horizontal reference pressure) fields at constant height.
Figure 7: Azimuthal wind velocity as a function of radius and height
Figure 8: Radial wind velocity as a function of radius and height
Figure 9: Relative humidity as a function of radius and height
Figure 10: Moist static energy as a function of radius and height
Figure 11: saturation moist static energy as a function of radius and height
Figure 12: Perturbation moist static energy as a function of radius and height