WHAT IS THE TRIGGER FOR TROPICAL CYCLOGENESIS?

David S. Nolan*
Division of Meteorology and Physical Oceanography
Rosenstiel School of Marine and Atmospheric Science
University of Miami, Miami, Florida

1. MOTIVATION

This study was inspired by the results of simulations which were originally intended to study the effects of asymmetric convection on a developing tropical cyclone as previously considered for linear dynamics by Nolan and Grasso (2003). For such a purpose, I had designed an idealized WRF simulation which I expected to illustrate the steady development of a tropical storm. The initial condition was a balanced vortex with cyclonic winds of 10 m/s at a radius of 100km, maximum at the surface and decreasing with height. The environmental sounding was that of the Jordan mean hurricane season (Jordan 1958). Convection was initiated with random noise in the low-level wind field and sustained by the frictionally generated secondary circulation. A vertical cross section of the initial wind field is shown in Figure 1a.

The simulation behaved quite differently than I expected. Rather than increasing steadily, the vortex intensity appeared to remain flat or even weaken for about 60 hours, and then suddenly began to intensify very rapidly. The minimum surface pressure from t = 36 to 72 hours is shown in Figure 1b. The “transition” from a steady or weakening vortex to a rapidly developing tropical cyclone appears to occur in just a few hours!

In this presentation I will explore the model output of idealized tropical cyclone simulations to determine what physical processes and changes occur leading up to the moment when rapid pressure fall begins. In these simulations, this moment is concurrent with the appearance of a smaller but strong surface vortex, and shall hereafter be referred to as “genesis.” The results will be compared to recent studies of tropical cyclogenesis.

2. MODEL PARAMETERS AND INITIAL DATA

All simulations were performed with WRF version 2.1.1. Two grids were used: the inner grid had 180x180 grid points with 2km resolution, and the outer grid had 240x240 points with 6km resolution, and doubly periodic boundary conditions. 40 vertical levels were evenly spaced in the WRF hydrostatic pressure (η) coordinates between the surface and z = 18 km. The YSU planetary boundary layer scheme (Noh et al. 2003) and the WRF 6-species microphysics schemes (Hong et al. 2004) were used. The Rayleigh damping layer was used above 14 km. A time step of 20 s was used on both grids.

The initial azimuthal wind profiles $v(r)$ were generated from Gaussian vorticity profiles, which were then extended into the vertical with functions of the form

* Corresponding author address: Prof. David S. Nolan, RSMAS/MPO, 4600 Rickenbacker Causeway, Miami, FL 33149. email: dnolan@rsmas.miami.edu

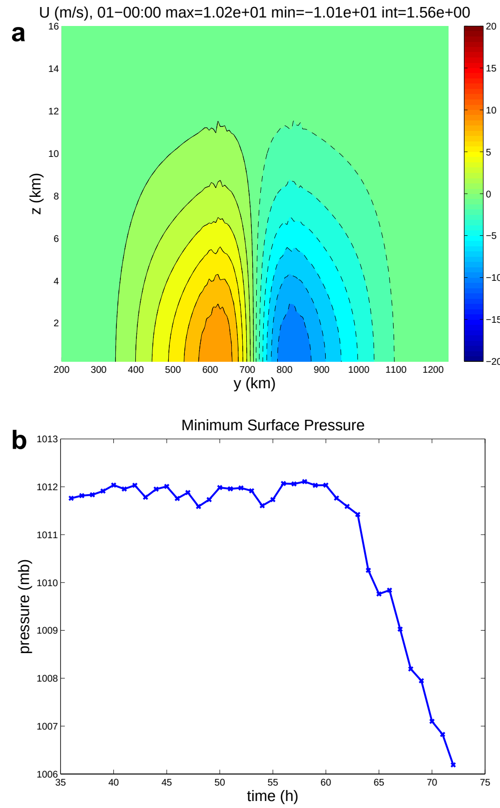

Figure 1: The control simulation: a) north-south cross-section of the initial zonal wind field; b) minimum surface pressure every hour from t = 36 to 72 h.

$$V(r, z) = v(r)e^{-|z - z_0|^{\alpha}} \tag{1}$$

The pressure and temperature fields which hold these fields in hydrostatic and gradient wind balance were computed with the iterative scheme of Nolan et al. (2001). The surface boundary condition is an ocean with a fixed SST of 29 C, and the Coriolis parameter is set to f = 5.0×10^{-5} s^{-1}. From hereafter, the wind field described in section 1 and shown in Figure 1a will be referred to as the “control” simulation.

3. INNER CORE EVOLUTION

The inner-core wind field goes through significant structural changes in the 60 hours before intensification begins. Due to surface friction, the surface wind field decays steadily. At the same time, a mid-level vortex with a smaller RMW develops and slowly strengthens, presumably due to the effects of convection, latent heat release, and mid-level inflow. The mid-level vortex is clearly seen in the azimuthally averaged azimuthal wind

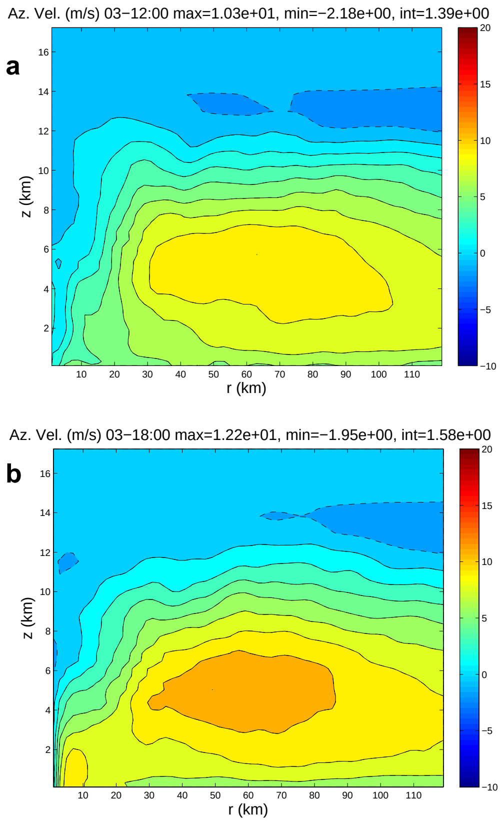

Figure 2: Azimuthally averaged azimuthal wind field in the control simulation at a) t = 60 h, and b) t = 66 h.

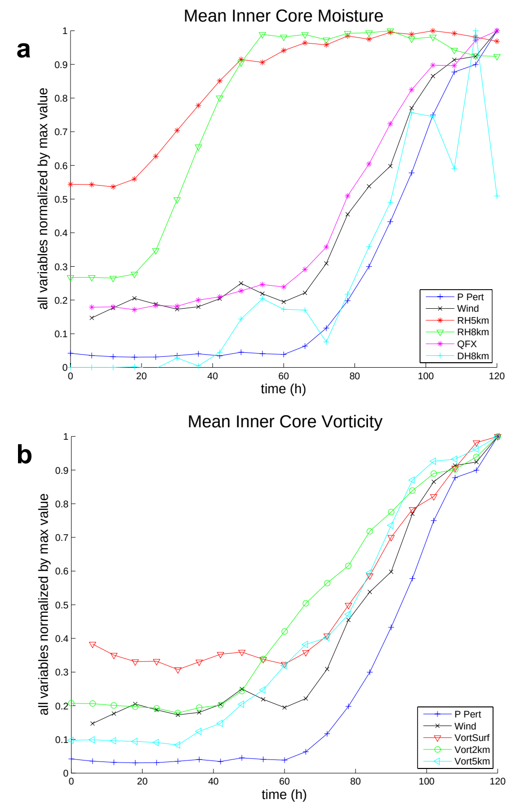

Figure 3: Mean inner core a) moisture variables (see text for details) and b) vorticity; also shown are negative surface pressure anomaly and maximum surface wind speed. Note that all variables are normalized by their maximum value over the interval shown.

field at t = 60 h (Figure 2a). Just 6 hours later, the mid-level vortex has intensified further, but in addition a much smaller and quite strong vortex has appeared at the surface (Figure 2b).

Cross-sections of relative humidity through the center of the vortex show that the middle and upper levels are steadily moistening in the pre-intensification period (not shown), and rapid intensification begins shortly after the vortex core is nearly saturated from the surface all the way up to z = 12 km.

The evolution of the wind field suggests that the mid-level vortex is more "important" for genesis than the surface vortex. However, a simulation with an initial vortex similar to the control, but with its 10 m/s winds maximized at z = 4 km, took almost exactly the same amount of time to develop. As I will show below, near-saturation of the vortex core was necessary before rapid intensifcation in this case as well.

4. **ANALYSIS**

Consideration of the above leads naturally to two questions: 1) What can these simulations tell us about the necessary conditions for tropical cyclogenesis? 2) Are there any observable trends that could be used to predict genesis? To address these questions I devised some quantitative measures of the changes in the vortex core. The first set of measures describe the moisture budget. As shown in Figure 3a, they are the mean values of surface moisture flux (QFX), RH at z = 5 km, RH at z = 8km, and diabatic heating at z = 8km, all averaged over a 100km x 100km box around the center of the vortex. The second set of measures describe the vorticity; Figure 3b shows the vertical vorticity at z =10m, 2km and 5km, averaged over the same region. Also shown in both figures are the surface wind speed and the surface negative pressure anomaly. In both figures, each variable is normalized by its maximum value over the time interval.

Several conclusions can be drawn from these figures. Regarding the vortex dynamics (Figure 3b), the surface vorticity is steady or even declining in the hours before genesis. The 2 km and 5 km vorticity increase steadily before the time of genesis, but neither show a significant "jump" beforehad; however, it is interesting to note that the 2km and 5km vorticity values reach the surface value just before genesis.

Regarding the moist dynamics, the surface moisture flux (QFX) follows the wind speed closely, but its increase does not precede genesis. RH at 5km and 8km do rise substantially, in order, and indicate near-saturation in the 6-12 hours before genesis. In addition, the sudden drop in surface pressure is preceded by a "burst"

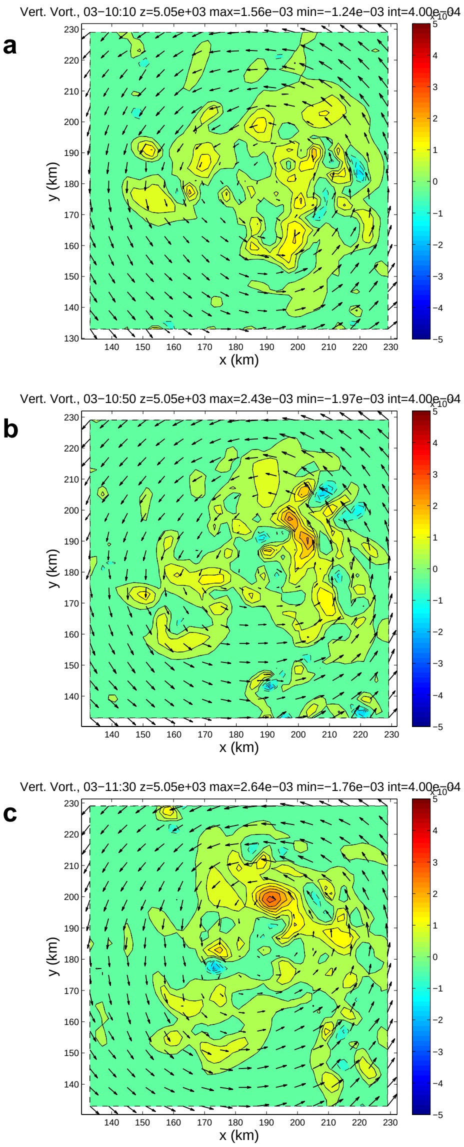

Figure 4: Horizontal sections of vertical vorticity overlaid with horizontal wind vectors at z = 5 km, a) t = 58 h, 10 min; b) t = 58 h, 50 min, c) t = 59 h, 30 min.

of latent heat release associated with deep convection.

A preliminary conclusion one might draw from these analyses is that genesis does not occur until the mid-level vortex has achieved some (as yet unknown) sufficient strength, and the entire column in the vortex core is nearly saturated. This is consistent with the genesis mode proposed by Bister and Emanuel (1997).

5. VORTEX MERGER

The merger and axisymmetrization of pre-existing vortices is a central theme to a number of studies of tropcial cyclogenesis. Earlier papers focused on the formation and merger of "mesoscale" vortices, i.e., on scales of 10-100 km (Ritchie and Holland 1997; Simpson et al. 1997; Davis and Bosart 2001). More recently, the vortex

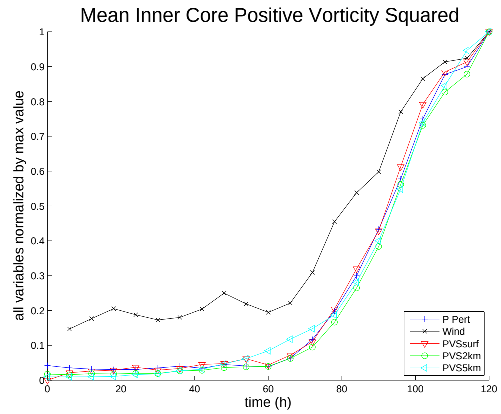

Figure 5: Mean inner core positive vorticity squared for the control simulation (see text for details).

meger concept has focused more closely on smaller scale vortices generated by individual convective updrafts, on scales down to 2 km or even smaller (Davis and Bosart 2002; Hendricks et al. 2004; Reasor et al. 2005; Montgomery et al. 2006).

The idea that vortex merger is *critical* to the structure changes that lead to genesis is not supported by these simulations. Low-level (surface to z = 2km) vorticity does not appear to increase through vortex merger events; rather, vorticity increases "in place" due to stretching from above. None of the positive vorticity anomalies at low or and mid-levels are closed circulations, suggesting that axisymmetrization by the the broader circulation is the organizing principle, rather than "merger." The increase in strength of the vorticity anomalies at z = 5km as seen from Figure 4a to 4b is caused by embedded convection (not shown). However, some "conglomeration" of vorticity *after* this convection has dissipated is also apparent in the 30 minutes before genesis occurs, as shown in Figure 4b and 4c.

In an effort to quantify the importance of strong vorticity anomalies in the genesis process, I devised an alternative measure of vorticity, which is the mean value of the positive vorticity *squared* in the same 100km x 100km box around the cyclone center (hereafter, PVS). Values for the control simulation are shown in Figure 5. Two remarkable facts can be noted: 1) The PVS at all three altitudes (10m, 2km, and 5km) follow each other very closely, and also lie nearly on top of the surface pressure anomaly; 2) The PVS at 5 km "jumps up" relative to the other curves in the hours before genesis.

Point (1) is quite interesting, because I am not aware of, nor have yet to find, a straightforward relationship between squared vorticity (or enstrophy) and hydrostatic pressure perturbation for *balanced* vortices. Point (2) suggests that an increase in intensity of the mid-level (as opposed to low-level) vorticity precedes genesis, but whether this is necessarily due to merger, or stretching, or some combination of both remains for further analysis.

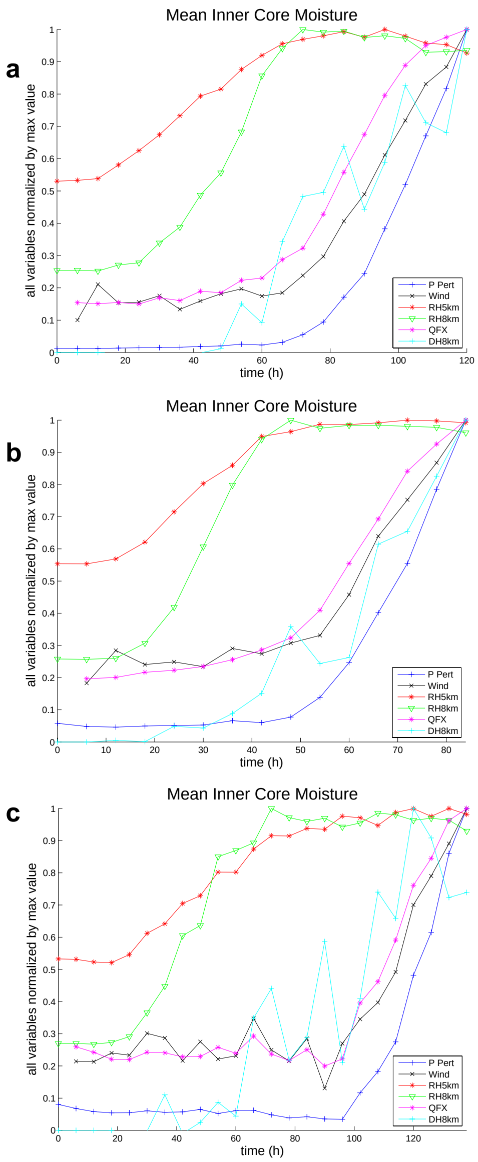

Figure 6: Inner-core moisture analyses for three different initial conditions: a) an elevated 10 m/s vortex with 5 m/s surface winds; b) A deeper 10 m/s surface vortex; c) A shallow 10 m/s surface vortex.

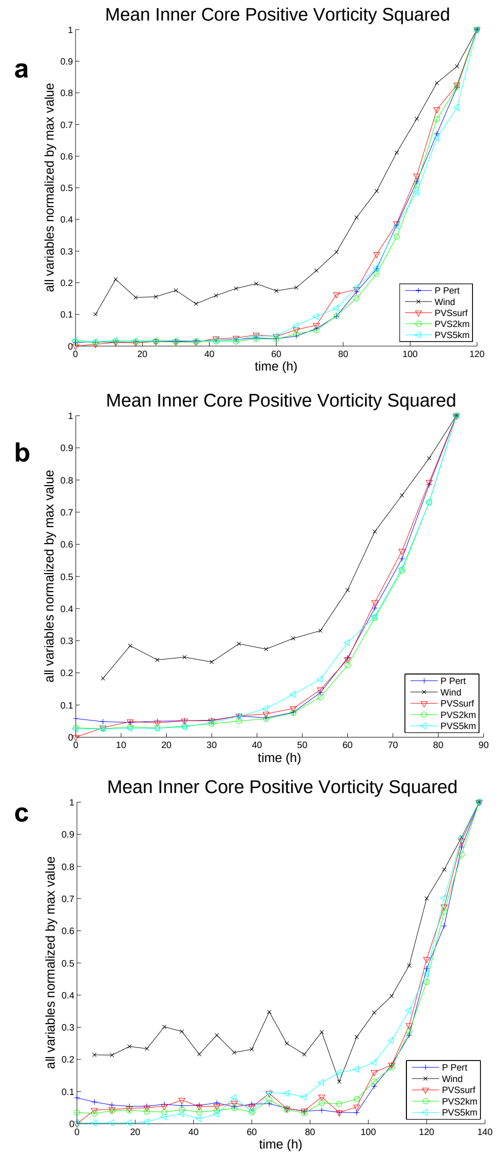

Figure 7: Inner core PVS analyses for the same simulations shown in Figure 6.

6. LESSONS FROM VARIATIONS ON THE INITIAL VORTEX

I have applied the same analysis to a number of simulations with initial conditions that are variations on the control simulation. Figure 6a and 7a show moisture analyses and PVS for the aforementioned "mid-level" vortex with 10 m/s maximum winds aloft and 5 m/s surface winds. The same changes in the inner-core variables occur as was shown for the control simulation.

Figure 6b and 7b shown the development of a vortex with an initial wind field whose low level barotropic region is essentially twice as "deep" as the control simulation. We see again that saturation of the inner core, a burst of convective activity, and a small jump in 5km PVS precede genesis. An even more "extreme" genesis transition is created when we start with a wind field that is more shallow (about half the depth) of the control simulation (Figure 6c, 7c). In this case, the surface vortex is clearly weakening up to the moment of genesis, and we again see the saturation of the core, increases in convective activity, and increases in PVS at 5km in the hours before genesis.

Simulations with a "pre-moistened" atmosphere with RH = 90% up to z = 12 km were also intriguing. A 10 m/s vortex elevated such that its surface wind speed was only 1 m/s was able to transition to a tropical cyclone in closely related.

ACKNOWLEDGEMENTS:

I would like to acknowledge the contribution of Mr. David Kofron who performed preliminary analyses of these simulations. This work was supported by NOAA through an ongoing collaboration with CIMAS and the University of Miami. I would like to thank Prof. Shu-hua Chen for developing the code to incorporate idealized initial conditions into the WRF model, the WRF developers for their ongoing assistance, and Dr. Eric Rappin for his comments on this document.

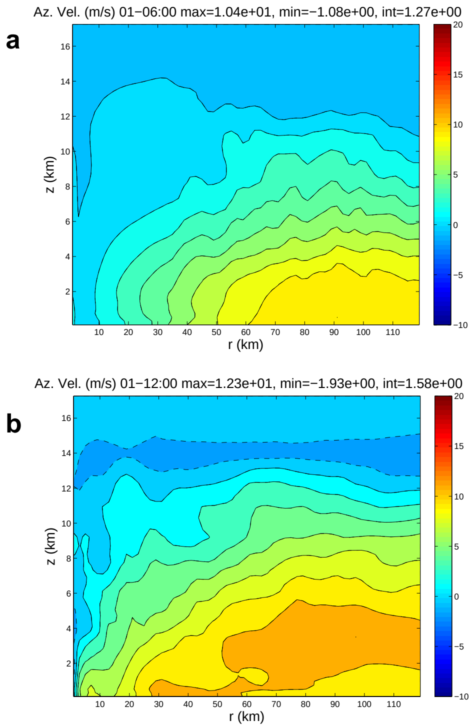

Figure 8: Azimuthally averaged azimuthal wind fields for a simulation with the same initial vortex as the control simulation but with the RH set to 90% from z = 0 to 12 km, a) t = 6 h; b) t = 12 h.

this environment, through the same processes identified above (not shown). However, a pre-moistened simulation with the same initial vortex as the control appeared to bypass the mid-level vortex phase of development, with a small-scale surface vortex appearing within 12 hours, as shown in Figure 8. This evolution is quite similar to that seen in the control simulation of Montgomery et al. (2006) (see their figure 4; their atmosphere was also "pre-moistened" but not to the same degree), and suggests that in "extremely favorable" conditions the mid-level, cold-core vortex phase required by Bister and Emanuel (1997) in not necessary.

7. CONCLUSIONS

These high-resolution simulations of tropical cyclogenesis suggest that a number of elements must come together before "genesis" and rapid intensification can begin. The middle and upper levels of vortex core must become nearly saturated; a coherent mid-level vortex must develop and contract; and there must also be a surface circulation, though it does not need to be very strong. When the atmosphere is "pre-moistened," perhaps to an unrealistic degree, the mid-level vortex phase may be bypassed. Genesis occurs after one or series of strong convective bursts that occur after near-saturation has been achieved. Genesis is also preceded by a sudden rise in the number and/or intensity of mid-level vorticity anomalies. Of course, these last two processes are closely related.

REFERENCES:

Bister, M., and K. Emanuel, 1997: The genesis of Hurricane Gillermo: TEXMEX analysis and modeling study. *Mon. Wea. Rev.*, **125**, 2662-2682.

Davis, C. A., and L. Bosart, 2001: Numerical simulations of the genesis of Hurricane Diana (1981). Part I: Control simulation. *Mon. Wea. Rev.*, **129**, 1859-1881.

Davis, C. A., and L. Bosart, 2002: Numerical simulations of the genesis of Hurricane Diana (1981). Part I: Sensitivity of track and intensity prediction. *Mon. Wea. Rev.*, **130**, 1100-1124.

Hendricks, E. A., M. T. Montgomery, and C. A. Davis, 2004: On the role of "vortical" hot towers in tropical cyclone formatioin. *J. Atmos. Sci.*, **61**, 1209-1232.

Hong, S.-Y., J. Dudhia, and S.-H. Chen, 2004: A revised approach to ice microphysical processes for the parameterization of clouds and precipitation. *Mon. Wea. Rev.*, **132**, 103-120.

Jordan, C. L., 1958: Mean soundings for the West Indies area. *J. Meteor.*, **15**, 91-97.

Montgomery, M. T., M. E. Nicholls, T. A. Cram, and A. B. Saunders, 2006: A vortical hot tower route to tropical cyclogenesis. *J. Atmos. Sci.*, **63**, 355-386.

Noh, Y., W.-G.Cheon, S.-Y. Hong, and S. Raasch, 2003: Improvement of the K-profile model for the planetary boundary layer based on large eddy simulation data. *Bound.-Layer Meteor.*, **107**, 401-427.

Nolan, D. S., and L. D. Grasso, 2003: Nonhydrostatic, three-dimensional perturbations to balanced, hurricane-like vortices. Part II: Symmetric response and nonlinear simulations. *J. Atmos. Sci.*, **60**, 2717-2745.

Nolan, D. S., M. T. Montgomery, and L. D. Grasso, 2001: The wavenumber-one instability and trochoidal motion of hurricane-like vortices. *J. Atmos. Sci.*, **58**, 3243-3270.

Ritchie, E. A., and G. Holland, 1997: Scale interaction during the formation of Typhoon Irving. *Mon. Wea. Rev.*, **125**, 1377-1396.

Reasor, P. D., M. T. Montgomery, and L. F. Bosart, 2005: Meoscale observations of the genesis of Hurricane Dolly (1996). *J. Atmos. Sci.*, **62**, 3151-3171.

Simpson, J., E. Ritchie, G. J. Holland, J. Halverson, S. Stewart, 1997: Mesoscale interactions in tropical cyclone genesis. *Mon. Wea. Rev.*, **125**, 2643-2661.