SOME OBSERVATIONAL AND THEORETICAL PROCESSES RELATED TO RAPIDLY INTENSIFYING TROPICAL CYCLONES

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1. INTRODUCTION

The largest errors in tropical cyclone intensity forecasting are usually the result of rapid intensification (RI) events. Holliday and Thompson (1979) have defined rapid intensification as a change in minimum central pressure of at least 42 mb in a 24 hour period. When forecast intensity errors for all storms are compared to forecast intensity errors for rapid intensifiers (as shown in Table 1), it is obvious that there is a lower percentage of accurate forecasts for RI storms (Mundell, 1990). Table 1 shows that the mean 24 hour error for forecasting RI's is nearly twice as large as the non-RI forecast errors. The strong negative bias of

Table 1: Comparison of JTWC 24 hour forecast errors (in m s^{-1}) for a period of rapid intensification versus all intensity forecast errors for the 1980-85 seasons (from Mundell).

Class	Mean	Bias	St. Dev.	Count
Rap Int (72-87)	12.4	-9.7	11.2	468
All Storms (80-85)	6.4	-0.3	8.5	2891

-9.7 m s^{-1} reflects an inability to accurately predict RI change, resulting in under-forecasting of tropical cyclone intensity just 24 hours later.

Two recent storms epitomize the lack of skill in forecasting RI events. One is the unanticipated strengthening of Typhoon Omar from 8/27/12Z to 8/28/12Z as it approached Guam. The Joint Typhoon Warning Center issued a forecast for Omar to increase in this time period from 70 Kts to 85 Kts. Instead, Omar increased to 115 Kts ($\Delta p \approx 45$ mb). The other cyclone was Hurricane Andrew which devastated south Florida. As shown in Fig. 1, the 24 hour forecast valid for 8/23/18Z (23.75 on graph) was an increase from 80 kts to 95 kts; unfortunately, the storm increased to 135 Kts ($\Delta p = 47$ mb). Andrew did weaken a little before making landfall on 8/24/09Z, and subsequent forecasts improved, but the storm's rapid strengthening had not been anticipated.

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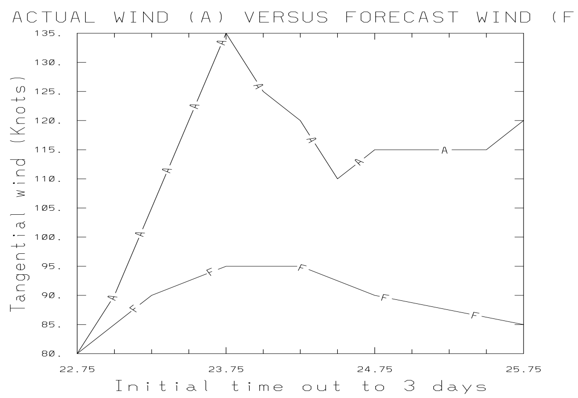

Figure 1: Plot of forecast winds initialized at 8/22/18Z out to 8/25/18Z (labeled F) versus actual winds (labeled A).

Obviously, further research in hurricane intensity change is needed. This paper addresses some of the theoretical aspects of hurricane intensity change and offers some observational evidence supporting these ideas.

2. THEORY

Most current intensity change research deals with the interplay of outflow channels such as a tropical upper tropospheric trough (TUTT) or a mid-latitude trough. Sadler (1978) noted the proper positioning of outflow channels with vigorous TUTT cells may encourage RI by efficient removal of mass and heat from the eyewall region, especially for a double outflow situation. Chen and Gray (1986) also found that double outflow channels were associated with the largest intensification rates. Using an axisymmetric model, Challa and Pfeffer (1980) showed that outflow jets with eddy angular momentum flux convergence ($-r^{-2}\partial/\partial r(r^2\overline{u_r'v_t'})$, a quantified measurement of trough interaction) can enhance the intensification process. They postulated that eddy flux convergence (EFC) provides a mechanism to remove high enthalpy air from the system due to the jets low inertial stability. Molinari and Vollaro (1989) made a similar argument of Hurricane Danny's intensification upon its encounter with a mid-latitude trough.

However, some researchers have questioned these ideas. Shapiro and Willoughby (1982) could not reproduce Challa and Pfeffer's result using their version of an axisymmetric model. Merrill (1988) found no significant relationship between EFC and intensity change in a five year composite of Atlantic hurricanes. DeMaria and Kaplan (1991) found that EFC's have no correlation with intensity change (except at 48 hours, perhaps indicating a time lag) and suggested that other environmental factors, such as SST's and vertical wind shear are more important.

In fact, many tropical cyclones (including RI cases) develop without any trough interaction. Andrew's period of rapid deepening occurred when it passed under a region of low vertical shear (a ridge). In fact, Shea and Gray (1973) demonstrated that the most intense tropical cyclones have minimum shear at the radius of maximum wind. And while it may be true that intensification corresponds to increased outflow activity, it can be argued that these jets are the *result of the intensification* in many cases, not the cause. In a region of minimum shear, multidirectional outflow will follow as a natural consequence, regardless of whether or not strong channels initially existed. Indeed, tropical cyclone axisymmetric model simulations, which inherently experience no environmental influence, readily produce stronger outflow jets during intensification. In situations where a TUTT is located poleward, it can be argued that it creates a region of minimum ventilation over the cyclone in the transition from easterly winds aloft on the equatorward side to westerly winds on the poleward side.

Incorporated into these theories is an explanation for the following paradox: convection is needed near the center for intensification, but the release of latent heat aloft stabilizes the atmosphere. Somehow, this liberalization of heat in the eyewall cloud must be removed for intensification to continue. This warming paradox is a key to RI cases. Merrill (1988) showed that intensifying storms were cooler aloft in the eyewall than non-intensifying storms. Fig. 2 shows temperature deviations along a constant pressure surface for nine hurricanes' inner core (5-50 n.m.) in the upper troposphere. These measurements were taken by upper tropospheric aircraft penetrations in the 1950's and 1960's.

Figure 2 shows that cyclones with the lowest surface pressure tend to develop strong temperature gradients from the eye to eyewall cloud. These data suggest that baroclinicity is a key factor to explain why deep cumulus convection can be maintained in the eyewall cloud as the eye undergoes strong warming.

This baroclinicity can be related to vorticity and vertical wind shear using the thermal wind equation in cylindrical coordinates

$$(f + \frac{2v_t}{r})\frac{\partial v_t}{\partial p} = -\frac{R}{p}\frac{\partial T}{\partial r} \qquad (1)$$

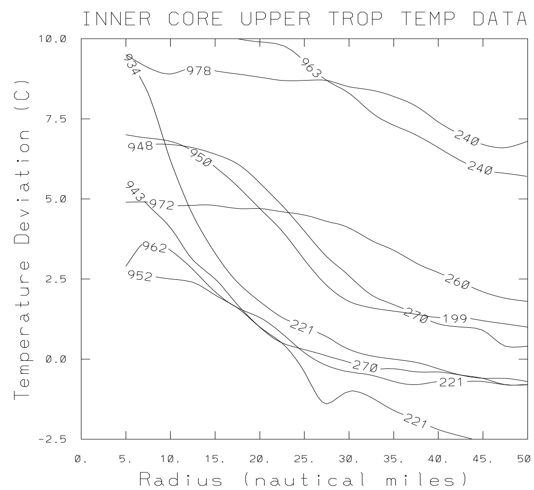

Figure 2: Inner core upper tropospheric temperature anomaly from the mean summer tropical atmosphere as measured by National Hurricane Research Laboratory reconnaissance flights. Numbers on left side are surface pressures, and on right are pressure flight levels.

which Gray (1967) has simplified to

$$WS = B$$

where W is an "inertial parameter," S is a vertical shear parameter, and B is a horizontal baroclinicity term.

It is our belief that inner-core processes are often more important to tropical cyclone intensification than outer core processes, including trough interaction. Most researchers have failed to differentiate between the inner core (0-1° radius) and outer core (1-2.5° radius) wind spin-up. Weatherford and Gray (1988) have shown that during the RI process the inner core and outer core winds are often decoupled – the inner core winds increase much faster than the outer core winds during the intensification process. An RI event can be hindered if the "outer core wind strength," or inertial stability, is too large. They showed that RI cases typically exhibit weaker outer core winds than do non-RI cases because strong outer core winds restrain momentum import into the center of the storm. In other words, strong outer core winds will weaken eyewall convection and enhance outer core convergence, impeding the intensification process. Mundell (1990) used a pixel count satellite routine of cloud tops to show that rapid intensification is preceded by deep eyewall convection and weak convection outside 2°.

It is hypothesized that RI cases can occur independent of environmental influences such as trough interaction, and instead are most dependent on vertical wind shear. Troughs and double outflow channels are often indicators of minimum shear over the system. Assuming low vertical wind shear is present, then inner core updrafts transport cyclonic momentum vertically to the upper circulation. For RI cases with vigorous inner core updrafts, this upper level cyclonic circulation should extend almost to the tropopause. In theory, the vertical shear of v_t up to 200 mb (S) would be relatively small. Assuming that the wind velocity goes to zero at the tropopause, most of the vertical wind shear of v_t would be confined to near the tropopause.

A concentration of v_t at the top of the troposphere has important thermal consequences. According to Eq. (1), thermal wind balance is achieved by an increase in B above the region of maximum shear. In a more common type of intensification, (top of Fig. 3) the inner core warming outside the eye is located in the middle troposphere where the vertical wind shear is greatest. Since convection acts to warm and stabilize the upper troposphere, this over time tends to reduce further deep convection through vertical stabilization.

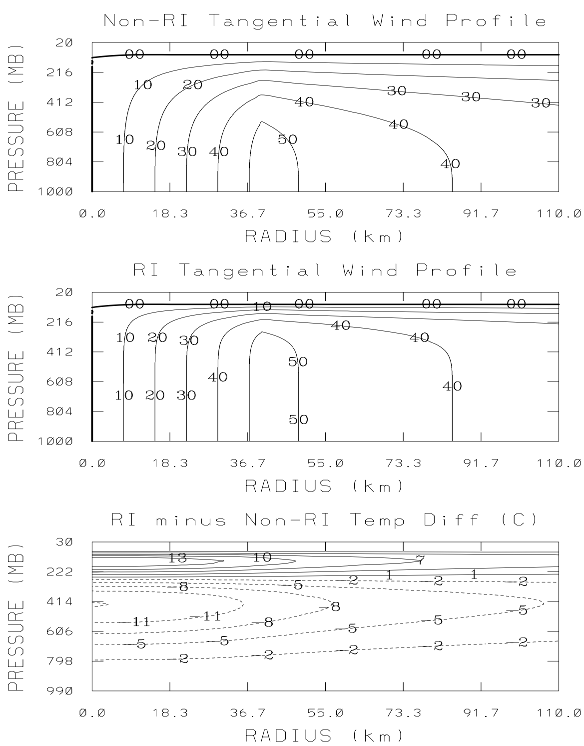

Figure 3: Tangential wind profile for a high value of $\partial v_t/\partial p$ (top figure), low value of $\partial v_t/\partial p$ (middle figure), and ΔT between the two $[T(\mathrm{RI}) - T(\mathrm{non-RI})]$.

The middle diagram of Fig. 3 corresponds to a hypothetical RI case. It is postulated that rapidly deepening storms limit the stabilizing effects of deep convection by concentrating the vertical shear of v_t near the tropopause so that the warming is as high and as close to the center as possible. The bottom diagram of Fig. 3 shows the temperature difference for these two cases, confirming that the RI storm is generally cooler than the non-RI case except near the tropopause.

In addition to the thermodynamic aspects of this problem, an increase in upper level eyewall vorticity can also act to dynamically intensify the storm. Hack and Schubert (1986) showed in a 2D model that such a high level wind profile acts to increase the ability of the storm to convert latent heat into kinetic energy since the Rossby radius of deformation has been decreased.

3. OBSERVATIONS

Figure 4 shows the inner core tangential winds corresponding to the storms in Fig. 2. In general, the most intense storms contain the strongest winds, and are located at smaller radiuses. This follows from our previous thermal wind discussion.

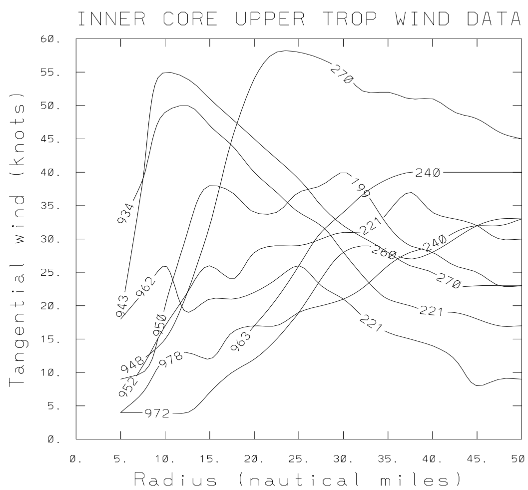

Figure 4: Inner core upper tropospheric tangential winds as measured by National Hurricane Research Laboratory reconnaissance flights.

Figure 5 shows the average inner core wind profile for storms where the change in surface pressure is $\Delta p \geq 35$ mb/day, and storms where $10 \leq \Delta p \leq 20$ mb/day at the beginning of their pressure drop. Note that the faster developing case is less sheared, with higher upper-tropospheric v_t. It is also interesting that the winds are the same near the surface. The faster developing case is 7 C cooler in the upper troposphere (figure not shown), consistent with the previous thermal wind discussion.

These results appear to offer forecast potential, since there seems to be a time lag between the increase in upper-level vorticity in the eyewall and the central pressure fall. Should high tangential winds be observed in the upper eyewall, one may be able to predict that future rapid deepening will occur. For example, at the beginning of Supertyphoon Flo's 35 mb pressure drop (in 24 hours), 90 kt inner core winds were observed at 200 mb in the eyewall (Fig. 6). Surface winds at that time were 115 kts, indicating a low $\partial v_t/\partial p$. The 200 mb baroclinicity is evident by the 9° C

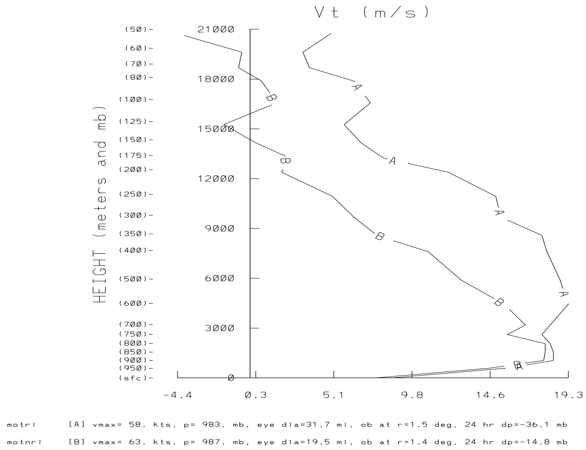

Figure 5: Average inner core ($0° \leq r \leq 2°$) v_t profiles for faster developing (A) and slower developing (B) storms at beginning of pressure drop.

temperature decrease from the eye to the eyewall, consistent with the reconnaissance plane observations discussed previously on intense storms.

This report indicates that it would be very beneficial to have upper tropospheric aircraft reconnaissance in critical intensity change forecast situations. This has been suggested by Gray (1993) in another paper of this conference.

Figure 6: Measured inner core flight level winds of Supertyphoon Flo at 200 mb on 0720 UTC 16 September 1990.

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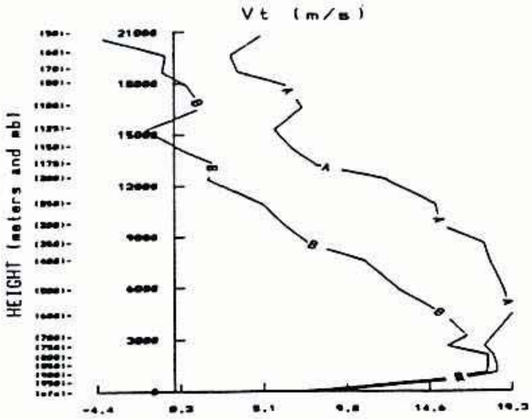

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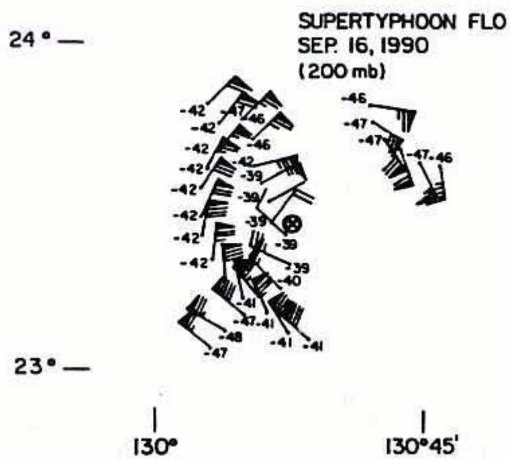

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