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DYNAMICS AND THERMODYNAMICS OF TROPICAL CYCLOGENESIS

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1. INTRODUCTION

It has been known that tropical cyclones (TCs) develop from preexisting finite amplitude disturbances that exhibit positive vertical vorticity, such as MCSs, tropical waves, monsoons. These preexisting disturbances exhibit small inertial stability and therefore processes of probabilistic nature play important roles in TC formation. These processes are much less important with mature phase of a tropical cyclone, when the system exhibits large inertial stability. This makes modeling tropical cyclogenesis far more challenging then modeling mature tropical cyclones. Keeping this in mind plus the fact that there are not enough data of pre-depression tropical disturbances, it is not a surprise that today genesis is the least understood phase of a TC life cycle. Many such preexisting disturbances form during the hurricane season, but only small percentage of those develop into TCs, even if the environment is favorable for genesis (Grey, 1968).

Today there exist two types of hypotheses of tropical cyclogenesis. Dunkerton et al. (2009) hypothesized that the genesis is initiated within the westward moving tropical waves. The idea is that the waves provide sheltered region of low-level cyclonic vorticity where continuous convection is enabled, which further increases the cyclonic vorticity. Ultimately the wave-convection interaction leads to vortex intensification at low levels first, and subsequently at mid and upper levels. This hypothesis represents the so-called "bottom-up" development. The other type of hypothesis postulates a "top-down" development. There has been observational evidence of disturbances that first became cold-core vortices (featured strong mid-level vortex), and then transitioned to warm-core vortices (Harr and Elsberry, 1996; Bister and Emanuel, 1996; Raymond et al., 1998, Raymond and López, 2011; Davis and Ahijevych, 2011, and others). The question that arises here is how the mid-level vortex encourages vorticity intensification near the surface. Bister and Emanuel (1996) and others have hypothesized that the mid-level vortex is being "pushed" downwards by downdrafts, but this violates the laws of fluid dynamics. Raymond et al. (2011), suggest that it is via the thermodynamic state of the atmosphere that the mid-level vortex produces.

In the present paper we analyze real data from developing and non-developing disturbances that were observed during the 2010 hurricane season in the Northern Atlantic and the Caribbean. The results show support for top-down development, that establishment of a mid-level vortex precedes genesis. We are also exploring possible correlations between various dynamic and thermodynamic variables. For that part of the analysis we include data gathered in the West Pacific during the 2008 hurricane season. Based on the results we propose a chain of events which occurs during tropical cyclogenesis.

2. DATA AND METHODS

Two data sets are used for our analysis. One is dropsonde data gathered during the field campaign Pre-Depression Investigation of Cloud systems in the Tropics (PREDICT) that took place during the period August-September of 2010. Target areas were the North Atlantic and the Caribbean. NCAR/NSF's G-V aircraft deployed approximately 600 dropsondes in pre-storm disturbances from altitudes $11-13$ km. Twenty-six missions were conducted. Eight disturbances were observed, four of which developed into tropical storms. The other data set was gathered during TPARC/TCS08 that took place in the period August-September 2008. NRL's P-3 and two WC-130 aircraft deployed dropsondes over the West Pacific. See Raymond and López (2011) for more information on these data. The quality control on both data sets was done by EOL.

The horizontal resolution of the dropsonde data is coarse. On average there were launched 20 dropsondes during one mission and they were spread out quasi-equidistantly over $5^o \times 5^o$ longitude-latitude area. We apply a three dimensional variational scheme (3D-Var) to these data. Detailed description of the 3D-Var technique used in the present work is given by López and Raymond (2011) and Raymond and López (2011). For

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each mission we obtain a 3D data file which is assumed to be a snapshot at a given reference time, usually taken halfway through the mission. Dropsonde positions are shifted using a Galilean transformation to the positions at the reference time, using the translational velocity of the disturbance. The snapshot assumption is adequate for studying mesoscale processes.

3. RESULTS

Figure 1 shows mean vertical profiles of three disturbances observed during PREDICT: Gaston, Karl and Matthew. The vertical profiles are obtained by horizontal averaging over area that we choose subjectively to best represent the overall disturbance at the time of observation. We make the choice based on the location of the circulation center and the saturation fraction distribution. Gaston was in a decaying phase as it was downgraded from a TS to a TD shortly prior the first mission into it and it never managed to redevelop. The decrease was attributed to dry Saharan air intrusion that was surrounding Gaston at that time (Davis and Ahijevych, 2011). Our analyses show misaligned circulation centers at different elevations (not shown). The circulation center at 5 km was shifted for more then a degree northward of the circulation center at 1 km, which implies strong vertical wind shear and therefore further dry air intrusion and subsequent decay of Gaston. Gaston 2 featured much weaker mid-level vortex already. After this the vorticity kept decaying at all levels (Gaston 3, 4, and 5). Six missions were conducted in Karl in a 5-day period. Karl started with a bottom-heavy vorticity profile (Karl 1 and 2). Then vorticity started increasing at mid levels while decreasing at low levels, so that Karl 4 featured very strong mid-level vortex. The low-level vortex started intensifying subsequently. Karl 5 exhibited a vorticity maximum at mid levels and it was classified as a low. One day later the vorticity had intensified at all levels (Karl 6) and Karl was officially a tropical storm. Matthew also developed mid-level vortex prior genesis. Matthew 3 was a low, while Matthew 4 was already a TS.

Evidently, these results support the top-down TS development. In order to address the question of how a mid-level vortex promotes a low-level vortex development/intensification we have examined possible relations of the mid-level vorticity to other dynamic and thermodynamic variables. Scatter plots between various variables using data from both PREDICT and TCS08 are given in Fig. 3. Each data point in these plots represents a single mission. The mid-level vorticity is calculated as an average in the layer between 3 and 5 km. The instability index is a measure of the atmospheric instability to moist convection. Larger instability index is associated with more unstable atmosphere. The saturation fraction is a proxy for vertically integrated relative humidity. The low-level vorticity tendency is an average over the lowest 1 km, and the mass convergence is averaged over the lowest 3 kilometers. See the appendix for definitions on the parameters.

4. DISCUSSION

Based on our results on the vertical structure evolution of the developing and the non-developing disturbances, and the suggested correlations between certain variables in the scatter plots we propose the following chain of events for tropical cyclogenesis:

The necessary ingredient is a well-developed mid-level vortex. This vortex maintains negative temperature perturbations in the lower and positive temperature perturbations in the upper troposphere (Fig. 2), which results in more stable thermodynamic stratification. The negative slope between the mid-level vorticity and the instability index in Fig. 3a implies that this is the case indeed. Such a troposphere is conducive to shallower convection that produces mass flux with a peak at lower elevations. Thus, the largest gradient of the vertical mass flux occurs in a shallow layer adjacent to the sea surface. This means that mass has been displaced vertically upwards from this layer and mass continuity then implies horizontal mass convergence and thus low-level vorticity convergence. This is reflected in a negative correlation between the instability index and low-level mass convergence (Fig. 3f), and also between the instability index and low-level vorticity tendency (Fig. 3e). As most of the water vapor is contained near the surface, moisture is also converging. The scatter plots in Fig. 3b and Fig. 3d support this picture. If the mid-level vorticity exists long enough, maintaining the thermodynamics for continuous shallower convection, i. e. for continuous low-level convergence of vorticity and water vapor import, the low-level wind speed will eventually reach the tropical storm threshold. In the Gaston case, the mid-level vorticity was destroyed too early and this is probably why it did not manage to redevelop.

5. REFERENCES

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7. ILLUSTRATIONS AND TABLES

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6. APPENDIX

Vorticity tendency equation:

$$\frac{\partial \zeta_z}{\partial t} = -\nabla_h \cdot (\zeta_z \vec{v}_h - \vec{\zeta}_h v_z + \hat{k} \times \vec{F}). \qquad (1)$$

Here, $\vec{v} = (\vec{v}_h, v_z)$ is the storm-relative velocity, $\vec{\zeta} = (\vec{\zeta}_h, \zeta_z)$ is the absolute vorticity, $\vec{F}$ is a force due to surface friction. Subscript h denotes horizontal components. The baroclinic term is omitted in our calculations, as in the tropics this term is insignificant in comparison to the terms kept in the equation.

The instability index is defined as:

$$\Delta s^* = s^*_{low} - s^*_{high}, \qquad (2)$$

where s^*_{low} is the virtual saturated moist entropy averaged over the area and over the layer between 1 to 3 km, and s^*_{high} is the virtual saturated moist entropy averaged over area and over the layer between 5 to 7 km. And the saturation fraction is calculated as follows:

$$SF = \frac{\int_0^h \rho r dz}{\int_0^h \rho r_{sat} dz}. \qquad (3)$$

Here ρ is the air density, r and r_{sat} are the mixing ratio and saturated mixing ratio, and h is the height of the observational domain.

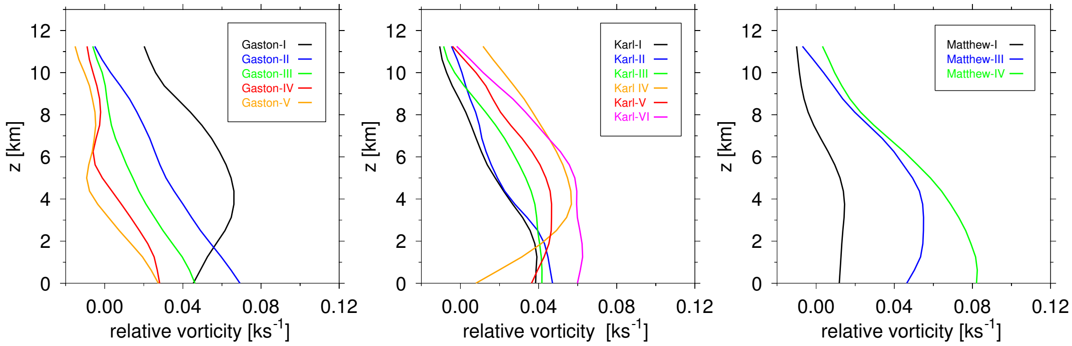

Figure 1: Area-averaged relative vorticity for Gaston, Karl, and Matthew.

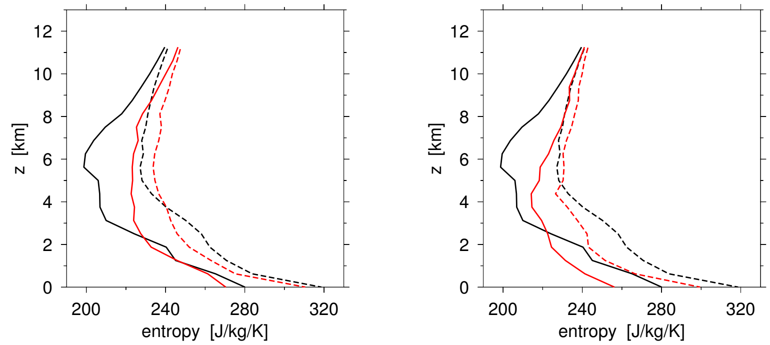

Figure 2: Vertical profiles of moist entropy (solid lines) and saturated moist entropy (dashed lines). The black lines in both panels are for the non-developer Gaston 5. The red lines are for the developers Karl 5 (left panel) and Matthew 3 (right panel).

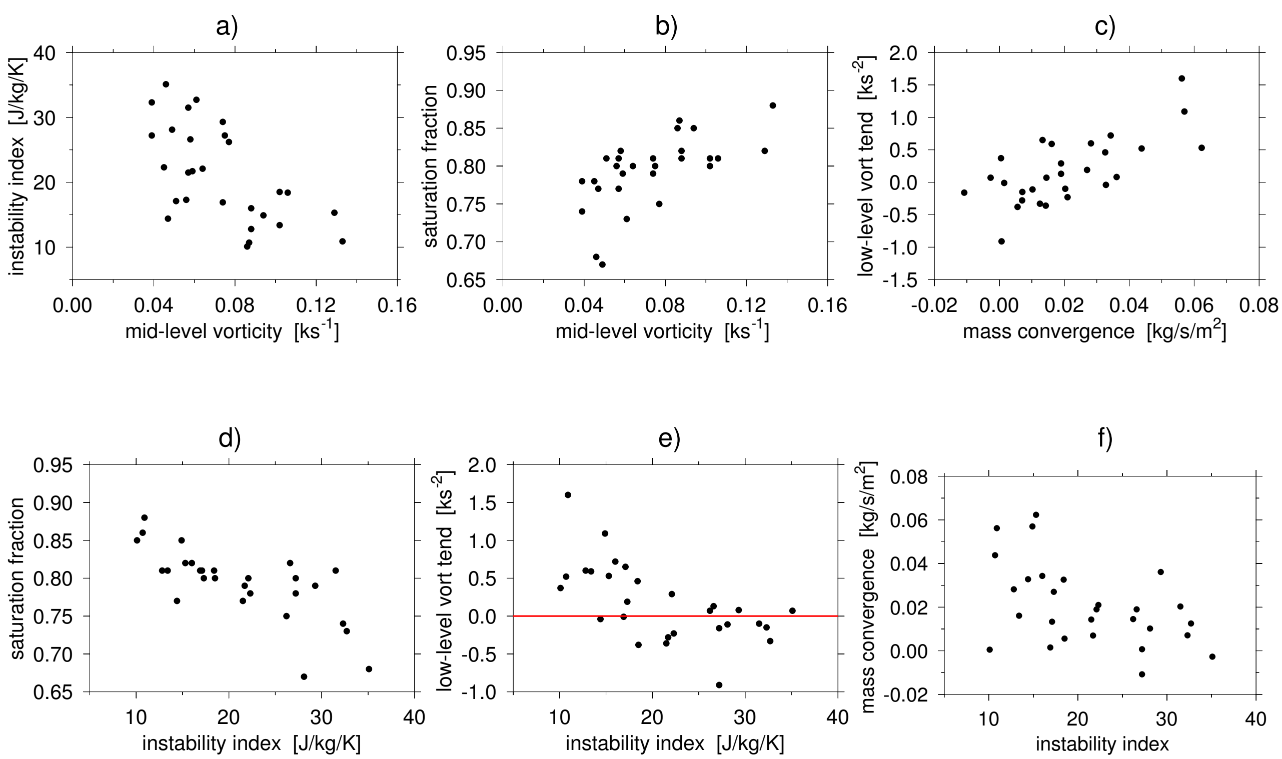

Figure 3: Scatter plots.