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Increased sensitivity of tropical cyclogenesis to wind shear in higher SST environments

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[1] A new method for evaluating the sensitivity of tropical cyclone (TC) genesis to environmental parameters involves the simulation of tropical cyclone development with a cloud-resolving model in environments of radiative-convective equilibrium (RCE) generated by the same model. This method is extended to allow for the incorporation of mean wind shear into the RCE states, thus providing much more realistic and relevant simulations of TC genesis. The ''finite-amplitude'' nature of tropical cyclogenesis is reproduced, with cyclogenesis resulting only when the initial vortex strength is sufficient, which in turn depends on the environmental parameters. For fixed thermodynamic parameters, the required initial vortex strength necessary for genesis increases with the mean wind shear. However, an unexpected result has been obtained, that increasing sea surface temperature (SST) does not allow TC genesis to overcome greater shear. In fact, the opposite trend is found, that shear is more effective in suppressing TC genesis when the SST is higher. This increased sensitivity can be explained by several factors, such as the higher altitude of the developing mid-level vortex, stronger downdrafts, and increased static stability, all of which allow the shear to be more effective in disrupting the developing cyclone. **Citation:** Nolan, D. S., and E. D. Rappin (2008), Increased sensitivity of tropical cyclogenesis to wind shear in higher SST environments, *Geophys. Res. Lett.*, *35*, L14805, doi:10.1029/2008GL034147.

1. Introduction

[2] In a recent study, *Nolan et al.* [2007] used a high-resolution (4 km), cloud-resolving model to simulate tropical cyclone (TC) genesis in environments of radiative-convective equilibrium (RCE) generated by the same model. These environments of RCE were defined by the sea surface temperature (SST), Coriolis parameter (f), and the mean surface wind (u_m). *Nolan et al.* [2007] found a very close correlation between the rate of TC development from a weak, mid-level vortex and the maximum potential intensity (MPI) determined by the SST and the overlying atmospheric sounding as computed from the theory of *Emanuel* [1995]. *Nolan et al.* [2007] found little sensitivity to other seemingly relevant parameters such as the Coriolis parameter, the mid-level relative humidity (RH), or the convective available potential energy (CAPE).

[3] An important limitation of that study was that the environments of RCE were completely free of vertical wind shear. Vertical shear of the horizontal wind is widely known to be a major controlling factor on the likelihood of TC genesis, both on seasonal and synoptic time scales [*Gray*, 1968; *Frank*, 1987]. *Nolan et al.* [2007] speculated that the absence of even a small amount of wind shear was the reason for the lack of sensitivity to RH and f in their simulations.

[4] Here, we present an extension of the approach of *Nolan et al.* [2007] that allows the specification of a prescribed wind as a function of height that remains approximately constant across the domain over the course of the simulation. This technique is used to generate states of RCE with shear, and these sheared RCE states are used as background conditions for simulations of TC genesis. The ''finite-amplitude'' nature of TC genesis, such that an initial disturbance of a sufficient strength is required to make the transition to a developing TC, is reproduced and used as a measure of the favorability of the environments for genesis. Furthermore, we find that as the SST underlying the RCE environment increases, the effect of shear in suppressing TC genesis actually increases. This is contrary to expectations, because it is widely assumed that factors favoring TC genesis can be considered separately and are additive or multiplicative. This assumption is fundamental to the genesis parameters of *Gray* [1968], *DeMaria et al.* [2001], and *Emanuel and Nolan* [2004], as well statistical hurricane intensity forecast models such as SHIPS [*DeMaria and Kaplan*, 1999].

2. Model and Simulation Design

2.1. Mesoscale Model and Radiative-Convective Equilibrium

[5] Except for the inclusion of mean wind shear the methodology is nearly identical to that described by *Nolan et al.* [2007]. The Weather Research and Forecast Model (WRF) version 2.1.2 [*Skamarock et al.*, 2005] is used to simulate environments of RCE on the doubly periodic f-plane, with 4 km horizontal grid spacing and 40 vertical levels equally spaced in pressure coordinates. The Yonsei University (YSU) planetary boundary layer [*Noh et al.*, 2003], WRF 6-class microphyiscs [*Hong and Lim*, 2006], Rapid Radiative Transfer Model (RRTM) longwave [*Mlawer et al.*, 1997], and Goddard shortwave [*Chou et al.*, 1998] parameterizations are used.

[6] To generate the RCE states, small domains are used with 50 × 50 horizontal grid points, with random convection simulated for 90 days. The spatial and temporal mean for 6-hourly data over the last 30 days are used as the initial environment for large-domain simulations of TC genesis.

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2.2. Mean Wind Shear on the Doubly-Periodic *f*-Plane

[7] Mean wind shear is the result of horizonal gradients in vertically averaged temperature. However, a mean temperature gradient cannot be defined in a periodic environment. To include the effects of wind shear on the RCE states and on TC genesis, we initialize the background state with a mean zonal flow as a function of pressure, $U(p)$. To balance this flow, we add terms to the equations which replicate the pressure gradient force that would exist if the temperature gradient were present, i.e.,

$$\frac{Dv}{Dt} + fu = -\frac{\partial \Phi}{\partial y} + fU(p) \tag{1}$$

By maintaining this artificial pressure force throughout the simulation, the desired mean wind shear is maintained throughout the domain. However, the local effects of convection and TC development can locally modify the mean flow, just as would occur in the atmosphere.

[8] It is important to acknowledge what is being neglected with this method: 1) the mean temperature gradient, and 2) adiabatic ascent and descent on sloped potential temperature surfaces, which might result in significant temperature changes for parcels moving on sloped potential temperature surfaces in a mid-level vortex. For a mean 850 to 200 hPa shear of 5 ms^{-1}, the horizontal temperature gradient amounts to 0.6 C per 1000 km; for the same shear, the adiabatic ascent and descent of parcels at the 500 hPa level in a 200 km diameter vortex would result in a change in temperature of 0.2 C. Thus we expect the kinematic effects of the shear to dominate over temperature advection and adiabatic ascent.

[9] For most of the simulations, a simple shear profile is used that has a linear increase in U in log-pressure height, increasing from a specified surface wind u_m:

$$U(p) = (u_m + C_1) + (C_2 U_{shear})[5 - \log_{10}(p)] \tag{2}$$

The constants in (2), are chosen so that U_{shear} is the zonal wind speed difference between the 850 and 200 hPa pressure levels, and u_m is the mean surface wind.

2.3. Simulations of TC Genesis

[10] Mean soundings from the RCE states are used as the background state for simulations of TC genesis. The genesis simulations use 300 × 300 grid points with the same resolution and parameterizations as the RCE simulations, and are initialized with a balanced, axisymmetric vortex modeled after observed precursors to TC genesis [e.g., *Raymond et al.*, 1998]. The azimuthal wind profiles have Gaussian vorticity profiles, with a radius of maximum winds (RMW) of 100 km, and specified values for the maximum azimuthal wind $v_{\max}$. The wind field also varies in the vertical, with a Gaussian decay both above and below $z_{\max}$ = 3720 m. The vertical decay parameter is set so that the initial maximum surface wind is exactly half of $v_{\max}$.

3. Results

3.1. Properties of RCE States With Mean Wind Shear

[11] For each RCE state we compute the convective available potential energy (CAPE), the downdraft CAPE (DCAPE [see, e.g., *Emanuel*, 1994, section 6.3], the MPI in terms of wind speed (V_{pot}) using the method of *Emanuel* [1995], and the genesis index of *Emanuel and Nolan* [2004], defined as

$$GP = \left|10^5 \eta\right|^{3/2} \left(\frac{H}{50}\right)^3 \left(\frac{V_{\text{pot}}}{70}\right)^3 (1 + 0.1 U_{\text{shear}})^{-2} \tag{3}$$

where η is the environmental absolute vorticity, H is the 600 hPa relative humidity, and V_{pot} and U_{shear} have been defined above.

[12] Values of CAPE, DCAPE, V_{pot}, and GP are shown in Table 1 for environments with U_{shear} = 5 ms^{-1} and varying values of SST and u_m. Variations in u_m can have large effects on both CAPE, DCAPE, and V_{pot}. Increasing the mean surface wind leads to more moisture in the boundary layer, and thus higher values of CAPE. As u_m increases from 1 to 3 ms^{-1}, CAPE more than doubles. However, CAPE does not increase again for u_m = 5 ms^{-1}, a result which is quite different from RCE without shear [*Nolan et al.*, 2007].

[13] As u_m increases, increasing boundary layer moisture leads to stronger convection, which then increases the temperature of the entire atmosphere, leading to a decrease in V_{pot}. Increasing SST or u_m also leads to an increase in the potential intensity of downdrafts, as indicated by the values of DCAPE.

3.2. Sensitivity of TC Genesis and Development to SST, Mean Wind, and Mean Wind Shear

[14] Figure 1 shows minimum surface pressure and maximum 10 m wind speed for simulations with f = 5.0 × 10^{-5} s^{-1}, SST = 30 C, u_m = 3.0 ms^{-1}, $v_{\max}$ = 10 ms^{-1}, and values of U_{shear} = 0, 2.5, 5.0, 7.5, and 10 ms^{-1}. As the shear increases, genesis is delayed and the intensification is slowed. The results for the "control case" with U_{shear} = 5 ms^{-1} are particularly interesting, with the minimum surface pressure nearly constant during the first 4 days, followed by intensification to hurricane strength. Genesis fails for U_{shear} = 7.5 ms^{-1} and 10 ms^{-1}.

[15] The finite amplitude sensitivity is further illustrated in Figure 2, which shows minimum surface pressures for simulations with varying values of the initial $v_{\max}$. Figure 2a shows the results for the control case and for a simulation with $v_{\max}$ = 7.5 ms^{-1}. This slightly weaker vortex fails to form a tropical cyclone. Evidently the finite-amplitude threshold for this environment lies between somewhere between $v_{\max}$ = 7.5 and 10 ms^{-1}.

[16] Table 1 shows that decreasing u_m leads to substantial increases in V_{pot} and GP, indicating more favorable environments. Figure 2b shows the results of simulations with varying values of $v_{\max}$ in an environment with u_m = 2 ms^{-1}, for which V_{pot} increases to 63.5 ms^{-1}. TC genesis and intensification occur in just a few days for $v_{\max}$ = 10 and 7.5 ms^{-1}, and after five days for $v_{\max}$ = 5 ms^{-1}. For $v_{\max}$ = 2.5 ms^{-1}, there are signs of the pressure beginning to fall after 5 days, but the very long incubation time before development makes genesis essentially irrelevant for this case. Nonetheless, by decreasing the mean surface wind of the RCE environment, the finite amplitude threshold for genesis has been lowered substantially. Similarly, increasing the mean surface wind makes the environment less favorable for

Table 1. Values of CAPE (J kg^{-1}), DCAPE (J kg^{-1}), V_{pot} (ms^{-1}), and GP for RCE With U_{shear} = 5.0 ms^{-1} and f = 5.0 × 10^{-5} s^{-1}

	SST = 27.5 C	30.0 C	32.5 C
u_m = 1.0 m/s	706	975	1671
	369	424	448
	70.2	74.7	78.0
	9.4	12.3	22.4
3.0	1624	2203	2862
	415	434	492
	51.4	55.5	57.1
	6.7	11.3	12.4
5.0	1558	2195	2832
	453	469	523
	44.2	48.4	49.5
	4.5	7.4	7.9

genesis. With u_m = 5 ms^{-1}, the v_{max} = 10 ms^{-1} vortex fails to develop (not shown).

[17] Table 1 also shows that V_{pot} and GP increase with SST. Thus, we expect that increasing SST shoud be more favorable for TC genesis. Figure 2c shows the control simulation and simulations with SST = 27.5 C and SST = 32.5 C. The results are exactly the opposite of what one would expect. The lower SST case intensifies and reaches hurricane strength sooner, while the SST = 32.5 C case fails to form a TC.

[18] This result depends critically on the presence of mean wind shear. *Nolan et al.* [2007] found that in the absence of shear, increasing the SST of an RCE environment universally increases the rate of development. These simulations indicate the opposite is true in the presence of a modest amount of mean wind shear. The result does *not* seem to be dependent, however, on the structure of the shear profile. This result has been reproduced for other shear profiles with the same U_{shear} value between 850 and 200 hPa: linear shear in log-pressure only between 850 and 200 hPa, and constant wind above and below; a similar profile but with the directions of the shear and the mean surface wind reversed; and a parabolic shear profile. For all of these profiles, TC genesis occurred for v_{max} = 10 ms^{-1} and SST = 30 C, but failed for SST = 32.5 C.

4. Discussion

[19] Simulations with the same initial vortex and the same shear profile, but for TC genesis with SST = 27.5 C, and for TC failure with SST = 32.5 C, looked amazingly similar for the first two days and genesis seemed imminent in both cases. However, for the 32.5 C case, a large convective burst near the vortex center at t = 2.2 days was followed by a large collection of downdrafts and cold outflows that substantially diminished the strength of the low-level vortex. A similar event occurred at t = 3.8 days, after which the mid-level circulation became decoupled from the residual surface vortex. While consistent with the increased downdraft activity, the modest increases in DCAPE (∼20%) for the higher SST environment do not seem indicative of the radically different results from these two simulations.

[20] Recent studies of TC genesis have identified the development of a moist, coherent, mid- level vortex as a necessary step toward TC genesis [*Ritchie and Holland*, 1997; *Bister and Emanuel*, 1997; *Nolan*, 2007]. However, these studies focused on vortices in favorable, low-shear environments. In a sheared environment, the mid-level, mesoscale convective vortex (MCV) is advected downstream from the surface vortex, and the two must remain dynamically coupled for development to proceed. The low-level vortex can slowly intensify by building upwards [e.g., *Davis and Bosart*, 2006; *Tory et al.*, 2007]; or a new low-level vortex can form underneath the mid-level vortex as a

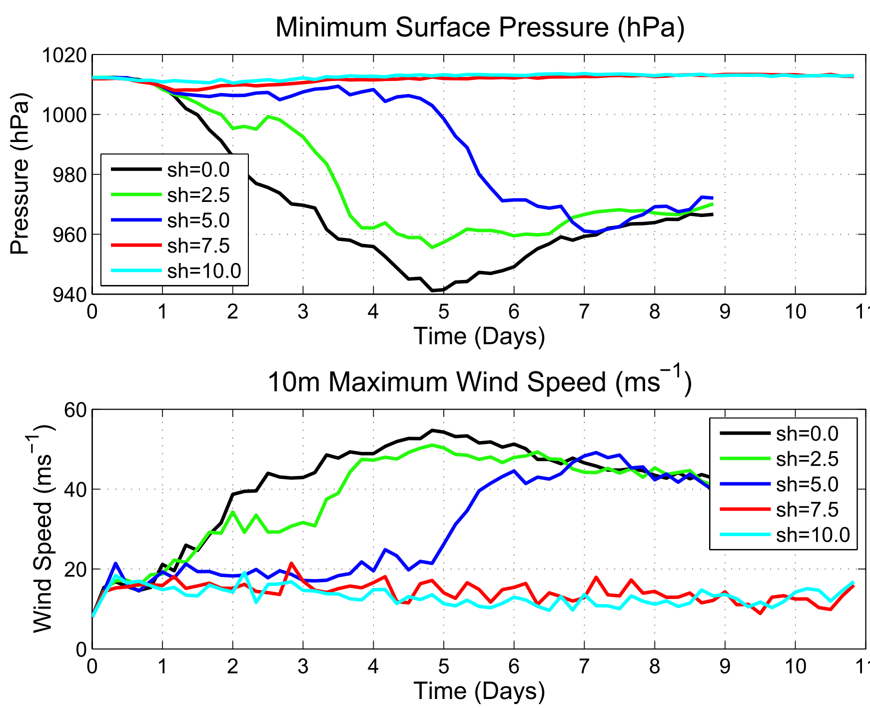

Figure 1. Minimum surface pressure and maximum 10 m wind speed for simulations with SST = 30 C, u_m = 30 ms^{-1}, and various values of 850 to 200 mb wind shear, as labelled, in ms^{-1}.

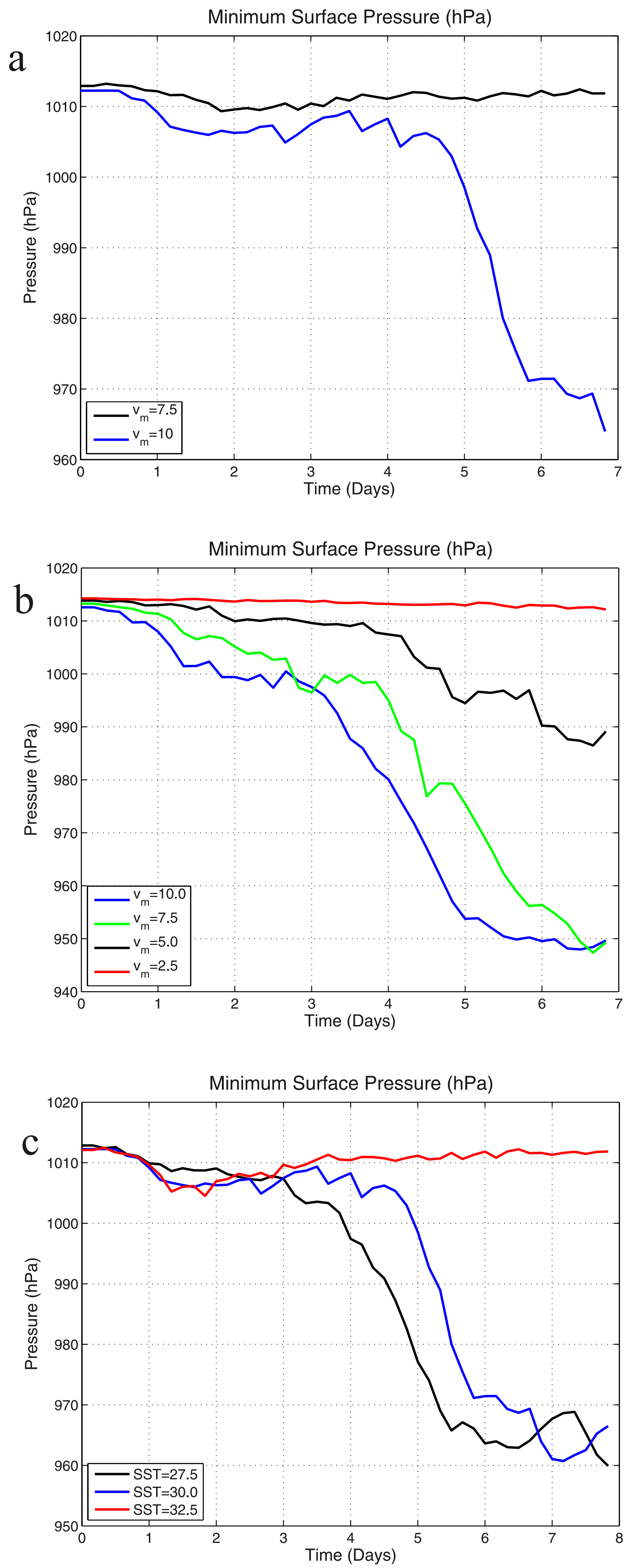

Figure 2. Minimum surface pressure for various sets of genesis simulations for (a) v_{max} = 10 and 7.5 ms^{-1} for the control case environment; (b) v_{max} = 10, 7.5, 5, and 2.5 ms^{-1} with u_m decreased to 2 ms^{-1}; and (c) for the control case with v_{max} = 10 ms^{-1} and for SST = 27.5 and 32.5 C.

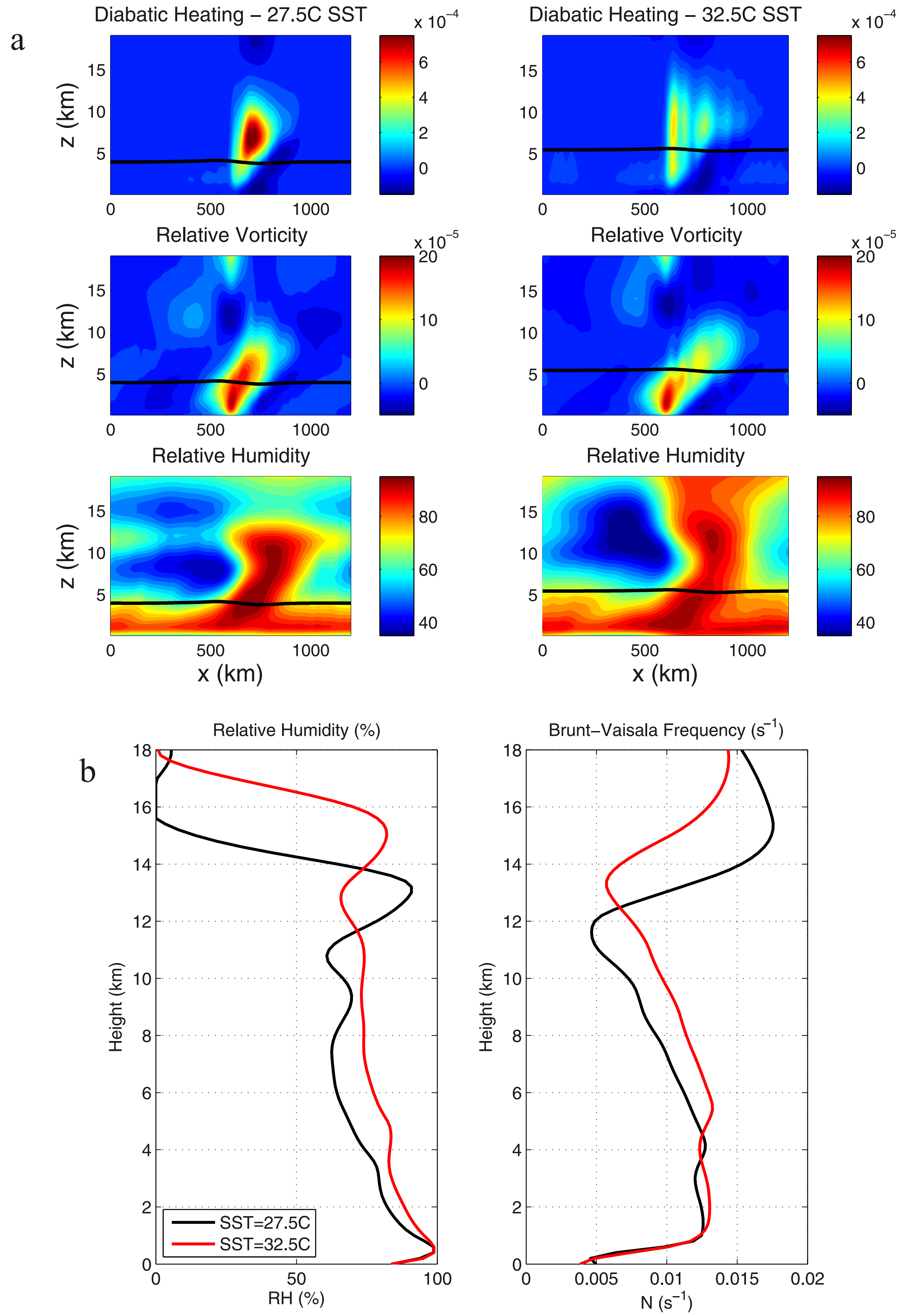

Figure 3. (a) Meridionally and temporally averaged fields of diabatic heating, relative vertical vorticity, and relative humidity in the 24 hours preceding TC genesis (or failure), for two environments: (left) SST = 27.5 C and (right) SST = 32.5 C. (b) Vertical profiles of RH and *N* for the background environments for RCE states for SST = 27.5 and 32.5 C.

result of a large burst of convection, the so called "downshear reformation" [*Molinari et al.*, 2006].

[21] As discussed by *Bister and Emanuel* [1997] and *Nolan* [2007], mid-level vortex intensification is driven in part by stratiform convective processes. These stratiform processes enhance the diabatic heating profile several kilometers above the melting level and diminish the heating around and below the melting level. The resulting mean diabatic heating profile creates a maximum in horizontal convergence between the melting level and the diabatic

heating maximum [*Mapes and Houze*, 1995]. In our RCE environments, the locations of the melting level and the altitude of the heating maximum are strongly controlled by SST and u_m, with both levels rising as each of these parameters are increased. If SST increases and the altitude of the MCV rises, the MCV will experience stronger advection away from the surface vortex since it is higher up into the shear profile. Also, the MCV is further away from the frictionally enhanced surface vortex, leading to less effective coupling between the two vortices.

[22] These effects are illustrated by the panels in Figure 3a. Each plot shows model output averaged in time and space and centered on the moving vortex. The spatial averaging is in the meridional direction from the center of the vortex to 200 km north and the time averaging is over 24 hours. The columns show moist heating, vertical vorticity, and relative humidity for simulations with $U_{\text{shear}} = 5$ ms^{-1}, $u_m = 3$ ms^{-1}, and SST = 27.5 C and SST = 32.5 C, respectively. The black line across each plot shows the melting level. For SST = 27.5 C, moist heating and vertical vorticity form a tilted but coherent vertical column. Dry air impinges on the column from the west, and it can be seen from careful inspection of the figures that there is a mimimum of RH $\sim$ 55% overlying the low-level vortex. For SST = 32.5 C, the melting level is 1.2 km higher. The vertical column of vorticity is more tilted and far less coherent above the melting level. The dry air penetrates further, such that lower values of RH $\sim$ 45% are over the low-level vortex.

[23] The vertical pentration depth associated with the potential vorticity of a finite-amplitude vortex can be written $H \sim [f_{\text{loc}}(f + \zeta)^{1/2}]L/N$, where ζ is the relative vorticity, $f_{loc} = f + 2V/r$, L is the horizontal length scale of the vortex, and N is the Brunt-Vaisala frequency [*Jones*, 1995]. For TCs developing from a pre-defined vortex, f, f_{loc}, and L are all equal across the simulations, at least before TC intensification begins. A factor which does suggest diminished dynamical coupling between the MCV and the surface vortex is the increased value of N in the warmer RCE environment (b). However, the mean values of N between the surface and the altitude of the MCV are 0.0114 and 0.0120 s^{-1}, respectively, only a 4.5% difference. The significant upward shift by 1–2 km of all the local minima and maxima of N and RH is evident in the warmer SST profiles. This is consistent with the upward shifts of the diabatic heating and vorticity generation as illustrated in Figure 3a.

[24] Figure 3b also shows that mid-level environmental RH is significantly higher for the warmer SST case. The larger DCAPE values in the higher SST environments are not due to increased dryness in the middle levels; rather, it is due to the higher *altitude* of the mid-level RH minimum, such that the cool, descending parcels have a longer distance over which to accelerate. Within the vortex, RH is lower and downdrafts are even stronger, due to the drier air generated by adiabatic descent of parcels on the upshear side of the tilted vortex [*Jones*, 1995]. The increased horizontal displacement of the upper and lower parts of the vortex in the higher SST case further enhances this effect in comparison to the lower SST case.

5. Conclusions

[25] The presence of low to moderate shears does not have large effects on the thermodynamics of an atmosphere that is in radiative-convective equilibrium. However, low to moderate shears do have a large effect on TC development. As the shear increases, the time required for a specified mid-level vortex to transition to a developing TC increases until, for sufficient shear, TC genesis does not occur. The inhibiting effects of shear on TC genesis can be overcome by a stronger initial vortex. A larger size may also favor development in shear, but we have not explored the sensitivity to size in this study.

[26] The new and surprising result of this study is that the effect of shear in suppressing TC genesis actually increases as the SST of the RCE environment is increased. A first analysis indicates the reasons for this increased sensitivity are, in decreasing order of likely importance: 1) The higher altitude of the melting level and the stratiform diabatic heating maximum generates a mid-level vortex that is more strongly advected away from the surface vortex; 2) increased shearing of the deeper vortex also leads to stronger adiabatic descent on the upshear side of the vortex, such that drier air is advected over the surface vortex; and 3) the increased dry static stability (as measured by the Brunt-Vaisala frequency, N) further inhibits vertical coupling between the mid-level and surface vortices.

[27] Do these results explain recent general circulation modeling studies predicting fewer tropical cyclones in a global warming world [e.g., *Bengtsson et al.*, 2007]? While the results here are certainly consistent with those studies, are the underlying mechanisms similar, given that the general circulation models cannot resolve convection and downdrafts? Our ongoing work will attempt to answer these questions.

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