Sensitivity of tropical-cyclone models to the surface drag coefficient

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Motivated by recent developments in tropical-cyclone dynamics, this paper re-examines a basic aspect of tropical-cyclone behaviour, namely, the sensitivity of tropical-cyclone models to the surface drag coefficient. Previous theoretical and numerical studies of the sensitivity in axisymmetric models have found that the intensity decreases markedly with increasing drag coefficient. Here we present a series of three-dimensional convection-permitting numerical experiments in which the intensification rate and intensity of the vortex increase with increasing surface drag coefficient until a certain threshold value is attained and then decrease. In particular, tropical depression-strength vortices intensify to major hurricane intensity for values of C_K/C_D as small as 0.1, significantly smaller than the critical threshold value of about 0.75 for major hurricane development predicted by Emanuel using an axisymmetric balance model. Moreover, when the drag coefficient is set to zero, no system-scale intensification occurs, despite persistent sea-to-air fluxes of moisture that maintain deep convective activity. This result is opposite to that found in a prior axisymmetric study by Craig and Gray.

The findings are interpreted using recent insights obtained on tropical-cyclone intensification, which highlight the intrinsically unbalanced dynamics of the tropical-cyclone boundary layer. The reasons for the differences from earlier axisymmetric studies and some potential ramifications of our findings are discussed.

The relative insensitivity of the intensification rate and intensity found for drag coefficients typical of high wind speeds over the ocean calls into question the need for coupled ocean wave–atmospheric models to accurately forecast tropical-cyclone intensity. Copyright © 2010 Royal Meteorological Society

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1. Introduction

In a seminal paper, Malkus and Riehl (1960) developed a theory for a steady-state hurricane in which the maximum swirling wind varied as the square root of the ratio of surface exchange coefficients, C_T and C_D, for sensible heat and momentum, respectively. Subsequent pioneering work by Ooyama (1969) and Rosenthal (1971) broadly confirmed the behaviour of the maximum tangential wind on the drag coefficient, but in addition found that the pertinent energy exchange at the ocean surface was associated with the flux of water vapour into the boundary layer. This transfer may be characterized by an exchange coefficient C_q.

In a tropical-cyclone vortex, the flux of moisture is typically almost an order of magnitude larger than the flux of sensible heat (Frank, 1977). In a particular experiment by Ooyama in which the drag coefficient was kept constant while C_q was allowed to increase with wind speed, the maximum swirling wind was found to exceed 95 m s^{-1}. In a complementary experiment in which C_D increased with wind speed while holding C_q constant, intensification ceased at a maximum swirling wind speed of 45 m s^{-1}. The results suggested that the maximum winds would generally increase with C_q and decrease with C_D. However, Ooyama cautioned that 'although such divergent behavior of C_D and C_q is unlikely to occur in nature, these experiments suggest that more definitive information on both coefficients is needed for the quantitative simulation of tropical cyclones'.

In the first of a series of highly influential papers, Emanuel (1986, henceforth E86) developed a steady-state axisymmetric hurricane model in which the maximum swirling wind is proportional to the square root of the ratio of the surface exchange coefficients for enthalpy*, C_K, and momentum. Like the Malkus–Riehl and Ooyama models, the maximum winds increase with C_K and decrease with C_D. Subsequent refinements of this model have been carried out to include eye dynamics (Emanuel, 1995), the effects of time dependence leading to a theory for intensification (Emanuel, 1997), and dissipative heating due to shear turbulence in the hurricane boundary layer (Bister and Emanuel, 1998). The E86 model formed the basis for a theory for the maximum tangential wind that a storm can achieve in a given environment, often referred to as the potential intensity (PI). As a result of the dependence of maximum tangential wind speed on the ratio of exchange coefficients, the PI increases as C_K increases and decreases as C_D increases†.

A curious result of the Emanuel (1995) study was that the development of intense hurricanes (category 3 and higher on the Saffir–Simpson scale) was curbed when the ratio C_K/C_D fell below a value of 0.75, thereby implying a threshold value of C_K/C_D below which incipient tropical depressions will not intensify to major storms. In a complementary and significant study using the Rotunno and Emanuel (1987) axisymmetric numerical hurricane model, Craig and Gray (1996) found that the rate of intensification increases with increasing values of the transfer coefficients for heat and moisture. They found further that the intensification rate is relatively insensitive to changes in the drag coefficient and noted that 'frictional convergence is of secondary importance [for intensification] but may represent a sink of energy that decreases the growth rate'. A puzzling result of Craig and Gray's study was the finding that the largest intensification rate was obtained with no surface friction at all (p. 3537 of their paper).

Braun and Tao (2000) found some conflicting results concerning the dependence of hurricane intensity on C_K/C_D in the course of a numerical simulation of the intensification of hurricane *Bob* (1991) using the Pennsylvania State University–National Center for Atmospheric Research fifth-generation Mesoscale Model (MM5; Dudhia, 1993; Grell *et al.*, 1995). The main focus of their study was to quantify the dependence of the predicted intensification on the representation of the boundary layer. Four boundary-layer schemes were examined: the Bulk scheme (Grell *et al.*, 1995), the Blackadar scheme (Blackadar, 1976), the Burk–Thompson scheme (Burk and Thompson, 1982) and the MRF scheme (Hong and Pan, 1996). Because these schemes use different formulations for C_K and C_D, calculations were compared in which a particular scheme used the formulations for C_K and C_D from the other schemes. It was found, *inter alia*, that, although the value of C_K/C_D increases from 0.7 to 1.0 to 1.3 between the Blackadar, Burk–Thompson, and standard bulk-aerodynamic schemes, the simulated minimum pressure did not decrease (nearly) linearly with C_K/C_D as predicted by theory (Emanuel, 1995b; his Figure 3, solid curve). They suggested that the reason for the discrepancy from theory 'is presumably because of the complex interactions between planetary boundary-layer processes, cloud microphysical processes, and storm dynamics' and noted that a similar argument can be made for the maximum wind speed. In particular, they noted that the standard bulk aerodynamic surface flux scheme without the wind speed dependence of surface roughness length, z_0, is characterized by the largest value of C_K/C_D, but it produced an intensity only slightly larger than that in the Blackadar case, in apparent contradiction to Emanuel's theory (Emanuel, 1995b).

When the wind speed dependence of z_0 was included in the bulk aerodynamic surface flux scheme, the bulk scheme did result in the strongest storm. However, other combinations of boundary-layer schemes and surface exchange coefficients with identical values of C_K/C_D resulted in very different intensities, leading Braun and Tao to conclude that the intensity is related not only to the magnitude of C_K/C_D, but depends significantly also on 'the wind speed dependence of the surface roughness parameter z_0'. These results alone call for a deeper understanding of the dependence of hurricane intensification on the surface exchange coefficients. It may be worth pointing out that Braun and Tao's calculations are based on a solution of the full vector momentum equation, unlike Emanuel's theory, a difference that we believe is important, as discussed below.

The perceived importance of the dependence of hurricane intensification on the ratio C_K/C_D emerging from Emanuel's axisymmetric theory, in particular, has led to a concerted effort to acquire measurements of the exchange coefficients at hurricane-strength wind speeds (Black *et al.*, 2007; Drennan *et al.*, 2007; French *et al.*, 2007; Zhang *et al.*, 2008). The new field observations suggest a value of C_K/C_D between 0.6 and 0.7 (Black *et al.*, 2007, their Figure 7; Zhang *et al.*, 2008, their Figure 4), but the data exhibit considerable scatter and are limited to wind speeds of marginal hurricane strength. Recent laboratory work by Haus *et al.*(2010) suggests a value of C_K/C_D more like 0.5 for winds greater than 30 m s^{-1} up to a maximum wind of nearly 40 m s^{-1}. There are differing views on how C_K/C_D might vary beyond 40 m s^{-1}, but the quantitative determination of this ratio at extreme wind speeds remains enigmatic (e.g. Andreas and

*If the transfer coefficients for sensible and latent heat are the same, then the total heat transfer is equivalent to the transfer of moist enthalpy (Emanuel, 1995). This equality will be assumed herein.

†A review of PI theory and its relation to the Malkus and Riehl theory is provided by Emanuel (2004).

Emanuel, 2001; Emanuel, 2003)[‡]. The most appropriate value at high wind speed is certainly considered important for accurate hurricane intensity prediction, but we believe it is equally important to determine and understand the sensitivity of hurricane behaviour to changes in C_K/C_D and furthermore to the individual exchange coefficients that comprise it.

Recent insights into the physics of tropical-cyclone intensification have drawn attention to the dynamical role of the boundary layer in the spin-up process (Smith *et al.*, 2009; henceforth M3). Here we define the boundary layer as the surface-based layer of air in which frictional stresses are important. The layer has a depth typically between 500 m and 1 km and is characterized by *strong* inflow with the radial wind component reaching 10–20 m s^{-1}[§]. In M3 we highlighted the fact that the maximum azimuthally-averaged tangential wind speed occurs at low levels near the top of the frictionally-induced inflow layer. This maximum is associated with radial convergence of absolute angular momentum *within the boundary layer*. Although absolute angular momentum is not materially conserved in the boundary layer, high tangential wind speeds can be achieved if the radial inflow is large enough to bring the air parcels to small radii with minimal loss of angular momentum. This spin-up mechanism is tied fundamentally to the dynamics of the boundary layer, where the flow is not in gradient-wind balance. Indeed, the strong frictionally-induced inflow near the surface leading to the spin-up is not captured by a balanced model (Bui *et al.*, 2009, their Figures 5 and 6).

It should be recalled that both the Ooyama (1969) and E86 models assume gradient wind balance everywhere, including the boundary layer (Smith *et al.*, 2008). However, this 'balanced boundary layer' assumption cannot be rigorously supported by a scale analysis of the boundary-layer equations (Smith and Montgomery, 2008; Vogl and Smith, 2009). For this reason, and because of the tendency for the boundary-layer inflow jet to produce supergradient winds for air parcels spiralling into the vortex core, it is possible that the dependence of the maximum azimuthally-averaged tangential wind on C_D may be different from that in the axisymmetric balance models discussed above. Since the frictionally-induced inflow arises from the agradient force[¶], its strength and that of the agradient force will increase with the drag coefficient. As shown in Smith *et al.* (2008), M3 and Bryan and Rotunno (2009b), the unbalanced dynamics in the inner-core region are generally important for determining the maximum tangential wind that can be attained and, indeed, the PI[‖]. In light of these findings, the foregoing results of Craig and Gray (1996) appear to be in conflict with those of M3. We believe this discrepancy may be explained largely on account of the coarse resolution of the boundary layer (1.25 km vertical grid spacing) and the overdiffusive nature of the Craig and Gray (1996) configuration of the Rotunno and Emanuel (1987) hurricane model (Bryan and Rotunno, 2009b, p. 3055). In such a coarse resolution and diffusive regime, the boundary-layer dynamics are less important in the spin-up process, a result that is brought out by an explicit comparison of balanced and unbalanced contributions to the secondary circulation of an idealized tropical-cyclone vortex undergoing rapid intensification (Bui *et al.*, 2009).

Current observational analyses of intense hurricanes (Kepert, 2006a,b; Bell and Montgomery, 2008; N. Sanger, personal communication; M. Bell, personal communication) as well as recent theoretical considerations (M3; Bryan and Rotunno, 2009a) suggest that the inertial (and hence unbalanced) effects are important in the inner-core region. A consequence of the results presented in M3 is that, in an inertial regime, where the inward advection of absolute angular momentum outweighs the frictional loss, an increase in drag could lead to an increase in intensity rather than a decrease because of an increase in the frictionally-induced inflow.

Given the goal to improve hurricane intensity prediction, it is important to have a basic understanding of the physical processes that govern intensity and their dependence (or lack thereof) on the exchange coefficients at high wind speeds. From the foregoing discussion concerning the potential importance of unbalanced effects associated with the frictional disruption of gradient wind balance in the boundary layer, we focus our attention here on examining the dependence of hurricane intensification on the surface drag coefficient in the context of idealized, three-dimensional, numerical model calculations.

The outline of this paper is as follows. In section 2 we describe briefly the numerical model used in this study. The calculations and their interpretation are described in section 3 and the conclusions are given in section 4.

2. The numerical model

The numerical model employed here is a modified version of the MM5 model, the same as used by Nguyen *et al.* (2008, henceforth M1) to study the tropical-cyclone intensification process in three dimensions. The model is configured with three domains: a coarse mesh of 45 km horizontal grid spacing and two, two-way nested domains of 15 and 5 km grid spacing, respectively. The domains are square and are 5400 km, 1800 km, and 600 km on each side. There are 24 σ-levels in the vertical, seven of which are below 850 mb and therefore adequate for representing the principal features of the boundary-layer flow as discussed above[**]. The calculations are performed on an f-plane centred at 20°N and neglect dissipative heating.

The numerical experiments presented are an extension of Experiment 12 discussed in the Appendix of Montgomery *et al.* (2009; henceforth M2). The specifics are as follows.

[‡]Some significant strides towards determining these coefficients at major hurricane wind conditions have been made recently by M. Bell (2010) as part of his Ph.D. dissertation.

[§]While there is inflow throughout the lower troposphere in the calculations presented in M3, by far the largest radial wind speeds are confined to the boundary layer. The lower-tropospheric inflow above the boundary layer results from a balanced response of the vortex to the negative radial gradient of the azimuthally-averaged diabatic heating rate in the eyewall clouds (Willoughby, 1979).

[¶]The agradient force is the sum of the radial pressure gradient force and the Coriolis and centrifugal forces, usually expressed per unit mass of air.

[‖]In theoretical studies, the maximum tangential wind is usually the preferred intensity metric. In operational communities, however, intensity is defined as a sustained horizontal wind speed at the surface or anemometer level (e.g. 10 m) over some time interval, either a 1 min or 10 min average. To keep with the basic research theme of this paper, we will adhere to the former definition unless noted otherwise.

[**]The main conclusions and interpretations presented below have been verified by repeating two of the main experiments (Experiments 2 and 4) using a higher vertical resolution of 36 σ-levels in the vertical.

Table I. List of first group of experiments, varying C_D from moderate to small values.

Experiment	Description
1	Model configuration as Experiment 12 from M2 (see text for details) in which C_K and C_D are defined through the default bulk-aerodynamic scheme in MM5
2	As Experiment 1, but surface exchange coefficients are set to a constant both in space and time with $C_K = 1.3 \times 10^{-3}$ and $C_D = 1.86 \times 10^{-3}$ ($C_K/C_D = 0.7$)
3	As Experiment 2, but with $C_D = 1.3 \times 10^{-3}$ ($C_K/C_D = 1.0$)
4	As Experiment 2, but with $C_D = 1.0 \times 10^{-3}$ ($C_K/C_D = 1.3$)
5	As Experiment 2, but with $C_D = 0.0$

To keep the experiments relatively simple, the main physics options chosen are the bulk-aerodynamic boundary-layer scheme and a warm-rain scheme. These schemes are applied in all domains. No cumulus parametrization is used in any domain[††]. The sea-surface temperature is set to a constant 27 °C. A simple radiative-cooling scheme is employed that relaxes the temperature at every grid point towards the initial sounding with a cooling time-scale of one day. The MM5 model is modified further to have periodic boundary conditions on the coarsest domain in both north–south and east–west directions as a simple expedient for ensuring global conservation of predicted fields in the absence of sinks and sources. In the diagnostic analyses presented, the vortex centre is defined as the centroid of relative vorticity at 1 km height over a circular region of 200 km radius from a 'first-guess' centre, which is determined by the minimum of the total wind speed at 1 km height. The maximum azimuthally-averaged tangential wind speed, V_{max}, and the maximum total wind speed, VT_{max}, are defined as global maxima in the innermost domain. These maxima occur mostly below an altitude of 1 km.

The initial vortex is axisymmetric with a maximum tangential wind speed of 15 m s^{-1} at the surface at a radius of 135 km. The strength of the tangential wind decreases sinusoidally with height, vanishing at the top model level (50 mb). The temperature field is initialized to be in gradient-wind balance with the wind field using the method described by Smith (2006). The far-field temperature and humidity are initialized with the so-called 'neutral' sounding from the Rotunno and Emanuel (1987) model (Figure 15 of M2). This sounding was generated by running the Rotunno and Emanuel model with an initially quiescent Jordan (1958) sounding environment to a state of near convective equilibrium[‡‡].

In this study, the surface exchange coefficients of heat and momentum are defined either through the default bulk-aerodynamic scheme in MM5 (Grell *et al.*, 1995, give details) or taken to be a constant in both space and time, whose quantitative value is comparable to known values. In the latter case, the moist enthalpy flux coefficient C_K* is set equal to 1.3×10^{-3}. This value is close to the mean value found using the default scheme in MM5 (M2 gives details) and is close to the mean value ($\approx 1.2 \times 10^{-3}$) derived from the Coupled Boundary Layers/Air–Sea Transfer (CBLAST) experiment (Figure 6 of Black *et al.*, 2007; Figure 4 of Zhang *et al.*, 2008) and a recent laboratory study (Figure 1 of Haus *et al.*, 2010) near and slightly above marginal hurricane wind speeds.

Unlike the Braun and Tao (2000) study in which the surface roughness was a variable that influenced the exchange coefficients, in these experiments surface roughness effects are circumvented by specification of the drag and enthalpy coefficients.

In the first group of experiments the drag coefficient is decreased from 1.86 to 1.3, 1.0, and 0.0×10^{-3}. These experiments will be referred to henceforth as Experiments 2, 3, 4 and 5, respectively (the experiments are listed in Table I). The results are presented in sections 3.1 and 3.2. In the second group of experiments (Experiments 6–10 in Table II), we increase C_D progressively until it reaches a value of 13.0×10^{-3}; the results are presented in section 3.3.

Some support for holding C_K constant while allowing C_D to vary is provided in the recent observations of Black *et al.* (2007, their Figures 5 and 6), Zhang *et al.* (2008, their Figure 4) and Haus *et al.* (2010, their Figure 1). These works suggest that the drag coefficient exhibits a steady increase to a wind speed of approximately 25 m s^{-1} with correspondingly little systematic increase in the moisture exchange coefficient above 15 m s^{-1}. We seek maximum simplicity by setting the heat and momentum exchange coefficients equal to a constant. This simplification eliminates
(1) the spatial variations in these quantities that would be expected to be significant in the vicinity of convective towers in the early stage of the intensification process, which locally enhance the surface wind speed (M1), and
(2) the more significant variations in these quantities in and near the eyewall at later times. This non-axisymmetric property is inherent in the calculations presented in Braun and Tao (2000) and all of the calculations presented in M2 (except for Experiment 11 wherein only the enthalpy coefficient was set to a constant).

The vertical and horizontal diffusion of heat and momentum that is included to parametrize subgrid-scale processes is determined from the model's default settings. For simplicity, all numerical integrations are terminated

[††]Strictly speaking, the MM5 model employs a dry convective adjustment scheme that contributes to the vertical mixing of potential temperature (see Grell *et al.*, 1995, for details). Whereas the focus of the current study is on the dependence of the simulated intensification rate and intensity on the surface drag coefficient and not vertical mixing processes *per se*, we have repeated two of the main experiments (Experiments 2 and 4) using 24 σ-levels in the vertical with the dry-adjustment scheme turned off. The general conclusions are unchanged despite the intrinsic stochastic element of the intensification process associated with the deep, rotating convective structures that drive the spin-up process.

[‡‡]We are of course aware of the fact that a neutral sounding in one model is not necessarily neutral in another, given the dependence of convective instability in a discrete model to model physics, to resolution, and even to dimensionality. For a hypothetical air parcel lifted from the lowest model level, the CAPE is approximately 100 J kg^{-1}. This value is considerably smaller than that of approximately 900 J kg^{-1} for the approximation to the Jordan sounding used in Experiments 1–7 of M2. The very small CAPE is thought to be sufficient to balance internal dissipation within the model's clouds and for this reason the sounding is considered 'neutral'.

*As noted in the Introduction, we assume $C_T = C_E$ and hence the moisture transfer coefficient is equivalent to the enthalpy coefficient, C_K.

Table II. List of second group of experiments. All are as Experiment 2, but C_D is varied to large values.

Experiment	C_D	C_K/C_D
6	2.6×10^{-3}	0.5
7	3.9×10^{-3}	0.33
8	5.2×10^{-3}	0.25
9	6.5×10^{-3}	0.2
10	13.0×10^{-3}	0.1

at 4 d, since longer time experiments may admit new evolutionary pathways for the hurricane vortex (such as secondary eyewall formation and subsequent evolution) that lie outside the scope of the present paper (e.g. Terwey and Montgomery, 2008, and references therein).

Experiments 2–10 provide a direct test of the hypothesized effect of intensification and mature intensity on the drag coefficient as articulated in the Introduction. Of course, the latest findings on the dependence of the drag coefficient with wind speed (Black *et al.*, 2007) could be incorporated to further test our findings. However, in view of the uncertainties in the standard parametrizations employed for the planetary boundary layer, the subgrid-scale parametrization, and the cloud microphysical processes, a systematic survey of all these issues lies outside the scope of the current study.

3. The calculations

Figure 1 shows a time series of the maximum azimuthally-averaged tangential velocity for the first five numerical experiments (the 'control' Experiment 12 from M2, hereafter Experiment 1, and the new Experiments 2–5 as described above). To provide a context for the Experiments 1–5, the principal characteristics of the control experiment (curve 1) are reviewed first.

3.1. *Small to moderate surface drag*

The intensification process is described in detail in M1 and M2. Briefly, the evolution begins with a gestation period, during which the vortex slowly decays because of surface friction, but moistens in the boundary layer because of evaporation from the underlying sea surface. This period lasts approximately 9 h, during which time the minimum surface pressure rises slightly from its initial value of 1004 mb to 1008 mb (not shown), but the maximum tangential wind speed decreases only by a small amount (by less than 0.5 m s^{-1}). The imposition of surface friction from the initial instant leads to an azimuthal-mean inflow in the boundary layer and a mean outflow above it, the outflow accounting for the decrease in tangential wind speed through the conservation of absolute angular momentum. The inflow is moist, and as it rises out of the boundary layer and cools, condensation progressively occurs in some grid columns interior to the corresponding radius of maximum tangential wind speed.

Existing relative vorticity is stretched and amplified in these columns, leading to the formation of localized rotating updraughts. Hendricks *et al.* (2004) and Montgomery *et al.* (2006a) coined the term 'vortical hot towers' (VHTs)

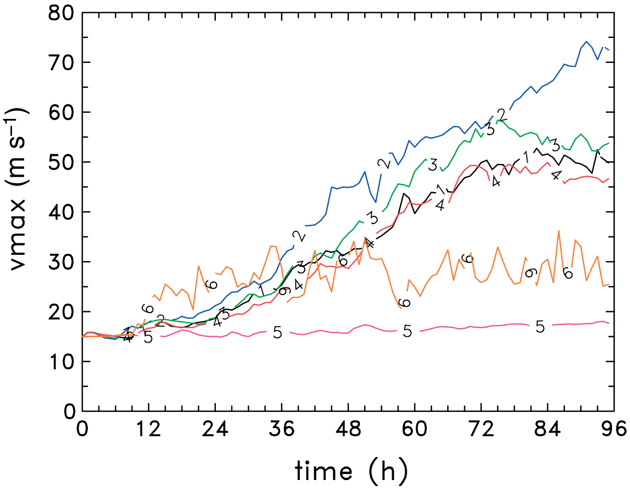

Figure 1. Time series of maximum azimuthally-averaged tangential wind speed in five experiments: curve 1, Experiment 12 of M2 (Experiment 1 here); curve 2, $C_K/C_D = 0.7$ (Experiment 2); curve 3, $C_K/C_D = 1.0$ (Experiment 3); curve 4, $C_K/C_D = 1.3$ (Experiment 4); curve 5, $C_D = 0$ (Experiment 5). Curve 6 shows the maximum total wind speed in Experiment 5.

to describe these rotating and locally warm updraughts. As the updraughts develop, there ensues a period lasting about 60 h during which the vortex rapidly intensifies and V_{max} increases from approximately 14.5 m s^{-1} to approximately 45 m s^{-1}. It was shown in M2 that, from a three-dimensional fluid dynamics perspective, the intensification mechanism is via the generation of locally buoyant VHTs and the near-surface convergence that the VHTs induce within the system-scale boundary layer. During the amplification process, it was found that the system-scale density temperature generally lagged behind the local density temperature within the VHT cores. It is these cores that drive the intensification process until the system-scale temperature along absolute angular momentum surfaces nearly coincides with the local temperature of the VHT cores. The azimuthally-averaged aspects of this spin-up process have been presented in M3.

Intensification in Experiments 2–4 is found to proceed by the same VHT pathway as described above for Experiment 1. Figure 1 summarizes the simulated intensification for each experiment by way of a time series for the maximum azimuthally-averaged tangential velocity. It shows also the maximum total wind speed in Experiment 5. While Experiments 2–4 intensify in a manner that is broadly similar to Experiment 1, we see that when the ratio C_K/C_D is reduced from 1.3 (Experiment 4) to 0.7 (Experiment 2), the intensification rate and maximum mean tangential wind speed increase noticeably (with the difference exceeding 15 m s^{-1} at mature intensity)†. To interpret this seemingly counter-intuitive result, it proves useful to adopt an axisymmetric flow perspective of the spin-up process as discussed in the Introduction and M3. In doing this, it should be borne in mind that the strong, unbalanced, frictional inflow that the VHTs generate within their own

†If a similar time series is constructed for the maximum horizontal wind speed at the surface ($z = 0$) and the global maximum horizontal wind speed (that occurs typically below 1 km altitude), we find a similar conclusion after 84 h. At earlier times, the difference in maximum surface wind speed between the small and large drag experiments is reduced, but the large drag experiment still yields maximum surface winds that are generally greater than those of the small drag experiment.

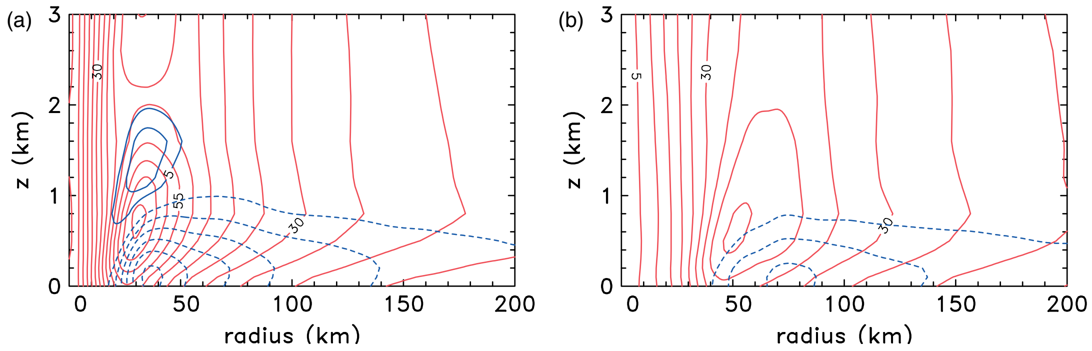

Figure 2. Radius–height cross-sections of azimuthally-averaged radial (dashed) and tangential (solid) wind speed components in the lowest 3 km at 96 h for (a) Experiment 2, $C_K/C_D = 0.7$, and (b) Experiment 4, $C_K/C_D = 1.3$. The contour interval for both components is 5 m s^{-1}, and the zero contour is not plotted.

boundary layers contributes to the unbalanced, azimuthally-averaged boundary-layer flow of the system-scale vortex. The contribution of the unbalanced dynamics to the azimuthally-averaged tendency of the tangential wind was quantified in Bui *et al.* (2009; their Figure 9).

Figure 2 shows radius–height cross-sections of the fields of azimuthally-averaged radial velocity (u) and tangential velocity (v) components in the lowest 3 km at 96 h for the two limiting values of C_K/C_D, namely, $C_K/C_D = 0.7$ (Experiment 2, Figure 2(a)) and $C_K/C_D = 1.3$ (Experiment 4, Figure 2(b)). As found in M3, the maximum azimuthally-averaged tangential wind speed occurs at low levels near the top of the frictionally-induced inflow layer for both cases. The radius of maximum tangential wind resides near a radius of 30 km for the stronger surface drag (Figure 2(a)), compared with approximately 50 km for the reduced surface drag (Figure 2(b)). A physical explanation for this behaviour follows directly from the ideas discussed in the Introduction. The increased friction leads to a larger gradient wind imbalance in the boundary layer. With the higher C_D value, inflowing rings of boundary-layer air are subjected to a greater agradient force and hence are converged farther inwards before rising out of the boundary layer and ascending into the eyewall updraught. The end result is an enhanced maximum tangential wind despite the loss of absolute angular momentum en route (M3).

One of the principal findings of M1 was the demonstration that the intensification process and maximum mean tangential wind possess an intrinsic stochastic element associated with the VHTs. In view of this result, each time series in Figure 1 should be regarded as a single member of an ensemble generated by, say, small moisture variations in the boundary layer. It is useful to recall the findings in M1 that show a spread of approximately 10 m s^{-1} in the maximum azimuthally-averaged tangential velocity amongst ensemble members during the intensification phase. On the basis of these results, the difference between Experiments 2 and 4 should persist clearly in an ensemble mean, though the difference between Experiments 1, 4 and 3 would likely be marginal for most of the evolution shown.

3.2. *The zero-drag limit*

For the limiting case of zero surface drag, $C_D = 0$, we find a stark contrast between previously published work using the Rotunno and Emanuel (1987) axisymmetric hurricane model (Craig and Gray, 1996) and the three-dimensional MM5 model. In the three-dimensional case with $C_D = 0$ (Experiment 5), we find no system-scale convergence produced by the boundary layer near or within the initial vortex's radius of maximum tangential wind ($\approx$ 135 km) and find only a few metres per second spin-up of the maximum (azimuthally-averaged) tangential wind field (Figure 1) in the outer circulation near 200 km radius after 4 days. The absence of any inner-core spin-up is unlike all other experiments conducted herein. Instead, what is observed is a temporally persistent, but spatially chaotic initiation of deep convective towers beyond the initial radius of maximum tangential wind ($\geq$ 200 km). These towers produce locally cyclonic–anticyclonic vorticity anomalies (i.e. vorticity dipoles) that originate from the tilting of the horizontal vortex lines of the mean swirling flow by the convective updraughts. The convectively-generated vorticity dipoles ultimately become sheared by the differential shear of the primary circulation and do not contribute much of a net increase in the system-scale circulation within a fixed closed loop surrounding the high wind region of the vortex.

Concurrent with the convective updraught/ downdraught cycle is the generation of equivalent potential temperature (θ_e) deficits in the vicinity of the convective cores near the surface ($z \approx$40 m). As discussed in M2 (and references cited therein), the surface θ_e deficits are created by precipitation-driven downdraughts that transport relatively low-θ_e air into the boundary layer, but generally diminish with time due to the persistent fluxes of water vapour from the underlying sea. Therefore, the deep convective activity can be supported without surface friction.

The picture that emerges is one in which the local convective towers produce regions of enhanced wind speed locally (curve 6 in Figure 1), but are unable to focus convection around the circulation centre and unable to foster any significant increase in the system-scale tangential wind field during the 4 d integration. This behaviour is strikingly different from that reported by Craig and Gray (1996). We have verified the basic result of Craig and Gray with our version of the Rotunno and Emanuel (1987) model using a configuration comparable to theirs. In this case we observe a gestation time of approximately 2 d (cf. M1) after which the system-scale vortex rapidly intensifies, reaching an intensity of approximately 80 m s^{-1} roughly 4 d later at 150 h (not shown). During the time interval in which the system-scale winds undergo intensification, the radius of maximum tangential wind decreases from 100 km to 80 km. At later times, the vortex continues to intensify and the radius of maximum tangential wind increases with time (M. Bell, personal communication). This experiment eliminates

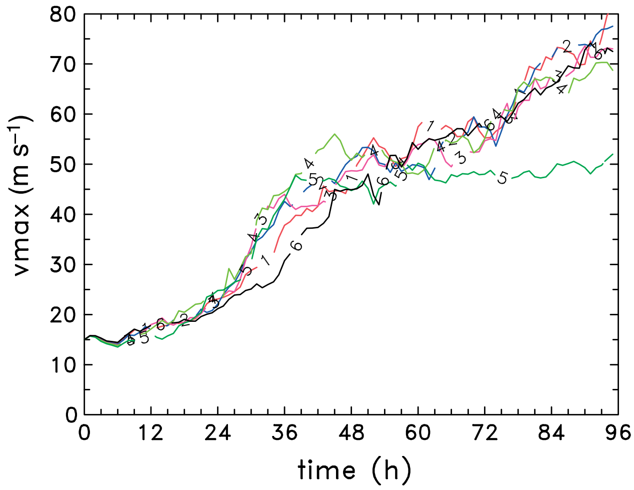

Figure 3. Time series of maximum azimuthally-averaged tangential wind speed in the set of Experiments 6–10 in Table II: curve 1, $C_K/C_D = 0.5$ (Experiment 6); curve 2, $C_K/C_D = 0.33$ (Experiment 7); curve 3, $C_K/C_D = 0.25$ (Experiment 8); curve 4, $C_K/C_D = 0.2$ (Experiment 9); curve 5, $C_K/C_D = 0.1$ (Experiment 10). Curve 6 is for Experiment 2 ($C_K/C_D = 0.7$), for reference.

the possibility of a coding error in the Craig and Gray (1996) study on this matter and points to a noteworthy difference in the intensification process in axisymmetry and three-dimensions for the case of zero surface drag, at least for the Rotunno and Emanuel hurricane model that has been used widely for idealized studies. While a zero drag is unrealistic, the ramifications are nonetheless interesting from a scientific standpoint.

3.3. *Large drag*

Having considered the zero-drag limit, we consider now the large-drag limit. To enter this regime, we increase the drag coefficient C_D incrementally above the values employed in Experiment 2 until reaching a maximum value of 13.0×10^{-3}. Figure 3 shows the results of this series of numerical experiments (Experiments 6–10 in Table II). When the drag coefficient is increased modestly above that of Experiment 2 (e.g. $C_D = 2.6 \times 10^{-3}$ and $C_D = 3.9 \times 10^{-3}$, equivalent to $C_K/C_D = 0.5$ or 0.33), the intensification rate and intensity at 4 d are essentially unaltered from similar characteristics observed in Experiment 2. At 96 h integration time, the intensity for these experiments is in the vicinity of 75 m s^{-1}. It is only when the drag coefficient is increased to 5.2 $\times 10^{-3}$ or beyond ($C_K/C_D = 0.25$ or less) do we see a slight increase in the intensification rate at early times. At yet higher values of surface drag ($C_D = 13.0 \times 10^{-3}$, or $C_K/C_D = 0.1$), the maximum intensity at 96 h reverts back to 50 m s^{-1}, an intensity comparable to that found in Experiment 1. The only curve in Figure 3 to differ significantly from the others is curve 5 (after 60 h).

In the light of the results and interpretation presented in section 3.1, it is not surprising to find an increase in the intensification rate with the drag coefficient. Nevertheless, it is surprising how resilient the intensification process is for surface drag coefficients larger than anything known over the open oceans! The largest value of C_D considered here is perhaps more appropriate for a land mass and may give a hint as to why storms can intensify as they make landfall.

The fact that a hurricane of major intensity can develop in these idealized experiments when C_K/C_D is as small as 0.1 is exactly opposite to the predictions of Emanuel (1995) who suggested a critical value of about three fourths (0.75) for major hurricane development and a 'strong dependence of hurricane structure and intensity on C_K/C_D'. Our findings are supported by the modelling results cited by Wang and Wu (2004, p. 269), Cram *et al.* (2007), and Hill and Lackman (2009, p. 762), all of whom neglected dissipative heating.

Within the context of the foregoing discussion and supporting results, it is pertinent to recall that the critical value of about 0.75 was obtained using the Emanuel (1995) model that neglects dissipative heating. When dissipative heating is represented following Bister and Emanuel (1998), the predicted dependence of the PI on the ratio C_K/C_D is unchanged. However, the threshold value of this ratio for major hurricane development *in this model* may be reduced by 50% since the thermodynamic efficiency parameter is increased by 50% via the replacement of the sea-surface temperature with the outflow temperature in the denominator. These modifications notwithstanding, the results and interpretations presented herein are significant because they show that the lack of a threshold in C_K/C_D is due to the unbalanced boundary-layer dynamics and not to the inclusion of dissipative heating, which serves to boost the simulated intensity by only a modest amount (e.g. Montgomery *et al.*, 2006b).

3.4. *Discussion*

On the basis of the above findings, in conjunction with recent insights in the dynamics of the hurricane spin-up process (M3), we offer two reasons why the discovery of the dependence of the maximum intensity on C_D for small to moderate values of C_D has gone unnoticed. The first is the reliance on models that adopt a *balanced boundary layer*, which for reasons articulated in the Introduction give the opposite dependence of intensity on drag coefficient. The second originates in the choice of the boundary-layer height or its equivalent numerical representation in a grid point (or spectral) model. Although there are differing definitions of the hurricane boundary layer, all of the definitions and observations suggest that the layer is confined to a shallow layer that is approximately 500 m to 1 km deep (Moss and Rosenthal, 1975; Montgomery *et al.*, 2006b; Bell and Montgomery, 2008; Marks *et al.*, 2008; Zhang *et al.*, 2009)[‡]. If the slab boundary layer is too deep, the vertical grid spacing is too large, or the vertical eddy diffusivity is too large, *the agradient force in the boundary layer will be suitably small* (Smith *et al.*, 2008; Smith and Vogl, 2008) and the dynamics will be slaved to behave like a *balanced boundary layer* (e.g. Bryan and Rotunno, 2009a, their Figure 6). To the best of our knowledge, many of the works cited herein examining the sensitivity of tropical-cyclone intensification and/or maximum intensity to surface exchange coefficients have fallen into one or both of these categories, presumably because of the limited computational resources available at the time of these studies and/or a lack of an appreciation of the important role of the unbalanced boundary-layer dynamics in the tropical-cyclone spin-up process.

[‡]The different definitions are discussed in Smith and Montgomery (2010).

4. Conclusions

Motivated by recent developments in tropical-cyclone dynamics, we have re-examined the sensitivity of tropical-cyclone models to a constant surface drag coefficient while keeping the coefficient of enthalpy transfer between sea and air unchanged and near observed values from the recent CBLAST experiment. Unlike the predictions of previous studies using axisymmetric models, we find that the intensification rate of the vortex and the intensity (up to 4 d) increase with increasing C_D up to approximately 2×10^{-3}. When C_D is increased further, no significant difference in the intensification rate or intensity is found until a threshold of approximately 1.3×10^{-2}, beyond which the intensity decreases. Although the latter drag coefficient is certainly not realistic over the open ocean, these findings nonetheless illustrate the relative insensitivity to the intensification rate and intensity for drag coefficients typical of high wind speeds over the ocean (Powell *et al.*, 2003; Donelan *et al.*, 2004; Black *et al.*, 2007). By relative insensitivity we mean variations that lie within the predictability envelope for intensity associated with the convective structures that operate in and around the eyewall region of the storm (M1).

For the regime of small to moderate C_D ($\lesssim 2 \times 10^{-3}$), the increase with C_D in the intensification rate and intensity up to 4 d is associated with the increase of frictional inflow on account of the increased agradient force in the boundary layer. The generalized Coriolis force associated with this inflow acts to accelerate the tangential flow also, leading to a more intense vortex. At this time, we are unable to offer an explanation for the relative insensitivity of the intensification rate and intensity for values of C_D greater than approximately 2×10^{-3}.

In the limiting case when the drag coefficient is set to zero, no system-scale intensification occurs, despite persistent sea-to-air fluxes of moisture that maintain deep convective activity. These results are in contrast to those of a significant study by Craig and Gray (1996), where the intensity decreased somewhat with increasing drag coefficient and where the intensification rate was a maximum in the presence of zero surface drag. Although Craig and Gray did not consider the zero-drag limit to be realistic, the latter result serves to highlight an interesting difference between this axisymmetric study and the current three-dimensional study for the hurricane intensification problem, at least in the limit of small surface drag.

We find also that the simulated intensity (at 4 d) does not depend markedly on the value of the ratio C_K/C_D below a value of approximately 0.7, as has been suggested by Emanuel (1995) using an axisymmetric hurricane model without dissipative heating. To wit: tropical depression strength vortices intensify to major hurricane intensity for C_K/C_D values as small as 0.1, appreciably smaller than the critical threshold value of about 0.75 suggested previously using the axisymmetric model of Emanuel (1995) employing a *balanced boundary layer* (cf. Smith and Montgomery, 2008). Since the current study neglected the dissipative heating effect, the results and interpretations herein demonstrate that the lack of a threshold in C_K/C_D for major hurricane development is due fundamentally to the unbalanced boundary-layer dynamics.

Although there are differing views on how C_K/C_D might vary beyond $40\,\mathrm{m\,s^{-1}}$, and while the quantitative determination of C_K/C_D at extreme wind speeds remains enigmatic, these findings suggest that the precise behaviour of C_K/C_D with surface wind speed may not be as critical as orginally perceived. For all intents and purposes, an approximate determination of C_K and C_D with suitable error bars may be sufficient without the need for complicated surface wave parametrizations in coupled air–sea modelling efforts. Accordingly, we question the viewpoint expressed in Hill and Lackman (2009, p. 763) that the wave-coupling component in coupled ocean wave–atmospheric models is necessary to accurately forecast tropical-cyclone intensity.

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References

Andreas EL, Emanuel KA. 2001. Effects of sea spray on tropical cyclone intensity. *J. Atmos. Sci.* **58**: 3741–3751.

Bell MM. 2010. Air–sea enthalpy and momentum exchange at major hurricane wind speeds. Ph. D. Dissertation, Naval Postgraduate School, Monterey, California, USA.

Bell MM, Montgomery MT. 2008. Observed structure, evolution, and potential intensity of Category 5 hurricane *Isabel* (2003) from 12 to 14 September. *Mon. Weather Rev.* **65**: 2025–2046.

Bister M, Emanuel KA. 1998. Dissipative heating and hurricane intensity. *Meteorol. Atmos. Phys.* **65**: 233–240.

Black PG, D'Asoro EA, Drennan WM, French JR, Niller PP, Sanford TB, Terril EJ, Walsh EJ, Zhang JA. 2007. Air–sea exchange in hurricanes. Synthesis of observations from the coupled boundary layer air–sea transfer experiment. *Bull. Amer. Meteorol. Soc.* **88**: 357–374.

Blackadar AK. 1976. Modelling the nocturnal boundary layer. Third Symp. on Atmospheric Turbulence, Diffusion and Air Quality. Raleigh NC, *Amer. Meteorol. Soc.* 46–49.

Braun SA, Tao W-K. 2000. Sensitivity of high-resolution simulations of hurricane *Bob* (1991) to planetary boundary-layer parameterizations. *Mon. Weather Rev.* **128**: 3941–3961.

Bryan GH, Rotunno R. 2009a. The maximum intensity of tropical cyclones in axisymmetric numerical model simulations. *Mon. Weather Rev.* **137**: 1170–1789.

Bryan GH, Rotunno R. 2009b. Evaluation of an analytical model for the maximum intensity of tropical cyclones. *J. Atmos. Sci.* **66**: 3042–3060.

Bui HH, Smith RK, Montgomery MT, Peng J. 2009. Balanced and unbalanced aspects of tropical-cyclone intensification. *Q. J. R. Meteorol. Soc.* **135**: 1715–1731.

Burk SD, Thompson WT. 1982. Operational evolution of a turbulence closure model forecast system. *Mon. Weather Rev.* **110**: 1535–1543.

Craig GC, Gray SL. 1996. CISK or WISHE as a mechanism for tropical cyclone intensification. *J. Atmos. Sci.* **53**: 3528–3540.

Cram TA, Persing J, Montgomery MT, Braun S. 2007. A Lagrangian trajectory view on transport and mixing processes between the eye, eyewall, and environment using a high-resolution simulation of hurricane *Bonnie* (1998). *J. Atmos. Sci.* **64**: 1835–1856.

Donelan MA, Haus BK, Reul N, Plant WJ, Stiassne M, Graber HC, Brown OB, Saltzman ES. 2004. On the limiting aerodynamic roughness in very strong winds. *Geophys. Res. Lett.* **31**: L18306, DOI: 10.1029/2004GL019460.

Drennan WM, Zhang JA, French JR, McCormick C, Black PB. 2007. Turbulent fluxes in the hurricane boundary layer. Part II: Latent heat fluxes. *J. Atmos. Sci.* **64**: 1103–1115.

Dudhia J. 1993. A non-hydrostatic version of the Penn State-NCAR mesoscale model: Validation tests and simulation of an Atlantic cyclone and cold front. *Mon. Weather Rev.* **121**: 1493–1513.

Emanuel KA. 1986. An air–sea interaction theory for tropical cyclones. Part I: Steady state maintenance. *J. Atmos. Sci.* **43**: 585–604.
Emanuel KA. 1995a. Sensitivity of tropical cyclones to surface exchange coefficients and a revised steady-state model incorporating eye dynamics. *J. Atmos. Sci.* **52**: 3969–3976.
Emanuel KA. 1995b. The behavior of a simple hurricane model using a convective scheme based on subcloud-layer entropy equilibrium. *J. Atmos. Sci.* **52**: 3960–3968.
Emanuel KA. 1997. Some aspects of hurricane inner-core dynamics and energetics. *J. Atmos. Sci.* **54**: 1014–1026.
Emanuel KA. 2003. Tropical cyclones. *Ann. Rev. Earth Planet. Sci.* **31**: 75–104.
Emanuel KA. 2003. A similarity hypothesis for air–sea exchange at extreme wind speeds. *J. Atmos. Sci.* **60**: 1420–1428.
Emanuel KA. 2004. Tropical cyclone energetics and structure. In: *Atmospheric Turbulence and Mesoscale Meteorology*, Fedorovich E, Rotunno R, Stevens B. (eds.) Cambridge University Press: 165–191.
Frank WM. 1977. The structure and energetics of the tropical cyclone. Part II. Dynamics and thermodynamics. *Mon. Weather Rev.* **105**: 1136–1150.
French JR, Drennan WM, Zhang JA, Black PB. 2007. Turbulent fluxes in the hurricane boundary layer. Part II: Momentum fluxes. *J. Atmos. Sci.* **64**: 1089–1102.
Grell GA, Dudhia J, Stauffer DR. 1995. A description of the fifth generation Penn State/NCAR mesoscale model (MM5). NCAR Tech Note NCAR/TN-398+STR, 138 pp.
Haus BK, Jeong D, Donelan MA, Zhang JA, Savelyev I. 2010. Relative rates of sea–air heat transfer and frictional drag in very high winds. *Geophys. Res. Lett.* **37**: L07802, DOI: 10.1029/2009GL042206.
Hendricks EA, Montgomery MT, Davis CA. 2004. On the role of 'vortical' hot towers in formation of tropical cyclone *Diana* (1984). *J. Atmos. Sci.* **61**: 1209–1232.
Hill KA, Lackmann GL. 2009. Analysis of idealized tropical cyclone simulations using the Weather Research and Forecasting model: Sensitivity to turbulence parameterization and grid spacing. *Mon. Weather Rev.* **137**: 745–765.
Hong S-Y, Pan H-L. 1996. Nonlocal boundary layer vertical diffusion in a medium-range forecast model. *Mon. Weather Rev.* **124**: 2322–2339.
Jordan CL. 1958. Mean soundings for the West Indies area. *J. Meteorol.* **15**: 91–97.
Kepert JD. 2006a. Observed boundary-layer wind structure and balance in the hurricane core. Part I. Hurricane *Georges*. *J. Atmos. Sci.* **63**: 2169–2193.
Kepert JD. 2006b. Observed boundary-layer wind structure and balance in the hurricane core. Part II. Hurricane *Mitch*. *J. Atmos. Sci.* **63**: 2194–2211.
Malkus JS, Riehl H. 1960. On the dynamics and energy transformations in steady-state hurricanes. *Tellus* **12**: 1–20.
Marks FD, Black PG, Montgomery MT, Burpee RW. 2008. Structure of the eye and eyewall of hurricane *Hugo* (1989). *Mon. Weather Rev.* **136**: 1237–1259.
Montgomery MT, Nicholls ME, Cram TA, Saunders AB. 2006a. A vortical hot tower route to tropical cyclogenesis. *J. Atmos. Sci.* **63**: 355–386.
Montgomery MT, Bell MM, Aberson SD, Black ML. 2006b. Hurricane *Isabel* (2003): New insights into the physics of intense storms. Part I: Mean vortex structure and maximum intensity estimates. *Bull. Am. Meteorol. Soc.* **87**: 1335–1347.
Montgomery MT, Nguyen SV, Smith RK. (M2) 2009. Do tropical cyclones intensify by WISHE? *Q. J. R. Meteorol. Soc.* **135**: 1697–1714.
Moss M, Rosenthal S. 1975. On the estimation of planetary boundary layer variables in mature hurricanes. *Mon. Weather Rev.* **103**: 980–988.
Nguyen SV, Smith RK, Montgomery MT. (M1) 2008. Tropical-cyclone intensification and predictability in three dimensions. *Q. J. R. Meteorol. Soc.* **134**: 563–582.
Ooyama KV. 1969. Numerical simulation of the life cycle of tropical cyclones. *J. Atmos. Sci.* **26**: 3–40.
Powell, MD Vickery PJ, Reinhold TA. 2003. Reduced drag coefficient for high wind speeds in tropical cyclones. *Nature* **422**: 279–283.
Rosenthal SL. 1971. The response of a tropical cyclone model to variations in boundary layer parameters, initial conditions, lateral boundary conditions, and domain size. *Mon. Weather Rev.* **99**: 767–777.
Rotunno R, Emanuel KA. 1987. An air–sea interaction theory for tropical cyclones. Part II: Evolutionary study using a nonhydrostatic axisymmetric numerical model. *J. Atmos. Sci.* **44**: 542–561.
Smith RK. 2003. A simple model of the hurricane boundary layer. *Q. J. R. Meteorol. Soc.* **129**: 1007–1027.
Smith RK. 2006. Accurate determination of a balanced axisymmetric vortex. *Tellus* **58A**: 98–103.
Smith RK, Montgomery MT. 2008. Balanced boundary layers in hurricane models. *Q. J. R. Meteorol. Soc.* **134**: 1385–1395.
Smith RK, Montgomery MT. 2010. Hurricane boundary-layer theory. *Q. J. R. Meteorol. Soc.* **136**: 1665–1670, DOI: 10.1002/qj.679.
Smith RK, Vogl S. 2008. A simple model of the hurricane boundary layer revisited. *Q. J. R. Meteorol. Soc.* **134**: 337–351.
Smith RK, Montgomery MT, Zhu H. 2005. Buoyancy in tropical cyclones and other rapidly rotating vortices. *Dyn. Atmos. Oceans* **40**: 189–208.
Smith RK, Montgomery MT, Vogl S. 2008. A critique of Emanuel's hurricane model and potential intensity theory. *Q. J. R. Meteorol. Soc.* **134**: 551–561.
Smith RK, Montgomery MT, Nguyen SV. (M3) 2009. Tropical cyclone spin-up revisited. *Q. J. R. Meteorol. Soc.* **135**: 1321–1335.
Terwey W, Montgomery MT. 2008. Secondary eyewall formation in two idealized, full-physics modeled hurricanes. *J. Geophys. Res.* **113**: D12112, DOI: 10.1029/2007JD008897.
Vogl S, Smith RK. 2009. A simple model of the hurricane boundary layer revisited. *Q. J. R. Meteorol. Soc.* **135**: 337–351.
Wang Y, Wu C-C. 2004. Current understanding of tropical cyclone structure and intensity changes – A review. *Meteorol. Atmos. Phys.* **87**: 257–278.
Zhang JA, Black PB, French JR, Drennan WM. 2008. First direct measurements of enthalpy fluxes in the hurricane boundary layer: The CBLAST result. *Geophys. Res. Lett.* **35**: L14813, DOI: 10.1029/2008GL034374.
Zhang JA, Drennan WM, Black PB, French JR. 2009. Turbulence structure of the hurricane boundary layer between the outer rainbands. *J. Atmos. Sci.* **66**: 2455–2467.

