SENSITIVITY OF TROPICAL CYCLONES TO SURFACE EXCHANGE COEFFICIENTS AND A REVISED STEADY-STATE MODEL INCORPORATING EYE DYNAMICS

Kerry A. Emanuel

Center for Meteorology and Physical Oceanography
Massachusetts Institute of Technology
Cambridge, MA 02139

Abstract

Numerical and theoretical models of tropical cyclones indicate that the maximum wind speed in mature storms is sensitive to the ratio of the enthalpy and momentum surface exchange coefficients, and that the spin-up time of tropical cyclones varies inversely with the magnitude of these coefficients. At the same time, the Carnot cycle model developed by the author predicts that the central pressure of mature cyclones is independent of the the magnitude of the exchange coefficients. We here present numerical simulations that prove this last prediction false and suggest that the reason for this failure is the neglect of eye dynamics in the steady-state theory. On this basis, the existing theory is modified to account for eye dynamics, and the predictions of the revised theory are compared to the results of numerical simulations.

Both the revised theory and the numerical modeling results, when compared to observations, suggest that the ratio of enthalpy to momentum exchange coefficients in real hurricanes lies in the range 0.75 to 1.5, contradicting published speculations about the behavior of this ratio at high surface wind speed.

1. Introduction

Tropical cyclones intensify and maintain themselves against dissipation by extracting heat from the ocean at high temperature, and exporting it at the low temperature of the tropical lower stratosphere. One would thus expect the characteristics of these storms to be somewhat sensitive to the details of the heat exchange process at the sea surface. In spite of this, the author (Emanuel, 1986, 1988) suggested that a reasonable lower bound on central pressure could be calculated without knowing the values of the exchange coefficients. The bounds so calculated agree quite well with the observed central pressures of the very most intense storms on record.

At the same time, scale analysis of the governing equations (Emanuel, 1989, 1995) suggests that the time scale over which tropical cyclones develop depends inversely on the magnitude of the surface exchange coefficients. Also, an analytic model of the structure of mature cyclones (Emanuel, 1986; hereafter E86) predicts that the maximum azimuthal wind speed varies as $\sqrt{C_k/C_D}$, where C_k is the exchange coefficient of heat and water (assumed equal), and C_D is the surface drag coefficient. The model also predicts that the ratio of an outer scale to the radius of maximum winds and other aspects of the structure of mature cyclones are sensitive to C_k/C_D.

Ooyama (1969) performed two experiments with his two-layer axisymmetric model, allowing C_k and C_D to differ. In one experiment, C_k was allowed to increase with wind speed while C_D was held fixed, and in the other, the wind speed dependence of C_D was retained while C_k was held fixed. The maximum wind speed in the first case exceeded 95 m s^{-1} but was only 45 m s^{-1} in the second case. When the exchange coefficients were equal the maximum wind speed was 60 m s^{-1}. This suggests a strong sensitivity of maximum wind speed to C_k/C_D.

Rosenthal (1971) undertook a series of numerical experiments to test the sensitivity of the development and intensity of tropical cyclones to individual values of the heat, moisture and momentum exchange coefficients. He used a seven-layer axisymmetric, hydrostatic primitive equation model with a Kuo-like representation of moist convection. In experiments in which the drag coefficient was systematically increased and decreased while holding fixed the heat and moisture exchange coefficients, the *rate* of intensification varied directly with C_D, as expected from scale analysis of the equations, while the final intensity varied inversely with C_D. Likewise, increasing the heat and moisture transfer coefficients increased the maximum wind speed of the mature storm, while decreasing them had the opposite effect.

Research on the sensitivity of tropical cyclones to surface exchange coefficients is severely hampered by lack of surface flux data at very high wind speeds. There have been several suggestions (e.g., Liu et al., 1979) that as wind speeds exceed about 12 m s^{-1}, the drag coefficient increases relative to the heat and water exchange coefficients, owing to the effects of form drag by ocean surface waves. Measurements at wind speeds near 15 m s^{-1} (e.g., Geernaert et al., 1987) seem to suggest that the drag coefficient is nearly twice that of the heat and moisture exchange coefficients at these wind speeds, although there is a great deal of scatter in the data.

On the other hand, there is no indication that the heat and water exchange coefficients differ appreciably from each other in the range of wind speeds within which they have been estimated (Katsaros, personal communication). This is an important point, for the energetically important transfer at the sea surface is of enthalpy, k, defined

$$k \equiv \left[c_{pd}(1-q) + c_l q\right] T + L_v q, \tag{1}$$

where k is the enthalpy per unit mass of dry air, c_{pd} and c_l are the heat capacities of dry air at constant pressure and of liquid water, respectively, T is the absolute temperature, q is the specific humidity, and L_v is the latent heat of vaporization. According to the aerodynamic flux formulae, the flux of enthalpy from the surface is given by

$$F_k = \rho_a |\mathbf{V}_a| \left[C_T(c_{pd}(1-q_a) + c_l q_a)(T_s - T_a) + C_q(L_v + c_l T_a)(q_s^* - q_a)\right], \tag{2}$$

where ρ_a is the density of dry air at the sea surface, $|\mathbf{V}_a|$ is the wind speed at some predefined level, T_s and T_a are the temperature of the sea surface and the air temperature, q_s^* and q_a are the saturation specific humidity at sea surface temperature and pressure and the actual mixing ratio of the air, and C_T and C_q are the transfer coefficients for heat and water. These coefficients are functions of surface roughness and stability.

If, as observations suggest, C_T can be set equal to C_q, (2) can be rewritten

$$F_k = \rho_a |\mathbf{V}_a| C_k (k_s^* - k_a), \tag{3}$$

where k_s^* is the saturation enthalpy at the sea surface, k_a is the enthalpy of the air and C_k is defined as the enthalpy transfer coefficient. This shows, among other things, that evaporation of spray cannot directly affect the enthalpy transfer from the ocean if $C_T = C_q$, since this process affects neither k_s^* nor k_a. For the purposes of the present work, we shall assume that $C_T = C_q$ and refer instead to the enthalpy transfer coefficient C_k that appears in (3).

We begin by reviewing the steady-state theory of E86, which predicts that the maximum wind speed, but not the central pressure, of mature hurricanes depends on the ratio C_k/C_D. In section 3 the predictions of this theory are compared to the results of numerical simulations with two different models, and it is shown that while the predictions of maximum wind speed are quite good, those of central pressure are poor. This leads us, in section 4, to revise the theory to account for the special dynamics of the eye; the revised theory is then shown to yield much better predictions of central pressure. These results are summarized in the concluding section.

2. Review of steady-state theory

The possible effect of differing heat and momentum exchange coefficients on tropical cyclones was addressed by the author (E86) and we begin by giving a somewhat simplified account of that analysis. I assumed that tropical cyclones could be divided into three regions (Fig. 5 of E86) categorized by the subcloud-layer entropy budget and eye dynamics. The first region, the eye, was assumed to be mechanically driven by flow in the eyewall. The second region, the eyewall, was characterized as a region in which the subcloud-layer entropy budget is dominated by surface heat fluxes and radial advection, with little turbulent flux through the subcloud-layer top. In contrast, in region 3 it was assumed that turbulent fluxes through the top of the subcloud layer nearly balance surface fluxes, as proved to be the case in the later modeling study of Rotunno and Emanuel (1987, hereafter RE87). For simplicity, E86 assumed that the net effect of this balance in region 3 was to maintain the subcloud-layer relative humidity at a value that remains constant with radius.

These assumptions about subcloud-layer thermodynamics, coupled with the assumption that the flow above the subcloud layer is neutral to slantwise convection and that the absolute vorticity at the storm top vanishes, allowed E86 to derive an expression for the maximum wind speed in mature tropical cyclones, and a relationship between the radius of maximum winds and an outer radius at which the storm-associated wind vanishes. In deriving these expressions, E86 ignored the special dynamics associated with the eye itself and assumed moist neutrality there.

For clarity, we briefly review the derivation of these relations, but use a nondimensional form of the E86 equations. The scaling factors appear in Table 1 of the companion paper (Emanuel, 1995, hereafter E95). The equations are phrased in the potential radius coordinate R, defined such that (in nondimensional terms)

$$\frac{1}{2}R^2 = rV + \frac{1}{2}r^2, \tag{4}$$

where r is physical radius and V is the azimuthal wind. R^2 is thus proportional to the absolute angular momentum per unit mass.

The nondimensional equations that are assumed to hold everywhere in steady conditions are as follows:

Thermal wind:

$$\frac{1}{r^2} = -\frac{2}{R^3}\frac{\partial \chi}{\partial R}, \tag{5}$$

Gradient wind relation (coupled with thermal wind):

$$P + \chi + \frac{1}{8}\left(\frac{R^4}{r^2} + r^2\right) = \frac{1}{4}r_0^2, \tag{6}$$

Saturation entropy of sea surface:

$$\chi_s^* = 1 - AP, \tag{7}$$

with

$$A \equiv \frac{T_s - T_o}{T_s} + \frac{\chi_s}{R_d T_s(1-\mathcal{H})}. \tag{8}$$

The relationship (7) has been linearized assuming small fractional departures of surface pressure from its ambient value. The nondimensional variables that appear in (5)–(8) are r, the radius of R surfaces in the subcloud layer, χ, the subcloud-layer entropy variable, and P, the departure of surface pressure from its ambient value. The parameters that appear in (5)–(8) are r_0, the outer radius at which the surface wind vanishes, T_s, the surface temperature, T_o, the outflow temperature, χ_s, the ambient entropy deficit of the subcloud layer with respect to the ocean, R_d, the gas constant of dry air, and $\mathcal{H}$, the ambient relative humidity. (See the companion paper, E95, for a complete description of these variables and parameters.)

In both (5) and (6), it has been assumed that the storm is mature, so that the radius of angular momentum surfaces at the tropopause is very large.

a. Inner region

In the inner region, it is assumed that radial entropy advection balances the surface fluxes. For the present, it will be assumed that this region covers both the eyewall and the eye itself, though the assumed balance is strictly valid only in the eyewall.

The dimensionless, steady entropy equation in the inner region, modified to account for $C_k \neq C_D$, is (see E95, Eq. A6)

$$\psi_0 \frac{\partial \chi}{\partial r^2} = -\frac{C_k}{C_D}(1 + c|V|)|V|(\chi_s^* - \chi), \tag{9}$$

where ψ_0 is the streamfunction at the top of the subcloud layer and c is a coefficient that determines the wind speed-dependence of the surface exchange coefficients.

This states that in the inner region's subcloud layer, radial entropy advection balances surface entropy flux.

Analogously, the balance of angular momentum in the subcloud layer is (see E86)

$$\psi_0 \frac{\partial R^2}{\partial r^2} = 2(1 + c|V|)|V|rV. \tag{10}$$

This states that radial advection balances frictional loss of angular momentum. Eliminating ψ_0 between (9) and (10) and using (4) gives

$$\frac{1}{2R}(R^2 - r^2)\frac{\partial \chi}{\partial R} = -\frac{C_k}{C_D}(\chi_s^* - \chi). \tag{11}$$

In the inner region, we assume that $r^2 \ll R^2$, which is equivalent to neglecting the Coriolis acceleration there. Using this approximation together with the thermal wind relation (5) in (11) gives

$$\frac{1}{4}\frac{R^4}{r^2} \simeq \frac{C_k}{C_D}(\chi_s^* - \chi). \tag{12}$$

Finally, using (4) and again assuming that $r^2 \ll R^2$ in (12) results in

$$V^2 \simeq \frac{C_k}{C_D}(\chi_s^* - \chi). \tag{13}$$

Thus in the inner region, the square of the azimuthal velocity is proportional to the *local* degree of thermodynamic disequilibrium between the subcloud layer and the sea surface.

Next, χ_s^* is eliminated from (13) using the saturation law (7), and from the resulting equation, pressure, P, is eliminated using cyclostrophic balance (6, with the approximation $r^4 \ll R^4$). This yields

$$V^2 \left[1 - \frac{1}{2} A \frac{C_k}{C_D}\right] \simeq \frac{C_k}{C_D} \left[1 - \frac{1}{4} A r_0^2 - \chi(1 - A)\right]. \tag{14}$$

Finally, to match this solution with that (yet to be derived) of the outer region, χ is assumed to decrease with radius such that the relative humidity becomes as low as its undisturbed value, $\mathcal{H}$, at the radius of maximum winds. To a good approximation, $\mathcal{H}$, by its definition, is related to χ and χ_s^* by

$$\chi = \mathcal{H}(\chi_s^* - 1). \tag{15}$$

Using this together with (7) and (6) in (14) yields an expression for the maximum azimuthal velocity:

$$V_m^2 = \frac{C_k}{C_D} \left[\frac{1 - \frac{1}{4}\gamma r_0^2}{1 - \frac{1}{2}\frac{C_k}{C_D}\gamma}\right], \tag{16}$$

where

$$\gamma \equiv A \frac{1 - \mathcal{H}}{1 - \mathcal{H} A}. \tag{17}$$

This is identical to Eq. (43) of E86, except that it is in nondimensional form.

Using (5) and (4) with the approximation $r^2 \ll R^2$ gives

$$\frac{\partial \chi}{\partial R} \simeq -\frac{2V^2}{R}.$$

By eliminating $\partial\chi/\partial R$ between this and (14), one obtains a simple differential equation for the radial distribution of V, whose solution is

$$V = V_m \left(\frac{R}{R_m}\right)^{\nu}, \tag{18}$$

where R_m is the potential radius of maximum wind and

$$\nu \equiv \frac{C_k}{C_D} \frac{1-A}{1-\frac{1}{2}A\frac{C_k}{C_D}}. \tag{19}$$

Thus the radial distribution of wind, pressure (and entropy) in the inner region depends on the ratio C_k/C_D, as does the magnitude of the maximum wind speed. But it must be noted that ν, as it appears in (18), cannot be greater than or equal to 2, else the coordinate transformation (4) fails. (This implies, from (19), that C_k/C_D cannot exceed 2.) In the numerical models, internal disspiation prevents discontinuities from forming. Thus we should expect some disagreement between this theory and the numerical models for C_k/C_D approaching or exceeding 2.

The minimum central pressure in this solution can be obtained directly from (6) and (7) at $R = r = 0$, by assuming that $\chi = \chi_s^*$ there. This gives

$$P_c = -\frac{1-\frac{1}{4}r_0^2}{1-A}. \tag{20}$$

This is the dimensionless equivalent of Eq. (27) of E86. Note that P_c *does not depend on* C_k *or* C_D in this formula.

For now, these solutions are assumed to be valid not only in the eyewall, where the propositions of slantwise neutrality and the near balance of surface enthalpy fluxes and radial advection are likely to be valid, but also in the eye itself, where they may not be. As will soon become apparent, we shall be forced to revise these relations where they pertain to the eye itself.

b. Outer region

In the outer region, we find an exact solution of (5)–(7) and (15) under the proposition that the relative humidity, $\mathcal{H}$, is constant, and satisfying $r = R$ ($V = 0$) at $R = r_0$, where r_0 is defined as the *outer radius.* This solution can be expressed as an algebraic relation between r and R:

$$R^4(2 - \mathcal{H}A) = 2r_0^4 \left(\frac{r}{r_0}\right)^{2\mathcal{H}A} - \mathcal{H}Ar^4. \qquad (21)$$

The solution of (21) for R can be used in (4) to give V as a function of r.

At the radius of maximum winds, this solution is matched to the solution in the inner region. The entropy is automatically matched as we have assumed a relative humidity of $\mathcal{H}$ at the radius of maximum winds in both solutions. We also demand that the dependent variable r (or equivalently, V) be matched across R_m, the potential radius of maximum winds. At the radius of maximum winds in a reasonably strong storm, the last term in (21) is negligible compared to the other terms. Using this and the relation from (4)

$$V^2 \simeq \frac{1}{4}\frac{R^4}{r^2}$$

to eliminate R^4 in (21) yields a specific relationship between the physical radius of maximum winds, r_m, and r_0:

$$r_0^{4-2\mathcal{H}A} = 4r_m^{2-2\mathcal{H}A}V_m^2\left(1 - \frac{1}{2}\mathcal{H}A\right), \qquad (22)$$

with V_m^2 given by (16). This is nearly the dimensionless equivalent of Eq. (46) of E86. Note that the dependence on C_k/C_D enters through the expression for V_m^2.

3. Model results

To test the predictions of the sensitivity of tropical cyclones to the ratio C_k/C_D, we ran both the simple, balance model of E95 and the nonhydrostatic, convection-resolving model of RE87 for different values of the ratio. In each case, the model was run for long enough that quasi-steady mature model storms were achieved. Average values of the maximum wind speed, radius of maximum winds, and central pressure during the mature stage of the model cyclones were estimated.

The very gradual radial decay of velocity made it nearly impossible to estimate r_0 in the solutions, so that (22) could not be tested in a reasonable way. Instead, (22) was used to predict r_0 given the value of r_m observed in the model and an initial estimate of V_m. This predicted value of r_0 was then used in the maximum wind equation (16); the solution for V_m was then used once again in (22) and a revised estimate of r_0 was produced, and so on. Since the terms involving r_0 in (16) proved to be quite small, this procedure always converged rapidly and made little difference to the results. The final estimate of r_0 was then used in the central pressure equation (20).

Note that the theoretical solution does not yield a specific scale (r_0 or r_m) for hurricanes; only a relationship (22) has been obtained between the two scales r_0 and r_m. Thus one scale (in this case, r_m) must be specified from the model results to complete the theoretical calculation. The indeterminacy of radial scales in the theory may correspond to the observation that real tropical cyclones occur with a wide variety of horizontal scales.

The comparison between the theoretical predictions and the two independent model results are shown in Figures 1 and 2 and in Tables 1 and 2 for six values of C_k/C_D. Since the model of RE87 is formulated in dimensional terms, the results of the theory and the simple model are here presented in dimensional form, using values of the scaling factors (Table 1 of E95) estimated from the environmental conditions of the experiments with the nonhydrostatic model. Note slight differences between the theoretical values listed in Tables 1 and 2; this is because r_m (and therefore r_0) differs between the models. (This small difference has been neglected in the graphs in Figures 1 and 2.)

Figure 1 shows that the model predictions of the maximum azimuthal velocity, V_m, agree quite well with each other and with theory, and that V_m varies roughly as $\sqrt{C_k/C_D}$. This is also in accord with the results obtained by Ooyama (1969) and Rosenthal (1971). The central pressures, on the other hand, agree less well between the two models and depart dramatically from the theoretical prediction except when $C_k/C_D \simeq 1.2$–1.5 (Figure 2). Both models show a strong sensitivity of central pressure to C_k/C_D, in contrast with the theoretical prediction of no dependence of P_c on C_k/C_D. Clearly, something is wrong with the theory as it pertains to the eye of the storm.

4. Revised theory of the eye region

The weakest aspect of the theory presented in E86 and repeated in section 2 of this paper is the assumption that the radial inflow continues to the center of the storm and that the eye itself is neutral to slantwise convection. In real storms, a strong temperature inversion is often present at low levels in the eye, and there is usually (though not always) a complete lack of deep convection. The theoretical model needs to be revised to account for the distinct dynamics of the eye.

One clue to a means of restructuring the theory is that the theoretical prediction of central pressure matches the modeling results when C_k/C_D is in the range 1.2–1.5. Now note, from (18) and (19), that the azimuthal velocity will depend on the first power of physical radius when

$$\frac{C_k}{C_D} = \frac{1}{1 - \frac{1}{2}A}, \tag{23}$$

which for the parameters used in the models, has a value of 1.37. Thus it happens that the prediction of central pressure is quite good for that particular value of C_k/C_D, given by (23), that gives a linear dependence of velocity on radius within the eye and eyewall.

Why should other values of C_k/C_D, which yield nonlinear dependence of V on radius in the eye, give poor predictions of central pressure? We hypothesize that in real storms and in the numerically modeled storms, *a viscous, cyclostrophic adjustment occurs that preserves the value of the maximum azimuthal wind and relaxes the velocity profile toward a linear dependence on radius.* Such a profile has zero divergence of the radial turbulent flux of angular momentum, according to virtually all formulations of three-dimensional turbulence including those used in both models. Thus in our revised formulation of the eye, *the azimuthal velocity is assumed to vary linearly with radius, up to a maximum value given by (16), and to be in cyclostrophic balance.* The preservation of V_m during the hypothesized adjustment is merely a guess, given the generally good fit between the original theoretical predictions of V_m and the values given by both models.

Accordingly, we accept the predictions of both V_m and P_m, the azimuthal velocity and surface pressure at the radius of maximum winds, given by the theory of E86 and recapped in section 2 of this paper. The value of P_m is given by (6) and (7) together with the relation (15), the neglect of r_m^4 compared to R_m^4, and the approximation $V_m \simeq \frac{1}{2}R_m^2/r_m$:

$$P_m \simeq \frac{\frac{1}{4}r_0^2 - \frac{1}{2}V_m^2}{1 - \mathcal{H}A}, \tag{24}$$

with V_m^2 given by (16).

Now, based on the proposed viscous adjustment, we assume that the azimuthal velocity in the eye ($R < R_m$) is given by

$$V = V_m \frac{R}{R_m}. \tag{25}$$

(Note that through (4), this also implies that V varies linearly with r.) Using the cyclostrophic relation

$$\frac{d}{dR}[P + \frac{1}{2}V^2] = 2\frac{V^2}{R}$$

in (25) and integrating with radius to R_m gives

$$P_c \simeq P_m - \frac{1}{2}V_m^2. \tag{26}$$

Finally, combining (26) with (24) gives a revised prediction of central pressure:

$$P_c \simeq -\frac{V_m^2\left(1-\frac{1}{2}\mathcal{H}A\right)-\frac{1}{4}r_0^2}{1-\mathcal{H}A}, \tag{27}$$

with V_m^2 given by (16). Because of its dependence on V_m^2, P_c is now dependent on C_k/C_D, unlike the original formula, (20).

Table 3 and Figure 3 compare revised predictions of the central pressure from (27) with results of the two models. Clearly, the agreement is quite good and much improved over the previous model.

The sensitive dependence of both the maximum wind speed and minimum pressure on C_k/C_D can be understood through the subcloud layer entropy budget. When this ratio is small, the Ekman inflow is relatively large compared to the rate of transfer of enthalpy from the ocean, and the total enthalpy gain is thus relatively small. The reduced heat input to the system results in smaller intensity. Conversely, when C_k/C_D is large, the enthalpy gain is larger, and so is the storm intensity. Note that in the revised model, most of the enthalpy gain in the subcloud layer occurs in the relatively narrow ring of the eyewall, with little contribution from the outer region and none from the eye itself. This is consistent with observed θ_e distributions in tropical cyclones (e.g. see Hawkins and Imbembo, 1976).

5. Discussion and summary

Results of integrating two very different numerical models show that *both* the maximum azimuthal velocity and the central pressure deficit of tropical cyclones depend on the ratio of the enthalpy to momentum exchange coefficients, C_k/C_D. This dependence of the central pressure on C_k/C_D is inconsistent with the steady-state model of E86, but these and previous modeling results are consistent with a revised analytic model of mature hurricanes that divides them into three regions: an outer region in which the interior flow is nearly neutral to slantwise moist convective and in which turbulent entropy fluxes through the top of the subcloud layer keep the subcloud layer relative humidity nearly constant with radius; a narrow eyewall that is also neutral to slantwise convection but which has a subcloud-layer entropy distribution determined by a balance between surface fluxes and radial advection; and an eye in solid body rotation and cyclostrophic balance.

Using the unrevised theory, E86 and Emanuel (1988) showed that theoretical predictions of the lower bound on central pressure of tropical cyclones, made using climatological conditions, agree quite well with the central pressures of the very most intense tropical cyclones on record. The direct implication of this result in the context of the present study is that the ratio C_k/C_D is most likely to lie in the range 1.2–1.5 in the high wind region of these intense storms. *In no event are the results from either model consistent with values of C_k/C_D less than about 3/4;* otherwise, the wind speeds would be much weaker than observed. This result contradicts current thinking about the behavior of the heat and momentum exchange coefficients at high wind speeds (e.g., Liu et al., 1979), though few if any measurements have been made at wind speeds much in excess of 20 m s^{-1}. It might be speculated that the observed downward trend of C_k/C_D with increasing surface winds reverses at higher wind speeds, though it is not obvious why this should happen. In any event, a reconciliation of the present results with ideas pertaining to C_k/C_D based on existing measurements would no doubt result in improved understanding of both tropical cyclones and air-sea interaction.

Given the strong dependence of hurricane structure and intensity on C_k/C_D indicated here, and the very incomplete understanding of the physics of heat and momentum exchange between the ocean and atmosphere at very high wind speeds, it is not inconceivable that real tropical cyclones are affected by naturally occurring or man-made surfactants, which are known to affect transfer rates at lower wind speeds. This offers the small hope that tropical cyclone intensity may be artificially reduced through the application of an appropriate surfactant to the sea surface along the path of the high wind region of the storm, as first suggested by Simpson and Simpson (1966).

Acknowledgments. This research was supported by the National Science Foundation under Grant ATM–9216906. The author thanks Alan Betts, David Raymond and an anonymous referee for their thoughtful critiques, and Kristina Katsaros for helpful discussions of surface fluxes at high wind speeds.

References

Emanuel, K. A., 1986: An air-sea interaction theory for tropical cyclones. Part I. *J. Atmos. Sci.,* **42,** 1062–1071.

Emanuel, K. A., 1988: The maximum intensity of hurricanes. *J. Atmos. Sci.,* **45,** 1143–1155.

Emanuel, K. A., 1989: The finite-amplitude nature of tropical cyclogenesis. *J. Atmos. Sci.,* **46,** 3431–3456.

Emanuel, K. A., 1995: The behavior of a simple hurricane model using a convective scheme based on subcloud layer entropy equilibrium. *J. Atmos. Sci.,* **52**. Submitted.

Geernaert, G. L., S. E. Larsen, and F. Hansen, 1987: Measurements of the wind stress, heat flux and turbulence intensity during storm conditions over the North Sea. *J. Geophys. Res.,* **92,** 13127–13139.

Hawkins, H. F., and S. M. Imbembo, 1976: The structure of a small, intense hurricane–Inez 1966. *Mon. Wea. Rev.,* **104,** 418–442.

Liu, W. T., K. B. Katsaros, and J. A. Businger, 1979: Bulk parameterization of air-sea exchange of heat and water vapor including the molecular constraints at the interface. *J. Atmos. Sci.,* **36,** 1722–1735.

Ooyama, K., 1969: Numerical simulation of the life-cycle of tropical cyclones. *J. Atmos. Sci.,* **26,** 3–40.

Rosenthal, S. L., 1971: The response of a tropical cyclone model to variations in boundary layer parameters, initial conditions, lateral boundary conditions and domain size. *Mon. Wea. Rev.,* **99,** 767–777.

Rotunno, R., and K. A. Emanuel, 1987: An air-sea interaction theory for tropical cyclones. Part II. *J. Atmos. Sci.,* **44,** 542–561.

Simpson, R. H. and J. Simpson, 1966: Why experiment on tropical hurricanes? *Trans. New York Acad. Sci.,* **28,** 1045–1062.

Table 1. Comparison between theory and the E95 model

		p_c (mb)		V_m (m s^{-1})		r_0 (km)
C_k/C_D	r_m (km)†	Theory	Model	Theory	Model	Theory
.5	35	926	989	44	42	374
.75	20	926	971	54	50	366
1.0	21	926	957	63	59	398
1.25	18	926	945	72	68	413
1.5	17	927	928	80	78	433
2.0	15	927	880	95	95	486

†Observed in model and used in theory.

Table 2. Comparison between theory and the RE87 model

		p_c (mb)		V_m (m s^{-1})		r_0 (km)
C_k/C_D	r_m (km)†	Theory	Model	Theory	Model	Theory
.5	12	925	972	44	49	265
.75	25	926	957	54	54	397
1.0	30	927	942	63	66	468
1.25	30	928	930	72	72	506
1.5	30	928	918	80	80	541
2.0	30	929	905	94	85	603

†Observed in model and used in theory.

Table 3. Comparison of revised theory of central pressure (p_c) with results of both models.

	p_c (mb) (E95)		p_c (mb) (RE87)	
C_k/C_D	Theory	Model	Theory	Model
.5	984	989	983	972
.75	968	971	968	957
1.0	952	957	953	942
1.25	935	945	936	930
1.5	918	928	919	918
2.0	881	880	883	905

Figure captions

Figure 1. Maximum azimuthal wind according to equation (16)(solid curve) and results of running the E95 (dots) and RE87 (+'s) models, as a function of the ratio of enthalpy to momentum surface exchange coefficients.

Figure 2. Same as Figure 1 but for central surface pressure according to equation (20).

Figure 3. Same as Figure 2 but showing the prediction of the revised model, equation (27).