CISK or WISHE as the Mechanism for Tropical Cyclone Intensification

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ABSTRACT

Examination of conditional instability of the second kind (CISK) and wind-induced surface heat exchange (WISHE), two proposed mechanisms for tropical cyclone and polar low intensification, suggests that the sensitivity of the intensification rate of these disturbances to surface properties, such as surface friction and moisture supply, will be different for the two mechanisms. These sensitivities were examined by perturbing the surface characteristics in a numerical model with explicit convection. The intensification rate was found to have a strong positive dependence on the heat and moisture transfer coefficients, while remaining largely insensitive to the frictional drag coefficient. CISK does not predict the observed dependence of vortex intensification rate on the heat and moisture transfer coefficients, nor the insensitivity to the frictional drag coefficient since it anticipates that intensification rate is controlled by frictional convergence in the boundary layer. Since neither conditional instability nor boundary moisture content showed any significant sensitivity to the transfer coefficients, this is true of CISK using both the convective closures of Ooyama and of Charney and Eliassen. In comparison, the WISHE intensification mechanism does predict the observed increase in intensification rate with heat and moisture transfer coefficients, while not anticipating a direct influence from surface friction.

1. Introduction

a. CISK and WISHE

There has been a long-running controversy in the study of tropical cyclones concerning the mechanism of intensification. The most frequently cited theory over the past 30 years has been conditional instability of the second kind, or CISK, introduced, in slightly different forms, by Ooyama (1964), and Charney and Eliassen (1964). Since that time, it has been modified and generalized to many other tropical phenomena, and some confusion has arisen regarding the precise meaning of the term (Ooyama 1982). This paper will not attempt a general review, preferring to consider CISK only as it has been applied to the intensification of warm-core vortices. This includes tropical cyclones but also certain intensely convective polar lows and Mediterranean cyclones, which resemble tropical cyclones and are believed to intensify by the same process (Rasmussen 1979, 1989; Rasmussen and Zick 1987).

A useful starting point for describing the CISK mechanism is given by Ooyama (1982, section 4), who describes CISK as a cooperative intensification involving organized moist convection and the cyclone-scale vortex. Radiation, surface fluxes and other forcings destabilize the tropical atmosphere, resulting in convection. If the convection is organized, it will lead to tropospheric warming and mass convergence in the lower levels that result in the warm-core vortex that defines the tropical cyclone. Cooperative intensification occurs when the vortex provides the necessary organization of the convection.

The spinup of the vortex in response to convective heating and mass fluxes occurs through a relatively well-understood process of adjustment to gradient balance. However, the control of the convection by the vortex-scale flow, the so-called cumulus parameterization or closure problem, is not well understood. Charney and Eliassen (1964) prescribed a closure that sets convective heating proportional to moisture convergence. Ooyama (1964, 1969) specified a convective mass flux proportional to frictional convergence in the boundary layer. In either case, instability results from a feedback, where a more intense vortex leads to stronger frictional convergence and increased convection, that in turn leads to further intensification of the vortex.

The term CISK is often used to denote the conceptual model described in the preceding paragraphs, independent of the assumptions about convective closure. However, as Ooyama (1982) notes, in order to obtain quantitative predictions of cyclone properties such as scale or growth rate, it is necessary to complete the theory by specifying a closure. In order to put the various theories to the test, such predictions are required. Therefore, for the purposes of this paper, CISK will be

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taken to represent the two theories of Charney and Eliassen (1964), and Ooyama (1964, 1969), including their respective closures.

The principal theory proposed as an alternative to CISK was introduced by Emanuel (1986) as air–sea interaction instability (ASII), but has more recently been renamed wind-induced surface heat exchange instability (WISHE) (e.g. Emanuel et al. 1994). In this theory the atmosphere is assumed to be neutrally stratified along sloping surfaces of constant angular momentum, although the difference between slantwise and vertical neutrality is not important for the present argument. Moist convection mixes air through the troposphere, but does not cause any temperature perturbation unless the boundary layer is anomalously heated as a result of surface fluxes of heat and moisture. The surface fluxes are wind speed dependent and, therefore, determined by the vortex-scale flow. Again, a feedback can occur, in which a more intense cyclone produces more rapid heating of the atmosphere organized in such a way as to intensify the cyclone, resulting in an instability. The key role of surface fluxes in controlling the intensification was earlier hypothesized by Malkus and Riehl (1960) on the basis of a model of the hurricane boundary layer.[1]

It may be seen that what distinguishes WISHE from CISK is the closure relating convective warming to the intensity of the cyclone vortex, which depends on surface heat and moisture fluxes and does not involve the organizing influence of frictional convergence (described as essential by Ooyama 1982).[2] If one chooses to regard CISK as a conceptual model independent of closure, WISHE could perhaps be taken to be another CISK theory, albeit one with a rather different closure from previous CISK theories. If, as in this paper, the closure is regarded as an intrinsic part of the theory, it is preferable to speak of the CISK theories of Ooyama, and of Charney and Eliassen, and the WISHE theory of Emanuel.

CISK and WISHE have existed in parallel for about ten years. The recent literature includes examples of both applied to tropical cyclones, both explicitly, and implicitly in analyses using moisture convergence or surface flux reasoning (e.g., McBride and Fraedrich 1995; Rogers et al. 1994; Barnes and Powell 1995). The debate has also extended to the hurricane-like polar lows mentioned previously (Rasmussen 1979; Emanuel and Rotunno 1989).

The persistence of the two modes of thought is due at least partly to the difficulty in distinguishing between them with currently available observations. Both theories predict the growth of warm-core cyclones satisfying thermal wind balance. While the descriptions of the processes leading to the cyclone have different emphases, strong surface fluxes, low-level convergence, and convection are all approximately collocated in space and time, making it difficult to identify cause and effect. Much interest has centered on the degree to which the atmosphere is unstable to moist convection, as measured by convective available potential energy (CAPE), but this has not provided a conclusive test. The WISHE theory assumes that the atmosphere is convectively neutral; however, this neutral condition will not correspond precisely to a moist adiabat obtained from undilute parcel ascent (zero CAPE). Although the energy source of the cyclone is through surface fluxes, a small amount of CAPE is not inconsistent with WISHE (Emanuel et al. 1994). In the CISK mechanism the vortex spins up from CAPE released in response to low-level convergence; however, the role of surface fluxes in maintaining the CAPE as the system develops is acknowledged (e.g., Ooyama 1969). Thus while CISK depends on the existence of CAPE, the values may not be large at any given moment. The nonzero, but not large, values of CAPE observed in tropical cyclones (e.g., Barnes and Powell 1995; Frank 1977) are consistent with either theory.

b. Timescales for intensification

If the predictions for the structure of the cyclone from the CISK and WISHE theories are not sufficiently different to decide which is acting, other predictions can nonetheless be considered. One possibility is to examine factors that control the rate of intensification. In particular, by identifying the rate-limiting process and the environmental factors that influence it, the dependence of growth rate on these factors can be anticipated. In nature the growth rate of a tropical cyclone is variable and difficult to observe precisely, but conditions that would lead to different growth rates may be possible to simulate in a numerical model.

Regardless of which theory is being considered, the ultimate energy source for a tropical cyclone is the sea surface. As depicted schematically in Fig. 1, this energy is transferred to the atmospheric boundary layer through fluxes of latent and sensible heat, and then to the troposphere by cumulus clouds. The collective effect of the clouds is represented in different ways in different cumulus parameterizations, for example as a vertical mass transfer (Ooyama 1964, 1969) or as a heat source (Charney and Eliassen 1964; Emanuel 1989), but in any case the result is a modified momentum or thermodynamic structure in the troposphere,

[1] As shown by Gray (1994), many aspects of the steady-state cyclone model of Emanuel (1986) were anticipated by Kleinschmidt (1951). However Kleinschmidt did not anticipate the WISHE mechanism, instead presenting moist symmetric instability as a necessary condition for cyclone intensification. An English translation of Kleinschmidt's paper is included as an appendix by Gray (1994).

[2] To our knowledge, low-level convergence, whether frictional or wave induced, has been present in all theories that have been described as CISK, including those not related to tropical cyclone intensification.

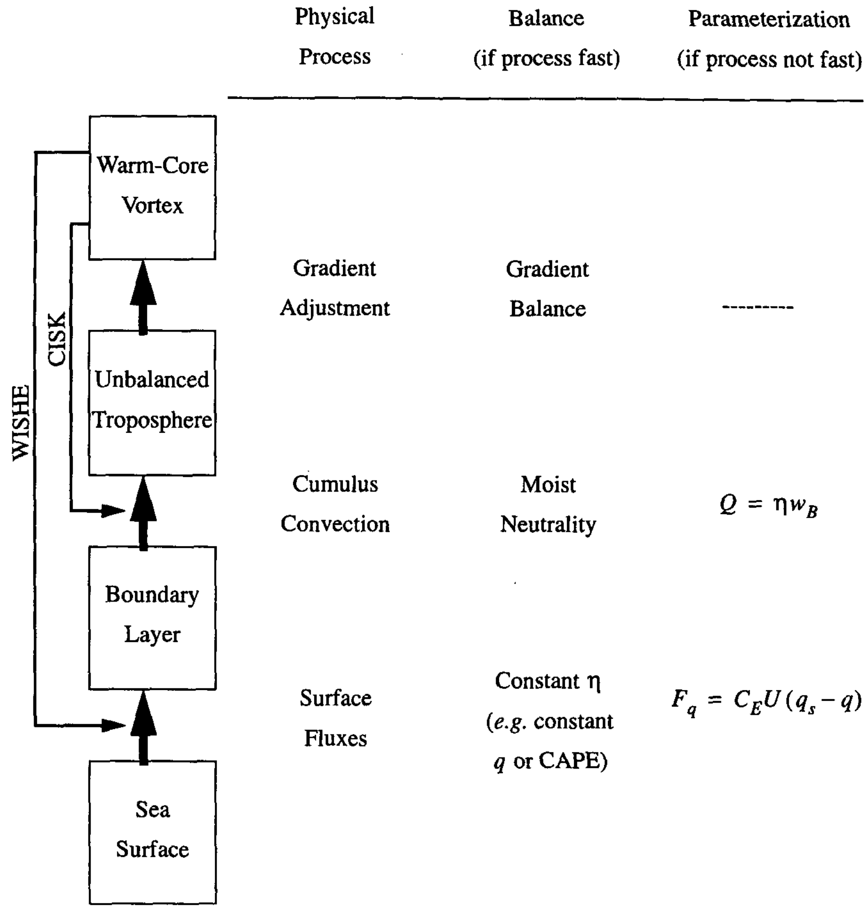

FIG. 1. Schematic diagram showing principle energy conversion processes in tropical cyclone intensification. For each of the three processes, the diagram shows the balance that would result if the process proceeded rapidly in comparison with the evolution of the large-scale vortex. The final column lists a simple parameterization that could be used to describe each process in terms of properties of the large-scale vortex (symbols defined in text). The principal rate-limiting processes in the CISK and WISHE theories are indicated by the arrows at the left of the diagram.

which is followed by an adjustment to gradient balance that produces the final warm-core vortex. Any of these processes, surface fluxes, convection, or gradient adjustment, could limit the rate of intensification of the cyclone. On the other hand, a process might evolve rapidly in comparison to the evolution of the cyclone as a whole and would not then be expected to influence the rate of intensification.

The adjustment of the atmosphere to a state of gradient balance can be described in terms of radiation of gravity waves as in the classical geostrophic adjustment problem (Schubert et al. 1980). Due to the rapid spread of the gravity wave response, the wind and pressure fields will be close to gradient balance at all times during the growth of the cyclone (Willoughby 1979), therefore it is not anticipated in either the CISK or WISHE theories that the gradient adjustment process will affect the rate of intensification. Indeed, models based on gradient balance have been used for both mechanisms (Charney and Eliassen 1964; Ooyama 1964; Emanuel 1989).

The rapidity of gradient adjustment also implies the equivalence of cumulus parameterizations that represent the effects of clouds as a heating or mass flux, at least for tropical cyclone intensification. If convection is assumed to heat the atmosphere, the adjustment to gradient balance will consist of rising motion that cools the atmosphere adiabatically, while simultaneously spinning up a vortex through the induced low-level convergence. The process stops when the cooler temperature anomaly and enhanced winds are in balance. If convection is assumed to transfer mass from the lower to the upper troposphere, the associated low-level convergence will spin up an unbalanced vortex. The adjustment to gradient balance will then consist of a subsidence compensating the convective mass flux that warms the core of the vortex adiabatically and spins down the wind until balance is achieved. Clearly, for any heat source it is possible to define a mass flux that will result in the same balanced vortex and vice versa.

Since gradient adjustment can be assumed to be fast, the distinction between CISK and WISHE comes in which of the remaining two processes influence the timescale for cyclone growth. A key assumption of the WISHE mechanism is that the convective mixing of boundary layer thermal energy through the troposphere is rapid. An anomalously warm moist region of the boundary layer will produce a conditionally unstable sounding (positive CAPE), which is quickly mixed out by moist convection. The eddy turnover time for deep convection is typically described as of the order of a half hour and, thus, is fast in comparison to the growth rate of the large-scale cyclone. In this case the atmosphere can again be characterized by an adjusted state, here a convectively neutral lapse rate, or zero CAPE. In the simple WISHE model of Emanuel (1989) the temperature is assumed to be equal to that of a boundary layer parcel that has undergone moist adiabatic ascent.

It is the rate at which latent and sensible heat is extracted from the sea surface that controls the rate of intensification of the cyclone in WISHE. A model that attempts to represent this mechanism must simulate or parameterize these fluxes in terms of the large-scale flow. This is most easily done using a bulk aerodynamic formula. For example, the moisture flux F_q, can be expressed as $F_q = C_E U(q_s - q)$, where U is the low-level (e.g., 10 m) wind speed, q the boundary layer specific humidity, q_s the surface value of the specific humidity, and C_E an empirical transfer coefficient. Indeed the scale analysis of Emanuel (1989, 1995) suggests that the timescale for vortex growth is inversely proportional to the transfer coefficient for heat and moisture fluxes (this is not immediately obvious since they are set equal to the momentum drag coefficient in those papers). Although the rate limiting process is the surface fluxes, the nonlinearity of the bulk aerodynamic formula implies that there is a possibility of an indirect influence of the convection on the fluxes through modification of the boundary layer temperature and moisture content. Emanuel (1989) has suggested that such an influence, associated with drying of the boundary layer by convective downdrafts, may prevent weak disturbances from intensifying.

In CISK, the transport of energy by moist convection is not assumed to be fast. Cumulus heating or mass flux, at least in an ensemble average sense, is controlled by the large-scale flow and, hence, evolves on that timescale. The heating rate or equivalent mass flux, which will be denoted Q, is calculated from a cumulus parameterization and, in particular, is often taken to be proportional to low-level convergence $Q = \eta w_B$, where w_B is the vertical velocity at the top of the boundary layer. The difference between the theories of Ooyama and Charney and Eliassen is in the definition of the coefficient η. Ooyama (1969) uses an energy balance argument to obtain η as a function of the conditional instability of the atmosphere (see appendix). Charney and Eliassen[3] assume it to be proportional to the boundary layer specific humidity q.

In a linear CISK model, the coefficient η is assumed to be constant, while the convergence is a perturbation quantity. This is equivalent to assuming that surface fluxes act rapidly to replenish any depletion of CAPE or boundary layer moisture by the convection, maintaining an equilibrium state characterized by constant η. In this case, the growth rate is limited by convective processes and is not restricted by the magnitude of the surface heat and moisture fluxes. Indeed, the growth rate in linear CISK models of tropical cyclones and polar lows is proportional to the Ekman layer drag coefficient (Charney and Eliassen 1964; Ooyama 1964; Bratseth 1985; McBride and Fraedrich 1995; Fraedrich and McBride 1989, 1995). In nonlinear CISK models (Ooyama 1969; Handel 1991), the rate-limiting step is still the convection, but the rate may be influenced indirectly by surface fluxes of heat and moisture, which may contribute to a time evolution of the coefficient η.

In essence, both CISK and WISHE describe a feedback in which the strength of the balanced vortex controls the energy input into the system (Fig. 1). The rate-limiting process in WISHE is the extraction of latent and sensible heat from the ocean surface. Once in the boundary layer, heat is rapidly mixed through the depth of the troposphere, which rapidly adjusts to maintain thermal wind balance with the warm core. Since the surface fluxes are dependent upon wind speed, a positive feedback occurs, resulting in instability. In contrast, for CISK, tropospheric heating is controlled by low-level convergence. In linear CISK, surface fluxes are assumed to act rapidly to maintain CAPE or boundary layer moisture content, so the rate of intensification is limited by the rate of frictional mass convergence in the boundary layer. In nonlinear CISK, surface fluxes of heat and moisture may influence the rate of intensification in combination with frictional convergence.

c. *A thought experiment*

An important conclusion from the preceding discussion is that for both WISHE and the CISK theories of Ooyama and Charney and Eliassen, the rate of growth is controlled by boundary layer processes: heat and moisture fluxes for WISHE, and frictional convergence for CISK. This suggests that the sensitivity of the growth rate of a tropical cyclone to surface properties, such as roughness and availability of moisture, will be different for WISHE and CISK, and could serve as the basis for an observable distinction. In nature, tropical

[3] Charney and Eliassen also include a term to represent moisture convergence above the boundary layer, but in their linear model it is proportional to the boundary layer term.

cyclones form only over the tropical oceans, where the range of surface properties other than temperature is rather small; however, one can easily devise a thought experiment where the theories will predict different results.

Such an experiment would involve triggering tropical cyclone-type disturbances over a variety of different surfaces to determine the effect on their intensification. The crucial element for distinguishing between the two theories is that the surface friction be varied independently of the heat and moisture fluxes. Although these processes are related, since the same eddies transport both momentum and heat (latent and sensible) through much of the boundary layer, they will not respond in an identical manner. Consider, for example, the virtually frictionless surface of an ice sheet covered by a thin melting layer, or since this is a thought experiment, replace the ice sheet with a temperature controlled metal plate with pores where water can be introduced to keep the surface wet. Heat and moisture will be transported away from the surface by buoyancy driven eddies, producing the usual well-mixed subcloud layer. The boundary layer eddies will mix momentum vertically, but there will be very little frictional loss at the surface, and even large horizontal wind speeds at low levels will result in very little frictionally induced convergence. In comparison with a cyclone forming over the ocean, CISK would predict that a disturbance triggered in such a low-friction environment would grow less rapidly due to a reduction in the boundary-layer convergence, which drives the cumulus convection. Note, however, that this effect might not be observed if CAPE or boundary layer moisture content were increased in the modified cyclone. WISHE, on the other hand, would predict that, since the heat and moisture fluxes are unaffected, the growth rate should also be largely unaffected.

The surface friction could then be varied by introducing roughness elements. The nonlinear dependence of drag on wind speed due to increasing wave height would be more difficult to arrange, but is not crucial for distinguishing between CISK and WISHE. An independent set of experiments could then be carried out by varying the permeability of the surface to heat and moisture. This would alter the fluxes of latent and sensible heat without corresponding changes in the friction. Such variations occur naturally as the surface changes from desert to partially vegetated marshland to open sea. A decrease in surface heat and moisture fluxes would reduce the growth rate according to WISHE, while CISK would predict little change, unless the reduction in fluxes was great enough to significantly reduce the CAPE or boundary layer moisture content. While the full range of situations envisaged in this thought experiment is unlikely to arise spontaneously in nature, and would be impractical to implement in a laboratory or field experiment, it is nonetheless consistent with the laws of physics and can easily be simulated in a numerical model.

In this work we examine the dependence of tropical cyclone and polar low intensification on the surface properties of moisture and roughness using a numerical model and relate the results to the predictions of the CISK and WISHE intensification mechanisms. It will be of interest to consider both the intensification of tropical cyclones and of those polar low systems that have similar structures to tropical cyclones. These two types of disturbance develop under very different atmospheric conditions and yet both have been hypothesized to develop via the same mechanism (be it CISK or WISHE).

2. Experimental design

a. Numerical model

In selecting a numerical model to compare the instability theories, the primary concern is that the model is able to represent either of the proposed mechanisms. A particular problem is the representation of cumulus convection. It is well known that models using parameterizations based on the work of Kuo (1965, 1974) will produce CISK instabilities (e.g., Rosenthal 1971). This is unacceptable if these parameterization schemes are incorrect, which is precisely what has been argued by critics of CISK (Emanuel et al. 1994). Similarly Emanuel (1989) has introduced a representation of convection that produces WISHE instabilities. To avoid the possibility of prejudicing the results through the choice of a particular parameterization scheme, a numerical model was used that explicitly simulates convection. The computational expense of such simulations made the use of a two-dimensional axisymmetric model preferable to the vastly more expensive three-dimensional alternative. The axisymmetric approximation should be acceptable since neither mechanism depends explicitly on the presence of vortex asymmetries and since such models have often been used successfully in the simulation of tropical cyclones (e.g., Willoughby et al. 1982; Baik et al. 1990).

The experiments were performed using modified versions of the nonhydrostatic, axisymmetric tropical cyclone model originally written by Rotunno and Emanuel (1987) and used to simulate polar low development by Emanuel and Rotunno (1989). The version used here includes the modifications described by Craig (1995, 1996). In brief, the nonhydrostatic, compressible equations of motion are integrated, with prognostic equations for momentum, potential temperature, nondimensional pressure, and mass mixing ratios of water vapor and ice. Cloud water and rain are also included in the tropical cyclone simulations. The turbulence parameterization is based on first-order Richardson-number-dependent eddy viscosity. The convection is treated explicitly rather than being parameterized.

Model physics such as radiation and microphysics can affect the growth rate of vortices (Craig 1995), but are not likely to affect their response to changes in surface boundary conditions. Hence for computational efficiency the effects of radiation were neglected here.

An extensive series of sensitivity tests was performed to determine polar low and tropical cyclone model configurations of which the model results would be independent. Following Craig (1995, 1996) a closed domain was employed with no-flux boundary conditions at both the outer and upper boundaries beyond which were placed "sponge" layers for the absorption of horizontally and vertically propagating gravity waves. After considerable experimentation with both this and radiation boundary conditions, it was concluded that the only way to ensure that the intensification rate was independent of the domain size was to use a very large domain (Table 1).

There is some evidence that the growth rate of a simulated cyclone will have converged for a horizontal grid size of a few kilometers (Tripoli 1992). Tests with grid sizes as small as 1 km did not reach convergence, but showed only a weak tendency for increasing growth rates with increasing resolution. Furthermore the sensitivity of the growth rate to changes in boundary layer parameters was similar for any grid size smaller than about 10 km. A grid size of 5 km was therefore considered acceptable for both tropical cyclone and polar low simulations. The model was found to be similarly insensitive to vertical resolution and the values chosen are shown in Table 1.

The simulations were initialized with either the mean hurricane season sounding of Jordan (1958), or a modified version of the 12 GMT 13 December 1982 sounding from Bear Island in the Norwegian Sea as used by Emanuel and Rotunno (1989). The model simulations were initialized with low-level vortices of the type described by Emanuel and Rotunno (1989). The initial tropical cyclone vortex was given a maximum azimuthal wind speed of 15 m s^{-1} at 75-km radius and the initial polar low vortex had a maximum wind speed of 10 m s^{-1} at 50-km radius. The initial conditions are described in greater detail by Craig (1995, 1996).

The most important aspect of the model for the experiments to be described here is the representation of the planetary boundary layer. The behavior of the boundary layer at high wind speeds is still a matter of great uncertainty due primarily to a lack of appropriate observations (Smith 1988). The approach taken here is to use a simple method and look for results that are robust to changes in the parameterization. Surface fluxes of heat, moisture, and momentum are modeled using bulk aerodynamic formulae and are given by

$$F_\theta = C_T U(\theta_s - \theta), \quad (1)$$

$$F_q = C_E U(q_s - q), \quad (2)$$

$$F_m = \rho C_D U^2, \quad (3)$$

TABLE 1. Model configuration for polar low and tropical cyclone simulations.

Model parameter values (km)	Tropical cyclone	Polar low
Horizontal domain size	3150	2200
Vertical domain size	22.5	14
Plus outer sponge layer	450	300
Plus upper sponge layer	5	4
Horizontal resolution	5	5
Vertical resolution	1.25	1

respectively, where ρ is density and U is the wind speed at the lowest model level; $(\theta_s - \theta)$ is the potential temperature difference between the sea surface and the lowest model level and $(q_s - q)$ is the corresponding difference for the water vapor mixing ratio. The momentum drag coefficient is given by Deacon's formula:

$$C_D = 1.1 \times 10^{-3} + 4 \times 10^{-5} U, \quad (4)$$

(Rotunno and Emanuel 1987). Following Smith (1988), C_T and C_E are assumed to be independent of wind speed, with C_E slightly larger; that is,

$$C_E = 1.2 \times C_T, \quad (5)$$

with $C_T = 1.00 \times 10^{-3}$ and $C_T = 1.06 \times 10^{-3}$ for the tropical cyclone and polar low simulations, respectively. A slightly larger transfer coefficient is appropriate for the vigorously convective boundary layer produced by the large air–sea temperature differences found where polar lows form. Emanuel and Rotunno (1987) set all three transfer coefficients equal to Deacon's formula (4), including the wind speed dependence. Experiments performed with this formulation resulted in faster growth rates and more intense cyclones, but showed the same sensitivity to the variations in parameters described in the next section.

The structure of the simulated cyclones will not be described in detail, as it is similar to that found in previous studies using similar axisymmetric, cloud-resolving models. In particular, the interested reader can refer to Willoughby et al. (1984) and Lord et al. (1984) for the tropical cyclone and Craig (1995) for the polar low structure. It is important to note that the tropical cyclone simulations include ice-phase microphysics, which has been shown by Willoughby et al. (1984) to produce convective structures and eyewall replacement events that are largely absent in simulations using only warm rain (e.g., Rotunno and Emanuel 1987).

A notable difference between the simulated polar lows and tropical cyclones is the azimuthal wind profile in the eye, shown in Fig. 2. The tropical cyclone eye is in solid body rotation, while the polar low eye is larger and almost at rest from the axis out to a region of strong shear at the eyewall. This difference can also be seen in the earlier studies cited above, although it has not been commented on in detail. The difference

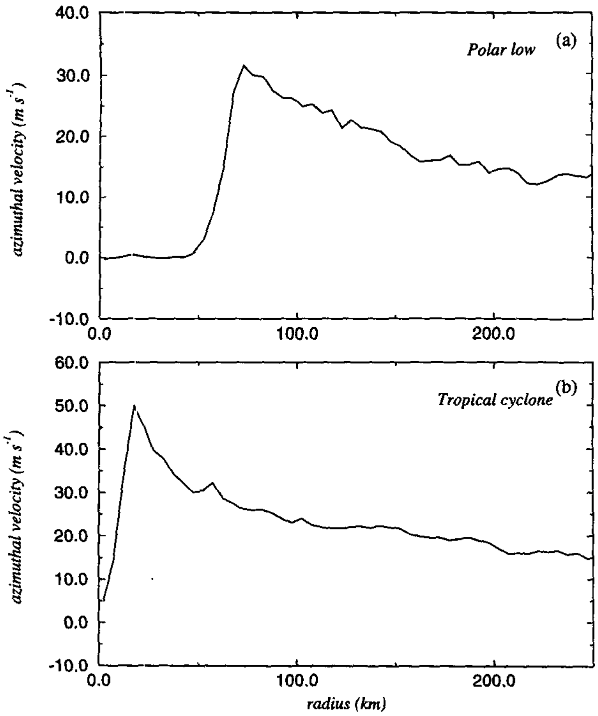

FIG. 2. Representative "snapshots" of boundary layer azimuthal velocity as a function of radius for (a) the polar simulation at 90 h and (b) the tropical cyclone simulation at 120 h.

in wind profile does not appear to be important for the present study since neither the polar low nor the tropical cyclone structure changes throughout the experiments, with one exception to be discussed in the next section.

b. Experimental design

The aim of the experiments described in the introduction is to simulate the spinup of a tropical cyclone or polar low over a variety of surfaces. In the bulk aerodynamic formulae, changes in surface roughness and availability of heat and moisture are represented by different values of the transfer coefficients. The experiments will, thus, consist of changing these values. The first set of experiments will involve varying C_T and C_E simultaneously, while the second involves variations in C_D. In each case the parameters will be scaled by a constant value everywhere in the domain, but in small increments to ensure that the resulting changes in growth rate are due to a simple parameter dependence, rather than a transition to a different regime of behavior where other physical processes dominate. One case will be shown, however, where the perturbation was sufficient to change the structure of the simulated cyclone. When the tropical cyclone experiment was performed without friction ($C_D = 0$), the wind profile in the eye was similar to that found in the polar low simulations. The structural change does not appear to produce any anomalous change in the rate of intensification; however, it must be noted that this simulation may not be directly comparable to the others.

The rate of intensification of the cyclones is described using plots of minimum central surface pressure as a function of time. Plots of maximum azimuthal wind showed similar results but with greater noise levels, and will not be reproduced. Conclusions will be based on the period of the simulations in which the cyclone is intensifying. Some problems have been noted with the simulations after maximum intensity has been reached (Craig 1995), so this phase will also be ignored.

The predicted results of the experiments by the intensification theories are summarized in Table 2. It is expected that boundary layer convergence will be proportional to the drag coefficient C_D (diagnostic calculations, not shown, verify that this is the case). CISK, therefore, predicts that growth rate should increase with increasing C_D, but should not be sensitive to changes in C_T and C_E, provided these changes are not accompanied by major changes in the boundary layer moisture content (Charney and Eliassen CISK) or in the instability of the atmosphere to convection (Ooyama CISK). In contrast, WISHE predicts that growth rate should increase with increasing C_T and C_E provided that the thermodynamic disequilibrium between ocean and atmosphere, $\Delta\theta_e$, is not significantly decreased. The frictional drag coefficient C_D does not directly affect the rate-limiting process in this theory and the effect of its variation should be comparatively small unless $\Delta\theta_e$ is affected by changes in C_D. Provided that such changes do not occur, the results of the numerical simulations will be consistent with the predicted results of only one of CISK or WISHE (Table 2).

The monitoring of boundary layer moisture content, conditional instability, and thermodynamic disequilibrium between ocean and atmosphere is complicated by nonlinearity, which implies that these variables may have a significant time dependence during the development of the cyclone. This is predicted, for example, by the nonlinear CISK model of Ooyama (1969),

TABLE 2. Change in rate of intensification predicted for both CISK theories and the WISHE theory in each experiment.

	Increase C_T, C_E	Increase C_D
CISK		
Charney and Eliassen $Q \sim w_B q$	no change (unless q changes)	faster growth (unless q decreases)
Ooyama $Q \sim w_B \eta$ (CAPE)	no change (unless CAPE changes)	faster growth (unless CAPE decreases)
WISHE $Q \sim C_{E,T} U \Delta\theta_e$	faster growth (unless $\Delta\theta_e$ decreases)	no change (unless $\Delta\theta_e$ changes)

which shows a decay with time in the convective instability of the atmosphere. To detect systematic differences in these variables between cyclones growing at different rates, it is necessary to compare their values at similar points in the development. This will be done by plotting each as a function of central pressure, so that the values in different runs can be compared at a given vortex intensity.

In detail, the following quantities are diagnosed: the specific humidity in the first model level (boundary layer), CAPE (calculated as explained in the appendix), and the difference in equivalent potential temperature between the sea surface and the first model level. These are calculated from an area-mean atmospheric sounding for the region of the eyewall. The tropical cyclone eyewall is defined as covering the radial width from 20–300 km outward from the central axis. The radius of the polar low eye was found to vary somewhat as the transfer coefficients were varied, and thus, the eyewall region in the polar low simulations was defined as a region of 100-km width, extending outward from the sharp gradient in azimuthal wind speed that marks the inner edge of the eyewall (Fig. 2b).

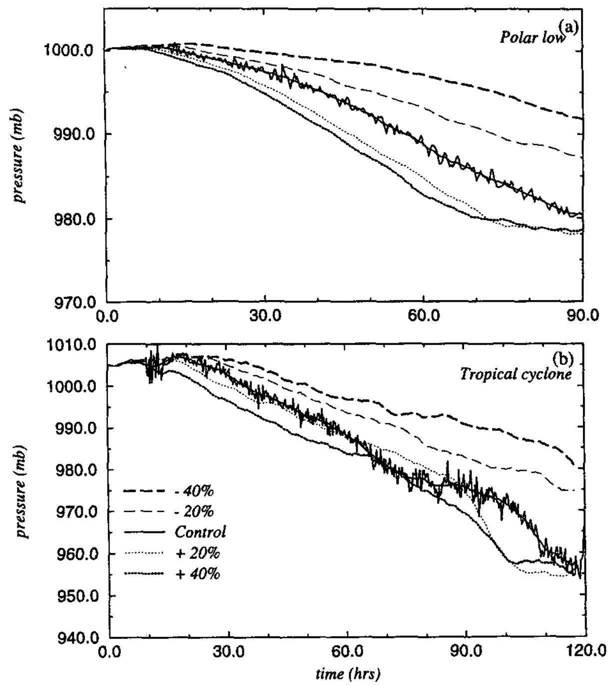

FIG. 3. Central surface pressure as a function of time for (a) polar low and (b) tropical cyclone model runs with varying heat and moisture transfer coefficients. For clarity, running mean data is plotted in all instances with 15-min, high temporal resolution data also shown for the control runs.

3. Results

a. Dependence on heat and moisture transfer coefficients

Figure 3a shows the time dependence of the central surface pressure in the polar low simulations for several values of the heat and moisture transfer coefficients. Over much of the range of C_T and C_E values, the rate of intensification (measured at a given pressure depression) increases approximately in proportion with C_T and C_E. An attempt was made to express this relation quantitatively, but the results were somewhat arbitrary owing to the difficulty in fitting the growth rate curves to a simple formula such as a linear or exponential function. It can also be seen that the proportionality breaks down when the transfer coefficients are increased by 40%.

The tropical cyclone experiments showed similar results to the polar low runs (Fig. 3b), with an increasing rate of intensification as C_T and C_E were increased. The pattern of increase is less clear, however, due to fluctuations in pressure on a 30–40 hour timescale that result from eyewall replacement events (Willoughby et al. 1982). In particular, such an event begins at approximately 60 h into the two runs with the largest values of C_T and C_E. After the initiation of eyewall replacement events (at pressures below about 980 mb), it is difficult to compare intensification rates and, hence, in discussing the intensification period of the tropical cyclone, comparisons will be restricted to the times before these events begin. As in the polar low model, the pattern of increasing intensification rate with heat and moisture transfer coefficients appears to break down for high values of these coefficients. The general pattern of faster growth with increasing heat and moisture transfer coefficients is nonetheless apparent for both the tropical cyclone and polar low experiments.

A comparison with the predictions in Table 2 suggests that this result is consistent with WISHE, but can only be reconciled with CISK if the enhanced transfer coefficients for heat and moisture fluxes lead to higher boundary layer relative humidities, or greater conditional instability. The variation of boundary layer specific humidity q, with central surface pressure, for both the tropical cyclone and polar low simulations is shown in Fig. 4 for the control and the runs in which the heat and moisture transfer coefficients have been varied by $\pm 40\%$. In all cases q increases substantially as the central pressure falls. However, during the period of interest in both the polar low (until $\sim$985 mb) and tropical cyclone (until $\sim$980 mb) vortices, the boundary layer specific humidity at a given central pressure is approximately independent of the value of the heat and moisture transfer coefficients. Large variations occur later in the development of the systems. This is due in part to the changes in intensity due to eyewall replacement events, but also due to the associated changes in the location of the eyewall which are not taken into account in the diagnostic calculations. A systematic increase in

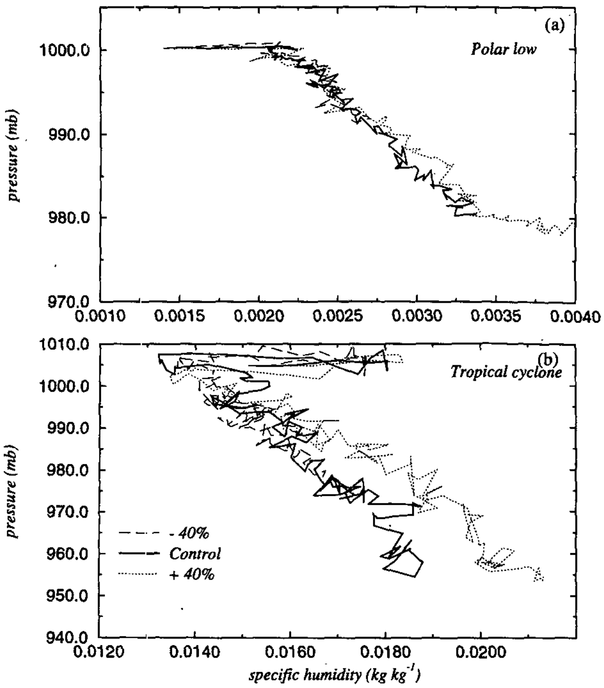

FIG. 4. Central surface pressure plotted against eyewall area-mean boundary layer specific humidity for (a) 90 h of polar low data and (b) 120 h of tropical cyclone data (at 15-min intervals). Control runs and runs with ±40% of the control heat and moisture transfer coefficients are shown.

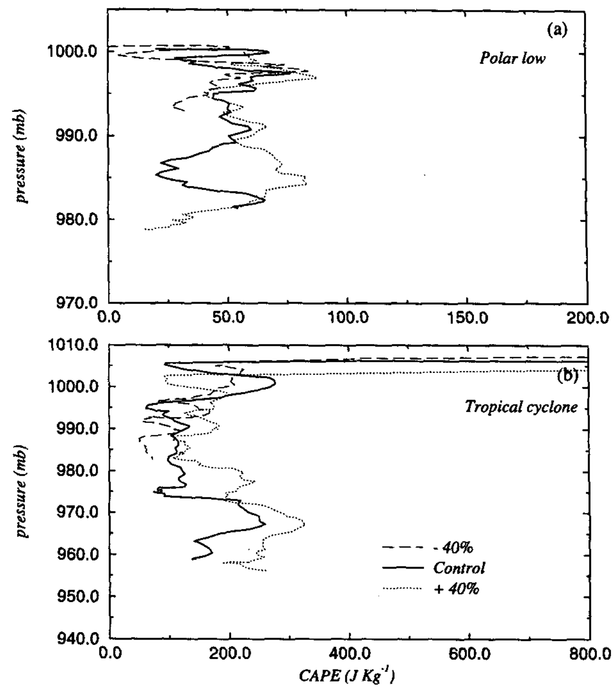

FIG. 5. Central surface pressure plotted against eyewall area-mean CAPE for (a) 90 h of polar low data and (b) 120 h of tropical cyclone data. Control runs and runs with ±40% of the control heat and moisture transfer coefficients are shown. For clarity, the data have been smoothed using a running mean.

boundary layer specific humidity with C_T and C_E, required by the CISK theory of Charney and Eliassen to explain the observed increase in intensification rate, is not found.

Figure 5 shows the evolution of CAPE in the same experiments. These data are extremely noisy (varying by an order of magnitude) and are therefore smoothed using a running mean. The unsmoothed data for the control run are shown in the appendix. Prior to the large variations induced by eyewall replacements, CAPE appears to be approximately independent of both time and the transfer coefficients C_T and C_E. The systematic trend required for consistency with the CISK theory of Ooyama (1964, 1969) is not visible.

The final plot in this series, Fig. 6, shows the variation of the ocean–atmosphere thermodynamic disequilibrium $\Delta\theta_e$. WISHE predicts the observed increase in intensification rate with increasing C_T and C_E provided $\Delta\theta_e$ is not significantly decreased. Figure 6 shows $\Delta\theta_e$ to be approximately independent of C_T and C_E during the period of interest.

b. Dependence on friction drag coefficient

The dependence of the intensification rate on the drag coefficient C_D is shown in Fig. 7. Variations in C_D of the same magnitude as those applied to C_T and C_E in the previous section (±40%) have almost no effect

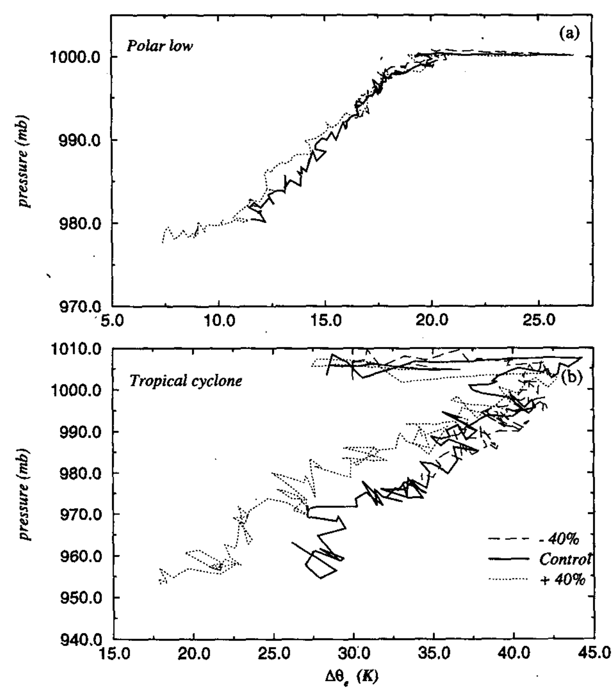

FIG. 6. Same as in Fig. 4 except central surface pressure plotted against eyewall area-mean ocean–atmosphere thermodynamic disequilibrium.

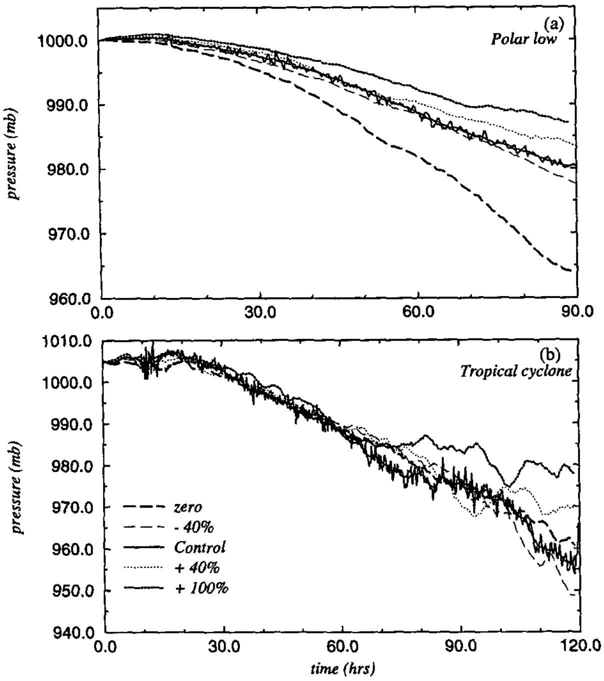

FIG. 7. Central surface pressure as a function of time for (a) polar low and (b) tropical cyclone model runs with varying frictional drag coefficient. For clarity, running mean data are plotted in all instances with 15-min, high temporal resolution data also shown for the control runs.

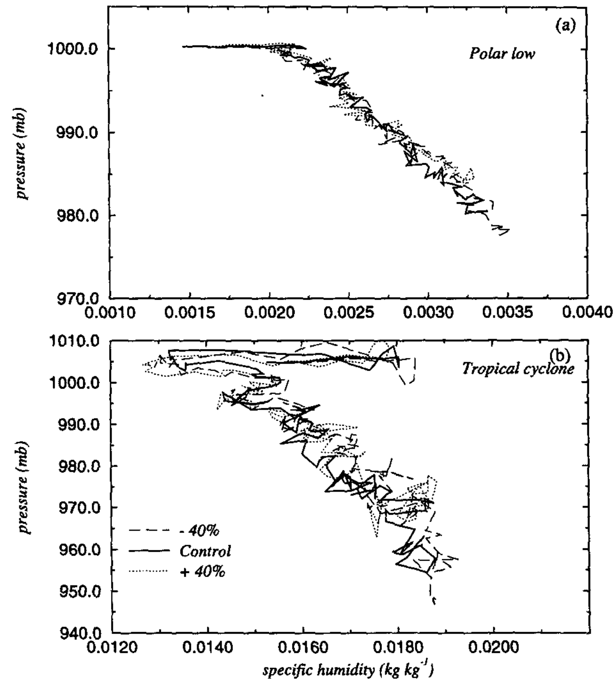

FIG. 8. Same as Fig. 4 except control runs and runs with ±40% of the control frictional drag coefficient are shown.

on either the polar low or tropical cyclone simulations. Imposing a larger variation shows that the growth rate of the polar low clearly decreases if C_D is doubled, although not by a factor of 2 (Fig. 7a). Perhaps the most striking result in Fig. 7a is that fastest growth is obtained with no surface friction at all. The intensification rate of the simulated tropical cyclone does not appear to respond even to these larger changes in C_D (Fig. 7b). However the tropical cyclone simulation with zero C_D may not be directly comparable due to the structural change described in section 2b.

The sensitivities to C_D of the boundary layer specific humidity, CAPE, and thermodynamic disequilibrium $\Delta\theta_e$, are monitored in Figs. 8–10. The results are the same as those found when C_T and C_E are varied. The specific humidity q increased and $\Delta\theta_e$ decreased systematically as the cyclone intensified, but at any given vortex intensity showed no dependence on C_D. CAPE remained approximately uniform both in time and with variations in C_D. Again, comparing the results of these experiments with the predictions in Table 2, shows that they are expected from WISHE, but are inconsistent with CISK since an increase in drag coefficient would be expected to produce an increase in growth rate unless a compensating decrease in q or CAPE is observed.

The results for both sets of experiments were found to be quite robust in a number of sensitivity tests. As

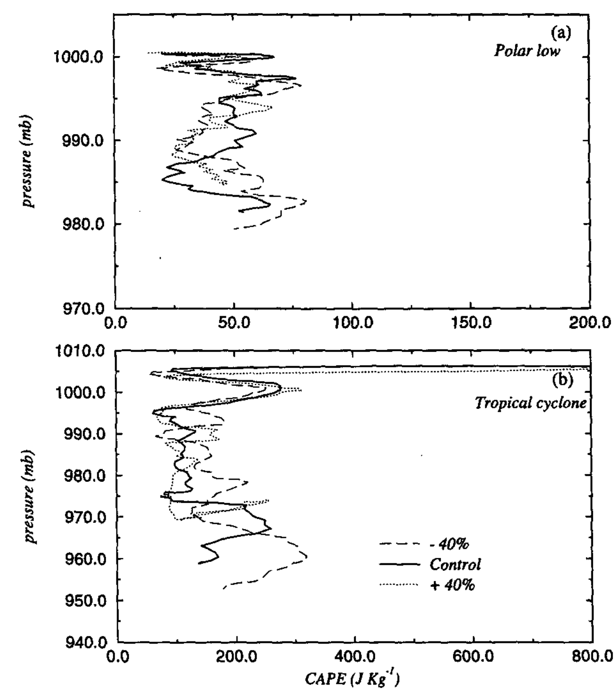

FIG. 9. Central surface pressure plotted against eyewall area-mean CAPE for (a) 90 h of polar low data and (b) 120 h of tropical cyclone data. Control runs and runs with ±40% of the control frictional drag coefficient are shown. For clarity, the data have been smoothed using a running mean.

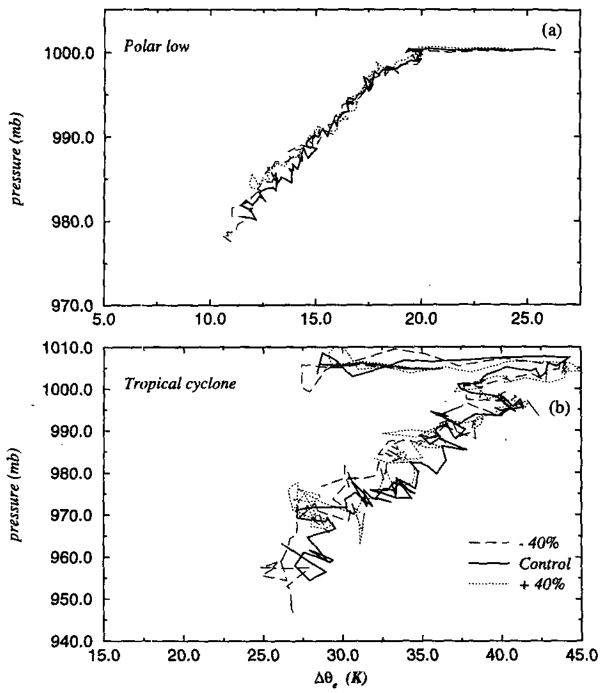

FIG. 10. Same as Fig. 6 except control runs and runs with ±40% of the control frictional drag coefficient are shown.

mentioned in section 2, the experiments were repeated with different horizontal resolutions and a different formulation for the heat and moisture transfer coefficients. In addition, experiments were carried out where the changes in transfer coefficients were introduced after 30 h of simulation to avoid changing the initiation of convection. In these sensitivity tests, there was some variation in the degree to which the growth rate changed in response to changes in C_T and C_E, and in some cases the slight tendency for a negative dependence of growth rate on C_D was enhanced, but the description of the results shown here remains valid.

4. Conclusions

Numerical experiments with a cloud-resolving, axisymmetric model show that the rate of intensification increases with increasing values of the transfer coefficients for heat and moisture, corresponding to increasing availability of heat and moisture at the sea surface. The model is relatively insensitive to changes in the momentum transfer coefficient (increasing surface friction), although the polar low version of the model exhibits a weak decrease in intensification rate with increasing frictional drag coefficient. This implies that the rate-limiting process for cyclone intensification is the rate of heat and moisture fluxes. Frictional convergence is of secondary importance but may represent a sink of energy that decreases the growth rate.

Diagnostic calculations showed that boundary layer specific humidity increases as the vortex intensifies, while the difference in equivalent potential temperature between atmosphere and sea surface decreases. Both show a one-to-one relationship with vortex intensity that is independent of the value of the transfer coefficients. The conditional instability (CAPE) shows no systematic change with cyclone intensity or transfer coefficients, although considerable variability is found on short space and timescales.

These results are predicted for a wind-induced surface heat exchange (WISHE) instability, and are contrary to the predictions of conditional instability of the second kind (CISK), in the forms proposed by either Charney and Eliassen (1964) or Ooyama (1964). It is concluded that the intensification of the numerically simulated tropical cyclones and polar lows is due to WISHE. It is anticipated that a similar conclusion would apply in nature.

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APPENDIX

Convective Available Potential Energy (CAPE)

This appendix describes the method of calculation of the CAPE values presented in section 3, and also compares CAPE with calculations of slantwise CAPE (SCAPE), and values of the parameter η defined by Ooyama (1969). CAPE and boundary layer θ_e were calculated using saturated vapor pressures calculated with respect to water for the tropical cyclone data and with respect to ice for the polar low data (for consistency with calculations within relevant the versions of the numerical model). CAPE was calculated for air parcels lifted from the lowest model level (0.625 km for the tropical cyclone simulations, 0.5 km for the polar low) and was defined by

$$\text{CAPE} = g \int_{\text{LCL}}^{\text{LNB}} \left[\frac{T_p - T_e}{T_e} \right] dz, \qquad \text{(A1)}$$

where T is temperature with the subscripts p and e referring to the parcel and environmental values, respectively, z is vertical height, g is acceleration due to gravity, LCL is the lifting condensation level, and LNB is the level of neutral buoyancy. SCAPE was defined by the same integral but calculated instead along momentum surfaces, and so in the initial stages of the simulations where the momentum surfaces are almost vertical, SCAPE is approximately equal to CAPE. The absolute angular momentum field at any given time was rather noisy and some smoothing was required before SCAPE could be calculated. This was particularly true for the tropical cyclone model where the smaller Coriolis parameter f reduced the dominance of the quadratic radial term ($fr^2/2$ where r is radius). A smoothing matrix was applied interatively to the raw data, with the number of iterations increasing linearly with model level as the strong eyewall velocities decreased and the momentum became more variable. This produced a smoothed momentum field to which the radius of absolute angular momentum contours could be traced uniquely with height from their surface values. SCAPE could then be calculated as the CAPE available for parcels lifted along these contours.

Fig. A1 shows the evolution of eyewall-averaged CAPE and SCAPE for the tropical cyclone simulation with control values of the transfer coefficients. The eyewall region is considered to extend from 20–300 km, and data are plotted at two hourly intervals. SCAPE values can not be obtained from a mean atmospheric sounding, so for consistency both the SCAPE and CAPE values in Fig. A1 are calculated for each (slantwise) column in the region and then averaged. The CAPE values can be compared to those in Fig. A1b, calculated from a mean sounding (as in Figs. 5 and 9, but without time smoothing).

After an initial surge associated with model initialization that extends until approximately 20 h, the mean CAPE and SCAPE values vary about a value of approximately 300 J Kg^{-1}. These calculations include very large values associated with individual convective updrafts that increase the average substantially. The values calculated from a mean sounding (Fig. A1b) show a very similar pattern but with about one-third the magnitude. Aside from the initialization phase, there is no systematic trend in any of the quantities as the cyclone grows and saturates at maximum intensity, as might be expected if an initial reservoir of CAPE were consumed by the cyclone growth as predicted by Ooyama (1969). The evolution of CAPE and SCAPE calculated from the polar low model is similar to that of the tropical cyclone but with a much smaller mean CAPE value of order 10 J Kg^{-1}.

The two-layer model of Ooyama (1969) did not explicitly invoke CAPE, but rather set the convective mass flux equal to ηw_B, where w_B is the vertical velocity at the top of the Ekman layer and η is defined as

$$\eta = 1 + \frac{\theta_{e0} - \theta_{e2}^*}{\theta_{e2}^* - \theta_{e1}},$$

where θ_e is equivalent potential temperature, θ_e^* saturation equivalent potential temperature, and the subscript 0 denotes boundary layer values, while subscripts 1 and 2 denote the lower and upper layers of the model, respectively. Clearly, η is a measure of the conditional instability of the atmosphere, which is equal to one if CAPE is zero and takes larger values if boundary layer air is buoyant in the upper troposphere.

As shown in Fig. A1b, the evolution of η is very similar to that of CAPE on all timescales. The η values in the figure were calculated by dividing the troposphere into two layers: surface to 600 mb and 600–200 mb. Other partitions produced similar results, although the absolute value of η is sensitive to the choice of layer boundaries. It is interesting that the η values in Fig. A1b are almost always less than one, indicating stability in Ooyama's model. This is due to outward spreading of θ_e lines in the upper troposphere, which suggests that a slantwise calculation of η might be more appropriate. It is anticipated from the comparison of CAPE and SCAPE in Fig. A1a that this would produce a modest increase in magnitude, but leave the variability largely unchanged.

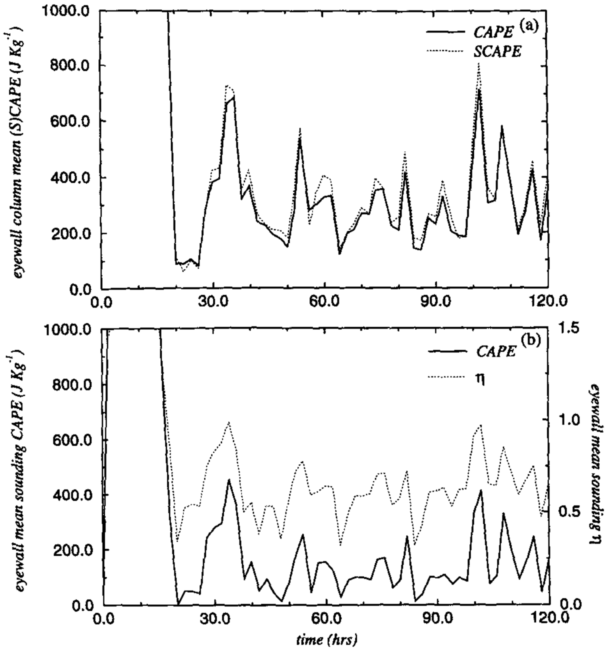

FIG. A1. Mean eyewall (a) CAPE and SCAPE, and (b) Ooyama's η parameter (see text) plotted as a function of time for the control run of the tropical cyclone model. Plotted values are (a) area means of (slantwise) column values and for (b) calculated from area-mean soundings, the area being the ring between 20- and 300-km radius.

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