The Effect of Surface Friction on the Development of Tropical Cyclones

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ABSTRACT

When tropical cyclones (hereafter referred as TCs) are over the ocean, surface friction plays a dual role in the development of TCs. From the viewpoint of water vapor supply, frictional convergence and Ekman pumping provide a source of moisture for organized cumulus convection and is propitious to the spin-up of TCs. On the other hand, surface friction leads to a dissipation of kinetic energy that impedes the intensification of TCs. Which role is dominant in the developing stage of TCs is a controversial issue.

In the present work, the influence of surface friction on the growth of TCs is re-examined in detail by conducting two sets of numerical experiments initialized with different cyclonic disturbances. Results indicate that, because of the inherent complexities of TCs, the impact of surface friction on the evolution of TCs can not be simply boiled down to being positive or negative. In the case that a TC starts from a low-level vortex with a warm core, surface friction and the resultant vertical motion makes an important contribution to the convection in the early developing stage of the TC by accelerating the build-up of convective available potential energy (CAPE) and ensuring moisture supply and the lifting of air parcels. This effect is so prominent that it dominates the friction-induced dissipation and makes surface friction a facilitative factor in the spin-up of the TC. However, for a TC formed from a mesoscale convective vortex (MCV) spawned in a long-lasting mesoscale convective system (MCS), the initial fields, and especially the low-level humidity and cold core, enable the prerequisites of convection (i.e., conditional instability, moisture, and lifting), to be easily achieved even without the help of boundary-layer pumping induced by surface friction. Accordingly, the reliance of the development of TCs on surface friction is not as heavy as that derived from a low-level vortex. The positive effect of surface friction on the development of TCs realized through facilitating favorable conditions for convection is nearly cancelled out by the friction-induced dissipation. However, as SST is enhanced in the latter case, the situation may be changed, and different development speeds may emerge between model TCs with and without surface friction considered. In short, owing to the fact that TC development is a complicated process affected by many factors such as initial perturbations, SST, etc., the importance of surface friction to the intensification of TCs may vary enormously from case to case.

Key words: surface friction, tropical cyclone, boundary-layer pumping, convective available potential energy

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1. Introduction

Surface friction plays a fundamental role in the development of tropical cyclones (hereafter referred as TCs). Although the intrinsic dissipation associated with surface friction is deemed to inhibit TCs from developing, the resultant low-level convergence and boundary-layer pumping export moisture and heat to the free atmosphere to impel the further development of the cyclone (Rosenthal, 1971; Zhang et al., 1994). Which role is more influential on the intensification of TCs is in debate. Substantial literature has been dedicated to this subject.

In the linear theory of conditional instability of the second kind (CISK) introduced by Ooyama (1964, 1969) and Charney and Eliassen (1964), the growth rate of a TC is found to be directly proportional to the surface friction and to the boundary layer humid-

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ity (Ooyama, 1969; Rosenthal and Koss, 1968). This result is entirely logical because the CISK mechanism takes the frictional convergence in the Ekman layer as the primary source of water vapor for organized cumulus convection. With a series of numerical experiments in which the surface friction was artificially changed, Rosenthal (1971) also found that the stronger surface friction is, the faster a TC develops in the earlier stage of the development. He ascribed this result to the CISK mechanism and argued that the friction-induced water vapor convergence dominates the kinetic energy dissipation during the intensifying process for TCs.

Since the late 1980s, the CISK mechanism has been challenged by wind-induced surface heat exchange instability (WISHE), a theory proposed as an alternative to CISK by Emanuel (1986). In WISHE, the atmosphere is assumed to be neutrally stratified along sloping surfaces of constant angular momentum. It is convection, rather than boundary-layer pumping, that plays a crucial role in the vertical transport of moist entropy, which is critical to the amplification of cyclones. Consistent with WISHE theory but contradictory to the CISK mechanism, the numerical results of Craig and Gray (1996) suggest that the spin-up of a TC is nearly insensitive to the changes in surface friction, except in the cases with particularly strong surface friction such that the development speed of a TC exhibits slightly negative dependence on surface friction. This finding implicates that during the development process of TCs, the moist entropy convergence caused by surface friction is dominant, or almost offset, by the friction-induced dissipation.

The theoretical investigations performed by Gray and Craig (1998) and Kieu (2004) further confirm the spin-down effect of surface friction on TCs. However, if the underlying assumptions of these theoretical works are reviewed, it can be determined that the negative dependence of the growth rate of TC on surface friction is a direct deduction from the assumptions. In Gray and Craig (1998), the thermodynamic equation is written as

$$\frac{\partial\theta_{\mathrm{e}}}{\partial t}=\frac{C_{\mathrm{E,T}}U(\theta_{\mathrm{es}}-\theta_{\mathrm{e}})}{h_{\mathrm{B}}} \tag{1}$$

where, θ_{es} and θ_{e} are the equivalent potential temperature near the sea surface and above the sea surface, respectively, and h_{B}, U, and $C_{\mathrm{E,T}}$ are the depth of the atmosphere, wind speed near the surface, and the heat and temperature drag coefficient, respectively. Equation (1) implies that an increase in boundary layer equivalent potential temperature increases θ_{e} uniformly in the whole convecting layer above it. Evidently, the relationship between the vertical transport of moist entropy and the vertical motion, including boundary-layer pumping, is not properly parameterized by Eq. (1). Unlike Gray and Craig (1998), Kieu (2004) took account of the vertical redistribution of thermodynamic energy induced by the vertical motion by expressing the thermodynamic and vertical motion equations as

$$\frac{\partial b}{\partial t}=\beta^2 w \tag{2}$$

$$\frac{\partial w}{\partial t}=b \tag{3}$$

where w and b are vertical velocity and buoyancy, respectively, and β^2 is a constant associated with the Brunt-Väisälä frequency. Obviously, only the vertical velocity linked with buoyancy is considered in the vertical transport of heat in Eqs. (2)–(3), and the heat transferred by the boundary-layer pumping is totally neglected. Since only the dissipative effect is taken into account in the theoretical models, the surface friction is automatically proved to be unfavorable to the intensification of TCs.

Using case studies of development of east Pacific easterly waves, Raymond et al. (1998) noticed that the spin-up of a TC depends ultimately on the balance between low-level vorticity convergence and surface friction. By comparing the respective tendencies of convergence and surface friction to spin up and spin down a developing disturbance, they proposed that friction is unimportant in the early stages of development. However, considering that the surface friction also makes a contribution to the low-level convergence, it can be inferred that the result of Raymond et al. (1998) only means that the dissipative effect of surface friction is not pronounced enough to apparently destroy the intensification of disturbances in the early stages of development.

Although moist entropy from the ocean and moist convective mixing are essential to the intensification of TCs, it is irrefutable that boundary-layer pumping resulting from surface friction also contributes to the vertical transport of moist entropy and thus also affects the evolution of TCs. An appropriate evaluation of these competing impacts on TC evolution is of interest and will improve the understanding of TC dynamics. In addition, similar to the situation for TCs, the role of surface friction in the Madden-Julian oscillation (MJO), or the tropical intraseasonal oscillation, is also in dispute (Chao and Chen, 2001). Since both MJOs and TCs take place in tropical ocean areas and are bound up with the convection, knowledge about the influence of surface friction on TC development may also be a good reference for discussions about the role of surface friction in the MJO. In view of the above reasons, the effect of surface friction on the evolution of TCs is re-examined with numerical simulations in the present work.

The outline of the paper is as follows. Section 2 presents the experimental design. The numerical results are displayed in section 3. Finally, section 4 reviews the main conclusions of the paper.

2. Experimental design

The numerical experiments were conducted using the nonhydrostatic, axisymmetric TC model originally written by Rotunno and Emanuel (1987). In brief, the nonhydrostatic, compressible equations of motion are integrated, with prognostic equations for momentum, potential temperature, nondimensional pressure, and mass mixing ratio of water vapor. Liquid water is also included. The turbulence parameterization is based on the first-order Richardson number-dependent eddy viscosity. The convection is treated explicitly rather than being parameterized.

The most important aspect of the model for the numerical experiments described here is the representation of the planetary boundary layer. For simplicity, surface fluxes of heat, moisture, and momentum are modeled using bulk aerodynamic formulae in the present work and are given by

$$F_\theta = C_{\rm E} U(\theta_{\rm s} - \theta) \quad (4)$$

$$F_q = C_{\rm E} U(q_{\rm s} - q) \quad (5)$$

$$F_{\rm m} = \rho C_{\rm D} U^2 \quad (6)$$

respectively, where ρ is density, U is the wind speed at the lowest model level, $(\theta_{\rm s} - \theta)$ is the potential temperature difference between the sea surface and the lowest model level, and $(q_{\rm s} - q)$ is the corresponding difference for the specific humidity. The momentum drag coefficient is given by Deacon's Formula:

$$C_{\rm D} = C_{\rm D0} \times 10^{-3} + C_{\rm D1} \times 10^{-5} U \quad (7)$$

where $C_{\rm D0} = 1.1 \times 10^{-3}$ and $C_{\rm D1} = 4 \times 10^{-5}$ (Rotunno and Emanuel, 1987). Following Craig and Gray (1996), the drag coefficient for heat (sensible and latent) $C_{\rm E}$ is assumed to be independent of wind speed and taken with a value of 1.0×10^{-3}.

The simulation is initialized with the mean hurricane season sounding of Jordan (1958). It is well known that the large low-level vortex disturbances related to easterly waves, disturbance in the Intertropical Convergence Zone (ITCZ), and/or baroclinic perturbations, etc. are prerequisite to the formation of TCs (Gray, 1968). In the present work, the low-level vortices are simply represented by two kinds of mesoscale vortices.

Since a low-level vortex in thermal wind balance has been widely used to initialize idealized simulation of TCs (Rosenthal, 1971; Rotunno and Emanuel, 1987; Craig and Gray, 1996), a low-level mesoscale vortex of the type depicted by Rotunno and Emanuel (1987) is adopted as an initial disturbance in Expt. A. The tangential wind field for the initial cyclonic vortex has a maximum wind speed of 14 m s^{-1} at a radius of 75 km at the surface and decreases with height (Fig. 1a). Given the tangential wind field for the initial cyclonic vortex, the mass and thermodynamic fields are obtained by using gradient and hydrostatic balance relationships. Figure 1b displays the initial potential temperature perturbation field, in which a warm core structure is conspicuous. For convenience, no moisture perturbation is introduced to the initial moisture field, which is given by the mean hurricane season sounding of Jordan (1958).

On the basis of the investigation of observational data, recent studies have suggested that Mesoscale Convective Vortexs (MCVs), which form in the stratiform precipitation region at midlevels of the troposphere in disturbed weather regions, are precursors to tropical cyclogenesis (Ritchie and Holland, 1997; Bister and Emanuel, 1997). Since a MCV has different structures in both the wind field and thermodynamics compared to a low-level vortex, the development of a TC originating from a MCV may proceed by a di-

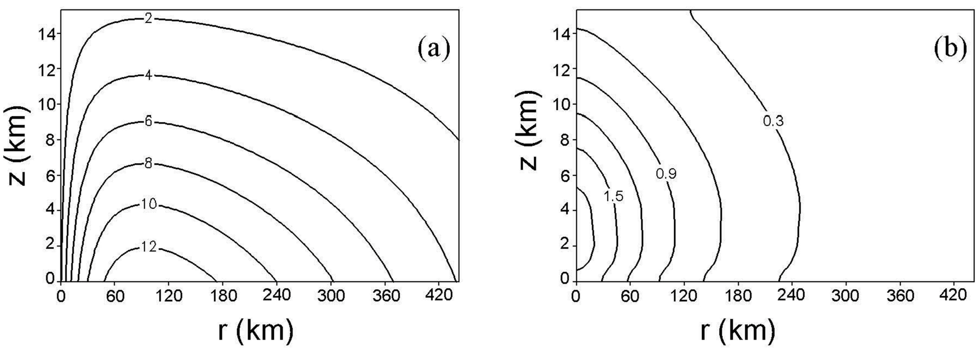

Fig. 1. (a) The initial azimuthal velocity and (b) potential temperature perturbation represented by contour lines in intervals of 2 m s^{-1} and 0.3 K, respectively.

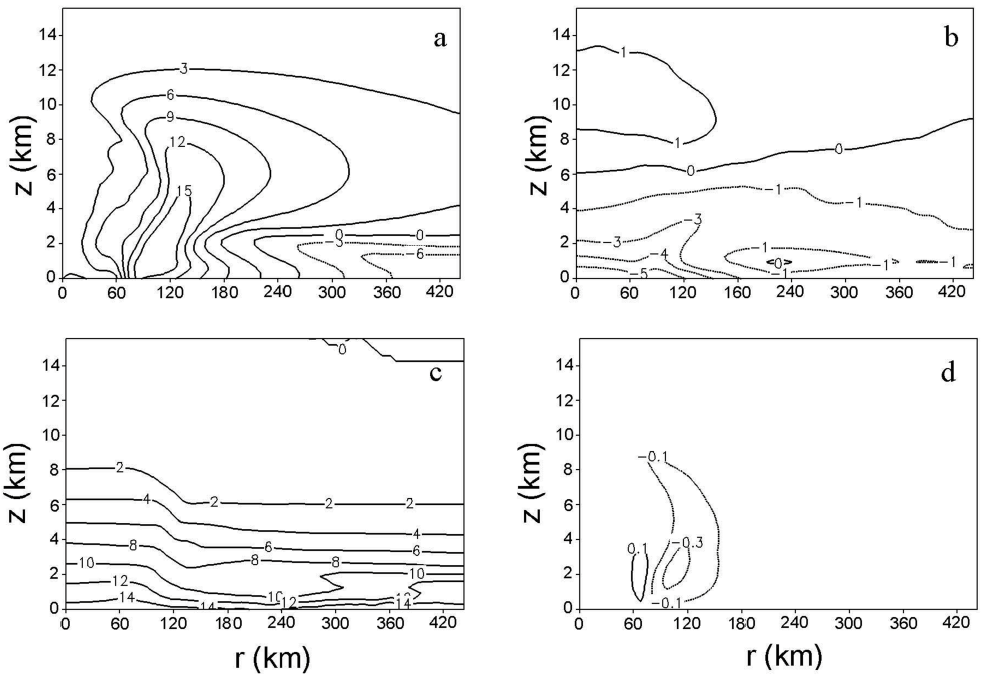

Fig. 2. The azimuthal velocity (a, units: m s^{-1}), perturbation of potential temperature (b, units: K), specific humidity (c, units: g kg^{-1}), and vertical velocity (d, units: cm s^{-1}) at the initial time of Expt. B.

fferent path. With the intention to roundly evaluate the role of the surface friction played in the tropical cyclogenesis, tropical cyclogenesis arising from a MCV is considered in Expt. B.

In order to produce the initial disturbance for Expt. B, a steady mesoscale "rain shaft" emanating from a prescribed altitude is switched on in the same way as in Bister and Emanuel (1997) at the beginning of the simulation. Any latent heating associated with the formation of this rain is ignored. Neither wind nor virtual temperature perturbations are present in the initial state. However, during the first 18 h of the simulation, the rainwater mixing ratio is set to

$$q_{\mathrm{r}} = 1.025\mathrm{e}^{-(r/150)^2-[(z-4.5)/3.75]^2} \tag{8}$$

where the units of rainwater mixing ratio and horizontal distance, altitude are g Kg^{-1} and km, respectively. After the "rain shaft" works for eighteen hours, a MCV with a certain amplitude has formed and the associated surface wind has a similar value as its counterpart in Expt. A. This moment is taken as the beginning of the numerical integration of Expt. B, and the corresponding initial fields are shown in Fig. 2. In fact, a MCV with humid and cold core similar to that displayed in Fig. 2 has been employed by Montgomery et al. (2006) in the investigation of the role of vertical hot towers in tropical cyclogenesis and by Nolan (2007) in the discussion of the trigger of tropical cyclogenesis.

With the aim to discussing the impact of surface friction on the spin-up of TCs, a series of numerical simulations with different surface friction are carried out both in Expt. A and Expt. B. Following Rosenthal (1971) and Craig and Gray (1996), changes in surface friction are represented by different values of the transfer coefficients in the bulk aerodynamic formula Eq. (7). $C_{\mathrm{D0}} = 1.1 \times 10^{-3}$ and $C_{\mathrm{D1}} = 4 \times 10^{-5}$ employed by (Rotunno and Emanuel, 1987) are taken as control values here. Based on this, four sensitivity experiments are conducted with surface friction approximately 80% (0.2F) or 40% (0.6F) reduced and 40% (1.4F) or 100% (2.0F) increased in Expt. A, and five sensitivity experiments are conducted with surface friction about 100% (0F), 80% (0.2F), or 40% (0.6F) reduced and 40% (1.4F) or 100% (2.0F) increased in

Table 1. The transfer coefficients used in Expt. A.

Expt. A-Ctrl	Expt. A-0.2F	Expt. A-0.6F	Expt. A-1.4F	Expt. A-2.0F
C_{D0}=1.1×10^{-3} C_{D1}=4×10^{-5}	C_{D0}=0.22×10^{-3} C_{D1}=0.8×10^{-5}	C_{D0}=0.66×10^{-3} C_{D1}=2.4×10^{-5}	C_{D0}=1.56×10^{-3} C_{D1}=5.6×10^{-5}	C_{D0}=2.2×10^{-3} C_{D1}=8.0×10^{-5}

Table 2. The transfer coefficients used in Expt. B.

Expt. B-Ctrl	Expt. B-0F	Expt. B-0.2F	Expt. B-0.6F	Expt. B-1.4F	Expt. B-2.0F
$C_{D0}=1.1\times10^{-3}$ $C_{D1}=4\times10^{-5}$	$C_{D0}=0$ $C_{D1}=0$	$C_{D0}=0.22\times10^{-3}$ $C_{D1}=0.8\times10^{-5}$	$C_{D0}=0.66\times10^{-3}$ $C_{D1}=2.4\times10^{-5}$	$C_{D0}=1.56\times10^{-3}$ $C_{D1}=5.6\times10^{-5}$	$C_{D0}=2.2\times10^{-3}$ $C_{D1}=8.0\times10^{-5}$

Expt. B (Tables 1 and Table 2). Considering that the changing behavior of moist convection as the cyclone evolves is central to the intensification of a TC (Raymond and Sessions, 2007), the attention of this work is primarily focused on the occurrence of convection in the sensitivity experiments designed above.

In the present work, the conclusion will be based on the period of the simulations in which the cyclone is intensifying. The rate of intensification of the TC is illustrated using plots of minimum central surface pressure as a function of time. Plots of maximum azimuthal wind show similar results but with greater noise levels, and will not be shown here.

3. Numerical results

3.1 *Impact of surface friction on the development of TCs in Expt. A*

Figure 3 shows the temporal variation of the minimum central surface pressure in Expt. A. Due to the inherent diffusion of the numerical model, the cyclone tends to decay at the beginning even when the surface friction is omitted, and the weakening is slightly aggravated when surface friction is taken into account. However, as the latent and sensible heat keeps flowing from the ocean to the atmosphere, the cyclone stops decaying and spins up gradually. With reference to Fig. 3, it is apparent that the stronger the surface friction is the faster the cyclone develops. After 12 hours of integration, the minimum central surface pressure in an experiment with significant surface friction (Expt. A-1.4F) ceases rising and starts to decreases little by

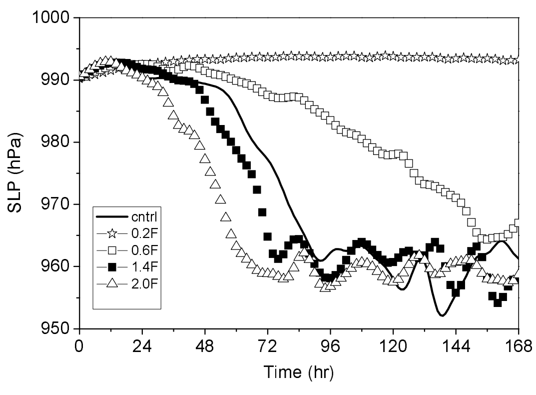

Fig. 3. The minimum central surface pressure as a function of time in Expt. A.

little. About 36 hours later, the cyclone undergoes rapid development and finally reaches steady state at t=72 h or so. A similar evolution also occurs in an experiment with weakened surface friction (Expt. A-0.6F), though the growth rate of the cyclone is reduced. However, when the surface friction is very weak (Expt. A-0.2F), the intensification is so trivial that the cyclone fails to become a TC within 240 hours. These results are consistent with those obtained by Rosenthal (1971) and have been explained from the viewpoint of the CISK mechanism, i.e., the primary source of water vapor for organized cumulus is the frictional convergence in the boundary layer, and accordingly, increased surface friction leads to increased water vapor convergence that dominates over the effects of the frictional dissipation and impels the TC to strengthen. Nevertheless, since the WISHE theory has been accepted as the leading explanation for the maintenance and intensification of existing TCs (Craig and Gray, 1996), the explanation given by Rosenthal (1971) about the reliance of the development of TCs on surface friction seems to be out of date to some extent.

In the framework of the WISHE mechanism, convection is critical to the development of cyclones. At earlier stages of development of a TC, before intensification, sporadic convection modifying the vortex structure in both the wind field and thermodynamics is necessary to the rapid intensification of a TC (Nolan, 2007). As rapid intensification begins, it is convection that keeps carrying the moist entropy upwards to fuel the amplification of the cyclone (Rotunno and Emanuel, 1987). Considering that the average state of the tropical atmospheric in summer, as represented, say, by Jordan (1958) sounding, is very nearly neutral to convection, it can be deduced that a key process related to the convection and then the development of TCs is the establishment of local conditional instability. It goes without saying that it is the surface flux of moist entropy that plays a principal role in such a process. However, this process is modified by the surface friction to some degree.

Figure 4 portrays the cross section of the initial 6-hour averaged vertical velocity, and deviations of specific humidity and potential temperature at t=6 h from the initial values in Expt. A-0.6F and Expt. A-2.0F. After a 6-hour integration, the specific humidity in the lower levels is conspicuously increased (Figs. 4c and

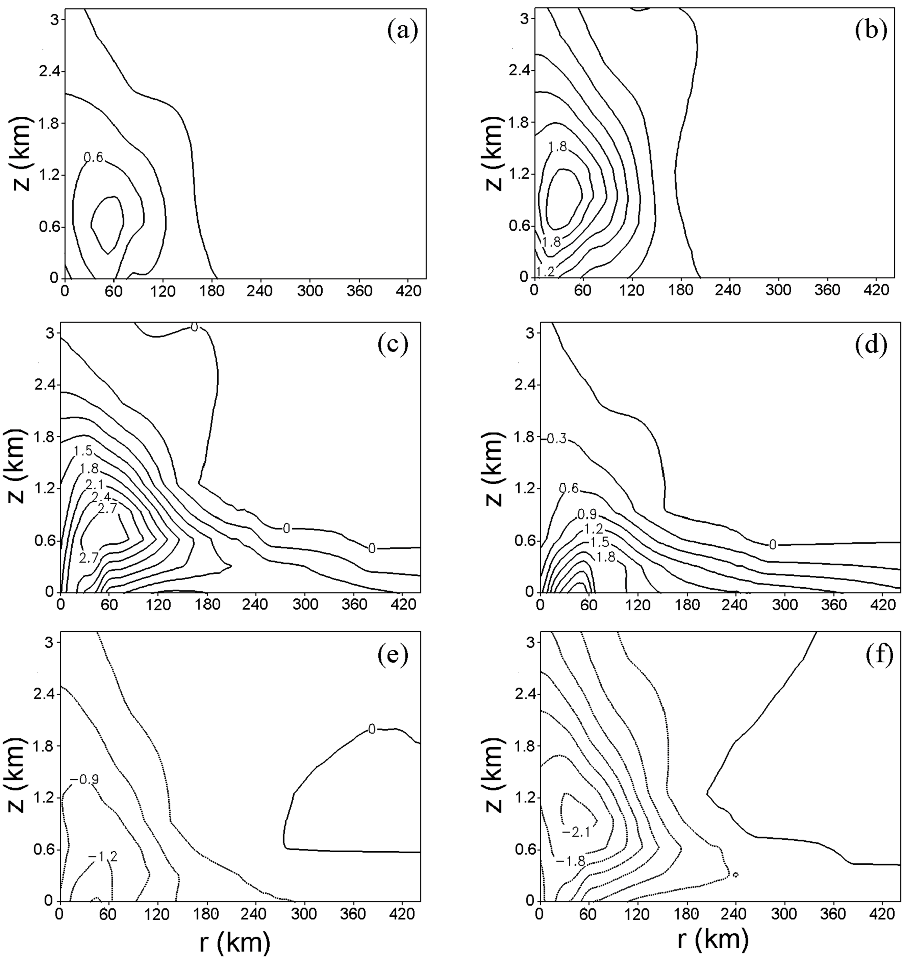

Fig. 4. The initial 6-hour averaged vertical velocity, and deviations of specific humidity and potential temperature at t=6 h from the initial values in Expt. A-0.6F (a, c, e) and Expt. A-2.0F (b, d, f). (a) (b): w (units: cm s^{-1}); (c) (d): $q_s - q_{s,\mathrm{initial}}$ (units: g kg^{-1}); (e) (f): $\Theta - \Theta_{\mathrm{initial}}$. (units: K)

4d). However, the increment of specific humidity near the ocean surface in Expt. A-2.0F is comparatively smaller than its counterpart in Expt. A-0.6F. This difference is clearly attributed to the different surface friction. The strong surface friction tends to reduce the surface winds, and thus cuts down the moisture flux from the ocean remarkably. Moreover, the moisture from the ocean is continuously conveyed upwards by the vertical motion. Considering that the initial fields in Expt. B satisfy the thermal wind equation, and with reference to the secondary circulation of the TC as inverted from the potential vorticity-vertical motion (PV-ω) system following Zhang and Kieu (2005) (that is, the vertical motion forced by latent heat and dry-dynamical processes are mainly situated in the free atmosphere while the vertical velocity associated with the boundary-layer frictional process is confined in the lower troposphere), it can be inferred that the vertical motion illustrated in Figs. 4a and 5b is largely ascribed to the boundary-layer pumping in the early stages of TC development[a]. Due to the strong surface friction, the vertical motion in Expt. A-2.0F is more prominent than that in Expt. A-0.6F, and consequently, more moisture is transferred to the interior and less is left at the bottom of the atmosphere. It is worth mentioning that the vertical transport of moisture accompanying the friction-induced ascent helps maintain the moisture gradient over the ocean surface,

[a]The maximum vertical velocity in Figs. 4a and 5b is not situated at the place predicted by classical Ekman theory (Ekman, 1905), i.e., at the center of the vortex, where the geostrophic vorticity is significant. The discrepancy is mainly ascribed to horizontal advection. It is found that boundary-layer pumping is not only dependent on the value of geostrophic vorticity, but also related to its horizontal distribution when the inertial force is introduced into the classical Ekman model (Wu and Blumen, 1982; Hart, 2000; Fang and Wu, 2005).

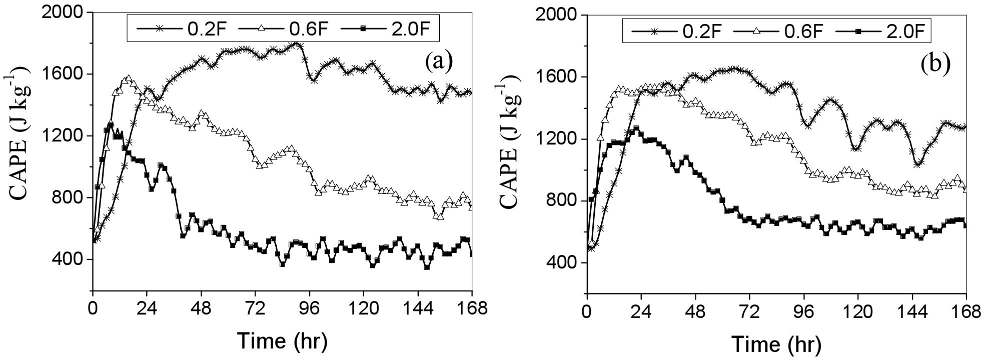

Fig. 5. Average CAPE measured across (a) 75 km and (b) 150 km sections from the center of the vortex as a function of time in Expt. A.

which guarantees the moisture continues to flow into the atmosphere from the ocean to support convective activity.

As a result of the increment of humidity in the lower troposphere, the low-level atmosphere is destabilized and the average convective available potential energy (CAPE) in the central area of the cyclone is enhanced. However, the variation of CAPE displayed in Fig. 5 seems to be inconsistent with the increment of specific humidity illustrated in Figs. 4c and 4d. In Expt. A-2.0F, the increment of specific humidity near the ocean surface is smaller than that in Expt. A-0.6F, while CAPE grows faster than its counterpart in Expt. A-0.6F at the beginning of the development of the TC. The inconsistency between the variations of the specific humidity and CAPE implies that CAPE may be affected by other processes more than just the surface flux of moisture from the ocean.

Just as noted by Holton (2004), the secondary circulation in the boundary layer induced by the surface friction not only reduces the azimuthal velocity field of the vortex through the action of the Coriolis force, but also changes the temperature distribution through adiabatic cooling of the air forced by boundary-layer pumping. A cold perturbation is located at the altitude of 1.0 km in Expt. A-2.0F, which nearly coincides with the position of the maximum vertical velocity (Figs. 4b and 4f). Inspection of Figs. 4d and 4f indicates that the cold perturbation happens to lie above the humid area. Colder and drier air just above a warm and moist air layer is favorable to the creation of unstable conditions and conducive to the formation of convection and then precipitation. In contrast, in Expt. A-0.6F, owing to the weak surface friction, the vertical velocity is smaller, and accordingly, the adiabatic cooling induced by the vertical motion is weaker (Figs. 4a and 4e). Therefore, the increase of CAPE is not as rapid as in Expt. A-2.0F. The substantial difference between the temporal variation of CAPE in the early development stage of a TC is shown by experiments with and without surface friction, as exemplified in Fig. 5. It is straightforwardly manifest that, apart from the surface flux of moist entropy, the boundary-layer pumping also plays a stimulative role in the establishment of the unstable environment through modification of the vertical temperature distribution.

The occurrence of convection demands not only instability, but also parcel saturation at the environmental temperature of the level where the convection begins (i.e., the parcel must reach the level of free convection). The mean relative humidity in the troposphere is well below 100%, even in the boundary layer. Thus, low-level convergence with resultant forced layer ascent or vigorous vertical turbulent mixing in the boundary layer is required to produce saturation (Holton, 2004). Since the initial fields employed in Expt. A are in thermal wind balance, only the boundary-layer pumping associated with the frictional convergence in the boundary layer can advect the parcel to the level of free convection in the early development stage of the cyclone. From this point of view, it is not difficult to understand why CAPE in Expt. A-2.0F is reduced earlier than in Expt. A-0.6F.

Based on the above analysis, the conclusion can be drawn that, in addition to conveying moist entropy into the interior to fuel convection as emphasized in CISK theory, the surface friction can further facilitate convection by adjusting the thermodynamic structure, through adiabatic cooling, to be advantageous to the formation of conditional instability and by providing lifting for the release of CAPE. Nonetheless, it is worthwhile to mention that the conclusion made above is tenable only under the condition that the surface friction is not very strong. If the surface friction is excessively strong, the surface wind will be enormously reduced, and correspondingly, the moist entropy flux from the ocean will not be abundant enough to overcome the intense kinetic energy dissipation. As a consequence, the intensification of the TC will be inhibited by surface friction.

3.2 *Impact of surface friction on the development of TCs in Expt. B*

Besides the low-level vortex, much observational research suggests that MCVs spawned in long-lasting MCSs can also be transformed into TCs under favorable circumstances (Bister and Emanuel, 1997; Raymond and Sessions, 2007). The effect of surface friction on the transformation of such kinds of disturbances into TCs is to be analyzed in this section.

In Expt. B, the spin-up of cyclones under the condition that SST equals 26.3°C is illustrated in Fig. 6. Contrary to Expt. A, the model TCs in Expt. B possess comparable growth rates in the first 96 hours whether the surface friction is considered or not. The insensitivity of the growth of TCs to surface friction implies that, at the early development stage of TCs, the impact of surface friction on the construction of an amicable environment for convection (as discussed in the Expt. A results) becomes less important to the spin-up of TCs, i.e., the local conditional instability, moisture supply, and lifting necessary to the occurrence of convection and development of TCs can be gained by other efficient means in Expt. B.

By referring back to Fig. 2, several special features can be identified in the initial vortex of Expt. B that contrast with the situation of Expt. A (Fig. 1). First of all, a cold core instead of a warm core appears in the lower troposphere (Fig. 2b). In comparison with the warm core, the cold core has advantages in at least in two aspects for the construction of locally unstable air conditions. On the one hand, the lower the atmospheric temperature is, the more sensible heat is transferred into the atmosphere from the ocean. On the other hand, the cold core aloft lowers the value of boundary layer equivalent potential temperature that

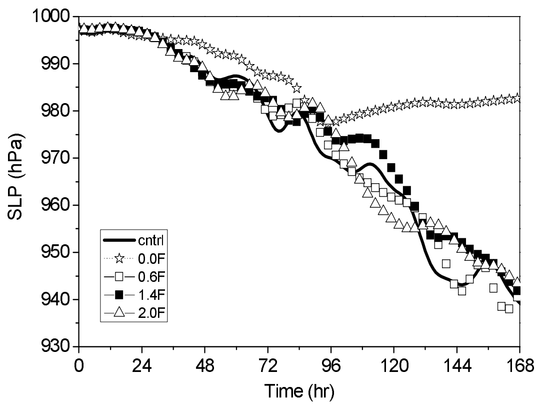

Fig. 6. Perturbation of the central surface pressure as a function of time in Expt. B.

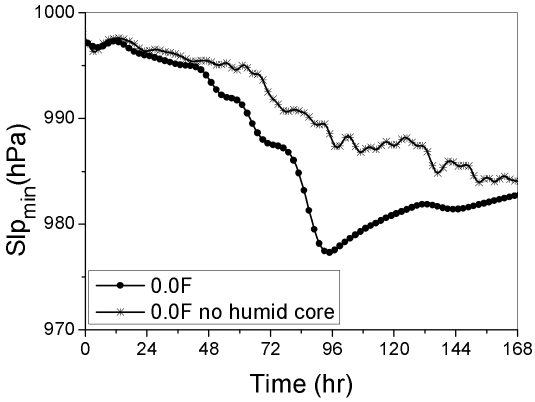

Fig. 7. Perturbation of the central surface pressure as a function of time in Expt. B-0F, with and without initial humid core.

is needed for the formation of conditional instability.

Secondly, unlike the uniform distribution of initial specific humidity in Expt. A, the atmosphere in the central area of the initial vortex is moister than its surroundings in Expt. B (Fig. 2c). Contrasting the development process of the TC in Expt. B-0F to its counterpart in the sensitivity experiment with the initial humid core artificially removed (Fig. 7), it is evident that the initial humid perturbation has an substantial influence on the spin-up of the model TC. This may be partially due to the fact that enriched moisture initially accumulated in the atmosphere assures convection of a moisture supply at the beginning of TC development. In addition, according to Bister and Emanuel (1997) and Raymond and Sessions (2007), a humid cold-vortex diminishes the evaporation of rain, and thus discourages convective downdrafts that plague a warm-core disturbance in an otherwise undisturbed environment, favoring the spin-up of the TC by facilitating the strengthening of low-level convergence.

Lastly, the vigorous vertical motion illustrated in Fig. 2c means that the initial mass and momentum fields do not satisfy the thermal wind balance in Expt. B. The vertical motion arising from the following geostrophic adjustment process may act as a substitute for the boundary-layer pumping in the vertical transport of moisture and lifting of air parcels at the early development stages of a TC.

By virtue of the special characteristics listed above, the initial fields of Expt. B have advantages for the formation of amicable conditions for convection and for lessening the convective downdrafts, which is consequently more beneficial to the development of TCs than conditions in Expt. A. The role of surface friction in moisture supply, the lifting of air parcels, and the establishment of conditional instability, which are in-

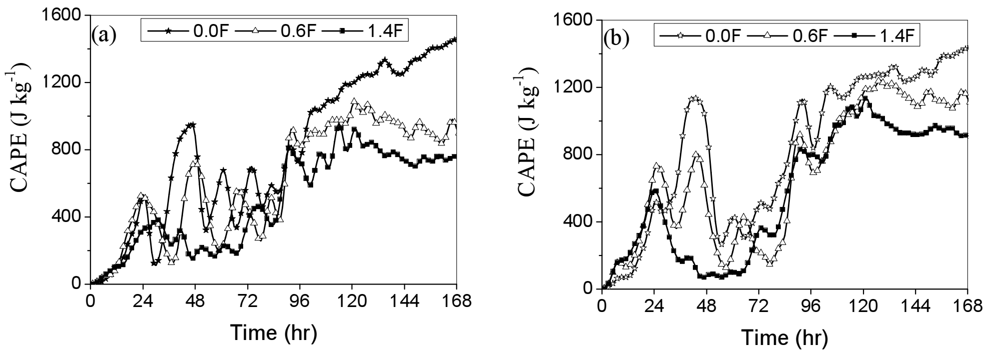

Fig. 8. Average CAPE measured across (a) 75 km and (b) 150 km sections from the center of the vortex as a function of time in Expt. B.

dispensable to the growth of TCs in Expt. A, is less essential in Expt. B, and in fact, it is almost cancelled out by the friction-induced dissipation as indicated by Fig. 6.

Figure 8 shows the average of CAPE measured across 75 km and 150 km sections from the center of the vortex as a function of time. In accordance with the variation of minimum central surface pressure delineated in Fig. 6, the variation of CAPE in Expt. B also exhibits noticeable differences from its counterpart in Expt. A. In Expt. A, CAPE undergoes an abrupt increase at first, and then gradually decreases during the spin-up of the TC (Fig. 5). It is in the decreasing period that the system is rapidly strengthened. At the beginning of the development of the cyclone in Expt. B, CAPE also demonstrates an increasing trend. But the increment is smaller than that in Expt. A because the vertical motion accompanying geostrophic adjustment tends to promote convectivity, and then the release of CAPE. It is the release of CAPE counteracting the friction and inherent dissipation that keeps the minimum central pressure in the early developing stage of the TC nearly constant in Expt. B instead of increasing as in Expt. A (Figs. 6 and 3).

Since the weak conditional instability cannot support long-lasting convection, convective activity slows gradually along with the release of CAPE until more moist entropy can be transferred into the atmosphere from the surface to fuel further convection. As a result, the values of CAPE begin to fluctuate with time after a short increasing period. In contrast to the sharp undulation in Expt. B-0F and 0.6F, the CAPE fluctuations in Expt. B-1.4F are rather mild. The reason is that the comparatively strong boundary-layer pumping facilitates the release of CAPE, more or less. Due to the failure to construct a veritable reservoir of CAPE, the model TCs in Expt. B experience relatively slow and steady spin-up instead of fast development as seen in Expt. A (Figs. 6 and 3).

Along with the intensification of the systems, the vortex winds gradually intensify and the sea-surface transfer of moist entropy is enhanced near the region of strongest wind and low pressure, as prescribed by the bulk aerodynamic formulas Eqs. (4)–(5), and then leads to considerable accumulation of CAPE. Meanwhile, the surface friction is also strengthened as a result of the increase of surface wind, and starts to play an apparent role in the release of CAPE when friction is strong enough. This is the reason why the variations of CAPE in Expt. B-0F, 0.6F, and 1.4F eventually present obvious differences. However, as a whole, the impact of surface friction on the accumulation and release of CAPE in Expt. B is not as distinct as in Expt. A, which further confirms that surface friction does not have a vital effect on the development of TCs in Expt. B as it does in Expt. A.

From the numerical results discussed above, it can be concluded that the importance of surface friction on the intensification of TCs varies considerably from case to case. In Expt. A, due to the warm core and thermal wind balance in the initial fields, the formation of favorable conditions for convection depends on the vertical motion caused by surface friction, which therefore plays an essential role in driving the cyclone to spin up. However, as Expt. B is concerned, the initial humid cold-vortex favors the development of the TC by facilitating the local construction of conditional instability and moisture supply, and by avoiding the downdraft problem that plagues the enhancement of low-level convergence. Furthermore, at the beginning of TC development, the lifting of air parcels and vertical transport of moisture can be fulfilled to a certain degree by the nontrivial vertical motion arising from geostrophic adjustment in Expt. B. Therefore, the importance of surface friction to the spin-up of the TC is substantially lessened in Expt. B. Nevertheless, because of the inherent complexities of TCs, the above statements cannot be expected to be robust, even in the case that the initial fields are similar to that in Expt. B.

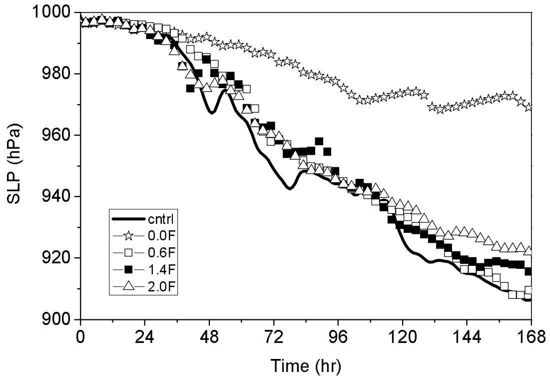

Fig. 9. Perturbation of the central surface pressure as a function of time in Expt. B (SST=28.3°C).

As an example, Fig. 9 shows the minimum central surface pressure as a function of time in experiments identical to Expt. B except that the SST is now set at 28.3°C instead of 26.3°C. It is obvious that although the growth rates of model TCs are close to each other in the experiments considering surface friction, the system develops at a comparatively low speed in the experiment with surface friction omitted. The reason can be simply explained as follows. Since the initial surface winds are almost identical, the higher SST indicates that more sensible heat is transferred into the atmosphere. However, owing to the weak vertical motion in the case without surface friction, the increased sensible heat cannot be advected upwards to play an effective role in the intensification of the system. Only under conditions when surface friction is taken into account can the additional sensible heat participate in the development of cyclones with the aid of vertical motion augmented by boundary-layer pumping.

4. Conclusion

Surface friction plays a dual role in the development of TCs. Regarding the moist entropy supply, the vertical motion induced by friction exports moisture and heat to the free atmosphere which impels the cyclone to intensify. On the other hand, the surface friction dissipates kinetic energy and tends to weaken the vortex. Which role is more influential to the spin-up of TCs has been the subject of much debate.

In the framework of the CISK mechanism, it is straightforward that surface friction has a positive effect on the growth of TCs because the frictional convergence in the boundary layer is taken as the primary source of water vapor for organized cumulus convection. However, since the WISHE theory, in which the importance of the collection and vertical transport of water vapor associated with surface friction is de-emphasized, has been accepted as the leading explanation for the maintenance and intensification of TCs, it is necessary to understand and evaluate the effects of surface friction on the evolution of TCs from the perspective of WISHE theory.

By performing two sets of idealized numerical experiments with different initial vortices, the role of surface friction in the intensification of TCs is assessed in the present work. It is found that, owing to the intrinsic complexity of TC development, the influence of surface friction on the growth of TCs may vary from case to case. In the case that a TC starts from a low-level vortex with a warm core, the surface friction helps the cyclone to spin up gradually through the promotion of moisture supply, lifting, and the establishment of conditional instability, which are essential ingredients for the occurrence of convection. On the contrary, if a TC commences from a MCV, which is a by-product of a mesoscale convective system, the importance of surface friction to the construction of favorable conditions for convection is significantly decreased, because the initial conditions (especially the humid cold-vortex) not only enable the local construction of conditional instability, moisture supply, and the lifting of air parcels to be easily accomplished, but also the conditions discourage convective downdrafts which inhibit the enhancement of low-level convergence. Consequently, the development of TCs is not as sensitive to surface friction under such conditions, in comparison to their counterparts evolving from low-level vortices. However, as SST is increased, the situation may be changed, and the difference between the growth rates of TCs modeled with and without surface friction may be noticeable once more. In summary, given the complicated nature of TCs, it is impossible to provide a unified description of the comprehensive effects of surface friction on the spin-up of TC under different conditions. Indeed, the influence of surface friction on TC development varies enormously depending on the circumstances of TC development involved, such as the initial fields, the SST, and so on.

It is worth mentioning that the sensitivity experiments on surface friction conducted in the present work are based on the bulk aerodynamic formula for surface flux of momentum via Eq. (6) and on Deacon's Formula for the momentum drag coefficient, Eq. (7). However, the parameterization of the momentum drag coefficient under high wind conditions is in dispute. It is proposed that the drag coefficients should become independent of wind speed at high wind speed. Smith (1988) argues that the drag coefficient strongly depends on wind speed up to 32 m s^{-1}. According to Global Positioning System dropwindsonde data, Powell et al. (2003) suggests that the wind dependence of the drag coefficient is reduced for 10 m height wind speed above about 40 m s^{-1}. Since the present work is mainly focused on the role of surface friction in the developing stages of TCs in which the maximum wind speed is usually much smaller than 40 m s^{-1}, Deacon's Formula can be taken as a reasonable parameterization for the drag coefficient in the numerical simulations carried out in this work. However, as mature TCs are concerned, further understanding about the influence of surface friction on the maintenance and decay of TCs relies on the development of new, accurate parameterization schemes for the drag coefficient based on measurements in high wind regimes.

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