Effects of Surface Exchange Coefficients for High Wind Speeds on Typhoon Structure: Numerical Simulations for Typhoon IOKE (2006)

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1 INTRODUCTION

Air–sea interaction is primarily important for the generation and development of tropical cyclones (TCs) (e.g., Emanuel 1986). The evaluation of heat and momentum fluxes at the sea surface is critical for simulating TCs; hence, the exchange coefficients for the fluxes are important. It has been recognized that the profile of drag coefficient (C_d) levels off as wind speed increases under strong wind conditions while decreases as wind speed further increases, from observational, experimental, theoretical, and numerical studies (e.g., Powell et al. 2003; Black et al. 2007; Donelan et al. 2004; Alamaro 2001; Makin 2005; Moon et al. 2004a,b,c). Moon et al. (2007) employed, in their hurricane simulations, the surface parameterization of Moon et al. (2004a) which incorporates the C_d behavior under strong wind conditions, and showed that modifying the plofiles of surface exchange coefficients for momentum and heat/moisture (C_k) changed the surface wind speed and momentum and heat fluxes, but made no difference in the minimum surface pressure. Although there are several studies that investigate the impact of the C_d and C_k behaviors on the prediction of hurricane track and intensity mainly for an operational point of view (Moon et al. 2007; Chen et al. 2007), there are few studies that deal with the influence of the C_d and C_k variability on TC intensity.

The present study examines the effects of surface exchange coefficients under strong wind conditions on typhoon intensity and structure. For this purpose, numerical simulations for Typhoon IOKE (2006) are conducted by using a model that incorporates the C_d and C_k behaviors in strong winds. We discuss the effects of exchange coefficients on typhoon structure as well as intensity.

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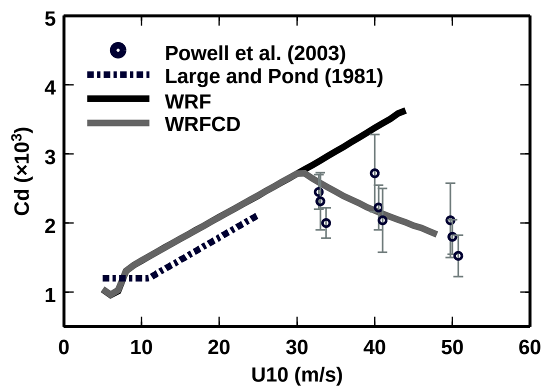

Figure 1: The relationship between 10–m wind speed and drag coefficient (C_d). The black and gray dots represent the relationships simulated by WRF and WRFCD, respectively. The solid line represents the formula derived from field measurements (Large and Pond 1981). The error bar shows the observational results of Powell et al. (2003).

2 MODEL AND EXPERIMENTAL SETTINGS

2.1 Modeling the effects of surface drag

In this study, we use the Advanced Research version 2.2 of the Weather Research and Forecasting Model (WRF, Skamarock et al. 2005), which is a fully–compressible, nonhydrostatic model. In order to examine the effects of surface drag, the roughness–length formulation in WRF is replaced with the equations obtained from the recent observations (Makin 2005) which include the effect of sea spray having an important effect on TCs (e.g., Andreas and Emanuel 2001; Wang et al. 2001).

The modified equations are shown as follows:

$$z_0 = c_l^{1-1/\omega} c_{z_0}^{1/\omega} \frac{u_*^2}{g}, \quad (1)$$

$$\omega = \min\left(1, \frac{a_{cr}}{\kappa u_*}\right), \quad (2)$$

where c_l(=$h_l g/u_*^2$) and c_{z_0}(=$z_0^l g/u_*^2$) represent the Charnock coefficient for suspension layer and local points, respectively. h_l is the height of suspension layer which is described in Makin (2005); κ is the von Karman constant; g is the gravitational acceleration; z_0^l(=ωz_0^w) is the local roughness length and z_0^w is the roughness length normally calculated in WRF. a_{cr} is the critical terminal velocity for typical sea spray droplet (we set a_{cr}=0.64 ms^{-1}). ω is a function which includes the effect of sea spray, and z_0^l becomes z_0^w in the case of no sea–spray effect (ω=1). We refer to the WRF model that includes the equations (1) and (2) as WRFCD. The drag coefficients with respect to the wind speed at 10–m height (U_{10}) in the simulations are shown in Figure 1. The profile of drag coefficient simulated in WRFCD agrees well with the observational results of Powell et al. (2003).

2.2 Experimental settings

The case examined here is Typhoon IOKE (2006). NCEP final analysis meteorological data were used for the initial and boundary conditions. The computational domain covered a 1500 km by 1500 km area (with horizontal grid spacing being 5 km) extending up to 50 hPa in the vertical (the grid number was 45). The domain was centered at 175E, 17N. The simulation period was from 1200UTC 28 August to 0000UTC 1 September and the time step was 30 second. The boundary and surface layer scheme used here were based on the Monin-Obkhov similarity theory (Janjic 2001), and the ice microphysics (Hong et al. 2004) and atmospheric short–wave and long–wave radiation schemes (Dudhia 1989 and Mlawer et al. 1997) were included. The cumulus parameterization was not included. The best–track data of Regional Specialized Meteorological Center (RSMC) were used as a reference for the central pressures of this typhoon, while those of Joint Typhoon Warning Center (JTWC) were used as a reference for the maximum wind speeds. It is noted that the RSMC wind speeds are averaged for ten minutes and therefore may not be appropriate to compare with the simulated results. The initial disturbance due to the typhoon was created as a Rankin–like vortex by the hurricane bogusing scheme of MM5 (5th generation Mesoscale Model, (Dudhia, 1993).

3 RESULTS FOR TYPHOON IOKE (2006)

We have conducted two numerical simulations. Figures 2a and 2b show the time series of minimum surface pressure and maximum wind speed of the simulated typhoons. Although the minimum surface pressures are not so different between the simulations, the values in WRFCD correspond better with the best estimates from t=36h to t=60h (from 0000UTC 30 to 0000UTC 31 August) than those in WRF. The maximum wind speeds are higher in WRFCD than in WRF, which is obviously due to the reduced drag in WRFCD. The wind speeds in WRFCD well agree with the best estimates of JTWC. The differences between WRF and WRFCD amount to 7.05 hPa for the pressure and 20.69 ms^{-1} for the wind speed, respectively, during the steady–state period. In spite of the difference in wind speed, the radius of maximum wind speed (RMW) is similar with each other; the temporal means of the RMW in WRF and in WRFCD are 67.17 km and 71.48 km, respectively.

Figures 2c, 2d, and 2e represent the temporal variation of friction velocity, sensible heat flux and latent heat flux, respectively, averaged over the area within the radius of 100 km centered at RMW. It is seen that the fluxes in WRFCD are smaller than those in WRF; this is considered to be due to the reduced exchange coefficients in strong winds in WRFCD. Note that the ratio of the coefficients for transfers of enthalpy (C_k) and momentum (C_d) is almost constant in the models.

The reason why the minimum surface pressure does not change, in spite of the reduced surface fluxes, can be explained by the air–sea interaction theory of Emanuel (1986) who regards a hurricane as a Carnot heat engine. In this system, the heat source is sea–surface

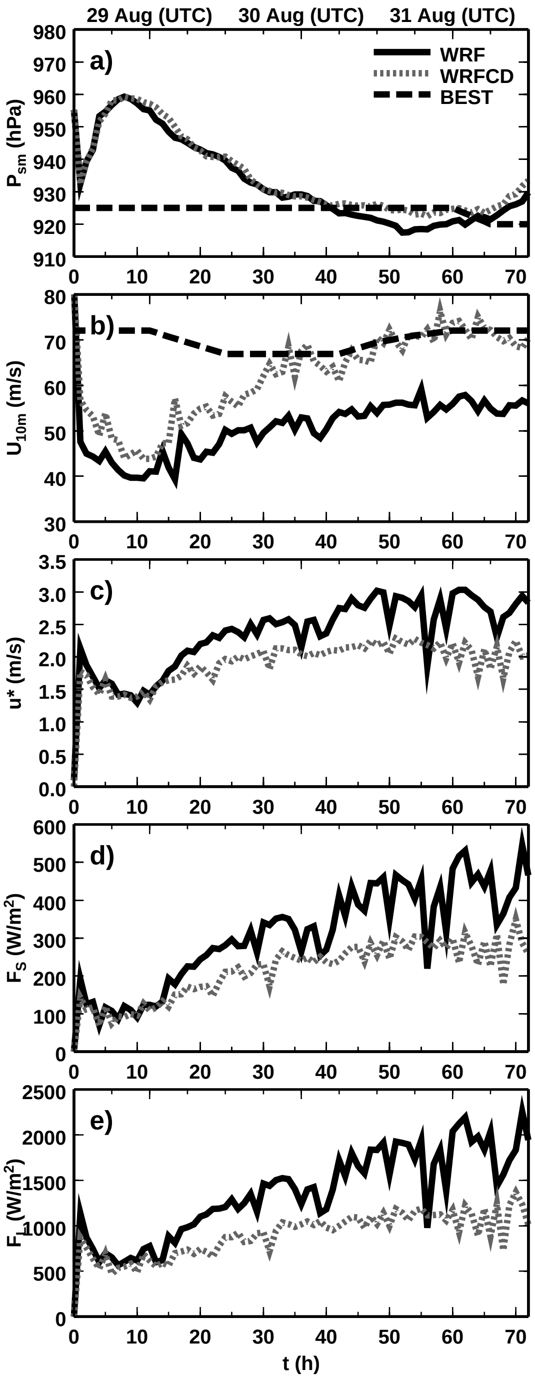

Figure 2: Time series of a) minimum surface pressure, b) maximum wind speed, the area-mean values of c) friction velocity, d) sensible heat flux and e) latent heat flux. The black solid, the black dotted and the dashed line show the results of WRF and WRFCD, and best track, respectively.

flux which is proportional to C_k and the work is produced by frictional dissipation which is proportional to C_d. Since in the present models C_k/C_d is almost constant, changing C_d and C_k simultaneously does not produce a remarkable difference in the typhoon intensity and thus the minimum surface pressure. In this sense, the results indicate that the important parameter is the ratio of the exchange coefficients (C_k/C_d) as presented by Emanuel (1995) and Bister and Emanuel (1998).

In addition to the qualitative explanation, the reason can be explained by using the energy balance equation of Emanuel (1986). In the system, the equation at the RMW, r_m, is shown as below:

$$\frac{1}{2}\left(V_m^2 + f r_m V_m\right) + C_p T_B \left(\ln \pi + \epsilon \ln \frac{\theta_{em}}{\theta_{ea}}\right) - \frac{1}{4} f^2 \left(r_0^2 - r_m^2\right) = 0, \quad (3)$$

where V_m is the maximum wind speed; f is the Coriolis parameter; C_p is the specific heat capacity at constant pressure; π is the Exner function; $\epsilon = (T_B - T_O)/T_B$ is the thermal efficiency; T_B and T_O are the absolute temperatures at the boundary–layer top and the outflow heigh (i.e., tropopause), respectively; r_0 is the outer radius at which the surface wind vanishes; θ_{em} and θ_{ea} are the equivalent potential temperatures of the boundary layer at the RMW and that of the ambient boundary layer, respectively. In the WRF and the WRFCD cases, T_B, r_m, θ_{ea} and r_0 are almost the same. Thus, the increased V_m in the equation (3) should require the decrease in θ_{em}. In the present results, θ_{em} actually decreases in the WRFCD (from 363.8 K and 361.0 K), which is due to the reduced surface fluxes in the WRFCD. Therefore, the increase in maximum wind speed produces no change in the central pressure because of the decreased surface fluxes.

In order to further demonstrate the sensitivity of typhoon intensity to the surface drag, the differences in water content and vertical velocity, as a measure of convective activity, between WRF and WRFCD are shown in Figure 3. The magnitudes of those in WRFCD are generally smaller than those in WRF, which are more focused around the eyewall region. It is also clear that the radius of maximum convective activity in WRFCD is larger than that in WRF. The results shown in Figures 2 and 3 seem to be consistent with each other, since the smaller surface fluxes reduce the strength of convective clouds.

One might argue, however, that the reduced convective development results from the decreased surface convergence, owing to the small C_d. Thus, we examine convergence as a function of a radius calculated by using the Gauss's divergence theorem in two dimensions outside the RMW region:

$$conv(r,t) = \oint u(x,y)\sin\theta dl/S, \qquad (4)$$

where $u(x,y)$ is the wind speed computed by the models; θ is the deviation of the angle between computed wind and gradient–wind balanced with the computed surface pressure gradient at each time. S is the surface area over a radius r. The reduced C_d at high wind speed is expected to reduce surface convergence. The result shown in Figure 4, however, indicates that the reduced C_d does not change the surface convergence, but rather makes the convergence slightly stronger just outside of the RMW. This may be explained as follows: the decreased surface drag makes the inflow angle smaller, while makes the wind speed larger; thus, the magnitude of convergence does not change in the decreased C_d cases. Consequently, the reduced convective activity around the eyewall region results from the decreased exchange coefficients in strong winds.

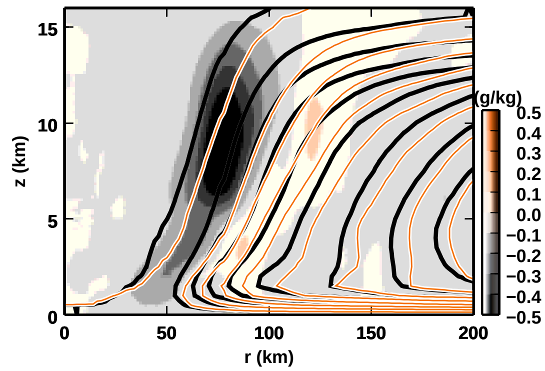

Figure 3: Temporal and azimuthal average of the difference in total water content (shaded) between WRFCD and WRF, and streamline. The orange and black lines show the streamline of WRF and WRFCD, respectively. The averaging period is from t=36h to t=60h (0000UTC 30 to 0000UTC 31 August).

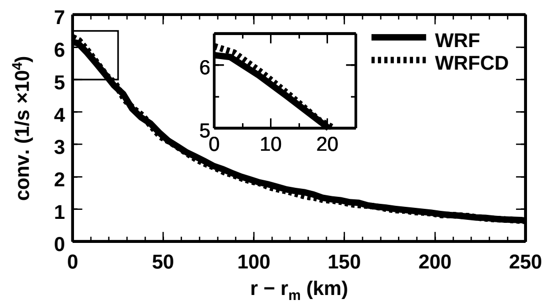

Figure 4: Temporal and areal mean convergence versus the radius in the outer region of RMW. The lines are the same as those in Figure 2 and the averaging period is same as Figure 3.

4 CONCLUSION

We have conducted numerical simulations for Typhoon IOKE (2006) by using the WRF model and that with the revised formulation of exchange coefficients, and examined the influences of surface exchange coefficients in strong winds upon the typhoon intensity. It is found that the reduced exchange coefficients in strong winds have little impact on the minimum surface pressure of the typhoon, while they lead to the significant increase in the maximum wind speed. In addition, the surface heat and moisture fluxes averaged within the radius of 100km centered at the RMW decrease in the reduced–C_d simulation. These results are qualitatively the same as indicated by Moon et al. (2007). Furthermore, it is found that the convective intensity represented in terms of water content and updraft velocity is significantly decreased in the reduced–C_d case specifically in the eyewall region. It is also shown that in spite of the reduced drag coefficient at high wind speed, the difference in surface convergence is not so significant between WRF and WRFCD. Thus, the wind convergence does not play a role in changing the convective intensity. These results suggest that the reduced C_d and C_k values under strong wind conditions have a negative impact on the convective development in the eyewall.

A future work should require a more in–depth study of the sensitivity of typhoon structure as well as intensity to the ratio of the exchange coefficients (C_k/C_d). Furthermore, it is interesting to examine how the difference in convergence outside the RMW region would produce the difference in the structure of outer rainbands.

Acknowledgments

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