# The impact of spray mediated enhanced enthalpy and reduced drag coefficients in modelling of tropical cyclones

N.C. Zweers‡, V.K. Makin, J.W. de Vries
*Royal Netherlands Meteorological Institute (KNMI), De Bilt, The Netherlands*

V.N. Kudryavtsev
*Nansen International Environmental and Remote Sensing Centre (NIERSC), Saint Petersburg, Russia*
*Russian State Hydrometeorological University (RSHU), Saint Petersburg, Russia*

**Abstract.** The impact of new parametrizations for drag and air-sea enthalpy exchange on modelling the intensity of tropical cyclones with a numerical weather prediction (NWP) model is examined. These parameterizations follow from a model for the marine atmospheric boundary layer (MABL) for high wind speed conditions in the presence of spray droplets that originate from breaking wave crests. This model accounts for the direct impact of these droplets on the air-sea momentum flux through action of a *spray force*, which originates from the interaction of the 'rain' of spray droplets with the vertical wind shear and is expressed in terms of the spray generation function (SGF). The SGF is cubic in the wind speed up to 50 m s$^{-1}$ beyond which its value increases less strongly. The drag coefficient ($C_D$) decreases from approximately 30 m s$^{-1}$, as in agreement with what the available measurements in these conditions indicate. The enthalpy exchange coefficient ($C_k$) increases with increasing wind speed and slowly decreases beyond a wind speed of about 40 m s$^{-1}$ due to the strong drop in $C_D$. The value for $C_k/C_D$ is in agreement with observational data for wind speeds up to 30 m s$^{-1}$; for higher wind speeds the value is in the range 1.2–1.5 as in agreement with a well established theory. The parametrization is tested in an NWP model. The tropical cyclones Ivan (2004) and Katrina (2005) in the Gulf of Mexico are simulated. To the sea surface temperatures (SSTs) from the European Centre archive that were prescribed to the NWP model, a parametrized cooling (based on estimations from theoretical studies and measurements) was applied during the model forecasts, as the NWP model does not resolve locally rather strong induced reductions in SSTs. The simulations show that realistic tropical cyclone wind speeds and central pressure can be obtained with the proposed drag and enthalpy parametrization. The results indicate that the value for $C_k/C_D$ at very high wind speeds in this study is in the correct range. Moreover, the results motivate the application of the parametrizations in atmosphere–ocean coupled models.

**Keywords:** Drag Coefficient, Enthalpy Exchange Coefficient, High Wind Speeds, Spray, Tropical Cyclones

## 1. Introduction

One of the challenges in tropical cyclone modelling is understanding and representing the physical processes near the sea surface that determine the surface fluxes of latent and sensible heat, and momentum. Tropical cyclones, also known as hurricanes, gain their intensity from the warm ocean through exchange of heat and moisture, while momentum is exchanged through the

‡ Current affiliation: MeteoGroup, Wageningen, The Netherlands
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drag that is exerted on the atmospheric flow. Numerical weather prediction (NWP) models normally compute the air-sea fluxes of momentum $\tau$, sensible heat $H_S$ and latent heat $H_E$ with a bulk type relation with an exchange coefficient according to,

$$\tau = \rho_a C_D U_L^2, \tag{1}$$

$$H_S = \rho_a c_p C_H U_L (\theta_0 - \theta_L), \tag{2}$$

$$H_E = \rho_a L_v C_E U_L (q_0 - q_L), \tag{3}$$

with $\rho_a$ the density of air, $c_p$ the specific heat of air (at constant pressure), $L_v$ the latent heat of vaporization, and $U$, $\theta$ and $q$ are respectively the wind speed, potential temperature and specific humidity, with subscript 0 the surface and $L$ a reference level height. $C_H$ and $C_E$ are the exchange coefficients for respectively heat and moisture, and $C_D$ is the drag coefficient.

For neutral stratification, the wind field in the marine atmospheric boundary layer (MABL) in NWP models is assumed to be logarithmic with height. Then,

$$C_D = \kappa^2 / \left[\ln^2\left((z + z_{0M})/z_{0M}\right)\right], \tag{4}$$

$$C_H = \kappa\sqrt{C_D} / \left[Pr_t \ln\left((z + z_{0H})/z_{0H}\right)\right], \tag{5}$$

$$C_E = \kappa\sqrt{C_D} / \left[Pr_t \ln\left((z + z_{0E})/z_{0E}\right)\right], \tag{6}$$

with $\kappa$ (=0.40) the von Karman constant, $z$ the height above the sea surface, $Pr_t$ the turbulent Prandtl number, and $z_{0M}$, $z_{0H}$, $z_{0E}$ the roughness lengths for momentum, heat and moisture, respectively.

As suggested by Emanuel (1986) and Emanuel (1995), the intensity of tropical cyclones in atmospheric models strongly depends on the ratio of the exchange coefficients for enthalpy $C_k$ and momentum $C_D$. Here, $C_k$ can be obtained from the enthalpy flux $H_k$, which corresponds to specific enthalpy $k = c_p\theta + L_v q$, according to

$$H_k = \rho_a C_k U_L (k_0 - k_L). \tag{7}$$

Emanuel (1995) concluded that $C_k/C_D$ should be in the range 1.2–1.5, in order to model realistic tropical cyclone intensity, with 0.75 as a lower bound. Measurement data in the tropical cyclone boundary layer are still limited. Hence, there is still a rather large gap in understanding what happens at the sea surface. The behaviour of $C_H$, $C_E$ and $C_D$ are still studied. Analyses by Powell et al. (2003) of measurements with Global Positioning System (GPS) drop sondes show that the magnitude of $C_D$ starts to decrease from a wind speed of 30–40 m s$^{-1}$. Analyses of Holthuijsen et al. (2012), which show a dependence of $C_D$ to the different azimuthal angles in tropical cyclones, confirm in general the findings by Powell. Analyses by Jarosz et al. (2007) of buoy data measuring ocean currents show a similar behaviour for $C_D$ at very high wind speeds.

Different reasons for a possible reduction in the drag at high wind speeds have been suggested. Makin (2005), Bye and Jenkins (2006), Kudryavtsev (2006) and Rastigejev et al. (2011), speculate on the impact of spray droplets, as suspended particles, on atmospheric stratification. Within the frame of this 'stratification mechanism', however, the observed suppression of the sea drag can only be attained at an unrealistically large volume concentration ($\mathscr{O}(10^{-3})$ or even larger) of sea spray (e.g. Kudryavtsev (2006), Rastigejev et al. (2011)). Another cause is the direct impact of droplets on the momentum balance through 'spray stress' (or 'spray force'). As illustrated by Andreas (2004) and Kudryavtsev and Makin (2011) this should be a more efficient way than through stratification. In addition, foam and streaks (of foam and spray) are suggested as a cause as well (Powell et al. (2003), Holthuijsen et al. (2012)).

The impact of reduced drag at high wind speeds on modelling tropical cyclone intensity has been examined by Zweers et al. (2010). They simulate severe tropical cyclones with reduced $C_D$, while their monotonically increasing drag coefficient results in significantly lower wind speeds.

The available measurement data are also questioned. Smith and Montgomery (2014) question the assumption of a logarithmic wind profile at very high wind speeds that Powell uses for computing $C_D$. Moon et al. (2007), who propose an empirical $C_D$ based on previous numerical calculations with a coupled atmospheric-wave model, suggest that $C_D$ saturates at very high wind speeds. Their $C_D$ shows poor agreement with available measurement data. In general, Richter and Sullivan (2013) mention that only the turbulent stress above the sea surface is measured at very high wind speeds. Therefore, the turbulent drag would decrease, but the atmospheric drag starts to saturate.

Analyses of CBLAST data by Zhang et al. (2008) indicate that with "95% confidence" $C_H = C_E = C_k$ for wind speeds up to about 30 m s$^{-1}$. Using momentum flux measurements by French et al. (2007), they show that $< C_k/C_D >$=0.63, which is smaller than Emanuel's critical value of 0.75. Zhang et al. (2008) speculate that "the enthalpy flux into the hurricane boundary layer may have come from sources other than air-sea turbulent fluxes, or the Emanuel model assumptions should be revisited". Andreas (2011) mentions Emanuel's criterion is often misinterpreted: for tropical cyclone wind speeds $C_k/C_D$ ¿ 0.75, while its value might be smaller for lower wind speeds. While $C_k/C_D$ obtained from Andreas' algorithm shows agreement with Emanuel's favourable range of 1.2–1.5 in the high wind speed regime, $C_D$ keeps increasing monotonically with increasing wind speed, which is not what availabbe measurement data suggest: $C_D$ would start to reduce.

In the present study the impact of an air-sea exchange parametrization on modelling tropical cyclone intensity with an NWP model is examined. The

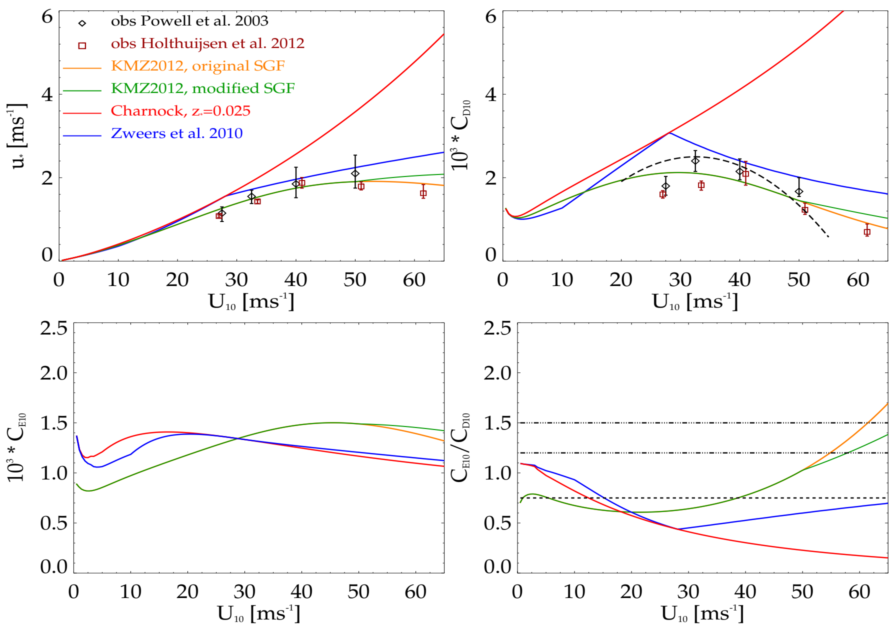


*Figure 1.* Friction velocity $u_*$ (top left), 10 m drag coefficient $C_{D10}$ (top right), 10 m humidity exchange coefficient $C_{E10}$ (bottom left), and ratio $C_{E10}/C_{D10}$ (bottom right) versus 10 m wind speed $U_{10}$, according to different parametrizations (*see legend*; KMZ2012 refers to Kudryavtsev et al. (2012)); observations are in black (Powell et al., 2003) and brown (Holthuijsen et al., 2012); the dashed black curve for $C_{D10}$ is from Jarosz et al. (2007); range for $C_{E10}/C_{D10}$ suggested by Emanuel (1995) in black dashed-dotted lines with the black dashed line is the 0.75 lower bound.

parametrization is based on the model by Kudryavtsev and Makin (2011), in which the impact of spray droplets on the momentum flux is directly accounted for in the way of a spray stress (‘vortex force’). The model produces a decrease in the drag at very high wind speeds, as in agreement with the available measurements in these conditions. In the model, further explored by Kudryavtsev et al. (2012), the air-sea enthalpy transfer is enhanced. Hence, the parametrized reduced $C_D$ and enhanced $C_k$ from Kudryavtsev et al. (2012) are tested. The parametrizations are used in the NWP model HiRLAM[1] (High Resolution Limited Area Model) to simulate the tropical cyclones Ivan (2004) and Katrina (2005) in the Gulf of Mexico. The reduction in the drag, and also the increase in $C_k$, is based on a different physical mechanism than in e.g. Zweers et al. (2010).

The outline is as follows. In Section 2 we present the parametrizations for $C_k$ and $C_D$ that are tested and explain on the simulations performed with the NWP model. Results are presented in Section 3, followed by a concluding section.

[1] see http://www.hirlam.org/ for more information on HiRLAM

## 2. Methods

A different air-sea exchange scheme is applied than Zweers et al. (2010) have used. They compute $C_H$ and $C_E$ in HiRLAM with formulations for $z_{0H}$ and $z_{0E}$ based on extrapolation of the empirical formulations proposed by Garratt (1992):

$$\ln \frac{z_{0M}}{z_{0H}} = 2.48 Re_*^{1/4} - 2, \tag{8}$$

$$\ln \frac{z_{0M}}{z_{0E}} = \ln \frac{z_{0M}}{z_{0H}} - 0.2 Re_*^{1/4}, \tag{9}$$

with $Re_* = z_{0M} u_* / \nu$ the Reynolds number, with $u_* = \sqrt{\tau/\rho_a}$ the friction velocity and $\nu$ the molecular kinematic viscosity of air. The behaviour of and the values for the exchange coefficients $C_H$ and $C_E$ are quite similar with (8) and (9). A change in the drag comes with changes in the exchange of heat and moisture. This is in fact a natural consequence (see Section 2.1). But with (8) and (9) these changes are not dramatic (see Figure 1). The drag parametrization by Zweers et al. (2010) together with (8) and (9) gives values for $C_E/C_D$ (or $C_H/C_D$) that are below Emanuel's critical value of 0.75 for all (high) wind speeds (see Figure 1).

In the present study (8) and (9) are not used. Here, $C_D$ is reduced more severely and $C_k$ is enhanced more strongly. A description is given in the next paragraph.

### 2.1. Drag and enthalpy exchange coefficient

At very high wind speeds the airflow near the air-sea interface is heavily loaded with spray droplets. In these conditions, the use of traditional equations for the MABL dynamics is not justified. Therefore, Kudryavtsev and Makin (2011) (hereinafter KM2011) derived equations in which spray effects on the MABL dynamics are directly accounted for. KM2011 used a modified form of the mass and momentum conservation equations that take into account the injection of droplets into the airflow at the height of breaking wave crests. In the momentum conservation equation they included the volume source of the droplet momentum. This is the rate of injection of momentum of the droplets, that are torn off from breaking wave crests, in a unit volume of air at height $z$ (see eq. (8) in KM2011).

Kudryavtsev et al. (2012) (hereinafter KMZ2012) present a simplified model (parametrizations) for the MABL that is based on the model by KM2011. In the present study the impact of these parametrizations on modelling tropical cyclone intensity is investigated. For a full and detailed description of the model and the parametrizations we refer to the original papers; here, we briefly present and explain the parametrizations.

The simplified model by KMZ2012 is a two layer model in which the MABL consists of a thin layer adjacent to the sea surface in which spray droplets are generated from breaking wave crests – the spray generation layer (SGL) – and the core of the MABL above the SGL. The spectral distribution of wave breaking parameters (e.g. length of breaking fronts, or distribution of white caps over wavenumbers) rapidly grows towards high wavenumbers, as suggested by Phillips (1985). This is why in the KMZ2012 model the SGL (associated with spume droplets generation) is a relatively shallow layer with depth $d$ that is proportional to the inverse wavenumber of shortest breaking waves $k_b$ that carry white caps ($k_b \simeq 10$ rad m$^{-1}$). These rather short breaking waves are strongly modulated by large dominant wind waves. Hence, the production of spume droplets largely takes place on top of the dominant waves and on their crests, as is discussed in Kudryavtsev and Makin (2009).

In the model by KMZ2012 the droplets act on the airflow dynamics through the 'spray force'. The interpretation of this mechanism is the following. At moderate and high wind speeds the surface stress is fully supported by the surface form drag. At higher wind speeds the crests of breaking waves can be disrupted aerodynamically (e.g. by Kelvin-Helmholtz type instability) and they are pulverized to droplets. The physical meaning of the mechanism suggested in KM2011 is that the form drag, initially supported by the wave crests, is transferred to the droplets once these crests have been subjected to the aerodynamical disruption and fragmentation. The droplets are produced at the height of the breaking crests.

Once generated, the droplets are embedded in the airflow, following the streamlines of the dominant waves. The droplets stay relatively close to the sea surface, as the airflow does not separate from the dominant waves. While falling back to the sea surface, the droplets transfer momentum to the airflow. Hence, the turbulent stress in the SGL increases and droplets will accelerate the airflow near the surface. As a result of this acceleration, in the core of the MABL (above the SGL) the vertical wind shear reduces and turbulence will be suppressed. This results in a reduction in the drag coefficient in the core of the MABL. Thus, while inside the SGL the turbulent mixing of momentum is enhanced, it is suppressed above the SGL. And the combination of aerodynamic disruption of wave crests (that initially provided the form drag) and injection of droplets into the airflow represents a coupled process that results into acceleration of the airflow and suppression of the sea drag at high wind speeds.

Furthermore, due to the enhanced turbulent mixing inside the SGL, spray droplets enhance the turbulent exchange of heat and moisture. This leads to reduced vertical gradients of these quantities in the SGL. This can be understood by considering the small upward shift of the surface values of temperature and humidity to the depth of the SGL. As a result, the vertical gradients of humidity and heat are enhanced above the SGL. Hence, through this mechanism spray droplets enhance the exchange of latent and sensible heat in the core of the MABL in this model.

A logarithmic wind profile is assumed. Then, the drag coefficient at height $h$ can be written as

$$C_{Dh} = \kappa^2[\ln(h/Z_{0M})]^{-2}. \tag{10}$$

Following KMZ2012, we write the effective roughness length for momentum, $Z_{0M}$, as

$$Z_{0M} = z_0 \exp(-\Delta_m). \tag{11}$$

For near-surface wind speeds up to approximately 30 m s$^{-1}$ the effective roughness length is equal to the aerodynamic roughness, $z_0$, which is computed with Charnock's relation ((Charnock, 1955))

$$z_0 = z_* u_*^2/g, \tag{12}$$

in which $z_*$ is the dimensionless aerodynamic roughness length, or Charnock's parameter. With further increase in the wind speed $Z_{0M}$ decreases due to the increasing impact of spray droplets on the MABL dynamics. This is represented by the function $\Delta_m$, which is determined by the spray generation function (SGF). The SGF, denoted as $\hat{F}_s(u_*, z)$, determines the vertical profiles of spray and the wind. We assume that inside the SGL, the vertical distribution of $\hat{F}_s(u_*, z)$ can be approximated by the spray flux at the surface, i.e. $\hat{F}_s(u_*, z) = F_s^0(u_*)$. KMZ2012 parametrize the SGF as

$$F_s^0 = c_s u_* (U_{10}/c_b)^3, \tag{13}$$

with $U_{10}$ the wind speed at 10 m height, $c_s$ a constant ($\simeq 10^{-9}$) and $c_b$ the phase velocity of the shortest breaking waves that produce droplets ($c_b \simeq$ 0.63 m s$^{-1}$). Note that KM2011 compute the total length of wave breaking fronts, which is cubic in the wind speed. This quantity is included here in terms of $(U_{10}/c_b)^3$. The function $\Delta_m$ then depends on $F_s^0$ through

$$\Delta_m = c_m (U_{10}/c_b)^3 = c_m F_s^0/u_* c_s, \tag{14}$$

with $c_m \simeq 10^{-6}$ a function that weakly depends on $u_*$.

With formulation (13) the SGF rapidly grows with increasing wind speed. KMZ2012 show there is fair agreement between (13) and conditions observed in laboratory experiments at very high wind speeds (see their Figure 1). As Kudryavtsev (2006) speculates, though, at very high wind speeds the total length of wave breaking fronts should be saturated. This seems plausible, as the sea surface must have a limited configuration in those extreme conditions, when almost each individual wave crest will break. This implies that formulation (13) overestimates the production of spray droplets in these conditions.

Moreover, a visual indicator of wave breaking is white capping. Holthuijsen et al. (2012) report on the generation of streaks of foam and spray at very high wind speeds. These streaks may originate from white caps and may also merge with these white caps. Their analysis and photographic evidence show that the white cap coverage remains almost constant from a wind speed of about 40 m $s^{-1}$. Formulation (13) leads to a decrease in the friction velocity and very low values for $C_D$ (see Figure 1, orange lines) at very high wind speeds; $C_D$ seems to drop even below the viscous drag. This implies that almost all the white caps are pulverized into droplets. That is in contradiction with the observed convergence in the degree of white capping mentioned above.

In the present study the rapid increase in the droplet production and the very strong drop in $C_D$ at very high wind speeds is prevented through reduction of the term $(U_{10}/c_b)^3$ from a certain threshold wind speed $U_T$. Hence, while relation (13) applies for $U_{10} \leq U_T$, for $U_{10} > U_T$ we propose the following relation:

$$F_s^0 = c_s u_* (U_{10}/c_b)^2 (U_T/c_b). \tag{15}$$

The precise value for $U_T$ may be a subject of future studies; here, we use $U_T =$ 50 m $s^{-1}$. With this modified SGF the friction velocity does not decrease anymore at very high wind speeds, but it starts to saturate with increasing wind speed (see Figure 1, green lines). Then, the dramatic drop in $C_D$ at very high wind speeds and the rapid growth in $C_k/C_D$ is avoided (see Figure 1).

The exchange coefficients for humidity $C_E$ and heat $C_H$ are assumed equal. This is based on extending the result by Zhang et al. (2008) to the hurricane wind speed regime (see Section 1). Focusing on the humidity exchange coefficient $C_E$, we note that the formulations also apply for $C_H$. The humidity exchange coefficient at height $z = h$, denoted as $C_{Eh}$, is computed as

$$C_{Eh} = \kappa^2 [Pr_t \ln(h/Z_{0E}) \ln(h/Z_{0M})]^{-1}, \tag{16}$$

which can also be written as

$$C_{Eh} = \frac{\kappa}{Pr_t \ln(h/Z_{0E})} \sqrt{C_{Dh}}. \tag{17}$$

The effective roughness lengths for humidity $Z_{0E}$ (and temperature $Z_{0H}$) used in this study do not take into account the impact of evaporation from droplets on the moisture and heat balance. The droplets, which originate from short breaking waves on top of large dominant waves (see Section 2.1), stay close to the sea surface, follow the streamlines of these large dominant waves and do not evaporate inside the SGL.

The effective roughness length for humidity, denoted as $Z_{0E}$, is parametrized as

$$Z_{0E} = z_{0E} (d/z_{0E})^{(1-\ln(1+\Delta_E)/\Delta_E)}. \tag{18}$$

The value for $Z_{0E}$ equals a constant background roughness length $z_{0E}$ in case there are no spray droplets, and grows with increasing wind speed towards a value close to the depth of the SGL $d$ ($\simeq$ 0.1 m), as a result of enhanced turbulent mixing inside the SGL. This is represented in the parametrizations by $\Delta_E$, which in a similar way depends on $F_s^0$ as $\Delta_m$, according to

$$\Delta_E = c_E(U_{10}/c_b)^3 = c_E F_s^0/u_* c_s, \tag{19}$$

with $c_E \simeq 4.5 \times 10^{-6}$.

The physical explanation for the parametrization in (18) is the following. Spray droplets enhance the turbulent exchange of heat and moisture in the SGL. This effect may seem to be inconsistent with suppression of the momentum exchange. However, such an effect has a clear physical meaning. Within the frame of the wind-over-waves coupling theory, the heat exchange coefficient is directly linked to the coupling parameter, which is the ratio of the surface stress supported by the form drag to the total stress in the marine atmospheric boundary layer (see e.g. Makin (1999)). At moderate and high wind speeds the coupling parameter tends towards the value of 1. This has to result in a rapid decrease of the roughness length for temperature ($z_{0H}$) and humidity ($z_{0E}$). Empirical formulations such as e.g. (8) and (9) do describe this effect: the values for $z_{0E}$ and $z_{0H}$ decrease with increasing wind speed. When crests of breaking waves are pulverized to droplets at very high wind speeds, the surface form drag is suppressed. This should inevitably result into enhancement of the heat transfer. Hence, both mechanisms – the suppression of the momentum exchange coefficient and the enhancement of the heat (and humidity) exchange coefficient due to the generation of spray droplets – are taken into account in the parametrizations by KMZ2012 that are tested here.

The humidity exchange coefficient depends on both $Z_{0M}$ and $Z_{0E}$ (see eq. (16)). Due to the rapid growth of $Z_{0E}$, $C_{E10}$ still increases for wind speeds larger than 30 m s$^{-1}$. For wind speeds larger than approximately 40 m s$^{-1}$ the strong reduction in $Z_{0M}$ starts to dominate the increase in $Z_{0E}$. Then, $C_{E10}$ starts to decrease with further increase in the wind speed (see eq. (16)).

Figure 1 shows the friction velocity $u_*$, 10 m drag coefficient $C_{D10}$, 10 m humidity exchange coefficient $C_{E10}$ and the ratio $C_{E10}/C_{D10}$ (or $C_{k10}/C_{D10}$) as a function of the 10 m wind speed $U_{10}$. The reduction in the drag shows good agreement with the reduction in $C_D$ indicated by observational data. The friction velocity levels off at very high wind speeds. The value for $C_{E10}/C_{D10}$ is in fair agreement with the experimental results presented by Zhang et al. (2008) (see Section 1) and is in good agreement with the range of 1.2–1.5 proposed by Emanuel (1995) for higher wind speeds.

## 2.2. Simulations

The tropical cyclones Ivan (2004) and Katrina (2005) in the Gulf of Mexico were simulated with the NWP model HiRLAM. The model grid has a horizontal resolution of $0.05^\circ$ and 40 levels in the vertical. The height of the lowest model level is approximately 30 metres. The wind speed at 10 m height is obtained through the logarithmic wind profile with eq. (11). HiRLAM uses a semi-Lagrangian discretization scheme for solving the primitive equations on all vertical levels, in which a dynamics time step of 2 minutes was used. The model uses a convective parametrization scheme. Lateral boundary conditions for the HiRLAM simulations were taken from the European Centre (ECMWF) model archive. The same model version and model grid were used as by Zweers et al. (2010), which allows for a fair comparison with results from that study.

Forecasts were performed with HiRLAM with model analyses every next six hours. These 6-hourly analyses are based on every previous forecast and the assimilation of observational data in a 3D-VAR assimilation scheme. These simulations were performed for the period 25–30 August 2005 (Katrina) and 11–17 September 2004 (Ivan).

For $z_*$ a value of 0.014 was used. It was verified that the values of $C_D$ and $C_k$ at very high wind speeds are not sensitive to this value (*not shown here*). Therefore, the results are not expected to be very sensitive to the choice for $z_*$. Next to this, we used a value of 1.0 for the turbulent Prandtl number $Pr_t$. This parameter, which describes the difference in turbulent transfers of momentum and sensible heat, shows to be particularly dependent on stability. Grachev et al. (2007) refer to several different studies on $Pr_t$ that indicate that its value should be around 1.0. Grachev et al. (2007) mention that 'on average' the value for $Pr_t$ decreases with increasing stability and that $Pr_t < 1.0$ in the very stable case. Based on their analysis of the SHEBA (Surface Heat Budget of the Arctic Ocean) experiment, they also conclude that this is "not a general result" for the MABL and they show that a value of 1.0 is a proper choice. Hence, for simplicity we use $Pr_t = 1.0$.

Finally, an important aspect of the model setup is how sea surface temperatures (SSTs) are prescribed to the NWP model. In the HiRLAM surface analyses SST fields from the ECMWF model archive are used. The default approach in HiRLAM, as in many other NWP models, is that during the forecasts these SSTs are used. In other words, the SSTs are fixed during the model forecasts and new SSTs are applied every next six hours. This approach is here the reference scenario, referred to as 'default SSTs'.

The approach described above is generally accepted a valid method for forecasting 'normal' weather conditions, as SSTs then usually do not rapidly change on local and small scales. Tropical cyclones cause a reduction in SSTs due to upwelling, currents generated in the ocean upper layer and because of heat uptake from the ocean. This reduction considerably affects the air-sea enthalpy flux. Cione and Uhlhorn (2003) mention differences between inner-core and ambient SSTs of 0–2 K and further state that SST changes on the order of 1 K lead to surface enthalpy flux changes of 40% or more with respect to the scenario in which SSTs do not change. D'Asaro et al. (2007) come to a similar conclusion: studying Category 4 tropical cyclone Frances (2004), they find a decrease in the strongest winds by about 5 m s$^{-1}$ to a measured wind speed of 60 m s$^{-1}$ due to a 0.4 K reduction in the SSTs under the storm core.

Higher wind speeds would intuitively result in tropical cyclone forecasts when the SSTs are overestimated. A 'compensation' for this increase in the wind speed will result if in the NWP model $C_H$ is not very large (see eq. (2)). Hence, an overestimation of the SSTs could partially explain why Zweers et al. (2010) obtain such positive results with $C_H/C_D$ below Emanuel's critical value of 0.75, despite the use of reduced $C_D$.

In the default SSTs approach described above new SSTs result from the surface analyses every next six hours. In these analyses the NWP model is not able to resolve the small scale SST reductions. For reasons described above, a second scenario is investigated in which SSTs under the tropical cyclones are subject to a simplified parametrized cooling. The aim was not to model SST reductions as accurate as possible, but simply to prevent an overestimation in the SSTs during the forecasts and to examine the impact of the parametrizations in the case that such a simplified SST cooling is applied. To that end, an SST reduction was imposed in a sea grid point, if in that grid point the 10 m wind speed is in the hurricane wind speed regime, i.e. if $U_{10} > 33$ m s$^{-1}$. Considering a total cooling of 2.0 K under the tropical cyclones (see Cione and Uhlhorn (2003)), a reduction $\delta T_0$ is applied every time step in the model simulation. Estimating the translation speed of the tropical cyclones from the experiments with default SSTs, the time interval in which they pass a model grid point was estimated to 12 hours. Therefore, we use $\delta T_0$ = (2 K/12 hrs) = $4.63 \times 10^{-5}$K s$^{-1}$. Thus, in this approach, referred to as 'reduced SSTs', the SSTs are determined by the ECMWF fields, the HiRLAM surface analyses, and additionally by the imposed SST reduction.

## 3. Results

### 3.1. Default SSTs

Figure 2 shows the modelled maximum wind speed ($U_{10_{max}}$) and the central pressure, and observed wind speed and pressure as described in the National Hurricane Center (NHC) tropical cyclone reports Stewart (2004) (Ivan) and Knabb et al. (2005) (Katrina). The NWP model overestimates the intensity

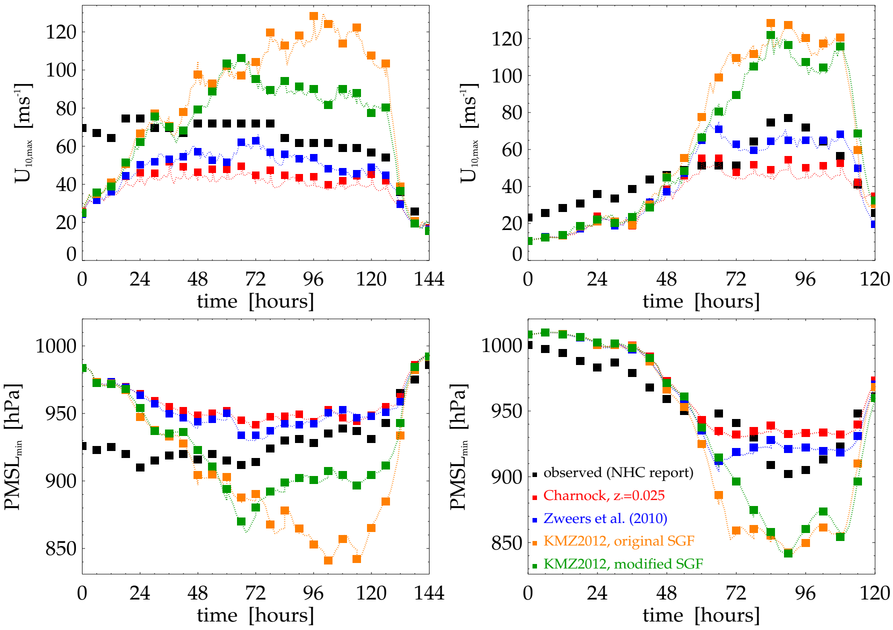


*Figure 2.* Maximum 10 m wind speed (top) and central pressure (bottom) for tropical cyclones Ivan (left) and Katrina (right). Results obtained with HiRLAM with the new parametrizations in orange (SGF from KMZ2012) and green (modified SGF, *see text*). Results from Zweers et al. (2010) are in red (Charnock's relation, $z_* = 0.025$) and dark blue (their $C_{D_{10}}$). $t = 0$ corresponds to 11 September 2004 0000 UTC (Ivan) and 25 August 2005 0000 UTC (Katrina). HiRLAM analyses are indicated by squares; observations from Stewart (2004) and Knabb et al. (2005) are in black.

with the drag and enthalpy exchange coefficients for both tropical cyclones (Figure 2, green lines). During Ivan wind speeds of about 100 m s$^{-1}$ are modelled, compared to observed wind speeds of $\simeq$ 70–75 m s$^{-1}$. The central pressure is approximately 870 hPa, compared to the measured value of around 915 hPa. The intensity drops to a rather constant 80–90 m s$^{-1}$ and central pressure of 900 hPa during the simulation, while observed wind speeds were 60–70 m s$^{-1}$ and the central pressure increased from about 925 to 935 hPa.

The differences between observed and modelled intensities are larger for Katrina than for Ivan. Knabb et al. (2005) report wind speeds up to 77 m s$^{-1}$, while the modelled wind speeds reach unrealistic values up to about 120 m s$^{-1}$. This is mainly due to the fact that the model does not reproduce the observed collapse of the eyewall (see Figure 2, *time* $\simeq$48-72 hrs). Rather similar behaviour was also found by Zweers et al. (2010). On the other hand, keeping the SSTs constant during the forecasts also significantly contributes to the fact that wind speeds larger than 100 m s$^{-1}$ are obtained (see also Section 3.2).

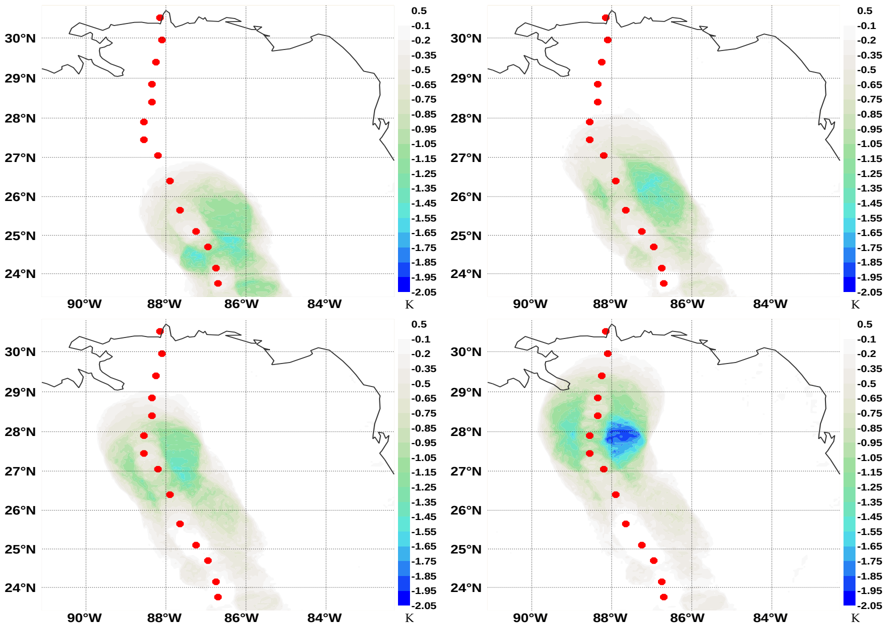


*Figure 3.* Difference between the SSTs that are kept constant during the forecasts and SSTs that are reduced underneath the cyclones during the forecasts, for tropical cyclone Ivan, at 15 September 2004 at 0600 UTC (top left), 1200 UTC (top right), 1800 UTC (bottom left) and at 16 September 2004 at 0000 UTC (bottom right). The data are from +6h forecasts.

As anticipated in advance, the use of the original SGF from KMZ2012 (see eq. (13)) in the new parametrizations – which leads to a dramatic drop in $C_D$ and decrease in $u_*$ above 50 m s$^{-1}$ – leads to even higher wind speeds in the eyewall up to 130 m s$^{-1}$ (Figure 2, orange lines). Especially during Ivan the differences are very large: the highest wind speeds are larger by approximately 40 m s$^{-1}$. This illustrates the significance of the SGF in the air-sea exchange parametrizations and shows that a 'run-away' effect may arise in case of a strong increase in the value of $C_k/C_D$ at very high wind speeds (see Figure 1, orange lines).

## 3.2. Reduced SSTs

Figure 3 shows the differences between the perturbed SSTs and the unperturbed SSTs at a fixed time during Ivan. As explained in Section 2, in the simulations with reduced SSTs, these SSTs follow from the archived ECMWF SST fields, the HiRLAM surface analyses and the reduction imposed per time step, if the wind speed exceeds the imposed criterion (see Section 2.2). Hence, the SST reductions shown in Figure 3 are almost, but not exactly, equal to the reduction applied in the model. The temperature difference has quite an irregular spatial shape, which is due to several reasons. First, the track of the

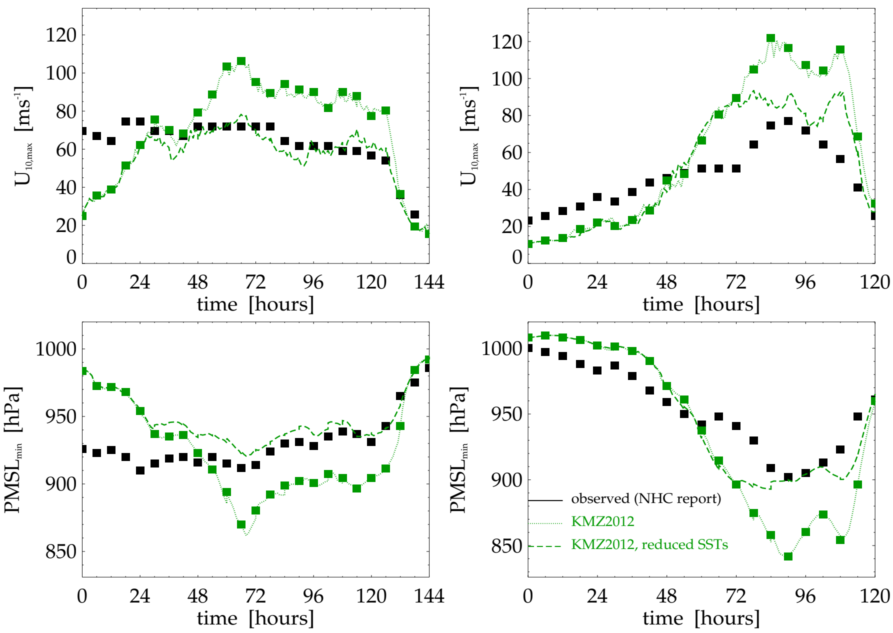


*Figure 4.* As in Figure 2, but then for the new parametrizations only, with unperturbed SSTs (solid green) and reduced SSTs (dashed green).

cyclones is not a straight line. The translation speed is not exactly constant during the simulation. Additionally, in model grid points farther away from the eye the temperature reduction is smaller than in those locations right under the eyewall. Moreover, the wind field is not symmetric.

Figure 4 shows the modelled wind speeds and pressure when SSTs are reduced in the model. These results are much more realistic. The model produces the observed central pressure of 902 hPa during Katrina, while in the simulations with default SSTs a value of 840 hPa was obtained. This evidently illustrates the impact of the SST modification. During the Category 5 stage the modelled wind speeds vary from 70 m $s^{-1}$ up to 90 m $s^{-1}$. This is about 30 m $s^{-1}$ lower than the result obtained without the SST modification.

The modelled wind speeds during Ivan show very good agreement with observed values. Occasionally, the obtained wind speeds are even lower than their observed equivalents. Considering the first day in the simulations as spin-up, we see that on average the model gives an almost reproduction of the observed wind speeds. The central pressure is throughout the entire simulation slightly too high by 5–10 hPa. The difference with the central pressure obtained with (default) unperturbed SSTs varies by about 30–40 hPa.

## 4. Discussion and conclusions

The impact of new parametrizations for drag and air-sea enthalpy exchange on the modelling of tropical cyclone intensity has been examined. These parametrizations are based on a model that directly accounts for the impact of spray droplets on the marine atmospheric boundary layer momentum balance. This model produces a reduction in $C_D$ at very high wind speeds as in agreement with what measurements indicate. The value for $C_k/C_D$ is in agreement with observational data for wind speeds up to 30 m s$^{-1}$. For higher wind speeds the value for $C_k/C_D$ is in the favourable range of 1.2–1.5 suggested by Emanuel (1995).

It was shown that the use of the parametrizations in an NWP model leads to an overestimation of the intensity of the tropical cyclones Ivan (2004) and Katrina (2005). An important model aspect that relates to this result, which has been addressed in this paper, is the way in which SSTs are prescribed and used in the NWP model. Due to strong winds, upwelling and ocean currents, and the heat uptake by tropical cyclones, the SSTs below tropical cyclones should reduce. The NWP model that was used, does not resolve these small scale reductions. In the traditional approach, in which SSTs are kept constant during forecasts, the winds are overestimated with the model. The application of a rather simple reduction in the SSTs underneath the cyclones during the model forecast, of which the magnitude is based on results in other scientific literature, leads to much better and realistic results. Even good agreement is found with observed wind speeds and central pressure. This indicates that the parametrizations may give realistic results for other tropical cyclones as well. Also, the result supports that $C_k/C_D \simeq$ 1.2–1.5 at high wind speeds.

The sensitivity shown here of eyewall wind speed and central pressure to the SSTs is in line with results by Cione and Uhlhorn (2003). A decrease in the SSTs of 1–2 K (see Figure 3) during Ivan leads to a drop in the eyewall wind speed from about 90 m s$^{-1}$ to 65 m s$^{-1}$ (see Figure 4). A similar result is found for Katrina. Regarding this sensitivity to the SSTs, and given the fact that the result significantly improves when the SSTs are decreased, the parametrizations are recommended for the modelling of tropical cyclones with NWP-ocean coupled models.

The results are expected not to be very sensitive to the choice for Charnock's parameter, $z_*$. At very high wind speeds $C_k/C_D$ is not sensitive to $z_*$ (*not shown here, but verified*). Hence, results obtained with different $z_*$ are expected not to be significantly different than shown here. The intensity of the cyclones is expected to be rather sensitive to the value for the turbulent Prandtl number, $Pr_t$, though. While the value for $Pr_t$ at very high wind speeds is still rather uncertain, the impact of the parameter is quite large though (see $Pr_t$ in eq. (17)). Using a value larger than 1.0 will result into even better agreement with observed conditions than found here, although such a value may be considered as rather unrealistic. Application of a value smaller than 1.0 is realistic, on the contrary, although this will result in (significantly) higher wind speeds.

Compared to Zweers et al. (2010) – who reduced $C_D$ at high wind speeds and obtained maximum wind speeds of about 60 m s$^{-1}$ (Ivan) and 70 m s$^{-1}$ (Katrina) – much higher wind speeds are obtained in the present study. This is due to the combined effect of the exchange coefficients (see Figure 1). The behaviour of $C_k$ is much different and $C_D$ is larger in Zweers et al. (2010). In the present study, in the end $C_k/C_D$ is almost twice as large at 60 m s$^{-1}$. This illustrates the importance of the value for $C_k/C_D$ and not simply $C_k$ or $C_D$ individually, and explains the much higher wind speeds obtained in this study. Even with reduced SSTs during the forecasts, which Zweers et al. (2010) did not consider in their work, the obtained wind speeds are still higher, although they are realistic.

Another study about spray mediated air-sea enthalpy and momentum transfer is Bao et al. (2011). Their approach is based on a physical mechanism (stratification) different from the direct impact of spray on the momentum balance. Also different are their use of constant SSTs throughout the entire model domain ($T$ = 302.16 K) and the simulation of an idealized tropical cyclone with their NWP model. The behaviour of $C_k/C_D$ does show similarities with our $C_k/C_D$, although the values are slightly different: $C_k/C_D$=1.6 at 60 m s$^{-1}$, while our $C_k/C_D$ is nearly 1.3. They obtain wind speeds up to 90 m s$^{-1}$. Our simulations with unperturbed SSTs result in even higher wind speeds. This is partly due to higher SSTs. Our simulations with reduced SSTs result into wind speeds lower than those obtained by Bao et al. (2011). While the SSTs are now comparable, our $C_k/C_D$ is still slightly smaller. The wind speeds during Katrina are still higher than observed. To a great extent this is due to the fact that our NWP model is not capable of resolving the collapse of the eyewall at 27 August 2005 (see Figure 4, *time*=48–72hrs) properly. This could be due to the fact that our NWP model uses a convective parametrization scheme. Although very good results for Ivan are obtained, it would be interesting to see the impact of the parametrizations in a non-hydrostatic NWP model, as in Bao et al. (2011).

Finally, tropical cyclone intensity in NWP models is also sensitive to the formulation used for the spray generation function (SGF). In the present study, we anticipated that the drag coefficient becomes too small when the SGF by Kudryavtsev et al. (2012) is used. It was shown that this results into unrealistically high wind speeds and extremely low central pressure. Our modification in the SGF is a step towards the concept that the length of wave breaking fronts and the amount of white capping should saturate at very high wind speeds, as is suggested in both theoretical and experimental studies (see Section 2.1). Although it is a step in the good direction, possibly a stronger limitation to the droplet production is required. To that end, better understanding of wave breaking at very high wind speeds is required. This could lead to better representation of wave breaking statistics and spray droplet production from individual breaking crests at very high wind speeds in models. This should in the end lead to better tropical cyclone forecasts. Eventually, with climate models indicating that future tropical cyclones will become more severe Knutson et al. (2010), it becomes more significant to aim at better understanding (and modelling) of the processes that dominate the dynamics at the air-sea interface in tropical cyclone conditions.

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