An Analytical Framework for Rapid Estimate of Rain Rate during Tropical Cyclones

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Abstract: An analytical framework for rapid estimate of rain rate during a translating tropical cyclone was proposed in this study. The efficient analysis framework for rain field is based on the observation that rain-induced momentum flux at Earth’s surface cannot be ignored. The total surface stress results mainly from momentum flux contributions of wind and rain. A height-resolving wind field was utilized during the model construction leading to a linear, analytical solution of the surface rain rate. The obtained rain rate model explicitly depends on parameters for a typical tropical cyclone wind field simulation, namely storm location, approach angle, translation speed, radius of maximum wind, pressure profile, surface drag coefficient, and turbulent diffusivity. Hence, it could be readily implemented into state-of-the-art tropical cyclone risk assessment using the Monte Carlo technique. The rainbands in the proposed methodology were simulated using a local perturbation scheme. Sensitivity analysis of the rainfall field to the abovementioned parameters was comprehensively conducted. The results generated by the present analytical framework for rapid estimate of rain rate during tropical cyclones are consistent with field measurements.

Keywords: Tropical cyclone, rain rate, wind field simulation, rain-induced stress.
1. Introduction

Tropical cyclones are responsible for the substantial part of natural hazard-induced economic and life losses through high winds, torrential rain and wind-driven storm surge. Among these, the rainfall-induced inland flooding contributes to a significant portion of the tropical cyclone related damages (e.g., Landsea 2000; Rappaport 2000). Therefore, the rain field simulation inside the tropical cyclone has attracted interest of a number of researchers for a better rainfall hazard assessment. While there have been considerable advances in improving the simulation accuracy of tropical cyclone rain field based on high-fidelity numerical weather prediction models, they are not practical for risk assessment due to their high computational demands. Usually, the rainfall distribution can be efficiently characterized based on probabilistic, parametric or physically-based schemes.

The probabilistic models give good insights on the exceedance rate of specific rainfall intensities, and are often used to predict the extreme rain rates. The development of this type of models usually suffer from a lack of a large number of historical data that are needed to fit the selected distributions. In addition, they are generally unable to represent the most important physics governing the rain field inside the tropical cyclone (e.g., sea surface temperature, moisture distribution, vertical wind shear, hurricane intensity and translational velocity). Actually, no physical justification has been provided for the use of the popular distributions such as lognormal, mixed-exponential and Gamma curves to fit the data (e.g., Woolhiser and Roldan 1982; Groisman et al. 1999; Wilson and Toumi 2005).

The construction of the parametric models also requires a huge amount of rain field measurements. Recently, the Tropical Rainfall Measuring Mission (TRMM) (Huffman et al.,
a joint satellite mission of the National Aeronautics and Space Administration (NASA) and the Japan Aerospace Exploration Agency (JAXA), has released a significant amount of tropical cyclone rainfall data. The goal of TRMM is to provide good estimates of global precipitation using satellite observations. TRMM contains several instruments, namely the TRMM Microwave Imager (TMI), the Precipitation Radar (PR), the Visible Infrared Scanner (VIRS), the Lightning Imaging Sensor (LIS), and the Clouds and Earth's Radiant Energy System (CERES). Details of the TRMM instruments are given in Kummerow et al. (1998). Several empirical models have been developed based on the TRMM database. For example, Lonfat et al. (2004) acquired the spatial distribution of the rain field over the ocean using the TMI data from 1998 to 2000. The rain rates were found to be heavily dependent on the sustained surface wind speed. The rain rate achieved the maximum value near the radius of maximum winds and then decayed exponentially. According to the hurricane intensities grouped into three categories, i.e., tropical storms, category 1-2 tropical cyclones, and category 3-5 tropical cyclones, different radial variations of the rain rate were obtained. Based on the findings from Lonfat et al. (2004) together with the surface rain gauge data, Tuleya et al. (2007) proposed the Rainfall Climatology and Persistence (R-CLIPER) model. In this parametric model, the rain rate presented a Rankine-like profile with a linear variation from the tropical cyclone center to the radius of maximum rain rate, followed by an exponential decay. In addition, it has been proved using the findings of Kaplan and DeMaria (1995) that the hurricane rain rates and wind speeds are always highly correlated before/after landfall. While the R-CLIPER model could be employed over both the ocean and land, it assumed a symmetric distribution of the rain rate inside the tropical cyclone. Lonfat et al. (2007) improved the spatial variation of the rain field by introducing a modified version of the R-CLIPER model known as the Parametric Hurricane Rainfall Model (PHRaM) with consideration of the wind shear effects. However, both
the R-CLIPER and PHRaM models were found to underestimate the maximum rain rate since they are based on the ensemble averages of numerous hurricanes (Tuleya et al. 2007).

Very few physics-based rain rate models have been introduced in the technical literature. In the theoretical model proposed by Langousis and Veneziano (2009a), it is assumed that all the upward moisture flux at the top of the tropical cyclone boundary layer is converted into rainfall. The vertical moisture flux was evaluated from the vertical winds at a reference height, generated by a modified version of the wind field model proposed by Smith (1968), along with the depth-averaged temperature and saturation ratio. Although observations from Hurricane Frances presented a maximum correlation (around 0.85) between the surface rain and vertical wind at an elevation of 2-3 km, more comprehensive data may be necessary for the selected reference height. Also, it is not easy to implement the modified version of Smith’s non-linear model in the Monte Carlo technique with a large number of simulations needed. While the model by Langousis and Veneziano (2009a) has been demonstrated to provide good estimates of the tropical cyclone rain rate, it cannot account for post landfall scenarios.

In this study, a new, analytical framework for tropical cyclones rain rate estimate will be developed for high-efficiency simulations. The rain rate analysis framework is based on the observation that rain-induced momentum flux at Earth’s surface cannot be ignored (e.g., Caldwell and Elliot 1971; 1972; Zhao et al. 2013). The total surface stress results mainly from momentum flux contributions of wind and rain. A general formula of the rain intensity has been first derived in which the iteration approach was utilized in the computation. The proposed methodology is able to effectively integrate an efficient wind field model for rapid estimation of rain intensity during tropical cyclones. Specifically, a height-resolving scheme recently developed by Snaiki and Wu (2017a; 2017b) was utilized leading to a linear, analytical solution of the surface rain rate. The
rainbands in the proposed methodology were simulated using a local perturbation scheme (e.g., Samsury and Zipser 1995; Holland et al. 2010; Li and Wang 2012). Sensitivity analysis of the rainfall field to several essential parameters in the rain rate simulation was comprehensively conducted. The present analytical framework for rapid rain rate estimation has been validated using observation data obtained from various hurricanes.

2. An analytical framework for rain rate estimation

The heat and moisture fluxes that are considered as the fuel and the surface stress contributing to the dissipation essentially govern the intensity of a tropical cyclone (Chen et al. 2007), as illustrated in Fig. 1.

![Diagram of major processes governing tropical cyclone intensity](image)

**Fig. 1.** Major processes governing tropical cyclone intensity

In this study, the modeling of rain intensity is based on the dissipation process of tropical cyclones where the total momentum flux is decomposed into several stress contributions. The general practice to consider the total surface shear stress or equivalently the momentum flux density \( \tau \) is based on the following parameterization (Andreas 2004; Donelan et al. 2004; Jarosz et al. 2007; Huang 2012):

\[
\tau = \rho_a u^*_{\text{effective}}^2
\]  (1)
where \( u_{\text{effective}} \) = effective frictional velocity during rain; and \( \rho_a \) = air density. This parameterization has been used in a number of applications such as the Fifth-generation Pennsylvania State University–National Center for Atmospheric Research Mesoscale Model (MM5) and the fully coupled atmosphere–wave–ocean model (AWO) (e.g., Chen et al. 2013). Furthermore, the effective frictional velocity can be related to the effective drag coefficient \( C_{d,\text{effective}} \) and the wind speed \( V_{\text{wind}} \) based on the following formula:

\[
\frac{u_{\text{effective}}}{V_{\text{wind}}} = \frac{C_{d,\text{effective}}}{2}
\]

Therefore, the total shear stress can be expressed as:

\[
\tau = \rho_a C_{d,\text{effective}} V_{\text{wind}}^2
\]

The square law as indicated in Eq. (3) has been comprehensively validated, especially at high wind speeds, with a number of observations (Garratt 1977).

Various factors may contribute to the total surface stress, namely the turbulent fluxes, rain effects, spray, airborne sediment (Saltation theory) and convection-induced stress. It is assumed here that no airborne sediment exists, therefore its momentum flux contribution is disregarded as well as the convection-induced stress (Huang 2012). The spray actually does not add extra momentum to the system but redistributes the wind stress near the surface (Andreas 2004). Since the droplets around the spray evaporate quickly (Emanuel 1995), its contribution is typically negligible (e.g., Wu 1973; Fairall et al. 1994). On the other hand, the rain-induced momentum flux can be significant as the raindrops interact with the near-surface wind and transfer momentum to the surface (Caldwell and Elliot 1971; 1972). Zhao et al. (2013) compared the rain-induced horizontal stress and the wind stress at various wind speeds and rain rates. It was demonstrated
that the horizontal stress by rain can have the same order of magnitude with that by wind. Therefore, it is important to consider the rain-induced stress contribution to the total stress near the surface. The total stress is then partitioned between two momentum contributions as:

\[
\tau = \tau_a + \tau_r \tag{4}
\]

where \(\tau_a\) = momentum flux contribution from wind; and \(\tau_r\) = momentum flux contribution from rain. The wind stress \(\tau_a\) can be expressed, in a similar way as in Eq. (3), in terms of the drag coefficient \(C_d\) without rain effects:

\[
\tau_a \equiv \rho_a C_d V_{\text{wind}}^2 \tag{5}
\]

A widely used parametrization of the rain stress relates it to the rain rate and wind speed as (Caldwell and Elliott 1971; 1972):

\[
\tau_r \equiv \gamma_r \rho_r V_{\text{wind}}^2 R \tag{6}
\]

where \(\rho_r\) = density of rainwater; \(\gamma_r\) = empirically determined factor varying between 0.8 and 0.9; and \(R\) = rain rate. In this study \(\gamma_r = 0.85\) will be adopted (Caldwell and Elliott 1971; Wong and Toumi 2016). The parameterization of Eq. (6) has been incorporated in several models, e.g., the one-dimensional mixed-layer model developed by Clayson and Kantha (1999), the bulk parametrization outlined by Fairall et al. (1996), and the Regional Ocean Modelling System (ROMS). Equations (3), (4), (5) and (6) lead to the following formula of the rain rate:

\[
R = \frac{\rho_a C_{d,\text{effective}} V_{\text{wind}}^2}{\gamma_r \rho_r V_{\text{wind}}^2} - \frac{\rho_a C_d V_{\text{wind}}^2}{\gamma_r \rho_r V_{\text{wind}}^2} \tag{7}
\]

It is interesting to note that the obtained rain rate formula implies that the rain-induced stress is proportional to the wind-induced one. The wind stress is well known to be responsible to maintain
the observed wind velocity profile in the atmospheric boundary layer (Taylor 1916). Since the wind and rain horizontal velocities are proportional to each other (e.g., Guo et al. 2001; Fu et al. 2015), the relation between wind and rain-induced stresses indicated in Eq. (7) is reasonable. The effective drag can be related to the roughness length based on the logarithmic law of wind profile in the vicinity of the surface as (Meng et al. 1995; Bryant and Akbar 2016):

$$C_{d,\text{effective}} = \frac{\kappa^2}{\ln\left(\frac{z_{10}}{z_{0,\text{effective}}^{\kappa}}\right)}$$  \hfill (8)

where $\kappa = \text{von Karman coefficient}$; $z_{0,\text{effective}} = \text{the effective roughness length}$; and $z_{10} = 10\, \text{m}$. Similarly, the drag coefficient $C_d$ (no rain effects) can be expressed in terms of the roughness length $z_0$ as $C_d = \kappa^2\left[\ln\left(\frac{z_{10}}{z_0}\right)^\kappa\right]^2$.

The studies of characterizing the rain-induced roughness length over land are very limited. On the other hand, the logarithmic profile well representing the lower part of the boundary-layer winds over both the ocean and land as indicate by a large number of field measurements (e.g., Powell et al. 2003; Vickery et al. 2009; Tse et al. 2013, Shu et al. 2017) provides a good approach for the estimation of the effective frictional velocity. As a result, the effective drag coefficient $C_{d,\text{effective}}$ may be obtained through the least squares fit of the measured or simulated (considering rain effects) vertical wind speed profile in the linear logarithmic space (e.g., Powell et al. 2003; Holthuijsen et al. 2012; Vickery et al. 2009; Bell et al. 2012). In this study, the height-resolving, wind-rain interaction model developed by Snaiki and Wu (2017c) (see Appendix A) is utilized. Since the rain rate is a necessary input to take the rain effects into account, the iteration technique is employed in the simulation as illustrated by Fig. 2.
The initial guess of the rain profile could be based on an empirical model as follows (Snaiki and Wu 2017c):

\[ R = R_{\text{max}} \left( \frac{r_m}{r} \right)^b \exp \left[ 1 - \left( \frac{r_m}{r} \right)^b \right]^{0.5} \]  

(9)

where \( R_{\text{max}} \) = maximum rain rate located at the radius of maximum wind \( r_m \); \( r \) = radial distance from the tropical cyclone center; and \( b \) = scaling parameter that adjusts the profile shape and depends on the radial extent of the tropical cyclone rain field. The maximum rain rate \( R_{\text{max}} \) was estimated with an empirical formula available in literature (Tuleya et al. 2007):

\[ R_{\text{max}} = a + b \left[ 1 + \left( V_m - 35 \right)/33 \right] \]  

(10)
where $a$ and $b =$ constants from the least squares fit of the TRMM radial rainfall rates; and $V_m =$ maximum wind speed. Although the iterative calculation is required in the analysis, it converges quite rapidly. With a good initial guess of the rain profile based on Eq. (9), two or three iterations are needed with a prescribed threshold $\varepsilon = 5\%$ for all simulations in the present study. The first-order closure of the rain rate simulation in Fig. 2 results in a simplified model of Eq. (7):

$$ R = \frac{\rho a k_m dV_{wind, modified}}{dz} - \frac{\rho b k_m dV_{wind}}{dz} $$

where $k_m = 100 \, m^2/s$ is the eddy viscosity; $V_{wind, modified} =$ modified wind speed due to the rain effects.

In the case of the marine conditions, the effective roughness length $z_{0, effective}$ could be expressed as a combination of the aerodynamic roughness length $z_0$ and the rain-induced one $z_{0,r}$ (Kumar et al. 2009):

$$ z_{0, effective} = z_0 + z_{0,r} $$

Several formulas have been proposed in the literature to characterize the rain-induced roughness length $z_{0,r}$ over ocean. For instance, Kitaigorodskii [referred in Houk and Green (1976)] defined $z_{0,r}$ as:

$$ z_{0,r} = 0.03\sigma $$

where $\sigma$ is the standard deviation of the experimentally obtained mean water level:

$$ \sigma = \beta \frac{kD}{g\nu} $$
where $\beta$ = a constant of 0.01; $D$ = drop diameter; $g$ = gravitational acceleration; $\nu$ = kinematic viscosity of water; and $k$ is the rain kinetic energy flux that can be related to the rain intensity $R$ and the terminal velocity of the rain drop $V_t$:

$$k = \frac{RV_t^2}{2}$$  \hfill (15)

As a result, the rain-induced roughness can be expressed as:

$$z_{0,r} = \frac{0.015 \beta D V_t^2 R}{g \nu}$$  \hfill (16)

Accordingly, the rain rate over ocean can be calculated as:

$$R = \frac{\rho_w k^2 V_{wind} \left( \ln \left( \frac{z_{10}}{z_0 + z_{0,r}} \right) \right)^{-2} - \ln \left( \frac{z_{10}}{z_0} \right)^{-2}}{\gamma \rho_r}$$  \hfill (17)

It should be noted that the use of Eq. (17) requires an iteration process since the rain-induced roughness length $z_{0,r}$ depends on the rain intensity.

### 3. Wind field model

The proposed analytical framework for rapid estimate of the rain rate requires the horizontal wind speed components as input. A recently developed height-resolving model by Snaiki and Wu (2017a; 2017b) is utilized here due to its high efficiency to obtain the wind field. A brief discussion of the wind field model is presented in this section for the sake of completeness. The governing equation of the wind field is described as follows:
\[
\frac{\partial \mathbf{v}}{\partial t} + \mathbf{v} \cdot \nabla \mathbf{v} = -\frac{1}{\rho_a} \nabla p - f \mathbf{k} \times \mathbf{v} + \mathbf{F}
\]  

(18)

where \( f \) = Coriolis parameter; \( p \) = pressure; \( \mathbf{k} \) = unit vector in the vertical direction; and \( \mathbf{F} \) = frictional force. In order to solve Eq. (18), the decomposition method is used where the wind velocity \( \mathbf{v} \) is expressed as:

\[
\mathbf{v} = \mathbf{v}_g + \mathbf{v}'
\]

(19)

where \( \mathbf{v}_g \) = gradient wind in the free atmosphere; and \( \mathbf{v}' \) = frictional component near the ground surface. The solution of the gradient wind speed could be solved straightforwardly in the cylindrical coordinate system (Georgiou 1986; Meng et al. 1995):

\[
v_{og} = \frac{(-\sin(\theta - \nu) - fr)}{2} + \left[ \frac{(-\sin(\theta - \nu) - fr)^2}{4} + \frac{r}{\rho_a} \frac{\partial p}{\partial r} \right]^{1/2}
\]

(20)

where \( \nu \) = approach angle (counter clockwise positive from the East); \( c \) = translation speed of the tropical cyclone; and \( \theta \) = azimuthal angle. The radial velocity \( \mathbf{v}_r \) = \(-\frac{1}{r} \frac{\partial \mathbf{v}_r}{\partial \theta} \) obtained from the continuity equation is usually disregarded due to its insignificant effects (Meng et al. 1995).

To obtain the frictional wind speed component, the nonlinear governing equation is first simplified using the scale analysis approach, and then linearized leading to the following frictional wind speed formulas as:

\[
u'(\theta, z')= (\alpha/\beta)^{1/2} \times \text{Real} \left\{ A_0 \times e^{(q_i z')} + A_i \times e^{(q_i z'+i\theta)} + A_{-1} \times e^{(q_i z'-i\theta)} \right\}
\]

(21a)

\[
u'(\theta, z')= \text{Imag} \left\{ A_0 \times e^{(q_i z')} + A_i \times e^{(q_i z'+i\theta)} + A_{-1} \times e^{(q_i z'-i\theta)} \right\}
\]

(21b)
where \((u', v')\) = frictional components of the wind velocity; 

\[
q_i = -(1 + i) \left[ \gamma + \sqrt{\alpha \beta - \phi} \right]^{1/2};
\]

\[
q_{i_1} = -(1 + i) \left[ -\gamma + \sqrt{\alpha \beta - \phi} \right]^{1/2}; \quad q_o = -\left( \alpha \beta \right)^{1/4}; \quad \alpha = \frac{1}{2k_m} \xi_x; \quad \beta = \frac{1}{2k_m} \xi_y; \quad \xi_z = \frac{2\nu_{\theta g}}{r} + f\]

is the absolute angular velocity; 

\[
\xi_{ag} = \frac{\partial \nu_{\theta g}}{\partial r} + \frac{\nu_{\theta g}}{r} + f
\]

is the vertical component of absolute vorticity of the gradient wind; 

\[
\gamma = \frac{1}{2K} \frac{\nu_{\theta g}}{r}; \quad \phi = \frac{1}{2Kr} \frac{\partial \nu_{\theta g}}{\partial \phi};
\]

and \(z'\) = new vertical coordinate used as the base of the computation scheme where \(z'=0\) is located above \(z_{10}\) (i.e., the 10 m height above the mean height of roughness elements) (Meng et al. 1995). The other necessary parameters needed for the simulation can be acquired in Snaiki and Wu (2017a; 2017b).

4. Model validation

The necessary parameters needed for the tropical cyclone wind field simulation are: \(\nu\) approach angle; \(c\) translation velocity of the hurricane; \(p_c\) central pressure; \(\Delta p\) central pressure difference; \(R_{max}\) radius of maximum winds; \(B\) Holland’s parameter; \(\psi\) latitude; and \(\lambda\) longitude. These wind parameters can be obtained from the National Hurricane Center’s North Atlantic Hurricane Database (HURDAT). On the other hand, the tropical cyclone rainfall data can be acquired from the high-fidelity numerical weather prediction models, the Tropical Rainfall Measuring Mission (TRMM) database or the Radar observations.

4.1. Comparison with MM5 model

In this section, the ensemble average of 12 surface rain fields corresponding to hurricane Frances (2004) in the period from 29th August to 1st September 2004 (6-h intervals) were simulated based
on the Fifth-Generation Pennsylvania State University/NCAR Mesoscale Model (MM5) (Langousis and Veneziano 2009a). A resolution of 1.67 km was used to generate the azimuthally averaged rain rates in the MM5 simulations. The rain rates obtained from MM5 were compared with those based on the proposed analytical framework, where the necessary parameters for the rain intensity simulation were extracted from the HURDAT database (Table 1). As shown in Fig.3, a good agreement between the MM5 and present simulation results is achieved.

**Table 1.** Tropical cyclones characteristics of the selected 12 rain fields

<table>
<thead>
<tr>
<th>Mon</th>
<th>Day</th>
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<th>Hurricane Center</th>
<th>Latitude (deg)</th>
<th>Longitude (deg)</th>
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<th>Storm Direction (deg)</th>
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<th>Δp (hpa)</th>
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**Fig. 3.** Comparison of the azimuthally-averaged, rain-rate radial profile of Hurricane Frances
4.2. Comparison with PR/TRMM rain fields

To validate the proposed analysis framework for rapid rain rate estimation during tropical cyclones, 38 PR/TRMM rain frames with a spatial resolution of $5\ km \times 5\ km$ were utilized. The parameters of the data collected from the PR/TRMM rain frames are summarized in Table 2 (Langousis and Veneziano 2009b). These 38 frames were selected from eight tropical cyclones that cover a wide range of storm intensities. Fig. 4(a) represents a scatterplot of the ratios between the PR/TRMM and simulated rain intensities $i_{PR}/i_{sim}$. A total number of 73819 points were selected from the 38 TRMM frames, covering a wide range of tropical cyclone spatial locations. All the necessary parameters for the rain rate simulation are taken from Table 2 (Langousis and Veneziano 2009b).

A large dispersion similar to the finding of Langousis and Veneziano (2009a) is observed in Fig. 4(a). It mainly results from the significant small-scale variability of rain rate due to the rainbands and local intensifications. Figure 4(b) depicts the local average and standard deviation of the ratio $i_{PR}/i_{sim}$ with a moving window of 2000 points. Clearly, the simulation based on the proposed analyses framework of rain field is unbiased since the moving average is close to unity. The moving standard deviation on the other hand is quite large highlighting the significance of the small-scale variability of the tropical cyclone rain field (Powell 1990; Molinari et al. 1994; Langousis and Veneziano 2009a).
Table 2. Tropical cyclones characteristics of the PR/TRMM rain fields (Langousis and Veneziano 2009b)

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Fig. 4. Comparison of the PR/TRMM and simulated rain rates: (a) Scatter plot of $i_{PR}/i_{sim}$; (b) Local average and standard deviation of $i_{PR}/i_{sim}$.

4.3. Comparison with TMI/TRMM and radar observations

The rain distribution of hurricane Dennis (1999) obtained from TMI/TRMM (Lonfat et al. 2004) was compared with the simulation result from the present methodology. As shown in Fig. 5, a good agreement between the simulations and observations is highlighted except in some local intensification regions caused by rainbands. TMI measurements are known to be more accurate for rainfall estimates over ocean than land. On the other hand, the Hydro-Next-Generation Doppler Radar system can measure the rainfall distribution over both ocean and land with very good accuracy (Lin et al. 2010). Hurricane Isabel (2003) was employed here to validate rain field simulation over land based on the present analysis framework. Isabel made landfall in North Carolina on 18 September 2003 as a category 2 tropical cyclone and caused widespread damages from storm surge flooding, wind and riverine flooding (Lin et al. 2010). Figure 6 presents a good agreement between the simulated and observed rainfall distribution up to 200 km radius. The orographic enhancement of rainfall due to the interaction between the tropical cyclone and the...
complex terrain conditions of the central Appalachians Mountains, not considered in the present
model, resulted in large differences beyond the range of 200 km (Lin et al. 2010).

Fig. 5. Comparison of the azimuthally-averaged, rain-rate radial profile of Hurricane Dennis

Fig. 6. Comparison of the azimuthally-averaged, rain-rate radial profile of Hurricane Isabel

5. Rainband simulation

Tropical cyclones exhibit significant rain rate in the eyewall and a set of rainbands (Willoughby
et al. 1984; Willoughby 1988). The complicated underlying mechanisms governing the rainbands
make their pattern vary from one tropical cyclone to another (Houze et al. 2006). Hence, it is
extremely challenging to systematically simulate these rainbands. The rainbands outside the
eyewall are usually characterized by secondary horizontal wind maxima (Samsury and Zipser
1995). A number of field measurements indicate that the local boundary-layer wind maxima in the rainbands are typically associated with pressure perturbations (e.g., Powell 1990; Yu and Tsai 2010; Lin et al. 2010; Sitkowski et al. 2011; Li and Wang 2012). Since the wind field here is simulated by a large-scale model (Snaiki and Wu 2017a; 2017b) and hence does not account for local intensifications, considerable differences between the simulated and observed rain rates are typically observed in the rainband regions as discussed in the preceding sections.

A good approach to improve the simulation would be based on a proper perturbation of the original wind profile that could characterize the rainbands. The rain rate heavily depends on the wind speed, hence, a perturbation in the wind field will generate a corresponding perturbation in the rain rate and hence the local maxima of the rain intensity (local perturbation scheme). This approach can be more convenient to be implemented by introducing a perturbation in the pressure profile that will generate a corresponding wind perturbation (Holland et al. 2010). To illustrate the adopted scheme, the rain field of hurricane Dennis (Fig. 5) was revisited to incorporate the contribution of the rainbands. The rainband locations were first identified from the TMI imagery [Fig. 7(b)]. Then, the simulated wind speeds at a 10-meter height were compared with those from the Remote Sensing System QuikScat (version-4) data [Fig. 7(a)] (Ricciardulli and Wentz, 2015) to obtain the perturbation values. Table 3 compares and presents the observed and simulated wind speeds in three different rainband regions. The consideration of perturbations in the wind field leads to an improved rain rate, as shown in Fig. 8. A good agreement between the observed and simulated profiles is observed. Figure 9 presents a comparison of the simulated rain rates of Hurricane Dennis with and without considerations of the rainbands. The corresponding correlation coefficients between the observations and simulations are 0.96 and 0.92, respectively. While the results demonstrate that the rain intensities predicted by both approaches match reasonably well
with the measured data, the simulation accuracy is certainly improved by integrating the rainband effects.

Fig. 7. Hurricane Dennis (1999) wind and rain fields: (a) Surface wind speed distribution obtained from QuikScat on August 28 (left) and August 29 (right); (b) Surface rain rate provided by TMI/TRMM on August 28 (left) and August 29 (right)
Table 3. Comparison between the observed (QuikScat) and simulated surface wind speed of Hurricane Dennis in the rainband regions

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Fig. 8. Comparison of the improved, azimuthally-averaged, rain-rate radial profile of Hurricane Dennis

Fig. 9. Comparison of the azimuthally-averaged rain rates of Hurricane Dennis
6. Sensitivity analysis

Numerous studies have demonstrated based on field measurements and numerical simulations that the rainfall distribution varies with a number of environmental factors (e.g., roughness) and inherent tropical cyclone features (e.g., intensity and translation speed) (Lonfat et al., 2004). The present rain rate analysis framework explicitly depends on these parameters. Figure 10 illustrates the sensitivity of the rain intensity to several selected parameters, namely $B$ Holland parameter, $r_m$ radius of maximum wind, $c$ translation speed, $\Delta p$ central pressure difference, $k_m$ turbulent diffusivity, and $z_0$ equivalent roughness length. The base case scenario is taken as: $B=1$, $r_m=30km$, $c=10m/s$, $\Delta p=70hpa$, $k_m=50m^2/s$, and $z_0=0.001m$. The radial profile of the rain rate was taken at $\theta=0^\circ$ (counterclockwise positive from the East) for all simulations.

As indicated in Fig. 10, the central pressure difference $\Delta p$, translational tropical cyclone speed $c$ and surface roughness $z_0$, have significant effects on the rain rate. Actually, it has been widely reported that the surface roughness can substantially alter the rain intensity (e.g., Trenberth et al. 2007; Langousis and Veneziano 2009a; and Lin et al. 2010). Also, a higher value of $\Delta p$, corresponding to a larger maximum wind $v_m$ (Holland et al. 2010), leads to more intense rain. This result agrees with the observations and findings of the literature (e.g., Lonfat et al. 2004; Tuleya et al. 2007; Langousis and Veneziano 2009a). Similarly, an increase of the translational tropical cyclone speed results in the enhancement of the total precipitation. Figure 10 also implies that the smaller radius of maximum wind is associated with a more peaked profile shifted to the tropical cyclone center. The Holland’s parameter $B$ mainly modifies the decay rate of the rainfall intensity profile and presents small effects on the maximum rain rate. In addition, low sensitivity of the rain rate to the vertical turbulent diffusivity is noted in the figure.
The spatial distribution of the rain intensity was investigated based on the following case study parameters: $B = 1.2$; $c = 3 \text{ m/s}$; $r_m = 40 \text{ km}$; $\Delta \rho = 60 \text{ hpa}$; $z_0 = 0.01 \text{ m}$. Figures 11(a) and 11(b) depict the three-dimensional shaded surface of the rain intensity and their corresponding contours, respectively. Due mainly to the tropical cyclone motion, the rain field is asymmetric. This result conforms with the finding of Langousis and Veneziano (2009a), highlighting the maximum rain rate location near the radius of maximum winds in the eyewall region (Lonfat et al. 2004; Tuleya et al. 2007).

**Fig. 10.** Sensitivity analysis of the rain rate at $\theta = 0^\circ$. 
7. Concluding remarks

A new, analytical framework for rapid rain rate estimation was proposed in this study. The rain rate formula was essentially developed based on the observation that rain-induced momentum flux at the surface cannot be ignored. The total surface stress results mainly from momentum flux contributions of wind and rain. The obtained results indicate that the spatial distribution of the rainfall field is governed by the wind field inside the tropical cyclone. This observation confirms the findings of several previous studies in which the rain intensity is shown to be highly correlated with the horizontal wind speed. A recently developed, height-resolving wind field model was utilized during the model construction leading to a linear, analytical solution of the surface rain rate. The obtained analysis framework for rain field explicitly depends on parameters for a typical tropical cyclone wind field simulation, namely storm location, approach angle, translation speed, radius of maximum wind, pressure profile, surface drag coefficient, and turbulent diffusivity. The
sensitivity analysis was extensively carried out to investigate the effects of several tropical cyclone and environmental parameters on the rain rate. It has been demonstrated that the rain rate heavily depends on the central pressure difference, translational tropical cyclone speed and surface roughness. The proposed analysis framework for the rain field is based on the large-scale horizontal wind field, therefore, it does not account for local rainfall intensifications due to rainbands. A plausible approach based on the local perturbation scheme was introduced to simulate the rain rates inside the rainband regions. The present analytical framework for rapid estimate of rain rate offers good simulation results that are consistent with tropical cyclone observations. It can be readily used in conjunction with the Monte Carlo techniques for risk analysis of tropical cyclone hazards.

Acknowledgments

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Appendix A

The drag force \( f_i \) exerted by one raindrop on air in the horizontal direction is introduced:

\[
f_i = \frac{1}{2} \rho_a C_{d,r} V_{rel} (V_{\text{rain}} - V_{\text{wind}}) \pi r_d^2 \tag{A.1}
\]

where \( C_{d,r} = \) drag coefficient for a raindrop of radius \( r_d \); \( V_{\text{rain}} = \) raindrop horizontal velocity; \( V_{\text{wind}} = \) wind velocity in the horizontal direction; and \( V_{rel} = \) the total relative speed of the raindrop. As a result, the total drag force \( f_d \) applied on a tiny volume of air \( V = A \Delta z \) can be obtained as follows:

\[
f_d = \frac{3R \rho_a C_{d,r} V (V_{\text{rain}} - V_{\text{wind}})}{4d} \tag{A.2}
\]

where \( d = \) the raindrop diameter. The total drag force is then integrated into the governing equation of the wind velocity [Eq. (18)]. With several mathematical manipulations, one can obtain the gradient wind speed as follows (Snaiki and Wu 2017c):

\[
v_{ag} = \frac{2}{r} \frac{A^2}{\xi_{ag}} \left( 1 + \frac{A^2}{\xi_{ag}^2} \right) + \frac{c \sin (\nu - \theta)}{r} \left( 1 + \frac{A^2}{\xi_{ag}^2} \right) - \left( f + \frac{A^2}{\xi_{ag}^2} \right) + \frac{c \sin (\nu - \theta)}{r} \left( 1 + \frac{A^2}{\xi_{ag}^2} \right) \left( 1 + \frac{A^2}{\xi_{ag}^2} \right) + \frac{4 \varphi_p}{\rho_a r} \left( 1 + \frac{A^2}{\xi_{ag}^2} \right) \right]^{1/2} \tag{A.3}
\]

where \( A = \frac{\pi C_{d,r} N r_d^2}{2} \gamma_1 \); and \( N = \frac{3R}{4 \pi r_d^3} \) is the number of raindrops per second on a unit area. On the other hand, the frictional wind speed components can be obtained as follows (Snaiki and Wu 2017c):
\[ v'_r = Y e^{-\alpha r} \left[ W_1 \cos(bz') + W_2 \sin(bz') \right] \]  
(A.4a)

\[ v'_\phi = e^{-\alpha r} \left[ W_2 \cos(bz') - W_1 \sin(bz') \right] \]  
(A.4b)

where \( Y = \sqrt{\frac{p}{q}} \); \( p = \frac{1}{2k_m} \zeta_g \); \( q = \frac{1}{2k_m} \xi_{eg} \); \( A' = \frac{N C_d \pi v_{d}^2}{2} \gamma_2 \). The variables \( a \) and \( b \) are defined as:

\[
 a = \sqrt{\frac{A'/k_m + \sqrt{A'^2/k_m^2 + 4pq}}{2}} \]  
(A.5a)

\[
 b = \sqrt{\frac{A'^2/k_m^2 + 4pq - A'/k_m}{2}} \]  
(A.5b)

The variables \( W_1 \) and \( W_2 \) are determined from the boundary conditions and presented as follows:

\[
 W_1 = -\left( \frac{a}{b} + \chi - Z \right) \left( \frac{\chi}{\chi'} - \frac{Z}{\chi'} \right) v_{rg} + (\chi - Z) v_{bg} \]  
\[ \frac{1}{1 + \left( \frac{a}{b} + \chi - Z \right)^2} \]  
(A.6a)

\[
 W_2 = \left( \frac{\chi}{\chi'} - \frac{Z}{\chi'} \right) v_{rg} - \left( \frac{a}{b} + \chi - Z \right) (\chi - Z) v_{bg} \]  
\[ \frac{1}{1 + \left( \frac{a}{b} + \chi - Z \right)^2} \]  
(A.6b)

where \( \chi = \frac{C_d d, v}{kb} \); \( Z = \frac{\gamma \rho R}{\rho_a bk} \); and \( v_s \) = total wind velocity near the ground surface.

Hurricane Katrina (2005) is employed here for the validation purpose. The anemometer was located on the 42003 station at (26°0'25" N, 85°38'54" W). The 10-min averaged time was
used for the observed wind data at approximately 10 m height. As shown in Fig. 12, the results generated by the present model are consistent with hurricane Katrina wind field observations.

**Fig. 10.** Observed and simulated wind speeds (top) and directions (bottom) of Hurricane Katrina
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