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Microphysical-Dynamical Interaction in Tropical Cyclone Intensification

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FLORIDA INTERNATIONAL UNIVERSITY

Miami, Florida

MICROPHYSICAL-DYNAMICAL INTERACTION IN TROPICAL CYCLONE
INTENSIFICATION

A dissertation submitted in partial fulfillment of the

requirements for the degree of

DOCTOR OF PHILOSOPHY

in

EARTH SYSTEMS SCIENCE

by

Zongyao Yang

2020

To: Dean Michael R. Heithaus
College of Arts, Sciences and Education

This dissertation, written by Zongyao Yang, and entitled Microphysical-Dynamical Interaction in Tropical Cyclone Intensification, having been approved in respect to style and intellectual content, is referred to you for judgment

We have read this dissertation and recommend that it be approved.

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ABSTRACT OF THE DISSERTATION

MICROPHYSICAL-DYNAMICAL INTERACTION IN TROPICAL CYCLONE
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Florida International University, 2020

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Professor Ping Zhu, Major Professor

Producing timely and accurate tropical cyclone (TC) intensity forecasts remains one of the most difficult challenges facing meteorologists today. The state-of-the-art three-dimensional (3D) full physics operational models, in particular, have problems in simulating rapid intensification (RI), a situation where a TC intensifies dramatically in a short period of time. For example, Hurricanes Patricia (2015) and Maria (2017) increased their maximum sustained winds by 90 knots and 70 knots within 24 hours, respectively. The major objectives of this dissertation are to (a) explore the underlying reasons why the Hurricane Weather Research and Forecasting (HWRF) system, one of the operational models used at the Environmental Modeling Center (EMC), NOAA for TC prediction, often fails to predict RI of major hurricanes; (b) improve turbulent mixing and hydrometeor sedimentation parameterizations in HWRF so that the model can improve the capture of the dynamical-microphysical interaction in the TC inn-core region; and (c) advance our understanding of the internal dynamics that governs the RI of TCs. Our investigations show that HWRF is unable to generate appropriate sub-grid-scale (SGS) turbulent forcing above

the boundary layer in the eyewall of a TC as a result of the lack of consideration of the cloud induced buoyancy. Incorporating a saturated Brunt-Väisälä frequency (BVF) in static stability calculation allows HWRF's turbulent mixing scheme to successfully generate the buoyancy production of turbulence in the eyewall, which notably improves HWRF's skill in predicting RI of TCs. Our analyses show that the HWRF microphysics scheme fails to produce hydrometeors with fall speed smaller than 0.2 ms^{-1} , which are abundant in TC clouds according to in-situ aircraft measurements. The failure stems from the fact that the hydrometeor fall speed parameterization used in HWRF was developed in quiescent conditions that neglects the strong impact of convective currents on falling hydrometeor particles. To fully consider the impact of dynamic-microphysical interaction on hydrometeor sedimentation in the eyewall, a new parameterization of particle fall speed in non-quiescent conditions is developed and implemented in HWRF. The new scheme successfully generated the lightest hydrometeors that the operational HWRF fails to produce and allows the HWRF to realistically simulate the convection in the TC inner-core region. Using the model output, this dissertation clarifies the role of eyewall convection in modulating the outflow temperature of a TC and in triggering the positive feedback mechanism among surface winds, evaporation, and storm intensification underlying the RI of TCs.

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1 Introduction

Tropical cyclones (TCs) pose a significant risk to loss of life and damage in properties in coastal areas. The strong and deadly landfalling TCs can kill thousands of people, cause billions of dollars in property damage, and result in immense human suffering. One example of such deadly storms is Hurricane Patricia (2015), which caused 13 fatalities (8 direct and 5 indirect) and \$463 million damages in the central America, Mexico, and Texas. In 2017, the total damage loss of Atlantic hurricane season was over \$400 billion, and essentially all of which was caused by three of the season's major hurricanes --- Harvey, Irma, and Maria. All these deadly storms share the same characteristics in a period during their lifetime when storm intensity strengthens dramatically known as rapid intensification (RI) defined as an increase in the storm maximum sustained wind speed at least 30 knots within 24 h. Some major TCs even undergo very rapid intensification (VRI, ~ 30 knots in 12 h) and extreme rapid intensification (ERI, ~ 40 knots in 12 h), for example, Hurricane Patricia (2015), super Typhoon Meranti (2016), and Hurricane Maria (2017) increased their intensities by 90 knots in 24 h, 120 knots in 60 h, and 70 knots in 24 h, respectively. Timely and accurate forecasts of these deadly storms allow for advanced warning that can save lives and property. The National Hurricane Center (NHC) has increasingly relied upon the guidance from the numerical forecasts made by operational numerical weather prediction (NWP) models to issue TC track (position) and intensity (strength) advisory. However, operational numerical forecasts only show a very limited success in predicting the intensity change of these hurricanes. Figure 1 shows the intensity forecasts of hurricanes Patricia (2015), Harvey (2017), Irma (2017), and Maria (2017) at the different forecasting lead time by the

Hurricane Weather Research & Forecasting (HWRF) modeling system, one of the state-of-the-art three-dimensional (3D) full physics operational numerical models at the Environmental Modeling Center (EMC) NOAA, compared with the NHC best-track data. In all cases, the observed RIs and the peak intensities that the storms reached were not successfully predicted.

Figure 1.1: Intensity forecast of hurricanes Patricia, Harvey, Irma, and Maria at the different forecasting lead time by the operational HWRF compared with the NHC best-track data.

The poor intensity forecasts of RI storms by HWRF shown in Figure 1.1. reflects the current status of TC forecast, that is, while TC track forecasting has been improving steadily over the past couple of decades, no comparable improvement is seen in TC intensity forecasts (Rappaport et al. 2009). Since TC movement is primarily governed by the large-scale steering flow, the steady improvement in TC track forecast can be largely attributed to the ever increasing model resolution and advances in observational network, which allows the large-scale atmospheric fields to be better resolved by operational NWP models. Then, if a 3D TC vortex can undergo a RI, VRI or ERI in the real atmosphere, what prevents a full physics operational model from capturing RI, VRI or ERI of TCs provided that modern operational models have a sufficient resolution to resolve environmental conditions and advanced data assimilation packages to appropriately initialize a TC vortex? The reason is that TC intensification is not only modulated by the large-scale forcing, such as wind shear and underlying sea surface temperature (SST), but also depends largely on the TC internal dynamics and thermodynamics that involve a complicated interplay between dynamic and physical processes spanning over a spectrum of scales from the TC vortex-scale flow down to sub-grid scale turbulence and microphysics (Marks and Shay 1998). To date, how the TC internal processes interact with external forcing to govern the rate of the TC intensification remains poorly understood. Improving TC intensity forecast has been identified as one of the highest priorities of NHC (Kaplan et al. 2015).

1.1 Basic Dynamics of TC Intensification

Tropical cyclones (TCs) intensification may be understood in a cylindrical coordinate with its origin set at the center of a TC vortex in which the governing equations of the azimuthal-mean model-resolved tangential wind, radial flow, and vertical velocity of a TC may be written as,

$$\frac{D\bar{v}}{Dt} = -\bar{u} \left(f + \frac{\bar{v}}{r} \right) + F_\lambda + F_{sgs_\lambda} + \varepsilon_\lambda, \quad F_\lambda = -\overline{\tilde{u}' \frac{\partial \tilde{v}'}{\partial r}} - \overline{\tilde{v}' \frac{\partial \tilde{v}'}{r \partial \lambda}} - \overline{\tilde{w}' \frac{\partial \tilde{v}'}{\partial z}} - \frac{\overline{\tilde{u}' \tilde{v}'}}{r}, \quad (1.1)$$

$$\frac{D\bar{u}}{Dt} = C - \frac{1}{\bar{\rho}} \frac{\partial \bar{p}}{\partial r} + F_r + F_{sgs_r} + \varepsilon_r, \quad C = \frac{\bar{v}^2}{r} + f \bar{v}, \quad F_r = -\overline{\tilde{u}' \frac{\partial \tilde{u}'}{\partial r}} - \overline{\tilde{v}' \frac{\partial \tilde{u}'}{r \partial \lambda}} - \overline{\tilde{w}' \frac{\partial \tilde{u}'}{\partial z}} - \frac{\overline{\tilde{v}'^2}}{r}, \quad (1.2)$$

$$\frac{D\bar{w}}{Dt} = -\frac{1}{\bar{\rho}} \frac{\partial \bar{p}}{\partial z} - g + F_w + F_{sgs_w} + \varepsilon_w, \quad F_w = -\overline{\tilde{u}' \frac{\partial \tilde{w}'}{\partial r}} - \overline{\tilde{v}' \frac{\partial \tilde{w}'}{r \partial \lambda}} - \overline{\tilde{w}' \frac{\partial \tilde{w}'}{\partial z}}, \quad (1.3)$$

where r , λ , and z represent the radial, azimuthal, and vertical coordinate axes; $\tilde{u}$, $\tilde{v}$, and $\tilde{w}$ are the model-resolved radial, tangential, and vertical wind components, respectively; f is Coriolis parameter; $\bar{\rho}$ is the air density; Overbar and prime indicate the azimuthal-mean and the perturbation away from the azimuthal-mean; F_λ , F_r , and F_w are the azimuthal-mean tangential, radial, and vertical eddy correlation term resulting from the model-resolved asymmetric eddy processes; F_{sgs_λ} , F_{sgs_r} , and F_{sgs_w} are the azimuthal-mean tangential, radial, and vertical SGS tendency resulting from the parameterized SGS eddy processes (or turbulence); ε_λ , ε_r , and ε_w are the viscous dissipation in three directions, and $\frac{D}{Dt} = \frac{\partial}{\partial t} + \bar{u} \frac{\partial}{\partial r} + \bar{w} \frac{\partial}{\partial z}$ is the material derivative operator following air particles along the model-resolved axisymmetric flow. In classic TC theories, Defining the azimuthal-mean model-resolved absolute angular momentum per unit mass as $\bar{M} = r\bar{v} + \frac{1}{2}fr^2$, it is easy to show that Eq. (1.1) becomes,

$$\frac{D\bar{M}}{Dt} = r(F_\lambda + F_{sgs_\lambda}). \quad (1.4)$$

where $r(F_\lambda + F_{sgs_\lambda})$ represents the torque per unit mass resulting from the model-resolved and SGS eddy forcing acting on air parcels. It is clear from Eq. (1.4) that the asymmetric eddy processes (both resolved and parameterized) provide an important forcing for the acceleration or deceleration of the primary circulation of a TC. In addition, for negligible $F_\lambda + F_{sgs_\lambda}$, $\bar{M}$ is conserved. Thus, as air parcels move radially inward (decrease of r), they must spin up in order to conserve their $\bar{M}$. Conversely, air parcels must spin down as they move radially outward. The correlation is essential for the spin-up process of a TC vortex. In other words, the evolution of the primary circulation of a TC must be accompanied by a secondary overturning circulation. Physically, this overturning circulation can be induced by the friction at the ocean surface and diabatic heating of convection. The interaction among the primary TC vortex, secondary circulation, surface friction, and diabatic heating of convection that governs the evolution of a TC vortex may be schematically illustrated by Figure 1.2. To the first-order approximation, the main body of a TC is assumed to satisfy the gradient wind balance in the radial direction and hydrostatic balance in the vertical, i.e., Eq. (1.2) and Eq. (1.3) reduce to,

$$\frac{1}{\bar{\rho}} \frac{\partial \bar{p}}{\partial r} = C, \quad (1.5)$$

$$\frac{1}{\bar{\rho}} \frac{\partial \bar{p}}{\partial z} = -g. \quad (1.6)$$

Near the surface, the turbulence induced friction slows down the tangential wind so that the outward Coriolis force and centrifugal force can no longer balance the inward pressure gradient force, which breaks down the gradient wind balance and results in a net radial

inflow. As the radial inflow reaches the eyewall, it has to move swiftly upward along the eyewall. The boundary layer air moving out of the boundary layer into the free atmosphere is known as the Ekman pumping. As the eyewall upward flow hits the tropopause, it has to diverge outward to conserve its mass. As the flow diverges out, it quickly spins down the primary circulation as required by the conservation of angular momentum. Thus, a spin-up mechanism is required for a TC vortex to evolve. And it is the eyewall convection that serves as a converging spin-up mechanism to the spin-down process to counteract the diverging spin-down process induced by the Ekman pumping caused by surface friction.

Figure 1.2: A schematic illustration of the interaction among the primary TC vortex, secondary circulation, surface friction, and diabatic heating of convection during the evolution of a TC vortex.

Mathematically, an analytical expression known as the Sawyer-Eliassen equation (Eliassen 1951; Shapiro and Willoughby 1982) that depicts the secondary circulation in Figure 1.2 can be derived from the gradient wind balance Eq. (1.5), the hydrostatic balance Eq. (1.6), and the governing equation of energy conservation,

$$\frac{D\bar{\theta}}{Dt} = \bar{\theta} + F_{\theta} + F_{sgs_{\theta}}, \quad F_{\theta} = -\overline{\tilde{u}' \frac{\partial \bar{\theta}'}{\partial r}} - \overline{\tilde{v}' \frac{\partial \bar{\theta}'}{r \partial \lambda}} - \overline{\tilde{w}' \frac{\partial \bar{\theta}'}{\partial z}}, \quad (1.7)$$

where $\bar{\theta}$ is the model resolved potential temperature, $\bar{\theta}$ is the diabatic heating, F_{θ} and $F_{sgs_{\theta}}$ are the azimuthal-mean eddy forcing terms resulting from the model-resolved and parameterized SGS asymmetric eddy processes. The angular momentum conservation and the Sawyer-Eliassen equation provide a fundamental physical model for understanding the intensification of a rapidly-rotating vortex.

If a model can realistically capture the interaction among the primary vortex, secondary circulation, surface friction, and diabatic heating, it should have no problem to simulate the evolution of a TC. Then, why do three-dimensional full physics models not show the expected ability to predict TC intensification, in particular, RI? The weakest links here are the parametric representation of the SGS turbulence and microphysical processes in numerical models, which introduce a major uncertainty in TC simulations and reducing such a uncertainty has been one of the greatest challenges for TC studies. In the next two sections, I will make a brief review on turbulent and microphysical processes in the evolution of a TC.

1.2 Turbulent Boundary Layer Processes of TCs

Turbulence is commonly regarded as a flow feature pertaining to the planetary boundary layer (PBL), a shallow turbulent layer adjacent to Earth's surface with a depth typically about 1 km, often with a distinct diurnal variation in non-TC conditions. The importance of PBL turbulence to TC evolution has been recognized for a long time. Both the Conditional Instability of the Second Kind (CISK) and cooperative-intensification

mechanism (Ooyama 1982), the two early theories for TC intensification, recognized the role of the PBL in converging moisture to sustain deep convection of a TC. The Ekman pumping resulting from the PBL turbulence in association with the secondary circulation of a TC is clearly illustrated in Figure 1.2. In the meantime, the PBL turbulence also serves as a mechanism to transport the latent heat from ocean surface upward to foster the development of moist convection that generates the converging spin-up to counteract the Ekman pumping induced diverging spin-down process. As Charney and Eliassen (1964) pointed out, friction performs a dual role; it acts to dissipate kinetic energy, but because of the frictional convergence in the moist PBL, it acts also to supply latent heat energy to the system of TC. Later it was Emanuel's evaporation-wind feedback mechanism (Emanuel 2003) that first articulated the critical role of air-sea interaction in generating a positive feedback among the near-surface wind speed, the rate of evaporation from the underlying ocean, and vertical turbulent transport during the intensification process of a TC.

Figure 1.3: Composite TKE derived from airborne radar data from 116 radial legs of P3 flights in the 2003-2010 hurricane seasons as a function of height and the radius normalized by the radius of maximum wind (RMW, adopted from Zhu et al. 2019).

However, in all the previously mentioned theories, the TC-PBL was treated the same as that in non-TC conditions with a typical depth about 1 km. By doing so, they implicitly adopted the basic assumptions of the classic PBL theory: (1) turbulent mixing is responsible for the vertical transport of momentum, heat, and moisture; (2) vertical turbulent transport becomes negligible above the PBL; and (3) the mean vertical velocity $\bar{w}$ in the PBL is negligible compared with the vertical velocity fluctuations w' (Stull 1988). But these basic assumptions in fact are not all valid in TC conditions.

First, turbulence and the resultant turbulent transport are always negligible above the PBL. Intense turbulent mixing within the deep convective clouds has been widely

observed by aircraft, Doppler radar/lidar, and other advanced remote sensing instruments (e.g., LeMone and Zipser 1980; Marks et al. 2008; Hogan et al. 2009; Giangrande et al. 2013). In particular, using the turbulent kinetic energy (TKE) derived from the airborne radar data collected in Hurricane Rita (2005), Lorsolo et al. (2010) showed that large TKE exists above the PBL in the eyewall and rainbands. Figure 1.3 shows the composite of TKE derived using the Lorsolo et al. (2010) method derived from the airborne radar observations from 116 radial legs of P3 flights in the 2003-2010 hurricanes seasons. It clearly demonstrates that the intense turbulence exists above the PBL in the conventional definition all the way up to over 10 km in the eyewall. Recognizing the deep convective nature of TCs, Smith et al. (2008) and Smith and Montgomery (2010) warned that the conventional PBL theory may become invalid in the TC inner-core region as the low-level radial inflow ascends swiftly within the eyewall as indicated in Figure 1.2. In fact, the problem of applying conventional PBL theory to a deep convective regime had been recognized early in the 1970s and 1980s. Deardorff (1972) noted, *“The definition of PBL has not included the region of turbulence within towering cumuli but only the average height of surface induced turbulent fluxes outside of such clouds”*. Moss and Rosenthal (1975) added, *“The method (of defining the PBL) contains several elements that may or may not be applicable under hurricane conditions”*. Shapiro (1983) wrote, *“As the radius of maximum tangential wind is approached, the boundary layer itself becomes ill defined, as air is pulled up into the active convection”*. Stull (1988) also acknowledged that *“the conventional definition of PBL is not applicable to the intertropical convergence zone (ITCZ), where the air ascends into deep convective clouds.”* The fact that the turbulent transport is not negligible above the PBL in deep convection challenges if the PBL schemes

developed in non-TC conditions can appropriately account for the turbulent transport in the eyewall and rainbands.

Second, most of the PBL schemes were developed for unsaturated conditions where the turbulent PBL is often cleanly separated from the free atmosphere above by a capping inversion. However, in the deep convective eyewall and rainbands, there is no such a capping inversion available. As a consequence, the concept of PBL may be still applicable in the context that the turbulent layer is directly affected by the surface processes, for example, the inflow layer of the secondary circulation. But such defined PBL is by no means a physical layer that separates the turbulence generated by surface processes and by cloud processes aloft. How to appropriately treat the turbulence in the eyewall and rainbands generated by the surface processes and cloud processes is a question that has not been thoroughly addressed. These issues will be tackled in this dissertation study.

1.3 Microphysical Processes in TCs

The importance of microphysics to the formation and evolution of TCs has long been recognized. Previous studies have helped us understand the fundamental role of microphysical processes in TC development but also raised some unanswered questions regarding how ice-phase processes modulate TC intensification. For example, Lord et al. (1984) showed that cooling caused by melting ice particles can initiate and maintain downdrafts, and thus, facilitate TC intensification. Willoughby et al. (1984) further demonstrated that ice-phase processes can lead to more realistic simulation of concentric eyewalls and inner-core structures compared with the warm-rain processes only. On the basis of their simulations, Zhu and Zhang (2004) argued that the release of latent heat of

fusion above the freezing level associated with the Bergeron process provides extra energy for TC intensification. The existence of graupel also enhances the fallout of hydrometeors and reduces the water loading effects on the eyewall updrafts, which favors TC intensification. However, other studies (e.g., Wang 2002; Sawada and Iwasaki 2007) suggested that the inclusion of the ice-phase microphysics might weaken TC intensity since cooling from melting and sublimation of solid phase hydrometeors can enhance the downdrafts but weaken the secondary circulation and inward transport of absolute angular momentum, which may result in an inhibitive impact on the TC intensification. The large sensitivity (or discrepancy) of TC intensification to ice-phase microphysics suggests that the microphysical schemes used in simulations may not realistically represent the real microphysical processes in TC clouds.

The problem may be inherent to the infrastructure of numerical models, which use standalone physics modules to represent interconnected physical processes. In a real TC, the microphysical processes interact directly with the dynamical processes, such as coherent up-/down-drafts and in-cloud eddy circulations, to regulate the magnitude and spatial distribution of diabatic heating that ultimately determine the intensification rate of a TC. While all variables in a numerical model are somewhat connected, how to generate robust microphysical-dynamical interaction in TC clouds via standalone physics modules is an important scientific and technical problem. Previous studies mainly focused on individual microphysical schemes, few of them attempted to tackle the problem from the interaction between the microphysics and other physical modules and model-resolved dynamic and thermodynamic fields. In my dissertation study, I explore the problem from the perspective of microphysical-dynamical interaction.

1.4 Objectives and Organization of the Dissertation

The objectives of the dissertation are to (1) identify the key issues regarding the representation of turbulence and microphysical processes in TC simulations; (2) improve turbulence and microphysics schemes in operational HWRF; and (3) understand the role of microphysical-dynamical interaction in the inner-core region in TC intensification.

The dissertation is organized as follows. Chapter 2 presents a sensitivity study of TC intensification to various PBL and microphysics schemes used in numerical models. Chapter 3 investigates the problems of current PBL scheme in estimating static stability in eyewall and rainbands of TCs and provides an accurate estimation of static stability for the saturated atmosphere following the thermodynamic law. Chapter 4 examines the problems in determining the hydrometeor sedimentation in TC conditions and proposes a new parameterization of hydrometeor terminal velocity that considers the impact of convective currents and eddy circulations on particle sedimentation. Chapter 5 investigates and discusses the role of microphysical-dynamical interaction in TC inner-core region in modulating TC intensification. A summary this dissertation research is provided in Chapter 6.

2 Sensitivity of TC Intensification to PBL and Microphysics Schemes

As shown in Figure 1.1, operational HWRF appear to have problems capturing some of the observed RIs of major hurricanes. Since in most of these cases HWRF is able to predict fairly decent storm tracks, the large-scale conditions are reasonably captured by the model. It may be argued that the poor intensity forecasts in these cases may be attributed to the inappropriate internal processes simulated by the model. It raises a question if physical processes are realistically represented in HWRF. In my dissertation study, I focus on the turbulence and microphysical processes during the evolution of a TC. As a first step to answer the question, it is useful to examine how the simulated TC intensity changes in response to the changes in the PBL and microphysical parameterizations. However, as an operational model, HWRF does not provide options of model physics, which is understandable since it is not practical to switch model physics options during operational forecasts of TCs. A good modeling platform for physics sensitivity tests is the WRF model with the Advanced Research WRF (WRF-ARW) dynamic core (Shamarock et al. 2008), since it comes with multiple options of model physics, so that the performance of different parameterization schemes can be compared with the same model dynamic core. Using WRF-ARW for sensitivity tests also allows us to look into if the inability of operational HWRF to simulate some of the RIs of major hurricanes is the common problem of other numerical models, which may help us identify the underlying problems.

2.1 PBL Schemes

The PBL scheme is an important part of model physics that is designed to account for SGS vertical turbulent mixing and the resultant transport. There are numerous PBL

schemes available used in numerical simulations. They may be classified into three groups defined by the ways that the schemes are formulated: (a) K-closure (or first-order-closure) vs. high-order closure schemes. In K-closure schemes, the vertical turbulent fluxes of a generic scalar are related linearly to the vertical gradient of the scalar via eddy exchange coefficient, whereas in a high-order closure scheme one or more high-order turbulent moments are prognostically determined on the basis of their governing equations depending on the complexity of the scheme. The simplest high-order closure scheme is the so-called TKE scheme, also known 1.5-order scheme, in which turbulent eddy exchange coefficients are parameterized in terms of the prognostically or diagnostically determined TKE. (b) Local vs. nonlocal closure schemes. In local closure schemes, all turbulent fluxes are assumed to be associated with the down-gradient transport similar to molecular diffusion. All K-closure and high-order closure schemes are local closure schemes since they are formulated with the vertical gradient of variables. However, not all turbulent transport is induced by local down-gradient diffusion. Large turbulent eddies, such as convective plumes or cells, roll vortices, can yield non-local mixing that cannot be represented by the traditional method involving vertical gradient of variables. (c) “Dry” vs. “moist” closure schemes. A “dry” scheme refers to a scheme that is formulated with the non-conservative thermodynamic variables, such as, potential temperature, virtual potential temperature, and water vapor mixing ratio. In contrast, a “moist” closure scheme refers to a scheme that is formulated with the thermodynamic variables that are conserved for the moist adiabatic processes, such as, liquid water potential temperature and total water mixing ratio.

The operational HWRF uses the modified NCEP Global Forecast System (GFS) PBL schemes [Tallapragada *et al.* 2015], which is a typical “dry” K-closure turbulent mixing scheme. Although there have been modifications to the scheme throughout the years, the basic formulae used to determine eddy exchange coefficients are kept the same as those originally proposed by Hong and Pan (1996). Although there are many sophisticated methods to parameterize SGS turbulent mixing summarized previously, the K-closure scheme is arguably the best choice for operational models at the current stage as it requires the least computational resource. In the K-closure scheme, the eddy exchange coefficients are determined separately using a diagnosed PBL height estimated iteratively from the bulk Richardson number over the PBL depth and the buoyancy of surface-driven thermals. Within the PBL, the momentum eddy viscosity is calculated by a simple formula as function of height, friction velocity and static stability in the surface layer, and the diagnosed PBL height. Above the diagnosed PBL height, the turbulent eddy exchange coefficients are calculated as a function of wind shear and gradient Richardson number

defined as $Ri_g = \frac{g}{\theta_v} \frac{\partial \widetilde{\theta_v}}{\partial z} / S^2$, where $S = \sqrt{\left| \frac{\partial \widetilde{u}}{\partial z} \right|^2 + \left| \frac{\partial \widetilde{v}}{\partial z} \right|^2}$ is the vertical wind shear.

The Advanced Research WRF (WRF-ARW) has multiple options of PBL schemes. Table 2.1 summarizes the PBL schemes available in WRF-ARW and the basic information of how these schemes are formulated in terms of the three types of schemes classified previously. All these three types of PBL schemes will be examined in the sensitivity tests presented in this section. Since all these schemes are developed in non-TC conditions, it is interesting to see if more sophisticated PBL schemes perform better than the simple

K-closure schemes in capturing RI storms. If not, what is the main issue of PBL schemes in simulating TCs.

Table 2.1: WRF-ARW PBL schemes tested in the sensitivity experiments

PBL schemes	Order of closure	Local / nonlocal	Dry / Moist
Yonsei University Scheme (YSU)	First-order	Local	Dry
Mellor–Yamada–Janjic Scheme (MYJ)	Second-order	Local	Dry
Quasi–normal Scale Elimination (QNSE) Scheme	1.5-order	Local	Dry
Mellor–Yamada Nakanishi Niino (MYNN) Level 2.5 Scheme	Second-order	Local	Moist
Mellor–Yamada Nakanishi Niino (MYNN) Level 3.0 Scheme	Partially third-order	Local	Moist
Asymmetric Convection Model 2 Scheme (ACM2)	First-order	Local + nonlocal	Dry
Bougeault–Lacarrere Scheme (BouLac)	1.5-order	Local	Dry
University of Washington (TKE) Boundary Layer Scheme	1.5-order	Local	Moist
TEMF Scheme	1.5-order	Nonlocal	Dry
Grenier–Bretherton–McCaa Scheme	1.5-order	Local	Moist
MRF Scheme	First-order	Nonlocal	Dry

2.2 Microphysics Schemes

Microphysics is another key SGS process that needs to be parameterized in models. An appropriate representation of microphysical processes, such as condensation, riming, melting, evaporation, sublimation, autoconversion, accretion, and sedimentation, which directly involve in generating diabatic heating and cooling, is critical to the determination of the rate of TC intensification in models. A number of studies (e.g., Lord et al. 1984, Zhu and Zhang 2004, Wang 2002, Sawada and Iwasaki 2007) have shown the simulated storm

intensity is sensitive to the detailed representation of microphysics. Since hydrometeors have a wide spectrum of scales, a straightforward way to parameterize microphysical processes is to divide hydrometeor particles into a number of bins of different sizes and explicitly compute the evolution of particles in each bin separately, a method known as spectral bin microphysics. In a bin microphysical scheme, the change in particle size distribution (PSD) is directly predicted by the scheme. Consequently, spectral bin microphysics schemes are computationally too expensive for convection simulations. To reduce computational requirement, bulk microphysics schemes are often used in numerical simulations. In a bulk microphysics scheme, the details of cloud particles are not considered, rather, it predicts one or more bulk cloud quantities (e.g., cloud water mixing ratio and rain water mixing ratio) using the assumed PSD, such as, the beta distribution or the gamma distribution.

The simplest bulk microphysics scheme is the liquid microphysics in which hydrometeors are separated into cloud water and rain water. The basic microphysical processes that need to be considered are accretion of cloud water by existing rain water; autoconversion of cloud water to form rain water; growth of cloud water; and hydrometeor sedimentation. For convection exceeds freezing level, ice microphysics must be considered. The ice microphysics is important mainly because of the slower fall speeds of snow and ice flakes compared to rain droplets and extra latent heating (cooling) associated with freezing (melting). However, ice microphysics is much more complicated than liquid microphysics because of the wide variety of ice particle characteristics, such as pristine ice crystals grown by diffusion of water vapor; snowflakes grown by aggregation; rimed ice crystals grown by accretion of supercooled cloud droplets; graupel (heavily rimed ice

crystals); and hail. The basic ice microphysical processes need to be considered include: ice particle initiation (nucleation); freezing; melting; collection of rain and cloud water (riming); aggregation (autoconversion and accretion); diffusional growth/sublimation; and sedimentation. In practice, different ice hydrometeors are partitioned into different species whose characteristics (e.g., number if concentration, particle density, fall speeds) are determined a priori.

Table 2.2: WRF-ARW microphysics schemes tested in the sensitivity experiments

Microphysics schemes	Type	Mass variables (mixing ratio)	Number of concentration
Kessler Scheme	Single-moment	q_v, q_c, q_r	
Purdue Lin Scheme	Single-moment	$q_v, q_c, q_r, q_i, q_s, q_g$	
WRF 3-class Scheme	Single-moment	q_v, q_c, q_r	
WRF 5-class Scheme	Single-moment	q_v, q_c, q_r, q_i, q_s	
WRF 6-class Scheme	Single-moment	$q_v, q_c, q_r, q_i, q_s, q_g$	
Goddard Scheme	Single-moment	$q_v, q_c, q_r, q_i, q_s, q_g$	
Stony-Brook U. Scheme	Single-moment	$q_v, q_c, q_r, q_i, q_{ir}$	
NSSL 6-class Scheme	Single-moment	$q_v, q_c, q_r, q_i, q_s, q_g$	
NSSL 7-class Scheme	Single-moment	$q_v, q_c, q_r, q_i, q_s,$ q_g, q_h	
Thompson Scheme	1.5-moment	$q_v, q_c, q_r, q_i, q_s, q_g$	N_r, N_i
Milbrandt–Yau Scheme	Double-moment	$q_v, q_c, q_r, q_i, q_s, q_g,$ q_h	$N_c, N_r, N_i, N_s, N_g,$ N_h
Morrison Scheme	Double-moment	$q_v, q_c, q_r, q_i, q_s, q_g$	N_r, N_i, N_s, N_g
CAM 5.1 Scheme	Double-moment	$q_v, q_c, q_r, q_i, q_s, q_g$	N_r, N_s, N_i, N_g
WRF 5-class Scheme	Double-moment	q_v, q_c, q_r, q_i, q_s	N_{cnn}, N_c, N_r
WRF 6-class Scheme	Double-moment	$q_v, q_c, q_r, q_i, q_s, q_g$	N_{cnn}, N_c, N_r
NSSL Scheme	Double-moment	$q_v, q_c, q_r, q_i, q_s, q_g,$ q_h	$N_c, N_r, N_i, N_s, N_g,$ N_h
NSSL-CNN Scheme	Double-moment	$q_v, q_c, q_r, q_i, q_s,$ q_g, q_h	$N_c, N_r, N_i, N_s, N_g,$ N_h, N_{cnn}
HUJI SBM fast Scheme	Spectral-bin	$q_v, q_c, q_r, q_i, q_s, q_g$	$N_c, N_r, N_i, N_s, N_g,$ N_{cnn}

On the basis of the complexity, bulk microphysics schemes can be divided into single-moment and multi-moment schemes. A single-moment scheme only predicts mass variables of hydrometeors, such as cloud, rain, ice, snow, and graupel mixing ratios; whereas a multi-moment scheme not only predicts mass variables but also provides additional quantities for each species, such as, particle number of concentration, reflectivity, etc. For example, a two-moment scheme provides both mixing ratios and particle number of concentrations for each hydrometeor species. The Ferrier-Aligo microphysical scheme (Aligo et al. 2018) used in the HWRF is a single-moment scheme that predicts mixing ratios of water vapor, cloud water, rain water, and ice water. Table 2.2 lists a wide variety of microphysics schemes available in WRF-ARW from single-moment warm water microphysical scheme to sophisticated spectral bin microphysical schemes. These schemes will be tested in our sensitivity experiments.

2.3 Sensitivity Experiment Setup

The HWRF used in our study is version 3.9a (2018 operational model). The major feature of version 3.9a includes the use of the upgraded model grid spacing for domains 1 – 3 of 0.099, 0.033, and 0.011 degree, corresponding to approximately 11 km, 3.67 km, and 1.22 km respectively. The horizontal grid-meshes for the three domains are 390 x 780, 268 x 538, and 268 x 538, and there are 74 model levels in the vertical. In addition to the modified NCEP GFS PBL schemes (Tallapragada et al. 2015) and Ferrier-Aligo microphysical scheme (Aligo et al. 2018), other model physics packages include the Rapid Radiative Transfer Model (RRTMG) longwave/shortwave radiation, scale-aware Arakawa-Schubert cumulus scheme, and modified GFDL surface layer scheme.

The Advanced Research WRF (WRF-ARW) version 3.8.1 is used in our study. In addition to the PBL and microphysics schemes listed in Tables 2.1 and 2.2, other physics packages used in our study include the 2D Smagorinsky (1963) turbulence scheme for horizontal diffusion; the Rapid Radiative Transfer Model (RRTM) radiation scheme (Mlawer et al. 1997) and the Dudhia scheme (Dudhia 1989) for longwave and shortwave radiation respectively; Kain-Fritsch cumulus scheme (Kain 2004), which is only activated in domain 1; 5-layer thermal diffusion land-surface model (Dudhia 1996); and the revised MM5 surface layer scheme (Jimenez et al. 2012). Recent studies (e.g., Powell et al. 2003 and Donelan et al. 2004) showed the Charnock formula (Charnok 1955) is unable to depict the observed variation of drag coefficient over ocean surface in high wind conditions. As a consequence, the Charnock formula is replaced by the Davis formula (Davis et al. 2008) when estimating surface drag coefficients over oceans to match the findings of Donelan et al. (2004). These schemes are kept the same during the sensitivity tests except for the PBL scheme and microphysics scheme that are changed for individual experiment. All WRF-ARW simulations contain three two-way nested square domains with a horizontal resolution of 18, 6, 2 km, and grid-mesh of 301X301, 181X181, 313X313, respectively. There are 61 levels in the vertical with a model top approximately at 26 km (20 hPa). Vertical grids with relatively high vertical resolutions are concentrated in the PBL and then gradually stretched out with height to the model top.

Hurricane Patricia (2015) is simulated in our sensitivity test. The initial and lateral boundary conditions were supplied using three-hourly GFS data with 0.25° grid spacing. Although both HWRF and WRF-ARW were not run coupled to an ocean model, daily real-

time global (RTG) SST data were used for creating the surface conditions in the simulations. All simulations were run for a period of 78 hours from 06 UTC 21 October 2015 to 12 UTC 24 October 2015.

2.4 Results of Sensitivity Tests

The main purpose of the sensitivity tests is to examine if there is any WRF-ARW version that uses different PBL scheme and microphysics scheme can generate the better intensity forecast than the operational HWRF. Two sets of WRF-ARW experiments are performed in which all member simulations use the same model configuration and model physics. In the first set of experiments, all WRF-ARW simulations use the Morrison microphysics scheme (Morrison et al. 2009) but activate different PBL schemes in each member simulation; whereas in the second set of experiments, all WRF-ARW use the same MYJ PBL scheme (Janjic, 1994), but activate different microphysics scheme in each member simulation. Figure 2.1 shows the maximum surface wind speed, the minimum storm surface central pressure, and storm track of Patricia (2105) simulated by HWRF and WRF-ARW with different PBL schemes compared with the NHC best-track data. The results show that while there are inter-model differences in storm intensity among WRF-ARWs with different PBL schemes, none of the PBL schemes show noticeable improvements in capturing Patricia's RI regardless of different assumptions and complexities used in formulating these schemes. Similar results are shown in the microphysics sensitivity tests (Figure 2.2). Microphysics schemes with different complexity including the spectral bin microphysics (HUJI SBM fast Scheme, Khain et al. 2010) do not appear to have a noticeable improvement on the intensity simulation. Despite

the poor intensity simulations, all experiments predict fairly consistent tracks. Since storm movement is largely steered by the large-scale flow, it suggests that the environmental conditions are reasonably resolved by the simulations. The success in the storm track prediction excludes the possibility that the poor simulations of storm intensity are caused by the incorrect external forcing but suggests that the failure in capturing the Patricia's RI is likely attributed to model's inability to simulate the right storm internal dynamics that governs the storm intensification.

Figure 2.1: Storm surface central pressure, maximum surface wind speed, and track of Patricia (2015) simulated by HWRf and WRF-ARW with different PBL schemes compared with the NHC best-track data.

Figure 2.2: The same as Figure 2.1 but for WRF-ARW with different microphysics schemes.

To verify that models have problem to simulate the right TC internal dynamics, the simulated storm structures are examined. Figure 2.3 compares the simulated horizontal structure of Patricia (2015) with the satellite observations. The 91-GHz GPM image clearly shows that Patricia is nearly axisymmetric with a double eyewall structure prior to landfall. In contrast, all simulations with different PBL and microphysics schemes yield a highly asymmetric structure with condensate concentrated in the northeast sector and fail to produce the observed double eyewall structure.

Figure 2.3: Comparison of horizontal structure of Patricia (2015) between satellite observations and HWRF and WRF simulations. Left panel: image of 91-GHz of NASA Global Precipitation Mission (GPM) from NOAA polar-orbiting satellites showing Patricia with a double eyewall structure at 20:23 UTC 23 October prior to landfall along the southwestern coast of Mexico (adopted from the NHC TC report of Patricia). Middle panel: Composite 5-km altitude total condensate mixing ratio at 20:00 UTC 23 October from all runs of the PBL experiments. Right panel: The same as the mid-panel but for all runs of the microphysics experiments.

The comparison of radial wind profile of Patricia between observations and simulations at round 17:30 UTC when Patricia is near its peak intensity is shown in Figure 2.4. Both in-suit aircraft observations and retrievals from the stepped frequency microwave radiometer (SFMR) show that Patricia is a small hurricane in terms of the area covered by hurricane force winds. It has a narrow high wind zone and wind speed decreases sharply

with the increase of radius outward from its peak value. Since the average cruising speed of NOAA WP-3D is about 460 km/h, and it took about three minutes flying through from one peak to the other, such estimated radius of maximum wind (RMW) is only about 12 km. On the other hand, the RMW near Patricia's peak intensity from simulations is about twice larger than that of observations. These comparisons suggest that the significantly underestimated intensity of Patricia by numerical models may be attributed to the poor simulated storm structure that fails to generate the right TC internal dynamics required for the RI of Patricia (2015).

Figure 2.4: Surface wind speed profiles from observations and HWRf and WRF-ARW simulations. Left panel: Wind profiles obtained from the NOAA reconnaissance flights and the stepped frequency microwave radiometer (SFMR) retrievals along the radial flight legs passing through the eye around 17:00 UTC 23 October, 2015 (adopted from the NHC TC report of Patricia). Middle panel: Surface wind profiles along the cross section from southwest to northeast from all runs of the PBL experiment averaged between 17:00 UTC and 18:00 UTC 23 October. Right panel: The same as the mid-panel but for all runs of the microphysics runs.

In summary, numerical experiments on Hurricane Patricia are performed to test if any PBL and microphysics schemes can yield better intensity simulations by the WRF-ARW modeling system than the operational HWRf. Total 11 PBL schemes and 18 microphysics schemes are tested. While the storm track is reasonably simulated in all

sensitivity experiments, none of the PBL schemes and microphysics schemes show improvements on the intensity forecast. Since the environmental conditions are well resolved by the simulations in this case, the poorly simulated storm intensity may be attributed to the incorrect TC internal dynamics generated in simulations. Indeed, all simulations produce a TC structure far away from that obtained from satellite and aircraft observations. In the rest of this dissertation, problems associated with the PBL and microphysics schemes currently used for TC simulations will be investigated. Methods to mitigate the identified problem will be proposed. Based on the improved PBL and microphysics schemes, the pathway to TC intensification through microphysical-dynamical interaction will be explored.

3 Parameterization of Turbulent Processes in the Eyewall and Rainbands of TCs

As discussed in Chapter 1, the turbulent processes in the eyewall and rainbands of TCs have unique characteristics. They do not disappear above the PBL due to the buoyancy generated by clouds processes and there is no physical interface to separate the turbulence generated by the surface PBL processes and cloud processes aloft in the eyewall and rainbands. In this chapter, first, I will illustrate the problem associated with the PBL scheme used in the operational HWRF, then, I'll explore the underlying reason that causes the problem and present the method to remediate the problem, and finally I'll present the results of improved TC intensity simulations by the modified PBL scheme.

3.1 The PBL Scheme Used in the Operational HWRF

The HWRF PBL scheme is a typical K-closure (or first-order-closure) turbulent mixing scheme. Although there have been modifications to the scheme throughout the years, the basic formulae used to determine eddy exchange coefficients are kept the same as those originally proposed by Hong and Pan (1996). In this scheme, the eddy exchange coefficients are determined separately based on a diagnosed PBL height. Within the PBL, the momentum eddy viscosity is calculated as:

$$k_m = \kappa \frac{u_*}{\phi_m} \alpha z \left(1 - \frac{z}{h}\right)^2, \quad (3.1)$$

where κ is the von Karman constant, u_* is the friction velocity, z is the height above the ground surface, ϕ_m is the surface layer stability function obtained by Businger et al. (1971), and h is a diagnosed PBL height calculated iteratively based on the bulk Richardson number over the PBL depth and the buoyancy of surface-driven thermals. Although there are many sophisticated methods to parameterize SGS turbulent mixing, such as TKE closure, high-order closure, nonlocal mixing, and schemes formulated using variables conserved for moist reversible adiabatic processes, the K-closure scheme is arguably the best choice for operational models at the current stage as it requires the least computational resource. Above the diagnosed PBL height, the momentum eddy viscosity is calculated as:

$$k_m = l^2 f_m(R_i) \sqrt{\left|\frac{\partial \bar{u}}{\partial z}\right|^2 + \left|\frac{\partial \bar{v}}{\partial z}\right|^2}, \quad (3.2)$$

where l is the mixing length, $f_m(R_i)$ is the stability function of Richardson number R_i , and vertical wind shear. Once k_m is determined, the eddy viscosity for heat and moisture is calculated by $k_{t,q} = k_m P_r^{-1}$, where P_r is the Prandtl number.

For fair-weather conditions, the parameterization formulated by Eq. (3.1) and Eq. (3.2) provides a practical way to appropriately parameterize the SGS turbulent mixing within and above the PBL since the turbulent layer resulting from the surface processes is often cleanly separated from the free atmosphere by a capping inversion. The mid-point of the inversion zone (or entrainment zone) is naturally defined as the PBL height (Stull 1988). In a TC environment, as stated previously, turbulence is no longer solely generated by the shear production and buoyancy production associated with the surface processes; it can also be generated by cloud processes aloft due to cloud radiative cooling, evaporative cooling, and inhomogeneous diabatic heating and cooling in the clouds. Thus, although the concept of PBL is still applicable, it becomes ambiguous from the turbulent mixing perspective. In many TC studies, the PBL is defined either as the turbulent layer that is directly affected by the surface processes or as the inflow layer of the secondary circulation. But in either case, the so-defined PBL height is by no means a physical interface that separates the turbulence generated by surface processes and by cloud processes. This is particularly true in the eyewall and rainbands of a TC, where intense turbulence can extend from the surface all the way up to the upper troposphere, as was illustrated in Fig. 1.3. Thus, from the nature of turbulent mixing, an artificial separation of turbulence using a diagnosed “PBL” height is not a physically sound method to parameterize the internally connected SGS turbulent mixing in the eyewall or any deep convective areas in a TC. Moreover, an artificial separation of the PBL from the free atmosphere above may create an unrealistic discontinuity in the vertical profile of eddy viscosity in this method. According to Eq. (3.1), as height z approaches the diagnosed “PBL” height h , eddy viscosity K_m becomes zero to result in zero turbulent mixing at a certain model grid level if the diagnosed PBL height

falls exactly at this level. Above the diagnosed PBL, the turbulent mixing jumps to whatever value estimated by Eq. (3.2). This singular point in the vertical profile of eddy exchange coefficient is unrealistic and could cause serious problems.

To diagnose the problem, the eddy exchange coefficients in TC simulations by the operational HWRF are examined. It is found that the default PBL scheme is unable to generate intense turbulent mixing above the PBL in the eyewall and rainbands. As an example, Figure 3.1 shows the horizontal distribution of the HWRF simulated eddy exchange coefficients for momentum (k_m) at different altitudes of Hurricane Patricia (2015) and the corresponding azimuthal-mean of k_m between the radii of 19 km -2 km where the eyewall is located at 06:00 UTC October 23, 2015 when Patricia (2015) underwent its RI. Within the PBL, the magnitude and horizontal spatial distribution reflects well the strong turbulent mixing in the eyewall and rainbands, but above the PBL, the HWRF generated eddy exchange coefficients are virtually zero. This result suggests that the PBL scheme used in the operational HWRF fails to capture the intense turbulent mixing above the PBL in the deep convective eyewall and rainbands. I hypothesize that the lack of appropriate SGS eddy forcing associated with deep convection above the PBL in the eyewall and rainbands is one of the culprits for the intensity forecast failure by HWRF in this case.

Figure 3.1: (a) – (d): Horizontal distribution of HWRF simulated eddy exchange coefficients of momentum (k_m) at the altitudes of $z = 0.5, 1.0, 2.5$, and 5.0 km respectively at 06:00 UTC, October 23, 2015 when Patricia underwent its RI. (e): The corresponding azimuthal-mean k_m averaged over the radii of $19 - 25$ km where the eyewall is located.

As indicated by Eq. (3.2), the turbulent eddy exchange coefficients above the PBL is parameterized based on the Richardson number defined as $R_i = N^2/S^2$, where $S = \sqrt{\left|\frac{\partial \bar{u}}{\partial z}\right|^2 + \left|\frac{\partial \bar{v}}{\partial z}\right|^2}$ is the vertical wind shear and N^2 is the Brunt-Väisälä frequency (BVF), which measures the static stability of the atmosphere. R_i provides an ideal predictor of the expected atmospheric turbulence as it includes both buoyancy production and mechanical (shear) production of turbulence. Since vertical wind shear can be readily obtained from either observations and numerical simulations, a correct determination of R_i largely

depends on how BVF is estimated. In many numerical models including HWRF, BVF is

simply calculated as, $N^2 = \frac{g}{\theta_v} \frac{\partial \theta_v}{\partial z}$, where θ_v is the virtual potential temperature.

Figure 3.2: BVF estimated as $N^2 = \frac{g}{\theta_v} \frac{\partial \theta_v}{\partial z}$ at different altitudes at 06:00 UTC October 23, 2015 from the HWRF simulation of Patricia (2015).

Figure 3.3: R_i at different altitudes at 06:00 UTC October 23, 2015 from the HWRF simulation of Patricia (2015).

To diagnose the problem, the BVF and the corresponding R_i from the HWRF simulation of Patricia (2015) are examined. As an illustration, Figures 3.2 and 3.3 show the BVF and R_i at different altitudes above the PBL at 06:00 UTC October 23, 2015. The results show that such calculated BVF causes the simulated eyewall and rainbands above the PBL to be static too stability to generate turbulence in the convective eyewall and rainband clouds. This is clearly seen from the simulated R_i that is well over its critical value for turbulence generation. Moreover, the R_i distribution does not reflect the TC inner-core

structures well, such as the eyewall and rainbands, indicating that such as calculated BVF and R_i do not represent the static and dynamic instability well in the eyewall and rainbands. The reason is that the BVF calculated by $N^2 = \frac{g}{\theta_v} \frac{\partial \theta_v}{\partial z}$ does not consider the buoyancy generated by the clouds. In the rest of this chapter, I'll discuss how to appropriately determine the static stability of the eyewall and rainbands.

Quantitatively, static stability of the atmosphere may be measured by the vertical acceleration of an air parcel in the environment in which the parcel is embedded, $a = g \left(\frac{\rho^e - \rho^p}{\rho^p} \right)$, where ρ is the air density and g is the gravitational acceleration. Superscripts “ p ” and “ e ” are reserved in this dissertation to denote air parcel and environment respectively hereafter. Applying the ideal gas law, $p = \rho R_d T_v$, where p is the air pressure, T_v is the virtual temperature, and R_d is the gas constant for dry air, and assuming that an air parcel undergoes an adiabatic process, which requires the pressure of the air parcel to be the same as that of the environment (i.e., $p^p = p^e$), the acceleration of the air parcel may be rewritten as $a = g \left(\frac{T_v^p - T_v^e}{T_v^e} \right)$. Considering a small vertical displacement δz of the air parcel from its equilibrium, says, $z = 0$ where $T_v^p(0) = T_v^e(0)$, the virtual temperature of the air parcel and environment at height $z = \delta z$ may be expressed as $T_v^p = T_v^p(0) + \frac{\partial T_v^p}{\partial z} \delta z$ and $T_v^e = T_v^e(0) + \frac{\partial T_v^e}{\partial z} \delta z$. This yields $a = -N^2 \delta z$, where

$$N^2 = \frac{g}{T_v^e} \left(\frac{\partial T_v^e}{\partial z} - \frac{\partial T_v^p}{\partial z} \right) \quad (3.3)$$

is known as the BVF. While conceptually it is easy to understand, an accurate determination of BVF using Eq. (3.3) is not a trivial thermodynamic problem. In the next two sections, I'll discuss how to calculate BVF in the unsaturated and saturated atmosphere.

3.2 BVF in the Unsaturated Atmosphere

To calculate BVF using Eq. (3.3), one needs to define virtual temperature T_v . For the unsaturated atmosphere, T_v may be expressed as,

$$T_v = T \left(1 + \frac{1-\epsilon}{\epsilon} q_v \right) = T(1 + 0.608q_v), \quad (3.4)$$

where q_v is the water vapor mixing ratio, and ϵ is the ratio of gas constant of dry air R_d to gas constant of water vapor R_v . Then, the adiabatic lapse of virtual potential temperature of an air parcel in Eq. (3.3) may be written as,

$$-\frac{\partial T_v^p}{\partial z} = -\frac{\partial}{\partial z} \left[T^p \left(1 + \frac{1-\epsilon}{\epsilon} q_v^p \right) \right] = \left(1 + \frac{1-\epsilon}{\epsilon} q_v^p \right) \Gamma_d \approx \Gamma_d, \quad (3.5)$$

where $\Gamma_d = -\frac{\partial T^p}{\partial z} = \frac{g}{C_{pd}}$ is the dry adiabatic lapse of temperature. Note that to obtain Eq. (3.5), we have used the assumption that water vapor mixing ratio is conserved during an adiabatic process. Thus, Eq. (3.3) may be rewritten as,

$$N^2 = \frac{g}{T_v^e} \left(\frac{\partial T_v^e}{\partial z} + \frac{g}{C_{pd}} \right). \quad (3.6)$$

Defining virtual potential temperature as $\theta_v = T_v \left(\frac{p_0}{p} \right)^{R_d/C_{pd}}$, where p_0 is a reference pressure, usually taken as 1000 hPa, and C_{pd} is the specific heat of dry air at the constant pressure, then, applying it to the environment and taking its derivative with respect to height z , it yields,

$$\frac{1}{\theta_v^e} \frac{\partial \theta_v^e}{\partial z} = \frac{1}{T_v^e} \frac{\partial T_v^e}{\partial z} - \frac{R_d}{c_p} \frac{\partial p^e}{p^e \partial z} = \frac{1}{T_v^e} \left(\frac{\partial T_v^e}{\partial z} + \frac{g}{c_{pd}} \right). \quad (3.7)$$

Note that to obtain Eq. (3.7), we have assumed that the environment is in the hydrostatic balance $\frac{\partial p^e}{\partial z} = -\rho^e g$ and used the ideal gas law. Thus, BVF for the unsaturated atmosphere can be expressed as,

$$N^2 = \frac{g}{\theta_v^e} \frac{\partial \theta_v^e}{\partial z}. \quad (3.8)$$

For the pure dry atmosphere, Eq. (3.8) reduces to $N^2 = \frac{g}{\theta^e} \frac{\partial \theta^e}{\partial z}$, where θ^e is the potential temperature of the environment. As stated previously, Eq. (3.8) is the one used in various meteorological applications including the PBL schemes used HWRF and WRF-ARW to estimate the static stability of the atmosphere. But it is not appropriate to use Eq. (3.8) to represent the static stability in the saturated eyewall and rainbands since it fails to consider the latent heating during water phase change. As a result, it significantly overestimates the static stability in eyewall and rainband clouds as it is seen in Figures 3.2 and 3.3.

3.3 BVF in the Saturated Atmosphere

When an air parcel becomes saturated, Eq. (3.8) is no longer valid since $-\frac{\partial T_v^p}{\partial z}$ deviates substantially from Γ_d owing to the latent heating release that heats the saturated air parcel. In this case, one needs to correctly calculate the virtual temperature of the saturated atmosphere, which can be derived directly from the ideal gas law as follows,

$$\rho = \frac{M_d}{V} + q_s \frac{M_d}{V} + q_c \frac{M_d}{V} = \frac{M_d}{V} (1 + q_t) = \rho_d (1 + q_t) = \frac{p-E}{R_d T} (1 + q_t), \quad (3.9)$$

where V is the volume of the saturated atmosphere; q_s , q_c , and q_t are the mixing ratios of the saturated water vapor, condensate, and total water defined as $q_s = \frac{M_v}{M_d}$, $q_c = \frac{M_c}{M_d}$, and $q_t = \frac{M_v + M_c}{M_d}$, respectively; M_c is the mass of condensate; E is the saturated water vapor pressure. Applying the ideal gas law to dry air and water vapor respectively, it is easy to show $q_s = \frac{\epsilon E}{p - E}$, then, Eq. (3.9) can be rewritten as, $p = \rho R_d T \frac{1 + \frac{q_s}{\epsilon}}{1 + q_t}$. It is clear that virtual temperature of the saturated atmosphere can be defined as,

$$T_v = T \frac{1 + \frac{q_s}{\epsilon}}{1 + q_t}. \quad (3.10)$$

In many meteorological applications, Eq. (3.10) is often simplified to the well-known formula, $T_v \approx T(1 + \frac{q_s}{\epsilon})(1 - q_t) \approx \left[1 + \left(\frac{1}{\epsilon} - 1\right)q_s - q_c\right] = T(1 + 0.608q_s - q_c)$. Here, we'd like to use Eq. (3.10) since it can yield simpler mathematical expressions of BVF. With Eq. (3.10), Eq. (3.3) may be written as,

$$N^2 = g \left\{ \frac{\partial}{\partial z} \left[\ln \left(T^e \frac{1 + \frac{q_s^e}{\epsilon}}{1 + q_t^e} \right) \right] - \frac{T_v^p}{T_v^e} \frac{\partial}{\partial z} \left[\ln \left(T^p \frac{1 + \frac{q_s^p}{\epsilon}}{1 + q_t^p} \right) \right] \right\}. \quad (3.11)$$

Note that the total water of an air parcel is conserved during an adiabatic process, i.e.,

$\frac{\partial q_t^p}{\partial z} = 0$, Eq. (3.11) may be simplified as,

$$N^2 = g \left\{ \left[\frac{1}{T^e} \frac{\partial T^e}{\partial z} + \frac{1}{(\epsilon + q_s^e)} \frac{\partial q_s^e}{\partial z} - \frac{1}{(1 + q_t^e)} \frac{\partial q_t^e}{\partial z} \right] - \frac{T_v^p}{T_v^e} \left[\frac{1}{T^p} \frac{\partial T^p}{\partial z} + \frac{1}{(\epsilon + q_s^p)} \frac{\partial q_s^p}{\partial z} \right] \right\}. \quad (3.12)$$

$\frac{\partial q_s}{\partial z}$ in Eq. (3.12) may be calculated from the definition of q_s and the Clausius-Clapeyron

equation, $\frac{1}{E} \frac{\partial E}{\partial T} = \frac{l_v}{R_v T^2}$, where l_v is the specific latent heat of vaporization. From $q_s = \frac{\epsilon E}{p - E}$,

it is ready to show,

$$\frac{\partial q_s}{\partial z} = q_s \left(1 + \frac{q_s}{\epsilon} \right) \left(\frac{l_v}{R_v T^2} \frac{\partial T}{\partial z} - \frac{1}{p} \frac{\partial p}{\partial z} \right). \quad (3.13)$$

Applying Eq. (3.13) to the air parcel and environment respectively and inserting them into Eq. (3.12), it yields,

$$N^2 = g \left[\left(1 + \frac{l_v q_s^e}{R_d T^e} \right) \frac{1}{T^e} \frac{\partial T^e}{\partial z} - \frac{q_s^e}{\epsilon p^e} \frac{\partial p^e}{\partial z} - \frac{1}{(1+q_t^e)} \frac{\partial q_t^e}{\partial z} \right] - g \frac{T_v^p}{T_v^e} \left[\left(1 + \frac{l_v q_s^p}{R_d T^p} \right) \frac{1}{T^p} \frac{\partial T^p}{\partial z} - \frac{q_s^p}{\epsilon p^p} \frac{\partial p^p}{\partial z} \right]. \quad (3.14)$$

During an adiabatic process, the pressure of an air parcel is the same as that of the environment and the environment is in the hydrostatic balance, i.e., $\frac{\partial p^p}{\partial z} = \frac{\partial p^e}{\partial z} = -\frac{p^e}{R_d T_v^e} g$.

Then, Eq. (3.14) becomes,

$$N^2 = g \left[\left(1 + \frac{l_v q_s^e}{R_d T^e} \right) \frac{1}{T^e} \frac{\partial T^e}{\partial z} - \frac{T_v^p}{T_v^e} \left(1 + \frac{l_v q_s^p}{R_d T^p} \right) \frac{1}{T^p} \frac{\partial T^p}{\partial z} + \frac{(q_s^e - q_s^p)g}{\epsilon R_d T_v^e} - \frac{1}{(1+q_t^e)} \frac{\partial q_t^e}{\partial z} \right]. \quad (3.15)$$

Using the approximations $\left(1 + \frac{l_v q_s^e}{R_d T^e} \right) \frac{1}{T^e} \approx \frac{T_v^p}{T_v^e} \left(1 + \frac{l_v q_s^p}{R_d T^p} \right) \frac{1}{T^p}$ and $\frac{q_s^e g}{\epsilon R_d T_v^e} \approx \frac{q_s^p g}{\epsilon R_d T_v^e}$, Eq. (3.15)

simplifies to,

$$N^2 \approx g \left\{ \left(1 + \frac{l_v q_s^e}{R_d T^e} \right) \frac{1}{T^e} \left(\frac{\partial T^e}{\partial z} + \Gamma_m \right) - \frac{1}{1+q_t^e} \frac{\partial q_t^e}{\partial z} \right\}, \quad (3.16)$$

where $\Gamma_m = -\frac{\partial T^p}{\partial z}$ is the moist adiabatic lapse rate of temperature. Thus, the determination of Γ_m becomes a key to the success of estimating static stability of the saturated atmosphere.

While there are many rough estimates of Γ_m available in literature and textbooks, Here following Bohren and Albrecht (1998), I provide a derivation of Γ_m from the conservation of the total entropy of a saturated air parcel, which may be written as,

$$\frac{\partial H^p}{\partial z} = 0, \quad H^p = M_d^p H_d^p + M_v^p H_v^p + M_c^p H_c^p, \quad (3.17)$$

where M_d^p , H_d^p , M_v^p , H_v^p , M_c^p , and H_c^p are the mass and entropy of dry air, water vapor, and condensate of an air parcel, respectively. For an adiabatic process, the total water mass and dry air mass are conserved, i.e., $\frac{\partial(M_v^p + M_c^p)}{\partial z} = 0$ and $\frac{\partial M_d^p}{\partial z} = 0$. With the definition of $q_s = \frac{M_v}{M_d}$ and $q_c = \frac{M_c}{M_d}$, Eq. (3.17) becomes,

$$(H_v^p - H_c^p) \frac{\partial q_s^p}{\partial z} + \frac{\partial H_d^p}{\partial z} + q_s^p \frac{\partial H_v^p}{\partial z} + q_c^p \frac{\partial H_c^p}{\partial z} = 0, \quad (3.18)$$

Applying the entropy change during the phase change, $T^p(H_v^p - H_c^p) = l_v$, it yields,

$$\frac{l_v}{T^p} \frac{\partial q_s^p}{\partial z} + \frac{\partial H_d^p}{\partial z} + q_s^p \frac{\partial H_v^p}{\partial z} + q_c^p \frac{\partial H_c^p}{\partial z} = 0. \quad (3.19)$$

From the first law of thermodynamics, it is readily to show that the entropy change of ideal dry air and saturated water vapor can be written as,

$$\begin{cases} \frac{\partial H_d^p}{\partial z} = \frac{C_{pd}}{T^p} \frac{\partial T^p}{\partial z} - \frac{R_d}{(p^p - E^p)} \frac{\partial (p^p - E^p)}{\partial z} \\ \frac{\partial H_v^p}{\partial z} = \frac{C_{pv}}{T^p} \frac{\partial T^p}{\partial z} - \frac{R_v}{E^p} \frac{\partial E^p}{\partial z} \end{cases}, \quad (3.20)$$

where C_{pv} is the specific heat of water vapor at the constant pressure. The entropy change of condensate may be written as,

$$\frac{\partial H_c^p}{\partial z} = \frac{C_c}{T^p} \frac{\partial T^p}{\partial z}, \quad (3.21)$$

where C_c is the specific heat of condensate. Inserting Eq. (3.20) and Eq. (3.21) in Eq. (3.19),

and then, grouping all terms containing $\frac{\partial T^p}{\partial z}$ together, it yields,

$$\frac{l_v}{T^p} \frac{\partial q_s^p}{\partial z} - \frac{R_d}{(p^p - E^p)} \frac{\partial (p^p - E^p)}{\partial z} - \frac{R_v q_s^p}{E^p} \frac{\partial E^p}{\partial z} + \frac{(C_{pd} + C_{pv} q_s^p + C_c q_c^p)}{T^p} \frac{\partial T^p}{\partial z} = 0. \quad (3.22)$$

Using $q_s = \frac{E}{p - E}$, Eq. (3.22) can be further simplified to,

$$\frac{l_v}{T^p} \frac{\partial q_s^p}{\partial z} - \frac{R_d}{(p^p - E^p)} \frac{\partial p^p}{\partial z} + \frac{(C_{pd} + C_{pv} q_s^p + C_c q_c^p)}{T^p} \frac{\partial T^p}{\partial z} = 0. \quad (3.23)$$

Applying Eq. (3.13) to the air parcel and inserting it in the first term of Eq. (3.23) and regrouping all terms containing $\frac{\partial T^p}{\partial z}$, it yields,

$$\left[\frac{c_{pd} + c_{pv}q_s^p + c_c q_c^p + \frac{(\epsilon + q_s^p)l_v^2 q_s^p}{R_d T^{p2}}}{T^p} \right] \frac{\partial T^p}{\partial z} - \frac{(\epsilon + q_s^p)l_v q_s^p}{\epsilon T^p p^p} \frac{\partial p^p}{\partial z} - \frac{R_d}{(p^p - E^p)} \frac{\partial p^p}{\partial z} = 0. \quad (3.24)$$

For an adiabatic processes, the pressure of an air parcel is the same as that of environment

and the environment is in a hydrostatic balance, i.e., $\frac{\partial p^p}{\partial z} = \frac{\partial p^e}{\partial z} = -\rho^e g$. Further using the

ideal gas law, $p^e = \rho^e R_d T_v^e$; the definition of T_v ; and $q_s = \frac{\epsilon E}{p - E}$, the last two terms of Eq.

(3.24) can be simplified as, $-\frac{(\epsilon + q_s^p)l_v q_s^p}{\epsilon T^p p^p} \frac{\partial p^p}{\partial z} - \frac{R_d}{(p^p - E^p)} \frac{\partial p^p}{\partial z} \approx \frac{(1 + q_t^e)g}{T^e} \left(1 + \frac{l_v q_s^p}{R_d T^p} \right)$, where the

approximation $\left(1 + \frac{q_s^p}{\epsilon} \right) \approx \left(1 + \frac{q_s^e}{\epsilon} \right)$ was used. Then, from Eq. (3.24), it is easy to show

that Γ_m may be expressed as,

$$\Gamma_m = -\frac{\partial T^p}{\partial z} = \frac{g}{c_{pd}} \cdot \frac{(1 + q_t^e) \left(1 + \frac{l_v q_s^p}{R_d T^p} \right)}{1 + \frac{c_{pv} q_s^p + c_c q_c^p + \frac{(\epsilon + q_s^p)l_v^2 q_s^p}{R_d T^{p2}}}{c_{pd}}}. \quad (3.25)$$

Eq. (3.25) is similar to that of Durran and Klemp (1982, c.f., Eq. 19 in their paper).

However, we explicitly distinguish the variables of the air parcel and environment. This is important for understanding the underlying physics and assumptions made during the derivation.

3.4 Stability Correction in Existing PBL Schemes

As stated previously, PBL schemes can be classified into “wet” and “dry” schemes.

The “wet” PBL schemes are formulated with the variables conserved for moist adiabatic processes, such as total mixing ratio and liquid water potential temperature. The schemes

developed by Bretherton et al. (2009) and Nakanishi and Niino (2009) are the examples of such “moist” turbulent mixing schemes. Since the schemes are formulated with the variables conserved for moist adiabatic processes, presumably they can appropriately account for the buoyancy induced by clouds. However, there are limitations of this method. First, these schemes are ideal for treating shallow non-precipitation PBL clouds, such as fair-weather cumulus and stratocumulus, but they will have problems to account for the buoyancy induced by the precipitating clouds since liquid water potential temperature is not really conserved when water falls out as rain. Thus, it is not an ideal method for treating convective clouds in the eyewall and rainbands. Second, since the basic structure of a “dry” scheme is different from that of a “wet” scheme, a significant computer coding is needed when converting an existing “dry” scheme into a “moist” scheme. Thus, for a “dry” scheme, such as the HWRF PBL scheme, correct formulae are needed to explicitly account for the buoyancy induced by clouds.

Most PBL schemes examined in Chapter 2 are “dry” schemes in which all of them simply use $\frac{g}{\theta_v^e} \frac{\partial \theta_v^e}{\partial z}$ to estimate static stability except the YSU scheme (Hong et al. 2006), which is often used in TC simulations. The YSU scheme uses the following expression (cf. Eq. A16b in Hong et al. 2006) to represent the Richardson number in clouds,

$$R_i \approx (1 + B) \left[\frac{g}{\theta_v} \frac{\partial \theta_v}{\partial z} - \frac{g^2}{c_p T} \left(\frac{A-B}{1+A} \right) \right] / S^2, \quad (3.26)$$

where $A = \frac{L^2 q_s}{c_p R_v T^2}$ and $B = \frac{L q_s}{R T}$. However, the assumptions that YSU made to obtain Eq. (3.26) may not be appropriate. Let’s start with Eq. (3.25). With the defined parameters A and B, it may be rewritten as,

$$\Gamma_m = \Gamma_d \frac{(1+q_t)(1+B)}{1 + \frac{C_{pv}q_s + C_wq_L}{C_p} + A(1 + \frac{q_s}{\varepsilon})} \quad (3.27)$$

With the first assumption that $\frac{C_{pv}q_s + C_wq_L}{C_p} \ll 1$ and $\frac{q_s}{\varepsilon} \ll 1$, Eq. (3.27) becomes,

$$\Gamma_m = \frac{g}{C_p} \frac{(1+q_t)(1+B)}{1+A}. \quad (3.28)$$

Inserting Eq. (3.28) into Eq. (3.16), it yields,

$$N^2 = g \left\{ (1+B) \frac{1}{T} \left[\frac{\partial T}{\partial z} + \frac{(1+B)(1+q_t)g}{(1+A)C_p} \right] - \frac{1}{1+q_t} \frac{\partial q_t}{\partial z} \right\}. \quad (3.29)$$

YSU further assumed that $q_t \ll 1$ and $\frac{\partial q_t}{\partial z}$ is negligible, then, Eq. (3.29) becomes,

$$N^2 = g \left\{ (1+B) \frac{1}{T} \left[\frac{\partial T}{\partial z} + \frac{(1+B)g}{(1+A)C_p} \right] \right\}. \quad (3.30)$$

Using the relation $\frac{1}{\theta} \frac{\partial \theta}{\partial z} = \frac{1}{T} \frac{\partial T}{\partial z} + \frac{g}{TC_p}$, Eq. (3.30) may be rewritten as,

$$N^2 = g(1+B) \left\{ \frac{1}{\theta} \frac{\partial \theta}{\partial z} - \frac{A-B}{1+A} \frac{g}{C_p T} \right\}. \quad (3.31)$$

YSU simply replaced $\frac{1}{\theta} \frac{\partial \theta}{\partial z}$ with $\frac{1}{\theta_v} \frac{\partial \theta_v}{\partial z}$, then, dividing Eq. (3.31) by the square of wind shear

S^2 , it yields Eq. (3.26), a formula that YSU used to compute R_i in saturated conditions.

Now let's make a rough estimate. For θ_v in a range of $\theta_v \propto [300 \text{ K}, 240 \text{ K}]$, the parameters

A and B should be in a range of $A \propto [3.69, 0.06]$ and $B \propto [0.71, 0.01]$, then, term

$\frac{g^2}{C_p T} \left(\frac{A-B}{1+A} \right)$ would be in a range of $\frac{g^2}{C_p T} \left(\frac{A-B}{1+A} \right) \propto [2.02, 0.19] \times 10^{-4}$. In the eyewall, $\frac{1}{\theta_v} \frac{\partial \theta_v}{\partial z}$

is typically in a range of $\frac{1}{\theta_v} \frac{\partial \theta_v}{\partial z} \propto [-1, 1] \times 10^{-4}$, then, it is easy to see that the YSU

correction significantly over-reduces the static stability in clouds! Because of this, it

generates unrealistically large eddy exchange coefficients in clouds, $k_{m,t}$. What YSU did

is to artificially reduce $k_{m,t}$ by averaging in-cloud eddy exchange coefficients and the entrainment eddy exchange coefficients at the top of the PBL via,

$$k_{m,t} = \sqrt{k_{m,t}^{cld} \cdot k_{m,t}^{ent}}, \quad (3.32)$$

where $k_{m,t}^{ent}$ is parameterized as,

$$k_{m,t}^{ent} = Pr_{m,t} \frac{-\overline{w'\theta'_{v,h}}}{(\frac{\partial\theta_v}{\partial z})_h} \exp \left[-\frac{(z-h)^2}{\delta^2} \right], \quad (3.33)$$

where $Pr_{m,t}$ is the Prandtl number, h is the PBL height, δ is the depth of the entrainment zone at the top of the PBL, which is usually 10 % of h , $-\overline{w'\theta'_{v,h}}$ is the entrainment buoyancy flux, and $(\frac{\partial\theta_v}{\partial z})_h$ is the vertical gradient across the entrainment zone. As indicated by Eq. (3.33), the Dirac delta function causes $k_{m,t}^{ent}$ to decrease dramatically away from the entrainment zone. In this way, it forces to reduce $k_{m,t}$ above and below the diagnosed PBL height via Eq. (3.32). But such a method was not developed based on sound physics. First, the averaging method of Eq. (3.32) and the Dirac delta function in Eq. (3.33) are purely from a mathematical consideration. Second, as discussed previously, there is no inversion existing in the eyewall and rainbands to separate the turbulence generated by the PBL processes and cloud processes aloft. Thus, the entrainment process at the top of the PBL, which plays a key role in the development of PBL in fair-weather conditions, is minor in the eyewall and rainbands.

In short, while individual assumptions that lead to Eq. (3.26) may be legitimate, the accumulated error resulting from the assumptions overwhelm the true static stability in clouds, which causes Eq. (3.26) to be inappropriate for estimating the stability in clouds.

In fact, such assumptions are not needed since static stability in clouds can be accurately determined by Eq. (3.16) and Eq. (3.25).

3.5 Static Stability in the Eyewall and Rainbands

With Eq. (3.16) and Eq. (3.25), the BVF in convective eyewall and rainband clouds may be appropriately determined since all variables in the formulae are available in numerical simulations. However, the calculation is not as straightforward as BVF calculated as $\frac{g}{\theta_v^e} \frac{\partial \theta_v^e}{\partial z}$ because the formulae involve variables of both the air parcel and environment. Recall that BVF is defined as a vertical derivative of virtual temperature difference between the environment and an air parcel (Eq. 3.3) and a real convective plume does not follow the strict adiabatic assumption throughout the troposphere. Thus, an appropriate way to calculate BVF using Eq. (3.16) & Eq. (3.25) is to divide an entire vertical column into a number of small segments based on the data points. Eq. (3.16) & Eq. (3.25) are, then, applied to each data point along a vertical profile. In other words, the data points along a vertical profile are considered as a series of equilibrium points where the parcel's thermodynamic properties equal to those of environment, and an air parcel in each segment is assumed to follow an individual adiabatic process from its equilibrium. With this assumption, BVF can be readily calculated from the model output.

Figure 3.4: The same as Figure 3.2 but for the BVF estimated using Eq. (3.16) and Eq. (3.25) that consider the buoyancy induced by clouds. To help distinguish the positive and negative BVF, the contours with zero value (black contours) are added in the figures.

Figure 3.5: The same as Figure 3.3 but for R_i estimated using Eq. (3.16) and Eq. (3.25) that consider the buoyancy induced by clouds. The black contours indicate the value of 0.25, which is assumed to be the critical Richardson number for turbulence generation.

Figures 3.4 and 3.5 show N^2 and R_i calculated by Eq. (3.16) & Eq. (3.25) at the same time and same height as those of Figures 3.2 and 3.3. The consideration of cloud induced buoyancy substantially reduces the overestimated static stability by $\frac{g}{\theta_v^e} \frac{\partial \theta_v^e}{\partial z}$. The horizontal distribution of negative N^2 and R_i below the critical Richardson number matches well with the eyewall and rainbands, suggesting that the static and dynamic

stability in the eyewall and rainbands reasonably estimated by the formulae derived following the thermodynamic law.

Figure 3.6: (a) – (d): Composite profiles of azimuthal-mean temperature, mixing ratios of water vapor and condensate, and wind shear averaged over the radii from 19 to 25 km where the eyewall is located during the RI (from 12 UTC 22 to 23 UTC 23 October, 2015) of Hurricane Patricia (2015) simulated by HWRF. (e) and (f): N^2 and R_i calculated by $\frac{g}{\theta_v^e} \frac{\partial \theta_v^e}{\partial z}$ and by Eq. (3.16) & Eq. (3.25) for the saturated atmosphere. The gray vertical line indicates the zero value.

To further illustrate the impact of clouds on the static stability in the eyewall, the vertical profiles of BVF and R_i are examined. Figures 3.6a – 3.6d show the composite vertical profiles of azimuthal-mean temperature, mixing ratios of water vapor and condensate, and vertical wind shear averaged over the radii from 19 to 25 km where the eyewall is located during the RI (from 12 UTC 22 to 23 UTC 23 October, 2015) of Patricia

(2015) simulated by HWRF. These composite soundings are taken as the representative of the ambient thermodynamic condition in the eyewall. The BVF and the corresponding R_i calculated using $\frac{g}{\theta_v^e} \frac{\partial \theta_v^e}{\partial z}$ and Eq. (3.16) & Eq. (3.25) are compared. The results are shown in Figures 3.6e & 3.6f, respectively. The static stability estimated by $\frac{g}{\theta_v^e} \frac{\partial \theta_v^e}{\partial z}$ causes the simulated eyewall to be statically too stable to generate any turbulent mixing in moist convection. The consideration of the buoyancy induced by clouds significantly reduces the stability in the eyewall and rainbands so that turbulence can be appropriately generated through buoyancy production in the convective clouds.

3.6 Impact of Static Stability on Intensity Forecast

We have implemented the BVF calculation for the saturated atmosphere by Eq. (3.16) & Eq. (3.25) in HWRF and upgraded the HWRF default K-closure PBL scheme to a TKE-closure scheme. Using the modified HWRF, we re-simulated Hurricane Patricia (2015) with the same model configurations and the same initial and boundary conditions as the default HWRF simulation. Figure 3.7 shows the horizontal distribution of eddy exchange coefficients for momentum at different altitudes at 06:00 UTC, October 23, 2015. Because of the accurate estimation of static stability, the buoyancy production of turbulence is well predicted by the TKE scheme. As a result, the modified HWRF successfully generates turbulent mixing above the PBL in the eyewall and rainbands. Figure 3.8 shows the storm track and intensity simulated by the default HWRF and the modified HWRF with BVF calculation upgrades compared with the NHC best track data. Note that these simulations are different from the official operational HWRF forecasts in that the coupled

ocean model is not activated, instead, the simulations are forced by the daily real-time global (RTG) SST data. This is to reduce the complications from the coupled ocean. Like the HWRF official operational forecasts, the default uncoupled HWRF is able to predict a decent storm track close to the best-track data, but it fails to simulate the observed RI of Patricia. The modified HWRF predicts a slightly better storm track in the later part of the simulation although the improvement is only marginal. But the BVF upgrades substantially improve the HWRF's skill in intensity prediction, suggesting the importance of representation of turbulent mixing in the eyewall and rainbands to TC intensity forecast. The reasons underlying such an improvement will be discussed in detailed in Chapter 5.

Figure 3.7: Horizontal distribution of eddy exchange coefficients for momentum (k_m) generated by the modified HWRF with BVF upgrades at the altitudes of $z =$

0.5, 1.0, 2.5, and 5.0 km respectively at 06:00 UTC, October 23, 2015 during the RI of Patricia (2015).

Figure 3.8: Storm surface central pressure, maximum surface wind speed, and track of Patricia (2015) simulated by the default HWRF and the modified HWRF with the BVF upgrades compared with the NHC best-track data.

To further examine impact of BVF on TC intensification and the performance of the modified HWRF on simulating RI of major hurricanes, we simulated additional two Category-4 major hurricanes, Jimena (2015) and Harvey (2017), as the operational HWRF had problems to predict their intensification rate and peak intensity. Figures 3.9 and 3.10 shows the intensity of Jimena (2015) and Harvey (2017) of four cycles with different lead time simulated by the default HWRF and the modified HWRF with the BVF upgrades compared with the NHC best track data. In Both cases, the observed RIs of Harvey (2017) and Jimena (2015) are largely missed by the default HWRF, but the intensity simulations, either the surface maximum wind speed or storm central pressure, are improved due to the BVF upgrades. Similar to what is shown in Figure 3.7, the modified HWRF is able to generate robust turbulent mixing the eyewall and rainbands of both storms. I will show in Chapter 5 that such an improvement in storm intensity simulation can be largely attributed

to the improved storm inner-core structure and eyewall/rainband eddy forcing simulated by the modified HWRF, which are needed for TC vortex spin-up.

Jimena (2015)

Figure 3.9: Maximum surface wind speeds and storm central pressure of hurricane Jimena (2015) simulated by the default HWRf and the modified HWRf with the BVF upgrades for four cycles with different lead time (18 UTC Aug. 26, 00 UTC, 06 UTC, and 12 UTC Aug. 27, 2015) compared with the NHC best-track data.

Harvey (2017)

Figure 3.10: Maximum surface wind speeds and storm central pressure of hurricane Harvey (2017) simulated by the default HWRf and the modified HWRf with the BVF upgrades for four cycles with different lead time (18 UTC Aug. 22, 00 UTC, 06 UTC, and 12 UTC Aug. 23, 2017) compared with the best-track data.

Figure 3.11: Storm central pressure, maximum surface wind speeds, and track of hurricane Hermine (2016) simulated by the default HWRf and the modified HWRf with the BVF upgrades for three cycles with different lead time (06 UTC, 12 UTC, and 18 UTC Aug. 31, 2016) compared with the NHC best-track data.

To further examine the performance of the modified HWRf on simulating weak storms, we simulated hurricane Hermine (2016) whose peak intensity only exceed Category-1 intensity for a shortly period of roughly 12 h from 18:00 UTC September 1st to 6:00 UTC September 2nd. Despite its weak intensity, Hermine (2016) has affected Florida, east coast of US, and Atlantic Canada and caused 5 fatalities and about \$550 M Damage. Figure 3.11 shows the simulated intensity and track of Hermine (2016) by the default HWRf and the modified HWRf with the BVF upgrades for three cycles with different lead time compared with the NHC best-track data. It shows that the BVF upgrades does not

appear to degrade HWRF's performance on this case, nor generate the false RI. In fact, it slightly improves the prediction of storm central pressure for all three cycles.

In summary, the static stability of the atmosphere is an important concept used for describing the expected moist convection and buoyancy production of turbulence. Quantitatively, it is measured by the vertical acceleration of an air parcel in the environment in which the parcel is embedded, which leads to an important stability parameter known as BVF. This chapter reviews and derives the mathematical expressions of BVF for the unsaturated atmosphere and the saturated atmosphere. Using the model output from HWRF, the static stability in deep convective eyewall and rainbands is examined. The analyses show that the formula of BVF for the unsaturated atmosphere significantly overestimates the static stability in moist convection of eyewall and rainbands. We implemented the BVF calculation for the saturated atmosphere in HWRF. Numerical simulations of several major hurricanes by the modified HWRF show that BVF upgrades combined with a TKE-cloud turbulent mixing scheme notably improves HWRF's intensity forecasting skill.

4 Hydrometeor Sedimentation in the Eyewall and Rainbands and its

Parameterization

The evolution of a TC vortex is intimately involved with a secondary overturning circulation induced by friction and diabatic heating. In real a TC, the microphysical processes directly interact with the dynamical processes, such as convective up/downdrafts and turbulent eddy circulations in clouds, to generate the diabatic heating and cooling that govern the intensification of a TC. A key microphysical process that depends strongly on

microphysical-dynamical interaction is the hydrometeor sedimentation. In real clouds, hydrometeor particle fall velocities are determined by both particle properties (e.g., mass density, size, and shape) and the cloudy turbulent flow in which hydrometeors are embedded. In numerical models, however, microphysics, turbulent mixing, and the dynamic core are handled by separate modules. How to appropriately generate robust internal interaction by the separated physics modules in models is an important scientific question that has not been thoroughly addressed. In this chapter, I'll discuss and investigate issues regarding hydrometeor sedimentation of TCs and their impact on TC intensification.

4.1 Diagnosing Hydrometeor Fall Velocities in Simulated TCs by HWRF

The sensitivity tests on microphysics presented in Chapter 2 show none of the microphysics schemes used in WRF-ARW is able to reproduce Patricia (2015)'s RI. Understanding the underlying reason is the motivation for this study. Despite their different complexities from spectral bin microphysics to single moment bulk microphysics, all these schemes exclusively use the following simple empirical relation in terms of particle diameters to parameterize hydrometeor fall velocities,

$$\begin{cases} V_t(D) = aD^b & (\text{Locatelli and Hobbs 1974}) \\ V_t(D) = aD^b \exp(-fD) & (\text{Ferrier 1994}) \end{cases} \quad (4.1)$$

where D is the equivalent diameter of a particle, a , b , and f are the empirical coefficients obtained from the best fitting between terminal velocity and particle diameters. It remains to be a question if such a simple empirical relation obtained in non-TC conditions can well represent the hydrometeor fall velocities in the eyewall and rainbands where the particle sedimentation depends strongly on the microphysical-dynamical interaction.

Figure 4.1: Hydrometeor number of concentration as function of diameter and time since the start of penetration for a constant altitude of a flight leg made through Hurricane Emily for a time period between 2015:58 and 2021:35 UTC on Sept. 22, 1987. Size distributions are obtained from all frozen and liquid hydrometeors. Solid curve indicates the vertical velocity deduced from aircraft measurements (after Marfarquar and Black 2004).

Figure 4.2: Hydrometeor fall velocities as function of particle diameters for different types hydrometers including unrimed radiating assemblages of dendrites (open circles), densely rimed radiating assemblages of dendrites (solid circles), and densely rimed dendrites (squares). The empirical relation (dashed curve) for unrimed radiating assemblages of dendrites is from Zikmunda and Vali (1972), and the empirical relation (dash-dotted curve) for unrimed dendrites is from Brown (1970). The figure is adopted from Locatelli and Hobbs (1974).

Our investigation is also inspired by the observational study by Marfarquar and Black (2004), who examined the particle size distributions of liquid and frozen hydrometeors including both snow and graupel near the melting layer using in-situ microphysical data collected during Hurricane Norbert (1984) and Emily (1987). Figure 4.1 shows a time series of number of concentration for various sizes of liquid and solid hydrometeors collected from Hurricane Emily (1987) adopted from Marfarquar and Black (2004). The observations clearly show that there were a large number of hydrometeor particles with diameters smaller than 300 μm (0.3 mm). In a highly-referenced study, using observations, Locatelli and Hobbs (1974) examined the relationship between fall velocities

and particle diameters. Figure 4.2 shows the fall speeds of various classes of unrimed and rimed hydrometeors as a function of particle size diameters (adopted from Locatelli and Hobbs, 1974). Based on this empirical relationship, the large number of particles with diameters smaller than 0.3 mm observed in Marfarquar and Black (2004) would correlate to fall velocities less than 0.2 ms^{-1} .

Figure 4.3: Histogram of snow fall speeds (ms^{-1}) from the HWRF simulations of Patricia (2015) averaged from 12 UTC 22 to 23 UTC 23 October, 2015; Harvey (2017) averaged from 06 UTC 24 to 18 UTC 25 August, 2017; and Jimena (2015) averaged from 18 UTC 27 to 06 UTC 29 August, 2015. Note that the default microphysics used in HWRF is the Ferrier-Aligo scheme.

Given the expected abundance of hydrometeors with fall speeds less than 0.2 ms^{-1} in real TCs, it is interesting to see if HWRF can generate a distribution of particle fall speeds as a function of particle diameters similar to observations. To do so, we examined the hydrometeor fall velocities in HWRF simulations. Figure 4.3 shows the Histogram of snow fall speeds (ms^{-1}) from Patricia (2015), Harvey (2017), and Jimena (2015) simulated by the default HWRF that uses Ferrier-Aligo microphysics. In contrast to the abundance of small hydrometeors with fall speeds less than 0.2 ms^{-1} shown observations, few particles with fall speeds less than 0.2 ms^{-1} are generated in the HWRF simulations. To further look into the problem, the fall velocities from the composite of the three storms are sorted based on the grid-box mean particle diameters and model resolved vertical velocities. The results

are shown in Figure 4.4, which clearly indicates that the fall speeds in the updraft are greater than those in the downdraft for the particles with same diameters in all ranges. Two features are unphysical. First, there is no mechanism that can explain why for a given particle diameter, fall velocity in the updraft is substantially larger than that in the downdraft, particularly for large particles greater than 550 μm . Second, small particles are presumably to be more subjected to convective currents in real clouds. However, the simulated small particles do not appear to be affected by the convective currents. The statistics shown here suggest that the impact of microphysical-dynamical interaction on hydrometeor sedimentation in the convective eyewall and rainbands is not captured by the simple fall speed parameterization of Eq. (4.1).

Figure 4.4: Statistics of snow fall speeds (ms^{-1}) from the composite of HWRP simulations of Patricia (2015) from 12 UTC 22 to 23 UTC 23 October, 2015; Harvey (2017) from 06 UTC 24 to 18 UTC 25 August, 2017; and Jimena (2015) from 18 UTC 27 to 06 UTC 29 August, 2015. (a): Snow fall speeds as function of grid-box mean particle diameter and vertical velocity. (b) Histogram of snow fall speeds in convective updrafts and downdrafts.

4.2 Sensitivity of TC Intensification to Hydrometeor Fall Speeds

In Chapter 3, we showed that an accurate estimation of static stability in the eyewall and rainbands improves HWRF's ability to capture Patricia (2015)'s RI. However, as indicated by Figure 3.8, the modified HWRF still misses the observed peak intensity of Patricia by a large margin, suggesting that while much improved, the inner-core dynamics associated with the microphysical-dynamical interaction that governs TC intensification is still not fully captured by the HWRF. We hypothesize that the under-predicted peak intensity of Hurricane Patricia (2015) by HWRF is likely caused by the lack of hydrometeors with fall speeds less than 0.2 ms^{-1} , which are abundant in real TCs according to observations. To test this hypothesis, we artificially reduce snow fall speeds V_{sf} using a weighted function in terms of particle diameter as follows,

$$\begin{cases} V_{sf} = V_{sf0}(\alpha + \frac{|d-d_1|}{d}\beta)^\gamma, & \text{for } d \leq d_2 \\ V_{sf} = V_{sf0}, & \text{for } d > d_2 \end{cases}, \quad (4.2)$$

where d_1 and d_2 are the cut off particle diameters at the both ends; and α , β , and γ are arbitrary coefficients to determine the weights of reducing snow fall speeds as a function of particle diameters. Note that in the sensitivity experiment Eq. (4.2) is only used to artificially reduce snow fall speeds to mimic the presence of snow particles with fall speeds less than 0.2 ms^{-1} , which are abundant in real TCs. We leave the fall speeds of rain droplets unchanged during the HWRF simulation as larger rain droplets are presumably less affected by the convective currents.

We performed a sensitivity experiment using the modified HWRF with the BVF upgrades by taking $d_1 = 8\mu\text{m}$, $d_2 = 1000 \mu\text{m}$, $\alpha = 0.1$, $\beta = 0.9$, and $\gamma = 3.84$. Figure 4.5 shows the storm track and intensity of Patricia (2015) simulated by the default HWRF,

the modified HWRF with BVF calculation upgrades, and the modified HWRF with the reduced snow fall speeds by Eq. (4.2) compared with the NHC best track data. The HWRF with the reduced snow fall speeds not only fully captures the observed RI of Patricia (2015) but also successfully reproduces the storm's peak intensity. To confirm the importance of light solid-phase hydrometeors with fall speeds less than 0.2 m/s to TC intensification, we examined the distribution of snow fall speeds in the sensitivity experiment (Figure 4.6). It clearly shows that majority of snow particles have fall speeds less than 0.2 m/s consistent with observations. However, it should be keep in mind that Eq. (2) has no physical basis. The only purpose is to illustrate that the light hydrometeors with small fall speeds play an important role in Patricia (2015)'s RI. In the next section, I'll review how the current particle fall speed parameterization was formulated and why it causes problems when applying to the eyewall and rainbands. Based on the discussion, I'll propose a new scheme that fully considers the microphysical-dynamical interaction in the eyewall and rainbands.

Figure 4.5: Storm surface central pressure, maximum surface wind speed, and track of Patricia (2015) simulated by the default HWRF, the modified HWRF with the BVF upgrades, and the modified HWRF with reduced snow fall speeds compared with the NHC best-track data.

Figure 4.6: Histogram of snow fall speeds (ms^{-1}) from the sensitivity experiment of Patricia (2015) simulated by the modified HWRF with the reduced snow fall speeds. The data are averaged from 12 UTC 22 to 23 UTC 23 October, 2015.

4.3 Hydrometeor Fall speeds in Quiescent Conditions

The particle fall speed parameterization currently used in microphysics schemes are derived in quiescent conditions in which the three forces acting on a particle, namely, the gravitational force, the buoyancy force, and drag are assumed to be balanced as illustrated by Figure 4.7. Mathematically, it can be represented by,

$$\frac{1}{2}\rho_e C_D A_h \omega_t^2 = V_h \rho_h g - V_h \rho_e g, \quad (4.3)$$

where V_h , A_h and ρ_h are the volume, projected area with respect to the fall speed, and density of a particle, respectively, ρ_e is the density of the environmental atmosphere, C_D is

the drag coefficient, and ω_t is the fall speed (or terminal velocity) of the particle. The three terms in Eq. (4.3) from left to right represent the drag, gravitational force, and buoyancy force, respectively. From Eq. (4.3), the particle fall speed may be expressed as,

$$\omega_t^2 = \frac{2V_h(S-1)g}{C_D A_h}, \quad (4.4)$$

where $S = \frac{\rho_h}{\rho_e}$ is the ratio of the particle density to air density. The drag coefficient C_D depends on the viscosity and turbulent motion of the air. By defining the particle Reynolds number R_e as,

$$R_e = \frac{\omega_t d}{\nu}, \quad (4.5)$$

where d is the equivalent diameter of the particle and ν is the kinematic viscosity of the air, Stokes (1851) solved the Navier-Stokes equations by neglecting all inertia terms and theoretically derived a relation between C_D and R_e for a spherical particle in a streamline regime ($R_e < 0.5$) as,

$$C_D = \frac{24}{R_e}. \quad (4.6)$$

For high R_e , which is the case of the turbulent atmosphere, C_D has to be determined empirically, often as a function of R_e , $C_D = f(R_e)$.

Figure 4.7: Schematic illustration of particle fall in a quiescent condition in which the downward gravitational force balances the upward buoyancy force and drag. ω_t is the fall speed (or terminal velocity) of a particle.

While the underlying physics is clear, a direct use of Eq. (4.4) to calculate particle fall speed is difficult because variables required for the calculation are not easy to obtain. Mitchell (1996) proposed a method to simplify the problem. He defined a new variable, called Davies number, as,

$$X = C_D R_e^2. \quad (4.7)$$

He further assumed that R_e may be related to X as,

$$R_e(X) = aX^b, \quad (4.8)$$

where a and b are empirical coefficients. Inserting Eq. (4.6), Eq. (4.7), and Eq. (4.8) into Eq. (4.4), it yields,

$$\omega_t = av^{1-2b} \left[\frac{2V_h g}{A_h} (S - 1) \right]^b d_n^{2b-1}. \quad (4.9)$$

Further assuming that the volume and projected area of the particle can also be represented in terms of the equivalent diameter in the form of,

$$V_h = \lambda d_n^\beta, \quad A_h = \gamma d_n^\varepsilon, \quad (4.10)$$

then, Eq. (4.9) may be regrouped as,

$$\omega_t = A_t d_n^{B_t}, \quad (4.11)$$

where $A_t = av^{1-2b} \left[\frac{2\lambda g}{\gamma} (S - 1) \right]^b$ and $B_t = (\beta - \varepsilon)b + 2b - 1$ are the empirical coefficients that need to be determined from either lab or field experiments. Eq. (4.11) is the same as Eq. (4.1) and explains why it is widely used in microphysics schemes.

4.4 Hydrometeor Fall Speeds in Non-Quiescent Conditions

Eq. (4.11) can be readily implemented in microphysics schemes to provide a practical way to parameterize particle fall speed based on the particle diameters. However, the quiescent condition of the background flow is not always valid, particularly in the convective eyewall and rainbands. As illustrated by Figure 4.8, there are strong updrafts and downdrafts with magnitude exceeding 10 m/s associated with the eyewall and rainbands. Surely, the impact of an updraft as strong as 10 m/s on a hydrometeor particle cannot be neglected. It will exert a tremendous dynamic pressure force on a falling particle. But such a dynamic impact on hydrometeors resulting from convective currents is not considered by the three-force balance represented in Eq. (4.3). As a result, significant errors in microphysics will be generated in the TC inner-core region as hydrometeor sedimentation is treated everywhere the same in terms of particle diameters regardless of the strong convective current associated with the eyewall and rainbands.

Figure 4.8: Vertical velocity (m/s) at 6 km altitude at 1200UTC October 23, 2015 of Patricia (2015) simulated by the modified HWRF with the BVF upgrades.

For coarse resolution simulations in which the up-/down-drafts induced by moist convection are not resolved, the application of Eq. (4.2) or Eq. (4.11) is acceptable since the weak vertical velocity associated with the large-scale flow resolved by models only has a negligible impact on particle fall velocity. As model resolution increases to cloud resolving grids, neglecting the impact of the strong convective currents in the eyewall on particle fall velocities could lead to severe bias in simulations of convection. In this case, the dynamic pressure force resulting from the background convective drafts acting upon particles should be considered in the force balance that governs the terminal velocity of a

particle. We hypothesize that neglecting the impact of strong eyewall convective currents on particle fall speed is the culprit for HWRF not producing hydrometeors with fall speed smaller than 0.2 m/s since smaller particles are presumably more subjected to the background flow acting upon them.

Figure 4.9: Schematic illustration of particle fall in a non-quiescent flow with an ambient speed of W_c . The particle fall speed (or terminal velocity) ω_t is governed by a balance among the downward gravitational force, the upward buoyancy force, drag, and a dynamic pressure force.

The previous argument suggests that the quiescent condition is not an appropriate assumption for hydrometeor sedimentation in the eyewall and rainbands. As illustrated by Figure 4.9, in a non-quiescent flow with an ambient speed of W_c , The particle fall speed (or terminal velocity) ω_t is governed by a balance among the downward gravitational force, the upward buoyancy force, drag, and a dynamic pressure force. Mathematically, the balance may be written as,

$$\frac{1}{2}\rho_e C_D A_h (W_c - \omega_t)^2 + \frac{1}{2}\rho_e A_h (W_c - \omega_t)^2 = V_h \rho_h g - V_h \rho_e g. \quad (4.12)$$

The terms in Eq. (4.12) from left to right indicate the drag, dynamic pressure force, gravitational force, and buoyancy force. Here, we distinguish drag and dynamic pressure force in that the former results from the viscosity and turbulence of the fluid. For inviscid and laminar flow, C_D is zero, but dynamic pressure force remains as long as velocity is not zero. Defining an equivalent diameter $d = (abc)^{1/3}$, solving Eq. (4.12), it yields,

$$(W_c - \omega_t)^2 = \frac{2V_h(S-1)g}{(C_D+1)A_h} = \frac{4(S-1)g}{3(C_D+1)} d. \quad (4.13)$$

To obtain particle fall speed, C_D must be determined empirically. Let's first consider two extreme cases. At one end, for small Re less than 0.5, C_D may be calculated in terms of Re via Eq. (4.6). At the other end, when Re approaches infinity, C_D is commonly considered as a constant (Rubey 1933). For Re in-between, Cheng (1997) proposed a generalized formula for C_D in the form of,

$$C_D = [(M/Re)^{1/\zeta} + N^{1/\zeta}]^\zeta, \quad (4.14)$$

where M , N , and ζ are the empirical coefficients. Among many empirical relations between C_D and Re , Eq. (4.14) is attractive since it consists of two asymptotic behaviors of C_D : $C_D = \frac{M}{Re}$, when $Re \rightarrow 0$; and $C_D = N$, when $Re \rightarrow \infty$. Another advantage of Eq. (4.14) is that combining with Eq. (4.13), an analytical expression of particle fall speed ω_t can be derived (Cheng, 1997; Wu and Wang, 2006; Riazi and Turker, 2019). Defining a generalized drag coefficient $C'_D = C_D + 1$ and further assuming that C'_D also follows Eq. (4.14) but with different coefficients, M' , N' , and ζ' , then, it can be shown that the particle fall speed ω_t may be expressed as,

$$|W_c - \omega_t| = \frac{M'v}{N'd} \left[\sqrt{\frac{1}{4} + \left(\frac{4N'}{3M'^2} D_*^3\right)^{1/\zeta'} - \frac{1}{2}} \right]^{\zeta'}, \quad (4.15)$$

where $D_* = d[\frac{(S-1)g}{\nu^2}]^{1/3}$.

To appropriately calculate ω_t , several coefficients needs to be determined. As discussed earlier, for spherical particles in a flow with $R_e < 0.5$, the theoretical derivations by Stokes (1851) showed $M = 24$. This result was confirmed by Rubey (1933), Van Rijn (1989) and Julien (1995). However, other studies (e.g., Zhang, 1961; Raudkivi, 1990; Cheng, 1997) showed that for non-spherical particles M has values between 32 and 34. While it is a consensus that C_D is a constant as R_e approaches infinity, different studies yield different values of N ranging from 1.0 to 2.1 (Rubey, 1933; Zhang, 1961; Rijn, 1989; Raudkivi, 1990; Julien, 1995; and Zhang, 1997). Coefficient ζ also varies from 1.0 to 1.5 in these studies. Here, we treat M' , N' , and ζ' as three tunable coefficients. But following the previous studies, here, we choose $M' = 32$ (Raudkivi, 1990; Cheng, 1997), $N' = 2.1$ (Camenen B. 2007), and $\zeta' = 2.0$ (Dallavalle, 1948).

Another important variable that can affect particle fall speed substantially is the ratio of hydrometeor density to air density S . Depending on the characteristics of hydrometeors, such as, rimed or unrimed snow flake and ice crystals, the density of hydrometeors can have a wide range from roughly 170 kg/m^3 to 900 kg/m^3 . Here following Ferrier (1994), we assume S may be parameterized as a function of particle diameters in the form of,

$$S = \max \left\{ 170.0 \cdot \left[1 + \delta_3 \left(\frac{d}{1000.0} \right)^{\delta_2} \right]^{\delta_1}, 900.0 \right\}, \quad (4.16)$$

where δ_1 , δ_2 , and δ_3 need to be determined empirically, here we treat them as tunable coefficients and taken as $\delta_1 = 5.0$, $\delta_2 = 3.9$, and $\delta_3 = 0.395$, respectively. The unit of particle equivalent diameter d is μm . Previous researches show that the viscosity of the

atmosphere depends strongly on temperature. The kinetic theory of gasses (Kauzmann 1966) allows the temperature-variation of gaseous viscosity to be accurately determined but involves complicated calculations. In practice, simple formulae were proposed based on different models, such as the so-called hard-sphere, power-law (Chapman and Cowling 1970), to calculate viscosity as function of temperature. Here, we retain the formula used in Ferrier (1994) to express the dynamic viscosity of atmosphere as,

$$\mu = \mu_0 \frac{T^{3/2}}{(T+120.0)}, \quad (4.17)$$

where $\mu_0 = 1.496e^{-6} \frac{kg}{ms}$ is a reference viscosity. Then, the kinematic viscosity of the atmosphere is, $\nu = \frac{\mu}{\rho}$, where ρ is the air density.

We have implemented the fall speed scheme represented by Eq. (4.15) in HWRF and re-simulated Hurricane Patricia (2015). Figure 4.10 shows the storm track and intensity simulated by the HWRF with the new fall speed parameterization compared with those from the default HWRF, the HWRF with BVF updates alone, and the best-track data. The modified HWRF with the new fall speed parameterization is able to nearly reproduce the observed peak intensity of Patricia (2015) although it appears to over-estimate the rate of intensification slightly during the RI period. This issue will be investigated in future study.

To confirm our hypothesis that the lack of particles with fall speeds less than 0.2 m/s is one of the culprits for HWRF's inability to reproduce Patricia's RI and peak intensity, Figure 4.11 shows the distribution of snow fall speeds in the HWRF simulation with the new particle fall speed parameterization. Majority of snow particles now have fall speeds less than 0.2 m/s consistent with observations.

The results shown in Figures 4.5 – 4.6 and 4.10 – 4.11 suggest the importance of turbulent and microphysical processes to storm intensification. However, how the turbulent processes interact with the microphysical processes in the inner-core region to regulate the storm intensification remains poorly understood. In the next chapter, this issue will be explored using the HWRF simulations performed in this study.

Figure 4.10: Storm surface central pressure, maximum surface wind speed, and track of Patricia (2015) simulated by the default HWRF, the modified HWRF with the BVF upgrades, and the modified HWRF with the BVF updates and new particle fall speed parameterization compared with the NHC best-track data.

Figure 4.11: Histogram of snow fall speeds (ms^{-1}) from the sensitivity experiment of Patricia (2015) simulated by the modified HWRF with the BVF upgrades and new particle fall speed parameterization. The data are averaged from 12 UTC 22 to 23 UTC 23 October, 2015.

5 TC Inner-core Dynamical-Microphysical Interaction and Intensification

The focus of this chapter is to explore how TC internal processes involving the dynamical-microphysical interaction in the inner-core region regulate the TC intensification. To do so, using model output I analyzed the two HWRF simulations of Patricia (2015) that uses HWRF's default physics and the modified physics that includes the BVF correction and the new hydrometeor fall speed parameterization. First, I will compare the storm structure simulated by the two HWRFs. Second, I will explore how eyewall convection triggers the positive feedback mechanism among surface winds, surface evaporation, and TC intensification leading to the RI of TCs. Finally, I will discuss how eyewall convection affects the outflow temperature, a key component of the Carnot heat engine of a TC, and its role in the RI of TC.

5.1 The Inner-Core Structure of the Simulated Patricia (2015) by HWRF

To better understand why the two simulated Patricia (2015) by HWRFs underwent different intensification pathways (blue and green in Fig. 4.10), I first examined the inner-core structure of the simulated Patricia (2015). Figure 5.1 compares the horizontal plane structure of Patricia (2015) retrieved from satellite observations at the initiation of Patricia's RI with that simulated by the default HWRF and the HWRF with the TKE scheme, BVF correction, and the new particle fall speed parameterization. A well-defined inner-core structure including a quasi-closed ring feature around the storm center (somewhat broken in the south quadrant) is clearly shown in the satellite image. A broken outer concentric eyewall is somewhat visible is seen at this time. From a large amount of 37 GHz microwave color products, Kieper

and Jiang (2012) showed that the first appearance of a cyan color ring around the storm center is highly correlated to subsequent RI, provided that environmental conditions are favorable. This result is consistent with the later analyses of Tropical Rainfall Measuring Mission (TRMM) 29 Precipitation Radar (PR) data (Jiang and Ramirez 2013; and Tao and Jiang 2015), which showed that nearly 90% of RI storms in different ocean basins formed a precipitation ring around the storm center prior to RI. The relationship between the ring feature and the subsequent RI obtained from these observational studies is consistent with the theoretical finding of Nolan et al. (2007), who demonstrated that the intensification processes of a balanced, baroclinic TC-like vortex is mainly driven by the TC symmetric response to the azimuthally-averaged diabatic heating, rather than to the heating directly associated with individual asymmetries distributed around the TC vortex.

To verify if Patricia's RI possesses the similar RI signature found in these observational and theoretical studies, we carefully examined the inner-core structure of the simulated Patricia (2015) prior and during the early stage of RI. Figures 5.1b and 5.1c show the horizontal distribution of hydrometeor mixing ratio averaged over 5 – 8 km altitude from the two HWRF simulations with default physics and the modified physics including the TKE scheme, BVF correction, and the new particle fall speed parameterization. The vortex inner-core structure simulated by the default HWRF is poorly organized and the simulated eyewall appears to be much larger in size than the satellite observed eyewall (Fig. 5.1a vs. 5.1c). It suggests that HWRF with operational model physics is unable to generate the right vortex inner-core structure needed for the subsequent RI. In contrast, the modified HWRF and physics

upgrades produces a well-defined convective ring around the storm center that is clearly shown in the hydrometeor mixing ratio (Fig. 5.1b). The size of the simulated convective ring by the modified HWRF is similar to that shown in the satellite image. The simulation somehow is able to reproduce the outer concentric convective ring shown in the satellite observation. In addition, the simulated asymmetric rainband structure with the strongest convection occurring in the southeast quadrant is consistent with the satellite observation. The similar vortex inner-core structure shown in both satellite observations and the simulation with the modified model physics implies that the RI of Patricia (2015) is likely governed by the axisymmetric dynamics. This result is consistent with the finding by Vigh et al. (2018) who showed that some of the VRI and ERI storms, such as Hurricane Patricia (2015), Typhoon Meranti (2016), and Hurricane Maria (2017), can be well captured by an axisymmetric two-dimensional (radius-height) model of the Coupled Hurricane Intensity Prediction System (CHIPS) developed by Emanuel (2004) by prescribing zero-wind shear. The fact that the observed TC inner-core structure including the quasi-closed convective ring feature is reproduced by the modified HWRF with upgraded physics suggests that the SGS physics involving with the in-cloud turbulent mixing above the PBL and the microphysical-dynamical interaction in the inner-core region facilitates the realization of the axisymmetric dynamics underlying the RI of a TC.

Figure 5.1: (a) Satellite image of Patricia (2015) at 14:45:00 UTC, Oct. 22, 2015 from the Advanced Microwave Scanning Radiometer 2 (credit to NRL Satellite Team). (b) and (c): Hydrometeor mixing ratio (g kg^{-1}) averaged over the altitudes of 5 – 8 km at 15:00 UTC, Oct. 22 of Patricia (2015) simulated by the HWRFBVF-STV with the TKE scheme, BVF correction, and the new particle fall speed parameterization (HWRFBVF-STV) and the HWRF with the default physics.

Figure 5.2: Simulated azimuthal-mean radius-height structure of updrafts (thick grey contours, ms^{-1}), downdrafts (magenta contours, ms^{-1}), diabatic heating (color shading, K h^{-1}), radial inflow (red contours, ms^{-1}), outflow (white contours, ms^{-1}), and radial flow convergence (black contours, s^{-1}) averaged over Patricia's RI period from 15:00 UTC 22 to 03:00 UTC 23 October, 2015. (a): default HWRF; (b): modified HWRF with the TKE scheme, BVF correction, and the new particle fall speed parameterization.

It should be pointed out that the upper limit on intensification is determined by the conversion of latent heating to kinetic energy. It has been argued that this conversion is often less efficient in a 3D intensification process than that in a 2D axisymmetric model. This is likely due to the fact that the convection in a 3D model is often not organized into an annular ring, and thus, the azimuthally averaged heating rate in a 3D model is smaller than that in a 2D axisymmetric model. This probably is the case in Vigh et al. (2018)'s 2D simulations that well reproduce the RI of some of the major hurricanes. However, there are examples in literature that showed the energy conversion in 3D model simulations can be as efficient as that in an axisymmetric model. For instance, Persing et al. (2013) reported that "there is a short period of time when the rate of spin-up in the 3-D model exceeds that of the maximum spin-up rate in the axisymmetric model, and during this period the convection is locally more intense than in the axisymmetric model and the convection is organized in a quasi ring-like structure resembling a developing eyewall" (see Page 12336 of their paper). This is a situation similar to our HWRF simulation with the inclusion of TKE turbulent mixing, BVF correction, and the new fall speed parameterization where the robust microphysical-dynamic interaction in the inner-core region appears to facilitate the realization of the axisymmetric dynamics underlying the RI of Patricia (2015).

Figure 5.2 shows the simulated azimuthal-mean radius-height structure of vertical velocity, hydrometeor mixing ratio, radial inflow/outflow, and radial flow convergence averaged over the Patricia's RI period from 15 UTC 22 to 03 UTC 23 October, 2015. Compared with the default HWRF, the modified HWRF generated much stronger updrafts

(thick gray contours) in the eyewall, stronger radial inflow (red contours) within the PBL, and outflow (white contours) near the tropopause, which are consistent with the strong storm intensity simulated by this experiment. Furthermore, the radial flow convergence (black contours) shows an interesting bimodal structure. In the modified HWRF run, the primary radial flow convergence matches well with the eyewall updrafts. This feature facilitates an efficient transport of moisture into the eyewall to result in a large diabatic heating (color shading) in the eyewall, which fosters the rapid converging spin-up processes as air parcels move radially inward and ascend swiftly within the eyewall. This result suggests the importance of microphysics-turbulence interaction in TC intensification. The secondary radial flow convergence is seen along the interface where radial inflow changes to outflow, which should be related to the development of rainbands. In this case, it might be linked to the outer concentric convective ring. In contrast, in the default HWRF simulation, the radial flow convergence in the eyewall is very weak, particularly in the low troposphere it appears to be overwhelmed by the radial flow convergence along the interface of radial inflow and outflow. Such a structure is apparently unfavourable to the rapid development of the vortex, since it cannot generate the efficient converging spin-up processes. Therefore, the simulated storm intensity difference by the two HWRFs may be largely attributed to the differences in the strength and structure of the secondary overturning circulation in this case. However, we note the depth of the radial inflow layer is similar in both HWRF simulations. It suggests that the upgraded physics does not alter the basic structure of the PBL in the TC vortex inner- and outer- core region except for the eyewall.

In short, Figures 5.1 and 5.2 show that under the same external condition, changes in turbulent mixing and microphysics scheme are sufficient to alter the intensification pathway and lead the vortex to evolve in a different direction. In the next two sections, I'll explore the underlying reasons why the simulated TC vortex in the two HWRF experiments underwent different intensification pathway.

5.2 Triggering Mechanism of Positive Feedback Underlying the Air-Sea Interaction for TC Intensification

Emanuel (2003) first articulated the critical role of air-sea interaction in generating a positive feedback between the near-surface wind speed and the rate of evaporation from the underlying ocean during the TC intensification process, which is now known as the wind-induced surface heat exchange (WISHE) feedback mechanism, which states a positive feedback between the azimuthal-mean boundary-layer equivalent potential temperature and the azimuthal-mean surface wind speed underneath the eyewall of the storm. This mechanism seems to work well for the simulated Patricia (2015) in the modified HWRF experiment where the storm intensifies at a rate close to the observed RI. Then, why this positive feedback mechanism between surface winds and surface evaporation fails to work in the default HWRF simulation? And what triggers the WISHE mechanism in the modified HWRF experiment, leading to the RI of Patricia?

Figure 5.3: (a) and (b): Radius-height structure of tangential wind (V_t) at 12 UTC Oct. 23, 2015 from the default HWRf (HWRf-DEF) and the modified HWRf (HWRf-MOD). (c) and (d): Time evolution of V_t averaged over the radii of 19-22 km from the two HWRf runs. (e): Time evolution of V_t acceleration averaged over the radii of 19-22 km and heights of 450-800 m.

To answer this question, first let us look at how the azimuthal-mean tangential winds evolve in the two HWRf experiment. Figures 5.3a and 5.3b compares the radius-height structure of azimuthal-mean tangential wind at 12 UTC October 23, 2015 (a time when Patricia is near its peak intensity) from the two HWRf simulations. It shows that

both simulations produce similar radius of maximum wind (RMW), which is about 20 km in radius, and similar height of the peak wind, which is located in the upper PBL around 500 – 800 m in altitude. For this reason, Figures 5.3c, 5.3d, 5.3e depict the time evolution of azimuthal-mean tangential wind at the radii of 19 – 22 km, and the mean acceleration rate of tangential wind at the height of 450 – 800 m, respectively. It shows that the default HWRF significantly under-predicts Patricia’s intensity. While both simulations show large fluctuations of tangential wind acceleration throughout the evolution of the storm, the modified HWRF predicts a consistent large acceleration of tangential winds from 15 UTC Oct. 22 to 3 UTC Oct. 23. This is the period that separates the two simulations in terms of the intensification. It appears that the air-sea interaction WISHE feedback mechanism for RI works well in the modified HWRF simulation. But it fails to work in the default HWRF simulation.

Figure 5.4: Azimuthal-mean radius-height structure of SGS tangential eddy forcing averaged over Patricia’s RI period from 15 UTC 22 to 03 UTC 23 October, 2015 from the two HWRF simulations (DEF-HWRF and TL-HWRF). Note that the SGS TEF smaller than $-1.0 \times 10^{-3} \text{ (ms}^{-2}\text{)}$ is shaded with white color for a clear illustration of SGS eddy forcing above the PBL.

To investigate this issue, we analyzed the azimuthal-mean tangential wind budget expressed in Eq. (1.1). We examined the tangential eddy forcing ($F_\lambda + F_{sgs_\lambda}$) for the primary circulation of the TC vortex. The model-resolved tangential eddy forcing F_λ is calculated using the wind fields in the HWRF output. The SGS tangential eddy forcing F_{sgs_λ} is calculated from the tendencies directly generated by the horizontal and vertical turbulent mixing scheme. Figure 5.4 compares the SGS tangential eddy forcing averaged over the RI period from 15 UTC 22 to 03 UTC 23 October, 2015 between the two HWRF simulations. There are a couple of interesting features shown in the figure. First, the radial-height structure of SGS tangential eddy forcing generated by the default HWRF (Fig. 5.4a) is very similar to that from a 3D full-physics TC simulation shown in Persing et al. (2013, their Figs. 10f & 11f). The SGS eddy forcing above 2 km in the eyewall region is virtually zero because the default HWRF is unable to generate appropriate in-cloud turbulent mixing. In contrast, the modified HWRF allows the model to successfully generate the SGS tangential eddy forcing associated with the eyewall convection above the PBL (Fig. 5.4b). Such a SGS eddy forcing in the eyewall region from the layer just above the PBL to the upper troposphere has not been shown and discussed in previous numerical studies. Second, in addition to the expected strong negative SGS tangential eddy forcing within the PBL, the in-cloud turbulent mixing parameterization generates an interesting vertical structure of SGS tangential eddy forcing above the PBL in the eyewall region. Although it is much weaker than that in the PBL, the SGS tangential eddy forcing in the eyewall does show positive values at the heights just above the inflow layer as well as above the mid

troposphere, suggesting that the eyewall SGS tangential eddy forcing above the PBL is indeed involved in the vortex spin-up processes during the RI. What remains unclear is the fidelity of the parameterized SGS eddy forcing above the PBL and its sensitivity to specific turbulent mixing parameterization. This constitutes one of the uncertainties in storm intensity simulation.

Figure 5.5: (a) and (b): Radius-height structure of resolved eddy forcing and diabatic heating averaged over the RI period of Patricia (2015) from 15 UTC Oct. 22 to 3 UTC Oct. 23, 2015 from the two HWRf runs. (c) and (d): Time-height

variation of diabatic heating averaged over the radii of 20-29 km from the two HWRF runs. (e): Time evolution of resolved eddy forcing averaged over the radii of 17-21 km and heights of 450-800m. The vertical lines indicate the RI period.

The radius-height structure of the model-resolved tangential eddy forcing averaged over the RI period from 15 UTC 22 to 03 UTC 23 October, 2015 from the two HWRF simulations is shown in Figures 5.5a and 5.5b. The basic radial-height structures of the resolved eddy forcing generated by the two simulations are similar to a certain extent, and share similar features to those from Persing et al. (2013)'s 3D full-physics TC simulation (cf. their Figs. 10g & 11g). But the resolved eddy forcing in the modified HWRF is much stronger than that in the default HWRF. A robust feature shown in both simulations is the positive eddy forcing right above the inflow layer (or upper PBL) in the vicinity of the eyewall. From the perspective of absolute angular momentum conservation, this positive tangential eddy forcing is directly linked to the vortex spin-up.

Comparing Figure 5.4 with Figures 5.5a and 5.5b, it is easy to see that the model-resolved eyewall eddy forcing above the PBL in the modified HWRF experiment has a magnitude 20 more times larger than the corresponding SGS eddy forcing, suggesting that the resolved eddy processes provide a major forcing that drives the primary circulation of the TC vortex in this case. As model resolution keeps increasing, we expect that the resolved eddy forcing will become more dominant. This is certainly a promising result, implying that numerical forecast of TC intensification may be ultimately a resolution problem. The difficulty, however, stems from the strong dependence of model-resolved eddy forcing and TC inner-core structure on the parameterized SGS eddy processes at the current resolution. As we showed here, the modifications in SGS turbulent mixing scheme and hydrometeor sedimentation parameterization result in substantial differences in the

vortex structure, secondary overturning circulation, and model-resolved eyewall eddy forcing. Such a dependence of model-resolved TC fields on physics parameterization is currently not fully understood. It could stem from the fact that the large energy-containing turbulent eddies, such as kilometre and sub-kilometer convective elements or roll vortices (evidenced in the LESs), are not resolved by the current model resolution of 1.5 km used by HWRF, and could also result from the dynamical-microphysical interaction in TC clouds. The strong dependence of the resolved TC vortex on SGS parameterization poses a great challenge for accurate prediction of TC intensity change.

The radius-height structure of eddy forcing and diabatic heating shown in Figs. 5.5a and 5.5b is consistent with the finding by Zhu and Zhu (2014) who showed that the diabatic heating of eyewall convection tends to accelerate the tangential wind in the upper boundary layer in the vicinity of the eyewall using the Sawyer-Eliassen diagnoses. The time evolution of the diabatic heating over the radii of 20 – 29 km from the two HWRF simulations are shown in Figs. 5.5c and 5.5d. It clearly shows that in the modified HWRF simulation, low-level convection starts to develop right before the RI of Patricia (Fig. 5.5d). It is this low-level convection leads to the large positive eddy forcing (Fig. 5.5e, red line), which accelerates the tangential wind in the boundary layer and kicks off the subsequent RI via the positive feedback between surface wind and evaporation. On the other hand, the low-level convection (indicated by the diabatic heating) stays weak in the default HWRF (Fig. 5.5c). Although the convection always tends to generate positive eddy forcing to accelerate the tangential wind, it is not sufficiently large (Fig. 5.5e, blue line) to overcome the viscous dissipation at the surface to result in the net acceleration of the tangential wind. This result suggests that there exists a critical strength of low-level convection, below this critical level,

convection cannot generate large enough eddy forcing to initiate the positive feedback mechanism between surface wind and evaporation for TC intensification.

Figure 5.6: (a) Radius-height structure of saturated equivalent potential temperature (θ_{se}) at 12 UTC Oct. 23, 2015 from HWRF-MOD. (b): Vertical profiles of θ_{se} averaged over the radii of 17-21 km at 12 UTC Oct. 23 from the two HWRF simulations. (c) and (d): Time evolution of surface layer θ_{se} averaged over the radii of 17-21 km and 25-70 km, respectively. The vertical lines indicate the RI period.

To understand why the low-level convection can well develop in the modified HWRF simulation, I examined the saturated equivalent potential θ_{se} defined as,

$$\theta_{se} = \begin{cases} T \left(\frac{p_d}{p_0} \right)^{-\frac{R_d}{c_p + c_w q_w}} \exp \left[\frac{L_v q_s}{(c_p + c_w q_w) T} \right] & \text{for } T \geq 273.16 \\ T \left(\frac{p_d}{p_0} \right)^{-\frac{R_d}{c_p + c_i q_i}} \exp \left[\frac{L_s q_s}{(c_p + c_i q_i) T} \right] & \text{for } T < 273.16 \end{cases} \quad (5.1)$$

where L_v and L_s are the specific latent heat of vaporization and sublimation, respectively; C_w and C_i are the specific heat of liquid and solid water; q_w and q_i are the mixing ratio of liquid- and solid-phase water; p_d and p_0 are the dry air pressure and reference pressure. Figure 5.6a shows the azimuthal-mean radius-height structure of θ_{se} at 12 UTC October 23, 2015 from the modified HWRF simulation. Large θ_{se} is seen in the eyewall and the tropopause is around 17.5 km in altitude (Fig. 4a). The vertical θ_{se} profiles averaged over the radii of 17-21 km, where the eyewall is located, from the two HWRF simulations are shown in Figure 5.6b. It is clear that θ_{se} in the modified HWRF is consistently larger than that in the default HWRF throughout the vertical column in the vicinity of the eyewall. Figures 5.6c and 5.6d compare the time evolution of θ_{se} in the surface layer in the eyewall (17 – 21 km) as well as in the outer-core region of the vortex between the two HWRF simulations, respectively. In the modified HWRF, θ_{se} increases well before the RI in both inner and outer-core regions, which sets the favorable environment for the development of low-level convection and kicks off the WISHE mechanism for TC intensification. This is in stark contrast to the default HWRF in which θ_{se} remains steady throughout the simulation in both inner- and outer-core regions. This suggests that the positive feedback between the azimuthal-mean boundary-layer equivalent potential temperature and the azimuthal-mean surface wind speed does not occur underneath the eyewall and the outer-core region of the storm. This result suggests that the low level convection provides an important triggering to the WISHE mechanism underlying the TC intensification. The HWRF experiments with different turbulent mixing schemes indicate that the low-level convection is sensitive to how turbulent mixing is parameterized. Without a robust

turbulent mixing parameterization, particularly in the eyewall, the WISHE mechanism cannot be appropriately triggered.

5.3 Importance of Outflow Temperature to TC Intensification

One of the foundations for understanding TC set by Emanuel (e.g., Emanuel 1986, 1991) is that the energy cycle of a TC can be idealized as a Carnot cycle constituting the most efficient heat engine between the two temperature reservoirs at the surface layer close to the ocean and TC outflow aloft to convert the heat energy extracted from the ocean to mechanical energy. In the previous section, our discussion focused on the temperature reservoir at the surface layer. We showed the important modulation of low-level convection via turbulent mixing in the eyewall in triggering the WISHE mechanism for TC intensification. A question yet to be answered is: why hydrometeor sedimentation, particularly for snow fall aloft in the deep convective eyewall clouds, shows an important effect on the TC intensification in the HWRF simulations. This question may be addressed again in the context of the Carnot cycle of a TC in the control of outflow temperature. Emanuel (2011) showed that the importance of the outflow temperature in a TC system can be seen in the structure of the thermal wind relation for a balanced convectively vortex in which the winds in the boundary layer is approximately constrained by,

$$V_b^2 \cong \frac{C_k}{C_d} (T_b - T_o) \frac{(k_o^* - k)}{T_s}, \quad (5.2)$$

where V_b and T_b are the azimuthal velocity and temperature at which the angular momentum surface intersects the top of the boundary layer, and they are both the function of radius; T_o is the outflow temperature; C_k and C_d are the dimensionless exchange

coefficients for enthalpy and momentum; T_s is the sea surface temperature; and k_0^* and k are the saturation enthalpy at the sea surface and the actual enthalpy of air at the reference level at which the wind speed and exchange coefficients are defined, respectively. Note Eq. (5.2) is not closed solution for the gradient wind, since a) k is not known a priori. As shown in Figs. 5.6c and 5.6d, it changes as a TC vortex evolves; and b) the outflow temperature must be specified. In his early work, Emanuel assumed that the whole tropical troposphere is nearly neutral to moist convection and thus would have nearly constant saturation entropy. Since the absolute temperature can be nearly constant with height just above the tropopause, Emanuel treated the outflow temperature as a constant to the first approximation. But this create a dilemma, since the quantity $k_0^* - k$ usually increases monotonically with radius due to the fact that a TC vortex has a warm core, and thus, V_b would also have to increase monotonically with radius, which of course is not right. To address this obvious problem, the early work of Emanuel had to artificially set the boundary layer entropy low outside the storm's core. Emanuel (2011) recognized the problem of constant outflow temperature in this idealized model. Using numerical simulations by a nonhydrostatic, convection-permitting axisymmetric model, he showed that the outflow temperature, which is defined as the temperature at which the angular momentum surfaces pass through the zero contour of tangential winds, is not related in any obvious way to the ambient temperature at the tropopause, rather, the outflow temperature appears to be self-stratified. This is particularly true for the angular momentum surfaces originating outside the RMW. The temperatures at cross section where these angular momentum surfaces meet the zero tangential wind contour deviate substantially from the ambient temperature.

Emanuel (2011) and the subsequent Emanuel (2012) showed that the self-stratified outflow temperature poses a great impact on the structure and intensification of a TC.

Figure 5.7: Black contours: Azimuthal-mean absolute angular momentum (AAM) scaled by 10^7 (m^2/s); Color shades: Azimuthal-mean temperature (K); and Red contours: Azimuthal-mean tangential wind (m/s) at 12 UTC October 23, 2015 from the modified HWRf simulation of Patricia (2015).

Motivated by these studies, I examined the outflow temperature in the two HWRf simulations. Figure 5.7 shows the radius-height structure of the absolute angular momentum, tangential wind, and temperature at 12 UTC October 23, 2015 from the modified HWRf simulation of Patricia. It clearly shows that the outflow temperature of the absolute angular momentum surfaces erupting near the RMW is close to the ambient tropopause temperature, but the outflow temperatures of the absolute angular momentum surfaces originating outside the RMW have a wide range of temperature depending on the radii, and the values can deviate substantially from the tropopause temperature. In other

words, the outflow temperature of a TC is not a single constant, but has a stratification varying with the radius and height.

To understand the impact of the model physics on the outflow temperature, I examined the outflow temperature in the two HWRF simulations of Particia (2015). Figure 5.8 compares the outflow temperature of three absolute angular momentum surfaces originating from the eyewall and in the outer region of the vortex at different time during the evolution of the storm between the two HWRF simulations. Before the RI (Fig. 5.8a), all three absolute angular momentum surfaces have nearly the same outflow temperature, suggesting that the TC vortex in the two simulations roughly has the identical energy conversion efficiency provided that the surface layer temperature is also nearly the same at this time (Fig. 5.6). However, during the RI period (Figs. 5.8b and 5.8c), the outflow temperature of the absolute angular momentum surface originating from the eyewall in the modified HWRF simulation is notably lower than that in the default HWRF simulation. Since the surface temperature is also much higher in the modified HWRF simulation than that in the default HWRF simulation in this period, the energy conversion efficiency is higher in the modified HWRF simulation. This may explain why the storm underwent the RI in the modified HWRF simulation but does not in the default HWRF simulation. After the RI period, we see that the outflow temperature of three absolute angular momentum surfaces originating from the different radii again has similar value in the two HWRF simulations. It should be pointed out that the storm in the modified HWRF simulation is much stronger in the post-RI period than that in the default HWRF, and thus, the storm should gain more latent heating from the ocean surface in the former. However, since the energy conversion efficiency is low, the RI stops in the modified HWRF simulation despite

more energy gain from the ocean surface. Since the difference in outflow temperature is mainly associated with the absolute angular momentum surface originating from the eyewall, such a difference must result from the upper eyewall ice clouds. This explains why snow fall above the freezing level can have a large impact on TC intensification.

Figure 5.8: Azimuthal-mean absolute temperature (color shapes), three absolute angular momentum surfaces originating from the eyewall and in the outer region (thin red and blue dashed contours); and tangential wind (thick gray and black dashed contours) at different time before the RI (a), within the RI (b and c), and after the RI (d) from the two HWRf simulations. Solid and dashed contours indicate the modified and default HWRf, respectively. Symbol "x" and "o" indicates the point where absolute angular momentum surface crosses the zero tangential wind contour.

6 Summary and Conclusion

One of the greatest challenges in numerical weather prediction (NWP) is how to appropriately treat the sub-grid scale (SGS) processes, such as turbulent transport and microphysics. The problem is inherent not only in the use of discretized models to simulate the continuous atmosphere, which results in new second-order terms in the grid-box-mean governing equations of the atmosphere due to nonlinearity, but also in the latent heating release/absorption during water phase change. To close the system, these new second-order turbulence terms and microphysical properties must be determined empirically based on model-resolved quantities, a method known as parameterization. For operational NWP models with a horizontal grid-spacing at or greater than a few kilometers, the energy-containing turbulent eddies are not resolved, how to appropriately represent these unresolved eddies and microphysical processes operational models is a difficult problem facing meteorologists and computer scientists today. This dissertation research aims to address some of the key issues regarding the treatment of turbulent mixing processes and hydrometeor sedimentation to better represent the microphysical-dynamical interaction in the eyewall of TCs and how such an interaction affects the intensification process of storms. One of the specific problems that this dissertation tackles is how to appropriately parameterize the turbulent mixing in the eyewall of a TC. The problem stems from the unique characteristics of turbulence generation in the eyewall. Unlike the fair-weather conditions in which the turbulent PBL is cleanly separated from the free atmosphere above by a capping inversion, the turbulence in the eyewall is no longer solely generated by the shear production and buoyancy production associated with the PBL processes. It can also

be generated by cloud processes above the PBL in the eyewall clouds due to radiative cooling, evaporative cooling, and inhomogeneous diabatic heating and cooling. While the concept of PBL is still applicable in the eyewall as to the layer that is directly affected by the surface turbulent processes, the treatment of turbulent mixing must go beyond the conventional scope of the PBL since there is no physical interface that separates the turbulence generated by the PBL processes and cloud processes aloft. One of the focuses of this dissertation is to improve the vertical turbulent mixing scheme used in the Hurricane Weather Research and Forecasting (HWRF) system, one of the operational models used at the Environmental Modeling Center (EMC), NOAA for hurricane prediction. Our analyses show that the turbulent mixing scheme used in the operational HWRF is unable to represent the turbulence above the PBL in the eyewall. The problem arises from the fact that HWRF uses $N^2 = \frac{g}{\theta_v^e} \frac{\partial \theta_v^e}{\partial z}$ to calculate the Brunt-Väisälä frequency (BVF), a quantitative measure of the static stability of the atmosphere. However, this formula is only valid for the unsaturated atmosphere. It generates severe bias in static stability in the saturated eyewall clouds since it completely neglects the buoyancy production of turbulence induced by clouds. Based on the air parcel theory, we provide a detailed derivation of BVF for the saturated atmosphere and implemented formula in HWRF. Numerical experiments show that the static stability correction allows HWRF to successfully capture intense turbulent mixing above the PBL in the eyewall, and thus, substantially improves HWRF's skill in predicting rapid intensification (RI) of TCs.

Our analyses also show that the microphysics scheme used in HWRF is unable to produce hydrometeor particles with fall speeds less than 0.2 ms^{-1} , which are abundant in

real TCs according to the in-situ aircraft measurements (Marfarquar and Black 2004) collected in Hurricane Norbert (1984) and Emily (1987). This problem likely results from the current method of parameterizing hydrometeor particle fall speed, which was derived from a three-force balance among gravitational force, buoyance force, and drag in quiescent conditions. However, the quiescent condition of the environmental flow with zero vertical velocity is not valid in the eyewall and rainbands where the strong convective updrafts and downdrafts (e.g., 10 m/s) can exert a strong dynamic pressure force on particles. We hypothesize that neglecting the impact of strong eyewall convective currents on particle fall speed is the culprit for HWRF not producing hydrometeors with fall speed smaller than 0.2 m/s. This is because smaller particles are presumably more subjected to the background flow acting upon them. In this case, the fall speed of a particle should be governed by a four-force balance among gravitational force, buoyance force, drag, and dynamic pressure force. This four-force balance allows me to develop a new hydrometeor particle fall speed parameterization. Numerical experiments of HWRF show that the new scheme successfully produces hydrometeors with fall speed smaller than 0.2 m/s. The more realistic hydrometeor sedimentation in eyewall clouds allows HWRF to better simulate eyewall convection, and thus, improves model's skill in predicting RI of TCs. This improvement is rooted in the solid physical base of the new scheme that appropriately considers the impact of convective currents of convection on falling hydrometeor particles.

The results from this dissertation research suggests that TC intensification is sensitive to changes in model physics in HWRF simulations. Such a sensitivity is understandable since in a real TC the microphysical processes directly interact with the dynamical processes to determine TC inner-core structure and intensification rate. In

numerical models, however, microphysics and turbulent mixing are handled by separate modules, thus, how to robustly link the separated physics modules in a model becomes a key. In the PBL scheme currently used in HWRF, except for the water vapor mixing ratio, none of the microphysical properties are used when determining eddy exchange coefficients. Likewise, in the microphysical scheme, the model-resolved convective up-/down-drafts and the parameterized turbulent eddy circulations are not used when parameterizing various microphysical moments. Thus, there is no direct link between the microphysical and PBL modules at least at a model time step. The interaction between microphysics and dynamics is mainly indirect as all variables in a model are connected. We argue that lack of direct interaction between the separated microphysical and PBL modules prevents HWRF from simulating realistic inner-core structure of TCs. Our modifications to the model physics are developed from the fundamental physics underlying the processes to be parameterized. As it can be seen in Chapter 3, the calculation of in-cloud Brunt-Väisälä frequency requires inputs of mixing ratios of water vapor as well as hydrometeors. Similarly, the parameterization of particle fall speed requires inputs from the mode-resolved convective currents and parameterized turbulence. We believe that such a direct link among model-resolved convective currents, parameterized turbulent mixing and microphysics at a model time step alleviates the limitations of the PBL and microphysical schemes currently used in HWRF in terms of the direct interaction between them.

While the sensitivity to model physics is clearly demonstrated, the details of how microphysical processes interact with dynamical processes need to be explored. The HWRF simulations with the default and modified turbulent mixing scheme and

hydrometeor fall speed parameterization provide an avenue to look into the related issues. The budget analyses of azimuthal-mean tangential wind indicates that the model-resolved eddy forcing is the dominant forcing for the spin-up of the TC vortex at the current model resolution. However, the simulated TC inner-core structure, secondary overturning circulation, and the model-resolved eddy forcing show a strong dependence on the parameterized in-clouds SGS eddy processes above the PBL. The in-cloud turbulent mixing parameterization appears to facilitate the realization of axisymmetric dynamical mechanism underlying RI of TCs in 3D full-physics simulations. It suggests that the model-resolved and SGS eddy forcings are not independent, although they appear as two separate terms in the governing equations and are determined separately in numerical simulations. Such a dependence may result from the fact that the dynamical-microphysical interaction and large energy-containing turbulent eddies, such as kilometre and sub-kilometer convective elements and roll vortices, are not resolved but parameterized at a grid spacing of 2 km. Will further increasing of model resolution reduce the dependence of model-resolved fields on parameterized SGS processes? This question cannot be answered until the dynamical-microphysical interaction and large energy-containing eddies can be explicitly resolved. To do so, large-eddy resolution both horizontally and vertically is needed not only in the PBL (like classic LES) but also aloft in the eyewall and rainbands to resolve in-cloud turbulent eddies generated by cloud processes. This is not likely to happen in the near future for operational forecasts even with ever-increasing computational capability. Therefore, as model resolution keeps increasing, future research should be continuously devoted to improving parametric representation of model physics not only in the PBL but also above the PBL to appropriately account for microphysical processes, in-

cloud turbulent processes, and the interaction between microphysical and dynamical processes.

Using the HWRF experiments executed in this study, I analyzed the physical processes underlying the TC intensification. A key question I addressed here is why the air-sea interaction induced surface wind and heat exchange (WISHE) feedback mechanism works well to drive the RI of Patricia (2015) in the HWRF simulation with the upgraded model physics but fails to work in the default HWRF simulation. In other words, what is the triggering mechanism in the WISHE positive feedback underlying the TC intensification? I looked into this issue in the context of the Carnot cycle of a TC. At the temperature reservoir in the surface layer of a TC's Carnot cycle, our analyses show that the eyewall low-level convection is an important triggering. The budget analyses of tangential wind show that the model-resolved eddy forcing resulting from eyewall convection is a dominant contributor for accelerating the boundary layer winds. However, this acceleration by convection via eddy forcing may be overwhelmed by the deceleration of winds due to surface viscos dissipation, which is happened in the default HWRF simulation. In contrast, the acceleration of the boundary layer wind by the eyewall convection is large enough to overcome the surface viscos dissipation. Thus, we argue that there exists a critical strength of convection. Beyond this critical strength, low-level convection is able to kick off the WISHE positive feedback leading to the RI. Below this critical strength, the WISHE feedback cannot spontaneously occur of its own. In addition to the deceleration of boundary layer winds, the low-level convection also enhances the surface moisture convergence to increase the saturated equivalent potential temperature,

which ensure enhances the energy conversion efficiency of a TC's Carnot cycle. The combined dynamic and thermodynamic effect creates a pathway to the RI of Patricia (2015).

At the outflow temperature reservoir aloft of a TC's Carnot cycle, our simulations show that the outflow temperature of Patricia (2015) deviates substantially from the tropopause temperature depending on the absolute angular momentum surfaces originating at the different radii. This result is consistent with Emanuel (2011, and 2012) who regarded the outflow temperature of a TC as self-stratified. Our analyses show that the eyewall convection is able to modulate the outflow temperature of the absolute angular momentum surfaces erupting from the vicinity of radius of maximum wind (RMW). The strong eyewall convection evidenced from the large diabatic heating in the modified HWRF simulation notably reduces the outflow temperature of the absolute angular momentum surfaces originating near the RMW, and thus, provides another mechanism to enhance the energy conversion of the Carnot cycle of a TC.

The results obtained from this dissertation research depicts a pathway to RI of a TC. The steering mechanism of the pathway is the eyewall convection, which unfortunately shows a strong sensitivity to how turbulent mixing and microphysical processes in the eyewall are parameterized. In this study, I've been making effort to improve the representation of the physical processes related to turbulence and microphysics, but this is just a first step of a long match. I will continue devoting to my effort in improving model physics in numerical models.

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