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Influence of Cloud-Radiative Processes on Predecessor Rain Events

A dissertation submitted in partial satisfaction
of the requirements for the degree
Doctor of Philosophy in Atmospheric and Oceanic Sciences

by

Omar Angelo Nava

2016

ABSTRACT OF THE DISSERTATION

Influence of Cloud-Radiative Processes on Predecessor Rain Events

by

Omar Angelo Nava

Doctor of Philosophy in Atmospheric and Oceanic Sciences

University of California, Los Angeles, 2016

Professor Robert G. Fovell, Chair

Predecessor rain events (PREs) are coherent mesoscale rainstorms that occur well in advance of recurving tropical cyclones (TCs) and have a high potential to cause flooding and adverse societal impacts. In 2007, a PRE associated with TC Erin produced record-breaking rainfall (> 350 mm) across southern Minnesota and Wisconsin, contributing to seven fatalities and over \$170 million in property damage. A series of idealized numerical simulations is conducted using an aquaplanet version of the Weather Research and Forecasting model Advanced Research core (WRF-ARW) v.3.6 to examine the influence of cloud-radiative processes on the development of PREs.

This study finds that cloud-radiative feedback (CRF), the interaction of hydrometeor cloud species with longwave (LW) and shortwave (SW) radiation, produces a more robust PRE structure with stronger convective activity and, ultimately, more precipitation. It is demonstrated that LW cooling associated with clouds outside of the PRE region induces a stronger horizontal pressure gradient that enhances low level convergence and drives more vigorous ascent. Therefore, the primary radiation driver of PRE formation occurs outside of the PRE itself and is based on how the model responds radiatively to low clouds.

In addition, the warming component of CRF enhances parcel buoyancy and reduces vertical stabilities within the PRE structure. In particular, the distribution of cloud ice produces a greater depth of in-cloud, primarily LW warming conducive to stronger convective

processes. In contrast, SW CRF effects weaken PRE development through a combination of in-cloud cooling and low level cloud top warming, countering the LW CRF effects responsible for greater rainfall production. Moreover, PRE formation is found to be sensitive to the diurnal cycle, resulting in faster development at night and slower development during the day. A seasonal sensitivity also brings about weaker PRE intensification during the summer season; shorter nights and longer days minimize favorable LW CRF effects and maximize unfavorable SW CRF effects.

PRE development depends on the balance between subtle, competing processes involving radiation and microphysics schemes. The high sensitivity of PRE initiation and intensification to variations in CRF cannot be ignored and necessitates further research and observations to improve operational forecasts. Because the recipe for PRE development requires many individual ingredients to come together correctly, accurately predicting heavy rainfall events associated with tropical cyclones remains a daunting forecast challenge.

The dissertation of Omar Angelo Nava is approved.

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2016

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LIST OF ABBREVIATIONS

ARW	Advanced Research WRF	4
CAM	Community Atmosphere Model.....	48
CP	convective or cumulus parameterization.....	9
CRF	cloud-radiative feedback.....	6
FLG	Fu-Liou-Gu	48
HWRF	Hurricane Weather Research and Forecasting.....	6
JDN	Julian Day Number	14
LLJ	low-level jet	1
LW	longwave	6
MM5	fifth-generation Mesoscale Model	9
MP	microphysics parameterization.....	6
MSL	mean sea level.....	38
NASA	National Aeronautics and Space Administration	48
NCAR	National Center for Atmospheric Research.....	8
PRE	predecessor rain event	1
PV	potential vorticity	4
Q-G	quasi-geostrophic.....	3
QCLOUD	warm cloud	13
QICE	cloud ice	13
QIS	combination of QICE and QSNOW.....	59
QRAIN	warm rain.....	13
QSNOW	cloud snow	13
QVAPOR	water vapor	13

RRTMG	Rapid Radiative Transfer Model for General Circulation Models	9
SLP	sea level pressure	26
SST	sea surface temperature	12
SW	shortwave	6
TC	tropical cyclone	1
WRF	Weather Research and Forecasting	4
WSM5	WRF Single Moment 5-class	9
YSU	Yonsei University	9

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VITA

Major Omar A. Nava graduated from Jesuit College Preparatory School in Dallas, Texas in 2001. He entered undergraduate studies at the United States Air Force Academy in Colorado Springs, Colorado where he graduated with a Bachelor of Science degree in Physics in June 2005 and received his commission as a Second Lieutenant in the United States Air Force.

For his first assignment, Major Nava was selected to attend the Naval Postgraduate School in Monterey, California through the Air Force's Basic Meteorology Program. After graduating with a Bachelor of Science degree in Meteorology in June 2006, he was assigned to Scott Air Force Base in Belleville, Illinois as a Senior Duty Officer or Lead Meteorologist. In November 2008, Major Nava was assigned to Osan Air Base, South Korea where he served as the Wing Weather Officer for the 51st Fighter Wing. After 10 months in South Korea, he was selected to enter the Graduate School of Engineering and Management at the Air Force Institute of Technology at Wright-Patterson Air Force Base, Ohio in October 2009. After earning a Master of Science degree in Applied Physics in March 2011, he was assigned to the 614th Air and Space Operations Center at Vandenberg Air Force Base in Lompoc, California as the Flight Commander of the Weather Specialty Team.

Most recently, Major Nava was competitively selected to enter the Air Force Institute of Technology's Instructor Faculty Pipeline program and began his doctoral program in the Department of Atmospheric and Oceanic Sciences at the University of California Los Angeles in September 2013. Upon graduation, Major Nava will join the Department of Engineering Physics at the Air Force Institute of Technology as an Assistant Professor of Atmospheric Physics.

CHAPTER 1

Introduction

Predecessor rain events (PREs) are coherent mesoscale rainstorms that occur well in advance of recurving tropical cyclones (TCs), and possess significant potential to cause extreme rainfall totals and severe flooding (Cote 2007). PREs typically involve warm advection and frontogenetically forced ascent at the intersection of a low-level jet (LLJ) with a quasi-stationary low-level baroclinic zone. They form beneath the equatorward entrance region of an upper level jet streak and in the presence of low-level deep tropical moisture advection from a TC (e.g. Cote 2007; Galarneau et al. 2010; Bosart et al. 2012). Topographically influenced processes including orographic lifting, cold-air damming, and coastal frontogenesis can contribute to PRE development as well (Cote 2007; Srock and Bosart 2009). PREs are distinguished from other heavy rain producing weather phenomena by the following criteria determined by Cote (2007): 1) a rainfall rate ≥ 100 mm in a 24-hour period for a duration of at least 6 hours with accompanying radar reflectivity values greater than 35 dBz, 2) clear separation on radar imagery between the PRE and the main TC precipitation shield, and 3) the advection of deep tropical moisture away from the tropical storm and into the PRE region.

The high-impact nature of PREs and their inherent forecast challenge has been underscored by extreme rainfall associated with TC Frances (2004; Galarneau et al. 2010), TC Erin (2007; Galarneau et al. 2010; Schumacher et al. 2011; Schumacher and Galarneau 2012), and TCs Ike and Lowell (2008; Bosart et al. 2012; Schumacher and Galarneau 2012). The PRE initiated by TC Frances occurred over New York City and produced heavy rainfall of up to 175 mm, resulting in one death and \$6 million in damages (NCDC 2004). The TC Erin PRE generated a swath of record rainfall amounts exceeding 350 mm over Minnesota

(383.5 mm at Hokah), resulting in widespread severe flooding, over \$170 million in property losses, and seven fatalities (NCDC 2007). TCs Ike and Lowell also initiated three PREs that contributed to all-time daily record rainfall at many locations extending from northern Texas to the Great Lakes region (e.g. 189.5 mm at Lubbock, Texas; 261.9 mm at Wichita, Kansas; and 168.7 mm at Chicago O’Hare, Illinois). Because attention is usually directed at the adverse effects of landfalling TCs, it is possible for developing PREs to take operational forecasters by surprise (Moore et al. 2013).

The potential for PREs to cause severe flooding and adverse societal impacts necessitates accurate forecasts of their onset, duration, and rainfall intensity. However, operational numerical weather models may poorly predict the occurrence of PREs as a result of: 1) inadequately resolved mesoscale features and characteristics associated with heavy-rain producing convective systems (Davis et al. 2003; Jankov and Gallus 2004), and/or 2) misrepresentation of the dynamical and thermodynamic environmental influences of a transitioning TC moving into the midlatitudes (e.g. Atallah and Bosart 2003). This research uses an idealized numerical modeling approach to investigate the development of PREs, specifically with respect to the influence of cloud-radiative processes. While the characteristics, formation, and climatology of PREs have been documented extensively in recent literature, there have been limited studies using numerical models and no studies using an idealized modeling framework. The purpose of this research is to improve the understanding of PRE development in weather prediction models, thus giving operational forecasters the ability to provide more accurate and advanced warning to potentially affected communities.

1.1 Predecessor rain events

Even though the term “predecessor rain event” was only recently coined by Cote (2007), the concept has its origins in a study by Bosart and Carr (1978), which examined a heavy rain event ahead of TC Agnes (1972) that generated over 200 mm of rainfall across western New York and Pennsylvania. A distant region of heavy rainfall developed as moisture from TC Agnes streamed northward along the Appalachian Mountains towards a region of confluent

flow and forced ascent associated with an approaching mid-tropospheric shortwave trough. Their analysis demonstrated that the PRE formed remotely from the rain region directly associated with TC Agnes, and was dynamically driven by processes external to the TC.

These concepts were further expounded by Cote (2007), who conducted a climatology of 47 documented PREs associated with 21 Atlantic basin TCs between 1998-2006. The composite analysis and case studies showed that PREs tended to form in a region where a poleward moisture surge from a TC interacted with a region of forced lift along a low-level baroclinic zone and broad quasi-geostrophic (Q-G) ascent associated with an upper-tropospheric jet streak. In addition, the PREs were located about 1000 km poleward of the TC at the time of PRE initiation, and occurred approximately 36 hours prior to the arrival of the TC at the same latitude as the PRE. When the PREs were stratified by their position relative to the total observed track of the TC, the study noted that flooding impacts were exacerbated when the extreme rainfall occurred along the eventual track of the parent TC. Consecutive PREs that develop across the same general region over a period of several days can also produce widespread flooding (Bosart et al. 2012). In some cases, persistent moist upslope flow along orographic features, such as the Appalachian Mountains, served as a focal point for PRE development.

Galarneau et al. (2010) extended the work of Cote (2007) and presented a climatology that summarized the key characteristics of PREs through a composite analysis of 28 events from 1995-2008. The study found that a majority of PREs occurred under the equatorward entrance region of an anticyclonically curved 200-hPa jet and persisted for an averaged duration of 15 hours. PREs also formed most often during the months of August and September and were mostly initiated by TCs ranging in intensity from Tropical Storm to Category 2 according to the Saffir-Simpson scale. Galarneau et al. (2010) also conducted an observational case study of the TC Erin (2007) PRE that produced more than 350 mm of rainfall over southern Minnesota and documented the importance of the interaction between the low-level baroclinic zone and the poleward surge of moisture from the TC. They hypothesized that the presence of deep moisture from TC Erin likely evolved the PRE from a significant rain event into an unprecedented rain event.

Moore et al. (2013) also classified PREs associated with Atlantic basin TCs between 1988-2010 and identified three general categories: “jet in ridge”, “southwesterly jet”, and “downstream confluence” (see their Fig. 20). Brief case studies of PREs associated with TC Rita (2005), TC Wilma (2005), and TC Ernesto (2006) were presented as specific examples of the three categories, respectively. In contrast to the Galarneau et al. (2010) study, the classification procedure focused on the structure of the upper level jet streak and its location relative to the TC and other large-scale features (e.g. upper level troughs and ridges). The composite analysis indicated that the upper level low potential vorticity (PV) diabatic outflow from the PRE and/or the TC can strengthen the upper level jet streak by enhancing the meridional PV gradient (e.g. Atallah and Bosart 2003), reinforcing broad ascent over the PRE region. Overall, their results were still consistent with previous studies, indicating that PREs typically occurred near a low-level baroclinic zone, beneath the equatorward entrance region of an upper level jet streak, and in the presence of a stream of water vapor from a TC.

PREs have been documented in other parts of the world as well. Higgins and Shi (2005) examined the relationships between Gulf of California moisture surges and TCs in the eastern Pacific basin. Farfán and Fogel (2007), Corbosiero et al. (2009), and other studies have explored the role of eastern North Pacific TCs and their remnants on heavy rainfall events over the Southwestern United States and Northwestern Mexico. Waylen and Harrison (2005) studied the remote effects of TCs on the daily rainfall record in the Caribbean basins. Stohl et al. (2008) even described the transport of tropical and subtropical moisture associated with two former TCs that triggered an extreme precipitation event over the southwest coast of Norway.

1.2 Numerical modeling of PREs

Several recent numerical modeling studies have explored the dynamical processes and predictability of PREs. Wang et al. (2009) used the Advanced Research version of the Weather Research and Forecasting (WRF-ARW) model to investigate the remote effects

of Typhoon Songda (2004) on a heavy rainfall event in Japan. Two primary numerical experiments were performed. In the control experiment, a bogus vortex was inserted at an initial time similar to the observed TC intensity; in the no-typhoon experiment, the vortex associated with Typhoon Songda was removed by a smoothing algorithm. The study found that the presence of the tropical storm played a dominant role in the observed heavy precipitation in central Japan, even though the typhoon was more than 1200 km to the south over the western North Pacific. In another experiment, removing the terrain over Japan resulted in a 10% reduction in total rainfall, indicating that local orographic effects only played a secondary role in the particular case studied.

Schumacher et al. (2011) also used the WRF-ARW model to study the TC Erin (2007) PRE in an effort to quantify the distant effects of recurving TCs on midlatitude precipitation systems in the United States. As suggested by Galarneau et al. (2010), the presence of additional moisture from the tropical system served to significantly enhance the total storm rainfall accumulation. Instead of removing the entire TC as in the Wang et al. (2009) study, numerical simulations, in which the moisture plume associated with TC Erin was only removed, showed an approximately 25% reduction in the area-integrated PRE rainfall and a 50% reduction in the maximum PRE rainfall total compared to the control run. Even though the synoptic scale environment would have produced a heavy rain event across Minnesota and Wisconsin without the presence of an upstream TC, their results emphasized the importance of moisture originating from TC Erin in producing a record-breaking rain event.

Lastly, Schumacher and Galarneau (2012) examined differences in TC track predictions on the transport of moisture ahead of recurving TCs and their influence on precipitation forecasts. Ensemble forecasts from multiple operational forecasting centers were analyzed for PREs ahead of TCs Erin (2007) and Ike (2008). In order to isolate the effects of the TC position on PRE development, forecasts were placed in two groups: 1) members where the TC recurved at similar times and to similar locations as the observations, and 2) members where the TC took a substantially different track. Remarkably, probabilities of heavy rain between the ensemble groups were not significantly different, accentuating the forecast challenge of

predicting PREs at both the medium range (3-7 days) and short range (1-3 days) time scales. The study also noted that the timing, location, and magnitude of ascent associated with key synoptic scale and mesoscale features (e.g. baroclinic zones, upper level disturbances, and orography) were as important as the moisture transport from the TC in forecasting PRE development.

1.3 Cloud-radiative feedback

Cloud-radiative feedback (CRF) is defined as the modulation of atmospheric radiation due to presence of hydrometeors (e.g. cloud ice, snow, cloud droplets, rain, etc.). This sensitivity of radiation to microphysics, owing to how hydrometeors in different schemes interacted with radiative processes, was the finding of Fovell et al. (2010a). They conducted semi-idealized simulations using the WRF-ARW model and demonstrated that variations in the structure and motion of the TCs in their microphysics ensemble disappeared when clouds were made transparent to radiation. Since microphysics parameterization (MP) schemes differ with respect to the amount and relative distributions of hydrometeors (Fovell et al. 2010b), simulations can be sensitive to microphysical assumptions due to the interaction of different particle effective sizes with longwave (LW) and shortwave (SW) radiation (e.g. Dudhia 1989).

Using semi-idealized simulations with the Hurricane Weather Research and Forecasting (HWRF) model and an axisymmetric model, Bu et al. (2014) showed that CRF generates pronounced LW cooling along the anvil top of the TC which is partially countered by the absorption of SW radiation during the daytime. Weak LW warming within the cloudy regions essentially negated (and even reversed) the fairly large LW cooling in the troposphere under clear sky conditions. CRF also enhanced convective activity in the TC outer core, leading to a wider eye, broader tangential wind field, more radially extensive diabatic heating, and stronger secondary circulation (i.e. vortex “in-up-out” wind flow). The weak, primarily LW warming component of CRF within the cloud anvil directly forced a stronger upper tropospheric radial outflow as well as sustained ascent throughout the outer core. However,

CRF only had a minor and inconsistent impact on the maximum tangential wind speed.

Therefore, the influence of CRF on the TC structure and track can have an *indirect* influence on PRE development through differences in the magnitude of convergence against the low-level baroclinic zone and the interaction of the TC diabatic outflow with the upper level jet streak. However, we demonstrate that CRF also has a *direct* and positive influence on PRE development by modifying the environment surrounding the PRE region and enhancing convective processes within the PRE.

1.4 Organization of dissertation

This dissertation consists of this introductory chapter (Chapter 1), a description of the numerical model and experiment design used in this research (Chapter 2), two chapters detailing the influence of CRF on the initiation and intensification stages of PRE development (Chapter 3-4), and a summary (Chapter 5).

Chapter 2 introduces the WRF-ARW model employed in this study and motivates the “semi-idealized” approach to modeling. The experiment design and implementation of the PRE system components into the idealized modeling framework are also described in detail. In addition, the calculation of CRF effects in the model is briefly discussed.

Chapter 3 presents a numerical modeling examination of the influence of CRF on the initiation of PRE development. The contributions of the heating and cooling components of the CRF field are also examined, as well as the role of LW cooling and low-level clouds in driving processes culminating in PRE formation. The sensitivity of PRE development to variations in radiation schemes is also presented.

Chapter 4 investigates the influence of CRF on the intensification stage of PRE development. The effect of different hydrometeor species and their role in increasing PRE rainfall production is explored. Furthermore, the sensitivity of PRE development to SW radiation effects is examined, as well as the influence of diurnal cycle timing and seasonal length of day changes.

CHAPTER 2

Model and Experiment Design

This study employs a “semi-idealized” numerical modeling framework (e.g. Fovell et al. 2009; Hill and Lackmann 2009; Bu et al. 2014) to investigate the influence of CRF effects on PRE development. This approach facilitates analyses of weather phenomena while retaining the dynamical frameworks that might be employed in operational settings. This chapter briefly discusses the community-supported WRF model used for operational weather forecasting in the United States and describes the implementation of the PRE system components which were carefully designed to produce a realistic simulation of PRE development.

2.1 Weather Research and Forecasting model

The WRF-ARW model v.3.6 developed at the National Center for Atmospheric Research (NCAR) is a fully compressible, nonhydrostatic, three-dimensional model, with a terrain-following hydrostatic pressure coordinate in the vertical (Skamarock et al. 2008). The “real-data” version of the WRF-ARW model is rendered semi-idealized by removing all land (aquaplanet), initializing with a base state constructed from the Dunion (2011) moist tropical composite sounding, and retaining earth’s curvature. The model setup, similar to the methodology of Fovell and Su (2007) and subsequent related works, is guided by the famous adage attributed to Albert Einstein: “make things as simple as possible, but not simpler.”

The single domain consists of a 6000 km square region with 30 km horizontal resolution (200 x 200 grid), 31 irregularly spaced vertical levels, and a 50 mb model top. The 30 km resolution is too coarse to appropriately allow explicit convection to occur and necessitates the

specification of a convective (or cumulus) parameterization (CP) scheme to resolve subgrid convective processes (Skamarock et al. 2008). In fact, the use of a high resolution domain to explicitly resolve convection is not preferred for this study. Operationally, computational resources and higher resolution nests will be centered on the TC to improve its characterization as it pertains to forecasts of high winds, heavy rainfall, storm surge, and other pertinent threats to human lives and property. Therefore, it is imperative to investigate how CRF influences PRE development at a fairly coarse resolution. The use of periodic boundary conditions in the zonal direction and specified boundary conditions in the meridional direction creates a “tropical channel” favorable for the synoptic scale environment of the PRE system. The simulation is integrated over a period of 72 hours with a brief initialization period during the first 24 hours.

The RRTMG (Rapid Radiative Transfer Model for General Circulation Models) radiation scheme is specified for both LW and SW radiation (Iacono et al. 2008). In addition, the WRF Single Moment 5-class (WSM5) MP scheme is utilized, which explicitly predicts the mixing ratios of cloud water, rain, cloud ice, and cloud snow (Hong et al. 2004). Information from the radiation and MP schemes are used to calculate CRF effects, which are activated in the WRF-ARW model via the namelist parameter *icloud*. The Kain-Fritsch CP scheme (Kain 2004), Yonsei University (YSU) boundary layer parameterization (Hong et al. 2006), revised MM5 (fifth-generation Mesoscale Model) surface scheme (Jiménez et al. 2012), and Noah Land Surface Model (Chen et al. 1996) are also implemented in this study.

2.2 Components of the PRE system

The necessary components of the PRE system consist of an upper-tropospheric jet, low-level baroclinic zone, and a TC. Without an upper level jet and associated baroclinic zone, there is no focal point for convergence and forced ascent associated with the TC, and consequently, no PRE development. Conversely, without the presence of a TC, there is also no PRE development as indicated by the Wang et al. (2009) numerical modeling study. The PRE system components are implemented by specifying the idealized distribution of tem-

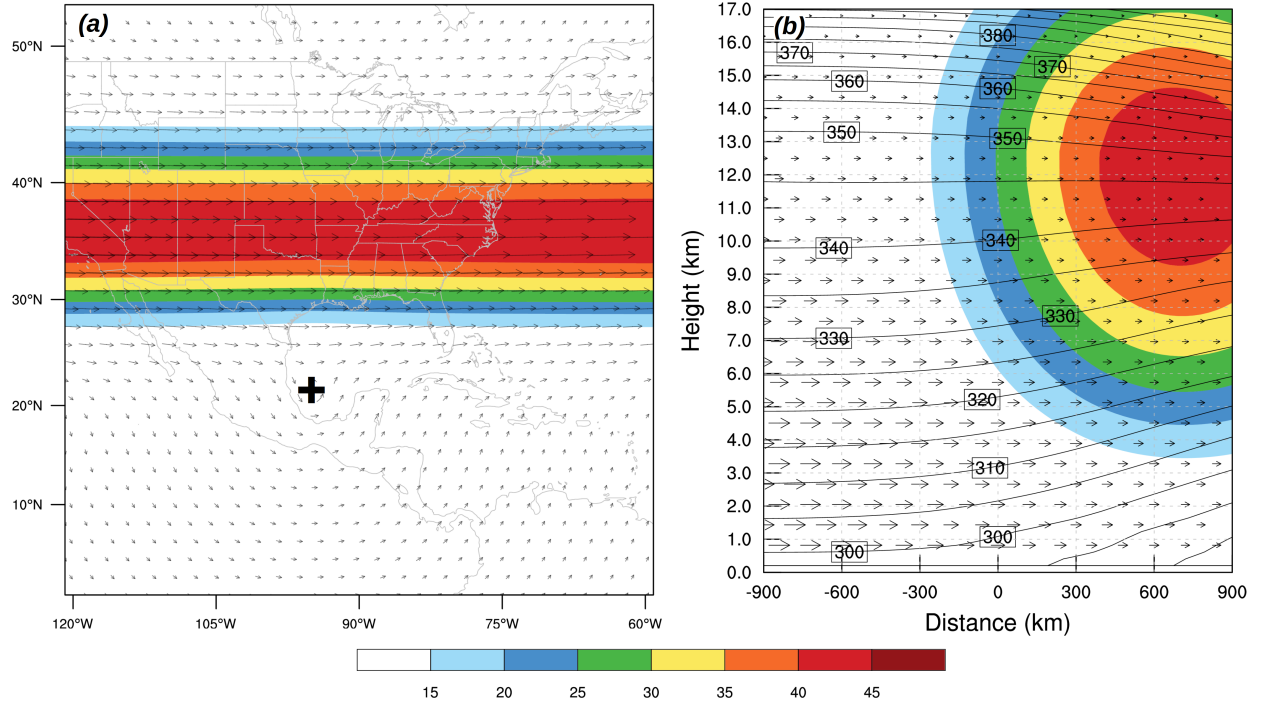


Figure 2.1: Initial (T+0) zonal wind speeds (m s^{-1} , shaded contours) (a) at 200 mb for the entire domain and (b) along a north-south cross-section at 70°W. The black plus sign (+) in panel (a) indicates the initial location of the TC bogus at 21°N and 95°W. Potential temperature (K, black contours every 5 K) in panel (b) shows the initial horizontal and vertical temperature distribution associated with the balanced upper level jet. Land outlines are for reference only; there is no land (aquaplanet).

perature, winds, pressure, and relative humidity at all grid points throughout the model domain.

2.2.1 Straight zonal jet and baroclinic zone

The balanced upper-tropospheric straight jet is centered at 36°N with zonal winds increasing with height to a maximum value of 50 m s^{-1} at the 175 mb pressure level (Fig. 2.1). Once the shape of the upper level straight jet is specified, the corresponding mass and temperature fields are derived to be in geostrophic and hydrostatic balance following the

Constant	Value	Units
u_0	50	m s ⁻¹
p_0	10 ⁵	Pa
Δz_0	5	km
z_0	22	km
z_1	30	km
a	6.371 x 10 ⁶	m
R	287	J K ⁻¹ kg ⁻¹

Table 2.1: Numerical values of the parameters used to characterize the zonal jet

methodology of Polvani et al. (2004). The zonal wind distribution is specified as

$$u(\phi, p) = u_0 \sin^{10}(\pi \mu^2) F(z), \quad (2.1)$$

$$\mu = \sin(\phi + \frac{\pi}{20}), \quad (2.2)$$

$$z = -H \log(\frac{p}{p_0}), \quad (2.3)$$

where u is the zonal wind distribution according to latitude ϕ and pressure p . The variable u_0 is the maximum zonal wind speed (m s⁻¹) and z is the height above the surface (km). Instead of assuming a constant scale height (H) of 7.34 km, as in Polvani et al. (2004) study, the scale height is calculated at each model level by the relationship

$$H = -(\frac{1}{p} \frac{\partial p}{\partial z})^{-1}. \quad (2.4)$$

The distribution of the zonal wind field $u(\phi, p)$ also deviates from the Polvani et al. (2004) study by raising the sine function to the 10th power in order to produce a more realistic meridional temperature gradient associated with the low-level baroclinic zone. In the original methodology, the sine function is only raised to the 3rd power which produces a zonal jet and associated horizontal temperature gradient that is too weak to be conducive to PRE development. The vertical structure of the zonal jet is characterized by the function $F(z)$, defined as

$$F(z) = \frac{1}{2} [1 - \tanh^3(\frac{z - z_0}{\Delta z_0})] \sin(\frac{\pi z}{z_1}), \quad (2.5)$$

where numerical values for the parameters u_0 , p_0 , Δz_0 , z_0 , and z_1 are listed in Table 2.1.

In order to calculate the temperature distribution associated with the balanced upper level jet, the meridional momentum and hydrostatic balance equations are used. The meridional momentum equation relates the zonal wind speed and geopotential height gradient as a function of latitude and can be written as

$$(af + u \tan \phi)u = -\frac{\partial \Phi}{\partial \phi}, \quad (2.6)$$

where a is the average radius of the Earth (m), f is the Coriolis frequency or parameter (s^{-1}), and Φ is the geopotential height (m). Hydrostatic balance is defined as

$$\frac{\partial \Phi}{\partial z} = \frac{R}{H}T, \quad (2.7)$$

where R is the gas constant for dry air ($\text{J K}^{-1} \text{kg}^{-1}$) and T is temperature (K). Combining both the meridional momentum and the hydrostatic balance equations yields

$$\frac{\partial T}{\partial \phi} = -HR^{-1}(af + 2u \tan \phi)\frac{\partial u}{\partial z}. \quad (2.8)$$

Integrating Eq. 2.8 results in a temperature distribution as a function of latitude and height:

$$T(\phi, z) = \int^{\phi} \frac{\partial T(\phi', z)}{\partial \phi} d\phi + T_0(z), \quad (2.9)$$

where $T_0(z)$ is the base vertical temperature distribution specified by the Dunion (2011) moist tropical sounding.

The balanced temperature distribution in Eq. 2.9 is calculated assuming a dry domain environment (no moisture). Experiments with no TC indicate that the unperturbed jet remains stable with no significant deviations throughout the 72-hour simulation, even with moist processes and sea surface fluxes (not shown). A time-invariant sea surface temperature (SST) gradient is specified to match the near-surface air temperature gradient, preserving the low-level temperature structure throughout the simulation (Ritchie and Elsberry 2001). Equatorward of the baroclinic zone, the maximum SST is 26.8°C and decreases in the poleward direction according to the meridional temperature gradient of the lowest model level (not shown). Overall, the synoptic scale arrangement in Fig. 2.1 closely resembles the “Southwesterly Jet” configuration described in Fig. 1b of Moore et al. (2013) and the TC Wilma (2005) PRE case study analyzed in the same paper.

2.2.2 Tropical cyclone bogus

To initiate PRE development, a TC is placed south of the upper level zonal jet at 21°N and 95°W using the TC bogus algorithm introduced in WRF version 3.0 for cold start runs with background fields (Davis and Low-Nam 2001). The TC bogus scheme generates a vertically erect axisymmetric modified Rankine vortex. The tangential wind V is specified as a function of radius R according to the equation

$$V(R) = V_0 \left(\frac{R}{R_m} \right)^{-\alpha}, \quad (2.10)$$

where the maximum wind speed V_0 is 20 m s⁻¹, radius of maximum wind R_m is 60 km, and wind profile decay factor α is 0.5 (e.g. Anthes 1982). The TC bogus is assumed to be in gradient and hydrostatic balance (i.e. no radial wind flow), except in the boundary layer due to frictional effects and in the outflow layer where strong divergence exists (Wang et al. 2013).

Cao et al. (2011) demonstrated that the WRF model had difficulty maintaining the initial TC bogus storm structure over time. However, the main purpose of the TC in this study is to advect deep tropical moisture towards the PRE region and interact with the mid-latitude baroclinic zone. The values associated with the modified Rankine vortex were selected because they produced a realistic looking simulation of PRE development for analysis. Investigating the influence of different TC characteristics is a possible topic for further research and is related to the PRE forecasting issue of representing the dynamical and thermodynamic environmental influences of a transitioning TC moving into the midlatitudes (Atallah and Bosart 2003).

2.3 CRF effects

The calculation of CRF effects by the RRTMG radiation scheme consists of LW radiation, SW radiation, warm cloud (QCLOUD), cloud ice (QICE), cloud snow (QSNOW), and cloud fraction (Iacono et al. 2008). Water vapor (QVAPOR) and warm rain (QRain) are not utilized in the calculation of the CRF field, in regards to the RRTMG radiation scheme. The

RRTMG LW and SW radiation schemes are called every 30 minutes (only during the day for SW), which is appropriate for the 30 km horizontal resolution of the domain (Skamarock et al. 2008). While the CRF calculation incorporates hydrometeor information from the WSM5 microphysics scheme, only the RRTMG radiation scheme code is modified in this study.

In the SW radiation scheme, the day of the year, or Julian Day Number (JDN), specifies the magnitude and distribution of incoming SW radiation associated with the seasonal cycle. The 72-hour simulation begins on 20 September (JDN 263) around the Autumnal Equinox, producing approximately equal lengths of day and night. Even though most PREs occur during the months of August and September (Galarneau et al. 2010), PREs can still develop through the month of October as evidenced by TC Wilma (2005) which produced a PRE on 23 October (Moore et al. 2013). The day of the year is not applicable to the LW radiation scheme since clear sky and LW CRF effects are always active.

CHAPTER 3

Influence of CRF on PRE Development

Bu et al. (2014) and Fovell et al. (2010a) demonstrated that CRF effects can have a significant influence on the TC structure, and consequently, TC motion. CRF enhances convective activity in the outer core of the TC, leading to a wider eye, a stronger secondary circulation, and a broader tangential wind field that also contributes to TC track variations via beta drift. Therefore, it stands to reason that CRF has an indirect influence on PRE formation through changes in the TC wind field and track. However, CRF also plays a *direct* role in a series of physical processes that drive PRE development.

The first section demonstrates that the WRF-ARW model and experiment design are able to produce a realistic-looking simulation of PRE development that is consistent with observations and climatology. Next, an experiment is conducted where CRF effects are deactivated and compared with the control experiment. The contribution of the CRF warming and cooling components are investigated individually, as well as the influence of low-level clouds on the initiation of PRE development. Lastly, variations in PRE development due to different radiation schemes are briefly explored. The experiments listed in Table 3.1 are described in more detail throughout the chapter.

3.1 Control experiment

The evolution of the total accumulated precipitation in the control experiment (labeled CRF-on) shows the development of a swath of heavy rainfall oriented in the zonal direction along the low-level baroclinic zone (Fig. 3.1). The precipitation distribution also bears strong resemblance to the PRE generated by TC Wilma (2005) documented in the Moore

Experiment label	Description
CRF-on	CRF effects active
CRF-off	No CRF effects, only clear sky radiation
CRF < 0	Only cooling component of CRF
CRF > 0	Only warming component of CRF
1km-LWCRFp0	No LW cooling below 1 km, only LW warming
CRF-south	Only LW cooling below 1 km south of 31.5°N
CRF-north	Only LW cooling below 1 km north of 31.5°N
Jet-control	CRF-on experiment with no TC
Jet-icloud0	CRF-off experiment with no TC

Table 3.1: List of PRE initiation experiments

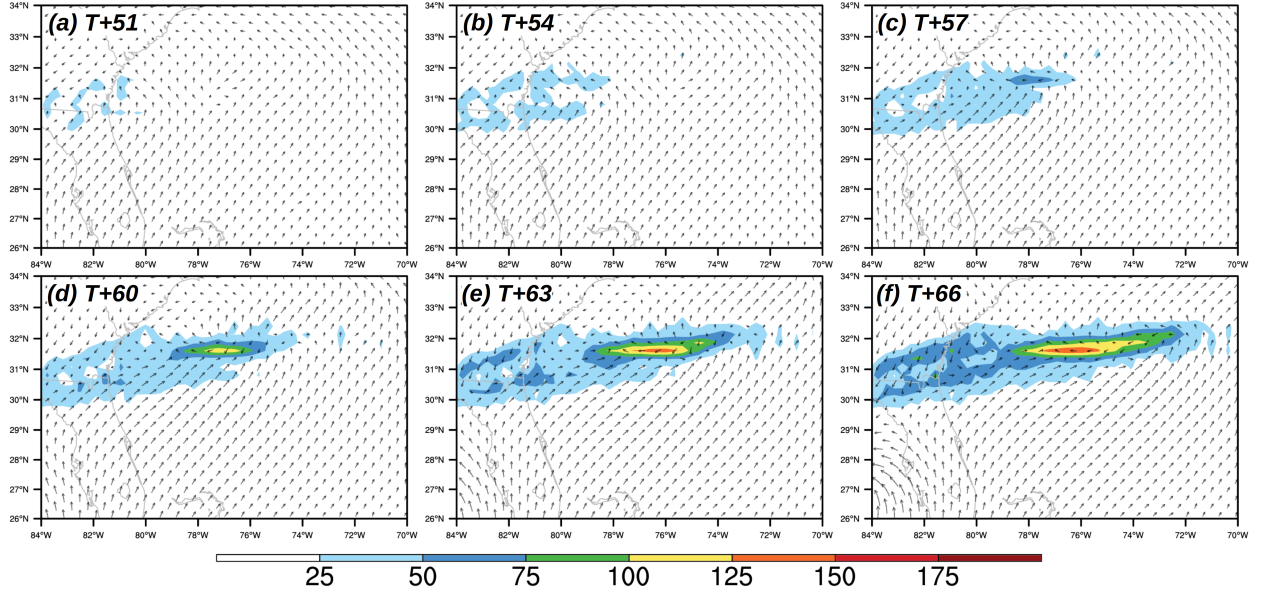


Figure 3.1: Total precipitation accumulation (mm) since the beginning of the simulation at (a) T+51 through (f) T+66 hours for the CRF-on experiment. Wind vectors represent surface winds.

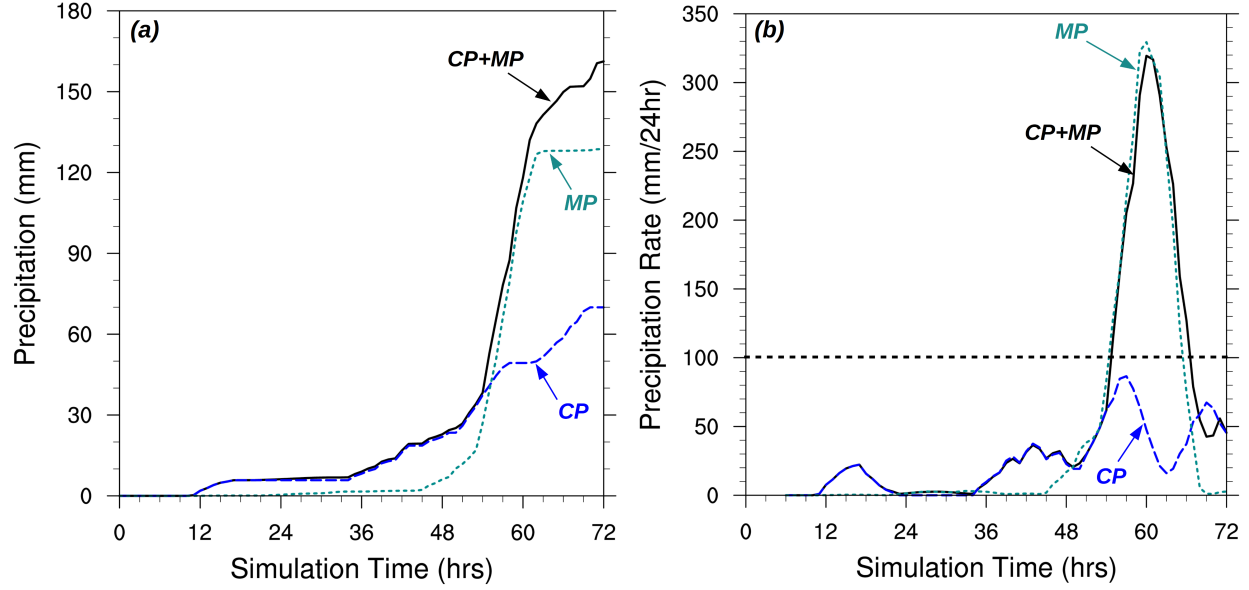


Figure 3.2: Line plots of (a) maximum PRE rainfall (mm) and (b) precipitation rate [mm (24 hr)^{-1}] over time for the CRF-on experiment produced by both CP and MP schemes (solid black), only the CP scheme (dashed blue), and only the MP scheme (fine-dashed cyan). The horizontal dashed line in panel (b) indicates the threshold rainfall rate for PRE classification.

et al. (2013) study. The first signs of PRE development appear at T+54 hours (Fig. 3.1b), and eventually produce over 150 mm of rainfall over a 12-hour period (Figs 3.1c-f). The separation distance between the initial location of PRE development and the TC center is about 1200 km, which is consistent with climatology (e.g. Galarneau et al. 2010).

The total precipitation is separated into contributions from the CP and MP schemes (Fig. 3.2a). Initially, the CP scheme generates most of the rainfall across the PRE region until T+54 hours, when the MP scheme becomes the dominant rain producing component and rapidly increases the maximum precipitation in the PRE over 90 mm in a 6-hour period. According to Cote (2007) and subsequent studies, PRE development initiates when the rainfall rate exceeds 100 mm in a 24-hour period for a duration of at least 6 hours (Fig. 3.2b). In the CRF-on experiment, PRE initiation occurs at T+54 hours. The period prior to this point in time is designated as the “PRE initiation” stage. After T+54 hours, the PRE rapidly intensifies until it reaches its peak rainfall rate at T+60 hours. This 6-hour period

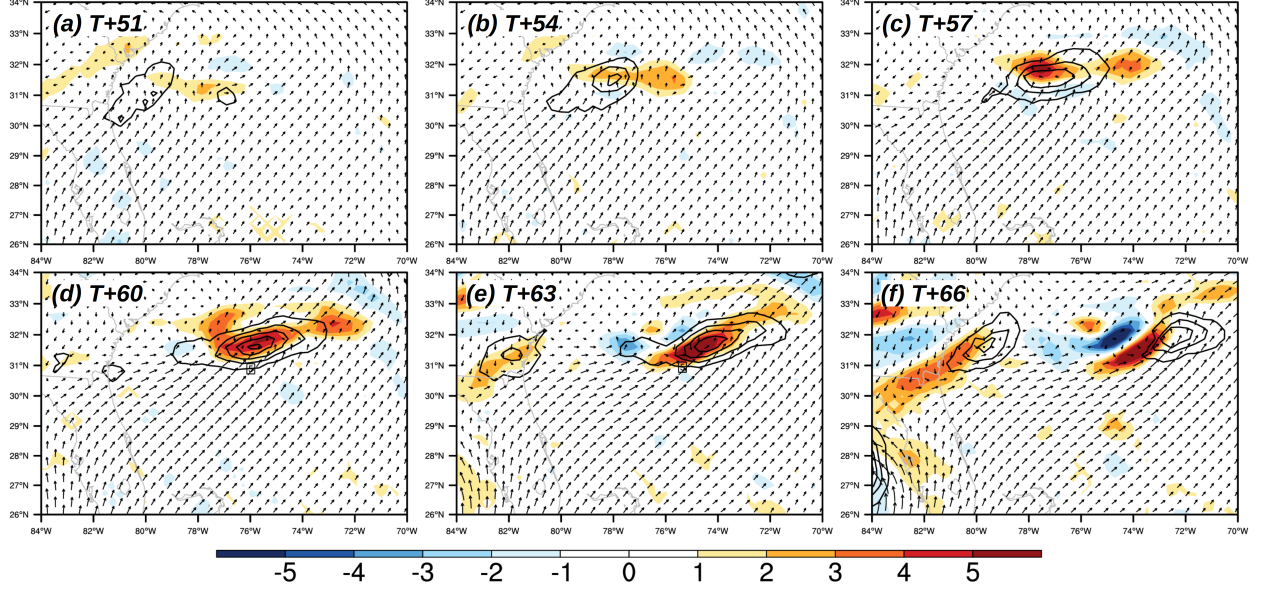


Figure 3.3: Frontogenesis [$\text{K (100 km}^{-1}) (3 \text{ hr}^{-1})$, shaded contours] averaged between the surface and 850 mb at (a) T+51 through (f) T+66 hours for the CRF-on experiment. Upward vertical motions (cm s^{-1} , black contours every 5 cm s^{-1}) are averaged from the surface to 200 mb. Wind vectors represent surface winds.

of time is denoted as the “PRE intensification stage”. The PRE then terminates at T+66 hours when the rainfall rate decreases below the PRE classification threshold. The 12-hour duration of the PRE in the CRF-on experiment is consistent with climatology (Galarneau et al. 2010).

A key feature of PRE development is strong frontogenesis along the low-level baroclinic zone (e.g. Galarneau et al. 2010; Moore et al. 2013), where the magnitude of the horizontal temperature gradient increases (e.g. due to low-level converging winds) and causes an imbalance in the thermal wind, or the vertical distribution of wind shear, associated with the upper level jet. Consequently, an adjustment to the geostrophic wind flow occurs at all levels, generating divergence aloft and convergence near the surface. Mass continuity then induces upward vertical motions along the baroclinic zone, resulting in adiabatic cooling that aims to relax the enhanced horizontal temperature gradient and restore thermal wind balance. This series of processes associated with frontogenesis occurs simultaneously and produces an

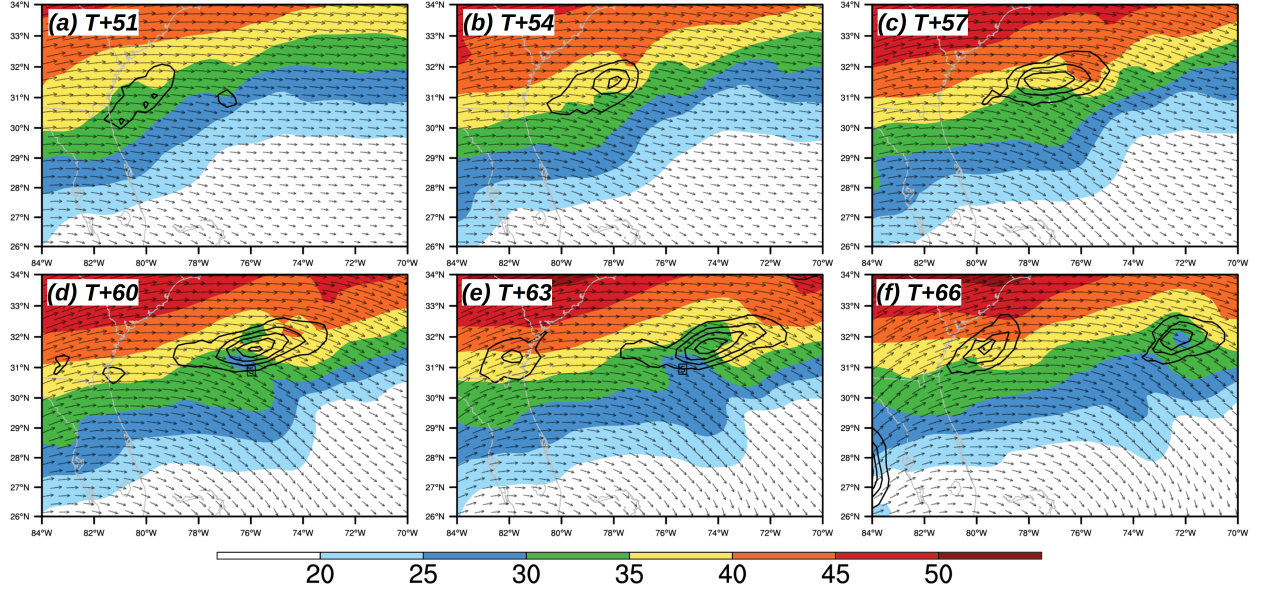


Figure 3.4: 200 mb wind speeds (m s^{-1} , shaded contours) at (a) T+51 through (f) T+66 hours for the CRF-on experiment. Upward vertical motions (cm s^{-1} , black contours every 5 cm s^{-1}) are averaged from the surface to 200 mb. Wind vectors represent 200 mb winds.

observable thermally direct circulation (Bluestein 1992; Holton and Hakim 2012). Furthermore, an enhancement to the horizontal temperature gradient is also characterized by an increase in the upper level jet wind speeds.

At the time of PRE initiation (T+54 hours), there is a clear signature of positive frontogenesis and a “bullseye” of strong upward vertical motions over the initial PRE formation location (Fig. 3.3b). At the peak PRE rainfall rate (T+60 hours), the magnitude of the upward vertical motions and frontogenesis are both at a maximum (Fig. 3.3d). As the precipitation rate in the PRE decays, the locations of frontogenesis and strong upward vertical motions move to the east (Figs. 3.3e,f), extending the PRE rainfall in a linear streak along the low-level baroclinic zone. Consistent with Q-G theory, frontogenesis in the CRF-on experiment also increases the wind speeds in the upper level jet, most notably on the eastern edge of the enhanced upward vertical motion signatures (e.g. Fig. 3.4c).

Strong frontogenesis is generated by converging southerly winds against the low-level baroclinic zone, which increases the near-surface horizontal temperature gradient over time

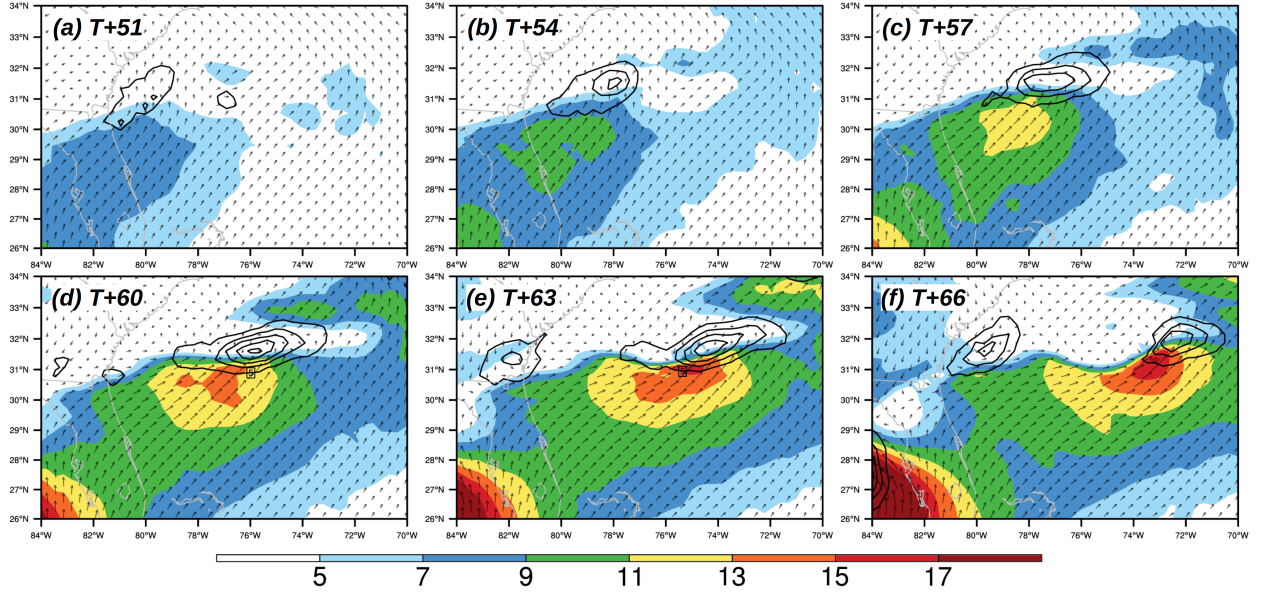


Figure 3.5: Surface-850 mb averaged wind speeds (m s^{-1} , shaded contours) at (a) T+51 through (f) T+66 hours for the CRF-on experiment. Upward vertical motions (cm s^{-1} , black contours every 5 cm s^{-1}) are averaged from the surface to 200 mb. Wind vectors represent surface winds. Strong wind speeds in the bottom left corner of the panels are associated with the TC cyclonic wind field.

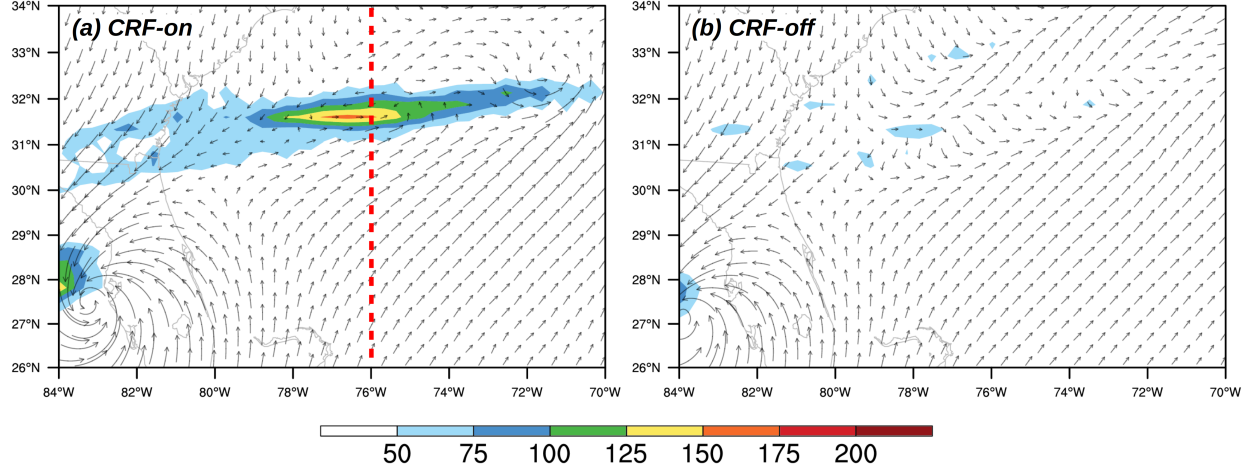


Figure 3.6: Total precipitation accumulation (mm) at T+72 hours since the beginning of the simulation for (a) CRF-on and (b) CRF-off experiments. Wind vectors represent surface winds. The dashed line (red) indicates the location of the vertical cross-section referenced in later figures.

(Fig 3.5). In fact, during heavy rain events, it has been documented that an LLJ oriented perpendicular to a surface boundary can enhance the magnitude of low-level warm advection, moisture convergence, and vigorous ascent driving strong convection (e.g. Maddox et al. 1979; Augustine and Caracena 1994; Trier et al. 2006). At T+54 hours, the increased low-level wind flow (i.e. LLJ) and convergence (Fig. 3.5b) coincide with stronger frontogenesis and upward vertical motions across the PRE region (Fig. 3.3b). The enhancement in the low-level wind speeds across the PRE region continues through the peak PRE intensity at T+60 hours, and propagates eastward along the low-level baroclinic zone (Figs. 3.5d-f), similar to the PRE rainfall signature. The presence of an LLJ on the leading edge of the developing PRE in the idealized simulation is consistent with observations in the PRE composite analyses of the Galarneau et al. (2010) study.

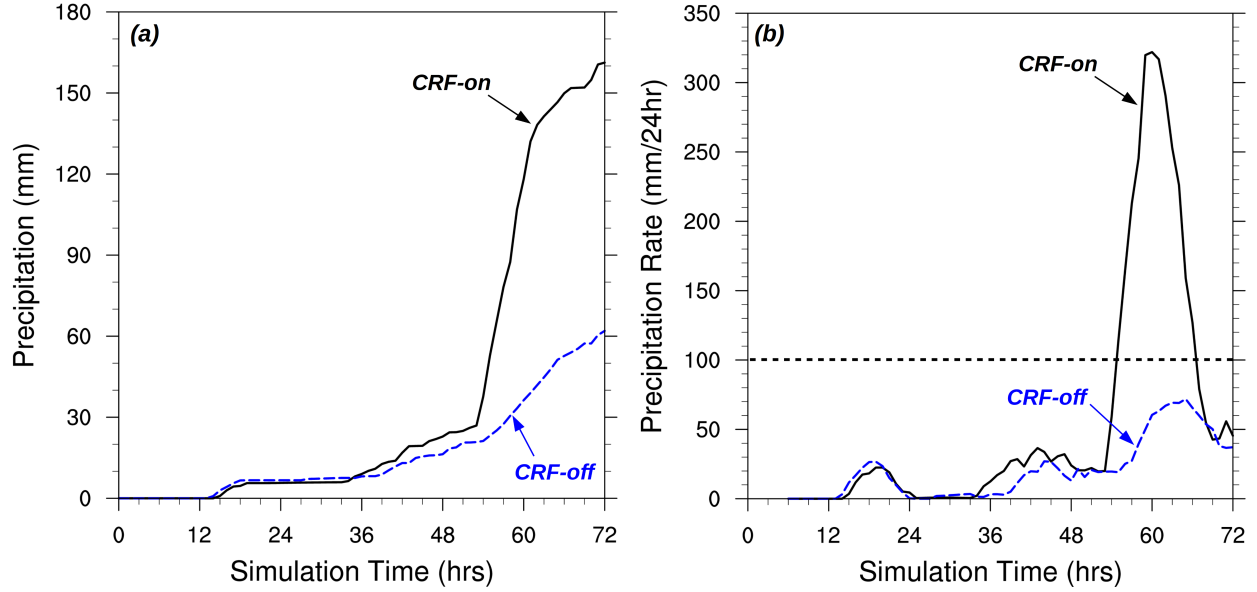


Figure 3.7: Line plots of (a) maximum PRE rainfall (mm) and (b) precipitation rate [$\text{mm} (24 \text{ hr})^{-1}$] over time for CRF-on (solid black) and CRF-off (dashed blue) experiments. The horizontal dashed line in panel (b) indicates the threshold rainfall rate for PRE classification.

3.2 Comparison with no CRF effects

Clearly, the idealized experiment design is able to produce a realistic-looking simulation of PRE development with CRF effects active (CRF-on). Next, an experiment where cloud radiative processes are deactivated (labeled CRF-off) is compared with the CRF-on experiment. Without CRF effects, the hydrometeor species generated by the MP scheme (e.g. warm cloud, cloud ice, cloud snow, etc.) are transparent to SW and LW radiation, resulting in only clear sky heating and cooling effects throughout the domain.

The difference in total accumulated precipitation at the end of the 72-hour simulation indicates a large decrease in PRE rainfall when CRF effects are disabled (Fig. 3.6b). While the precipitation in the CRF-on experiment is concentrated in a narrow swath along the low-level baroclinic zone, the CRF-off experiment only produces sparse regions of moderate rainfall (Fig 3.6). The significant difference in the precipitation distributions between the two experiments has serious implications for operational forecasts of PRE development.

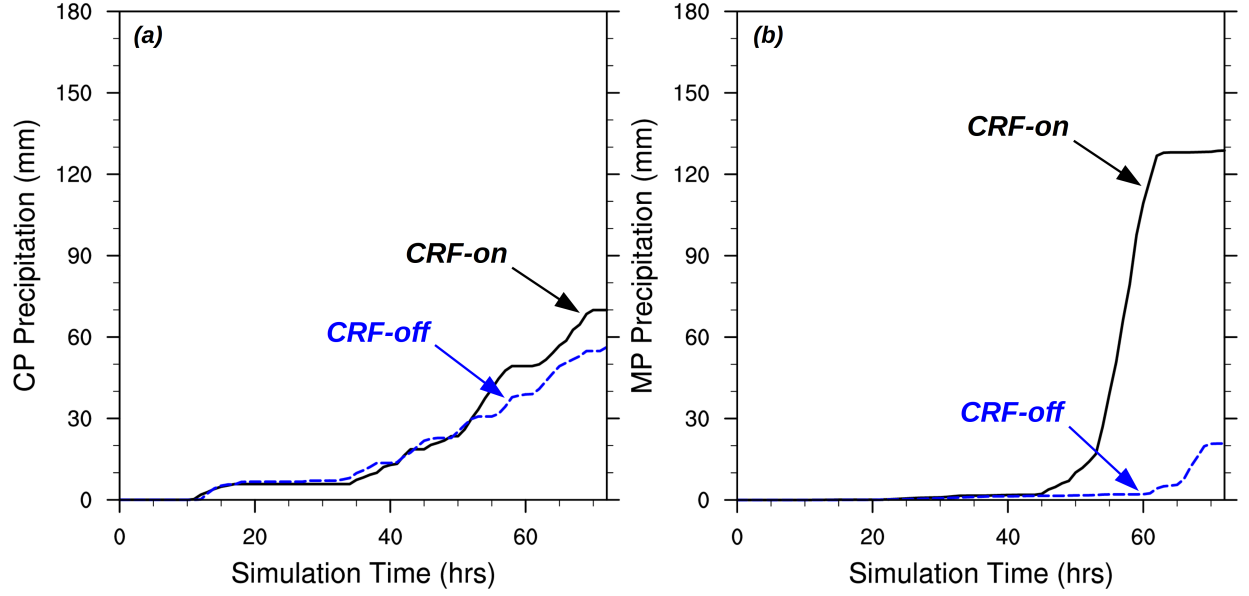


Figure 3.8: Line plots of maximum PRE rainfall (mm) over time produced by the (a) CP scheme and (b) MP scheme for CRF-on (solid black) and CRF-off (dashed blue) experiments.

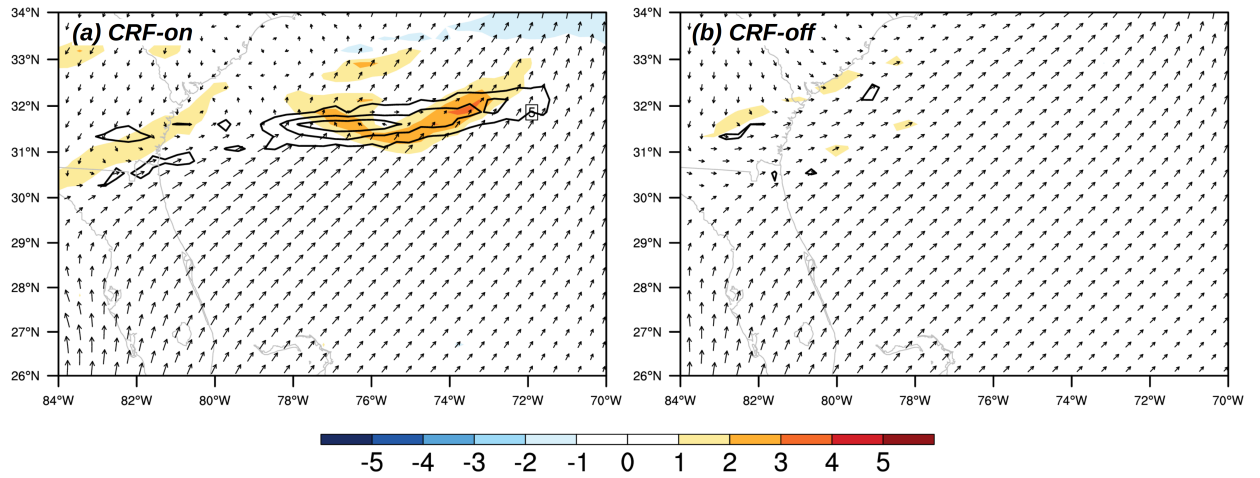


Figure 3.9: Frontogenesis [K (100 km⁻¹) (3 hr⁻¹), shaded contours] averaged from the surface to 850 mb between T+54 and T+66 hours for (a) CRF-on and (b) CRF-off experiments. Upward vertical motions (cm s⁻¹, black contours every 5 cm s⁻¹) are averaged from the surface to 200 mb. Wind vectors represent surface winds.

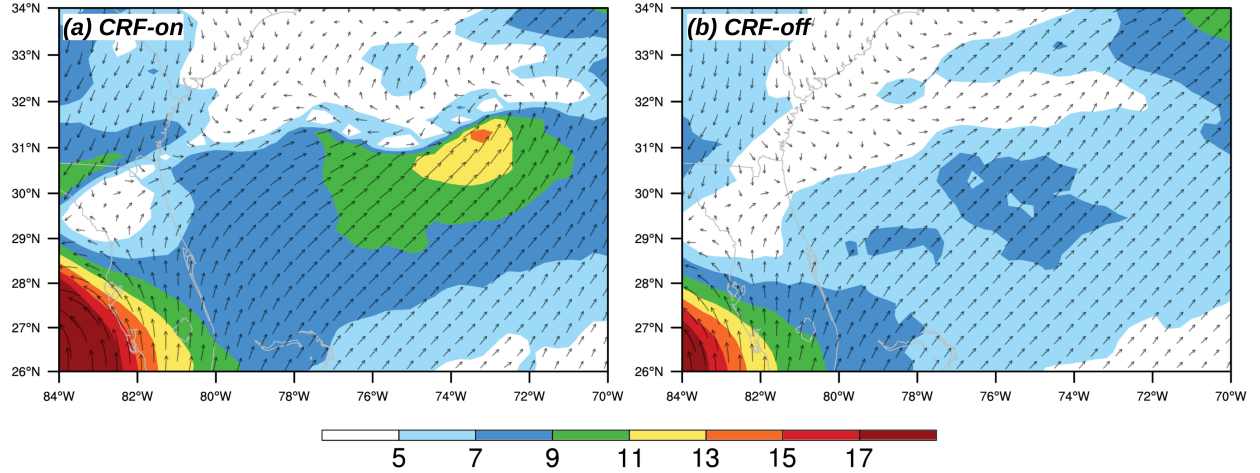


Figure 3.10: Averaged surface-850 mb wind speeds (m s^{-1}) between T+54 and T+66 hours for (a) CRF-on and (b) CRF-off experiments. Wind vectors represent surface winds. Strong wind speeds in the bottom left corner of the panels are associated with the TC cyclonic wind field.

The maximum rainfall in the CRF-on experiment is about 100 mm greater than in the CRF-off experiment (Fig. 3.7a). The reduced precipitation in the CRF-off experiment is associated with a lower rainfall rate that never crosses the threshold criteria for classification as a PRE (Fig. 3.7b). The large disparity in rainfall production by the MP scheme indicates that CRF effects have much more influence on microphysical processes compared to the subgrid convective processes associated with the CP scheme (Fig. 3.8b). This result highlights a strong link between the hydrometeor information from the MP scheme used in the calculation of CRF and the influence of CRF on the precipitation generation by the MP scheme.

The significant difference in the magnitude of frontogenesis between the experiments (Fig. 3.9) can be attributed to reduced low-level wind speeds and weaker convergence across the PRE region when CRF effects are ignored in the CRF-off experiment (Fig. 3.10b). The strong PRE inflow winds in the CRF-on experiment are clearly separate from the main TC cyclonic wind field and develop without an obvious direct extension from the TC circulation (Fig. 3.5). This suggests that a stronger TC may not necessarily produce a stronger PRE,

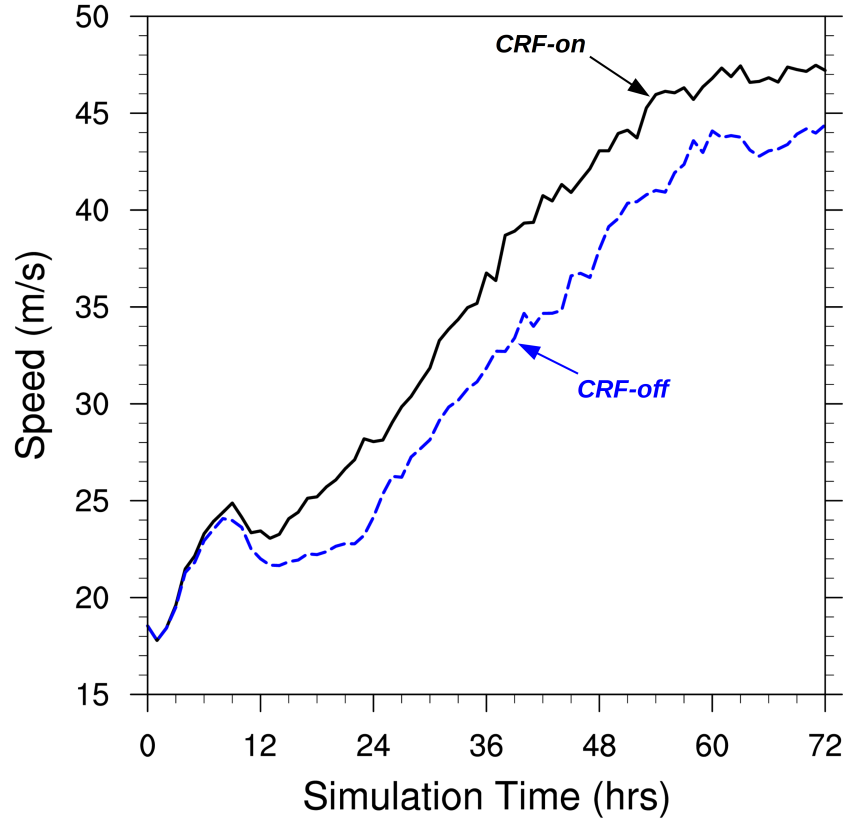


Figure 3.11: Maximum TC wind speed (m s^{-1}) over time for CRF-on (solid black) and CRF-off (dashed blue) experiments.

and that factors outside of the main TC cyclonic circulation may be responsible for the enhancement of the low-level winds across the PRE region.

Bu et al. (2014) demonstrated that CRF effects broaden the wind field in the TC outer core located 200 km or more from the storm center. In the CRF-on experiment, PRE development initially occurs about 1200 km from the TC center, much farther than the region of enhanced TC cyclonic winds. Comparison of the maximum TC wind speeds over time indicates a small difference of $4\text{--}5 \text{ m s}^{-1}$ (Fig. 3.11), which is most likely not responsible for the large 100 mm difference in PRE rainfall between the CRF-on and CRF-off experiments. The 100 km variation in TC track between the two experiments is also minimal (not shown). Therefore, the intensification of the low-level wind field and increased convergence across the PRE region must be generated locally by CRF effects.

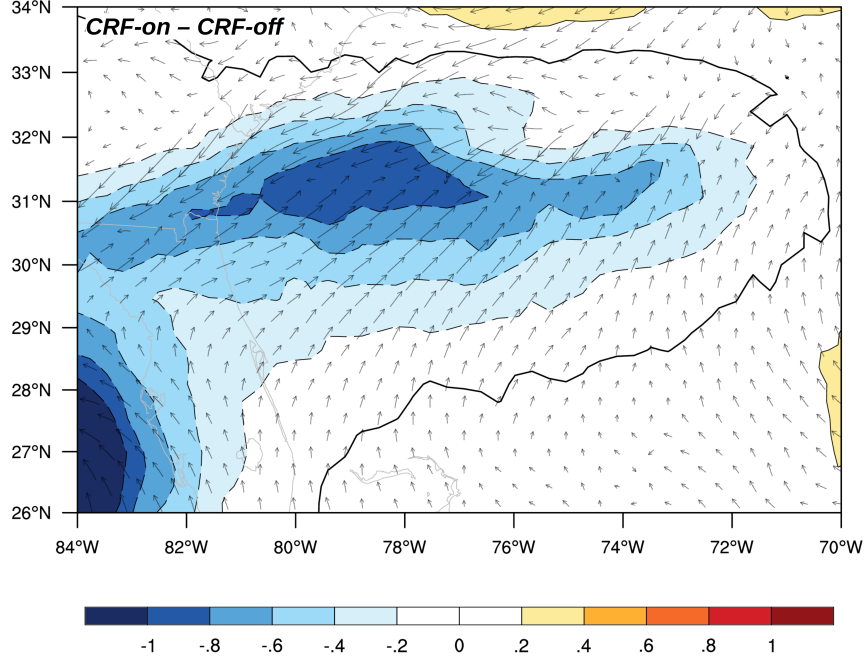


Figure 3.12: SLP difference (mb) and surface wind vector difference averaged between T+54 and T+66 hours for CRF-on and CRF-off experiments.

Differences in the magnitude of the PRE inflow winds can be attributed to differences in the pressure distribution across the PRE region. The CRF-on experiment indicates lower sea level pressures (SLPs) along the low-level baroclinic zone and higher pressures outside of the PRE region, extending several hundred kilometers to the north and south (Fig. 3.12). The locally enhanced pressure gradient accelerates the low-level wind flow, strengthens convergence across the PRE region, increases upward vertical motions, and leads to the production of more precipitation. Consistent with the results of the Schumacher and Galarneau (2012) numerical modeling study, the direct role of CRF on PRE development confirms that the TC may not have the largest influence on whether or not PRE formation occurs.

Cross-sections highlight the distribution of radiative heating and cooling between the CRF-on and CRF-off experiments in three general regions of the PRE structure (Fig. 3.13c): 1) LW cooling in the upper levels associated with the PRE cloud tops, 2) in-cloud warming in the lower and middle levels of the storm, and 3) strong LW cooling near the surface surrounding the PRE. The wind vector difference field also indicates increased convergence at

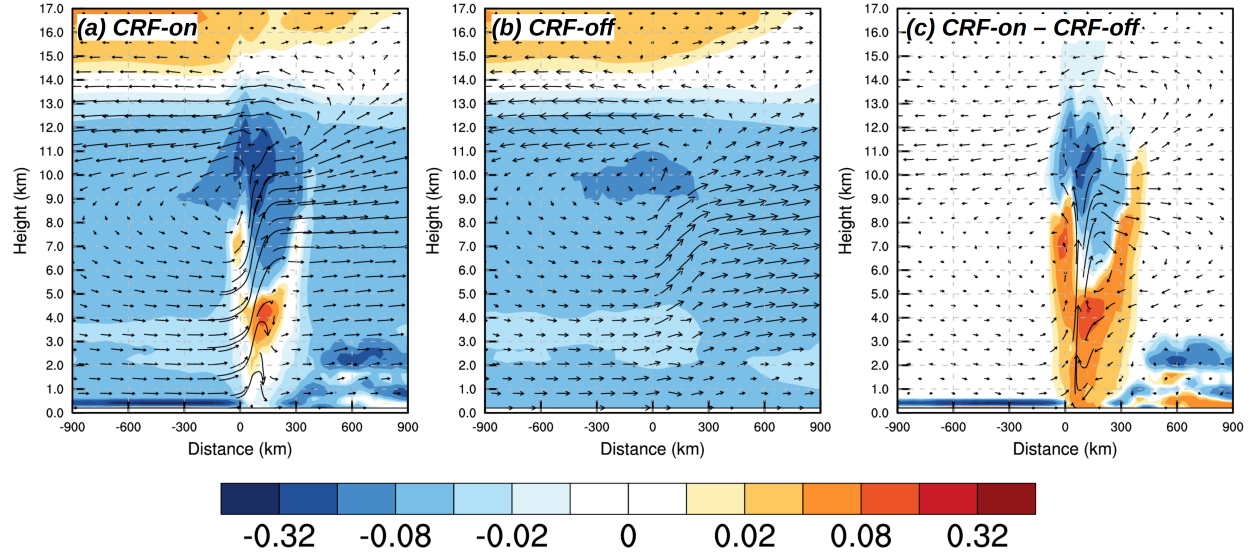


Figure 3.13: Radiative heating tendency (K hr^{-1}) cross-sections averaged between T+54 and T+66 hours for (a) CRF-on, (b) CRF-off, and (c) CRF-on - CRF-off experiments. The position of the north-south cross-section is indicated in Fig. 3.6a and centered on the region of heavy rainfall exceeding 100 mm in the CRF-on experiment at 31.5°N and 76°W . The cross-section is also averaged across a 300 km east-west longitude. Vertical winds are multiplied by a factor of 100 (cm s^{-1}) to emphasize upward vertical motions. Note the log2 scaling of the contour values.

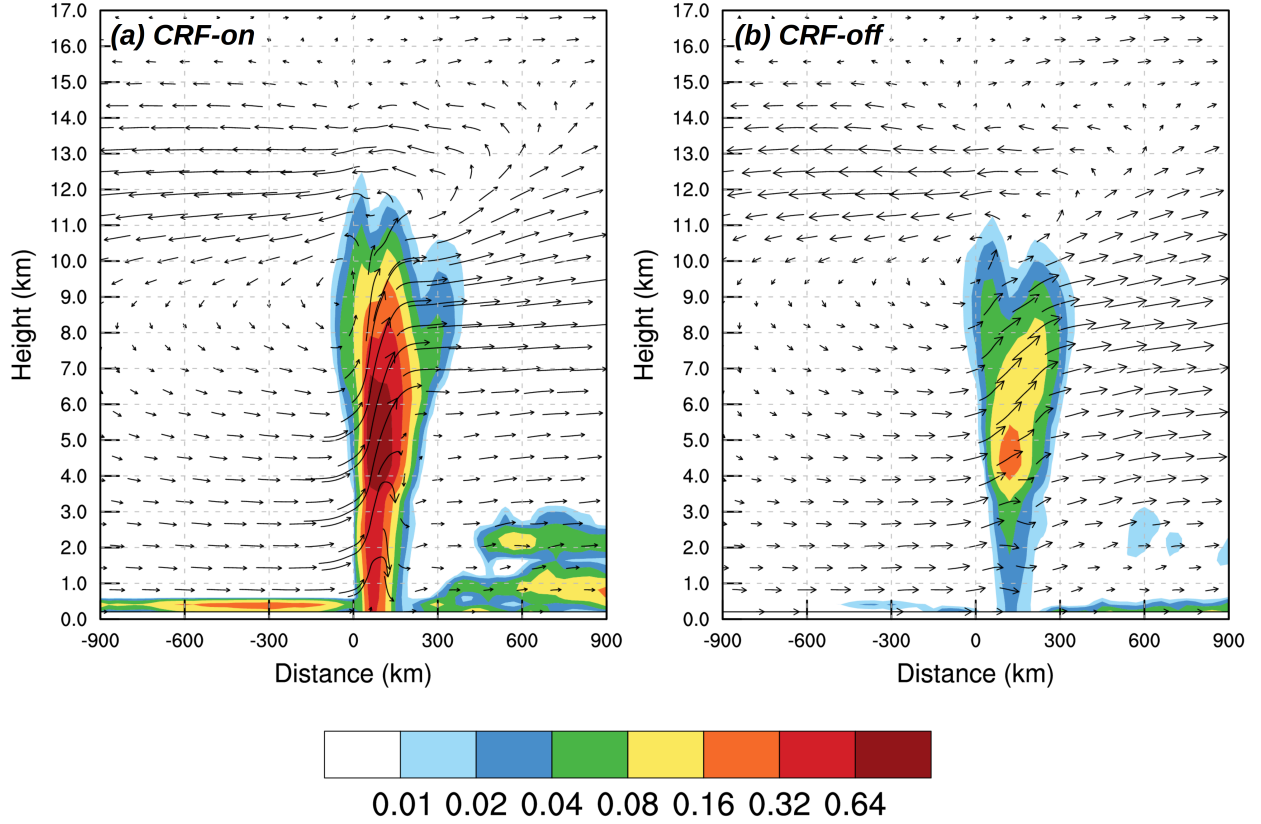


Figure 3.14: Condensate (g kg^{-1}) cross-sections as in Fig. 3.13 for (a) CRF-on and (b) CRF-off experiments. Condensate generated by the WSM5 microphysics scheme consists of warm cloud (QCLOUD), cloud ice (QICE), cloud snow (QSNOW), and rain (QRain).

the base of the PRE, more divergent wind flow aloft, and stronger upward vertical motions. Consequently, the enhanced convective secondary circulation associated with the CRF-on experiment generates much more condensate everywhere in the PRE structure (Fig. 3.14a). The low-level clouds to the north and south of the PRE region are also associated with the LW cooling signatures near the surface (Fig. 3.13a). It is important to note that differences in the distribution of cloud species can have a significant influence the calculation of the CRF heating and cooling field (Fovell et al. 2010a).

Enhanced condensate production by the MP scheme also releases more latent heat favorable for strong convective processes (Fig. 3.15c). In the CRF-on experiment, a strong MP diabatic heating signature extends from the surface to about 12 km and is mainly

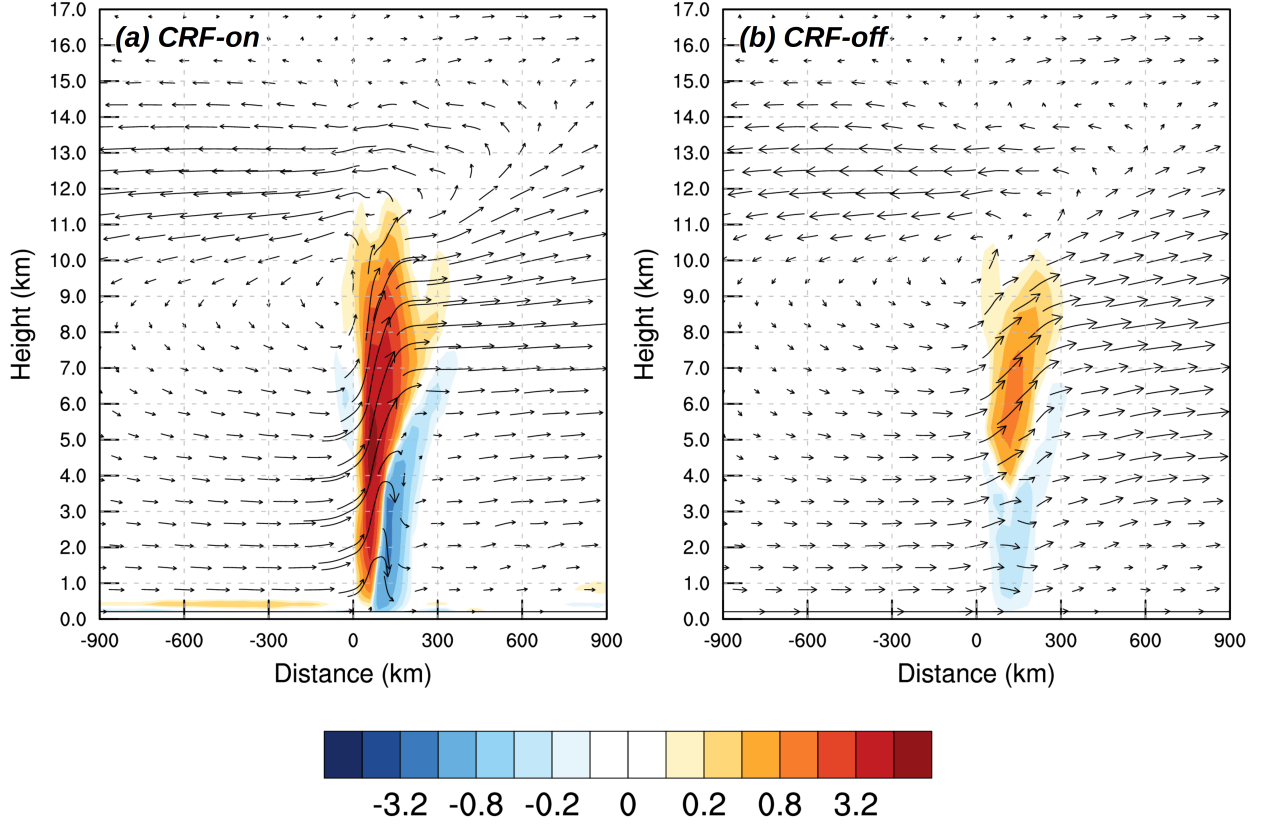


Figure 3.15: MP heating tendency (K hr^{-1}) cross-sections as in Fig. 3.13 for (a) CRF-on and (b) CRF-off experiments.

concentrated between 4-6 km, due to enhanced latent heat release as parcels ascend past the freezing level. In contrast, weaker upward vertical motions in the CRF-off experiment generate much less MP diabatic heating, only extending between 4-10 km within the PRE structure. Between the CRF-on and CRF-off experiments, the relatively small difference in radiative heating tendency drives a significant difference in the MP heating tendency, about one order of magnitude greater. Therefore, CRF effects elicit a notable response from the MP scheme that results in more condensate production, greater release of latent heat, and stronger convective processes. This key interaction must be accurately represented in order to improve forecasts of PRE development, and is directly related to the problem of properly representing mesoscale features and processes associated with heavy rain producing events in the model (Davis et al. 2003; Jankov and Gallus 2004).

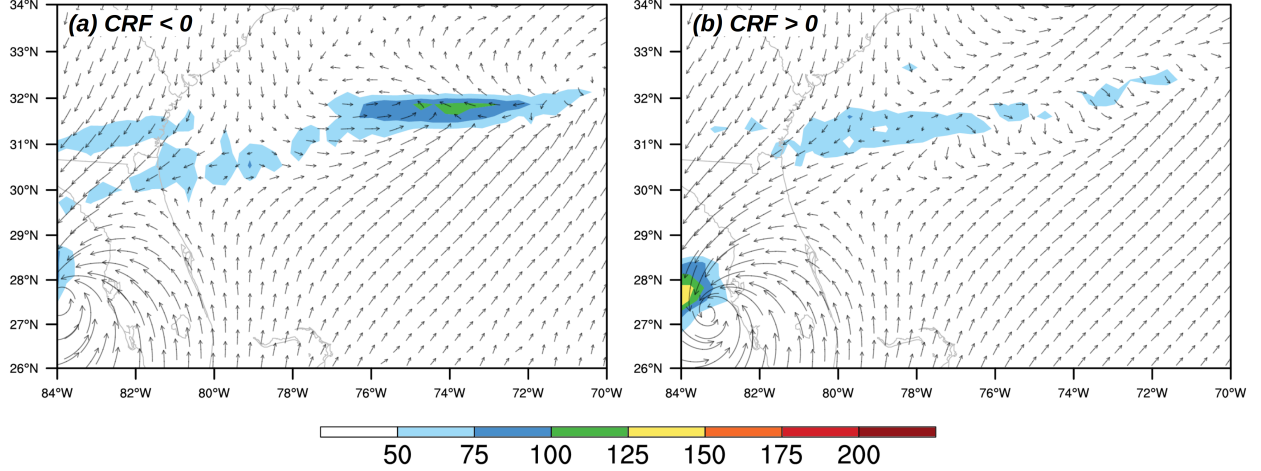


Figure 3.16: Total precipitation accumulation (mm) at T+72 hours as in Fig. 3.6 for (a) $CRF < 0$ and (b) $CRF > 0$ experiments.

3.3 PRE initiation

CRF clearly has a significant influence on the intensity and evolution of PRE development, resulting in two distinct stages of formation denoted as PRE initiation and PRE intensification. Through a series of experiments, this section explores the role of CRF in generating favorable environmental conditions that eventually manifest in the initiation of PRE development. The influence of CRF effects on PRE intensification will be investigated in the following chapter.

3.3.1 CRF heating and cooling components

CRF warming and cooling components in the CRF-on experiment are first isolated to investigate their influence on PRE initiation individually, similar to the methodology in the Bu et al. (2014) study. The experiment labeled $CRF > 0$ only retains the warming part of the CRF calculation from both LW and SW radiation schemes, and sets all CRF cooling (negative) values equal to zero. In contrast, the experiment labeled $CRF < 0$ only keeps the cooling component of the CRF calculation and sets the warming values to zero.

A comparison of the total precipitation accumulation shows that the cooling component

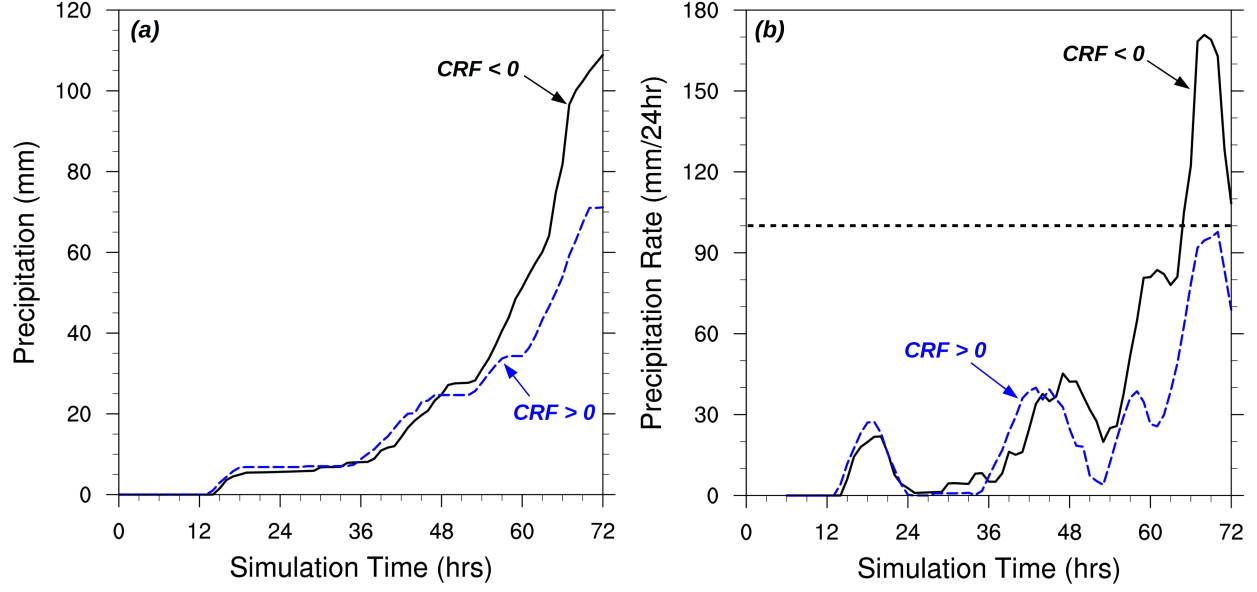


Figure 3.17: Line plots of (a) maximum PRE rainfall (mm) and (b) precipitation rate [$\text{mm} (24 \text{ hr})^{-1}$] over time as in Fig. 3.7 for $\text{CRF} < 0$ (solid black) and $\text{CRF} > 0$ (dashed blue) experiments.

of CRF ($\text{CRF} < 0$) produces a stronger PRE overall (Fig. 3.16), generating about 40 mm more rainfall (Fig. 3.17a) than the warming component of CRF ($\text{CRF} > 0$). This result is counter-intuitive, as local warming is typically associated with rising motions favorable for convection and precipitation production; local cooling induces downward vertical motions that hinder convective processes. The $\text{CRF} < 0$ experiment eventually intensifies into a PRE by the end of the 72-hour simulation, but the $\text{CRF} > 0$ experiment never crosses the threshold rainfall rate to be classified as a PRE (Fig. 3.17b).

The distribution of radiative heating in the $\text{CRF} < 0$ experiment indicates cloud top cooling in the upper levels of the PRE structure, as well as LW cooling along the low-level clouds both north and south of the PRE region (Fig. 3.18a). In contrast, warming in the $\text{CRF} > 0$ experiment is generated within the cloudy regions, extending vertically throughout the PRE structure (Fig. 3.18b). The difference field indicates stronger upward vertical motions associated with the $\text{CRF} < 0$ experiment, even with widespread cooling across the PRE region (Fig. 3.18c). LW cooling at the top of the PRE structure induces

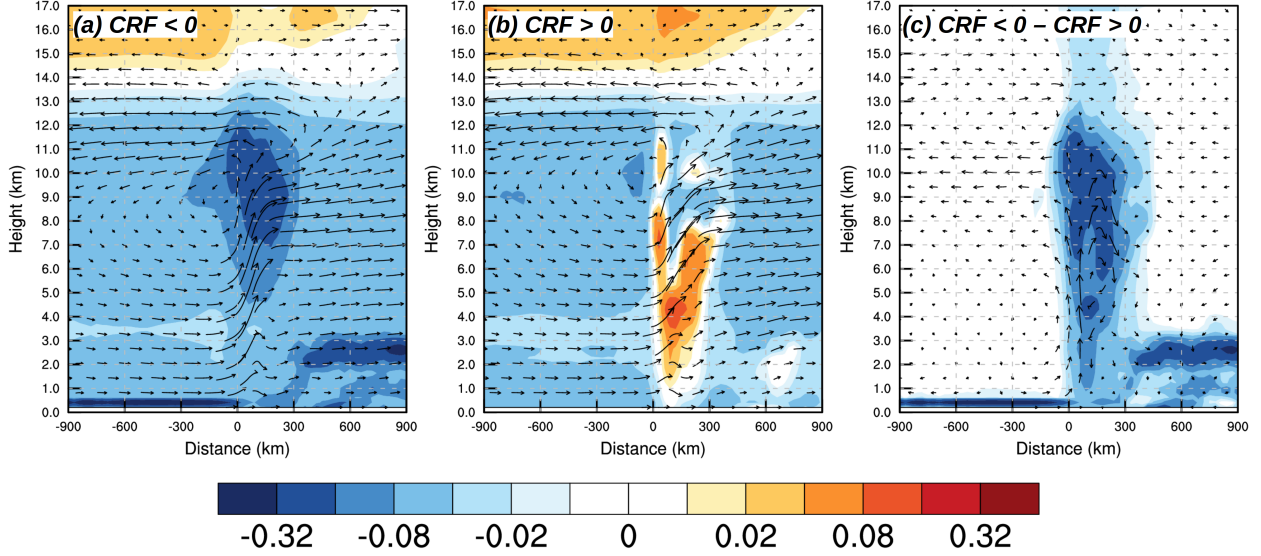


Figure 3.18: Radiative heating tendency (K hr^{-1}) cross-sections as in Fig. 3.13 for (a) $\text{CRF} < 0$, (b) $\text{CRF} > 0$, and (c) $\text{CRF} < 0 - \text{CRF} > 0$ experiments.

downward vertical motions that hinder convective processes. Therefore, the low-level LW cooling signatures outside of the PRE must be responsible for the enhanced upward vertical motions and stronger PRE development.

Similar to the CRF-on experiment, the LW cooling near the surface in the $\text{CRF} < 0$ experiment is associated with the low-level clouds south of the PRE region and more elevated clouds to the north (Fig. 3.19a). Whereas, in the $\text{CRF} > 0$ experiment, low cloud development is largely non-existent without LW cooling effects. The $\text{CRF} < 0$ experiment also indicates stronger low-level winds and enhanced convergence across the PRE region (Fig. 3.20a), which results in stronger frontogenesis favorable for PRE development (not shown). Even with strong CRF warming throughout the PRE structure in the $\text{CRF} > 0$ experiment, PRE development is much weaker without the favorable low-level LW cooling effects that enhance convergence across the PRE region and drive stronger upward vertical motions.

Another examination of the magnitude of the TC wind field confirms that TC intensity is not necessarily a driving factor in PRE development. Even though the $\text{CRF} < 0$ experiment

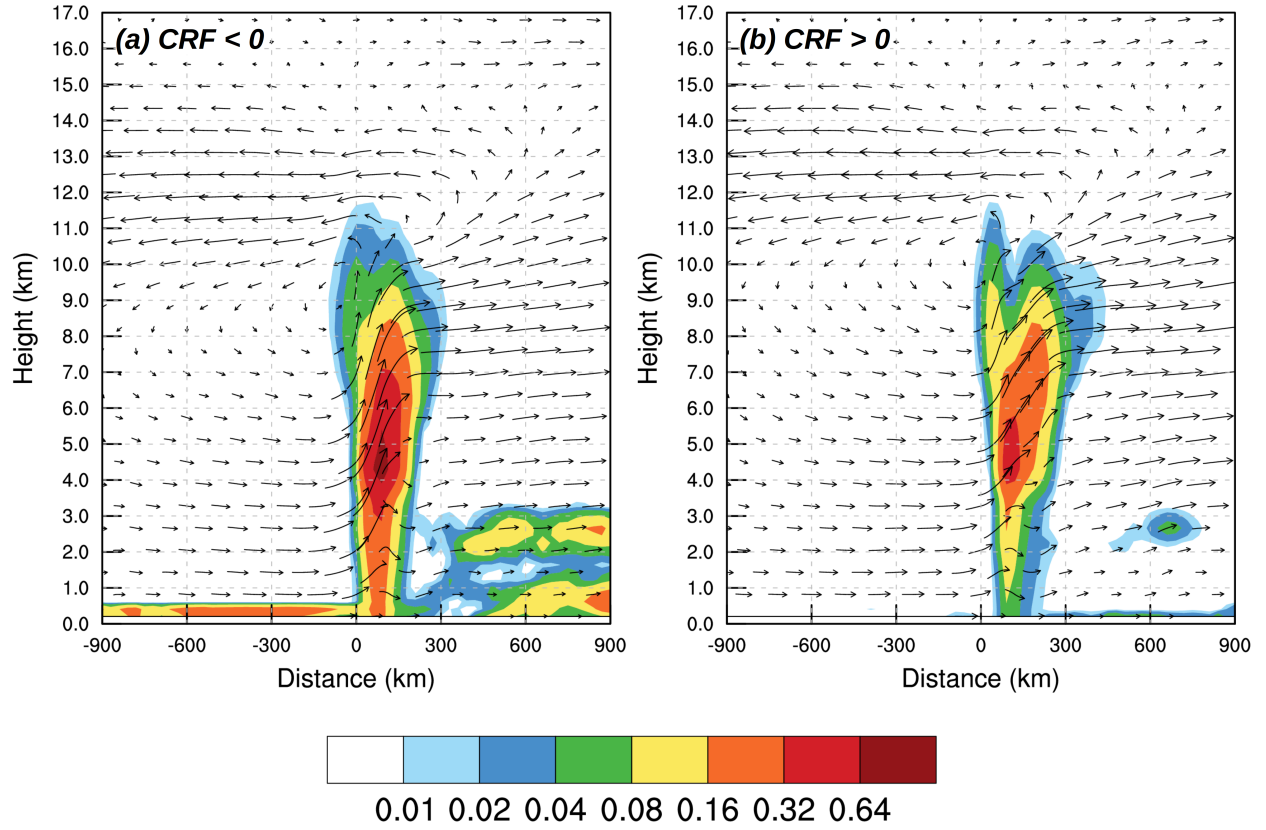


Figure 3.19: Condensate (g kg^{-1}) cross-sections as in Fig. 3.14 for (a) $CRF < 0$ and (b) $CRF > 0$ experiments.

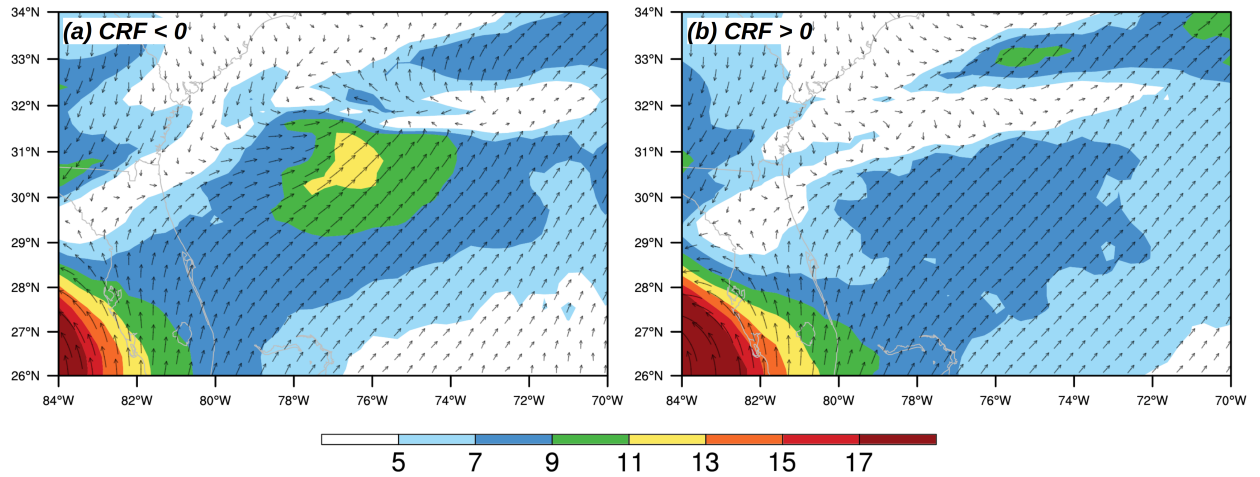


Figure 3.20: Surface-850 mb averaged wind speeds (m s^{-1}) as in Fig. 3.10 for (a) $CRF < 0$ and (b) $CRF > 0$ experiments.

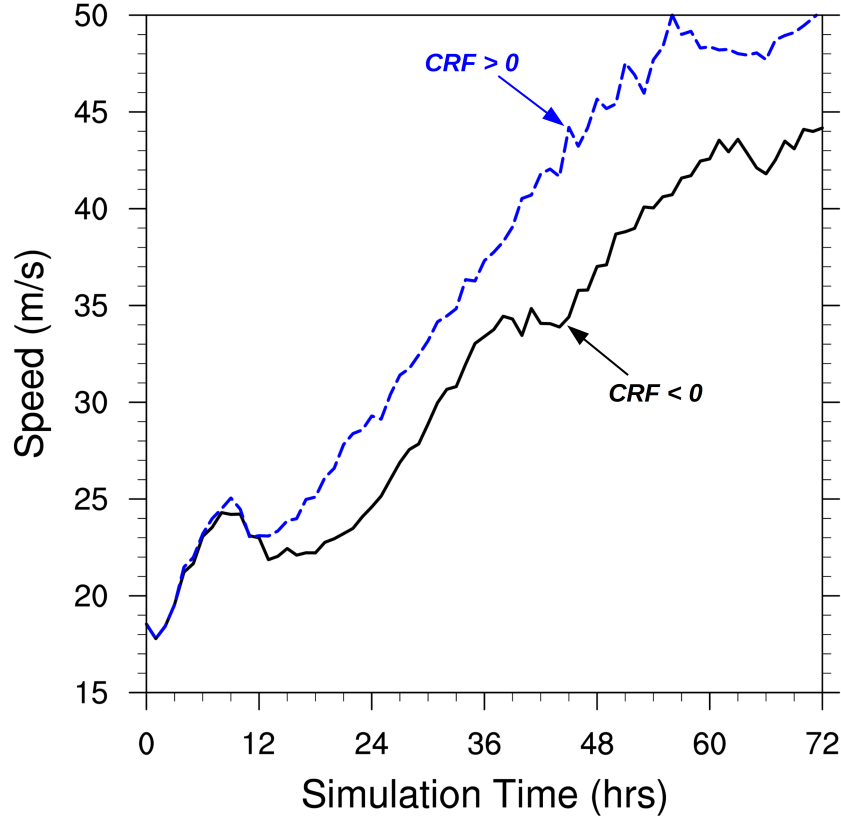


Figure 3.21: Maximum TC wind speed (m s^{-1}) over time as in Fig. 3.11 for $\text{CRF} < 0$ (solid black) and $\text{CRF} > 0$ (dashed blue) experiments.

produces a stronger PRE, it is associated with a significantly weaker TC at the time of PRE initiation (T+54), about 8 m s^{-1} less than the maximum TC winds in the $\text{CRF} > 0$ experiment (Fig. 3.21). Even though CRF warming effects are favorable for generating a broader TC structure and enhancing convective activity (Bu et al. 2014), it is actually the cooling component of CRF that is favorable for stronger PRE development. Therefore, it is suggested the influence of LW cooling along the low-level cloud tops outside of the PRE structure is primarily responsible for the initiation of PRE formation in the CRF-on and $\text{CRF} < 0$ experiments.

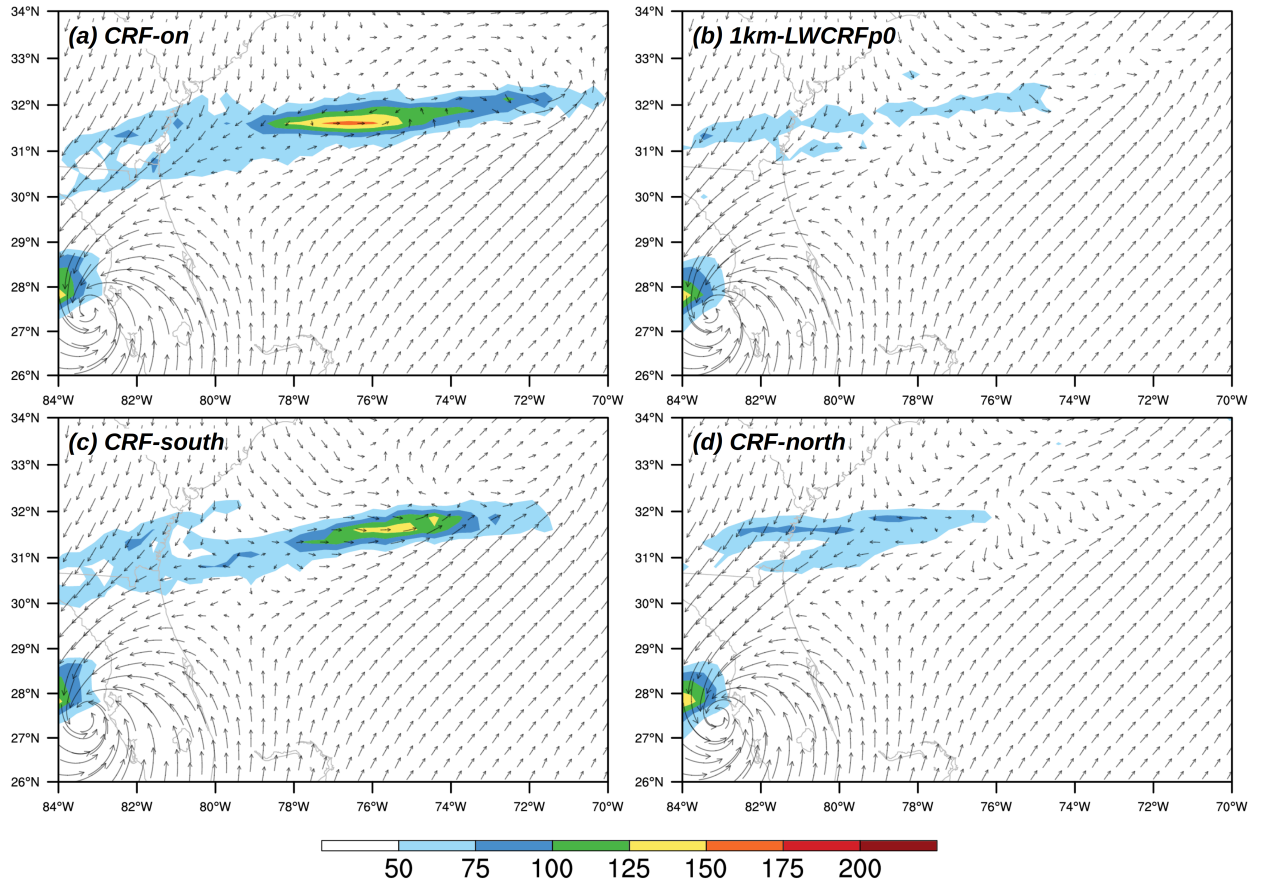


Figure 3.22: Total precipitation accumulation (mm) at T+72 hours as in Fig. 3.6 for (a) CRF-on, (b) 1km-LWCRFp0, (c) CRF-south, and (d) CRF-north experiments.

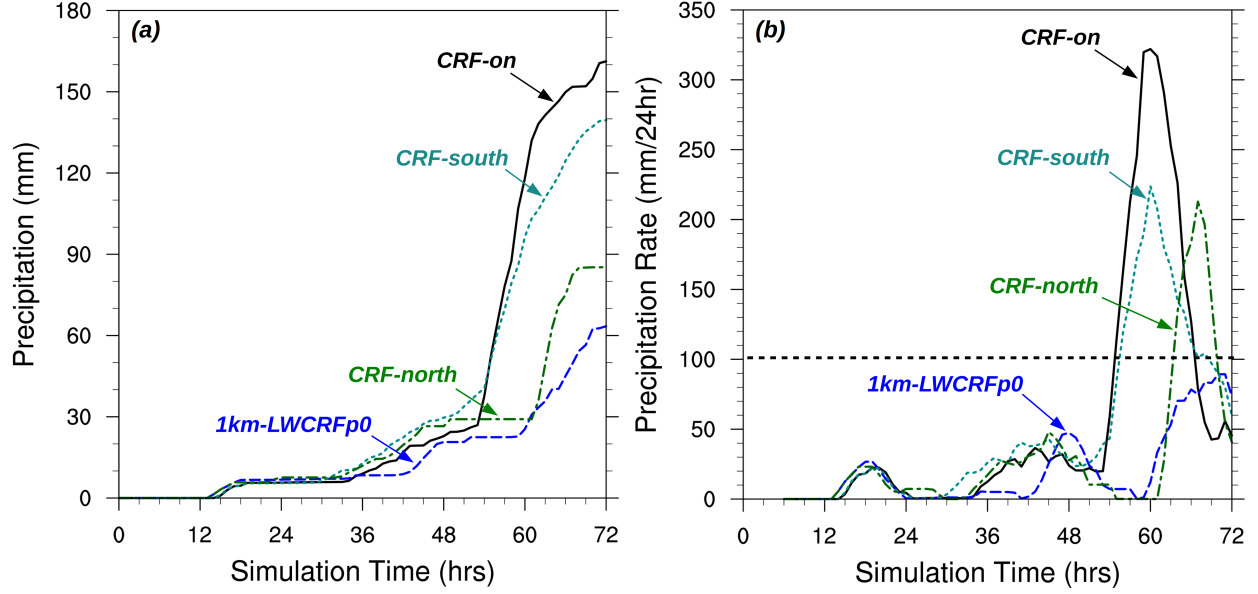


Figure 3.23: Line plots of (a) maximum PRE rainfall (mm) and (b) precipitation rate [$\text{mm} (24 \text{ hr})^{-1}$] over time as in Fig. 3.7 for CRF-on (solid black), 1km-LWCRFp0 (dashed blue), CRF-south (fine-dashed cyan), and CRF-north (dot-dashed green) experiments.

3.3.2 Role of low-level LW cooling

In order to isolate the influence of the low-level LW cooling component of CRF on PRE initiation, an experiment (labeled 1km-LWCRFp0) is conducted where LW cooling effects below 1 km mean sea level (MSL) are negated by modifying the LW radiation scheme. LW CRF warming effects (in-cloud) remain intact and, for less complexity, the SW component of CRF is unmodified.

As confirmation of the role of the LW cooling along the low-level cloud tops, PRE development is significantly weaker (Fig. 3.22b) and bears more resemblance to the rainfall distribution in the CRF-off experiment, even though full CRF effects are active above 1 km (Fig. 3.23a). The PRE rainfall rate in the 1km-LWCRFp0 experiment also never crosses the threshold for PRE classification (Fig. 3.23b).

In addition, minimal cloud development occurs to the north and south of the PRE region in the 1km-LWCRFp0 experiment (Fig. 3.24b) without the presence of strong LW cooling

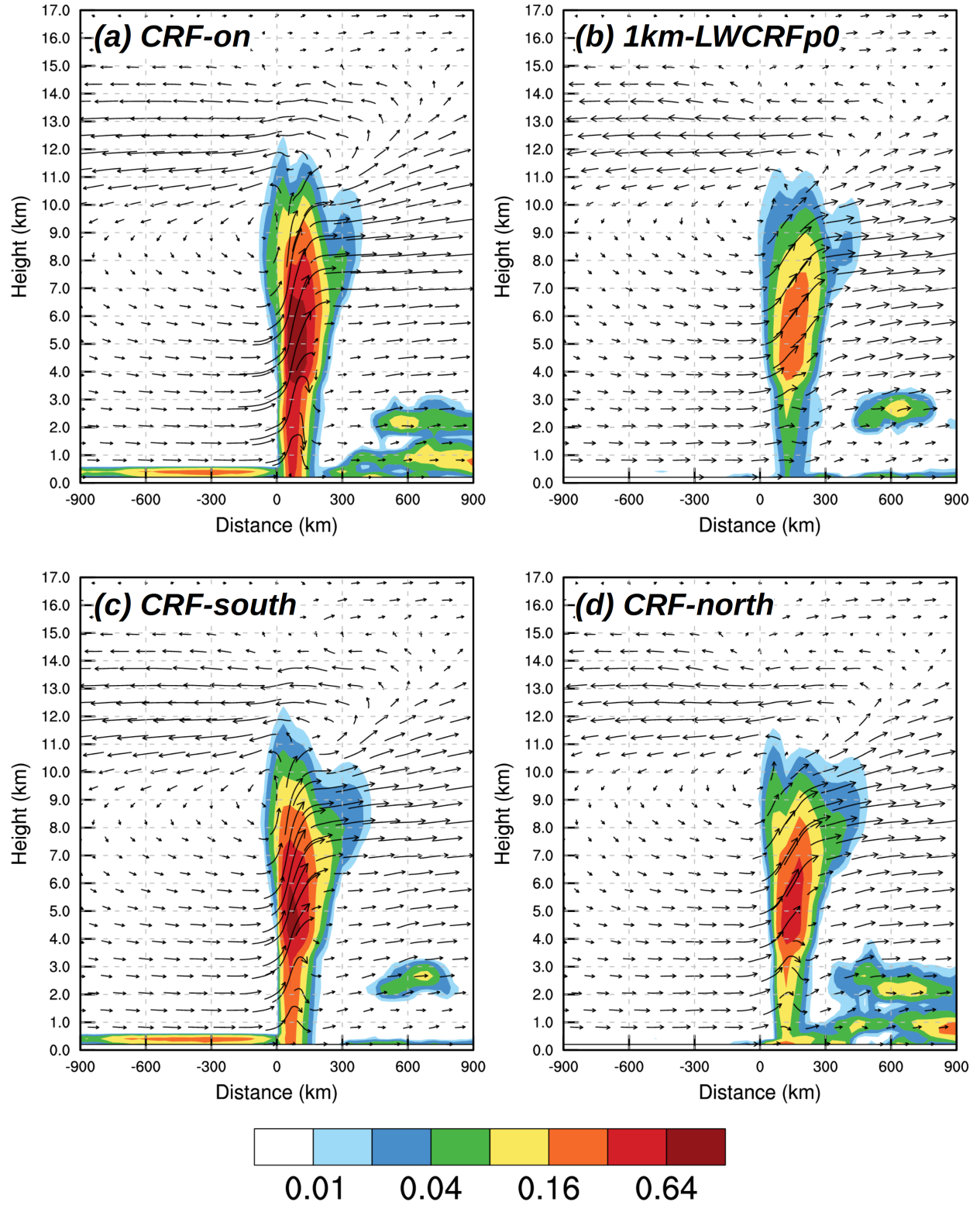


Figure 3.24: Condensate (g kg^{-1}) cross-sections as in Fig. 3.14 for (a) CRF-on, (b) 1km-LWCRFp0, (c) CRF-south, and (d) CRF-north experiments.

near the surface (Fig. 3.25b). Comparison of the CRF-on and 1km-LWCRFp0 experiments reveals that lower tropospheric cooling along the low-level cloud tops is indeed responsible for the enhanced SLP gradient (Fig. 3.26) and increased wind speeds (i.e. LLJ) across the PRE region (Fig. 3.27a,b). Therefore, LW cooling clearly plays a *direct* role in generating stronger convergence underneath the PRE, increased upward vertical motions, and more rainfall production.

Next, the influence of the latitudinal distribution of the lower tropospheric cooling is explored. Two experiments are conducted where LW cooling effects below 1 km mean sea level (MSL) are nullified north and south of the PRE structure, labeled CRF-north and CRF-south, respectively. The north-south separation is specified at the latitude of maximum rainfall in the CRF-on experiment (31.5°N) and extends to the east and west across all longitudes. All other LW and SW CRF effects are unmodified.

The total accumulated precipitation in the CRF-south experiment shows much stronger PRE development overall (Fig. 3.22c), producing about 55 mm more rainfall than the CRF-north experiment (Fig. 3.23a). The evolution of the PRE rainfall rate in the CRF-south experiment also more closely resembles the CRF-on experiment (Fig. 3.23b), especially with respect to the timing of PRE initiation at $T+54$ hours and termination at $T+66$ hours. In the CRF-north experiment, the rainfall rate is not only weaker, but PRE initiation is delayed over 8 hours and the duration of the PRE is almost halved.

LW cooling in the CRF-south experiment is also more concentrated equatorward of the PRE region (Fig. 3.25c), and has a greater influence on the modification of the low-level pressure gradient (Fig. 3.28a). The combination of southerly winds ahead of the approaching TC and a favorable pressure gradient south of the PRE results in an acceleration of the low-level wind flow, enhanced convergence (as well as frontogenesis) underneath the PRE (Fig. 3.27c), and stronger upward vertical motions throughout the PRE structure. In contrast, the enhancement of the pressure gradient poleward of the PRE region in the CRF-north experiment (Fig. 3.28b) mainly serves to increase convergence underneath the PRE without a significant acceleration of the winds across the PRE region (Fig. 3.27b). Therefore, the influence of LW cooling associated with the elevated clouds north of the PRE region is minimal.

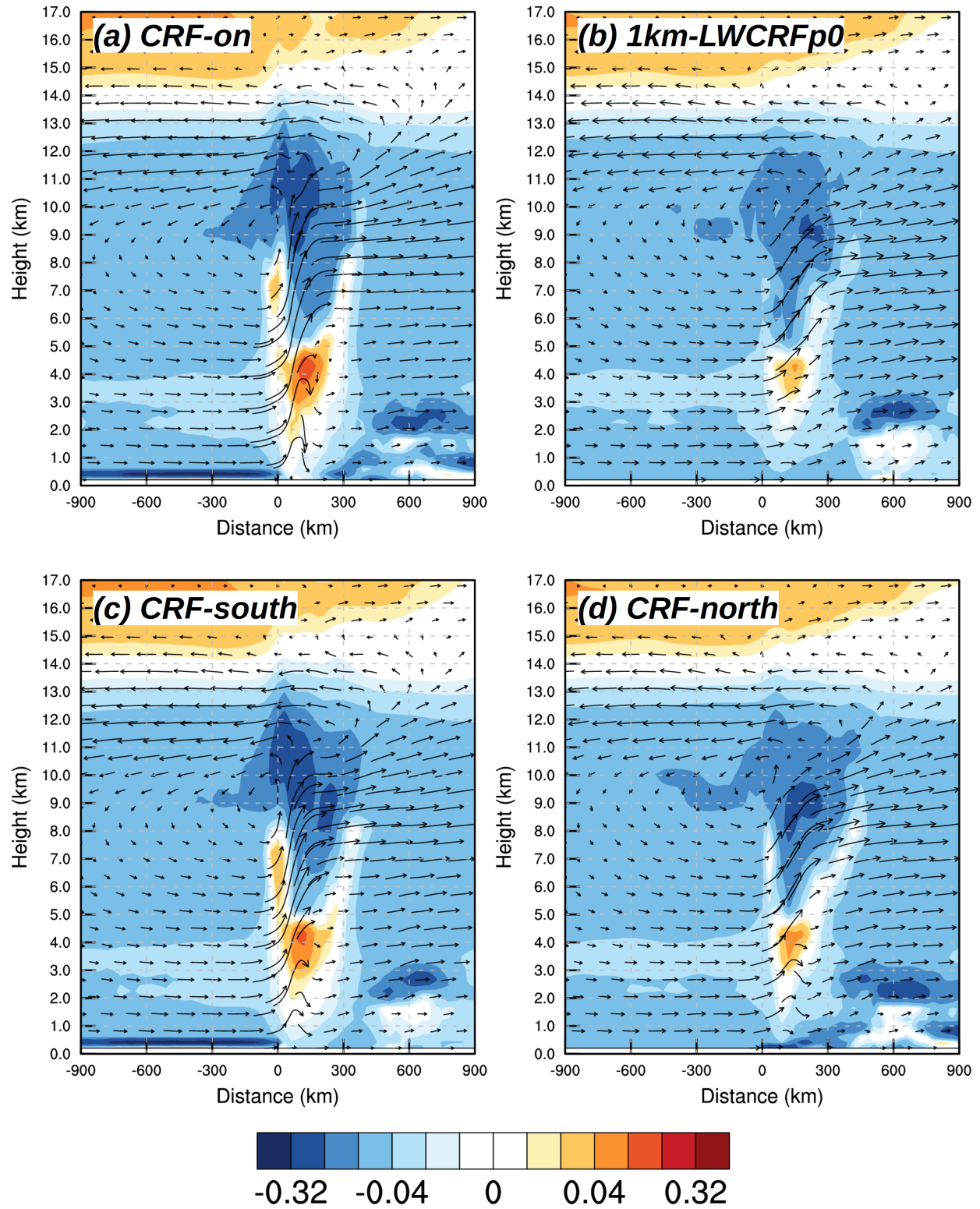


Figure 3.25: Radiative heating tendency (K hr⁻¹) cross-sections as in Fig. 3.13 for (a) CRF-on, (b) 1km-LWCRFp0, (c) CRF-south, and (d) CRF-north experiments.

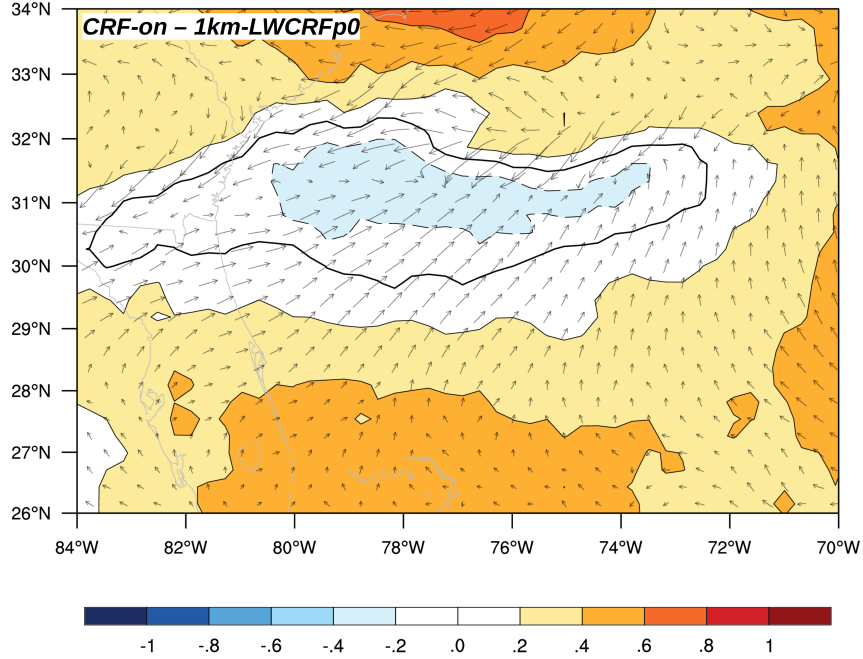


Figure 3.26: SLP difference (mb) averaged between T+54 and T+66 hours for CRF-on and 1km-LWCRFp0 experiments.

To confirm the role of the LW cooling distribution in the CRF-south experiment, a passive tracer is placed uniformly across the domain at the lowest model level at the start of the simulation (T+0). The presence of more lofted tracer material in the CRF-south experiment (Fig. 3.29a) indicates stronger upward vertical motions throughout the PRE structure, which is consistent with the production of more condensate (Fig. 3.24c), increased latent heat release (not shown), and stronger convection. In addition, the tracer distribution above 10 km is associated with the outflow layer of the upstream TC, demonstrating an upper level connection between components of the PRE system separated by a distance greater than 1200 km. Overall, CRF effects south of the PRE region play a more influential role in driving dynamical processes that eventually culminate in PRE initiation.

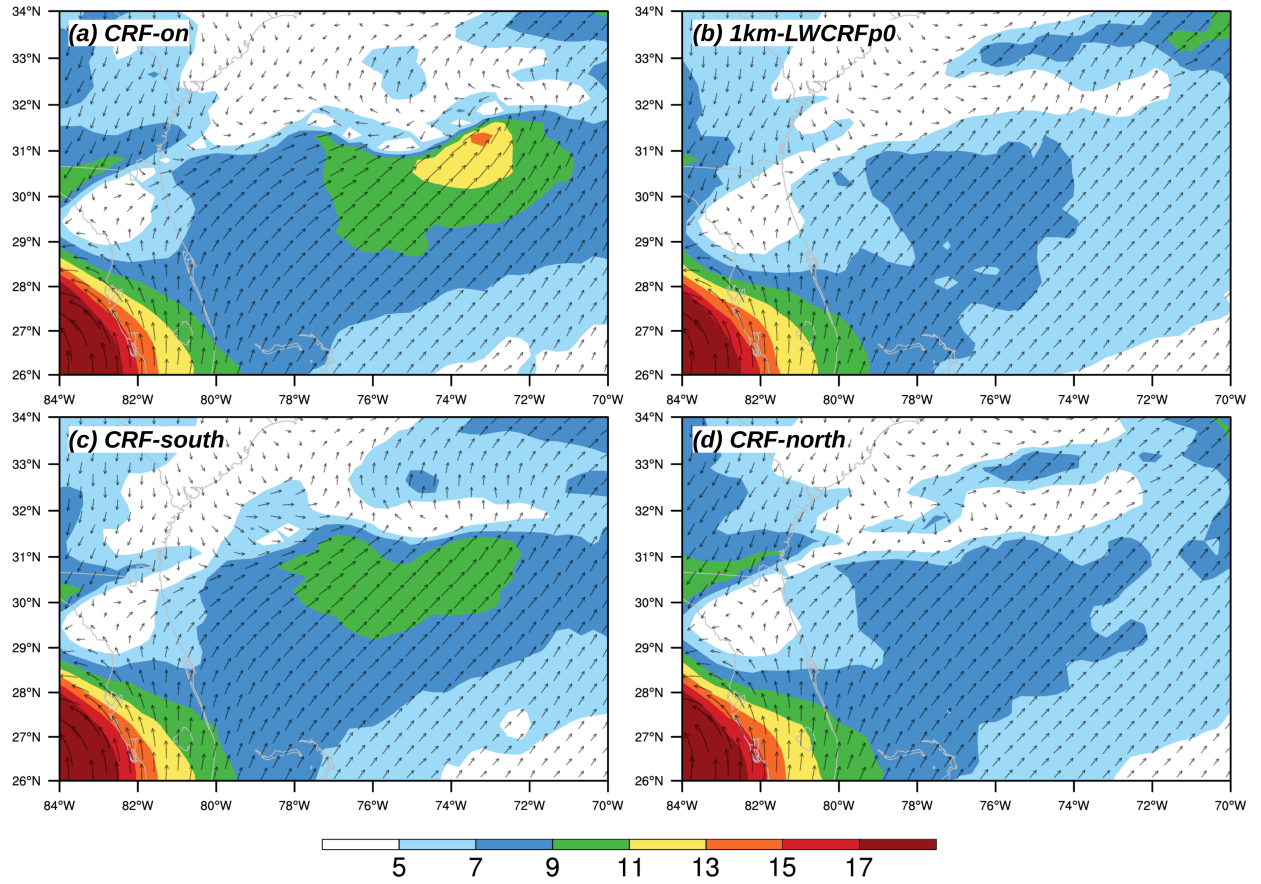


Figure 3.27: Surface-850 mb averaged wind speeds (m s^{-1}) as in Fig. 3.10 for (a) CRF-on, (b) 1km-LWCRFp0, (c) CRF-south, and (d) CRF-north experiments.

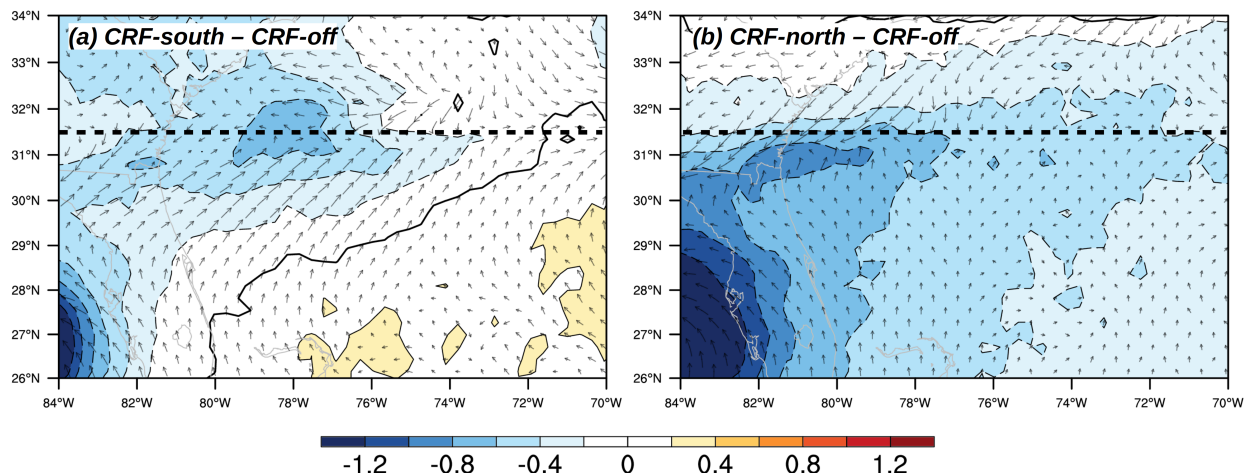


Figure 3.28: SLP difference (mb) as in Fig. 3.12 for (a) CRF-south - CRF-off and (b) CRF-north - CRF-off experiments. The dashed line (black) indicates the latitude of the north-south separation of CRF effects at 31.5°N.

3.4 Low-level cloud formation

The formation of low-level clouds in the CRF-on experiment can be attributed to a positive feedback between the radiation and MP schemes (Fig. 3.30). Once warm clouds form near the surface, LW cooling immediately occurs along the low-level cloud tops as well as LW warming within the clouds (Fig. 3.31a). Latent heat is also released during the formation of the low-level clouds (Fig. 3.31b). The combination of cloud top cooling, in-cloud warming, and MP diabatic heating are all absent in the CRF-off experiment. As a result, no low-level cloud development occurs.

The vertical distribution of the radiative and MP heating tendencies is inherently unstable as indicated by the magnitude of the lapse rates below 1.5 km MSL (Fig. 3.32). In the CRF-on experiment, warming near the surface and cooling above the cloud tops results in greater environmental lapse rates, compared to the CRF-off experiment. This leads to increased vertical instabilities and facilitates the formation of abundant low-level clouds. At a height of 0.5 km, lesser environmental lapse rates are produced by strong LW cooling along the low-level cloud tops and weaker clear sky LW cooling above (non-cloudy region). This vertical

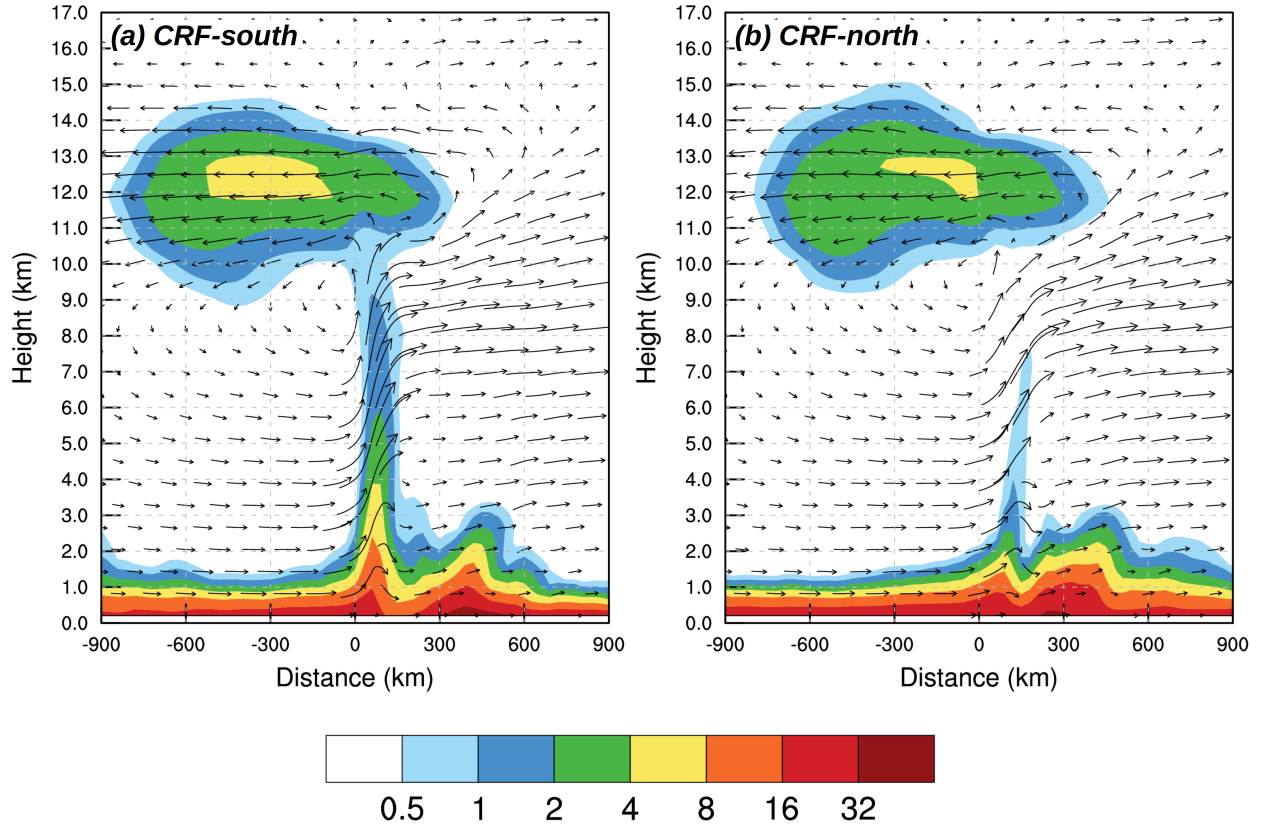


Figure 3.29: Passive tracer concentration (-) cross-sections averaged between T+54 and T+66 hours as in Fig. 3.13 for (a) CRF-south and (b) CRF-north experiments.

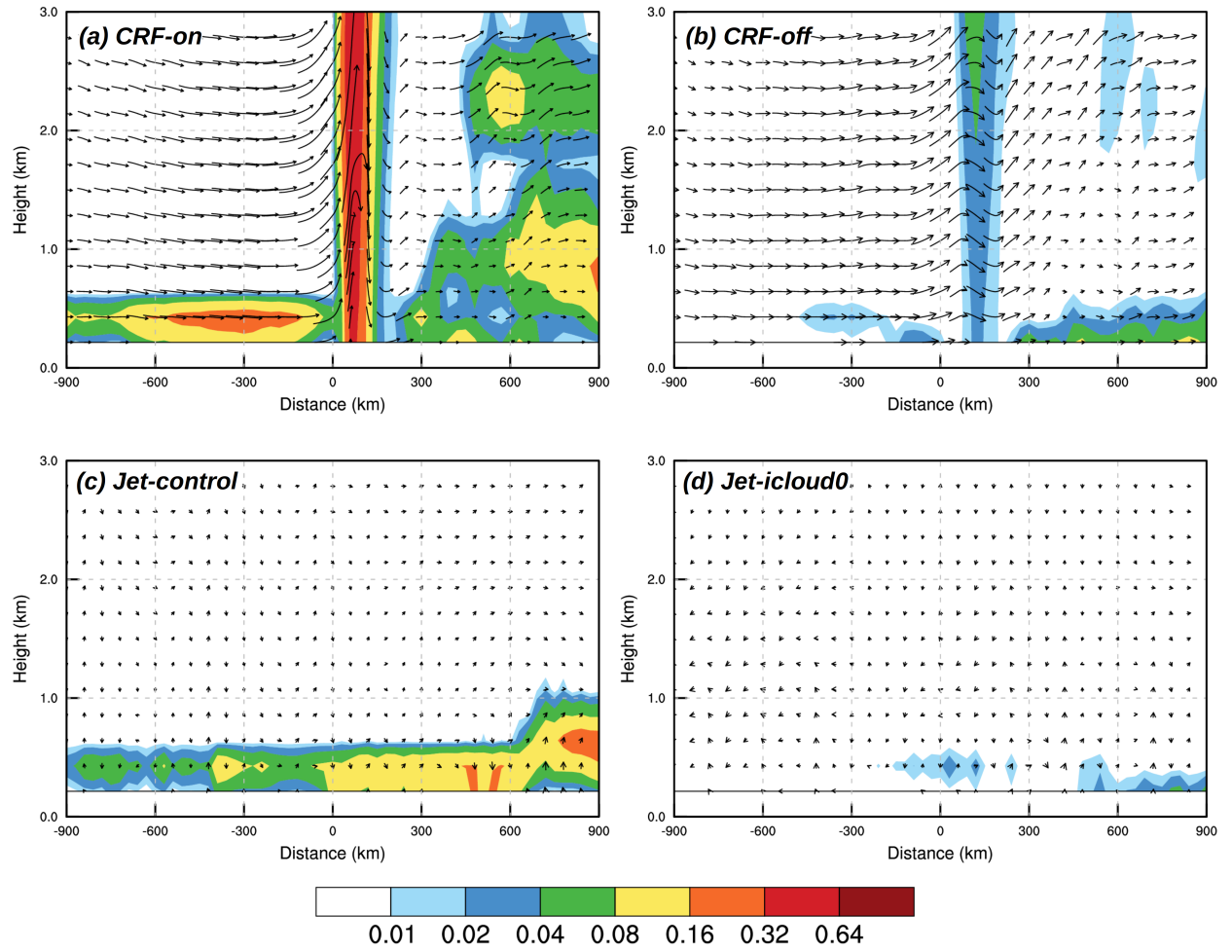


Figure 3.30: Condensate (g kg^{-1}) cross-sections as in Fig. 3.14 for (a) CRF-on, (b) CRF-off, (c) Jet-control, and (d) Jet-icloud0 experiments.

temperature distribution is associated with a more stable environment, effectively capping the height of the low-level clouds and limiting their vertical growth. If LW cooling below 1 km MSL is removed (e.g. 1km-LWCRFp0 experiment), the low-level clouds disappear, even with in-cloud warming and MP diabatic heating effects active (not shown). The lapse rates are also much weaker and more indicative of a stable environment less conducive to low-level cloud formation, similar to the CRF-off experiment (Fig. 3.32).

Another set of experiments is conducted where the TC is removed in both the CRF-on and CRF-off cases, labeled Jet-control and Jet-icloud0, respectively. Because the upper level jet and associated baroclinic zone are the only components of the PRE system in the domain, PRE development does not occur. Even without the presence of a TC, abundant low-level clouds still form near the surface in the Jet-control experiment (Figs. 3.30e,f). However, without increased moisture advection from the TC, there are less abundant clouds near the surface, especially south of the PRE region.

Without the presence of a TC in the Jet-control experiment, the distribution of cloud top cooling is much more uniform with respect to latitude (Fig. 3.31e), compared to the CRF-on experiment. As a result, there is no significant enhancement in the low-level horizontal pressure gradient that encourages strong convergence, frontogenesis, and upward vertical motions conducive to strong rainfall production (Fig. 3.33). Even though CRF is responsible for the formation of low-level clouds across the PRE region, it is the distribution of LW cooling along the cloud tops and the corresponding horizontal pressure gradient modification that are responsible for processes driving PRE initiation. However, it is important to note that the TC also plays a key role in influencing the development, concentration, and latitudinal distribution of the low-level clouds.

3.5 Radiation scheme variations

Lower tropospheric LW cooling is clearly important to PRE development and sensitive to different radiation parameterization schemes. Four radiation schemes in the WRF-ARW model are compared and contrasted (see Table 3.2): RRTMG (control CRF-on experiment),

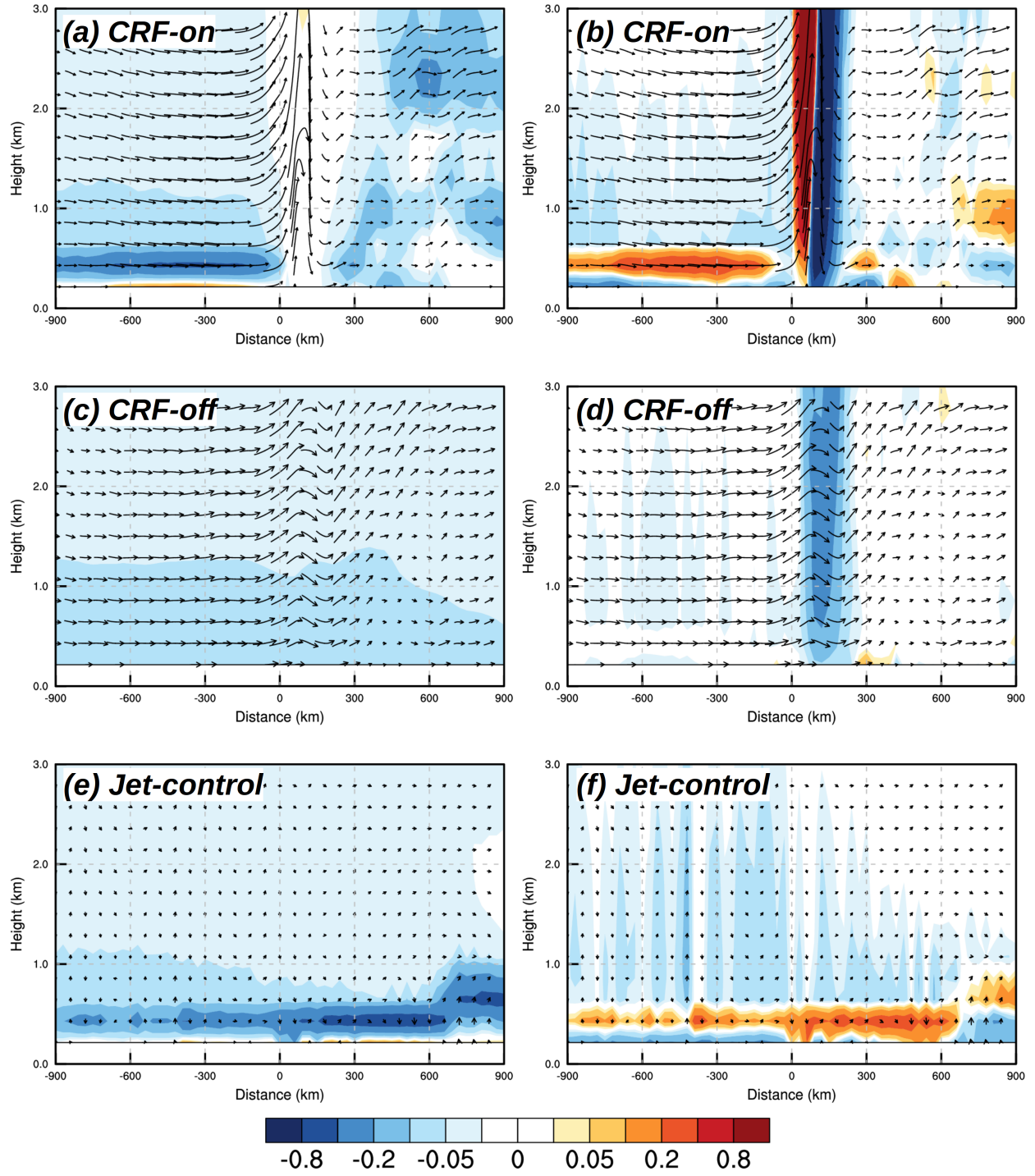


Figure 3.31: Radiative heating tendency (K hr⁻¹, first column) and MP heating tendency (K hr⁻¹, second column) cross-sections as in Figs. 3.13 and 3.15, respectively, for (a-b) CRF-on, (c-d) CRF-off, and (e-f) Jet-control experiments.

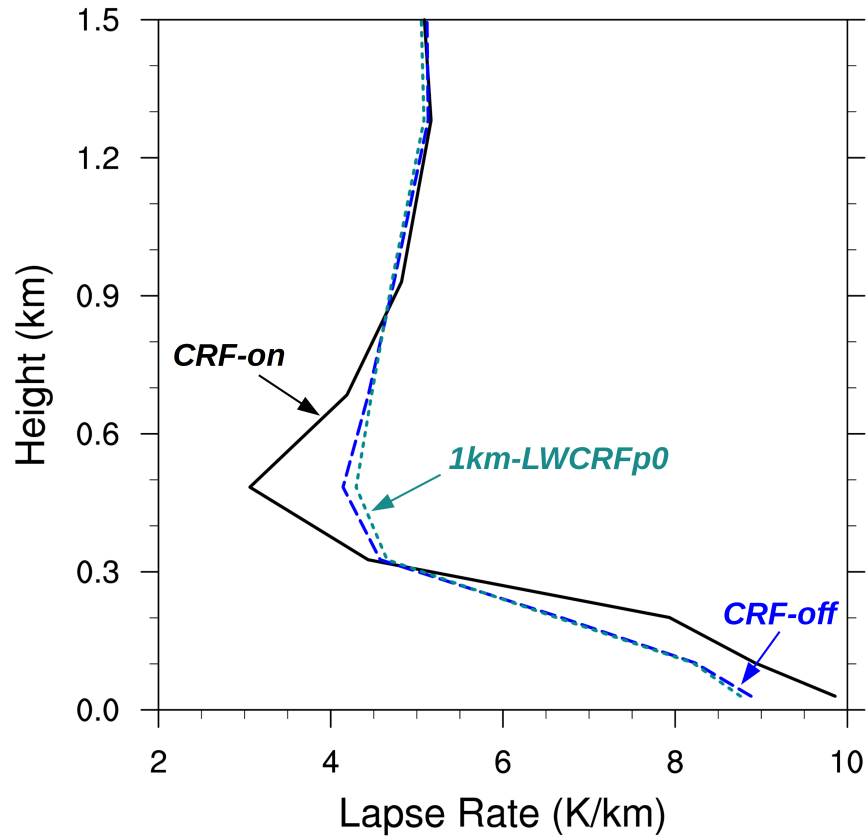


Figure 3.32: Lapse rate (K km^{-1}) averaged between T+54 and T+66 hours for CRF-on (solid black), CRF-off (dashed blue), and 1km-LWCRFp0 (fine-dashed cyan) experiments. Lapse rates are averaged across a 300 km latitude by 600 km longitude region centered on 30°N and 76°W .

Radiation Scheme	Reference
RRTMG (control)	Iacono et al. (2008)
CAM	Collins et al. (2004)
FLG	Gu et al. (2011), Fu and Liou (1992)
New Goddard	Chou and Suarez (1999)

Table 3.2: Radiation parameterization schemes

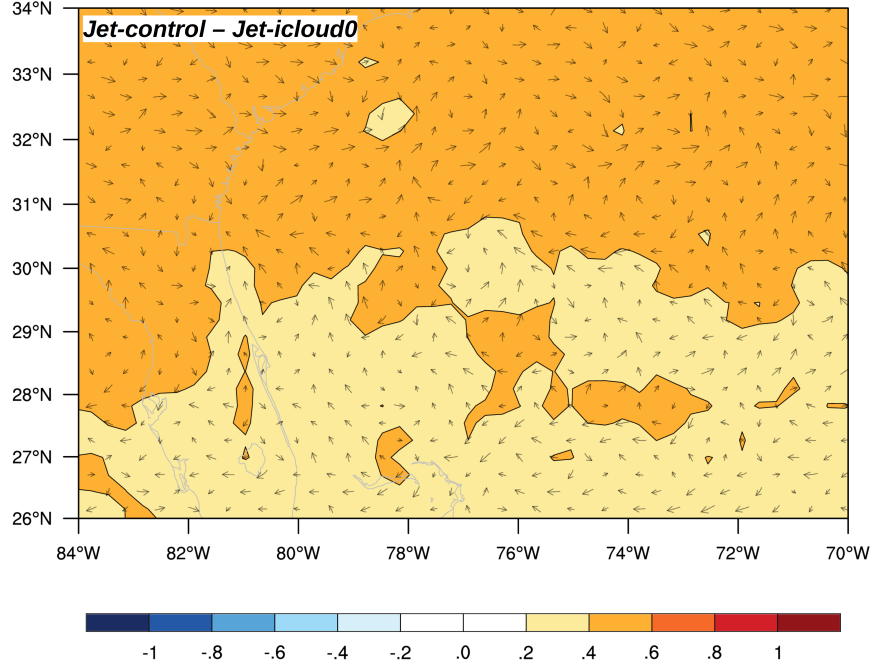


Figure 3.33: SLP difference (mb) and surface wind vector difference averaged between T+54 and T+66 hours for Jet-control and Jet-icloud0 experiments.

Community Atmosphere Model (CAM), Fu-Liou-Gu (FLG), and the National Aeronautics and Space Administration's (NASA) New Goddard. In general, there is a high sensitivity in the distribution of PRE rainfall among the experiments (Fig. 3.34), especially between the RRTMG and FLG radiation schemes. The precipitation intensity (Fig. 3.35a) and the timing of PRE development (Fig. 3.35b) vary considerably as well. Differences among the radiation schemes lead to changes in the SLP gradient, low-level convergence across the PRE region, and the resulting PRE rainfall distribution. Operationally, these large variations can have significant implications when forecasting the onset, duration, intensity, and even location of PRE development in an attempt to minimize the impact of heavy rainfall and severe flooding. Therefore, when predicting the formation of PREs, *radiation scheme differences are important!*

The schematic in Fig. 3.36 summarizes the influences of cloud-radiative processes on the generation of heavy PRE rainfall and how they interact between scales and parameterizations. Interestingly, the primary driver of PRE development actually resides outside of the

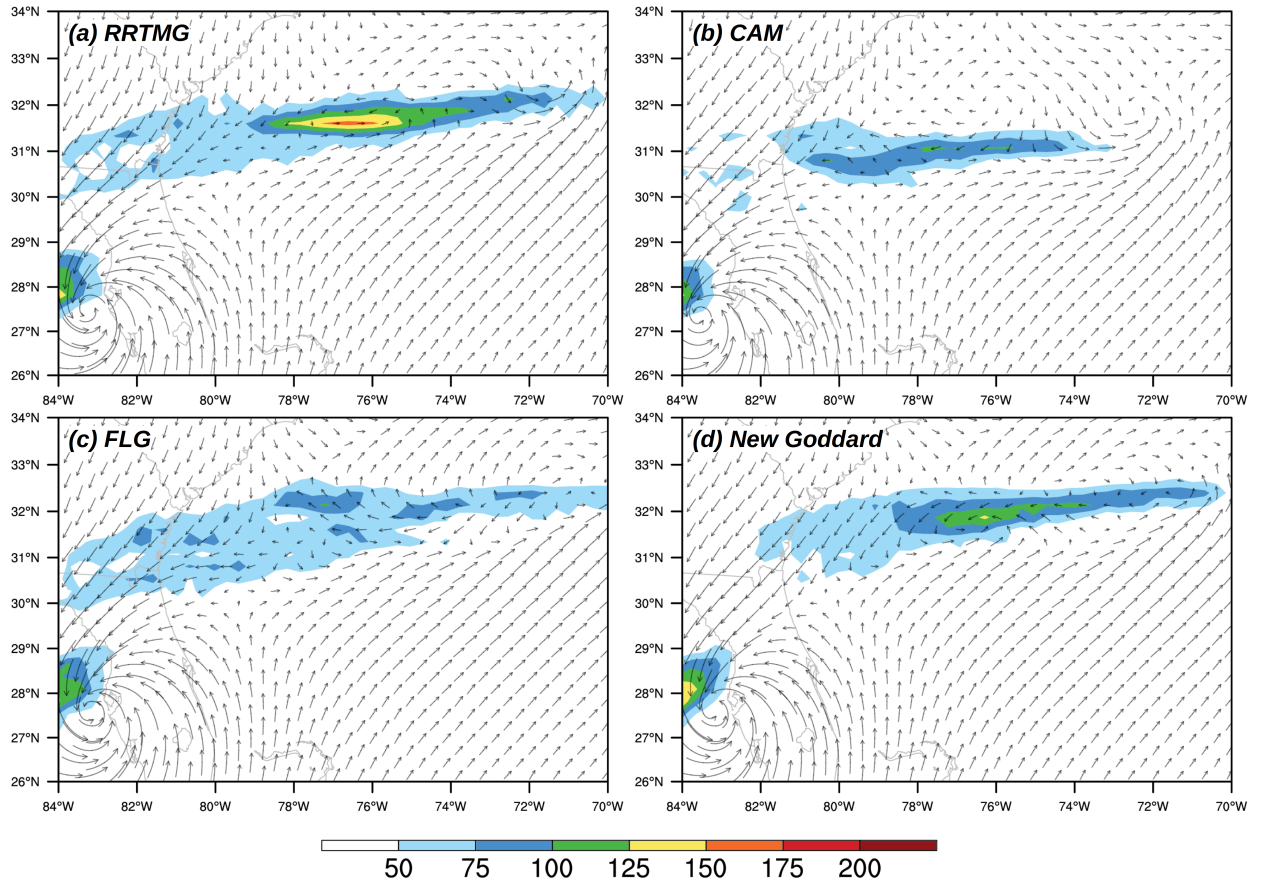


Figure 3.34: Total precipitation accumulation (mm) at T+72 hours as in Fig. 3.6, but for various radiation schemes readily available in the WRF-ARW v.3.6 model listed in Table 3.2.

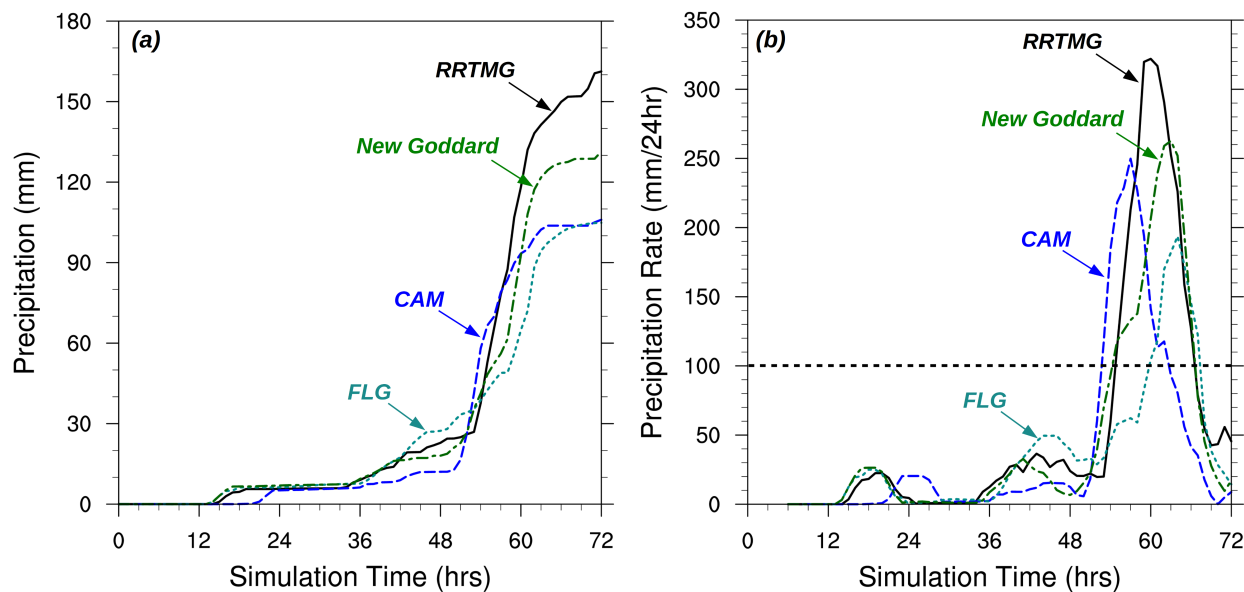


Figure 3.35: Line plots of (a) maximum PRE rainfall (mm) and (b) precipitation rate [mm (24 hr)^{-1}] over time as in Fig. 3.7 for RRTMG (solid black), CAM (dashed blue), FLG (fine-dashed cyan), and New Goddard (dot-dashed green) radiation scheme experiments.

PRE region itself and is based on how the model responds radiatively to low clouds. For further complication, CRF can also have a significant influence on TC structure and motion (cf. Bu et al. 2014). Overall, the high sensitivity of PRE development to different radiation schemes drives a need for observations to verify that the model calculations of the CRF field between the MP and radiation schemes are indeed accurate.

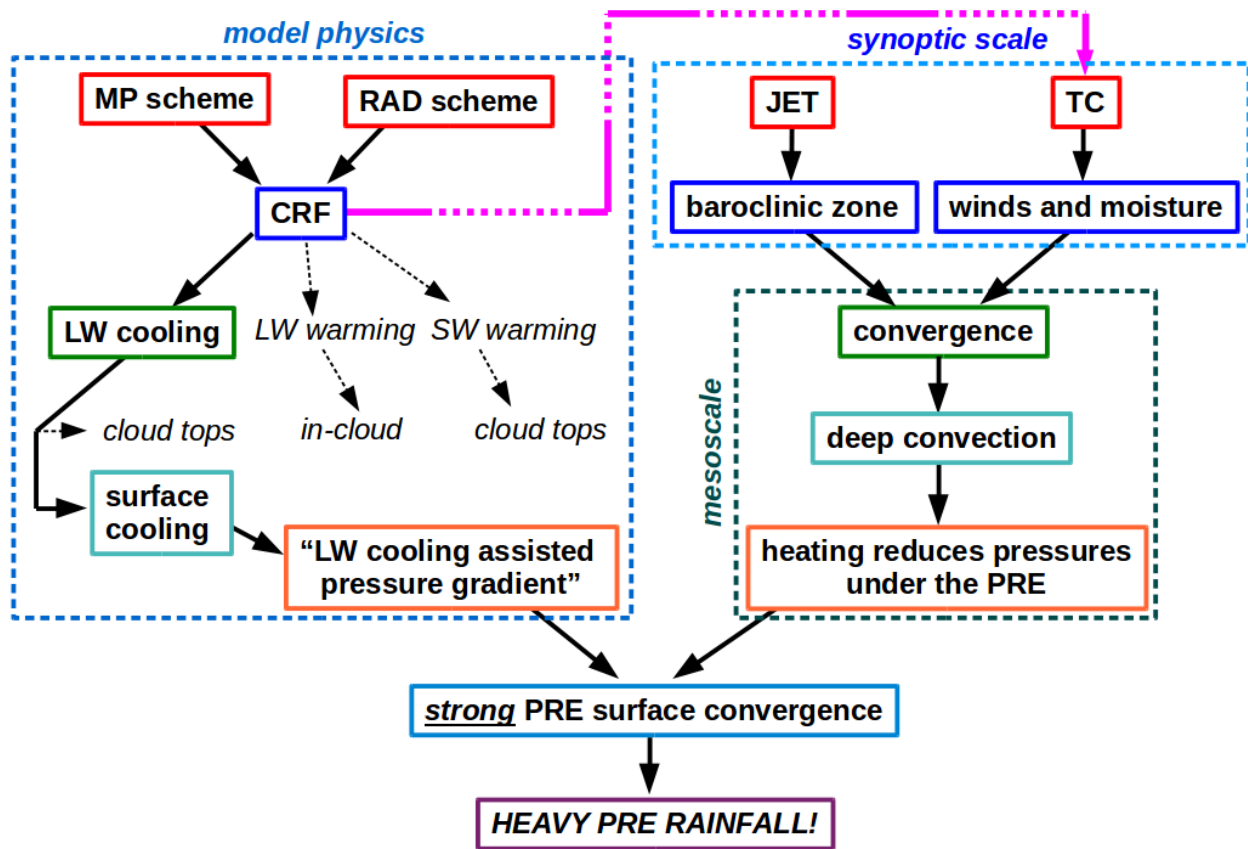


Figure 3.36: Schematic of the CRF influence on the initiation of PRE development.

CHAPTER 4

Influence of CRF on PRE Intensification

The first stage of PRE development is characterized by low-level LW cooling processes outside of the PRE region which play an important role in PRE formation. The second stage is distinguished by a rapid increase in the maximum precipitation and rainfall rate after PRE initiation has occurred. This chapter demonstrates that CRF effects internal to the PRE structure are primarily responsible for the increased rainfall production during the PRE intensification stage. The first section investigates the role of LW CRF effects and the contribution of different cloud species within the PRE structure. Next, the influence of SW CRF effects, as well as the sensitivity of PRE development to the timing of the diurnal cycle and seasonal length of day, are explored. The experiments listed in Fig. 4.1 are described in more detail in Table 4.1 and discussed in later sections.

4.1 PRE intensification process

An experiment, inspired by Fovell et al. (2010a), is conducted where CRF effects only associated with QCLOUD are allowed (labeled QC-only). Because QICE and QSNOW in the middle and upper levels of the PRE structure are effectively transparent to radiation, only 56% of the precipitation maximum in the CRF-on experiment is generated by the QC-only experiment (Fig. 4.2a). In addition, the rainfall rate in the QC-only experiment closely follows the CRF-on experiment until PRE initiation at T+54 hours when it decreases significantly during the intensification stage through T+60 hours (Fig. 4.2b). This suggests that CRF effects associated with QICE and QSNOW must be responsible for the rapid intensification of rainfall in the PRE.

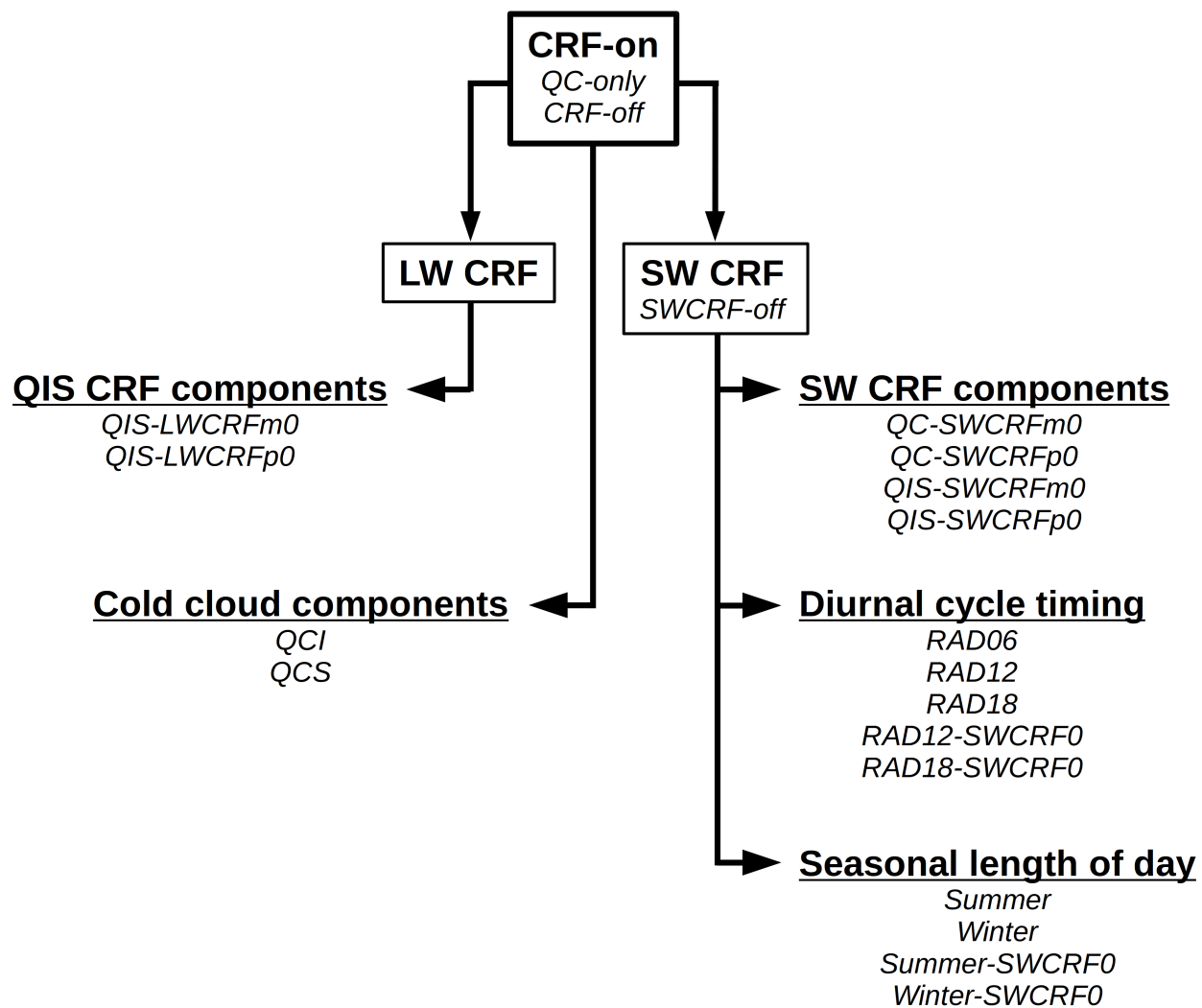


Figure 4.1: A schematic of a series of experiments in Chapter 4 focused on investigating the influence of CRF effects on the intensification stage of PRE development. Experiment names in italics are described in more detail in Table 4.1.

Experiment label	Description
CRF-on	CRF effects active
CRF-off	No CRF effects, only clear sky LW and SW radiation
QC-only	Only CRF effects due to QCLOUD
QIS-LWCRFm0	Only LW CRF cooling due to QICE and QSNOW
QIS-LWCRFp0	Only LW CRF warming due to QICE and QSNOW
QCI	Only CRF effects due to QCLOUD and QICE
QCS	Only CRF effects due to QCLOUD and QSNOW
SWCRF-off	No SW CRF effects, only LW CRF effects
QC-SWCRFm0	Only SW CRF cooling due to QCLOUD
QC-SWCRFp0	Only SW CRF warming due to QCLOUD
QIS-SWCRFm0	Only SW CRF cooling due to QICE and QSNOW
QIS-SWCRFp0	Only SW CRF warming due to QICE and QSNOW
RAD06	Diurnal cycle timing + 6 hours
RAD12	Diurnal cycle timing + 12 hours
RAD18	Diurnal cycle timing + 18 hours
RAD12-SWCRF0	Diurnal cycle timing + 12 hours with no SW CRF
RAD18-SWCRF0	Diurnal cycle timing + 18 hours with no SW CRF
Winter	December 20 length of day
Summer	June 20 length of day
Winter-SWCRF0	December 20 length of day with no SW CRF
Summer-SWCRF0	June 20 length of day with no SW CRF

Table 4.1: List of PRE intensification experiments

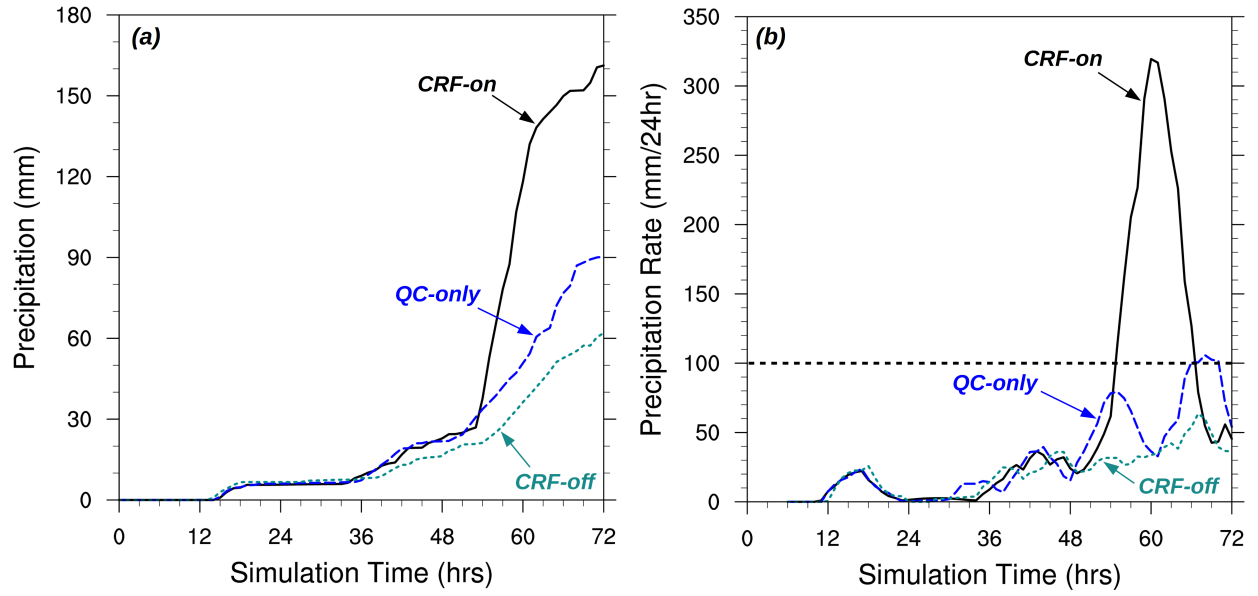


Figure 4.2: Line plots of (a) maximum PRE rainfall (mm) and (b) precipitation rate [$\text{mm} (24 \text{ hr})^{-1}$] over time as in Fig. 3.7 for CRF-on (solid black), QC-only (dashed blue), and CRF-off (fine-dashed cyan) experiments.

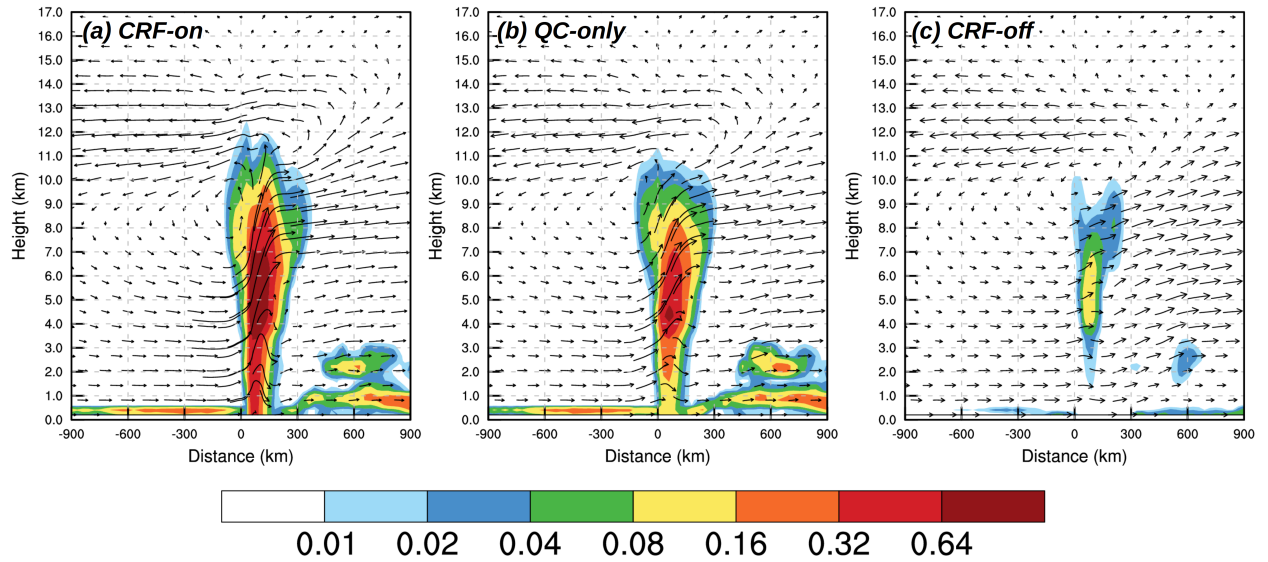


Figure 4.3: Condensate (g kg^{-1}) cross-sections as in Fig. 3.14 averaged between T+54 and T+60 hours for (a) CRF-on, (b) QC-only, and (c) CRF-off experiments.

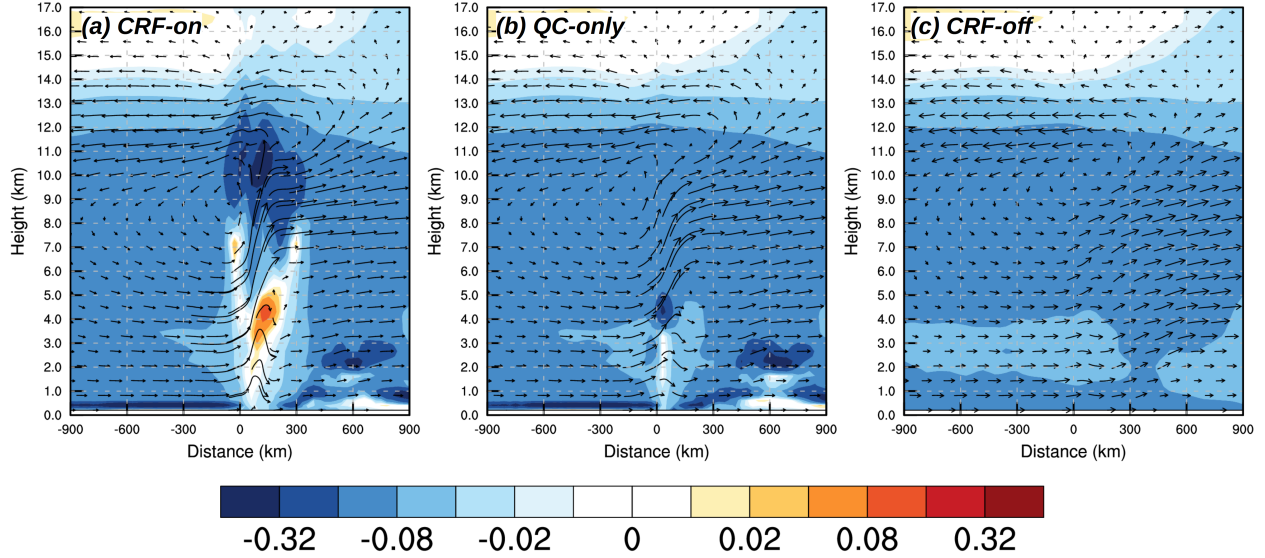


Figure 4.4: Radiative heating tendency (K hr^{-1}) cross-sections as in Fig. 3.13 averaged between T+54 and T+60 hours for (a) CRF-on, (b) QC-only, and (c) CRF-off experiments.

Cross-sections are constructed and averaged between T+54 and T+60 hours to examine key differences between the experiments during the PRE intensification stage. The weaker upward vertical motions in the QC-only experiment produce much less condensate throughout the PRE structure, compared to the CRF-on experiment (Fig. 4.3). Consequently, the combination of reduced CRF warming in the middle-levels of the PRE (Fig. 4.4) and less latent heat release throughout the PRE structure (Fig. 4.5) both contribute to a significant decrease in the production of rainfall in the PRE.

As in the PRE initiation stage, strong horizontal pressure gradients are also important during the PRE intensification stage. Among the experiments, a broad region of higher pressures is located both north and south of the PRE region with lower pressures across the PRE structure (Fig. 4.6a). The higher pressures are associated with clear sky LW cooling outside of the cloudy regions of the PRE. The lower pressures are generated by a combination of latent heat release and CRF warming within the PRE structure. However, in the CRF-on experiment, there is a local region of higher pressures underneath the PRE that can be attributed to evaporative cooling associated with the MP scheme (Fig. 4.5a). The rapid transition from lower to higher pressures creates an intense horizontal pressure gradient that

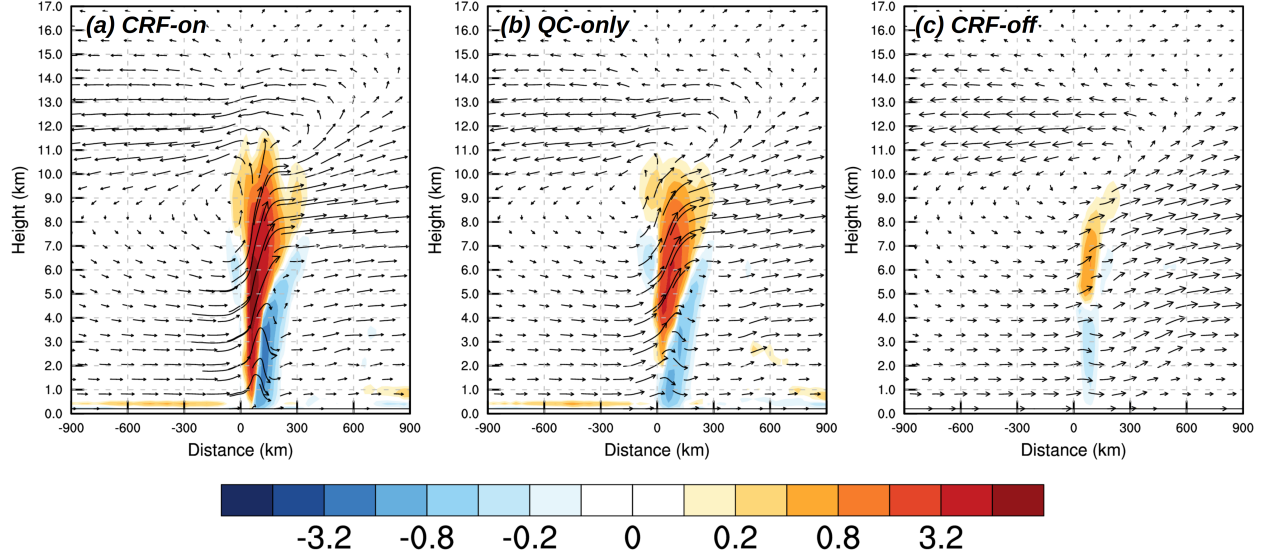


Figure 4.5: MP heating tendency (K hr^{-1}) cross-sections as in Fig. 3.15 averaged between T+54 and T+60 hours for (a) CRF-on, (b) QC-only, and (c) CRF-off experiments.

serves as a focal point for strong convergence on the equatorward edge of the PRE structure, extending from the surface to about 7 km. This suggests that evaporative cooling, which is typically unfavorable for strong convective processes, may play an important role in PRE intensification. In comparison, there is a weaker horizontal pressure gradient in the QC-only experiment underneath the PRE (Fig. 4.6b) associated with weaker upward vertical motions and decreased MP diabatic heating and cooling (Fig. 4.5b). In the CRF-off experiment, there is no obvious indication of an enhanced horizontal pressure gradient or a strong increase in upward vertical motions (Fig. 4.6c).

The intensity of the horizontal pressure gradient influences the meridional wind speeds and magnitude of convergence on the equatorward side of the PRE structure, driving strong upward vertical motions conducive to heavy rainfall production. In the CRF-on experiment, a region of strong convergence extends from the surface to the mid-levels of the PRE structure (Fig. 4.7a), compared to the QC-only experiment which has a similar vertical extent, but weaker intensity (Fig. 4.7b). In the CRF-off experiment, the wind enhancement and mid-level convergence is almost non-existent (Fig. 4.7c).

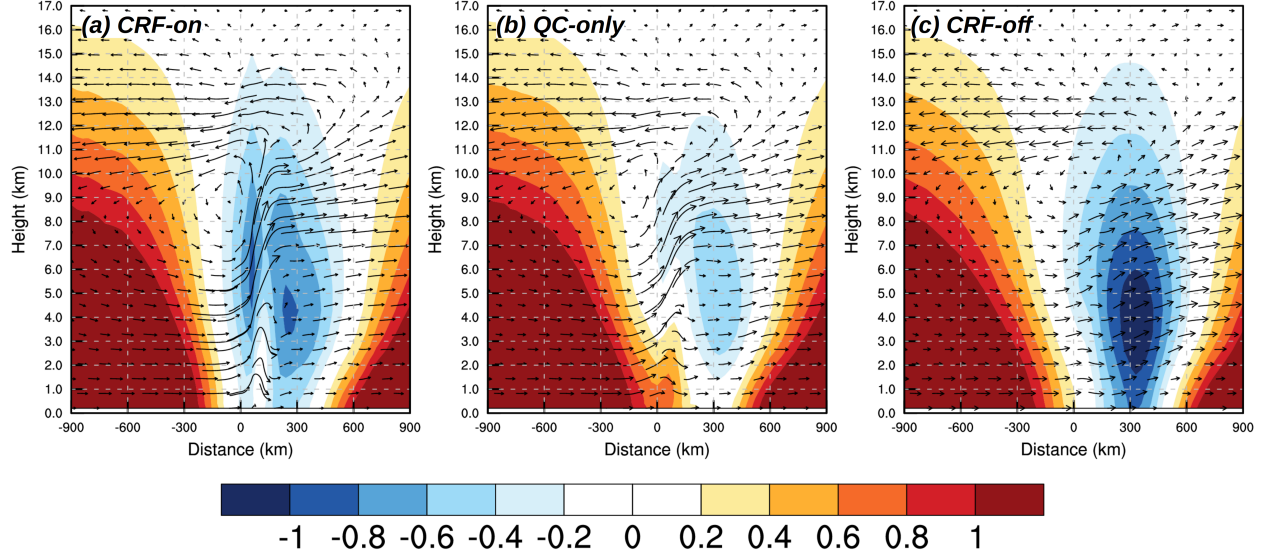


Figure 4.6: As in Fig. 4.5, but for cross-sections of pressure perturbation (mb) from T+0 and wind vector differences averaged between T+54 and T+60 hours for (a) CRF-on, (b) QC-only, and (c) CRF-off experiments.

The magnitude of the horizontal pressure gradient (and wind response) is largely influenced by the distribution of both MP heating and cooling, which is in-turn influenced by the distribution of CRF heating and cooling. Even though the interaction between LW radiation and low-level clouds plays a critical role in the initiation of PRE development, PRE intensification is driven by CRF effects associated with QICE and QSNOW in the middle and upper levels of the PRE structure. It is clear that strong upward vertical motions generated by various interconnected physical processes are the key to significant rainfall production during the PRE intensification stage.

4.2 Contribution of longwave CRF effects

The heating and cooling components of CRF are isolated to determine their influence on PRE rainfall production. For simplification, only the LW radiation component of CRF is investigated. The contribution from the SW CRF component will be explored in the following section. The experiment labeled QIS-LWCRFm0 incorporates only the cooling

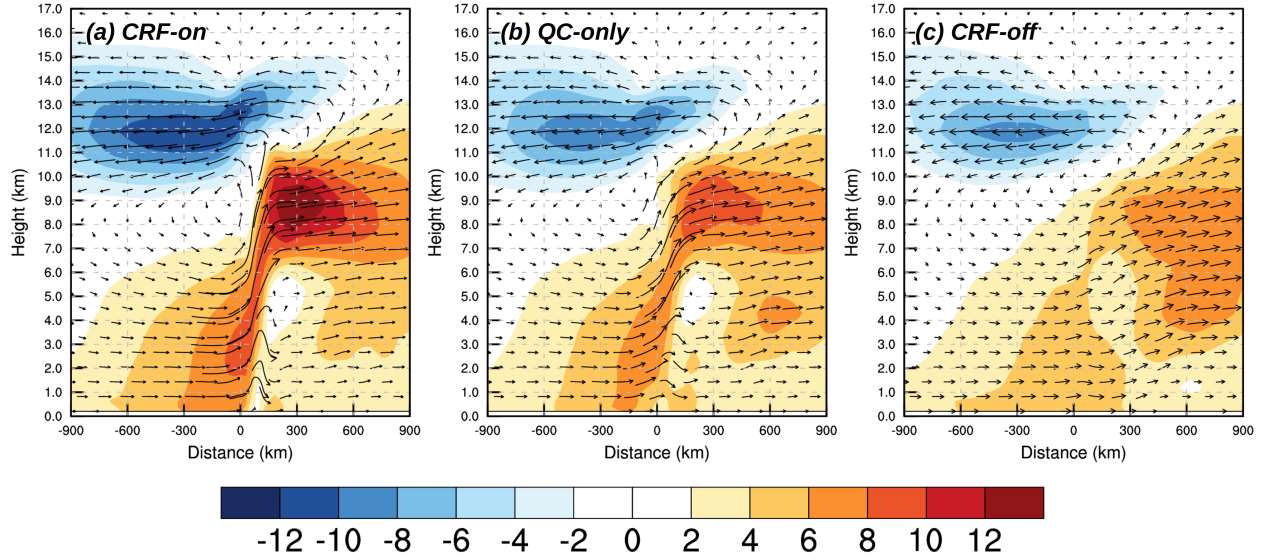


Figure 4.7: As in Fig. 4.6, but for cross-sections of meridional wind speed (m s^{-1}) between T+54 and T+60 hours for (a) CRF-on, (b) QC-only, and (c) CRF-off experiments.

component of LW CRF acting on the QICE and QSNOW cloud species (QIS); the experiment labeled QIS-LWCRFp0 only includes the warming component of LW CRF acting on the same hydrometeor species. In both experiments, the LW radiation interaction with QCLOUD near the surface is unmodified to preserve the low-level forcing processes responsible for initiating PRE development.

At the end of the 72-hour simulation, the LW CRF warming component (QIS-LWCRFp0) produces a more intense swath of heavy rainfall across the PRE region (Fig. 4.8b), as well as higher values of maximum precipitation (Fig. 4.9a), compared to the cooling component of LW CRF (QIS-LWCRFm0). The timing of PRE intensification and termination in the QIS-LWCRFp0 experiment is nearly identical to the CRF-on experiment, suggesting that the warming component of LW CRF is primarily responsible for the processes driving PRE intensification (Fig. 4.9b). In contrast, the cooling component of LW CRF not only decreases rainfall production in the PRE, but also delays development by about 10 hours.

LW CRF cooling in the QIS-LWCRFm0 experiment mostly occurs in the upper half of the PRE structure (Fig. 4.10a), while LW CRF warming in the QIS-LWCRFp0 experiment

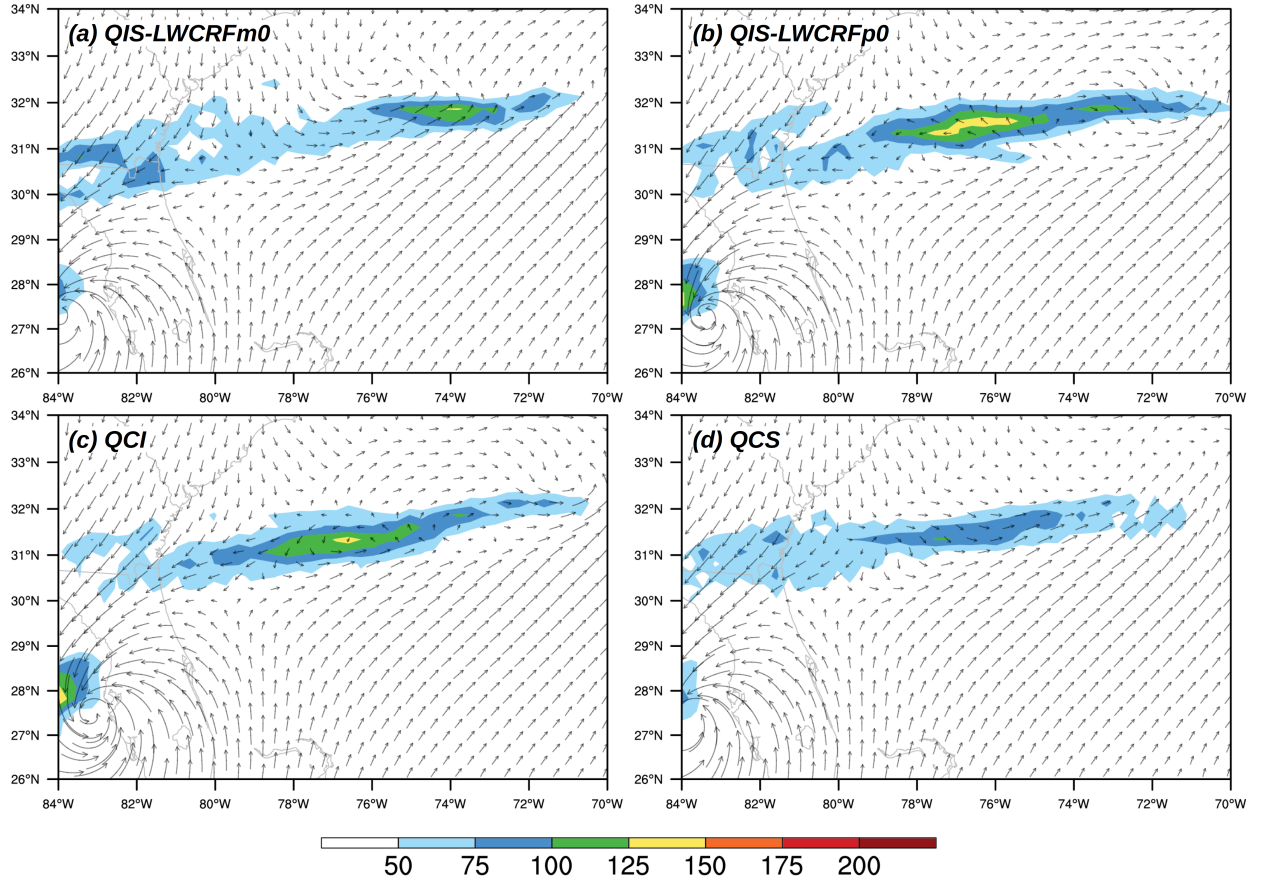


Figure 4.8: Total precipitation accumulation (mm) at T+72 hours as in Fig. 3.6 for (a) QIS-LWCRFm0, (b) QIS-LWCRFp0, (c) QCI, and (d) QCS experiments.

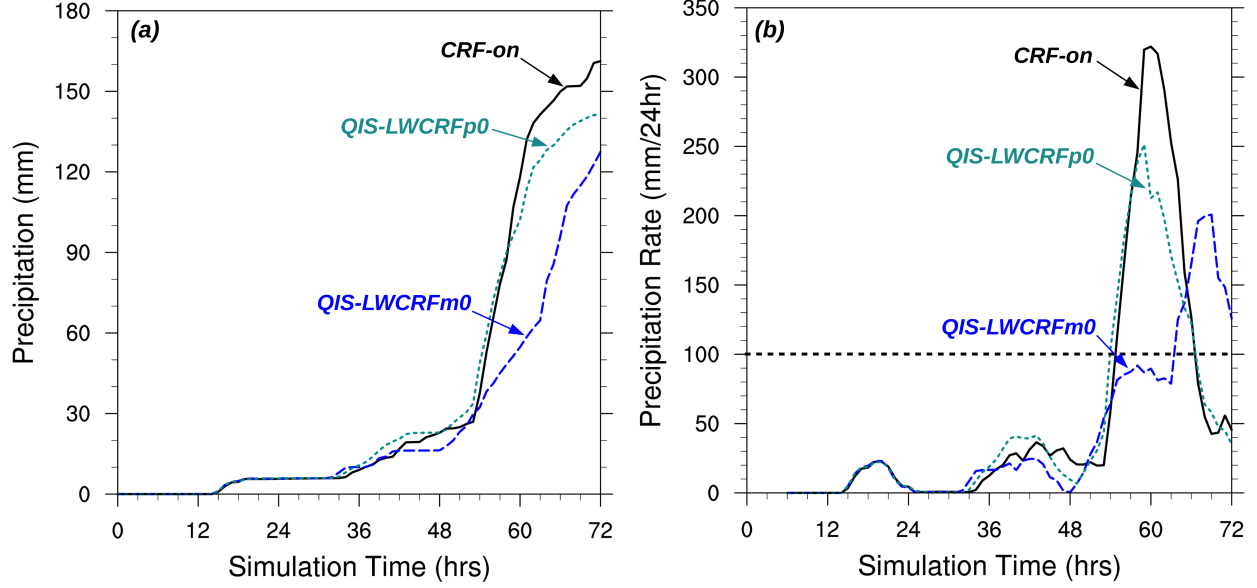


Figure 4.9: Line plots of (a) maximum PRE rainfall (mm) and (b) precipitation rate [mm (24 hr)^{-1}] over time as in Fig. 3.7 for CRF-on (solid black), QIS-LWCRFm0 (dashed blue), and QIS-LWCRFp0 (fine-dashed cyan) experiments.

is mainly concentrated in the lower half of the storm (Fig. 4.10b). Warming in the lower troposphere increases the buoyancy of air parcels and decreases the vertical stability within the PRE structure, encouraging stronger upward vertical motions. Consequently, the QIS-LWCRFp0 experiment produces more condensate (Fig. 4.11b) and generates more latent heat release (Fig. 4.12b). LW cooling at the top of the PRE structure, however, induces downward vertical motions (subsidence) which hinders convective processes and results in weaker PRE development (Fig. 4.10b).

Because QICE and QSNOW were both involved in the calculation of the LW CRF field, the next set of experiments separates the two cloud species and investigates their contribution to PRE intensification individually. The experiment labeled QCI only allows QCLOUD and QICE to interact with both LW and SW radiation, so that QSNOW is transparent to radiation. The experiment labeled QCS only allows QCLOUD and QSNOW to interact with radiation. Again, the interaction between radiation and QCLOUD is maintained for both experiments to ensure PRE initiation occurs.

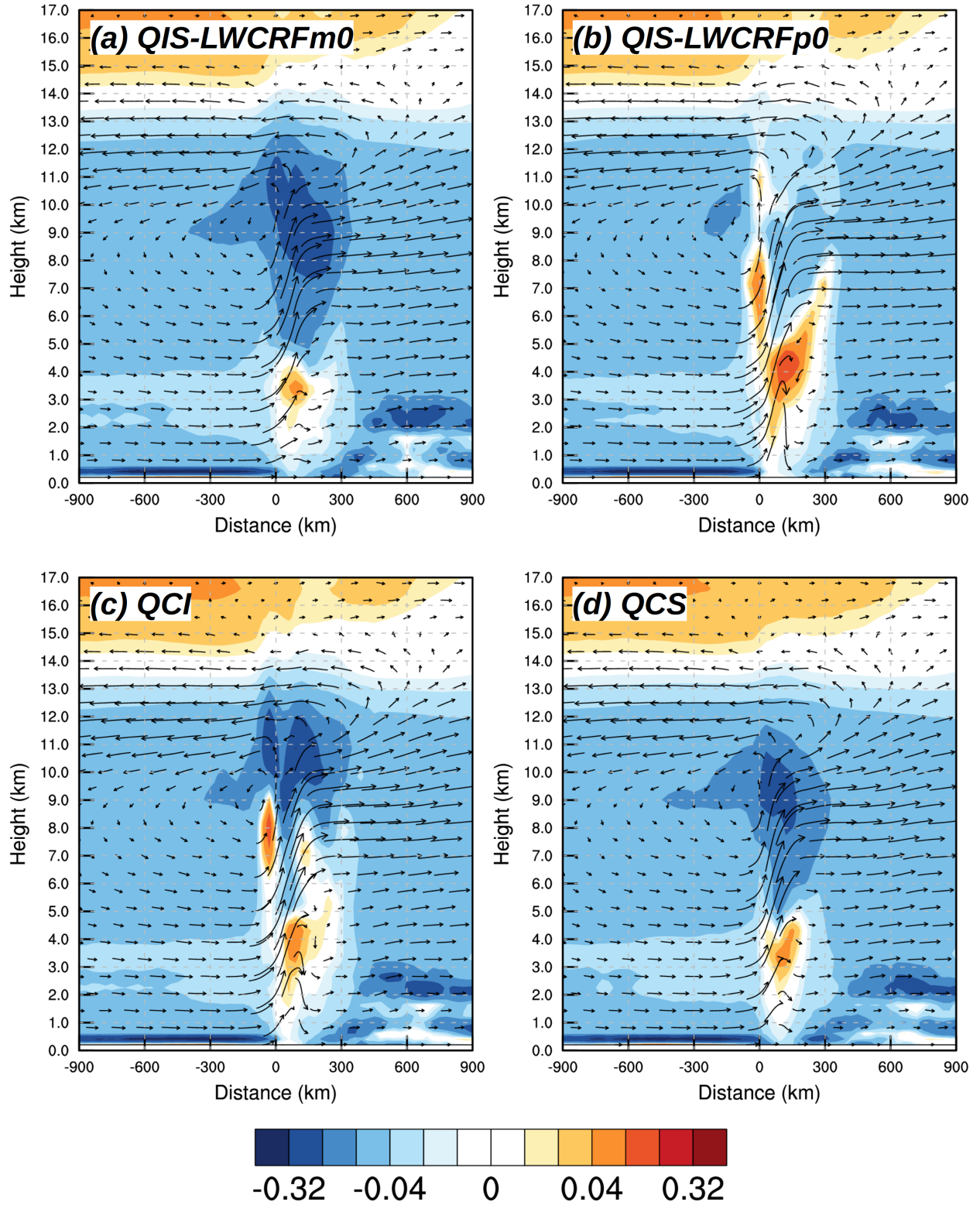


Figure 4.10: Radiative heating tendency (K hr^{-1}) cross-sections as in Fig. 3.13 for (a) QIS-LWCRFm0, (b) QIS-LWCRFp0, (c) QCI, and (d) QCS experiments.

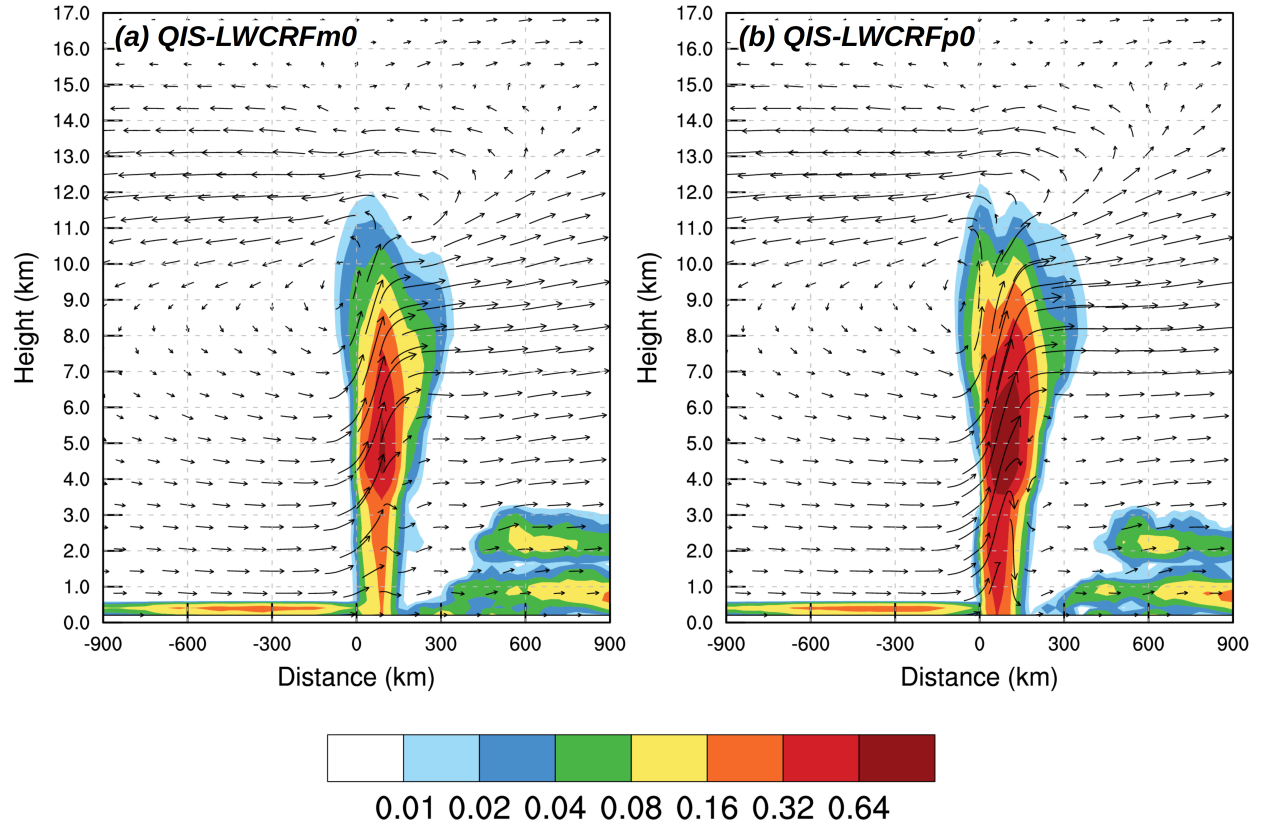


Figure 4.11: Condensate (g kg^{-1}) cross-sections as in Fig. 3.14 for (a) QIS-LWCRFm0 and (b) QIS-LWCRFp0 experiments.

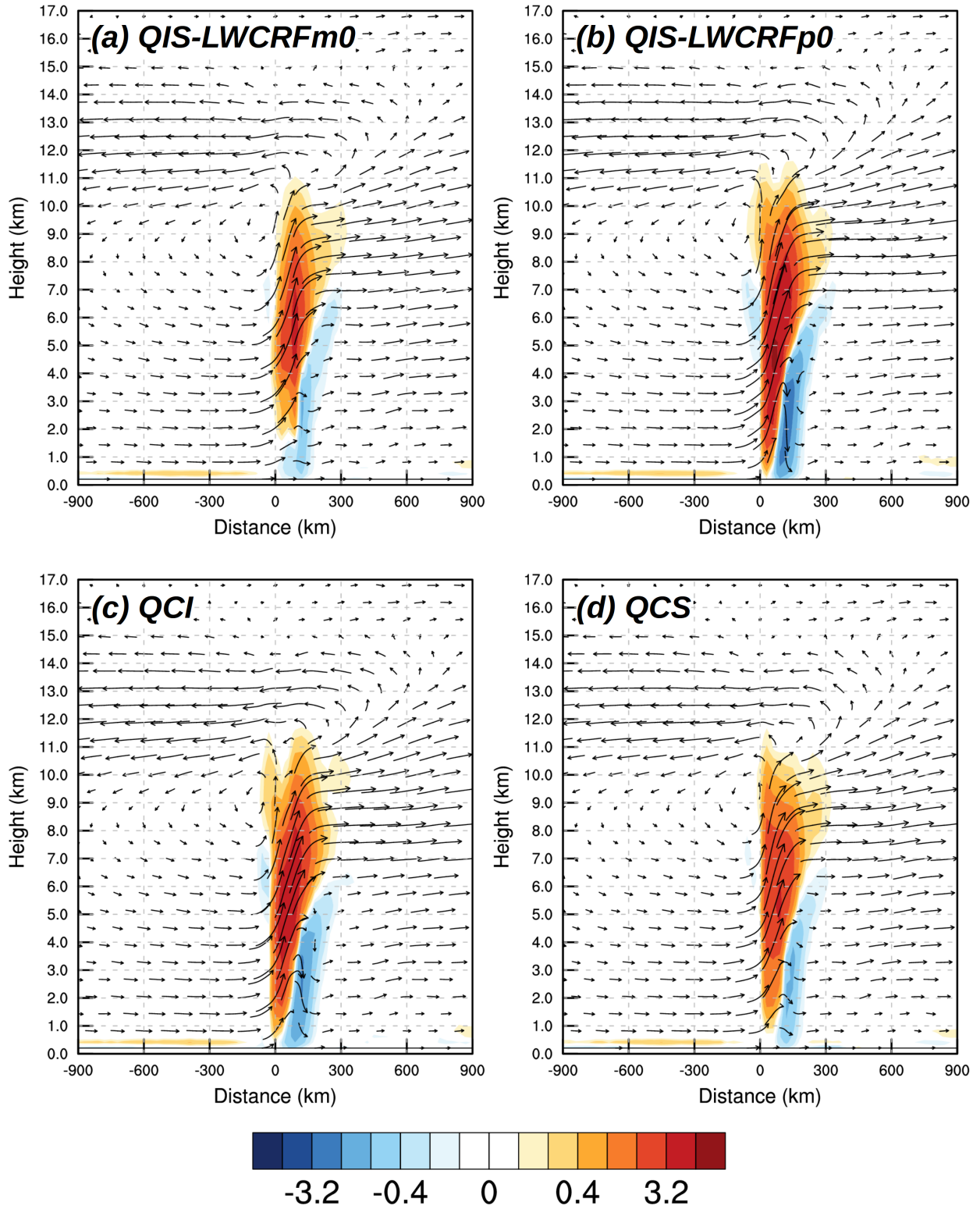


Figure 4.12: MP heating tendency (K hr^{-1}) cross-sections as in Fig. 3.15 for (a) QIS-LWCRFm0, (b) QIS-LWCRFp0, (c) QCI, and (d) QCS experiments.

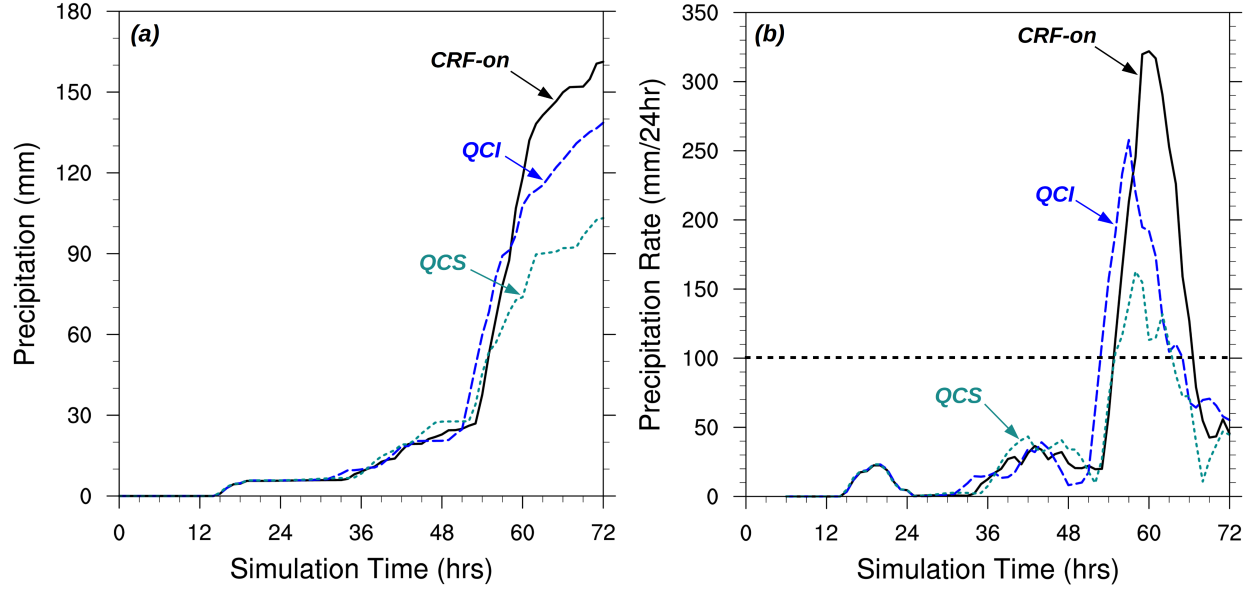


Figure 4.13: Line plots of (a) maximum PRE rainfall (mm) and (b) precipitation rate [$\text{mm}/(24 \text{ hr})^{-1}$] over time as in Fig. 3.7 for CRF-on (solid black), QCI (dashed blue), and QCS (fine-dashed cyan) experiments.

At the end of the 72-hour simulation, the interaction between radiation and the QICE component (QCI) produces a wider swath of heavy precipitation across the PRE region (Fig. 4.8c), as well as 40 mm more rainfall compared to QSNOW component in the QCS experiment (Fig. 4.13a). The QCI experiment also intensifies into a PRE at a similar (albeit slightly faster) rate as the CRF-on experiment, while the development in the QCS experiment is not only slower, but much weaker (Fig. 4.13b).

In the QCI experiment, the warming component of CRF extends vertically through a greater depth from the surface to about 9 km (Fig. 4.10c). In the QCS experiment, the vertical extent of the in-cloud warming is capped at 5 km (Fig. 4.10d). The location of QICE in the upper half of the PRE structure in the QCI experiment (Fig. 4.14a) results in a greater vertical depth of in-cloud warming favorable for enhanced upward motions, increased MP latent heat release (Fig. 4.12c), and more rainfall production. In contrast, QSNOW in the QCS experiment is generally concentrated in the middle levels of the PRE structure (between 3 and 10 km), limiting the vertical extent of the in-cloud warming (Fig. 4.14b).

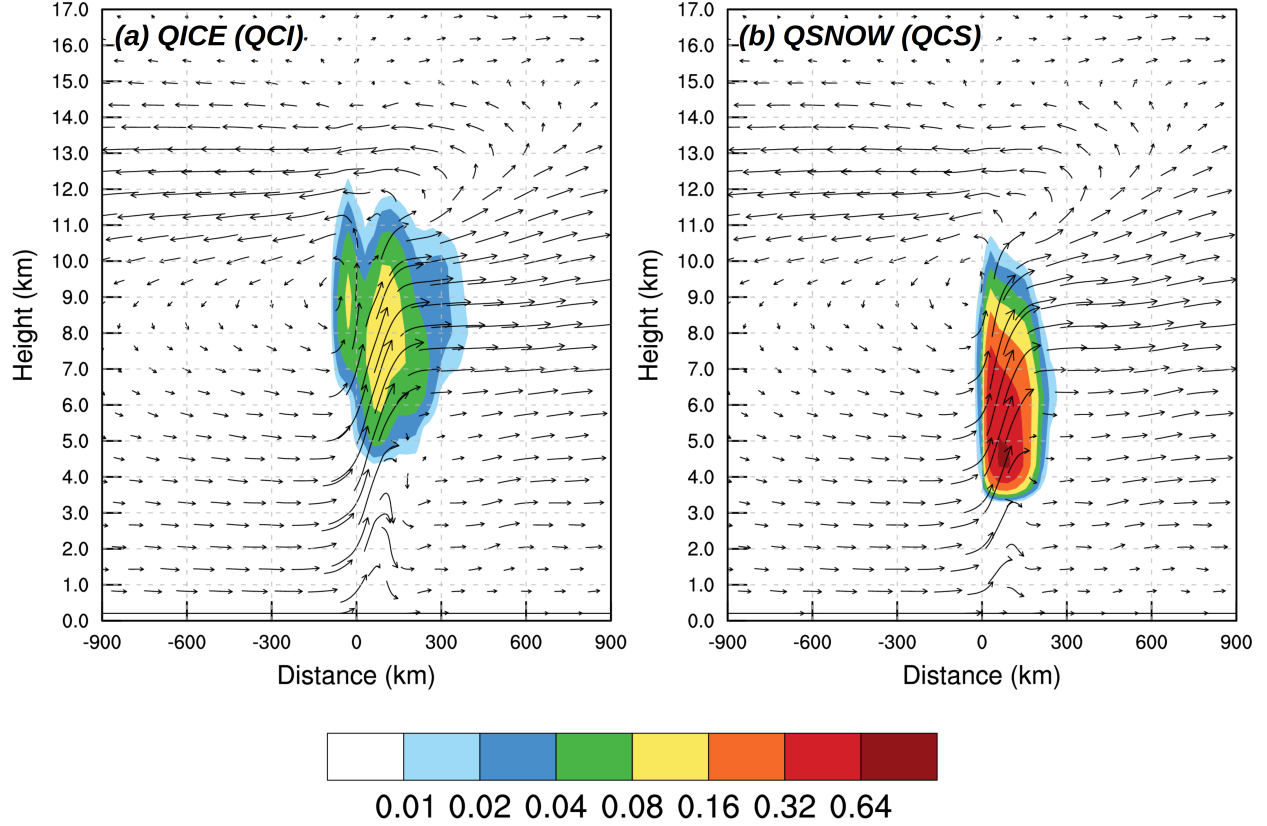


Figure 4.14: Condensate (g kg^{-1}) cross-sections as in Fig. 3.14 for (a) QICE in the QCI experiment and (b) QSNOW in the QCS experiment.

The location of QSNOW also produces a region of LW cooling at a lower height (Fig. 4.10d), hindering convective processes more than the higher altitude cloud top cooling associated with QICE. Overall, the LW warming component of CRF associated with the interaction between radiation and QICE is primarily responsible for the rapid intensification of heavy rainfall in the PRE.

4.3 Impact of shortwave CRF effects

In order to investigate the influence of CRF effects associated with SW radiation, an experiment is conducted where SW CRF effects are deactivated (labeled SWCRF-off). Consequently, SW radiation does not interact with any cloud hydrometeor species and only

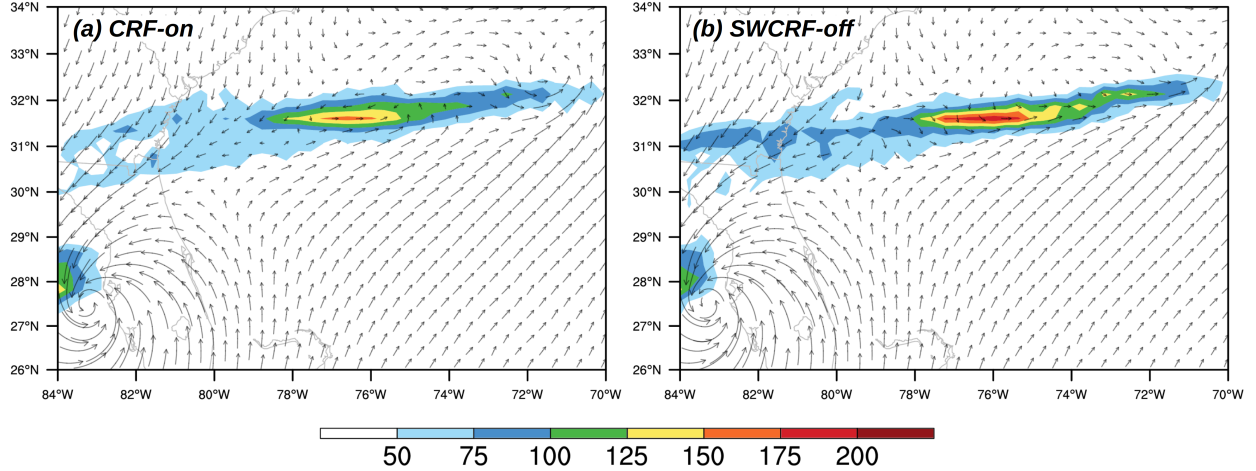


Figure 4.15: Total accumulated precipitation (mm) at T+72 hours as in Fig. 3.6 for (a) CRF-on and (b) SWCRF-off experiments.

produces clear sky SW warming throughout the domain according to the timing of the diurnal cycle. LW CRF effects are unmodified in order to maintain the critical processes responsible for both PRE initiation and PRE intensification.

Even though the distribution of the PRE rainfall is remarkably similar overall, the SWCRF-off experiment clearly produces a greater amount of precipitation than the CRF-on experiment (Fig. 4.15), about 35 mm more (Fig. 4.16a). The timing of PRE initiation (as well as termination) between the two experiments is also nearly identical, but the peak rainfall rate in SWCRF-off experiment is about 50% greater than the CRF-on experiment at T+60 hours (Fig. 4.16b). Consequently, these results suggest a diurnal cycle influence on the intensification of the PRE that is stronger at night and weaker during the day.

The radiative heating distribution indicates three main regions of the PRE structure affected by SW CRF effects (Fig. 4.17c): 1) warming along the upper level cloud tops, 2) in-cloud cooling in the middle levels, and 3) warming along the low-level cloud tops north and south of the PRE region. The in-cloud “cooling” signature in the CRF-on experiment (Fig. 4.17a) is actually a region of reduced clear sky SW warming (not cooling) compared to the clear sky SW warming that occurs everywhere in the SWCRF-off experiment (Fig. 4.17b). The interaction between SW radiation and clouds in the CRF-on experiment effectively

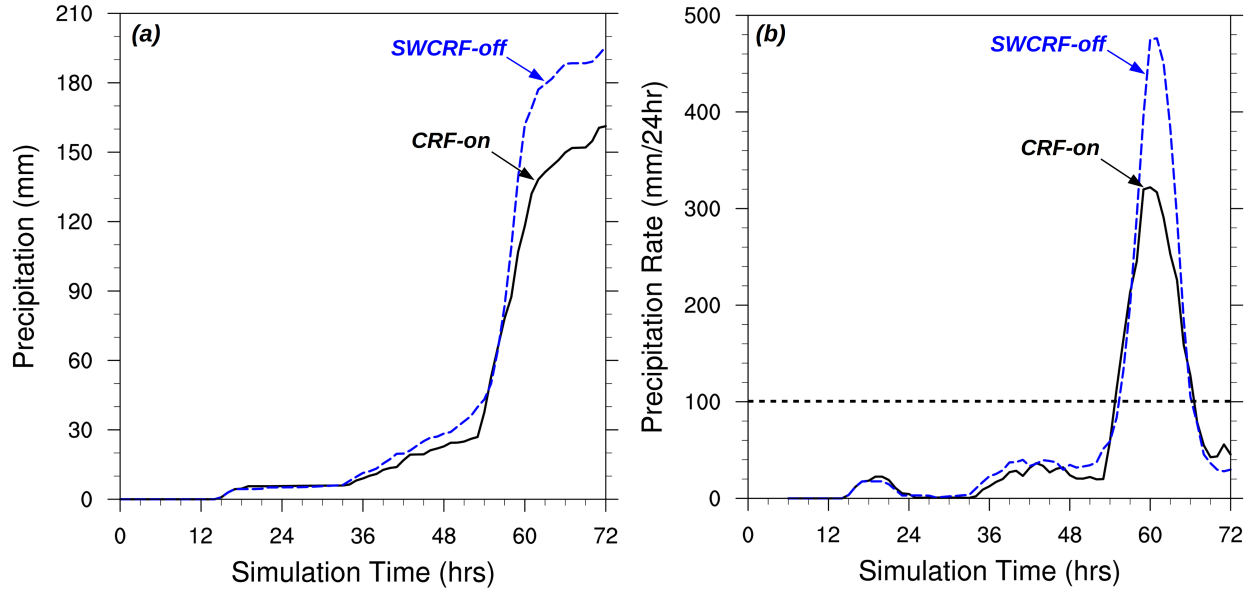


Figure 4.16: Line plots of (a) maximum PRE rainfall (mm) and (b) precipitation rate [$\text{mm}/(24 \text{ hr})^{-1}$] over time as in Fig. 3.7 for CRF-on (solid black) and SWCRF-off (dashed blue) experiments.

reduces the amount of clear sky warming within the cloudy regions of the PRE structure, resulting in relatively less warming or in-cloud “cooling” compared to the CRF-off (not shown) or SWCRF-off experiments.

Four experiments are conducted to investigate the influence of the different SW CRF heating and cooling components individually: 1) QCLOUD SWCRF cooling (QC-SWCRFm0), 2) QCLOUD SWCRF warming (QC-SWCRFp0), 3) QICE-QSNOW SWCRF cooling (QIS-SWCRFm0), and 4) QICE-QSNOW SWCRF warming (QIS-SWCRFp0). QICE and QSNOW are grouped together (QIS) because both cloud species significantly overlap in the middle and upper levels of the PRE structure. A significant rewrite of the RRTMG radiation code would be required to investigate the influence of the two hydrometeor species individually. The warm cloud component is easier to isolate and investigate since QCLOUD is mainly concentrated in the lower troposphere near the surface. As a note, the SW CRF modifications only act where QC or QIS is the sole cloud species component in the grid box for each respective experiment; otherwise, only clear sky SW radiation effects are calculated.

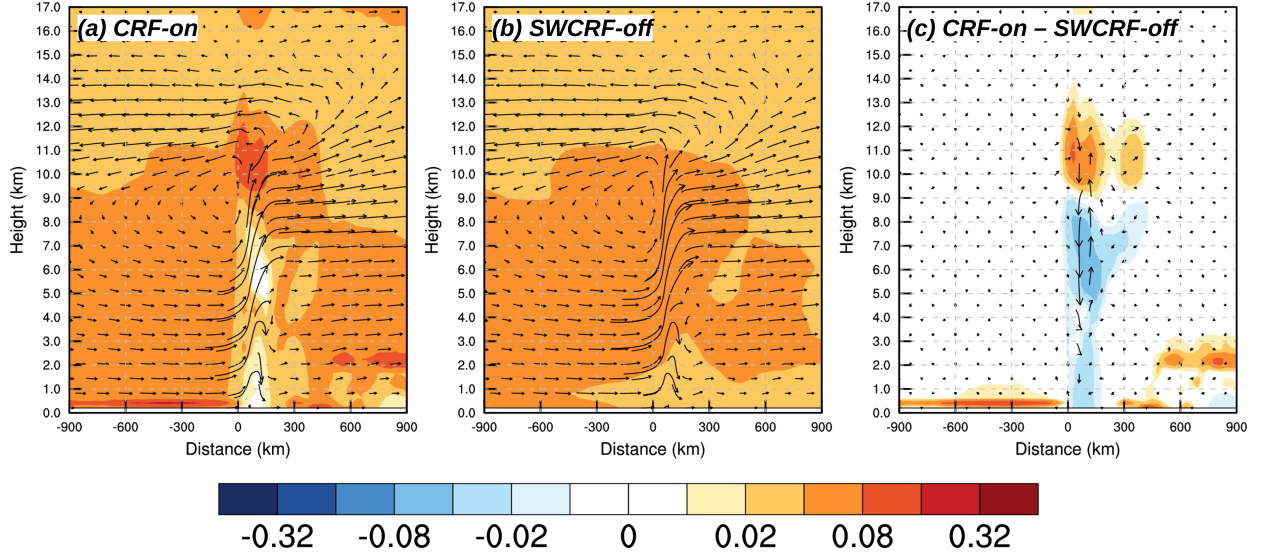


Figure 4.17: SW radiative heating tendency (K hr^{-1}) cross-sections as in Fig. 3.13 for (a) CRF-on and (b) SWCRF-off experiments.

The total precipitation accumulation at the end of the 72-hour simulation indicates that the QIS-SWCRFm0 experiment produces the largest decrease in PRE rainfall, suggesting that the SW in-cloud cooling component of CRF has the most negative influence on PRE development (Fig. 4.18b). In contrast, the upper level cloud top warming in the QIS-SWCRFp0 experiment has the least influence on weakening the intensity of the PRE. Both QCLOUD-related SWCRF warming and cooling experiments (QC-SWCRFp0 and QC-SWCRFm0) produce PREs with approximately the same rainfall distribution and intensity. The greater variation in the distribution of PRE rainfall in the QIS-SWCRF experiments arises because PRE intensification is more sensitive to the CRF effects associated with QICE and QSNOW (e.g. QC-only, QCI, and QCS experiments).

Local cooling in the QIS-SWCRFm0 experiment not only reduces in-cloud LW warming within the PRE structure (less parcel buoyancy), but also induces downward vertical motions which weaken convective processes (Fig. 4.19c). The QC-SWCRFp0 experiment produces the next weakest PRE by generating widespread warming along the low-level clouds to the north and south of the PRE region (Fig. 4.19b). During the day, the low-level SW warming counters the favorable low-level LW cooling primarily responsible for PRE initiation.

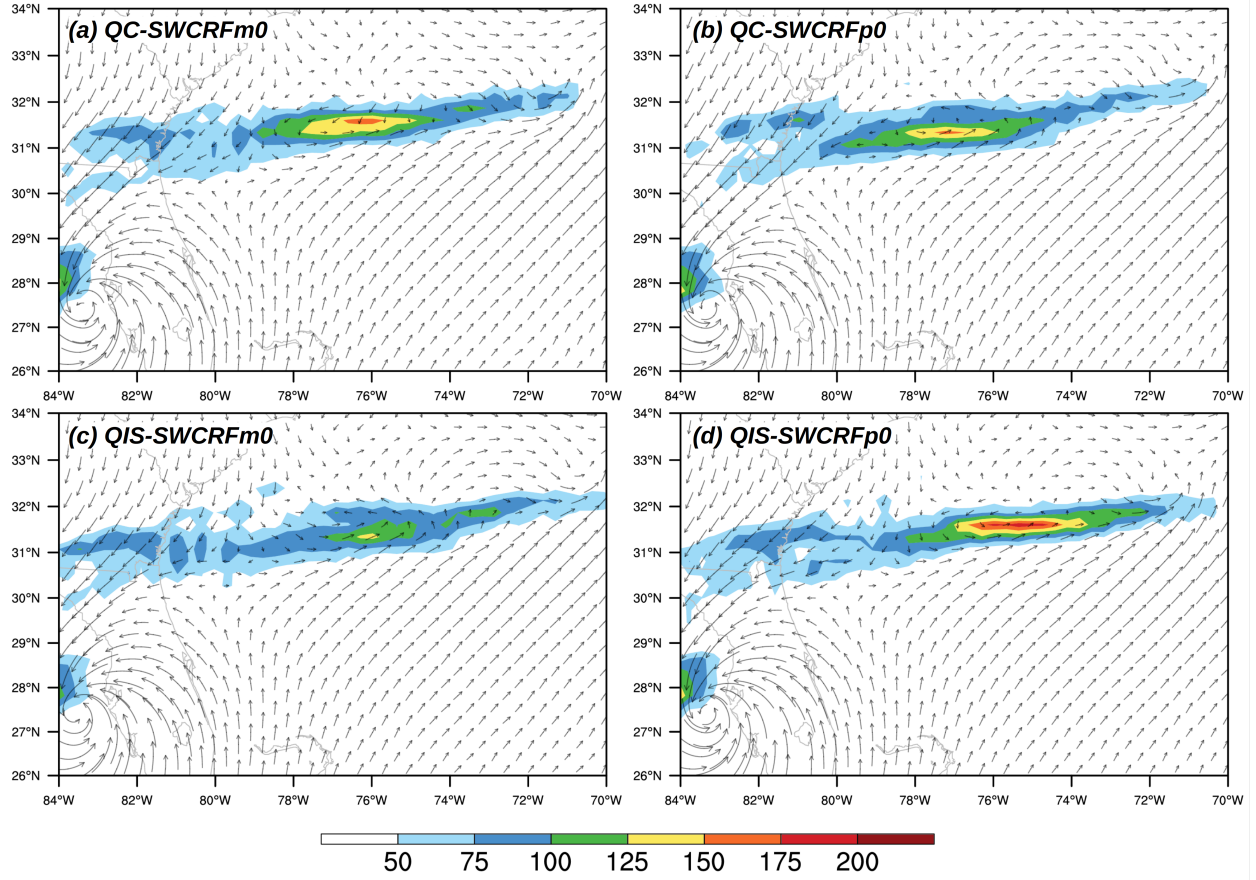


Figure 4.18: Total accumulated precipitation (mm) at T+72 hours as in Fig. 3.6 for (a) QC-SWCRFm0, (b) QC-SWCRFp0, (c) QIS-SWCRFm0, and (d) QIS-SWCRFp0 experiments.

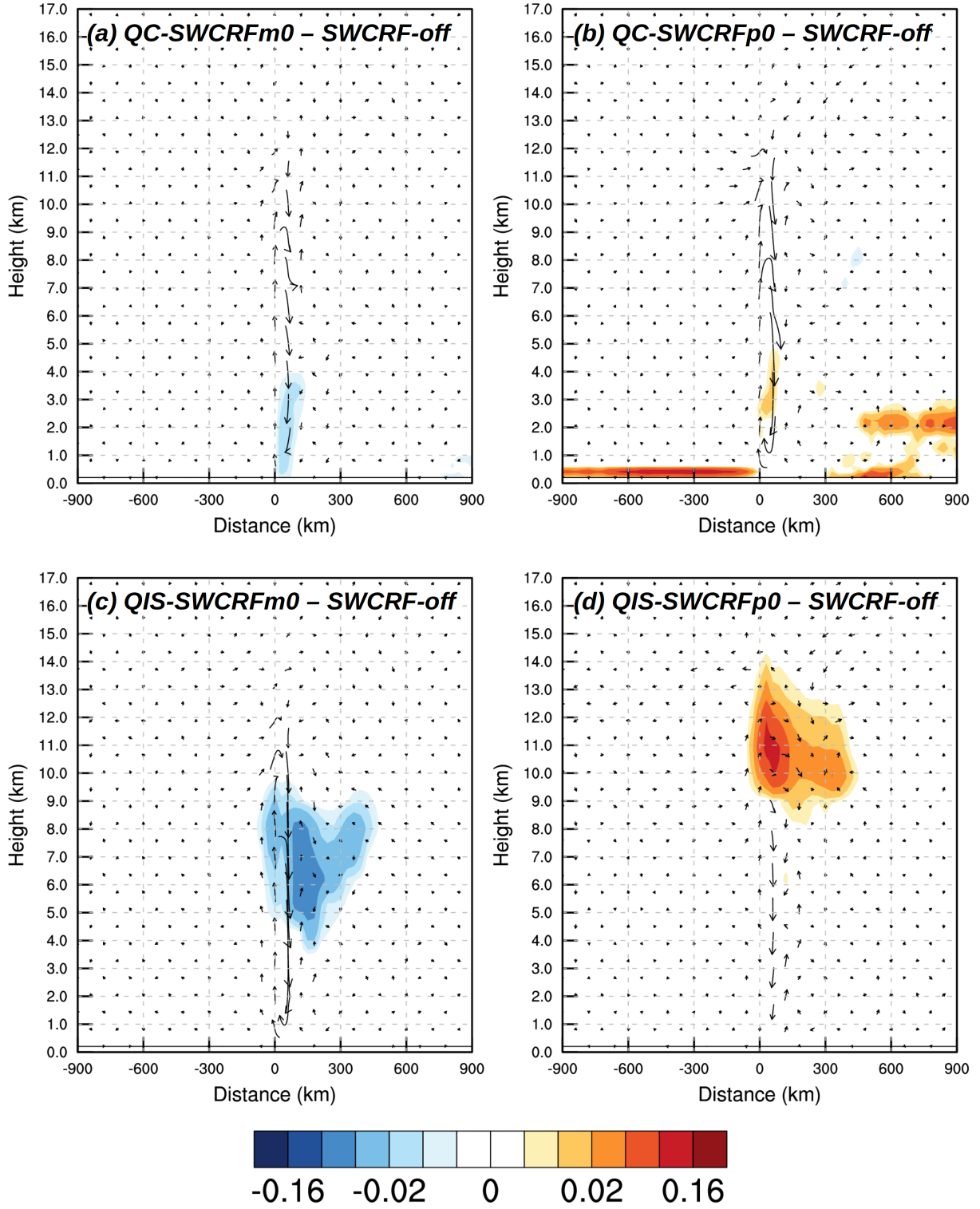


Figure 4.19: Difference cross-sections of SW radiative heating tendency (K hr^{-1}) from SWCRF-off as in Fig. 3.13 for (a) QC-SWCRFm0, (b) QC-SWCRFp0, (c) QIS-SWCRFm0, and (d) QIS-SWCRFp0 experiments.

Furthermore, the QC-SWCRFm0 experiment produces in-cloud SW cooling at the base of the PRE structure, generating localized subsidence that hinders upward vertical motions (Fig. 4.19a). Lastly, the upper level cloud top warming in the QIS-LWCRFp0 experiment increases the vertical stability within the storm, which is not conducive to strong convection and rainfall production (Fig. 4.19d).

Because many different components of SW CRF contribute to weaker PRE development, it is important to connect changes in the heating distribution to changes in the pressure gradients and wind fields across the PRE region. During the day, in-cloud SW cooling not only induces downward vertical motions throughout the PRE structure, but also produces divergence near the surface and convergence in the upper levels. Diverging winds at the base of the PRE (in combination with the SW warming along the low-level cloud tops) reduces the magnitude of the low-level convergence and may contribute to frontolysis (opposite of frontogenesis) or a weakening of the horizontal temperature gradient. In addition, converging winds at the top of the PRE structure entrain dry upper-tropospheric air into the core of the storm, encouraging evaporative cooling and downward vertical motions. The unfavorable secondary circulation associated with SW CRF effects is clearly not conducive to strong PRE intensification.

At night, however, there are no SW CRF effects to hinder PRE development. In-cloud LW warming encourages stronger upward vertical motions throughout the PRE structure, as well as convergence underneath the PRE (aided by LW cooling along the low-level clouds tops) and divergence at the top of the storm. The favorable secondary circulation associated with LW CRF effects complements the direct thermal circulation associated with frontogenesis, further enhancing convective processes within the PRE structure and producing more intense rainfall. Therefore, the influence of cloud-radiative processes (both LW and SW) on the convective secondary circulation suggests that PRE development may be sensitive to the diurnal cycle (e.g. Albrecht and Cox 1975; Gray and Jacobson 1977).

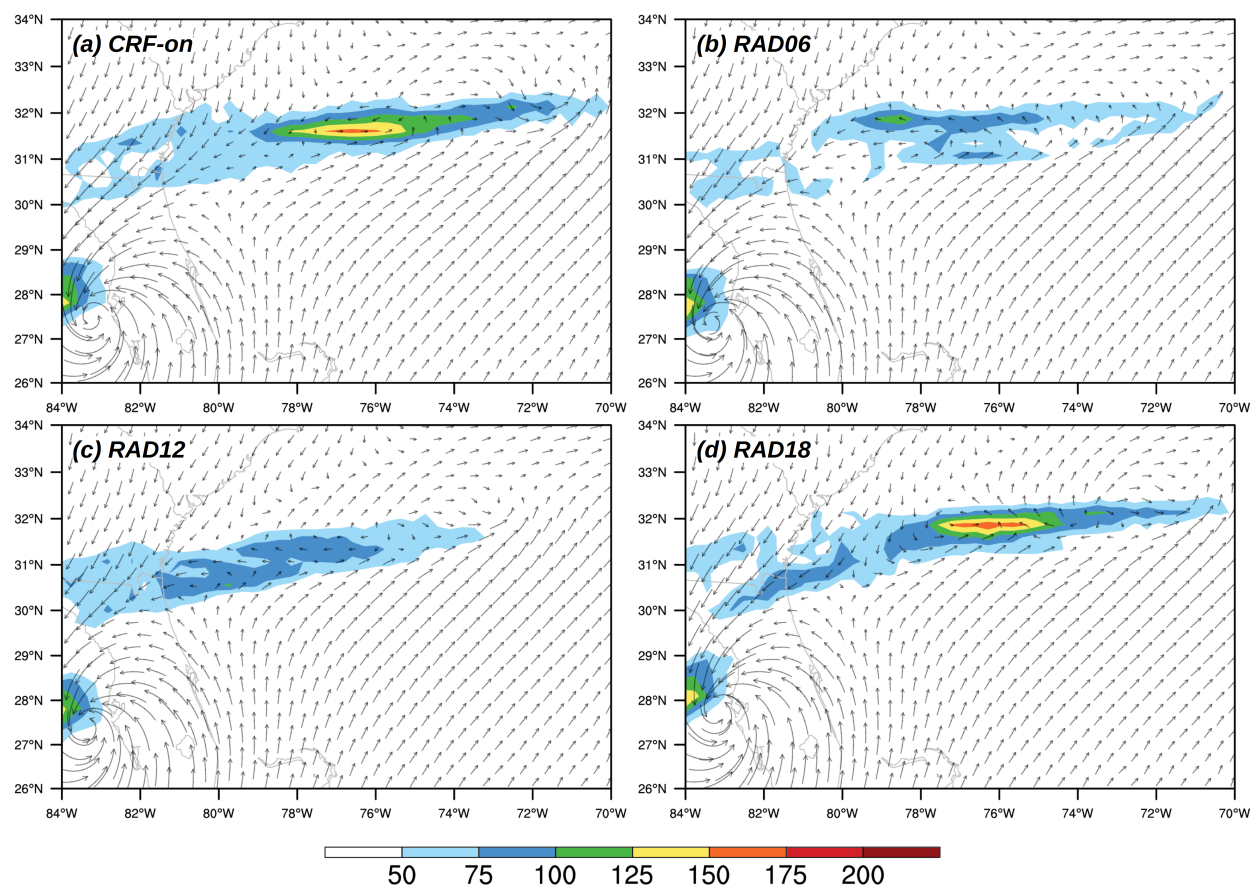


Figure 4.20: Total accumulated precipitation (mm) at T+72 hours as in Fig. 3.6 for (a) CRF-on, (b) RAD06, (c) RAD12, and (d) RAD18 experiments.

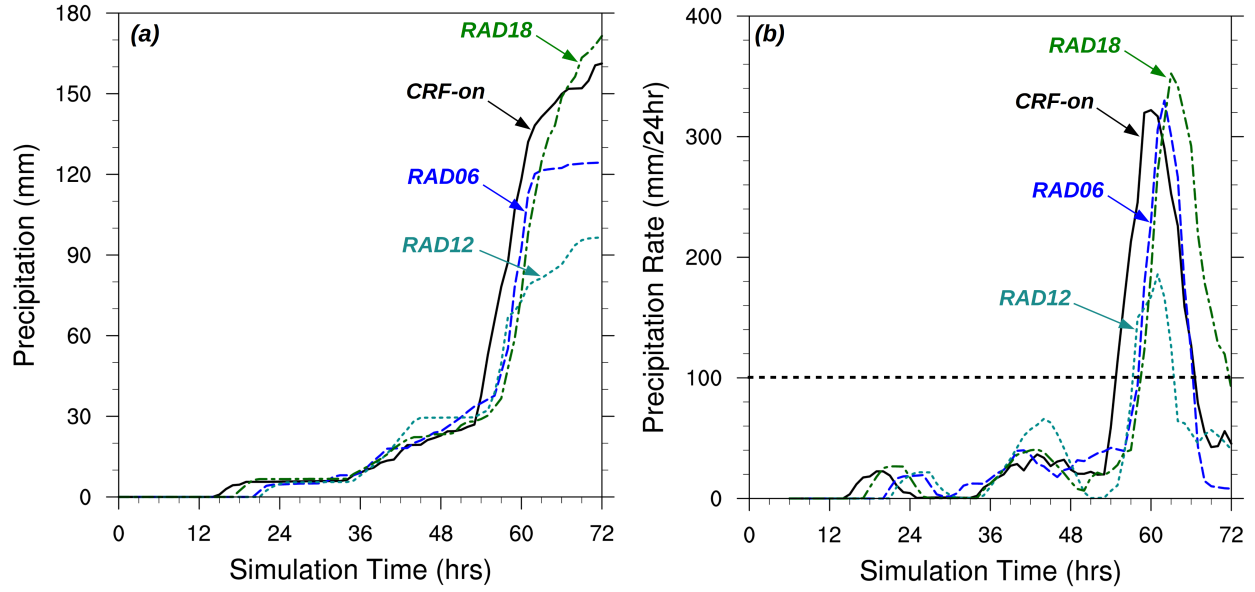


Figure 4.21: Line plots of (a) maximum PRE rainfall (mm) and (b) precipitation rate [mm (24 hr)^{-1}] over time as in Fig. 3.7 for CRF-on (sold black), RAD06 (dashed blue), RAD12 (fine-dashed cyan), and RAD18 (dot-dashed green) experiments.

4.3.1 Diurnal cycle sensitivity

Three experiments are conducted to examine the influence of the diurnal cycle on PRE intensification. The timing of the diurnal cycle in the control (CRF-on) experiment is offset by +6 (RAD06), +12 (RAD12), and +18 (RAD18) hours for each of the experiments. The differences in the timing of the diurnal cycle manifest in changes to the calculation of SW clear sky and SW CRF effects at different stages of PRE development throughout the 72-hour simulation. The LW radiation scheme and LW CRF effects are unmodified.

At the end of the 72-hour simulation, the CRF-on and RAD18 experiments produce the most rainfall, while the RAD06 and RAD12 experiments generate the least (Fig. 4.20). The largest difference in maximum precipitation is about 75 mm between the RAD18 and RAD12 experiments (Fig. 4.21a), suggesting that even a small 6-hour change in the timing of the diurnal cycle can have a significant influence on the overall intensification of the PRE. The timing of PRE initiation and peak rainfall also vary as much as 6 hours among

the experiments (Fig. 4.21b). Lastly, the CRF-on and RAD18 experiments produce longer duration PREs (12-15 hours), compared to the RAD06 and RAD12 experiments (6-9 hours).

Since LW CRF effects are always active, the variations among the experiments must be driven by differences in the timing of SW radiation effects (Fig. 4.22). From PRE initiation to PRE intensification (T+54 to T+60 hours), the CRF-on and RAD18 experiments are in nighttime, or near-nighttime, conditions while the RAD06 and RAD12 experiments are essentially in daytime conditions. This is consistent with the results of the SWCRF-off experiment, indicating that SW CRF effects weaken PRE development mainly through in-cloud cooling and low-level cloud top warming. Without SW CRF effects at night, the magnitude of low-level convergence underneath the PRE and in-cloud warming within the PRE structure is greater, driving stronger convective processes and more rainfall production. Even though the influence of the diurnal cycle timing on TC intensity is minimal (Fig. 4.23), the strong impact of SW radiation on PRE development cannot be ignored.

For brevity, the radiative heating tendency is compared and contrasted only for the RAD18 (strongest PRE) and the RAD12 (weakest PRE) experiments (Fig. 4.24). As indicated by the wind vector difference field, the RAD18 experiment (nighttime) produces a stronger secondary circulation favorable for PRE intensification (Fig. 4.24c), compared to the RAD12 (daytime) experiment. The RAD18 experiment also produces more LW cooling along the cloud tops outside of the PRE structure, which enhances low-level convergence across the PRE region. More in-cloud warming at the base of the PRE also reduces vertical stabilities, which encourages stronger upward vertical motions and increased latent heat release (Fig. 4.25c). Therefore, the timing of daytime and nighttime conditions relative to the stage of PRE development can have a significant influence on the overall magnitude of the heavy rain event.

The influence of clear sky SW radiation effects is also investigated. The diurnal cycle differences between the original RAD18 and RAD12 experiments are maintained, but SW CRF effects are deactivated. Without SW CRF effects, there is a significant increase in the amount of PRE rainfall (at least 50 mm) for both experiments (Fig. 4.26a). PRE initiation is also accelerated forward about 6 hours in the RAD18-SWCRF0 experiment, but

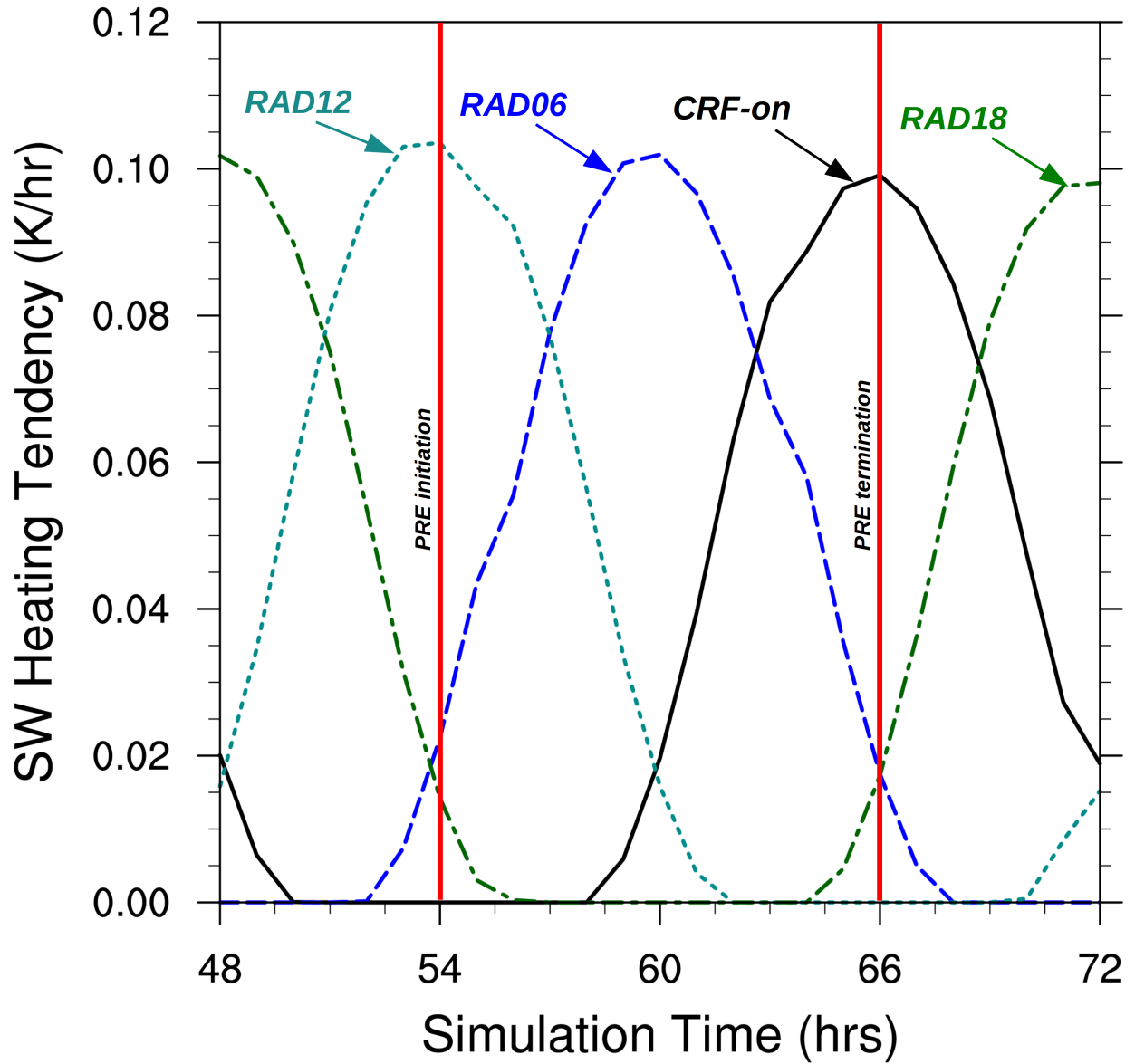


Figure 4.22: Average SW radiative heating tendency (K hr^{-1}) across the PRE region for (a) CRF-on (solid black), (b) RAD06 (dashed blue), (c) RAD12 (fine-dashed cyan), and (d) RAD18 (dot-dashed green) experiments. Solid vertical lines (red) indicate when PRE initiation ($T+54$) and PRE termination ($T+66$) occur in the CRF-on experiment.

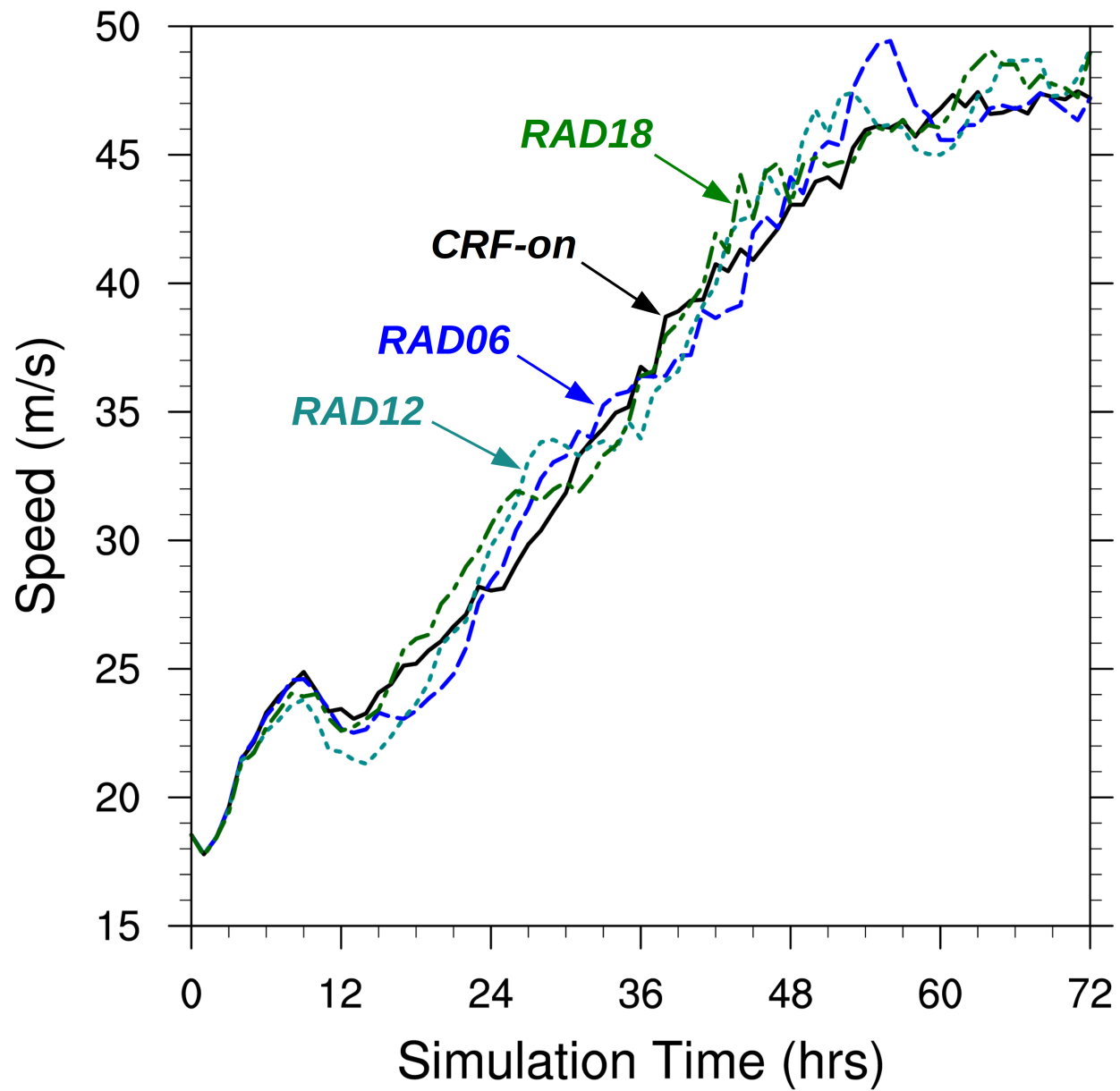


Figure 4.23: Maximum TC wind speed (m s^{-1}) over time for (a) CRF-on (solid black), (b) RAD06 (dashed blue), (c) RAD12 (fine-dashed cyan), and (d) RAD18 (dot-dashed green) experiments

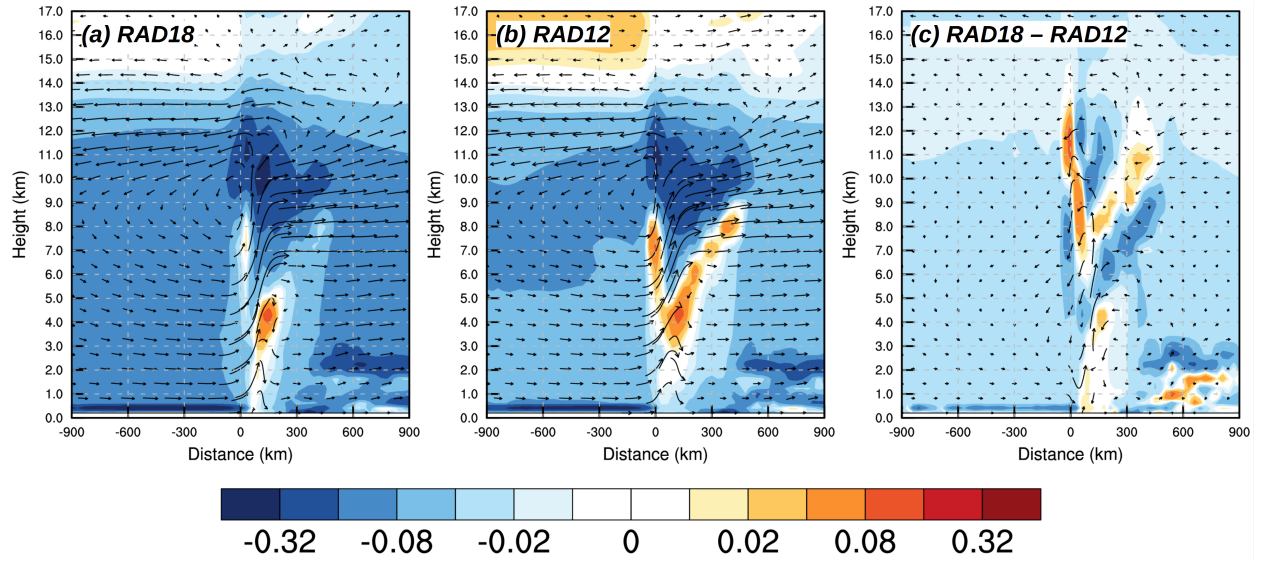


Figure 4.24: Radiative heating tendency (K hr^{-1}) cross-sections as in Fig. 3.13 for (a) RAD18, (b) RAD12, and (c) RAD18 - RAD12 experiments.

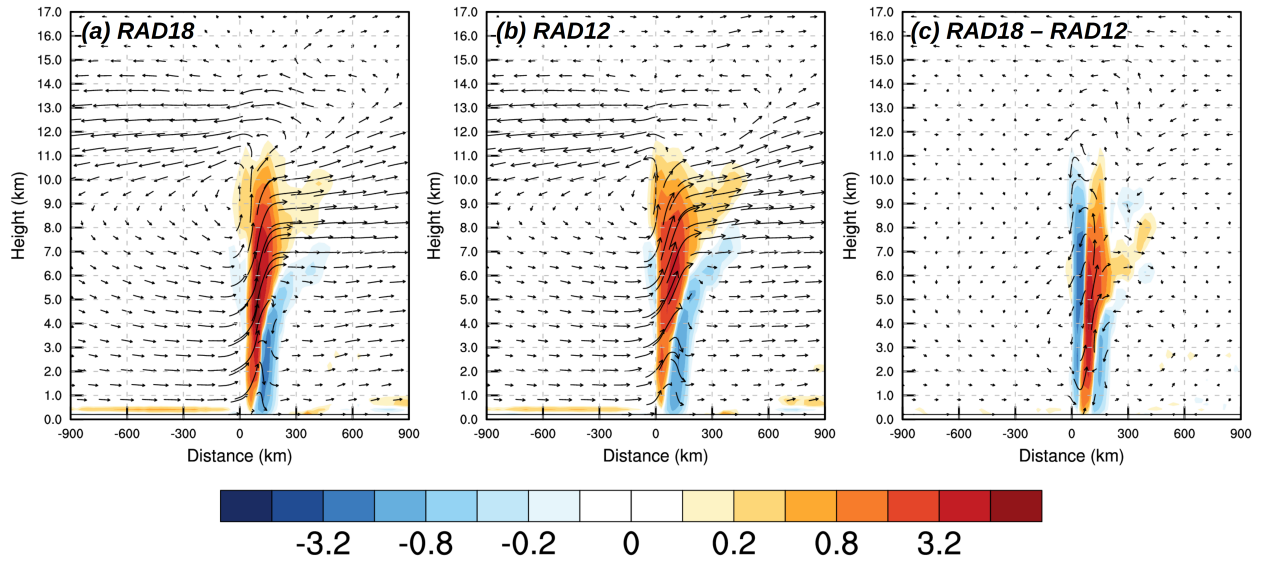


Figure 4.25: MP heating tendency (K hr^{-1}) cross-sections as in Fig. 3.15 for (a) RAD18, (b) RAD12, and (c) RAD18 - RAD12 experiments.

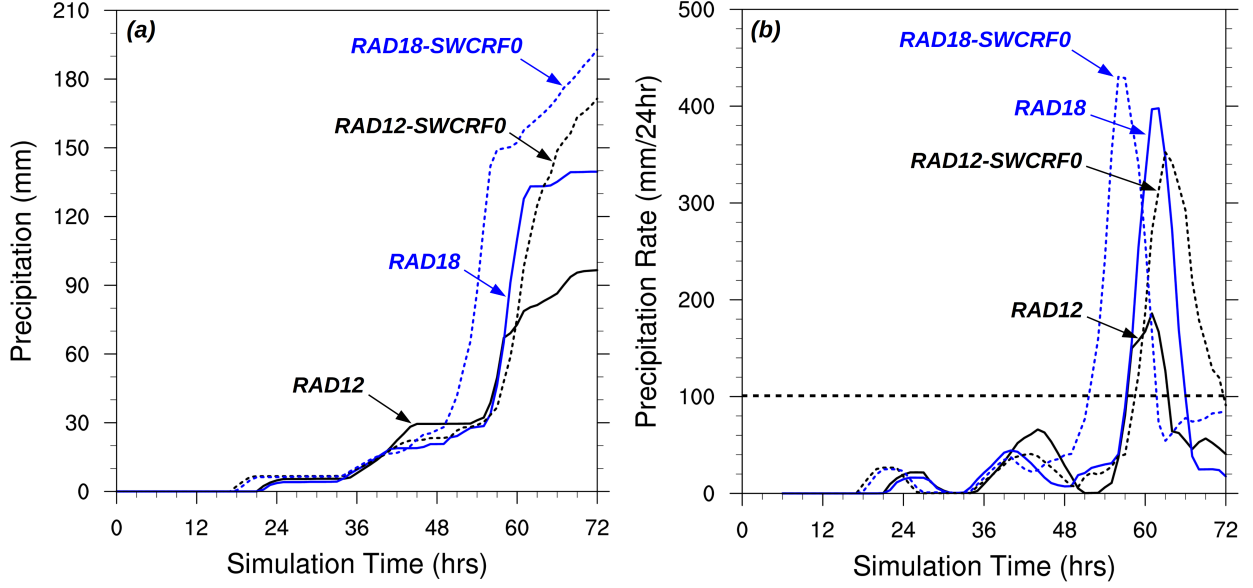


Figure 4.26: Line plots of (a) maximum PRE rainfall (mm) and (b) precipitation rate [mm (24 hr)^{-1}] over time as in Fig. 3.7 for RAD12 (solid black), RAD18 (solid blue), RAD12-SWCRF0 (dashed-black), and RAD18-SWCRF0 (dashed-blue) experiments.

slightly delayed in the RAD12-SWCRF0 experiment (Fig. 4.26b). Because the difference in TC intensity among the four experiments is not significant (not shown), the variations in the overall PRE intensity must be driven by the magnitude and timing of *clear sky* SW radiation. SW CRF effects, however, introduce more variability to the heating distribution, further complicating the PRE forecast.

4.3.2 Seasonal length of day sensitivity

The influence of SW radiation on PRE development also suggests a seasonal sensitivity associated with different lengths of day. In the SW radiation scheme, the JDN is modified to investigate the influence of different lengths of day associated with the Winter (20 December, JDN 354) and Summer (20 June, JDN 171) experiments. Because LW CRF effects are always active, the difference in the length of day only apply to calculations of clear sky SW radiation and SW CRF heating. All other seasonally-dependent environmental parameters (e.g. SST, temperature, winds, etc.) are unchanged in the summer and winter experiments, even if

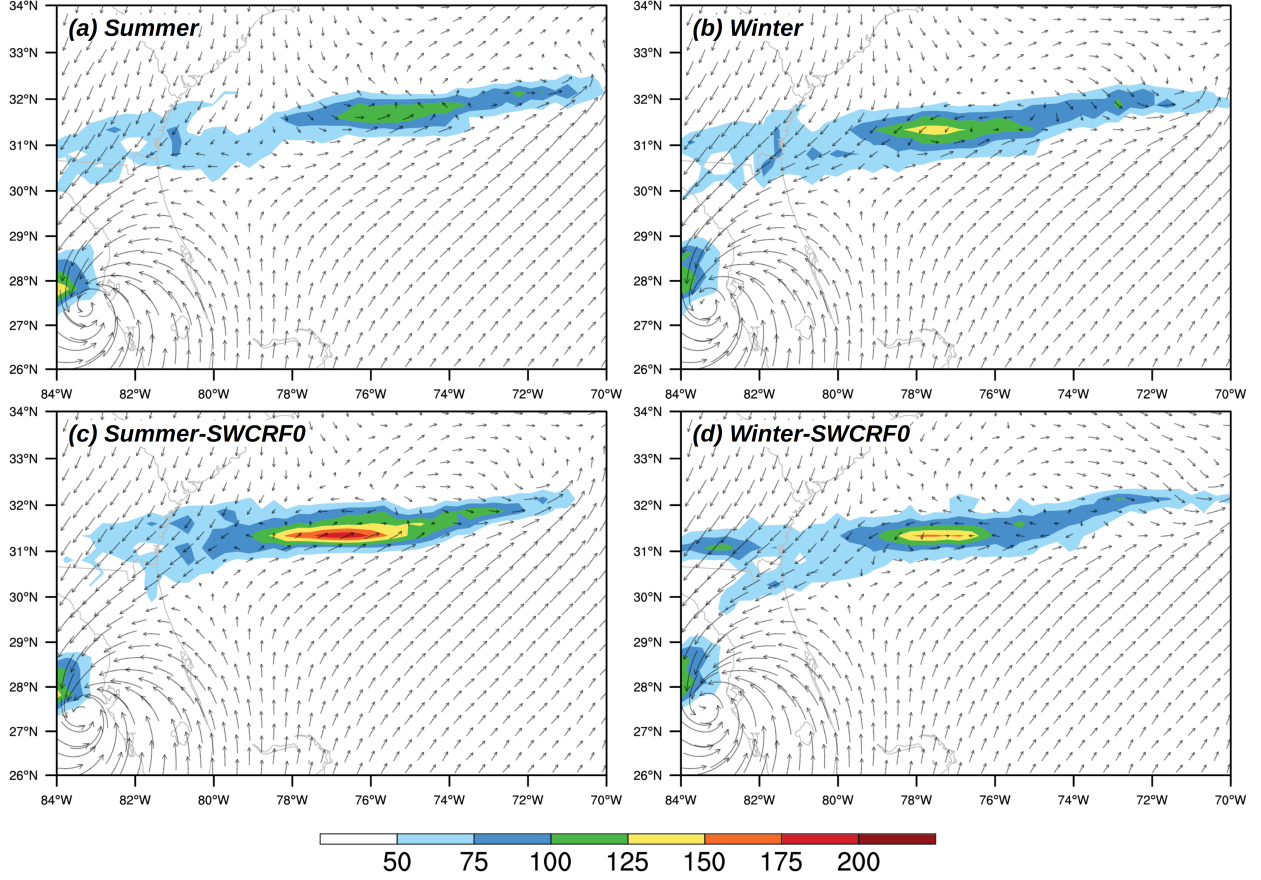


Figure 4.27: Total accumulated precipitation (mm) at T+72 hours as in Fig. 3.6 for (a) Summer, (b) Winter, (c) Summer-SWCRF0, and (d) Winter-SWCRF0 experiments.

they are not climatologically consistent. Furthermore, even though TCs have been observed between the months of April and December, no PREs have been documented in the month of December (Galarneau et al. 2010). TCs are less likely to make landfall in the United States during the winter season; colder SSTs and increased vertical wind shear are both unfavorable for TC genesis and storm longevity in the northern hemisphere (Chan and Kepert 2010).

The Winter experiment clearly produces a broader and more intense PRE, compared to the Summer experiment (Fig. 4.27a,b). Even though there is only about a 10 mm difference in maximum rainfall (Fig. 4.28a), the Winter experiment PRE is consistently stronger for the entire duration of the 72-hour simulation. Moreover, the PRE rainfall rate in the Winter experiment is not only greater, but also initiates, peaks, and terminates about 4

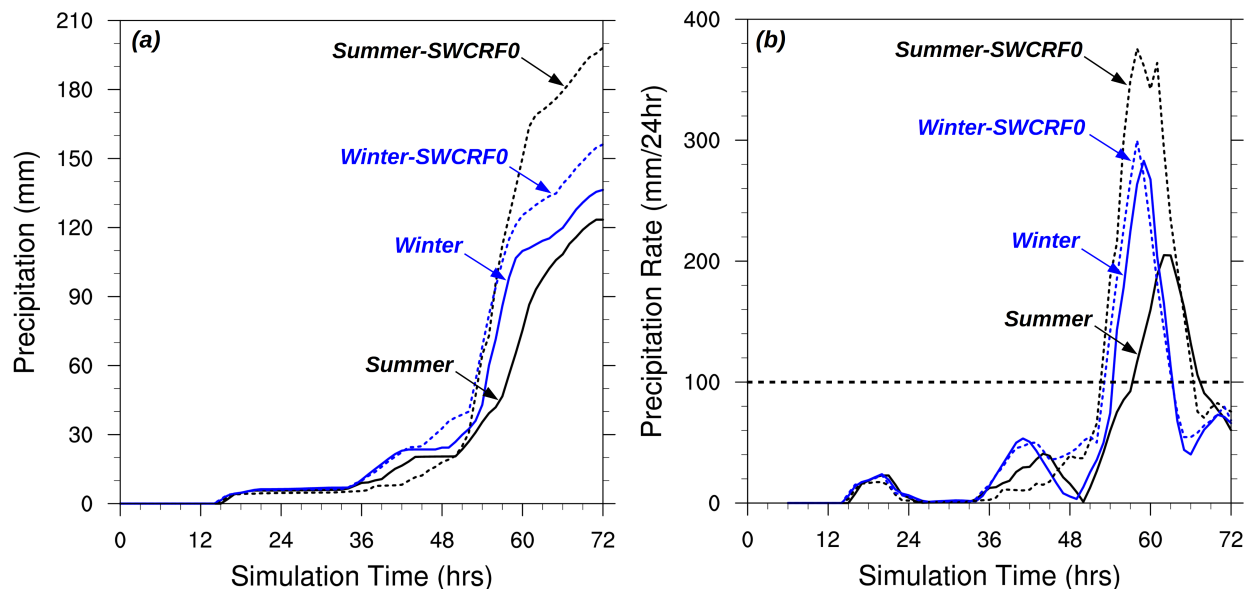


Figure 4.28: Line plots of (a) maximum PRE rainfall (mm) and (b) precipitation rate [mm (24 hr)^{-1}] over time as in Fig. 3.7 for Summer (solid black), Winter (solid blue), Summer-SWCRF0 (dashed black), and Winter-SWCRF0 (dashed blue) experiments.

hours earlier than in the Summer experiment (Fig. 4.28b). The longer days in the Summer experiment increase the amount of SW radiative heating across the PRE region (Fig. 4.30). This generates more in-cloud cooling within the PRE structure and more SW warming along the low-level cloud tops (Fig. 4.29a), which both contribute to weaker PRE development. Overall, the negative influence of SW CRF is maximized during the summer season (weaker PRE) and minimized during the winter season (stronger PRE).

The SWCRF-off versions of the Winter and Summer experiments are also compared and contrasted. For both experiments, labeled Winter-SWCRF0 and Summer-SWCRF0, the SW radiation scheme only produces clear sky SW radiative heating and no SW CRF effects. LW radiation and LW CRF effects are also unmodified. With SW CRF effects disabled, the Summer-SWCRF0 experiment now produces much more rainfall (about 40 mm) than the Winter-SWCRF0 experiment (Figs. 4.27c,d), a reversal of the trend with SW CRF effects activated (Fig. 4.28a). The Summer-SWCRF0 experiment also initiates earlier than the Winter-SWCRF0 experiment and terminates about 3 hours later, producing a longer

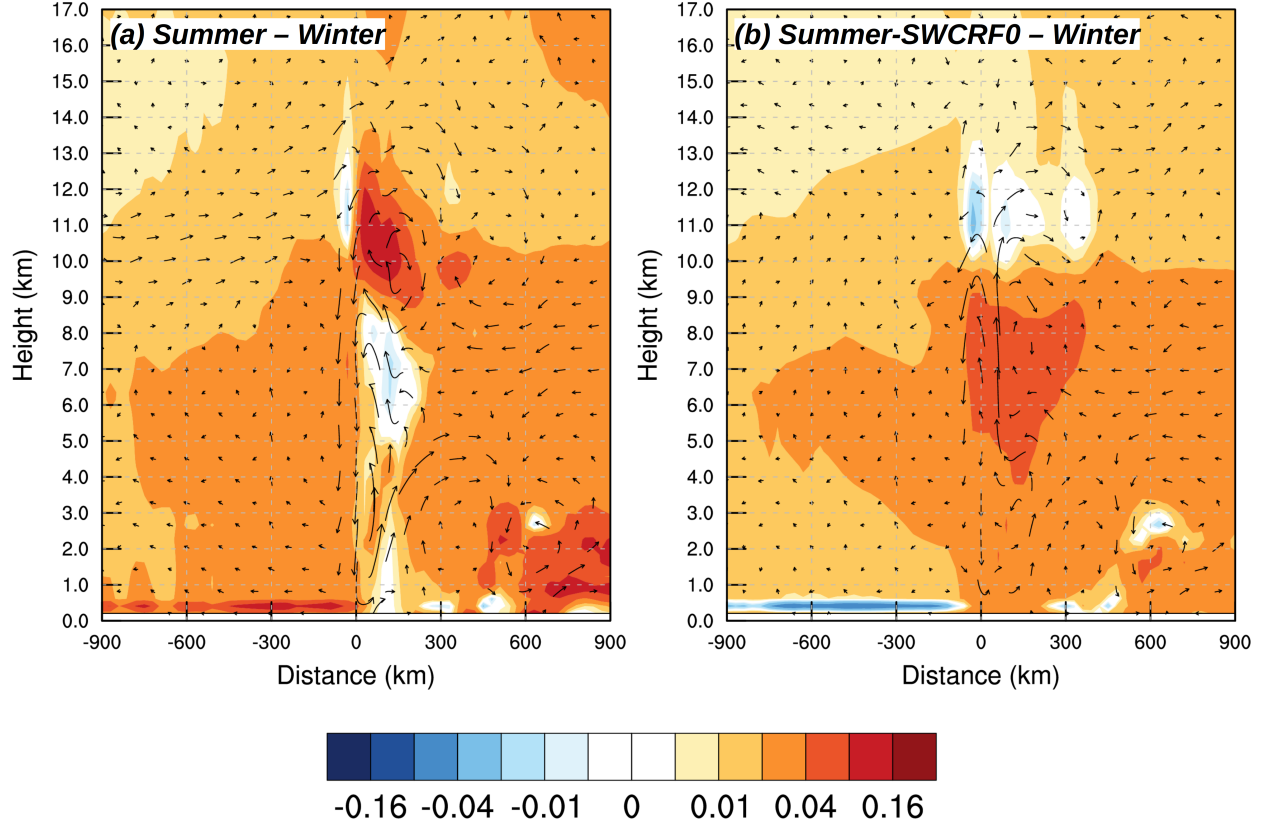


Figure 4.29: SW radiative heating tendency difference (K hr^{-1}) cross-sections as in Fig. 4.19 for (a) Summer - Winter and (b) Summer-SWCRF0 - Winter. Note the reversal in the SW heating difference when SW CRF effects are deactivated for the Summer experiment in panel (b).

duration PRE (Fig. 4.28b). Both experiments also generate more precipitation than their SWCRF-on counterparts, even though the magnitude of SW radiative heating across the PRE region is slightly decreased without SW CRF effects (Fig. 4.30).

The distribution of SW radiative heating indicates that the Summer-SWCRF0 experiment no longer generates SW warming along the low-level cloud tops and no in-cloud SW cooling within the PRE structure (Fig. 4.29b), which are both favorable for PRE initiation and PRE intensification, respectively. Longer days in the Summer-SWCRF0 experiment also maximize the magnitude of clear sky SW warming within the PRE structure, which complements convective processes and increases rainfall production. The stark difference in

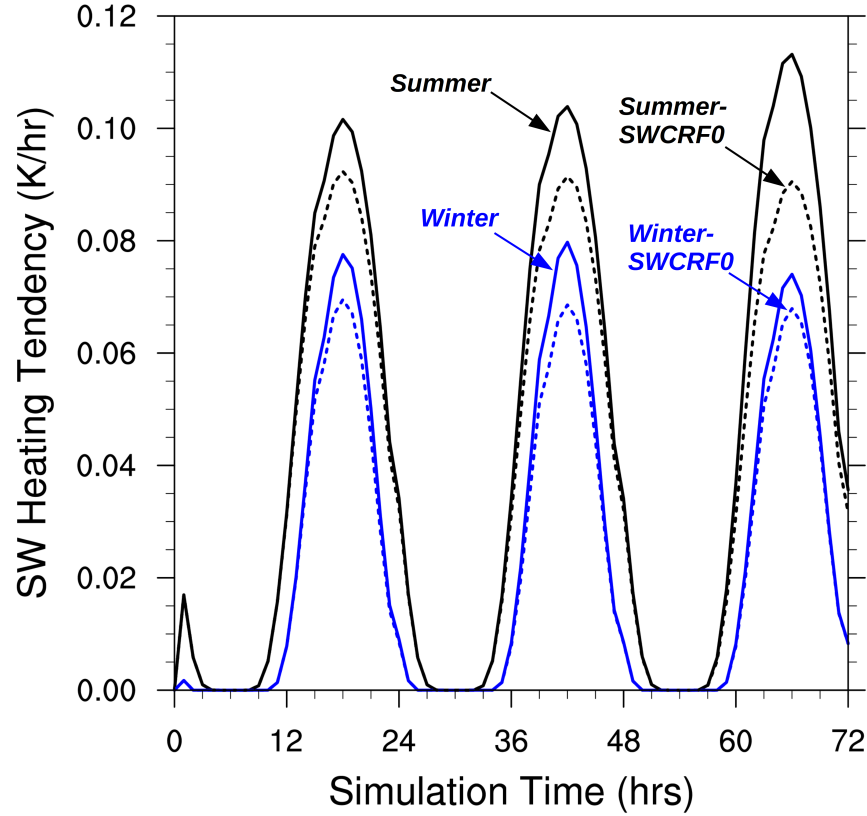


Figure 4.30: Average SW radiative heating tendency (K hr^{-1}) across the PRE region as in Fig. 4.22 for Summer (solid black), Winter (solid blue), Summer-SWCRF0 (dashed black), and Winter-SWCRF0 (dashed blue) experiments.

PRE intensity between the SWCRF-on and SWCRF-off Summer experiments emphasizes that SW radiative effects need to be properly represented to accurately forecast PRE development during the North American hurricane season, extending from June to November.

CHAPTER 5

Summary

This research represents the first known idealized modeling study to investigate the influence of cloud-radiative processes on the development of PREs. A series of numerical simulations is conducted with an aquaplanet version of the WRF-ARW model v.3.6 using a TC bogus and balanced upper-level straight jet. In the control simulation retaining cloud-radiative forcing (CRF-on), a realistic-appearing PRE forms just south of the jet axis, well in advance of the TC. Comparison with a CRF-off run reveals that CRF effects generate a more robust PRE structure with stronger convective activity and, ultimately, more precipitation. CRF produces LW cooling along the cloud tops of the PRE structure, in-cloud (primarily LW) warming in the mid-levels, and cooling above the low-level clouds to the north and south of the PRE. In contrast, the CRF-off experiment only possesses the rather smaller net clear sky cooling. The CRF-on storm's greater convective activity is revealed via difference fields and is also manifested as stronger boundary layer winds and greater low-level wind convergence immediately south of the PRE location.

Experiments separating the warming and cooling CRF components reveal that *LW cooling* is largely responsible for the enhanced convergence and convection. This result contrasts with the Bu et al. (2014) study, which showed that only LW warming directly impacted TC structure. In our case, however, only LW cooling associated with the *low clouds* south of the PRE region is relevant. Note that a version of the CRF-on experiment that neglects radiative cooling associated with hydrometeors in the lowest 1 km generates much less precipitation. This surprising result transpires because the LW cooling associated with shallow clouds raises the surface pressure, thereby increasing the SLP gradient across the PRE region. This key process establishes stronger low-level convergence and leads to more vigorous convection in

the PRE. Differences among the radiation schemes lead to changes in the SLP gradient, low-level convergence across the PRE region, and the resulting PRE rainfall distribution. Operationally, these variations can have significant implications when forecasting the onset, duration, intensity, and even location of PRE development in an attempt to minimize the impact of heavy rainfall and severe flooding.

CRF effects associated with cloud ice and cloud snow in the middle and upper levels of the PRE structure are responsible for driving stronger PRE intensification. In-cloud warming enhances parcel buoyancy and reduces vertical stabilities within the PRE structure, leading to stronger PRE intensification; cloud top cooling in the upper levels of the PRE structure induces downward vertical motions (subsidence) which hinders convective processes and results in weaker PRE development. Using the WSM5 microphysics scheme, the interaction between radiation and cloud ice contributes more to PRE intensification than cloud snow. The location of cloud ice in the upper half of the PRE structure results in a greater vertical depth of in-cloud warming favorable for enhanced upward motions, increased MP latent heat release, and more rainfall production. In contrast, cloud snow is generally concentrated in the middle levels of the PRE structure, limiting the vertical extent of the in-cloud warming. The location of the cloud snow also produces a region of LW cooling at a lower height, which hinders convective processes more than the higher altitude cloud top cooling associated with cloud ice.

Even though LW CRF effects are primarily responsible for PRE initiation and intensification processes, SW CRF effects also have a non-negligible influence on PRE development. In-cloud SW cooling plays the largest role in weakening PRE development by enhancing the stability within the PRE storm structure and inducing local subsidence. In addition, cloud top SW warming along the low-level clouds weakens the magnitude of the LW cooling-induced pressure gradient responsible for enhancing convergence across the PRE region. The negative influence of SW CRF effects on PRE development also suggests a sensitivity to the timing of the diurnal cycle. PRE intensification is stronger when PRE initiation occurs at night and weaker when initiation occurs during the day. The strong influence of SW CRF also causes PRE development to be sensitive to seasonal length of day differences. All other

factors being equal, the summer season produces a weaker PRE; shorter nights minimize the favorable influence of LW CRF effects and longer days maximize the unfavorable impact of SW CRF effects. During the winter season, these trends are reversed and result in stronger PRE development. The seasonal sensitivity has implications when forecasting the formation of PREs during the North American hurricane season, extending from June to November.

This study extends the understanding of CRF effects on weather phenomena beyond the outer core of the TC and demonstrates a far-reaching and direct influence on the development of PREs over a 1000 km away. The next logical step in this research is to obtain observations of the CRF heating and cooling field in order to verify the calculations of CRF effects in numerical weather models between different radiation and microphysics schemes. Additional topics of future study include investigating the specific influence of the MP scheme and other model physics parameterization schemes (e.g. cumulus, boundary layer, surface layer, etc.) on PRE development, and determining an optimized combination or “suite” of physics schemes in the WRF-ARW model that can be applied to weather forecasting operations. Lastly, exploring the influence of terrain effects with a non-aquaplanet version of the idealized experiment would be more applicable, since most PREs and severe flooding impacts occur over land.

Overall, accurately representing the influence of LW and SW CRF effects is necessary to improve forecasts of heavy rain events associated with recurving tropical cyclones. The high sensitivity of PRE development to variations in CRF effects cannot be ignored and necessitates further research to improve operational forecasts. Most importantly, direct observations of the CRF heating and cooling field are required to verify that model calculations are accurate. It is important to note that the influence of CRF also applies to other types and scales of weather phenomena (e.g. sea breeze thunderstorms, fog development, monsoons), as well as long-term global and regional climate studies. Because the recipe for PRE development requires many individual ingredients to come together in just the right way, accurately predicting the onset, intensity, and duration of heavy rainfall events associated with tropical cyclones remains a daunting forecast challenge.

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