On Shallow Cumulus Cold Pools

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ON SHALLOW CUMULUS COLD POOLS

By
Zhujun Li

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ON SHALLOW CUMULUS COLD POOLS

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Observations of organized precipitating trade-wind cumuli show convective invigoration at their cold pool boundaries. Convection and cold pools are studied using nested-WRF simulations of a day from the Rain in Cumulus over the Ocean experiment. The sub-cloud updrafts downwind and near the cold pool boundaries are statistically compared to updrafts not associated with cold pools. Cold pools affect cloud growth through altering the thermodynamic properties of the updrafts to include moister, cooler air and by enhancing the updraft speed; the latter is particularly important for encouraging deeper, high-liquid water path clouds.

It is examined in this study how different microphysics schemes affect the characteristics of simulated trade-wind cumulus convection and cold pools. The initial simulation is compared to an additional nested-WRF simulation using a different microphysics scheme. The differences in autoconversion and accretion rate lead to different precipitation production. The parameterization of raindrop terminal fall speed causes the discrete evaporation efficiencies and different surface cold pool properties under the same surface rainrate.
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Chapter 1

Introduction

Precipitation radar observations, both from space (Short and Nakamura, 2000) and collected at the surface during the Rain in Cumulus over the Ocean (RICO) experiment (Rauber et al., 2007; Snodgrass et al., 2009; Nuijens et al., 2009), reveal rainrates exceeding $2 \text{ mm hr}^{-1}$ for shallow precipitating cumuli in the Caribbean trade-wind region. Ship-based observations have associated such precipitation with a decrease in surface specific humidity, equivalent potential temperature, and increase in wind speed and subsequently the surface fluxes (Fig. 1.1), confirming the presence of large (40-60 km in diameter) cold pools visible from space as mesoscale arcs organized around cloud free areas (Fig. 1.2a). Unlike light ($\approx 1 \text{ mm day}^{-1}$) precipitation that generally cools and moistens the well-mixed sub-cloud layer while keeping the equivalent potential temperature ($\theta_e$) constant (Nitta and Esbensen, 1974; Albrecht, 1993), the heavier shallow precipitation appears capable of gener-
ating convective downdrafts that originate above cloud base and descend into the sub-cloud layer (Zuidema et al., 2012), similar to what has been observed for deeper convection (e.g., Zipser, 1969; Barnes and Garstang, 1982). If the downdrafts are able to impact the surface air, a pool of evaporatively-cooled air that is also drier than the surrounding surface environmental air is formed. Analysis of radar reflectivities collected during RICO find that the leading edges of precipitating shallow cumuli propagate at speeds higher than the mean low level wind speeds, and similar to the estimated propagation speed of the associated cold pool outflow (Zuidema et al., 2012). Cold pools are therefore considered to be responsible for the observed arc-shaped organizations of precipitating shallow cumuli, with the cold pools invigorating convection at their downwind boundary and suppressing thermals inside the stable cold pool area.

Cold pools occurring in the trade wind regions are worthy of research interest, not only for their participation within the global hydrological cycle, but also because of their relationship to the global radiative budget through their influence on cloud fraction. Cold pools influence cloud fraction both by representing areas of low surface buoyancy that discourage bottom-up shallow convection, and by encouraging secondary convection at the cold pool boundaries. An earlier study on feedbacks of shallow cloud precipitation upon cloud cover neglecting cold pools suggests light
Figure 1.1: Compare characteristics within and outside the cold pools identified from ship observation: (a) Maximum surface air specific humidity decrease, (b) maximum equivalent potential temperature decrease, and (c) maximum wind speed increase, all as a function of the maximum temperature drop, for all 37 rain events, with the 17 and 20 cases from undisturbed and disturbed days shown as filled and unfilled circles, respectively. (d) Changes in Bowen ratio within and outside the cold pools as a function of the change in latent and sensible heat fluxes (filled circles and open triangles, respectively). The Bowen ratio is the ratio of sensible to latent heat fluxes.
precipitation should reduce cloud cover by removing liquid water (Albrecht, 1993). Results from an ensemble of high-resolution simulations of precipitating cumuli, based on RICO soundings, indeed show a slight trend of decreasing shallow cloud cover with increase of the precipitation rate (vanZanten et al., 2011). RICO analyses of scanning surface-based radar data, however, find a slight positive correlation (Nuijens et al., 2009). The discrepancy can be explained by the shearing off of the upper portions of the more heavily precipitating clouds (4-5 km, Fig. 1.3), compared
Figure 1.3: Ceilometer-derived cloud-base heights (both the lowest and second-lowest cloud bases are shown if available) as a function of the LCL. Gray shading indicates the cloud base fraction contained within a 100 m × 100 m box. (left) Cloud fraction at each level at 100-m vertical resolution (black line; first cloud-base heights only), and the cumulative cloud fraction (blue line). Data from 9, 10, and 13 January were excluded. The ceilometer-derived cloud fractions are higher than those reported using aircraft aerosol lidar data (Nuijens et al. 2009; Medeiros et al. 2010), maybe because of different instrument thresholds or because near-surface aircraft lidar returns were excluded.
to the modeled cloud top heights of 2-3 km in vanZanten et al. (2011). More variation in the upper levels of shallow cumulus cloud cover rather than lower levels has also been noted in a longer time series at Barbados (Nuijens et al., 2013). The deeper clouds that are associated with cold pools suggest sub-cloud processes at the cold pool-precipitation interface, and the cold pool related processes are critical for understanding how the necessary cloud depth is achieved that can allow overall cloud cover to vary.

Cold pools are also becoming of increasing interest to the large-scale modeling community because their parameterization is suggesting new solutions to old modeling problems. For example, cold pools plays an important role in organizing convection according to the scale of the convection, provides an approach for overcoming inadequacies in entrainment assumptions (Mapes and Neale, 2011). The parameterization of cold pools is also one approach for improving the representation of the diurnal cycle in precipitation within large-scale models (Rio et al., 2009). Such parameterizations can build on the increasing ability to pursue model simulations possessing both the necessary resolution and domain size to represent cold pool processes (e.g., Seifert and Heus, 2013). By comparing high-resolution simulations to observations at similar process-level scales, as will be done here, greater confidence can be gained in the high-resolution simulations themselves.
Several mechanisms have been proposed to explain cold pool invigoration of further convection, both for shallow and deep clouds. One purely thermodynamic mechanism is highlighted within a cloud-resolving modeling study of deep convection, with early-stage sub-cloud evaporation providing the surrounding air with moisture that then enhances the convective available potential energy at the boundaries and thereby encourages convection without any other forcing (Tompkins, 2001). This mechanism for the mesoscale organization of shallow precipitating cumuli is further supported in the large-eddy simulations of Seifert and Heus (2013), with moist rings generated from the evaporation of rain around the cold pool boundaries.

The density-driven outflows can also contribute dynamically to encourage the convergence of moisture at the cold pool boundary, as mentioned in the large eddy simulation (LES) study on trade wind cumuli mesoscale organization by Xue et al. (2008). Dynamic lifting by the denser cold pool air occurs as the cold pool air subsides and spreads into the environment at the speed of gravity current, with a propagation speed close to the surface exceeding that of the mean wind. The propagation causes convergence at the leading edge that can forcibly lift air parcels. This can strengthen already-buoyant updrafts, or transport humidity to upper levels, preparing for future convection. In addition, dynamic lifting at the cold pool bound-
Figure 1.4: For 1540-1600 UTC January 19, 2005: a) the ship-board X-band radar signal-to-noise ratio (grey shaded), with lidar-observed positive (red arrow) and negative (blue arrow) vertical velocities superimposed. Filled yellow triangle indicates the downwind side surface cold pool boundary identified by the onset of surface rain. The ship observed b) surface rainrate, c) surface air temperature, and d) surface water vapor mixing ratio.
ary, by producing stronger and wider updrafts, can increase their available kinetic energy and ability to trigger further convection. The influence of such processes on deep convection has been explored as a parameterization within Rio et al. (2009); Grandpeix and Lafore (2010); Rio et al. (2013). The processes themselves have not yet been investigated in detail for shallow convection.

The influence of the typical wind shear pattern of the trade winds on cold pool induced updrafts should also be considered. Studies on mid-latitude squall lines have noticed that the lifting of updrafts due to cold pool propagation is strongest when the local circulation associated with the vertical variation of the cold pool boundary propagation is balanced by the circulation of low level environmental wind shear (Rotunno et al., 1988; Moncrieff and Liu, 1999; Weisman and Rotunno, 2004). When the environmental wind shear exceeds the vertical gradient in the cold pool propagation speed, the updraft tilts down shear of the environmental wind, and when the circulation within the cold pool boundary dominates, the updraft leans against the cold pool boundary. In the RICO cases, although the shallow cumulus cold pools are much weaker than the mid-latitude deep convection, the influence of shear on the lifting process should remain similar.

It is a challenge to disentangle the co-existing thermodynamic and dynamic aspects of cold pool effects, but proper attribution of causes holds the promise of im-
proving those convective parameterizations that can be explicitly formulated to depend on cold pool processes (e.g., Bretherton et al., 2004; Mapes and Neale, 2011; Rio et al., 2013). Simulation data provide four-dimensional fields that are more comprehensive for understanding the relevant processes, than the two-dimensional (time-height) column data fields acquired observationally from a ship. An example can be made using the shipboard data from a motion-compensated 10 – µm wavelength Doppler lidar, a vertically pointing X-band precipitation radar, and 15-m flux tower meteorological data across the leading edge of a raining cold pool boundary (Fig. 1.4). The updrafts are located on the downwind side of the major precipitation, and are associated with newly developed clouds, while the downdrafts that produce cooling and drying at the surface are associated with the rain shaft of the convection. The onset of surface precipitation, shown as the yellow triangle, coincides with the largest drop in surface temperature and water vapor. The in-situ observations are intriguing, but lacking Lagrangian tracking and the full three-dimensional temperature, water vapor, velocity, and precipitation fields, do not provide enough information for distinguishing between thermodynamic and dynamic mechanisms linking cold pools and convection. Such intriguing but inadequate pieces of information characterize field observations of spatially-inhomogeneous time-varying phenomena, reinforcing the need for complementary modeling studies.
Figure 1.5: The average profile of a) potential temperature, b) water vapor mixing ratio, c) zonal and d) meridional wind speed for the innermost model domain averaged over the 24-hour simulation period (black line; WRF); the average of six radiosondes launched from 0700 UTC January 19, to 0300 UTC January 20 from the ship (red line; 19RVSJ); all of the radiosondes launched from the ship January 9-24 (green dotted line; allRVSJ); and all radiosondes launched from Spanish Point from December 16, 2004 through January 8, 2005 (blue dotted line; SPNT).

Therefore, this study analyzes realistic, high-resolution model simulation to assess the processes by which shallow cumulus cold pools invigorate further shallow precipitation, motivated by observations of trade wind cumulus cold pools during the RICO experiment (Zuidema et al., 2012). The day of January 19, 2005 is selected for the simulation, in part because the day has been previously studied (Abel and Shipway, 2007; Snodgrass et al., 2009). Conditions on January 19, reflect the influence of a dissipated cold front (Caesar, 2005), with northeasterly winds rather than the easterly trades. The strong boundary layer northeasterly wind can explain the observed linear cloud lines (Fig. 1.2b), though no obvious gradient in tempera-
Figure 1.6: Shipboard measurements during 1200-1800UTC19 January: (a) surface air (black) and sea surface (blue) temperatures and accumulation mode aerosol concentration (green), (b) surface RH(black) and the LCL (blue), (c) rain rate (black, nonzero values highlighted in red) and LWP (blue), (d) surface air specific humidity (black, dashed line indicates mean value of 15.2 g kg\(^{-1}\)) and microwave-derived water vapor path (blue), (e) \(\theta_e\) (black, mean 342.1 K) and \(\theta_v\) (blue, mean 300.2 K), (f) latent (black, mean 111 W m\(^{-2}\)) and sensible (blue, mean 13 W m\(^{-2}\)) heat fluxes, and (g) zonal (red), meridional (blue), and absolute (black, mean 5.9 m s\(^{-1}\)) wind speed.
ture or moisture is documented in the surface meteorological measurements or the ship soundings. The observed liquid-phase-only precipitation (Fig. 1.4) is identified as "undisturbed" by Nuijens et al. (2009), meaning that the fractional area covered by precipitation deviates less than three standard deviations from the mean of RICO operating period (from November 24, 2004 to January 25, 2005). The influence of a dissipating cold front arriving from the north is likely a common occurrence for Caribbean trade-wind cumuli during the boreal winter, so that January 19, may represent one of several typical conditions for trade-wind cumulus. Cold pools generated by shallow precipitation are observed during this day (Fig. 1.6), with the rapid drop of surface air temperature and specific humidity earlier in the time series (the downwind side of the cold pool), corresponding to the onset of surface rain. The surface fluxes change with the variation of either surface wind direction or speed.

The simulation uses the Weather Research and Forecasting Model (WRF) (updated version 3.2) with multiple nested domains, a modeling setup that has been previously successfully applied to study continental stratocumulus (Zhu et al., 2010). The nesting technique allows the model to explicitly resolve turbulent scale processes in the innermost domain of 24 km by 24 km at 100 m resolution, while the open lateral boundary conditions allow a realistic large-scale forcing imposed
on the parent domain of size 972 km by 972 km to transmit to the innermost do-
main. As will be shown, by incorporating a sensitivity to the realistic, spatially-
inhomogeneous, time-varying large-scale forcing, cold pools can develop within the
relatively small inner domain of the nested-WRF simulation.

The ability of this simulation to model cold pools can be compared to the experi-
ence with large-eddy-scale simulations that typically apply doubly-periodic bound-
ary conditions, idealized initial conditions, and prescribed homogeneous large-scale
forcing (e.g., Matheou et al., 2011; Seifert and Heus, 2013). Such simulations as-
sume that the turbulent atmosphere at the inflow boundary is identical to that at
the outflow boundary, an assumption that is more applicable to the relatively homo-
geneous conditions of marine stratocumulus or non-precipitating shallow cumulus,
than to strongly precipitating shallow cumulus conditions. When nevertheless large-
eddy-scale simulations of strongly precipitating shallow cumulus are undertaken, a
sufficiently large domain and sufficiently high resolution of the turbulence processes
is required to allow such spatially inhomogeneous features as the asymmetric trade-
wind cold pools to develop. Indeed, Matheou et al. (2011) are not able to reproduce
precipitating mesoscale arcs until a resolution of 20 m is achieved within a compa-
rable domain size of 20 km by 20 km. Seifert and Heus (2013) also find a significant
time delay for the development of a cold-pool dominated regime within their larger
domain size of 50 km by 50 km at larger grid spacings (50 m and 100 m) compared to a grid spacing of 25 m. Our one-day hindcast simulation produces convection and cold pools that can be compared to the observations from this particular day, developing an understanding of the modeling strengths and weaknesses for non-homogeneous conditions.

Recent studies attempt to better understand the processes related to shallow cumulus cold pools through fine resolution ($dx = dy = 25 \sim 100$ m) model simulations, such as the mesoscale cloud organization due to cold pools (Seifert and Heus, 2013), and the aerosol effects on the organization of shallow cumulus convection through cold pools (Xue et al., 2008). Seifert and Heus (2013) and Xue et al. (2008) report a sensitivity of the cold pool mesoscale organization to imposed cloud drop number or aerosol number concentration, as well as domain size and grid spacing.

The parameterization of rain microphysics should influence the mesoscale organization of trade-wind cumulus through the impact of rain evaporation on the cold pool properties. Stevens and Seifert (2008) examine the sensitivity of simulations of shallow cumulus convection to bulk microphysics schemes, but have not examined rain evaporation and cold pools. Abel and Shipway (2007) improve the ability of model to capture trade-wind cumulus convection through facilitating the autoconversion process, and also do not focus on the relationship to cold pools. Shipway and
Hill (2012) use a one-dimensional (1D) kinematic framework to examine differences in warm rain microphysics schemes including rain evaporation, but microphysical-dynamical feedbacks are explicitly neglected by construction, thus no comparisons to observations could be made. The relationship between microphysical parameterizations and cold pools has been assessed for deeper convection, such as mid-latitude squall lines (Morrison and Milbrandt, 2011, and references therein), but not yet for cold pools produced by only warm rain.

In Chapter 3, I evaluate the impact of choice of microphysics scheme upon the simulation of trade-wind cumulus convection and precipitation-induced cold pools. This builds on the Chapter 2 (Li et al., 2014), in which a nested-Weather Research and Forecasting (WRF) simulation using a double-moment Thompson microphysics scheme is analyzed. The analyzed case of 19 January 2005 reflects the influence of a dissipated cold front (Caesar, 2005), with northeasterly winds rather than the easterly trades, containing cold pools embedded within more synoptically-defined cloud lines. The simulation captures cold pools that are comparable to the observation if weaker and smaller than the observed cold pools. I focus here on a further assessment of impact of the microphysics, by analyzing a second nested-WRF simulation that has the exact same setup as that discussed within Chapter 2, but with the Thompson microphysics scheme replaced by the double-moment Morrison
scheme (Morrison et al., 2005). This allows us to explore the relative strengths of a one-moment versus two-moment scheme for representing trade-wind cumulus rain evaporation in a three-dimensional (3D) setting in which the mesoscale organization may be assessed against observations.

In microphysics schemes, the conversion of water from cloud category to rain category by collision and coalescence is represented by the autoconversion parameterization, the raindrops then grow by collecting cloud drops (accretion) and raindrops (self-collection). The parameterization of autoconversion process in the Thompson scheme follows Berry and Reinhardt (1974), and Morrison scheme follows Beheng (1994). Shipway and Hill (2012) concludes that the different treatments of autoconversion process in the two schemes lead to the differences in the onset of rain and the rain mass production. These differences are important to the study of cold pools and associated mesoscale cloud feature because the cold pool size and strength is determined by the rain intensity.

Under similar rain intensity, greater evaporation rate may result colder cold pools. The factors that control rain evaporation processes are implemented in the parameterizations, including rainrate, drop size spectrum, turbulent effect, and environmental air temperature and humidity. For a isolated moving drop with mass $m$ and diameter $D$, the evaporation rate can be solved from the steady state convective
diffusion equation as:

\[
\frac{dm}{dt} = 2\pi D_d f_v (\rho_\infty - \rho_D) \tag{1}
\]

Where \(d_v\) is the diffusion coefficient for water vapor, \(f_v\) is the mean ventilation coefficient that parameterized as a function of Reynolds number, \(\rho_\infty\) and \(\rho_D\) are the water vapor density in the environment and at the drop surface respectively (Pruppacher and Klett, 1997). In parameterizations, this balance is solved numerically in order to be implemented in models. The Morrison scheme follow the solution in Rutledge and Hobbs (1983), and the Thompson scheme includes several higher order term of the supersaturation to minimize the error of the solution (Srivastava and Coen, 1992).

For the same rain mass, a distribution that has more smaller drops with higher total drop number concentration and lower mean drop size, would result in higher evaporation efficiency, since smaller drops evaporate more readily. In both the Thompson and Morrison schemes, the raindrop size distribution (RSD) is assumed to follow the Marshall-Palmer distribution: \(N(D) = N_0 e^{-\lambda D}\), where \(N(D) (m^{-4})\) represents the number of rain drops of diameter \(D (m)\), \(N_0 (m^{-4})\) is the intercept parameter, and \(\lambda (m^{-1})\) is the slope parameter. Both double-moment schemes predict the total number of rain drops along with rain mixing ratio, from which \(N_0\) and \(\lambda\) are diagnosed. The rain number concentration is alternated during the fall
of raindrops by the processes of raindrop collision, coalescence and breakup. The double moment schemes allow larger drops to fall faster by gravitational sorting. However, double moment schemes with exponential drop size distributions may overestimate the sedimentation of large drops, causing high number concentration at lower levels and steep surface rainrate peak (Wacker and Seifert, 2001; Milbrandt and McTaggart-Cowan, 2010). The Morrison scheme use implicit parameterization for raindrop breakup to reduce the production of unrealistic large mean drop size due to size sorting (Morrison et al., 2009)

After comparing the two three-dimensional nested-WRF simulations, the two microphysics schemes are also evaluated within the one-dimensional Kinematic Driver Model described within Shipway and Hill (2012), with prescribed realistic updraft velocities. This serves to confirm the dominant role of microphysics in WRF-simulated rain evaporation, by excluding influences from environmental conditions and turbulent mixing.

Shallow cumulus cold pools not only commonly observed in trade-wind region, but also over tropical ocean. The results of this thesis would be better interpreted if they are not regionally dependent, since one of the goals of this thesis work is to provide reference for climate models to better represent shallow precipitation processes in general. This calls for comparisons of trade-wind region shallow cumu-
lus cold pools with other regions. The Dynamics of the Madden-Julian oscillation (MJO) campaign (DYNAMO) (Yoneyama et al., 2013) is conducted over the Indian Ocean for an intensive observing period (IOP) from October 2011 to January 2012. Analysis on satellite data during the DYNAMO period indicates that the shallow convective rain contributes significantly to the total rain in the observation region, reaching its peak value before the intense rain event (Zuluaga and Houze, 2013). In Chapter 4, cold pools documented during this field campaign are compared to RICO cold pools, with the emphasis on cold pool characteristics and their influence on convection.
Chapter 2

Simulated convective invigoration processes at trade-wind cumulus cold pool boundaries

2.1 Characteristics of simulated convection and cold pools

WRF simulation setup

The January 19 case is simulated from 0000 UTC January 19, 2005, to 0600 UTC 20 January 2005, with only the last 24 hours analyzed. National Centers for Environmental Prediction (NCEP) (Final) Operational Global Analyses data (FNL) at 1 degree resolution, available at every 6-hour, supply the initial and lateral boundary conditions. The sea surface temperature is prescribed from the NCEP FNL and updates every 6 hours. One parent domain (972 km by 972 km) and four two-way nested domains centered at 61.7°W 18°N are configured with nesting ratio of 1:3 (grid spacing of each outer domain is 3 times the grid spacing of its next level
Figure 2.1: A snapshot of the innermost domain from the simulation, with the size and location as indicated by the pink square in Fig. 1.2b. a) WRF simulated 3-m level temperature (grey shaded) overlaid with the vertically integrated cloud water path (0-2 cm; white contours). b) WRF simulated 3-m level water vapor mixing ratio overlaid with 3-m level wind vectors.

The innermost domain covers the track of the Research Vessel Seaward Johnson (RVSJ), possessing a size of 24 km by 24 km with a horizontal spacing of 100 m. The vertical domain extends from the surface to 10 hPa with in total 77 levels, of which 48 levels are below 4 km with a vertical spacing varying from 6.5 m to 200 m. I refer to the 3-m first model level as the surface level of this simulation. The 3D Smagorinsky scheme (Smagorinsky, 1963) treats the sub-grid scale turbulent mixing for domains with 100 m and 300 m grid spacing, and the Mellor-Yamada-Janjic boundary layer scheme (Janjic, 2001) is used for treating the vertical
turbulent mixing for the coarse domains. The simulation uses the double-moment Thompson cloud microphysics scheme (Shipway and Hill, 2012) (the known bug of WRF v3.2 Thompson scheme is fixed with the updated code), with a total cloud droplet number concentration $N_c = 100 \times 10^6 \, m^{-3}$. Each of the 1441 one-minute simulation outputs from the innermost domain is treated as an individual sample for the statistical analyses.

Since the vertical resolution of the NCEP FNL analysis does not resolve the fine vertical structure of cumulus topped marine boundary layer, the thermodynamic profiles from radiosondes launched from the RVSJ every six hours are assimilated and nudged in the outer domains. Fig. 1.5 shows the average of the ship soundings from January 19 (19RVSJ) and the average of the profiles from the innermost model domain (19WRF). These are compared with the average of all soundings launched from RVSJ between 9-24 January (allRVSJ), and the average of the two to six-times-daily soundings from Spanish Point, Barbuda, between December 16, 2004 through 8 January 2005 (SPNT). Spanish Point is located to the south-south-west of the ship. One of the ship radiosondes launched on January 19 penetrates a cloud that reaches above 4 km, causing the 19RVSJ water vapor mixing ratio above 3 km to be significantly higher than the simulated domain-average 19WRF values. The composite ship sounding of January 19 also reveals a more moist atmosphere column above 1
km than the average of all soundings from the ship, indicating the additional moisture available for this day. The most significant difference is between the January 19 wind profiles and those of the four-week-mean SPNT profile and two-week January allRVSJ mean, with the allRVSJ wind profile more sheared than the SPNT profile. The winds are weaker on January 19 compared to allRVSJ, with the wind shear at 2.5-3 km. RICO-mean conditions at Spanish Point, Barbuda, show more easterly winds above 2 km, and drier and warmer air above 2.5 km compared to the January-mean ship profile. The simulations described in vanZanten et al. (2011) are initiated by the mean soundings from SPNT, producing cloud tops that extend up to 2.5 km, with weak precipitation and no cold pools.

Comparisons to observed convection and cold pool air properties

The simulated surface air temperature reveals cold pools embedded within the larger-scale defined cloud line (Fig. 2.1a). Three shallow precipitating convection cases along with associated surface cold pools are documented within these cloud lines by the RVSJ flux tower. The flux tower measurements are gathered \( \sim 15 \) m above sea level with more detailed information regarding the instruments and data processing available in the Appendix of Zuidema et al. (2012). The average and standard deviation of the 24-hour flux tower measurements are compared to the 24-hour simulation data from all grid points for air temperature, water vapor mix-
Table 2.1: Mean and standard deviation ($\sigma$) of the 13-m level simulated temperature, water vapor mixing ratio, $T_v$, $\theta_e$, wind speed, sensible and latent heat fluxes are compared to the corresponding ship measurements during RICO.

<table>
<thead>
<tr>
<th></th>
<th>$T$ (K)</th>
<th>$q_v$ (g kg$^{-1}$)</th>
<th>$T_v$ (K)</th>
<th>$\theta_e$ (K)</th>
<th>Wind speed ($m s^{-1}$)</th>
<th>SHF ($W m^{-2}$)</th>
<th>LHF ($W m^{-2}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>WRF mean</td>
<td>298.0</td>
<td>15.6</td>
<td>300.8</td>
<td>343.5</td>
<td>5.6</td>
<td>8.9</td>
<td>108.7</td>
</tr>
<tr>
<td>Obv Mean</td>
<td>298.0</td>
<td>15.6</td>
<td>300.8</td>
<td>343.5</td>
<td>5.9</td>
<td>7.6</td>
<td>103.2</td>
</tr>
<tr>
<td>WRF $\sigma$</td>
<td>0.3</td>
<td>0.5</td>
<td>0.3</td>
<td>1.4</td>
<td>2.5</td>
<td>4.3</td>
<td>37.1</td>
</tr>
<tr>
<td>Obv $\sigma$</td>
<td>0.5</td>
<td>0.3</td>
<td>0.6</td>
<td>1.4</td>
<td>1.3</td>
<td>4.7</td>
<td>20.6</td>
</tr>
</tbody>
</table>

Temperature and wind speed at 13-m level, as well as surface rainrate, sensible heat flux (SHF) and latent heat flux (LHF) (Table 2.1). The mean simulated air temperatures and water vapor mixing ratio match the average ship-board values well. The ship mean wind speeds are slightly higher than those of the WRF domain, yet the ship mean turbulent fluxes are less, implying higher transfer coefficients are used within the model. The relatively higher variability of the simulated wind speed is thought to reflect its variability across the domain, whereas the ship mostly remained within the linear cloud line, reporting mostly north-northeasterly winds (e.g., Fig. 1.2b, with a wind speed of 4.7 $m s^{-1}$ at 1606 UTC).

The 24-hour domain-averaged simulated surface rainfall rate is 2.1 mm day$^{-1}$, relatively close to the daily area-averaged rainfall rate 1.87 mm day$^{-1}$ derived from scanning precipitation radar reflectivities for this day (Snodgrass et al., 2009).
Figure 2.2: a) Probability density function of $RR_a$ with the dashed line indicating the threshold for 10% highest $RR_a$. b) Cloud cover as a function of $RR_a$ with the shading indicating the number density of output minutes. Dashed line indicates the mean cloud cover of the 24-hour period. Two dash-dotted lines extending to the right and left of the top 10% $RR_a$ threshold indicate the averaged cloud cover over the 10% output minutes with highest $RR_a$ and the other output minutes, respectively.
The domain-averaged non-zero surface rainrate, or $RR_a$, is calculated for each output minute by averaging the surface rainrate of all grid points with rainrates $\geq 0.1 \ mm \ hr^{-1}$, and $RR_a$ is set to zero when no such rainrates are present. $RR_a$ ranges up to $10 \ mm \ hr^{-1}$, with 95% of the values exceeding 0 and below $6 \ mm \ hr^{-1}$ (Fig. 2.2a). Rainier time periods were also cloudier, as shown in Fig. 2.2b. The cloud cover is defined as the fraction of grid columns that contain cloud at any level, and the mean cloud cover of all 1441 output minutes is about 0.13. This is comparable to the cloud cover reported by Matheou et al. (2011) using a grid spacing of 40 m. The average cloud cover of the top tenth $RR_a$ percentile is 0.05 higher than the average of the remaining output minutes (Fig. 2.2b). The cloud cover increases with higher $RR_a$ for a correlation coefficient of 0.47.

The simulated surface temperature and cloud water path in Fig. 2.1a reveal the circular cold pools embedded within the linearly-oriented cloud lines. The colder regions mostly correspond to drier air, as seen by the anomalous water vapor mixing ratio (Fig. 2.1b). The apparent change in surface wind can be spotted at some of these cold and dry boundaries (Fig. 2.1b). The simulated clouds also extend above 3 km (Fig. 2.3), similar to the observations (Fig. 1.4). For output minutes with $RR_a > 0$, the averaged vertical profile of cloud fraction shows three local peaks: at the cloud base (500 m), 1.6 km, and the level right below 3 km (Fig. 2.3a). The
Figure 2.3: For output minutes containing $RR_a > 0$, with one standard deviation shaded: a) Domain-mean cloud fraction (solid), with the mean buoyant cloud core fraction profile indicated separated (triangles); b) horizontally-averaged cloud water mixing ratio over cloudy region ($q_c$); c) horizontally-averaged water vapor mixing ratio over cloudy region ($q_v$); and d) horizontally-averaged rain water mixing ratio within rain shaft ($q_r$).

Two upper peaks coincide with layers of slight enhanced wind shear (Figs. 1.5c-d).

The buoyant cloud core fraction, defined as the portion of cloudy grid cells ($q_c > 0.1 \, g \, kg^{-1}$) that are positively buoyant relative to the domain-mean, increasingly deviate from the averaged cloud fraction up to 1.6 km (Fig. 2.3a) as environmental air is mixed into the cloud. The clouds that do reach up to 2 km provide the peak at 2 km in the averaged cloud and rain water mixing ratio (Fig. 2.3b and d), with drier conditions aloft.

One assessment of the simulated cold pools is shown in Fig. 2.4 as a composite of all 1441 output minutes centered upon each output’s most intense instant rain center, and the same composite for only those output minutes associated with the upper tenth percentile of $RR_a$. For each of these output minutes, I first locate the grid...
Figure 2.4: Composite of 13-m level air properties across the domain-maximum surface rainrate (0 on x-axis) from all 1441 simulation output minutes (black solid line with filled circle) and the output minutes of highest 10% $RR_a$ values (black dotted line), as compared to the observed flux tower measurements at 15 m averaged over the passages of the three cold pools documented on January 19 (red line with filled circle) of a) rainrate, b) wind speed, c) temperature, d) water vapor mixing ratio, e) virtual temperature, f) equivalent potential temperature, g) sensible heat flux and h) latent heat flux.
Figure 2.5: The maximum surface changes within individual cold pools, observed on January 19 (red filled circle) and other “undisturbed” RICO days (pink filled circle), and as contoured frequency distribution for the maximum anomalies within individual simulated cold pools represented by the composite of Fig. 2.4, for a) $q_v$, b) $\theta_e$, and c) wind speed, all as function of the change in $\theta$. d) The change in LHF as a function of SHF change using the same plotting conventions.

point containing the domain-maximum surface rainrate, then consecutively sample the nearby grid points that align with the domain-averaged surface wind vector within the same moment in time, and average these surface air properties from the selected output minutes. The negative distance on the x-axis corresponds to the downwind (southwest) side and positive distance to the upwind (northeast) side of the domain-maximum surface rainrate. Fig. 2.4 also shows the average of the three observed cold pool passages from this day. These correspond to maximum surface rainrates of 7.4, 7.5 and 45 mm hr$^{-1}$. The ship moves in the same direction as the surface wind during the three cold pool passages, meaning that the ship samples its cold pool for longer than if it was stationary.

Fig. 2.4 clearly shows that the simulated cold pools are, like the observations,
asymmetric along the mean wind direction. The temperature change is steeper on
the downwind side of cold pool, which is where the rain occurs. The wind increases
ahead of the rain, enhancing the surface fluxes, and indicating the cold pool outflow.
The mean rainrates in the simulation match observed values, while the average cold
pool is obviously weaker than observed. The simulated temperature depression is
at best 0.5 K and the change in surface water vapor mixing ratio is negligible. The
cold pools corresponding to the upper 10% of the \( RR_u \) are significantly stronger than
the average, but still weaker than the observations. To some extent the composite
may reflect the process of averaging, but nevertheless the weak simulated cold pools
also reflect on the ability of the simulation to produce cold pools. The January 19
observation captures only the change in wind direction at the cold pool boundary
rather than wind speed (see Fig. 12g in Zuidema et al., 2012), which I attribute to
ship location.

The decrease of the water vapor mixing ratio \( (q_v) \) and \( \theta_e \) with \( \theta \) in simulated cold
pool cases are consistent with the observed cases (Fig. 2.5a-b). The wind changes
in the simulation are stronger for the same change in \( \theta \) than in the available ob-
servations, resulting in slightly larger surface fluxes (Fig. 2.5c-d). The simulated
cold pools may be weaker than those observed in their thermodynamic properties,
but the gustiness changes at the cold pool boundaries and the resulting convergence
Figure 2.6: Snapshots focusing on the 80-m level cold pool downwind boundary (black stippled) from 2105 to 2123 UTC in 6 minutes intervals from a) to d). Plots depict the 80-m level updraft area (red stippled), positive and negative $\Delta q_v$ (light and dark shading respectively), $\Delta \theta_v = 2$ K contours (light green) coincide with $\Delta q_v = 0.7$ g kg$^{-1}$ light shaded contours, and areas with surface rainrates $\geq 2$ mm hr$^{-1}$ outlined in light blue contours. Dashed black lines correspond to the cross sections in Fig. 2.13.

appear adequately captured.

**Terminology and statistics of simulated cold pools**

Since objective of this study is to examine the ability of precipitation-induced cold pools to invigorate further convection at the downwind boundary, a necessary first step is to articulate how arc-shaped cold pool downwind boundaries and their influences are identified within the simulation. Strict criteria are imposed to appropriately identify precipitation-driven cold pool downwind boundaries. First, the cold pools’ characters as density currents are preserved through a criterion based on virtual potential temperature ($\theta_v$). Second, a negative anomaly of $\theta_v$ must be asso-
Associated with the cold pool, and not represent a preexisting condition. The horizontal \( \theta_v \) anomaly is calculated at each individual level below cloud base \( (\Delta \theta_v(x, y, z) = \theta_v(x, y, z) - \overline{\theta_v}(z)) \), the horizontal anomalies of other scalers \( (\Delta \chi) \) are calculated in the same manner. For output minute \( t \), the cold pool downwind boundary grid points must be less buoyant than the domain-mean \( (\Delta \theta_v(x, y, z)|_{t < 0}) \), and, must have been equally or more buoyant than the domain-mean within the previous minute \( (\Delta \theta_v(x, y, z)|_{t-1} \geq 0) \). Third, the grid point wind speed must be higher than the speed of domain-averaged wind \( (U(x, y, z) > \overline{U(z)}) \).

In addition, a connection of cold pool downwind boundary to the precipitation and its catchment area are maintained by only selecting grid cells that already satisfy the conditions above and are in the proximity of significant rain. A grid point is considered near significant rain if the rectangle horizontal area of 6 km by 6 km centered on the grid point contains precipitating points, and the average surface rainrate of these points are greater than 2 mm hr\(^{-1}\). The rainrate threshold, and the size of the precipitation catchment area are reasonable if subjective choices. The 6 km spatial scale estimate is shown adequate to encompass most of the related rain catchment area (including the top 10% strongest rain events) (Fig. 2.4), but still restrictive enough to exclude other irrelevant rainy points. The 2 mm hr\(^{-1}\) catchment-area-average threshold has been identified previously as a way to distinguish cold pools.
that are dried by convective downdrafts from those that are moistened (Barnes and Garstang, 1982). A catchment area criterion also excludes those boundaries of cold pools that are dissipating with the decaying of convection. Based on these conditions, the cold pool downwind boundary can be automatically identified as shown in the example of Fig. 2.6. Overall, the combined criteria provide us with confidence that I am examining the subset of cold pool boundaries that are actively involved with promoting secondary convection. 620 of the 1441 output minutes contain at least one cold pool boundary satisfying this criteria, or, 43% of the total. The depth of the cold pool boundaries ($h_{cpb}$) is estimated by the buoyancy of grid cells within the columns above the 3-m level cold pool boundary points. For each output minute, the $h_{cpb}$ is the level where the number of negative $\Delta \theta_v$ points within the columns becomes less than 10% of the 3-m level cold pool boundary points.

Based on these definitions, 91% of the cold pool boundaries have $h_{cpb}$ more than 80 m (Fig. 2.7), corresponding to 565 output minutes. The negative $\Delta \theta_v$ extends through the whole sub-cloud layer for 31% of the boundaries, indicating deeper cold pools than Seifert and Heus (2013). The length along the 3-m level cold pool downwind boundary arc ($L_{cpb}$) are below 20 km for all applicable output minutes, with 144 output minutes greater than 6.8 km (equivalent to cold pool diameters of $\sim 12.7$ and $\sim 4.3$ km, respectively), the latter corresponding roughly to the scale
Figure 2.7: The accumulated probability density function of length along cold pool boundary arc at the 3-m level (solid black) and 80-m level (dashed black), and the cold pool boundary depth (solid blue).
Figure 2.8: For the 80-m level: a) contoured frequency distributions of the $\Delta q^u_p$ as a function of $\Delta \theta^u$ for CPAR updrafts (red; based on the 560 output minutes) and non-CPAR updrafts (black; based on all the 1441 output minutes). b) Contoured frequency distributions of the $q^u_p$ difference between the CPAR updrafts and the non-CPAR updrafts within the same output minute, as a function of the $\theta^u$ difference, based on 560 output minutes containing CPAR updrafts. c) The probability distribution of $\Delta \theta^u$ for the CPAR updrafts (red; based on the 560 output minutes) and non-CPAR updrafts (black; based on all the 1441 output minutes).

of the highest 10% rain catchment area in Fig. 2.4. The smaller $L_{cpb}$ at the 80-m level relative to the surface level reflects the vertical reduction in cold pool size. Satellite images reveal arc-shaped organization of clouds spanning up to 40~60 km in equivalent diameter during RICO, and the ship-launched soundings identify some cold pools of 200 m in depth (Zuidema et al., 2012).

The difference between the wind averaged within the cold pool downwind boundary and the domain-averaged wind is defined as the expansion rate of cold pool downwind boundary ($C_e$). Its magnitude is similar to the propagation speed of a
Figure 2.9: For the 80-m level: contoured frequency distributions of the $\Delta q_v$ averaged over all CPAR points as a function of a) the $q_v^{\text{up}} - q_v^{\text{np}}$ difference between the CPAR updrafts and non-CPAR updrafts, and, b) the cold pool expansion rate. Both plots based on the 560 output minutes containing CPAR updrafts.

density current, in the way that the expansion rate also depends on the magnitude of negative buoyancy within the cold pools and the cold pool depth (Grandpeix and Lafore, 2010). For output minutes that have multiple cold pools present in the domain, $C_*$ of each output minute is the average rate for all the cold pool downwind boundaries. In the following section, a subscripted number indicates the altitude level at which the $C_*$ is estimated.

The simulated updrafts are identified as points with vertical velocity $w \geq 0.5 \text{ m s}^{-1}$. The updrafts that can potentially be influenced by the cold pool outflow by virtue
of proximity need to be distinguished from the updrafts that occur thermodynamically regardless of the presence of a cold pool. This is done by delineating the area that encounters the propagating cold pool downwind boundary. A “cold pool ambient region” (hereafter CPAR) is defined as the area within 1 km downwind of the cold pool boundary at the 80-m level (with the cold pool boundary not included). Almost all (560 of the 565) output minutes with 80-m level cold pool boundaries contain updrafts within 1 km of the boundary. Another distance choice of 1.5 km does not increase the sample size of updrafts within CPAR significantly, indicating the 1 km distance is capturing most of the updrafts. In contrast, a distance threshold of 0.5 km reduces the sample size significantly. Ultimately, while the choice of a 1 km distance is subjective, the choice is also reasonable. The potential reach of spatially-inhomogeneous cold pool outflow to the updrafts in their vicinity is a statistical correspondence. More distant updrafts may still be influenced by cold pool density currents, and some updrafts downwind of the cold pool will already be thermodynamically buoyant, but as our results will show, a clear statistical signal of the influence of the upwind cold pool can be inferred based on a 1 km choice for the CPAR.
Figure 2.10: For the 80-m level: contoured frequency distributions of the updraft vertical velocity ($w_{\text{up}}$) as a function of the buoyancy ($b_{\text{up}}$) for a) the updrafts, b) buoyant portion of the updrafts, and c) non-buoyant portion of the updrafts, within CPAR (red; based on the 560 output minutes) and outside CPAR (black; based on the 1441 output minutes). Buoyancy of each grid cell is assessed relative to the domain-mean: $b(x, y, z) = g [\Delta \theta(x, y, z)/\bar{\theta}(z) + 0.61 \Delta q_v(x, y, z) - q_e(x, y, z) - q_r(x, y, z)]$, where $g$ is the gravitational acceleration.

2.2 Cold pool effects

The cold pool effects on organizing subsequent precipitating shallow convection are examined in this section. The thermodynamic and dynamic properties of the updrafts within and outside the CPAR are compared to demonstrate the existence of cold pool effects, and the cold pool-affected updrafts are diagnosed thereafter for the mechanisms by which the convection is invigorated.

**Cold pool effects on low-level updrafts**

The 80-m level anomalies of $q_v$, $\theta$ and $\theta_e$ from the domain-mean values are averaged over the updraft points within and outside (the updrafts outside of CPAR include the
updrafts from output minutes lacking any cold pools) the CPAR for each applicable output minute (Fig. 2.8). Not surprisingly, updrafts tend to be moister and warmer than the domain-mean regardless of proximity to the cold pool boundary. The moisture content differs markedly between the two populations. The updrafts influenced by cold pool outflows are generally moister than for the other updrafts (Fig. 2.8a). When only compared to the updrafts within the same output minutes, the CPAR updrafts are moister than the other updrafts by a mean of 0.07 g kg$^{-1}$ and are also slightly warmer (Fig. 2.8b). The impact of both the enhanced moisture and warmth can be seen in the distribution of the updraft $\theta_e$ anomaly values from their domain-mean in Fig. 2.8c.

In Fig. 2.9, the relationships of the overall moisture content of the CPAR relative to the domain-mean with the difference between two populations of updrafts and the cold pool expansion rate are examined. On average, the CPAR updrafts only cover 16% of the CPAR area, CPARs that are uniformly more moist are more capable of supporting individual moister updrafts (Fig. 2.9a). More interesting is the finding that moister CPARs are also associated with faster-expanding cold pools (Fig. 2.9b), possibly due to the more efficient converging of moisture by the cold pool expansion.

Although many of the cold pool influenced updrafts are more buoyant than the other updrafts within the same moment (Fig. 2.8b), enhanced buoyancy is not the
main factor affecting the speed of the CPAR updrafts. This is made clear when the vertical velocities of CPAR and non-CPAR updrafts are compared to their buoyancy, and then each group of updrafts is divided into their buoyant and non-buoyant (relative to the domain-mean) portion (Fig. 2.10). Overall, the CPAR updrafts are capable of attaining higher vertical velocities than the non-CPAR updrafts (Fig. 2.10a). The buoyant portion non-CPAR updraft vertical velocities increase with positive buoyancy (Fig. 2.10b), and likewise as the stability of the non-buoyant portion increases (Fig. 2.10c) – indicating compensation between the buoyant and non-
buoyant portions (Fig. 2.10a). In contrast, the vertical velocities of buoyant CPAR updrafts do not show an increasing trend with the positive buoyancy (Fig. 2.10b), and the non-buoyant portion of CPAR updrafts are closer to neutrally buoyant than the non-CPAR updrafts (Fig. 2.10c), the latter contribute to the higher buoyancy of CPAR updrafts relative to the non-CPAR updrafts.

As a result of the CPAR updrafts not driven by buoyancy and the non-CPAR updrafts showing little effect of buoyancy on controlling the average updraft vertical velocities, the difference between the vertical velocity of the CPAR and non-CPAR updrafts taken from the same output minutes does not increase with higher buoyancy difference (Fig. 2.11a). However, I can infer from Fig. 2.11b that the dynamic lifting by the cold pool outflow contributes to strengthen the CPAR updrafts. The difference in vertical velocity between the CPAR and non-CPAR updrafts is enhanced with the cold pool expansion rate, as compared to the weak negative correlation of the relationship shown in Fig. 2.11a.

The cold pool expansion strengthens the updrafts by lifting air parcels preferably with high $\theta_e$. The example of a cold pool occurring between 2105 UTC to 2123 UTC provided in Fig. 2.6 helps to visualize these processes. In Fig. 2.6, the majority of the CPAR updrafts coincide with a high $\theta_e$ anomaly ($\Delta \theta_e > 2K$) as well as high $q_v$ anomaly ($\Delta q_v > 0.7 \text{ g kg}^{-1}$). A moisture patch located about 1-2 km to the
downwind (southwest) side of the cold pool boundary at 2105 UTC, eventually converges with the cold pool boundary, becoming the locus for further updrafts.

**Impact of ambient wind shear on the lifted updrafts**

The relationship between the cold pool boundary and the ambient vertical wind shear can influence the orientation of the lifted updrafts and thereby the ability to further propagate convection (e.g., Liu and Moncrieff, 1996; Weisman and Rotunno, 2004). Fig. 2.12 shows three possible scenarios of the relationship. The mean cold pool-ambient horizontal wind ($U$), which resembles the mean horizontal wind in this simulation, increases with height up to the cloud base level, with a difference $dU$ between vertical level $z$ and $z + dz$ of $dU = U(z + dz) - U(z)$. By the definition of $C_\ast$, the mean horizontal wind within the cold pool downwind boundary equals $C_\ast + U$, and $dC_\ast = C_\ast(z + dz) - C_\ast(z)$. Since the environmental wind below cloud base level enhances with height (e.g., Fig. 1.5), assume $dU > 0$, and cold pool expansion rate decreases with height, $dC_\ast$ and $dU$ are of opposite sign. The sign and magnitude of $dU$ and $(dC_\ast + dU)$ determine the direction and strength of the circulation associated with the ambient vertical wind shear and the wind shear within the cold pool downwind boundary respectively. When the two circulations compensate for each other $dU \sim -(dC_\ast + dU)$ and $dC_\ast \sim -2dU$ (Fig. 2.12b). When the ambient wind circulation is stronger, $dU > -(dC_\ast + dU)$, and $|dC_\ast| < |2dU|$
Figure 2.12: Schematics of the three scenarios of the relationship between cold pool downwind boundary and ambient wind shear that cause different orientations of the force lifted updrafts. Black arrows indicate mean/ambient horizontal wind ($U$); blue arrows indicate the cold pool expansion rate ($C_\ast$). The tilted solid blue lines represent the interface between cold pool downwind boundary and the ambience.

a) The cold pool downwind boundary circulation is too weak to counter the ambient wind circulation, lifted updrafts follow the ambient wind shear direction.

b) The two circulations are about the same strength, the lifted updrafts rise upright.

c) The cold pool downwind boundary circulation is stronger than the ambient wind circulation, the lifted updrafts follow the wind shear direction within cold pool boundary.
(Fig. 2.12a). $dU < -(dC_s + dU)$ and $|dC_s| > |2dU|$ are true when the cold pool boundary circulation dominates over that of the ambient wind (Fig. 2.12c). The force lifted updrafts follow the direction of the dominant circulation, or rise upright when the two circulations counter each other. This in turn has an impact on the air feeding further into the cloud.

The cold pool-ambient low level vertical wind shear is estimated as the difference of mean horizontal wind relative to the 3-m level ($U_{surf}$), at the 80-m level ($U_{80}$), 170-m level ($U_{170}$), and 300-m level ($U_{300}$) respectively. The ambient vertical wind shear in this simulation closely resembles the domain-mean vertical wind shear. The circulation that may balance the ambient vertical wind shear is associated with the difference in cold pool expansion rate at 80-m level ($C_{s80}$), 170-m level ($C_{s170}$) and 300-m level ($C_{s300}$) relative to the 3-m level ($C_{surf}$). The magnitude of $C_{s80,170,300} - C_{surf}$ is smaller than the magnitude of $U_{80,170,300} - U_{surf}$ for nearly all output minutes, with averaged $|C_{s80,170,300} - C_{surf}| \approx 0.6, 0.9, 1 \text{ m s}^{-1}$ compared to an average ambient wind shear of $|U_{80,170,300} - U_{surf}| \approx 1.8, 2.2, 2.3 \text{ m s}^{-1}$. Thus, the simulation most closely matches the condition shown in Fig. 2.12a, so that the lifted updrafts should rise along the down shear of the ambient or mean wind. The scenario may differ in nature, since the simulation underestimates the cold pool strength.
Figure 2.13: Snapshots of cross section along the dashed black line indicated in Fig. 2.6 for 2105-2123 UTC. a)-d): Cloud mixing ratio $q_c > 0 \text{ g kg}^{-1}$ (black contours); rain mixing ratio $q_r > 0 \text{ g kg}^{-1}$ (light blue contours) with denser contours indicating higher $q_r$; plane projected wind vectors relative to the domain-averaged wind for updrafts (red arrows) and downdrafts (deep blue arrows) below cloud base level (dotted line). Negative buoyancy below cloud base level (dark shaded). e)-h): The speed of domain-averaged horizontal wind projected onto the cross section. Note that the negative wind speed on x-axis suggests northeasterly.
Cross sections 2105-2123 UTC aligned with the domain-averaged surface wind vector illuminate how the vertical structure of the sub-cloud updraft relates to the cloud and rain field (Fig. 2.13a-d). At 2105 UTC, the cold pool boundary is still away from some updrafts extending southwestward to ~2 km in Fig. 2.13a (see also Fig. 2.6a) that support the convection without the cold pool influence. At 2111 UTC, the surface-based stable cold pool layer diminishes the updrafts, and the updrafts below 300 m along the section are too weak to detect (Fig. 2.13b and Fig. 2.6b). At 2117 and 2123 UTC, the cross sections are able to capture the rain and strong downdrafts reaching the surface and are expanding the cold pool downwind boundary (Fig. 2.13c-d). The horizontal wind anomalies of updrafts below 300 m change signs, from against the mean wind direction at 2105 UTC to along the mean wind and mean wind shear direction (Fig. 2.13c-d and g-h). The updrafts that advance faster than the mean wind have better access to the ambient environmental moisture that converged by expanding cold pools. In addition, as evident in Fig. 2.13, the wind shear above the cloud base tilts the cloud into the wind, allowing the precipitation to fall inside the cold pool, and thereby strengthen the already-existing cold pool with its evaporative cooling.
Cold pool effects at the cloud base level

The updrafts at the cloud base level are more directly related to the cloud and rain development than the low-level updrafts considered so far. The 450-m level in the simulation is below and most near the average model cloud base (Fig. 2.3a), and is hereafter referred to as the cloud base level. The thermodynamic and dynamic properties of the cloud base level updrafts are analyzed for evidence of cold pool effects. Similar to the updrafts at 80-m level, the cloud base level updrafts are also divided into two groups depending on whether or not they are directly above the 80-m level CPAR.
The CPAR updrafts retain their high $\theta_e$ anomalies relative to the domain-mean $\theta_e$ from 80 m to 450 m, shifting the number density peak from $\Delta \theta_e^{up} = 1.5 \, K$ to $2K$ (Fig. 2.8c and Fig. 2.14a). The higher $\Delta \theta_e^{up}$ of the CPAR updrafts is mainly due to their higher humidity, but the moister CPAR updrafts are also cooler than the non-CPAR updrafts (Fig. 2.14b). The high $\Delta \theta_e^{up}$ but cold CPAR updrafts are thereby neutrally even slightly less buoyant relative to non-CPAR updrafts (Fig. 2.14c), in contrast to a slightly positive relative buoyancy at the 80-m level (Fig. 2.11a).

The scale of vertical velocity differences (CPAR - non-CPAR) at cloud base level is three times of the 80-m level (Fig 2.14d versus Fig. 2.11b), attesting to the impact of the cold pool dynamic lifting. Although the correlation of cloud base level vertical velocities difference with the 80-m level cold pool expansion rate is even weaker than the 80-m level vertical velocity difference, attributed to the weaker correlation between the CPAR updraft vertical velocities at cloud base level and 80-m level ($r=0.3$) as compared to the non-CPAR updrafts ($r=0.7$). Possible interpretations include that the lifting force that depends on the cold pool strength varies over time with the precipitation fall upon the cold pool, and more mixing with environmental air, compared to updrafts without the influence of cold pools, takes place when air is dynamically lifted from 80 m to the cloud base level under the influence of wind shear interaction between the cold pool boundary and the ambient winds.
Figure 2.15: a) Contoured frequency distributions of the cloud water path of updraft columns as a function of the cloud base level vertical velocity for CPAR updrafts (red; based on the 560 output minutes) and non-CPAR updraft (black; based on the 1441 output minutes). b) Contoured frequency distributions of the difference in cloud water path of CPAR updrafts relative to non-CPAR updrafts as a function of the cloud base level vertical velocity difference, based on the 560 output minutes containing CPAR updrafts.

The cloud base level updraft coverage may also be affected by the presence of cold pool boundary in the domain. A $\Delta \theta_e$ distribution for the CPAR cloud base level updrafts that resembles that for the non-CPAR updrafts (Fig. 2.14a) can be achieved by removing some current CPAR updrafts that contribute to the excessive number density relative to the non-CPAR updrafts for higher $\Delta \theta_e$s, which include $\sim 50\%$ of the current CPAR updrafts. This estimation suggests that cold pool lifting approximately doubles the updraft coverage within CPAR at cloud base level by mechanically lifting air with wider range of $\theta_e$. However, this contribution to total
updraft coverage is very small, since the CPAR updrafts only make up $\sim 3.3\%$ of the total cloud base level updrafts.

Nevertheless, the CPAR updrafts are capable of generating more cloud than the non-CPAR updrafts. The average cloud water path (CWP) associated with CPAR updrafts is $254 \, \text{g m}^{-2}$, compared to $130 \, \text{g m}^{-2}$ for the non-CPAR updrafts. The CWP associated with non-CPAR updrafts depends strongly on the cloud base level vertical velocity (Fig. 2.15a), whereas the CPAR CWP\textsuperscript{up}s are higher in the mean for the same vertical velocity, if also more variable. The CWP\textsuperscript{up} differences (CPAR- non-CPAR) increase from $\sim -200 \, \text{g m}^{-2}$ to $\sim 400 \, \text{g m}^{-2}$ as the relative vertical velocity increases from $\sim -0.5 \, \text{m s}^{-1}$ to $\sim 0.6 \, \text{m s}^{-1}$ (Fig. 2.15b). A similar relationship with CWP does not occur for $\theta$ or humidity difference (not shown).
Chapter 3

The sensitivity of simulated shallow cumulus convection and cold pools to microphysics

3.1 WRF simulations with Thompson and Morrison microphysics

The two WRF simulations are configured in the same manner except for the choice of microphysics scheme. The simulations with one parent domain and four two-way nested inner domains are centered at 61.7°W 18°N, starting at 0000 UTC 19 January and ending at 0600 UTC 20 January (only the 1441 output minutes of the last 24 hours are used for analysis). Only the simulation results from the innermost domain are analyzed. This has a domain size of 24 by 24 km from the surface to 10 hPa, at a horizontal resolution of \(dx = dy = 100\ m\), and a vertical resolution varying from 6 m to 200 m between surface and the 4 km level. Both simulations prescribe
Figure 3.1: The a) potential temperature, b) water vapor mixing ratio, c) zonal, and d) meridional wind speed profiles averaged for the Thompson (black) and Morrison (red) simulation, compared to the averaged soundings from 0700 UTC January 19 to 0300 UTC January 20 (green). Blue dashed profiles in a) and b) indicate the initial profiles used in KiD experiments.

A total cloud droplet number concentration $N_c = 100 \times 10^6 \text{ m}^{-3}$. The averaged $\theta$, water vapor mixing ratio ($q_v$), and wind profiles of the innermost domain that are very similar for the two WRF simulations capture the basic thermodynamic and dynamic features of this day (Fig. 3.1). The daily area-averaged rainfall rate derived from a scanning precipitation radar for this day is 1.87 mm day$^{-1}$ (Snodgrass et al., 2009). An indication of the different responses of the simulated convection to the two microphysics schemes is given by their 24-hour-averaged domain-averaged surface rainrate of 2.1 mm day$^{-1}$ for the Thompson simulation and 2.4 mm day$^{-1}$ for the Morrison simulations, respectively.
Figure 3.2: For the Thompson (black) and Morrison (red) simulation: the vertical profiles averaged over all output minutes: a) cloud fraction (solid) and cloud core fraction (dashed), b) cloud water mixing ratio within cloud, c) rain fraction, d) rain water mixing ratio within rain shaft, and e) total liquid water ($L_t$) (solid) and the ratio of rain water in the total liquid water ($L_r/L_t$) (dash-dotted).

**Cloud and rain in WRF simulations**

Under resembled mean thermodynamic and wind condition (Fig. 3.1), the fractional area of the convective core is similar in the two simulations (Fig. 3.2a). The convective core contains grid points with cloud water mixing ratio $q_c > 0.1 \text{ g kg}^{-1}$, with positive vertical velocity ($w > 0$) and positive buoyancy relative to the domain mean ($b = g\left[\frac{\theta - \bar{\theta}}{\partial} + 0.61(q_v - \bar{q}_v) - q_L \right] > 0$; $g$: gravitational acceleration; $q_L$: liquid water mixing ratio). The average cloud base height in both simulations are $\sim 0.5 \text{ km}$, and the clouds extend up to 2 to 3 km in both simulations (Fig. 3.2 a). However, the Morrison simulation has more cloud above 1.5 km (Fig. 3.2a), and the rain mixing ratio peaks at a higher altitude (Fig. 3.2 c-d), consistent with Shipway and Hill.
Figure 3.3: For the Thompson (black) and Morrison (red) simulation: contoured frequency distribution of the cloud cover as a function of the surface rain fractional coverage. Values indicate the percentile of output minutes included within each contour.

Although the Morrison simulation produces more total liquid water, it retains greater portion of the total liquid water as cloud water. The ratio of rain water to the total liquid water is therefore higher for the Thompson simulation (Fig. 3.2e), reflecting a more efficient autoconversion process as also mentioned in Shipway and Hill (2012), as will also be shown in the next section within the one-dimensional
column model experiments.

The cloud cover correlates much more strongly with surface rain fraction in the Thompson simulation than the Morrison simulation (Fig. 3.3). The boundary layer wind speed increases over the day as observed (Zuidema et al., 2012), leading to the enhancement of the 1-2 km vertical wind shear in the later hours (Fig. 3.4a). Interestingly, only the convection in the Thompson simulation seems to respond to the change of wind shear. The higher autoconversion rate allows the Thompson simulation to produce rain from lower altitudes to cloud top. Before 1200 UTC, the rain originated below 2 km in the Thompson simulation falls into the updraft that created it, discouraging further convection. The increase of rain fraction with cloud fraction in the Thompson simulation is related to the increase of boundary layer vertical wind shear of this day that shears off the cloud layer and enhances the cloud cover, while the simulated rain intensifies accordingly in later hours of the day. On the other hand, since the generation of rain in Morrison simulation is more concentrated at higher altitude within the clouds, the greater wind shear and the consequential greater cloud cover do not correspond to greater rain fraction, therefore, little correlation is observed between the cloud cover and rain fraction for the Morrison simulation (Fig. 3.3). However, an additional Morrison simulation with cloud droplet number concentration $N_c = 250 \times 10^6 \ m^{-3}$ instead of $100 \times 10^6 \ m^{-3}$
Figure 3.4: For the Thompson (black) and Morrison (red) simulation: the one-hour smoothed a) the difference between the speed of domain-averaged horizontal wind at altitude of 2 km and 1 km (2 km - 1 km), and b) the domain-maximum surface rainrate. Dashed line in b) indicates the results from an additional simulation using Morrison scheme with cloud drop concentration of $N_c = 250 \times 10^6 \text{ m}^{-3}$, representing environment with higher aerosol condensation.
indicates weakening of rain intensity with the increase of 1-2 km vertical wind shear (Fig. 3.4b). This means that the depletion of liquid water is essentially effective at strong vertical wind shear for the microphysics scheme with lower autoconversion rate. In the observations, the ship-board rain measurements and scanning radar images show precipitation occurring most frequently between 1200 UTC and 1800 UTC, during the intermediate vertical wind shear (Zuidema et al., 2012).

**Cold pool properties and rain evaporation in WRF simulations**

The modification of surface air properties underneath the rain shaft are dependent on the rain intensity and coverage. Despite significantly different time evolution of rain produce by the two simulations (Fig. 3.4b), the probability density distributions of domain-maximum surface rainrate ($RR_{max}$) for the two simulations are more similar (Fig. 3.5a). I then estimate the impact of rain on the surface air buoyancy relative to the domain mean, represented by the horizontal anomaly of virtual potential temperature ($\Delta \theta_v$). The horizontal anomaly of a scalar is defined by $\Delta \chi(x, y, z) = \chi(x, y, z) - \overline{\chi}(z)$, where $\chi$ is the selected air property, and $\theta_v$ is calculated as $\theta_v = \theta(1 + 0.61q_v - q_L)$. The most noticeable differences in the fractional area of surface negative buoyancy between the two simulations are for $0.5K < \Delta \theta_v < 0.1K$, where the Thompson simulation produces larger area of negative buoyancy, corresponding to more frequent occurrence of $RR_{max} < 10 \text{ mm hr}^{-1}$. 
Figure 3.5: For the Thompson (black) and Morrison (red) simulation: a) The probability density function of $RR_{\text{max}}$ in 5 mm hr$^{-1}$ interval; b) the accumulated fractional area containing surface $\Delta \theta_v$ smaller than the values on x-axis.
In the same manner as Fig. 2.5 in Chapter 2, the maximum changes of the surface air properties are estimated and plotted as functions of the domain-maximum surface rainrate \((RR_{\text{max}})\) (Fig. 3.6). For majority of cases with \(RR_{\text{max}} < 20 \text{ mm hr}^{-1}\), the Thompson simulation produces slightly stronger decreases of surface temperature, hence buoyancy, on average. For the few instances with very intense rain \((RR_{\text{max}} > 40 \text{ mm hr}^{-1})\), however, the surface air temperature and buoyancy response is stronger in the Morrison simulation. These are reflected in the steeper linear best fit lines of \(\Delta \theta, \Delta \theta_v\), and \(\Delta U\) with \(RR_{\text{max}}\) for the Morrison simulation. The very intense \(RR_{\text{max}}\) value is comparable to the observed rainrates of this day, since one among only three ship documented rain events has maximum surface rainrate of 45 mm hr\(^{-1}\) (Zuidema et al., 2012; Li et al., 2014).

The evaporation process should explain the different cold pool temperature depressions produced by the two simulations for a given rainrate. I examine the the column-integrated evaporation rate \((E_{\text{col}}^r)\) and the column-integrated evaporation rate below 1 km altitude \((E_{\text{r}}^{1km})\) as functions of surface rainrate for the grid point containing the domain-maximum surface rainrate from each output minute (Fig. 3.7). Both \(E_{\text{col}}^r\) and \(E_{\text{r}}^{1km}\) are higher within the Morrison simulation than the Thompson simulation for very high rainrates \((RR_{\text{max}} > 40 \text{ mm hr}^{-1})\), although occurring less often. \(E_{\text{r}}^{1km}\) is higher within the Thompson simulation than the Morrison
simulation for columns with $RR_{\text{max}}$ below 20 mm $hr^{-1}$, corresponding to slightly stronger temperature depressions in Fig. 3.6. However the difference in $E_{r}^{\text{col}}$ between the two schemes is less consistent with the surface cold pool properties at moderate rainrates. The difference between the rain evaporation rate integrated along columns below 1 km and the entire modeled atmosphere column is likely caused by greater rain mixing ratio, hence greater rain evaporation, above 2 km (Fig. 3.2). The rain evaporation at lower altitudes seems more related to the properties of surface cold pools as may be expected.

In Fig. 3.8, the vertical profiles of the rain mixing ratio ($q_{r}$), evaporation rate
Figure 3.7: For the Thompson (black) and Morrison (red) simulation: contoured frequency distribution of the column-integrated evaporation rate as a function of the domain-maximum surface rainrate, a) integrated over the entire model column; b) integrated over the column below 1 km.
Figure 3.8: For the Thompson (black) and Morrison (red) simulation: the profiles of a) rain mixing ratio, b) evaporation rate, and c) evaporation efficiency, composite over grid columns containing the domain-maximum surface rainrate of selected moderate (solid) and very intense (dashed) rain cases.

\(E_r\), and evaporation efficiency, which is indicated simply as \(E_r/q_r\), are shown for the composite over columns containing domain-maximum surface rainrates of \(5 < R_{R_{\text{max}}} < 10 \, \text{mm hr}^{-1}\) (moderate) and \(40 < R_{R_{\text{max}}} < 80 \, \text{mm hr}^{-1}\) (intense) respectively. At the relatively moderate rainrates, the Thompson scheme clearly produces a higher evaporation rate below 1 km as well as a greater evaporation efficiency. At the higher rainrates, the rain mixing ratios and evaporation rates are less within the Thompson simulation than the Morrison simulation, and so does the evaporation efficiency. The inclusion of higher order terms in the numerical solution of evaporation rate within the Thompson scheme may benefit the high evaporation efficiency shown for the moderate rain cases, but can not explain the lower evaporation efficiency for intense rain cases. Notice that since the higher rainrates occur
much less frequently (Fig. 3.5a), the moderate cases should dominate the characteristics of rain evaporation in both simulations, and lead to greater rain evaporation at low altitudes in the Thompson simulation on average.

### 3.2 1D column model experiments

Experimentations with a one-dimensional model using the Thompson and Morrison microphysics schemes confirm that the microphysical feedbacks with the environment does not mask the difference in rain and evaporation that can be attributed to the microphysics schemes themselves. The KiD model is designed for intercomparisons of microphysics schemes, the microphysical processes in the KiD model are driven by prescribed vertical velocity profiles that meant to resemble the convective core (Shipway and Hill, 2012). All experiments are run for 5400 seconds, updating every 1 second, and outputting prognostic variables in every 10 seconds. The column is set to be 7 km high, with 120 full z-levels and 119 half z-levels at a vertical resolution of \( \sim 58m \).

The initial thermodynamic profiles resemble the average shipboard sounding during the day of 19 January 2005 (Figure 3.1). The \( \theta \) profile is held constant throughout the run and only the water vapor mixing ratio \( q_v \) is updated. The imposed vertical velocity profiles resemble an updraught core modeled on data from Abel and Shipway (2007) (test Case 5 in KiD user’s guide version 2.3.2625, avail-
Figure 3.9: For the Thompson (black) and Morrison (red) run driven by the same\,$w$ profiles: the profiles of a) cloud water mixing ratio, b) rain mixing ratio, c) auto-conversion rate, and d) accretion rate, composite over the 5400 seconds model run time.

able online at http://appconv.metoffice.com/microphysics/doc.shtml), varying exponentially with height and time. The profiles reach their peak value ($w_{\text{max}}$) at approximately the same altitude as the peak of averaged updraft profiles simulated by both the WRF simulations. For each microphysics scheme, experiments are carried out by varying the $w_{\text{max}}$ from 0.2 to 2 m s$^{-1}$, encompassing the range observed with Doppler lidar (Zuidema et al., 2012).

The generation of cloud and rain by the two schemes is evaluated by examining the two runs with Thompson and Morrison scheme respectively driven by the exact same vertical velocity profiles at all times. The time-averaged profiles of cloud mixing ratio and rain mixing ratio are shown in Fig. 3.9a-b to compare with the average profiles from 3D WRF simulations (Fig. 3.2). As expected, the averaged cloud
and rain mixing ratio profiles are of the same magnitude as the average profiles from WRF simulations, and the surface rainrate produced by these two 1D model runs are less than 10 \( \text{mm hr}^{-1} \), corresponding to the most frequent surface rainrate in the WRF simulations. Also consistent with the 3D simulations, the 1D model produces more cloud for the Morrison run, but greater ratio of rain to cloud water within the Thompson run. Further examination on the autoconversion rate and accretion rate for these two runs indicates that the significantly higher autoconversion rate of the Thompson scheme causes the slightly greater rain mixing ratio, especially below 1.5 km (Fig. 3.9c), despite the production of less cloud the Thompson scheme. This may also explain the greater ratio of rain water over total liquid water for the 3D Thompson simulation (Fig. 3.2e). The difference in accretion rate between the two scheme is very small below 1.5 km, but the Morrison scheme produces greater accretion rate at higher altitudes (Fig. 3.9d). Since the scale of accretion rate within the two schemes is two orders larger than the autoconversion rate, the small advantage of the Morrison scheme in the accretion rate may compensate or surpass its disadvantage in autoconversion rate. In these two KiD runs, the Morrison run matches the Thompson run in rain mixing ratio above 1.5 km; in the WRF 3D simulations, the Morrison simulation produces greater rain fraction and rain mixing ratio above 2 km.
Figure 3.10: The four time series of surface rainrate from the runs representing the moderate (solid) and intense (dashed) rain cases for Thompson (black) and Morrison (red) scheme respectively. Vertical lines mark the beginning and end of the 1000 seconds period over which the composite of Fig. 3.11 is made.
Although the 1D model runs can not generate cold pools, they can still be compared to the 3D simulations on the rain evaporation rate that directly associated with the cold pool generation. In a similar manner as the rain evaporation profiles of columns with domain-maximum surface rainrate are examined in the 3D simulations, the rain evaporation profiles produced by the two microphysics schemes are compared for the same surface rainrate in the 1D model runs for both moderate and intense rain scenarios. The same surface rainrate for Thompson and Morrison runs are achieved using different vertical velocity profiles. The four runs that are used to represent moderate (\( \sim 5 \text{ mm hr}^{-1} \)) and intense (\( \sim 60 \text{ mm hr}^{-1} \)) cases for the Thompson and Morrison scheme respectively are selected based on their surface rainrates (Fig. 3.10). Composite profiles of rain mixing ratio \( (q_r) \), rain evaporation rate \( (E_r) \), and \( E_r/q_r \) are constructed for each experiment (Fig. 3.11), by averaging over the 1000 seconds that correspond to the surface rainrate selections. The vertical structure of each profile from 1D runs differs from its corresponding composite profile from WRF simulations, likely due to the shearing off of rain shaft by vertical wind shear and wind shift within the WRF simulations that can not be represented by the 1D model. However, the important result is that the ordering of the four composites relative to each other is similar to that in Fig. 3.8, with the greatest rain evaporation efficiency occurs in the moderate Thompson case.
Figure 3.11: For the Thompson (black) and Morrison (red) scheme: composite profiles from the 1D KiD model output of: a) rain mixing ratio, b) rain evaporation rate, c) rain evaporation efficiency, averaged over the 1000 seconds indicated in Fig. 3.10 for moderate (solid) and intense (dashed) rain case.

The terminal fall speed of raindrops is different for the two microphysics schemes, as a result of different parameters used in the scheme (Shipway and Hill, 2012). Fig. 3.12 shows that the Thompson scheme parameterizes greater fall speed for raindrop size up to $\sim 3 \text{ mm hr}^{-1}$, and the difference increases with the drop size from $\sim 0.4 \text{ mm}$ to $\sim 2 \text{ mm}$. The effects of fall speed on the difference in producing rain and rain evaporation rate between the two schemes are investigated through the same KiD model experiments shown in Fig. 3.10 and Fig. 3.11, by replacing the fall speed parameters within the Thompson scheme with the ones used in the Morrison scheme. Only the results from Thompson runs may be affected.

Fig. 3.13 shows that, by replacing the terminal fall speed parameterization within the Thompson scheme with the one used in the Morrison scheme, the evaporation
Figure 3.12: The parameterized terminal fall speed of raindrops over size spectrum 100 µm to 4 mm for the Thompson (black) scheme, and 80 µm to 4 mm for the Morrison (red) scheme.
efficiency of the moderate Thompson run is significantly reduced to match that of the Morrison run, due to the increase of rain mixing ratio (Fig. 3.11b and Fig. 3.13b) and the decrease of evaporation rate (Fig. 3.11b and Fig. 3.13b). These changes in the results from Thompson moderate run may be explained through the changes in RSD caused by raindrop size sorting. Since the fall speed is reduced from the original Thompson run to the modified Thompson run, especially for larger drop sizes that experience greater difference (Fig. 3.12), more larger drops are retained in the air instead of falling to the ground. This causes a change in the RSD for the Thompson moderate run, with relatively more larger drops and less small drops (Fig. 3.14). This change in the RSD leads to the increase of rain mixing ratio, and the decrease of rain evaporation rate.

Figure 3.13: Same as Fig. 3.11, except the Thompson profiles are from runs with the raindrop terminal fall speed parameters replaced by the ones in the Morrison scheme.
Figure 3.14: At 530 m level, the RSDs calculated over a raindrop size spectrum, 100 $\mu$m to 4 mm for the Thompson (solid black) scheme, and 80 $\mu$m to 4 mm for the Morrison (solid red) scheme, averaged over the RSDs of selected 1000 seconds for the moderate KiD runs. The dash-dotted black line indicates the RSD calculated in the same manner, but from the moderate Thompson run with fall speed parameters replaced by the one used in the Morrison scheme.
Figure 3.15: At 500 m level, the RSDs calculated over a raindrop size spectrum, 100 µm to 4 mm for the Thompson (dashed black) scheme, and 80 µm to 4 mm for the Morrison (dashed red) scheme, a) in KiD runs, averaged over the RSDs of selected 1000 seconds for the two intense KiD runs; b) in WRF simulations, averaged RSD from grid columns possessing the domain-maximum rainrate between 50 mm hr$^{-1}$ and 70 mm hr$^{-1}$.

For the two KiD runs corresponding to intense rainrates that occur much less frequent in the 3D WRF simulations, the replacement of fall speed parameterization does not affect much of the rain mixing ratio and rain evaporation of the Thompson run (Fig. 3.13). The RSDs from the intense KiD runs show greater number of total raindrops as well as the smaller raindrops for the Morrison scheme, consistent with the greater rain mixing ratio, evaporation rate, and greater rain evaporation efficiency produced by the intense Morrison run (Fig. 3.15a). However, the RSDs
from the WRF simulations that correspond to intense rainrates do not show significant difference in the number concentration of smaller drops (Fig. 3.15b), therefore, can hardly explain the greater rain evaporation efficiency of intense Morrison cases for the WRF simulation. The different relationship of RSD produced by the two scheme in the KiD runs and the WRF simulations may infer the involvement of dynamical factors in the intense rain conditions generated by the WRF simulations. Nevertheless, for the same intense surface rainrate, the Morrison scheme produces greater rain mixing ratio, evaporation rate, and evaporation efficiency in both 3D simulations and 1D experiments.
Chapter 4

Comparisons on cold pools from DYNAMO and RICO

The study so far is based on observation and simulation restricted within the Caribbean trade-wind region, the following analysis will provide an different perspective on the characteristics of shallow cumulus cold pools, by examining the observation data from equatorial Indian Ocean. This analysis will further expand our knowledge on shallow cumulus cold pools for a more comprehensive view.

4.1 The DYNAMO data set

The DYNAMO observation array included land-based facilities, research vessels, soundings sites, and research flights. The Department of Energy Atmospheric Radiation Measurement (ARM) Mobil Facility 2 (AMF-2) was deployed at the Gan airport, located at Addu Atoll, Maldives (0.7°S, 73.15°E) (Long et al., 2011). The AMF-2 site on Gan contained meteorological measurement tower, a Ka-band (8.6-mm
wavelength) ARM Zenith Radars (KAZR), radiosondes launched every 3 hours, and instruments measuring atmosphere radiation (see Fig. 3 in Yoneyama et al., 2013). NCAR/EOL’s S-PolKa radar with 10 cm S-band and 0.86 cm Ka-ban transmission operated at Gan island, north west to the AMF-2 site, at 0.64°S, 73.1°E. The S-band radar operated in both Plan Position Indicator (PPI) mode and Range Height Indicator (RHI) mode. The surveillance scan began from the 0.5° elevation angle.

The central equatorial Indian Ocean (IO) during the IOP was favorable for convective development, judging from the sea surface temperature (SST) distributions (Gottschalck et al., 2013). Four MJO events were documented during the deployment of DYNAMO in late October, late November, late December and March, with large-scale convective events propagating eastward across the tropical IO to reach the Maritime Continent (Yoneyama et al., 2013). The two MJO events during October through December period are shown in the sounding time series of relative humidity (RH), SST and the surface precipitation rate (Fig. 4.1). These events are related to the mean state of each day over the DYNAMO observation array. Despite the active growth of deep convection, precipitation from shallow convection was also often observed. This provided an opportunity to compare the shallow precipitation and the associated cold pools under environmental conditions that favor deep convection, to those occur under environmental conditions that were less support-
Figure 4.1: (top) Integrated depiction of vertical structure of two MJOs during the period October 1 to December 15, 2011 based on mean fields over the northern sounding array. Green (yellow) shading denotes areas of relative humidity greater than 70% (less than 40%), respectively. Thin (thick) arrows denote locations of zonal wind (vertical motion) maxima. Centers of cool and warm temperature anomalies are shown. Relative magnitudes of features are represented by sizes of symbols. At upper levels, these anomalies have a tilted structure as do the easterly wind maxima.Daily averaged 0°C level (dashed line) and cold-point tropopause level (solid line) are indicated. (middle) Hourly SST time series at R/V Revelle. (bottom) TRMM 3B42 daily averaged rainfall rate over the northern sounding array. As Fig. 24 in Johnson and Ciesielski (2013).
Table 4.1: Mean and standard deviation of near surface atmosphere properties during DYNAMO (October 8, 2011 to January 15, 2012) and RICO (January 9, 2005 to January 24, 2005)

<table>
<thead>
<tr>
<th></th>
<th>T (C)</th>
<th>$q_v$ (g kg$^{-1}$)</th>
<th>RH (%)</th>
<th>Wind speed ($m s^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>RICO mean</td>
<td>25.2</td>
<td>14.9</td>
<td>74.1</td>
<td>7.9</td>
</tr>
<tr>
<td>DYNAMO mean</td>
<td>28.1</td>
<td>15.2</td>
<td>72.3</td>
<td>2.4</td>
</tr>
<tr>
<td>RICO $\sigma$</td>
<td>0.7</td>
<td>1.0</td>
<td>4.8</td>
<td>2.9</td>
</tr>
<tr>
<td>DYNAMO $\sigma$</td>
<td>0.9</td>
<td>0.9</td>
<td>4.2</td>
<td>1.1</td>
</tr>
</tbody>
</table>

The meteorological conditions over Addu Atoll is documented daily from the S-PolKa radar site. As a member of the science team at the S-PolKa radar site, I contributed to the composition of daily science report during the S-PolKa operational period, which is available online at [http://catalog.eol.ucar.edu/cgi-bin/dynamo/report/index](http://catalog.eol.ucar.edu/cgi-bin/dynamo/report/index) under "SPolKa Scientist science summary".

Significant difference between the two environmental conditions can be shown by simply examining the mean near surface atmosphere properties. The environmental condition during RICO is represented by the average of measurements taken by the flux tower on RVSJ, while the data from met tower on the AMF-2 site is used to calculate the mean state of near surface atmosphere during DYNAMO. As shown in Table 4.1, the DYNAMO mean state is warmer and moister, but the rel-
ative humidity is slightly lower, with notably lower surface wind speed than RICO condition. The lower and less variant wind speed may affect the propagation of cold pools and the interaction between cold pool and associated convection. In visible channel satellite imageries shown in Fig. 4.2, while the strong boundary layer wind condition in the trade-wind region is associated with arc-shaped cloud organization at the cold pool downwind boundary (Fig. 4.2a), the cloud organizations around cold pools under lighter wind condition in the tropical region are closer to being a full circle (Fig. 4.2b). This is likely because that the part of outflow against wind direction is much weaker than the strong wind in the RICO condition, but is more comparable to the environmental wind in the DYNAMO condition.

4.2 Cold pools in DYNAMO

Similar to (Zuidema et al., 2012), the air property measurements from instruments mounted on a met tower are used to identify and characterize the cold pools during DYNAMO. In Zuidema et al. (2012), cold pools are identified based on both surface temperature and rainrate, but no threshold of temperature depression is involved, since this study attempts to associate the changes in surface air properties with shallow precipitation. Now the connection has been made, one may assume a temperature depression associated with each precipitation event, and use identify the cold pools from the temperature depression only. To be more specific, a
Figure 4.2: a) GOES-12 visible channel 1 km resolution satellite imagery at 11625 UTC January 14, 2005, during RICO campaign, over the Caribbean. b) Aqua-MODIS visible channel 1 km resolution for 0855 UTC November 11, 2011, during DYNAMO campaign, over the observational array. The red triangle indicates the location of AMF-2 site. The yellow arrow in both figures indicate the boundary layer wind direction.
cold pools is identified when the surface temperature drops at least 1 K within a 7-minute window. The applied threshold for cold pool criteria automatically filters out the weaker cold pools with small temperature depression, but at the same time includes some temperature depression that may not necessarily be associated with precipitation. Similar criteria are used to identify cold pools documented on ship during DYNAMO (Simon de Szoeke, personal contact).

Base on the criteria described above, cold pools are identified through examining the met tower data from October 8, 2011 to January 15, 2012 in ~1 minute resolution. The identified cold pools during DYNAMO IOP are mostly associated with deep convection, since the environmental conditions during the period encourage the growth of deep convection. Nevertheless, some cold pools seem to be caused by shallow convective precipitation. The distinction is made through the KAZR profiling data from the same ~3 months period in ~4 seconds resolution, by using a simple algorithm to examine whether the first level cloud top height is lower than the 0°C level at ~5 km. The variable of cloud layer top height in KAZR Active Remote Sensing of Clouds (ARSCL) Value Added Product (VAP) is used to estimate the cloud top height. Visual examination through KAZR reflectivity plots of the interested period is also applied to further identify cold pools associated with shallow convection.
The time series of cold pool properties and associated convection of one example shallow cumulus cold pools are shown in Fig. 4.3. This cold pool passage over the observational site lasts over an hour, with the minimum temperature inside this cold pool about 3 K lower than the environment. The convection that corresponds to the surface temperature depression extends to about 4 km, with highest reflectivities observed at between 1 to 2 km, and about 3 km. Precipitation from the convection reaches the surface, judging from KAZR reflectivity, co-locating with the rapid drop of surface temperature and $\theta_e$.

The AMF-2 site is based on land rather than a ship, the land underneath may only contribute to the temperature recovery through sensible heat flux, but barely to the humidity recovery through latent heat flux. However, such difference is hard to be manifested in the time series data on one observation point, since the moisture advection and environmental mixing also take place. The wind gust at the cold pool front boundary is not distinguishable from the wind speed data in this case, as it also happens to some RICO cold pool cases (e.g., Fig. 12 in Zuidema et al., 2012). The specific humidity is not lower everywhere within the cold pool, but a depression in time series is obvious right about the time of the greatest temperature depression.

The changes of surface air properties within cold pools observed during DY-NAMO, including 20 shallow convection cold pool cases and 105 deep convec-
Figure 4.3: The time series of 0800 to 1100 UTC November 13, 2011 for a) KAZR reflectivity; b) surface temperature, the green and red circle stand for the beginning and end of the cold pool period respectively, the blue circle marks the point with lowest temperature within the cold pool period; c) surface specific humidity; d) surface $\theta_e$; e) surface wind speed.
Figure 4.4: The maximum changes of a) surface specific humidity; b) surface wind speed as functions of the surface temperature change within each cold pool. c) The cold pool duration as a function of surface temperature change. The probability density function of maximum changes with cold pools for d) surface specific humidity; e) surface wind speed; f) cold pool duration; g) surface temperature change. Red closed circle and black closed circle represent the cold pools generated by shallow and deep convection in DYNAMO. Blue triangle stands for the RICO cold pools.
tion cold pool cases, are compared to those observed during RICO (Fig. 4.4). The changes within each cold pools are calculated as the difference between the lowest or highest value within the cold pool period relative to the beginning of cold pool, where the beginning and end of the cold pool are identified as the data point from (until) which the successive decrease (recovery) of temperature begins (ends) (e.g., Fig. 4.3b). The beginning of cold pools is defined slightly different in Zuidema et al. (2012) that involves the criterion on surface rainrate, but the resulted cold pool lengths do not vary significantly. However, all the 20 shallow convection and 105 deep convection are co-located with the documented cold pools in Fig. 4.4, as the example shows in Fig. 4.3, similar to the RICO cold pools documented in Zuidema et al. (2012).

Despite the larger temperature drop within the cold pools, the DYNAMO cold pools have smaller surface specific humidity change and gust wind speed (Fig. 4.4a-b). The relationships between the cold pool duration and the surface temperature change of DYNAMO and RICO, however, are more similar (Fig. 4.4c). It is expected that the cold pools generated from shallow convection in DYNAMO are more often weaker than the ones from deep convection, as reflected in the higher frequency of occurrence in small surface changes (Fig. 4.4d-g). However, the relationships of the change in surface temperature with changes in specific humidity and wind
wind speed are not distinguishable between the cold pools from shallow and deep convection in DYNAMO.

The difference in the relationships shown in Fig. 4.4 may be explained by the origin of cold pools. The four radiosonde profiles from DYNAMO Gan island and three profiles from RICO RVSJ are shown in Fig. 4.5, each of these radiosondes is coincided with a shallow cumulus cold pool documented close to its launch time. Apart from the warmer and more moist atmosphere profiles of DYNAMO that may be expected from the aforementioned analysis on surface mean state, it is also noted that the DYNAMO shallow cumulus cold pools seem less pronounced in the vertical \( \theta \) profiles than the RICO cases (Fig. 4.5a). The \( \theta_e \) below 1 km during DYNAMO is much higher than RICO (Fig. 4.5b), and the vertical variation is smaller.
Figure 4.6: S-band radar reflectivity on 0.5° PPI scan from S-PolKa for a) 0900 UTC; b) 1015 UTC; c) 1145 UTC; d) 1300 UTC. The orange box follow the examined convection and cold pool outflow. Images available at http://www.eol.ucar.edu/rsf/image_archive/dynamo/spol/sur
The lack of vertical wind shear below 4 km in DYNAMO soundings is another notable feature that differentiate the DYNAMO condition from the RICO condition. As discussed in Zuidema et al. (2012) and Li et al. (2014), the wind shear in the cloud layer enables the rain evaporation in environmental air above the cloud base level, therefore generating convective downdrafts that originate from above the more well mixed sub-cloud layer. This may be another reason that the DYNAMO cold pools are not as dry as the RICO cold pools. In addition, the vertical wind shear in cloud layer help separate the precipitating downdraft region away from the convective updrafts that feed into the convection, allowing further growth of shallow convection and precipitation. The cold pools are more likely to be positioned directly underneath the shallow convection in DYNAMO (Fig. 4.3a and c), rather than behind the cloud front of the convection such as in Fig. 2.13.

Visual examination on S-PolKa surveillance scan images from October 8, 2011 to January 15, 2012 reveals many cases where new convection is invigorated upon the arrival of cold pool outflow. However, most of the visually significant cases are associated with cold pools generated by deep convection. A case from the day of November 13, 2011 is shown as an example. Around 0900 UTC, a cold pool is formed underneath precipitating convection on the northeast corner of the S-PolKa S-band surveillance scan (Fig. 4.6a). The outflow from this cold pool expands about
from 0900 to 1145 UTC, and encourages convection along the outflow boundary (Fig. 4.6b-c). The newly invigorated convection later intensifies the cold pool and sends out another outflow boundary (Fig. 4.6d). For shallow convection and associated cold pools, it is more difficult to distinguish them on the S-PolKa radar images, since the shallow clouds are only \( \sim 5 \) km radius, and the cold pool outflow boundaries are too weak to appear as the green curves on the images due to Bragg scattering signals.

Future work will be needed to further investigate the relationship of shallow cumulus cold pools to different environmental conditions. For example, are the gratin of shallow cumulus cold pools different during the suppression period of MJO when the mid-level atmosphere is relatively dry, comparing to the MJO dominated period when the deep convection precipitation is the primary source of surface precipitation and cooling? The results of this investigation would help better understand the role of shallow convection and precipitation in the MJO related processes.
Chapter 5

Conclusions and discussions

This thesis studies the cold pools generated from shallow cumulus convection by analyzing the observed and simulated cold pools. Chapter 2 attempts to identify the dominant mechanisms supporting the updrafts generating precipitation by shallow trade-wind cumuli organized in mesoscale arcs. The mechanisms that are considered include 1) thermodynamically-enhanced updrafts due to either environmental moisture from outside the cold pool or moisture brought to the sub-cloud layer by the evaporation of rain inside the cold pool (Tompkins, 2001), and 2) forced lifting by the expanding cold pool. These mechanisms are not completely independent of each other, since, for example, both mechanisms may be implicated when a cold pool boundary moving faster than the mean wind converges and further lifts high \( \theta_e \) air, as sketched in Fig. 5.1. The individual mechanisms are investigated through statistical comparisons of the updrafts within and outside of the regions near the
cold pool boundaries, as well as the time evolution of a selected case study (Li et al., 2014).

The analyses reveal that the updrafts in the proximity of the downwind cold pool boundary are typically moister than the other updrafts. The higher humidities associated with higher $\theta_v$ for the cold pool influenced updrafts. The temperatures of the CPAR updrafts are slightly warmer than those of the non-CPAR updrafts at lower sub-cloud layer, but are cooler at the cloud base level. The local dynamic processes due to cold pool propagation significantly modify the properties of CPAR updrafts. The updrafts near the cold pool boundaries generate more cloud water on average than the updrafts away from cold pools. The cold pool boundary collects environmental moisture as it propagates into new environments, the advection of moisture causes moisture convergence close to the boundary during the propagation. The moisture excess of the updrafts within CPARs correlates well with an overall positive moisture anomaly for the CPARs. The cold pool lifted air parcels originate from the moist air within CPAR, and contribute to the broader $\theta_v$ range for the CPAR updrafts. These results speak to a preferential sampling of pre-existing environmental moisture pools that may well reflect the remnants of other convective events, but little evidence is found to support the hypothesis that updrafts at the cold pool boundary arise from air pre-moistened by earlier sub-cloud evaporation
from the same convection (Tompkins, 2001). Instead, these results speak more to the ability of secondary convection to thrive and continue when propagating into relatively moist environments.

The strengthening of CPAR updrafts by cold pool dynamic lifting is evident
through their enhanced vertical velocities at the 80-m level, as compared to the non-CPAR updrafts. The dynamic lifting by the cold pool outflow slightly increases the area covered by the cloud base level updrafts, by introducing air parcels that may not rise otherwise. Since these parcels tend to be moist, they contribute to the higher cloud liquid water paths associated with the CPAR updrafts compared to non-CPAR updrafts.

This study relies on a nested-WRF simulation with an innermost domain of size 24 km by 24 km and a horizontal grid spacing of 100 m, and vertical resolution of 48 levels below 4 km. The nesting technique provides open lateral boundary conditions that allow the reanalysis-derived large-scale forcing imposed on the parent domain of 972 km by 972 km to be transmitted to the innermost domain capable of resolving large-eddy-scale circulations. This simulation is able to produce cold pools, as are observed for this day, if with weaker changes in surface air properties for rainrates that often exceed those observed. This ability to model cold pools can be compared with the behavior of large-eddy-scale simulations that typically apply doubly-periodic boundary conditions and idealized, homogeneous initial conditions and forcings. Such simulations require larger domains at a higher resolution to be able to capture the spatially-inhomogeneous, asymmetric trade-wind cumulus cold pools (e.g., Matheou et al., 2011; Seifert and Heus, 2013). Instead, this study’s
modeling approach demonstrates that the ability to explicitly simulate a spectrum of scales, from the large-scale flow, to the mesoscale, and further down to the largest of the turbulent eddies, is also effective for the simulation of cold pools. This connection between the synoptic scale and the mesoscale has typically been ignored for trade-wind cumulus, but likely represents a common influence on the trade-wind region by the mid-latitudes, of which January 19, 2005 is an example.

Wind shear is arguably overly efficient at shearing off upper-level cloud in the simulations compared to the observations (Fig. 2.13) and this may suggest that the updraft strength is also weaker than in nature. Nevertheless, the increase in cloud fraction with domain-averaged rainrates is in agreement with observation. The weak simulated cold pools in average could reflect issues with the spatial resolution, microphysical parameterization, and turbulent mixing. A reduced grid spacing would allow more resolution of the turbulent scale processes and amplify the downdrafts and updrafts for forming the cold pools and invigorating convection. This in turn may also affect the relationship of the cold pool circulation with that of the low-level environmental wind shear. In this simulation, the low-level environmental circulation is consistently stronger than that of the cold pools, and in the case study examination of updrafts lifted by a strong cold pool, the lifted updrafts point downshear, as would be expected.
The trade-wind cumulus cold pools are associated with stronger winds in lower boundary layer, which in turn will generate stronger surface fluxes (Fig. 2.5; see Zuidema et al. 2012 for a more complete analysis of the observed surface fluxes). The cold pools provide a mechanism for generating the stronger updraft velocities that can also deepen the boundary layer. As such the cold pools may help stabilize the relative humidity in the trade-wind regions, by encouraging the drying and warming of the boundary layer through entrainment of drier, warmer, air from aloft. This contrasts with the cooler, moister cold pools that have been observed for stratuscumulus (Terai and Wood, 2013), wherein increased stability traps the moisture introduced by the latent heat fluxes off of the ocean near the surface.

This study focuses on one day only, that of January 19, 2005. The convection generated for this day is associated with a dissipating cold front, but the associated cold pool strength falls within the envelope of values from more truly undisturbed days (Fig. 2.5). Further comparative modeling studies can help assess the representativeness of these results for days possessing cold pools within other typical trade-wind conditions.

Possible reasons for the weak cold pools in this and other simulations despite discrepancies in the imposed boundary conditions include numerical diffusion and grid resolution (Matheou et al., 2011), as well as the microphysical parameter-
ization (Seifert and Heus, 2013). The parameterized processes of rain production and the raindrop size distribution are crucial for the simulation of cold pools. More rain leads to larger and stronger cold pools. Since small drops evaporate more readily, distribution that favors smaller drops causes higher evaporation efficiency, thus stronger cold pools. This in fact is evident in a one-dimensional model inter-comparison where the Thompson scheme produces the largest rain evaporation rates of the microphysics schemes assessed for similar environmental moistures (Shipway and Hill, 2012). The modeled surface rainrates are larger than observed, and the use of assimilated soundings help ensure that the sub-cloud environmental moisture is realistic, so that it seems unlikely that the simulated rain evaporation rates are less than occurred in nature. This does not rule out the importance of turbulent mixing in changing the properties of the downdrafts and updrafts.

Microphysical parameterizations have a great impact on simulated shallow cloud and precipitation features, thereby affecting the characteristics of cold pools. The sensitivity of simulated cold pools to microphysics is important to the simulation of shallow cumulus, because cold pools suppress thermals within the area with cold and stable surface layer, and encourage secondary convection and impact cloud fraction through the cold pool propagation and expansion. In Chapter 3, I examine cold pool sensitivity to the hybrid single-moment Thompson scheme that allows
the intercept parameter to vary with rain mass and thereby produces more small drops at lower rainrates, and the double-moment Morrison scheme (Morrison et al., 2005). Both microphysics schemes assume exponential raindrop size distributions, matching distribution fits to observed RSDs from the RICO experiment (Geoffroy et al., 2014).

The findings of this study emphasize that the autoconversion and accretion processes are crucial to the production of rain in model simulations, affecting not only the amount of rain water, but also the spacial structure of precipitating convection. With the aim of varying vertical wind shear of this simulated day, the effect of autoconversion process on convection is able to manifest on the rain intensity caused by the interaction with environmental wind shear. The greater autoconversion rate allows the Thompson simulation to generate rain water starting from lower altitudes instead of concentrating on higher altitudes due to greater accretion rate in the Morrison simulation. The lower rain origin in the Thompson simulation renders the convection to be sensitive to the change of vertical wind shear, as the stronger vertical wind shear help avoiding the rain shaft from falling onto the lower updrafts, therefore allowing further growth of the convection. Additional simulation using Morrison scheme with greater cloud drop number concentration shows sensitivity of convection intensity to the vertical wind shear, with convection weakens
at stronger vertical wind shear condition, suggesting the more efficient depletion of rain due to increase of aerosol number concentration at stronger vertical wind shear condition for the Morrison scheme.

The relationship between the surface cold pool properties and rainfall is examined from two different aspects. The surface cold pool coverages within the two simulations are similar, since the occurrences of various rainrates do not differ much in the two simulations, despite the much stronger sensitivity to vertical wind shear in the Thompson simulation. The intensity of surface cold pools is then compared between the two simulation for the matching surface rainrates. For moderate rainrates, the Thompson simulation contains more colder and stronger cold pools, but the situation is reversed for the very intense rainrates that occur much less often. The surface cold pool properties are strongly controlled by the rain evaporation, especially the evaporation that occurs below 1 km. The Thompson scheme produces more rain evaporation for the moderate rain cases, and the Morrison scheme generates more rain evaporation for the very intense rain cases.

Experiments with 1D model find a similar order in rain mixing ratio, rain evaporation rate, and the ratio between the two microphysics schemes in moderate and intense rain conditions, confirming the dominance of microphysics in producing the difference in evaporation efficiency within the 3D simulations. Experiments with
replacing raindrop terminal fall speed parameters in the Thompson scheme with the ones used in the Morrison scheme prove the importance of the different fall speed parameterizations in differentiating the evaporation efficiency of the two scheme for moderate rain cases, since the fall speed alters the RSD and affects both the rain mixing ratio and evaporation rate. For the very intense rain cases that occurs much less frequent in the 3D simulations, however, fall speed does not seem to be generating the difference in rain mixing ratio, evaporation rate and evaporation efficiency of the two schemes.

In Chapter 4, the cold pools documented during DYNAMO campaign are compared with cold pools previously analyzed from RICO. The different thermodynamic and dynamic environmental conditions have effects on the average cold pool characteristics, especially the change of specific humidity and wind speed within cold pools. Despite greater surface temperature depression recorded in the DYNAMO cold pools, the cold pools seem slightly shallower and weaker. The lack of vertical wind shear in the cold pools’ generation process, and their effect on the further development of convection needs to be further examined.

This study attempts to elaborate processes associated with shallow cumulus precipitation that have been an interest of recent research efforts, but not been thoroughly understood. My approach to understand the cold pools generated by shallow
cumulus is through analyzing both in situ observation, and sophisticatedly facilitated model simulations. The results of this study not only contribute to understanding the mechanisms of cold pool associated mechanisms in invigorating further convection, but also provide reference for researchers sharing same interest with better perspective of modeling strategy.

This study may also shed some light on representing shallow cumulus cold pool process within large-scale models. A parameterization for cold pool dynamic lifting introduced by Grandpeix and Lafore (2010) and Rio et al. (2009, 2013) is based on the idea that greater expansion rates, and deeper and larger cold pools favorably induce higher updraft mass flux at the cold pool downwind boundary. Compare to mid-latitude, the tropics provide abundant moisture that is available for the cold pool enhanced updrafts. The cold pool effects on enhancing the updraft coverage and vertical velocity require consideration in GCM parameterizations of trade-wind cumulus. However, such parameterizations would need to be modified to account for the moderate depth and strength of shallow cumulus cold pools, as well as focusing on parameterizing the updraft properties and associated cloud water path rather than the cloud base mass flux. The cold pools produced by both the Thompson and Morrison scheme are still smaller and weaker than observed, but this may reflect an inclination for weak turbulence production within WRF-LES cloudy boundary layer
simulations (Yamaguchi et al., 2013; Mirocha et al., 2014). As effort increases to simulate cold pool processes more globally (e.g., Rochetin et al., 2014), more emphasis will also need to be placed on counting trade-wind cumulus cold pool processes.
References


