MICROPHYSICAL INFLUENCES ON COLD POOLS

BY

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THESIS

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ABSTRACT

Downdrafts extending from within convective clouds to the ground can produce cold pools, regions of air at the surface cooled by melting or sublimating ice and/or evaporation of rain within the downdrafts. These cold pools can propagate outward, sometimes initiating new convection along their leading edges. Models operating at scales requiring convective parameterizations usually lack a representation of this detail, and thus omit this convective regeneration and fail to predict longer episodes of convective activity (e.g. severe weather outbreaks). Recent studies have begun attempting to parameterize cold pools and the associated convection they can trigger, but a lack of understanding of the most important factors for cold pool strength, depth, and propagation speed hampers these efforts. Prior studies have investigated the influence of different hydrometeor types upon the formation of the initial cold pool but have reached drastically different conclusions.

This study uses CM1 (“Cloud Model 1”), a non-hydrostatic, fully compressible model, to produce a set of simulations in order to investigate the hydrometeor types and associated microphysical processes that are most important for determining cold pool initiation timing, strength, depth, and propagation speed. Idealized numerical simulations based upon deep convection observed on a single day during the MC3E field campaign are produced using the NSSL (6-class, double moment) microphysics scheme and a grid spacing of 250 meters. The simulations vary by altering the initial characteristics influencing warm-rain, ice processes, or secondary ice production, or the scaling factors in the underlying size distributions of hail. These simulations are all performed using the same environmental conditions.
Time-integrated microphysical budgets are calculated to quantify the contribution of each hydrometeor type (e.g. melting of graupel or hail, sublimation of graupel or hail, or evaporation of rain) to the total latent cooling occurring in the downdraft prior to the initiation of a -2K cold pool. The melting and sublimation of graupel in the downdraft dominates the integrated latent cooling terms for some runs, while the evaporation of rain dominates in others. However, the contribution from the melting or sublimation of hail is minimal.

Time-integrated microphysical budgets are also calculated to quantify the contribution of each hydrometeor type most responsible for sustaining the cold pool. Here, the latent cooling is calculated within all downdrafts that intersect it, for the 51 minutes after its initiation. Graupel sublimation always dominates the integrated latent cooling term in this case. Rain evaporation, while not dominant, is still an important contribution.

Microphysical factors affecting the initiation timing, speed, strength and depth of the cold pool through the latent cooling they promote in the downdrafts are also explored. Quickening or slowing the warm rain process respectively hastens or slows cold pool initiation. Slowing the warm rain process also limits the maximum cold pool strength by altering not only rainfall but also the amount of graupel and hail. On the other hand, the average cold pool propagation speed best correlates with the amount of latent cooling due to the sublimation and melting of graupel. The total time-integrated latent cooling best correlates with the average cold pool depth, as no single phase change term dominates that relationship.
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CHAPTER 1: INTRODUCTION AND STATEMENT OF RESEARCH OBJECTIVES

1.1 BACKGROUND AND MOTIVATION

Deep cumulus convection plays a crucial role in the energy and hydrologic balance of the Earth system. Cloud formation processes occur on scales ranging from sub-micrometer to thousands of kilometers, while atmospheric flows organize clouds into systems ranging from tens to thousands of kilometers. This range of scales presents significant modelling challenges for both cloud-resolving models (CRMs) and global climate models (GCMs). Aerosol-cloud interactions, cloud microphysics, and precipitation remain a weak point in all models due to the lack of basic understanding of the small-scale physics (IPCC 2013).

Within models, sub-grid scale processes (such as radiative transfer, microphysical processes, small-scale turbulence, and even cumulus convection in large-scale models) are represented through parameterizations. High-resolution CRMs have played an important role in improving parameterizations and have led to an increased understanding of the importance of small-scale interactions between precipitation and cloud dynamics (Kuang and Bretherton 2006), among others. Despite the progress that has been made in improving the representation of cloud processes, they remain the dominant source of uncertainty GCMs for understanding changes in the climate system (Randall et al. 2003, Arakawa 2004).

Clouds have the potential to cause significant climate feedbacks, but the sign of the feedback upon climate change (i.e. cooling or warming) depends on their properties that influence short- and long-wave radiative budgets. This in turn affects the global circulations, as regions of deep, thick clouds that have large, positive values of long-wave radiative effects have been shown to correspond to regions of precipitation, thus signifying the close connection between cloud
radiative effects and the hydrologic cycle (IPCC 2013). Accurate representation of cloud microphysical processes is crucial as it affects key aspects of the climate system such as distribution of precipitation, tropical variability, and the Hadley circulation. In particular, the representation of ice- and mixed-phase clouds in GCMs are poor due to the complexity of ice processes and the small scales at which microphysical processes can occur. As such, model sensitivity to the microphysical parameterization and its relation to the distribution and depth of these clouds exist (IPCC 2013). The parameterization of cumulus convection influences much of the variability seen in climate models.

Recently, models have begun adopting more sophisticated cumulus parameterizations. Some diagnose vertical velocities in updrafts and better representing the vertical variation of hydrometeors within convective clouds (e.g. Donner et al. 2011). Others couple shallow convection with moist boundary layer turbulence (e.g. Neggers 2009) to better represent cumuliform clouds in conditionally unstable environments. Still others include cold pools in instances of deep convection through the use of a simple model where precipitating downdrafts create a circular wake at the surface which then spreads out as a density current (e.g. Grandpré and Lafore 2010).

Convective storms also have a significant societal impact, and thus improving their prediction on shorter time scales is also important. Deep convective storms are the primary producers of hazards such as flooding, lightning, hail, severe winds, and tornadoes, all posing a threat to life and property. For the 10-year period from 2008-2017, these hazards combined to cause an average of 280 deaths a year (NOAA NWS 2017). Within this time period, there was also more then 70 flooding or severe weather events (excluding hurricanes) that produced more than $1 billion in damage (NOAA NCDC 2018).
Cold pools are one of the fundamental components of deep convection, and thus must be understood for prediction at all time scales. Cold pools form via latently-cooled downdrafts that reach the surface and spread outward as a density current (Simpson 1969, Charba 1974), displacing the surrounding warm, moist air (Goff 1976). The more buoyant environmental air is lifted along the edge of the cold front, known as the gust front or outflow boundary, and can be realized as a convective cloud (Goff 1976, Warner et al. 1980). Thus, cold pools are important for triggering new convection (Byers and Braham 1949, Purdom 1976, Weaver and Nelson 1982). Cold pools thus impact the radiative forcing of convective clouds, and modulate their role in the water cycle, by exerting control on the fundamental kinematic, thermodynamic, and diabatic processes that impact convective initiation, intensity, and lifecycle. Cold pools tend to suppress convection over their area by strongly stabilizing the lapse rate right above the ground. The propagation of their leading edge, however, can be a region of enhanced surface convergence capable of triggering new convection. While cold pools occur in both oceanic and continental environments, this study will focus on cold pools associated with land-based convection, where their societal impact is most direct, and for which more surface observations exist.

1.2 COLD POOL DEFINITIONS AND OCCURRENCE

Because there is no unified variable used for defining a cold pool in the literature (and sometimes a definition is not even explicitly identified), and because surface observations may be too sparsely-distributed to detect them, the frequency of cold pools can be difficult to ascertain. For those studies that do define it, most use a form of either potential temperature perturbation, $\theta'$ (e.g. James et al. 2006, James and Markowski 2010, Morrison and Milbrandt 2011, Van Weverberg et al. 2012, Kalina et al. 2014, Peters and Schumacher 2015) or equivalent potential
temperature perturbation, $\theta'_e$ (e.g. Dawson et al. 2010, Schlemmer and Hohenegger 2014). Of these studies, there is a split between those that use a minimum value threshold of -1 K (James et al. 2006, Dawson et al. 2010, James and Markowski 2010, Van Weverberg et al. 2012, Peters and Schumacher 2015) versus -2 K (Morrison and Milbrandt 2011, Kalina et al. 2014, Schlemmer and Hohenegger 2014). A considerable number of studies define the cold pool using variables more characteristic of the storm dynamics, such as density potential temperature perturbation, $\theta'_\rho$ as it directly relates to buoyancy (Grant and van den Heever 2014), or winds associated with the corresponding gust front (Corfidi 2003, James et al. 2005, Redl et al. 2014). Because this study primarily focuses on the thermodynamic properties of downdrafts in deep convection, a cold pool is defined as a parcel of air at the surface having a potential temperature perturbation less than or equal to -2 K or:

$$\theta'_{sfc} \leq -2 \, K \quad (1)$$

Limited observations of cold pools exist due to the sparse nature of surface observing networks. A number of studies have obtained point observations of cold pools (e.g. Goff 1976, Trier et al. 1991, Engerer et al. 2008). However, these fail to capture the evolution of the cold pool throughout the convective life-cycle. The Oklahoma Mesonet provides a source of high-resolution spatial and temporal observations of temperature, humidity, pressure, wind speed and direction, rainfall, solar radiation, and soil temperature at 120 sites across the state of Oklahoma (Brock et al. 1994). Engerer et al. (2008) used these data to investigate cold pool properties throughout the lifecycle of 39 different mesoscale convective systems (MCSs). It was seen that on average, potential temperature deficits associated with cold pools were nearly 10 K in the early stages of the storm and decreased to approximately 5 K in the dissipation stage. This decrease in potential temperature perturbation, coupled with a rise in surface pressure, led the authors to
suggest that the cold pool deepens as the MCS matures, but no observations exist to test this conjecture. Little variation was seen in the average surface wind gusts throughout the life stages of the MCSs, with all having gusts greater than 15 m s\(^{-1}\). Provod et al. (2016) used data from the African Monsoon Multidisciplinary Analysis campaign and found similar characteristics in terms of surface temperature deficits, wind gusts, and pressure increases.

Attempts have been made to identify cold pools in climatological data, numerical models, and in real-time using automated algorithms using multiple meteorological variables. Redl et al. (2015) created an algorithm to generate multi-year climatologies of convective events producing cold pools over northwest Africa. The algorithm uses surface observations of dewpoint temperature, wind speed and direction, and precipitation to detect the leading edge of the cold pool which is validated through detection of a nearby convective system using satellite-observed brightness temperature. More recently, Drager and van den Heever (2017) developed an algorithm for detecting cold pools within numerical simulations by analyzing the extremes of ten different meteorological variables. Surface rainfall rates and radial gradients in density potential temperature fields were found to be the most successful in identifying and tracking cold pools. Both of these studies, however, use arbitrary threshold values in the surface-observed variables, which can vary between environments, and neither algorithm is able to provide information on cold pool depth. They also are conducted in tropical environments which tend to have cold pools that are approximately circular in nature due to the lack of vertical wind shear. For these reasons it is not known how well these algorithms will perform in mid-latitude convective regimes which are characterized by strong wind shear, widespread precipitation, and asymmetrical cold pools.
1.3 COLD POOL INITIATION FROM DOWNDRAFTS

Because cold pools originate from convective downdrafts, the one-dimensional Srivastava downdraft model is useful to conceptualize their initiation (Srivastava 1978). Rain and ice hydrometeors contribute to negative buoyancy through precipitation loading, with larger loading being associated with larger raindrops and frozen hydrometeors (i.e. hail and graupel). Evaporation limits precipitation loading while contributing to downdraft cooling and downward-accelerating air. As the parcel descends, adiabatic warming opposes both of these processes. Adiabatic drying, i.e., the compressional warming of a descending adiabatic parcel with water vapor content held constant, allows the relatively dry downdraft to be even less dense than the surrounding environment, and accelerate downward until it reaches the surface, forming a cold pool. Srivastava found that downdrafts are stronger in environments with steep lapse rates and increased humidity, which allows for greater negative buoyancy as the air dries with descent. When ignoring ice particles, smaller raindrop sizes and larger rainwater mixing ratio at cloud base lead to stronger downdrafts by increased cooling through stronger evaporation. When including ice particles, the melting of hail and its subsequent evaporation creates the strongest downdrafts in environments with weaker lapse rates. In environments with stronger lapse rates, strong downdrafts are the product of sublimation of small, low-density ice hydrometeors. Srivastava concluded that hydrometeor properties such as phase and size play an important role in governing the formation of strong downdrafts, but their role is modulated by the environment. In this simple model, however, the respective roles of the environment versus internal storm processes (that control and determine hydrometeor properties), as well as the relationship between downdraft and cold pool properties, could not be examined.
Building off of early work from Charba (1974), Trapp (2013) gives a modified version of the equation for theoretical speed of a density current. Existing as a region of higher density air with respect to the undisturbed environment at the ground, the cold pool speed can be approximated by the theoretical speed of a density current ($V_{dc}$): 

$$V_{dc} = \sqrt{\frac{k h \Delta T_v}{T_v}}$$

(2)

where $h$ is the cold pool depth, $T_v$ is the virtual temperature, $\Delta T_v$ is the difference in the virtual temperature of the cold air and its environment (an indication of its relative density), and $k$ is a constant. The value of $V_{dc}$ determines the distance the gust front (at the leading cold pool boundary) can travel from its generating convection, and thus the horizontal extent of convective suppression within it, and the possible upscale feedback of the storm to forming new convection. Additionally, if its speed is too quick, it may undercut the inflow into the original storm, causing its demise (Engerer et al. 2008).

Downdraft size must ultimately also play a role in the cold pool properties, and multiple studies note influences upon updraft and/or downdraft sizes. Lucas et al. (1994) used boundary-layer depth to explain the greater updraft widths in continental midlatitude convection over than in tropical maritime environments Weckwerth (2000) linked a deeper boundary layer and wider updrafts, as a result of the action of horizontal convective rolls. Schlemmer and Hohenegger (2014) concluded that deeper cold pools force wider updrafts in subsequently generated storms. While wider updrafts and downdrafts are less susceptible to entraining drier ambient air, they do suffer from a reduction in buoyancy to account for the lateral displacement of air (Yuter and Houze 1995). Vertical wind shear can also promote wider updrafts and downdrafts (Marion and Trapp 2018), and also affects the overall organization of convective clouds into different convective
modes such as supercells and squall lines. Beyond this local area, convective feedbacks are comparatively small; although CAPE is eliminated in the wake of deep convection, the lapse rates are equivalent to those in the pre-storm environment (Weisman et al. 2014, Trapp and Woznicki 2017). Thus, the ability of the atmosphere to again support deep convection is mostly dependent upon how quickly the cold pool diminishes. However, convective interactions and feedbacks can also promote upscale growth into mesoscale convective systems (MCSs) such that even in a steady-state environment, draft size can evolve.

Coupling between the convective components has been studied mainly within the context and scale of individual cumuli; however, a larger-scale effect of coupling is also possible. Within the local area of active convection, adiabatic and diabatic processes modulate the temperature and humidity profiles such that the atmosphere is locally stabilized. In the wake of this local area, stabilization and elimination of instability is realized primarily through the presence of a cold pool.

1.4 COLD POOL TRIGGERING OF CONVECTION

As mentioned previously, cold pools can trigger new convection (Byers and Braham 1949, Purdom 1976, Weaver and Nelson 1982). Lift along the gust front could be due to mechanical forcing such as colliding cold pools, which are especially effective (Droegemeier and Wilhelmson 1985), or mechanical forcing due to the interaction between the cold pool and low-level wind shear (Knupp and Cotton 1982, Rotunno and Klemp 1985). The latter is typically seen in organized convective systems, such as squall lines, where it has been shown that cold pools play an important role in maintaining their longevity (Rotunno et al. 1988, Weisman and Rotunno 2004). In tropical, oceanic environments, thermodynamic forcing resulting from an accumulation of water vapor at the edge of a decaying gust front may force new convection but is not thought to be important for
land-based convection. As previously stated, Grandpriex and Lafore (2009) recently developed a parameterization of the convective cold pool for GCMs in order to generate new convection (when appropriate). While this is a positive step towards improving convective parameterizations, their approach is limited by several factors: (i) a prescribed cold pool height and radius which restricts cold pool properties such as propagation speed, and its potential to lift air ahead of it; (ii) the lack of feedback between the cold pool and land surface; (iii) the lack of influence of vertical wind shear on cold pool formation and convective triggering; and, (iv) the absence of communication of the cold pool to surrounding grid spaces, important for long-lived convective systems possibly triggering new convection over large distances. There is also no influence of the cold pool properties from the storm microphysics, despite the fact that latent cooling from hydrometeor phase changes is the origin of the cold air producing the cold pool itself, because no consensus exists regarding the most important microphysical influences, as discussed next.

1.5 DOUBLE-MOMENT MICROPHYSICS SCHEMES

Single- and double-moment microphysics schemes are the most popular options used in model simulations. Single-moment schemes predict the mixing ratio of each hydrometeor species while keeping the number concentration or mean diameter fixed so as to predict the other (Lin et al. 1983, Straka and Mansell 2005). Double-moment schemes predict mixing ratio and number concentration of each hydrometeor species while allowing for a fixed shape parameter (Ferrier 1994, Milbrandt and Yau 2005, Morrison et al. 2005, Seifert et al. 2006, Mansell et al. 2010). Single-moment schemes are computationally cheaper than their double-moment counterparts however extensive research has shown that double-moment schemes significantly outperform single-moment schemes, particularly in mid-latitude convective regimes such as supercells.
(Dawson et al. 2010, Jung et al. 2012), MCSs (Lee and Donner 2011), and squall lines (Morrison et al. 2009, Van Weverberg et al. 2012).

One of the major improvements seen with double-moment schemes is sedimentation (Wacker and Seifert 2001, Milbrandt and McTaggart-Cowan 2010, Milbrandt and Yau 2005). Despite the benefits of allowing diameter and number concentration to vary in time and space, the use of double-moment schemes presents challenges not seen with single-moment schemes due to the increased number of variables and assumptions. For instance, size sorting is able to be predicted with double-moment schemes however it has been shown to be exaggerated (Milbrandt and Yau 2005, Wacker and Lüpkes 2009). Poorly observed and understood processes such as raindrop breakup, hail shedding and melting, and Hallet-Mossop must also be parameterized in double-moment schemes but not in single moment schemes, leading to uncertainties. Despite these assumptions, it has still been shown that double-moment schemes are more successful than single-moment schemes and should be chosen whenever possible (Igel et al. 2015).

1.6 MICROPHYSICAL PROCESSES CONTRIBUTING TO COLD POOLS

Adding to the complexity of the connections between the updraft, downdraft and cold pool is the dependency of these components upon the storm microphysics. Latent heat release by hydrometeor growth in the updraft, and latent cooling from phase changes within the downdrafts, strengthens these vertical motions, and thus subsequently affect the cold pool. Numerous modeling studies (McCumber et al. 1991, Gilmore et al. 2004, van den Heever and Cotton 2004, Cohen and McCaul 2006) have shown the sensitivity of updraft and downdraft intensity to assumptions in the representation of ice phase hydrometeors, particularly graupel and hail. Hydrometeor drag upon the air also can weaken updrafts and strengthen downdrafts. Entrainment
also plays a role in influencing hydrometeor phase changes, as the mixing of drier environmental air into updrafts and downdrafts alters their relative humidity and temperature. It may be for any of these reasons that past idealized CRM simulations often produce cold pools that are less intense (Engerer et al. 2008), as well as shallower with a differing vertical structure (Bryan et al. 2005), than those observed.

Few studies have explicitly investigated the microphysical processes that influence downdrafts as they relate to cold pools or their properties. A few studies suggest that a stronger warm-rain process creates stronger downdrafts and subsequently stronger cold pools, by comparing numerical simulations with and without ice hydrometeors (Johnson et al. 1993) or in a warmer, future climate (Villanueva-Birriel et al. 2014). More thorough numerical studies of this topic, however, have yielded conflicting results regarding the importance of warm rain versus ice for cold formation.

Gilmore et al. (2004; hereafter referred to as G04) examined the sensitivity of accumulated precipitation to changes in the hail and graupel distributions and densities using a simple, single-moment microphysics scheme. Model runs are initialized for a single thermodynamic profile and three idealized shear profiles. It was found that the coldest cold pools (i.e. minimum temperature) occurred with smaller hail but the coldest area-averaged cold pools occurred in regimes of small graupel. Latent cooling calculated across the domain found that for smaller graupel, latent cooling is spread over a larger area and results in smaller rain mass that more readily evaporates and thus decreases precipitation. It was also found that downdraft strength is dependent on hail mass and distribution and that stronger near-surface downdrafts resulted in colder cold pools.

Van den Heever and Cotton (2004; hereafter referred to as VC04) also examined the impacts of the assumed mean hail diameter in a single-moment microphysical scheme upon
convective storm dynamics and precipitation. Sensitivity tests were conducted using the same environmental and initial conditions. They found that a smaller mean hail diameter led to increased melting, via surface area arguments, and subsequently increased rain evaporation, further strengthening the latent cooling that can contribute to downdrafts. They also linked the stronger downdrafts to cold pools that were deeper, more intense, and faster moving. Their results are consistent with G04’s attribution of the minimum temperature reached at the surface to small hail, but inconsistent with G04’s finding that the coldest area-averaged cold pool occurred with small graupel (with much lower density than hail).

More recently, Dawson et al. (2010; hereafter referred to as D10) compared model runs using a variety of single-moment and multi-moment microphysics schemes to analyze differences in rain evaporation and size sorting, and the subsequent effect on cold pool strength and overall storm structure. Microphysical budgets were computed for processes occurring in the region of low-level downdrafts. The model is initialized using a derived sounding from a May 1999 supercell in Oklahoma and verification is done using a combination of Doppler radar and Oklahoma Mesonet observations. In agreement with previous studies concerning only the warm rain process, D10 found that more evaporation of rain led to stronger downdrafts and subsequently stronger cold pools, but in contrast to G04 and VC04, rain evaporation was the most important process to cold pool development. D10 argued that because the multi-moment scheme allows drop size distributions and number concentrations to vary during evaporation, they are able to more accurately represent size sorting and evaporation, and thus the downdrafts contain less water mass and larger drop sizes, which limits evaporation and produces a more realistic cold pool when compared to observations.
The D10, G04, and VC04 studies have all contributed to increasing our knowledge of the link between microphysical processes and downdrafts, despite some differing conclusions. These three studies, as well as others such as Srivastava (1987), have all provided a conceptual framework including hydrometeor phase changes, the downdraft, and the cold pool. From these studies, we can conclude that: (i) melting from ice hydrometeors leads to an increase in latent cooling directly, and indirectly through increased evaporation after melting, (ii) the enhanced latent cooling within the low-to mid-levels causes stronger downdrafts, (iii) stronger downdrafts lead to stronger cold pools. The goal of these studies was to identify sources of model bias towards cold pools through microphysical parameterizations, however, rather than to learn more about the cold pool development and characteristics.

Thus, these studies have several shortcomings, and fail to provide detailed information about the cold pool that is needed in order to improve parameterizations, a major goal of this study. A major weakness among these studies is related to the choice of microphysics schemes. The use of single-moment microphysics schemes by G04 and VC04 leaves doubt regarding the accuracy of their findings (as also expressed by G04); D10 and others have found that multi-moment schemes significantly outperform single-moment schemes in convective regimes (see section 3.2.1). The representation of either graupel or hail by G04, rather than having both graupel and hail represented simultaneously, precludes a comparison of their relative importance. While the D10 and VC04 studies explicitly state their focus on changes due to the representation of rain and hail, they fail to address the impact this has on the graupel distribution, or to acknowledge the contribution of graupel to latent cooling in the downdrafts. The contribution to latent cooling from sublimation is also completely neglected in G04 and VC04, and only briefly mentioned in D10. Furthermore, while stronger downdrafts have been linked to stronger cold pools, a quantitative
relationship between downdraft properties and cold pool properties has not been established by these former studies.

1.7 RESEARCH OBJECTIVES

Modeled storms are sensitive to hydrometeor assumptions because their formation, transport, and latent cooling influence and are influenced by storm dynamics. Convectively generated cold pools are an inseparable component in the lifecycle of convective storms. Proper representation of cold pools in convection-permitting models await answers to unresolved questions regarding the complex ways in which cold pools are coupled both dynamically and microphysically to updrafts and downdrafts, as well as the atmosphere and land surface. Stemming from these complexities, questions regarding the microphysical processes most important for determining cold pool characteristics remain unanswered.

This research focuses on how microphysical phase changes affect downdrafts and cold pool properties by answering the following questions:

• What is the most important hydrometeor type(s) and associated phase change governing the downdraft that produces the initial cold pool?
• What is the most important hydrometeor type(s) and associated phase change occurring within downdrafts that sustain the cold pool?
• What is the most important hydrometeor type(s) and associated phase changes in determining cold pool characteristics such as propagation speed, depth, and strength?
CHAPTER 2: MIDLATITUDE CONTINENTAL CONVECTIVE CLOUDS
EXPERIMENT

The Midlatitude Continental Convective Clouds Experiment (MC3E) was conducted in April-June 2011 and centered near the Department of Energy’s Atmospheric Radiation Measurement (ARM) Southern Great Plains (SGP) research facility. At the time, the campaign leveraged a large ground-based observing infrastructure combined with an extensive sounding array, aircraft observations, and in-situ precipitation measurements (Jensen et al. 2014). The primary goal of the campaign was to provide a complete, three-dimensional characterization of convective clouds to provide constraints for cumulus parameterizations. As such, observations were focused on obtaining a high-resolution data set to better understand eight specific components of convective simulations and microphysical parameterizations: (i) pre-convective environment, (ii) convective initiation, (iii) updraft/downdraft dynamics, (iv) detrainment/entrainment, (v) precipitation and cloud microphysics, (vi) impact of the convective cloud system on the surrounding environment, (vii) impact of the convective cloud system on radiation, and (viii) large-scale forcing (Jensen et al. 2010). Jensen et al. (2010, 2016) provide a detailed summary of the radar, aircraft, and other in-situ observing platforms.

As this study exclusively uses numerical model simulations in order to test the stated research questions, the following section only includes information about the MC3E soundings which were used to initialize the model. MC3E conducted extensive sounding operations during the intensive observing periods (IOPs). During these IOPs, rawinsondes were launched 8 times a day from six different locations across north-central Oklahoma and southern Kansas. All sounding data were quality-controlled and bias-corrected as detailed in Jensen et al. (2014).
This study uses an observed sounding from the Purcell, OK site on 23 May 2011 to initiate the model simulations. On this day, a strong dryline coupled with surface boundaries stemming from a surface low pressure system forced convective development in the late afternoon. Strong vertical wind shear coupled with significant instability from daytime heating led to the initial development of strong supercell thunderstorms that propagated eastward, eventually merging to form a convective line (Fig. 1). Interestingly, a new Oklahoma state record for the largest hailstone (6 inches in diameter) was observed on this day (National Weather Service 2013).
3. METHODS

3.1 MODEL OVERVIEW

Cloud Model 1 (CM1) is a three-dimensional, non-hydrostatic numerical model that was designed primarily for idealized research of deep precipitating convection (Bryan and Fritsch, 2002). Total mass and energy for a moist atmosphere are conserved in the governing equations. CM1 has options to run either a compressible or incompressible set of governing equations. Additionally, this model possesses a large number of user-specified options for parameterizations including but not limited to radiation, turbulence and land-use. The specifications chosen for this research will be described in section 3.2.

The base state is assumed to be in hydrostatic balance. The model integrates the governing equations for velocity in the horizontal and vertical dimensions \( (u, v, w) \), non-dimensional pressure \( (\pi) \), potential temperature \( (\theta) \), and mixing ratios of three moisture variables- water vapor \( (q_v) \), liquid water \( (q_l) \), and ice \( (q_i) \) (if more detailed water categories are used, as described in Section 3.2, the hydrometeor masses within each liquid or ice category are simply combined for inclusion in the general equations below). A general equation for a variable \( \alpha \) can be defined as: \( \alpha(x, y, z, t) = \alpha_0(z) + \alpha'(x, y, z, t) \), where the subscript 0 denotes the base state values and is only a function of height \( (z) \), and perturbation from the base state is represented with a prime superscript.

The governing equations for velocity are:

\[
\frac{\partial u}{\partial t} + c_p \theta \frac{\partial \pi'}{\partial x} = \text{ADV}(u) + f v + T_u + D_u + N_u \tag{3}
\]

\[
\frac{\partial v}{\partial t} + c_p \theta \frac{\partial \pi'}{\partial y} = \text{ADV}(v) - f u + T_v + D_v + N_v \tag{4}
\]
\[ \frac{\partial w}{\partial t} + c_p \theta_p \frac{\partial \pi'}{\partial z} = ADV(w) + B + T_w + D_w + N_w \]  

(5)

where $T$ represents tendencies from sub-grid turbulence, $D$ represents diffusion, and $N$ represents Newtonian relaxation or Rayleigh dampening. $ADV$ is the advection operator given with a generic variable $\alpha$ as:

\[ ADV(\alpha) = -u \frac{\partial \alpha}{\partial x} - v \frac{\partial \alpha}{\partial y} - w \frac{\partial \alpha}{\partial z} \]  

(6)

and $B$ is buoyancy given as:

\[ B = g \frac{\theta_p - \theta_{\rho_0}}{\theta_{\rho_0}} \]  

(7)

The governing equations for the moisture components are:

\[ \frac{\partial q_v}{\partial t} = ADV(q_v) + T_{qv} + D_{qv} - q_{\text{cond}} - q_{\text{dep}} \]  

(8)

\[ \frac{\partial q_l}{\partial t} = ADV(q_l) + T_{ql} + D_{ql} - q_{\text{cond}} - q_{\text{frz}} + \frac{1}{\rho} \frac{\partial (\rho V_l q_l)}{\partial z} \]  

(9)

\[ \frac{\partial q_i}{\partial t} = ADV(q_i) + T_{qi} + D_{qi} + q_{\text{dep}} + q_{\text{frz}} + \frac{1}{\rho} \frac{\partial (\rho V_i q_l)}{\partial z} \]  

(10)

where the $\dot{q}$ terms represent the phase changes between vapor and liquid (subscript “cond”), vapor and ice (subscript “dep”), and liquid and ice (subscript “frz”).

The governing equation for $\theta'$ is:

\[ \frac{\partial \theta'}{\partial t} = ADV(\theta) - \Theta_1 \theta \left( \frac{\partial \theta'}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} \right) + T_{\theta} + D_{\theta} + N_{\theta} + \Theta_2 (L_v q_{\text{cond}} + L_s q_{\text{dep}} + L_f q_{\text{frz}}) + \Theta_3 (L_v q_{\text{cond}} + q_{\text{dep}}) + \Theta_2 \varepsilon + \dot{Q}_{\theta} + W_T \]  

(11)
where $Q_\theta$ represents external tendencies to internal energy, primarily radiative heating and cooling. $\Theta$ is dependent on user input (for the variable eqtset) on whether to have the equation set mathematically conserve mass and energy in moist environments. This study opts to conserve mass and energy (eqset = 2), so the equations used for $\Theta$ are:

$$\Theta_1 = \left( \frac{R_m}{c_{vm}} - \frac{Rc_{pm}}{c_p c_{vm}} \right)$$

$$\Theta_2 = \frac{c_v}{c_{vm} c_p \pi}$$

$$\Theta_3 = -\theta \frac{R_v}{c_{vm}} \left( 1 - \frac{Rc_{pm}}{c_p R_m} \right)$$

The governing equation for $\pi'$ is:

$$\frac{\partial \pi'}{\partial t} = ADV(\pi) - \Pi_1 \pi \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} \right) + \Pi_2 \left( L_v q_{\text{cond}} + L_s q_{\text{dep}} + L_f q_{frz} \right)$$

$$+ \Pi_3 (q_{\text{cond}} + q_{\text{dep}}) + \Pi_4 \left( T_\theta + D_\theta + N_\theta + \Theta_2 e + Q_\theta + W_T \right)$$

$$+ \Pi_4 (T_{qv} + D_{qv})$$

where the equations for $\Pi$ are determined similar to those of $\Theta$ and are given by:

$$\Pi_1 = \frac{R}{c_p} \frac{c_{pm}}{c_{vm}}$$

$$\Pi_2 = \frac{R}{c_p} \left( \frac{1}{c_{vm} \Theta} \right)$$

$$\Pi_3 = -\frac{R}{c_p} \left( \frac{R_v c_{pm}}{R_m c_{vm}} \right)$$
$$\Pi_4 = \frac{R \, \pi}{c_v \, \theta} \quad (19)$$

$$\Pi_5 = \frac{R \, \pi}{c_v \, \epsilon + q_v} \quad (20)$$

where:

$$c_{pm} = c_p + c_{pv} q_v + c_i q_l + c_i q_l \quad (21)$$

$$c_{vm} = c_v + c_{vv} q_v + c_l q_l + c_l q_l \quad (22)$$

$$R_m = R + R_v q_v \quad (23)$$

and the variable \(c\) represents the specific heat and the subscripts denote the hydrometeor (\(i\) representing ice, \(l\) representing liquid water, \(v\) representing water vapor, and no subscript representing dry air) and if the variable is isobaric (denoted by subscript \(p\)) or isochoric (denoted by subscript \(v\)). \(R\) is the gas constant with the same subscript convection as noted with \(c\).

3.2 MODEL SETUP

3.2.1 General specifications

Table 1 lists the model configuration used in this study. Variables denoted with an asterisk are defined within the namelist input file. The variable names defined in the “init3d” subroutine are listed in parentheses. All model runs use the listed specifications except where noted. The differences in model runs are discussed in section 3.3.

The model is run at high resolution both spatially and temporally. Grid spacing is 250 meters in both the horizontal and vertical with a 3 second time step. Data are output every 30 seconds. Run time for each simulation is 205 minutes, except in two instances where it was extended to 250 and 305 minutes (see section 3.3). Sub-grid scale turbulence uses the TKE scheme described in Deardorff (1980). Radiation is turned off in order to isolate the differences in cold.
pool properties between model runs to differences in the storm microphysics. A Rayleigh damping zone extending from 17 km to 20 km (top of the domain) was applied to eliminate the creation of artificial downdrafts created via strong upward vertical motions interacting with the top rigid boundary, as well as to eliminate spurious gravity waves. Horizontal boundary conditions are open (radiative). The lower boundary condition is without friction (i.e. free-slip) for simplicity; thus the cold pool propagation speeds should be considered upper limits for a given simulation.

As stated earlier, the model base state is initialized with the sounding observed from the Purcell, OK site on 23 May 2011 at 2030 UTC from the MC3E field campaign. Slight modifications were made to the boundary layer in order to remove convective inhibition (CIN) and allow convection to develop. Figure 2 shows the sounding used for all simulations presented in this study. The initial storm in the simulations was initialized with a warm bubble as in Klemp and Wilhelmson (1978) with specifics given in Table 1. All subsequent convection develops from the outflow of this initial storm.

3.2.2 NSSL Microphysics Scheme

In contrast to past studies, a major advantage of this study is that a single microphysics scheme is used for all model runs, with just certain parameters altered. Thus, the underlying microphysical assumptions remain the same, and any differences in storm characteristics between model runs is purely due to changes in these parameters.

This study uses the double-moment, 6-class NSSL microphysics scheme (Mansell et al. 2010). The hydrometeor species included are: cloud water, rain, ice, snow, graupel, and hail. The inclusion of both graupel and hail as separate categories in this scheme, as well as the prediction of graupel and hail density at each grid point that more accurately represents fall speed estimates,
makes it preferable for this study. A detailed explanation of the physical processes represented, and corresponding equations, can be found in Appendix A in Mansell et al. 2010.

3.3 GENERATION OF MULTIPLE REALIZATIONS

This study uses a set of 10 simulations to investigate the importance of hydrometeor types and associated microphysical processes for determining cold pool characteristics. This set of simulations produces variability in storm characteristics resulting purely from microphysical differences and is designed to help find relationships between the microphysical processes, and the downdrafts and subsequent cold pools. All runs are initiated using the same environmental profile (Figure 2) and microphysics scheme (section 3.2.2). Only slight alterations to the initial characteristics influencing precipitation processes were made: (i) the number of CCN, affecting the warm-rain process, (ii) the shape parameter of the hail distribution, altering the width of its underlying size distribution, and (iii) the number of ice-nucleating particles (INP), and disabling rime-splintering and immersion freezing, all altering ice processes. Table 2 shows the differences in initial characteristics between the model runs.

The control run uses a cloud condensation nuclei (CCN) value of 700 cm$^{-3}$ that was deemed reasonable for the Oklahoma area. The warm-rain process is altered by changing the initial amount of CCN. Smaller CCN concentrations produce a faster warm-rain process; drops grow larger due to less “competition” for available water vapor and thus expedite the collision-coalescence process. For larger CCN concentrations, drop size decreases and the collision-coalescence process is delayed. As such, the model run time had to be extended for the cases with increased CCN concentrations in order to see the formation and development of the cold pool. While the goal of modulating the warm-rain process is to analyze the influence of liquid
water, it is also necessary to consider the impact this change has on the production of ice hydrometeors. Smaller drops associated with increased CCN concentrations can be lofted higher in the storm, becoming supercooled and potentially creating additional graupel and/or hail.

Changes in the hail distribution were made by altering the shape parameter in the namelist input file. Mansell et al. (2010) relate the shape parameter to particle distribution in terms of diameter by:

\[
N(D) = 3A \frac{D^{\alpha}}{D^{\alpha+1}} \exp \left[ -B \left( \frac{D}{\bar{D}} \right)^{3\mu} \right]
\]

where:

\[
A = \frac{\mu N_t}{\Gamma \left( \frac{\nu + 1}{\mu} \right)} \left( \frac{\nu + 1}{\mu} \right)^{(\nu+1)}
\]

\[
B = \left[ \frac{\Gamma \left( \frac{\nu + 1}{\mu} \right)}{\Gamma \left( \frac{\nu + 2}{\mu} \right)} \right]^{-\mu}
\]

and \( \nu \) is the shape parameter in terms of volume, \( \mu \) is the exponent in the size distribution, \( D \) is the diameter, and \( \alpha \) is the shape parameter in terms of diameter. The bar indicates the mean of the variable. The graupel shape parameter is set to zero for all runs. Changes to the hail distribution were made for two simulations to create a broader and narrower hail distribution, by setting the shape parameter to 0.2 and 0.9, respectively. For all other runs the hail shape parameter is 0.5. Figure 3 shows the effect of these changes on the size distribution for a hail mixing ratio of 4 g kg\(^{-1}\) and total number concentration of 100 cm\(^{-3}\). The solid black line is equivalent to a negative-exponential distribution (i.e. Marshall-Palmer) and is included for
reference. It is expected that the broadening of the hail distribution will produce a larger variety of hailstone sizes and result in a variety of fall speeds and size sorting, whereas with the narrow distribution the similarity of hailstone sizes would lead to hail falling at the same rate.

Changes in ice processes are achieved through changes in the initial amount of INP: INP were increased or decreased by a factor of 10 for different simulations as noted in Table 2. The expectation is that less ice would translate to fewer graupel and potentially more hail. After initial testing between these runs, the decrease in ice expected was not seen. It is hypothesized that the model compensated for the fewer INP by other primary or secondary ice processes. To severely limit the amount of ice being produced, two additional simulations were run for the low INP case, with either immersion freezing turned off or both immersion freezing and rime-splintering turned off.

3.4 MICROPHYSICAL BUDGET CALCULATIONS

Similar to D10, this study calculates microphysical budgets to evaluate the relative quantity of latent cooling contributed by each hydrometeor. The latent cooling terms calculated in this study are: (i) graupel sublimation, (ii) graupel melting, (iii) hail sublimation, (iv) hail melting, and (v) rain evaporation. These are calculated by:

\[ L_x dq = c_p dT \]  \hspace{1cm} (27)

where \( L_x \) represents the latent heat due to the phase change and is a function of pressure and temperature, \( dq \) represents the cooling, \( c_p \) is the specific heat of air in terms of constant pressure,
and \( dT \) is the change in air temperature. Since cold pools are defined using \( \Theta \), the equation can be manipulated using the relationship:

\[
T = \theta \pi
\]

(28)

where \( \pi \) is the Exner function:

\[
\pi = \left( \frac{P_0}{P} \right)^{\frac{R_d}{c_p}}
\]

(29)

So, output values can be expressed as a change in potential temperature. Values are multiplied by -1 to be reported as positive quantities (i.e., \( d\theta \)), and can also be transformed into Joules (J) using an alternate form of the latent cooling equation 27:

\[
d\theta = \frac{L_x}{\pi c_p} dq_x
\]

(30)

inserting the full form of the Exner function (equation 29) yields:

\[
L_x dq_x = d\theta \cdot c_p \left( \frac{P}{P_0} \right)^{\frac{R_d}{c_p}}
\]

(31)

where \( p \) is total pressure, \( p_0 \) is a constant pressure of \( 10^5 \) Pa, and \( R_d \) is the dry gas constant. The latent cooling output is then multiplied by grid box density and volume to transform into Joules. Values can be interpreted as Joules of heat extracted by the phase change.
4. RESULTS

4.1 MODELED STORM DYNAMICS AND MORPHOLOGY

A series of tests was conducted to investigate if any changes made in the initial microphysical conditions drastically altered the storm dynamics. Simulated near-surface reflectivity and cold pool were analyzed for the duration of each run. All simulations began as a supercell thunderstorm before transitioning into a multicell cluster with a large, sweeping cold pool along its leading edge. Figure 4 illustrates the reflectivity and cold pool for each model run near the end of the simulation. The modeled storm evolution similarly follows MC3E observations for a subset of storms that initialized in north-central Oklahoma on 23 May 2011, further verifying the ability of the model to accurately resolve convection.

Vertical velocity extrema were also compared for each run to investigate any changes in storm dynamics. Figure 5 shows a time series of maximum updraft and downdraft velocities below 5 kilometers altitude from the beginning of the simulations until 205 minutes. More variance is seen in the downdrafts as time progresses, indicating the variance is due solely to microphysical effects, and further demonstrating the sensitivity of models to hydrometeor assumptions and the interconnections between microphysical processes and storm dynamics. The maxima for simulations run beyond 205 minutes (ModerateWR and SlowWR) did not show appreciable variation from that shown in Fig. 5. Thus, the variations to the model microphysics was successful in creating 10 different realizations of the 23 May case that did produce some differences in downdraft strengths that should be useful in diagnosing the effects on cold pools. Differences in model runs as they relate to the proposed research questions can be evaluated.
confidently, given that the general trends in storm dynamics and evolution remain relatively unchanged among them.

4.2 TIMING OF COLD POOL FORMATION

As previously stated, the first modeled storms began as supercells, during which time the cold pools produced were shallow, spatially small, and remained stationary or propagated opposite the storm motion. Since one of the underlying motivations of this study is to better understand how cold pools can trigger new convection, the small, transient initial cold pool in each simulation was not analyzed. The cold pools used for analysis in this study reach appreciable depths, spatial extent, temperature perturbations, and propagate with the storm in such a way that the leading edge of the cold pool can serve to initiate new convection (e.g. Fig. 4).

Identifying the start of the cold pool was done objectively using low-level vertical velocity in tandem with surface potential temperature perturbation from the environmental value. Working backwards in time from the end of the simulations where the cold pool is largest, the time and location at which a low-level downdraft initiated the cold pool were identified. An increase in cold pool area is observed shortly after the cold pool start time for each simulation and serves to validate these start times (Fig. 6). The ten different simulations produce cold pools of varying sizes and growth rates.

The start time of the cold pool for each simulation is presented in Figure 7. Although the cold pool begins at nearly the same time (between 100 and 150 min) in most of the simulations, a trend between cold pool initiation and the speed of the warm-rain process is seen. A faster warm-rain process produced an earlier start to the cold pool, whereas a slow-warm rain process
significantly delayed its onset. No clear relationships emerged between the cold pool initiation time and any other microphysical differences among the simulations. The simulations thus suggest that the warm-rain process plays a significant role in cold pool formation, although it is unclear at this point if this role is due to enhancement of rain evaporation, or some feedback into graupel or hail production (and thus their associated phase changes).

4.3 MICROPHYSICAL INFLUENCES UPON COLD POOL FORMATION

This section will address the following research question—what are the most important hydrometeor type(s) and associated microphysical processes governing the downdraft that produces the initial cold pool? Whereas previous studies looked at the microphysical processes over the entire convective system and assumed they were representative of the local area which first forms cold pools, this study specifically evaluates the microphysical processes contained within the single downdraft that initiates the cold pool.

For this analysis, latent cooling budgets are calculated for each of the five phase changes possibly contributing to cold pool formation (graupel melting/sublimation, hail melting/sublimation, and rain evaporation) within a subset of the domain ten minutes prior to the formation of the cold pool. This subset of the domain ranges between 200 and 560 square kilometers and was subjectively determined to encompass the strongest downdraft at the lowest model level that was persistent for the entire 10-minute period leading to cold pool formation. The ten-minute time frame is chosen based on the transient nature of the downdrafts feeding the initial cold pool. The vertical extent of the sub-domain over which the budgets are calculated is
capped at 5 km; the melting level exists at approximately 3.5 km in the model sounding so this altitude will include all melting as well as possibly sublimation of frozen hydrometeors.

Based on previous research summarized in Chapter 1, the strongest latent cooling should be contained within the strongest downdrafts, to form the strongest cold pools. This principle is applied into the framework of this analysis, but with a small caveat. During the specified time-frame, differences in downdrafts contributing to the initial cold pools among the model runs varied between 5 and 15 m s\(^{-1}\). In order to make a fair comparison across all the simulations, latent cooling was integrated within the less stringent downdraft threshold of 5 m/s. D10 used a similar method by integrating latent cooling within the 0.5 m/s downdrafts below 4 km for two different 30-minute time periods but performed the calculations across the entire model domain. The method used in this study make it superior to that of D10 because it specifically focuses on the latent cooling that occurs within the single downdraft that causes the formation of the cold pool, and does not assume that storm-scale microphysical processes necessarily represent properties of individual downdrafts.

To enable comparison among the different model runs, the integrated latent cooling is normalized by the volume of the downdraft, to account for possible differences in downdraft sizes, using the equation:

\[
\Delta \theta_{h,LC,t} = \sum_i \frac{\Delta \theta_i \cdot V_i}{V_{DD}}
\]

where \(\Delta \theta_{h,LC,t}\) represents the change in potential temperature due to the latent cooling \(LC\) of hydrometeor \(h\), at time \(t\), and \(V\) represents volume, with \(i\) denoting the number of grid boxes where latent cooling is occurring and \(DD\) denoting the downdraft.

To summarize, latent cooling is integrated 10-minutes prior to the formation of the cold pool in a subset of the model domain, within the 5 m/s downdrafts below 5 km, and normalized
by the volume of the downdraft. As such, the computation of latent cooling is proportional to the change in mass of each hydrometeor. Figure 8 shows the result of these calculations, accumulated over time, since the effects of latent cooling are not instantaneously transferred to the surface cold pool, but rather build in time.

As shown in the first row of Fig. 8, when the warm-rain process is slowed, the amount of graupel melting and sublimating with the downdraft contributing to the cold pool initiation increases, and the contribution of the evaporation of rain in the downdraft decreases. Comparing the ModerateWR and SlowWR cases, the rain evaporation term does not decrease at the same rate as between the FastWR and ModerateWR, because some of the rain evaporation results from the additional graupel melting. An increase in the contribution from hail also increases as the warm-rain process is slowed. As the hail size distribution is changed (Fig. 8, 2nd row), there is a slight increase (BroadHail) and decrease (NarrowHail) in its contribution to the integrated cooling in the downdraft. More notably from those experiments, the contribution from graupel changes significantly: a broad hail size distribution decreases its cooling effects in the downdraft, while a narrow size distribution increases it. It is hypothesized that the increase in graupel seen with the narrow hail distribution is the result of less competition with the smaller hail hydrometeors for collecting the available mass of supercooled water. When ice-nucleating particles were decreased in the simulations by various degrees (Fig. 8 bottom row), the contributions from the frozen hydrometeors (graupel and hail) decreased by varying degrees. In the most extreme case (LessIN_IFH Moff), the contribution of hail melting in the downdraft increases back to the control values, as it likely has much less competition for growth with other frozen hydrometeors collecting cloud water.
Figure 9 shows the contribution of each process to the total latent cooling to summarize these trends. There is variation among the runs regarding the most important cooling term for the initiation of the cold pool: graupel and its associated latent cooling dominates in some instances (sublimation ranging from 10 to 45%; melting ranging from 22-37%), while evaporation of rain is most important in other cases (15-64%). The contribution from hail is minimal in all model runs (less than 15%). Figure 9 also highlights a successful attempt at generating simulations with similar dynamics while having good microphysical variability in order to compare how variances in microphysics affects the cold pool.

4.4 HYDROMETEOR PROPERTIES AND MICROPHYSICAL PROCESSES SUSTAINING COLD POOLS

This section addresses the research question- what is the most important hydrometeor type(s) and associated microphysical processes occurring within the downdrafts that sustain the cold pool? This question has rarely been addressed in the literature; instead, focus has been placed upon the mechanisms forcing the initial cold pool. However, understanding the hydrometeor and microphysical processes that sustain the cold pool may provide more information useful for predicting cold pool properties, and how the cold pool can initiate new convection.

Each latent cooling process is calculated from the start time of the cold pool until 51 minutes afterward, the maximum amount of time over which all the simulations were conducted. In order to search the model domain but limit calculations only within downdrafts that influence the cold pool, a new algorithm was devised. First, at each time period for when a cold pool exists at the surface, latent cooling is calculated only within the 1 m/s downdraft(s) at the lowest
vertical level that directly contact the cold pool. Then, latent cooling is calculated in adjacent grid boxes (horizontal and vertical) where a 1 m/s downdraft exists. The algorithm continues searching additional vertical levels, moving upward for contingent grid boxes (touching those identified as downdraft at the level below) that occupy the downdraft with a threshold of 1 m/s. The calculation of the latent cooling continues until the vertical level at which the downdrafts (touching the cold pool surface) no longer extend. Unlike in section 4.3, latent cooling is not volume-normalized. Instead, the units for latent cooling are transformed into Joules (J) using an alternate form of (31):

$$d\theta = \frac{L_x}{\pi c_p} dq_x.$$  \hspace{1cm} (35)

Inserting the full form of the Exner function (33) yields:

$$L_x dq_x = d\theta \cdot c_p \left(\frac{p}{p_0}\right)^{\frac{R_d}{c_p}}.$$  \hspace{1cm} (36)

where $p$ is total pressure, $p_0$ is a constant pressure of $10^5$ Pa, and $R_d$ is the dry gas constant. The latent cooling calculated at each grid point is then multiplied by its air density and volume to transform into Joules of energy. Values can be interpreted as Joules of heat extracted by the phase change.

Figure 10 shows the integrated latent cooling using the algorithm for each model run over the first 51 minutes of their respective cold pools. Unlike Figure 8, there is now no variation between runs regarding the dominant latent cooling process; for all model runs graupel sublimation is the dominant term, consisting of approximately 50% of the total latent cooling in the downdrafts sustaining the cold pools. The other contributors in decreasing order of importance are rain evaporation (approximately 30% of total latent cooling), graupel melting (approximately 15% of total latent cooling), and hail melting and sublimation (combined,
approximately 5% of the total latent cooling). While the relative magnitudes of individual cooling terms differ among the 10 realizations, the importance of graupel sublimation appears to always be paramount, at least in this particular storm environment. The differences in these results between cold pool initiation (Fig. 8) and cold pool sustainment (Fig. 10) are interesting and warrant additional investigation in the future, and also question the findings of past studies that investigated storm-wide microphysical effects upon cold pool initiation and did not consider how they might evolve while sustaining the cold pool.

4.5 VARIATIONS IN COLD POOL PROPERTIES

4.5.1 Area and Propagation Speed

As previously noted, an increase in cold pool area is seen shortly after the start time of the cold pool for all cases (Fig. 5). An analysis of the total cold pool area 51 minutes after its start for each model run (Fig. 12) shows no discernable trends indicative of particular microphysical processes causing a more or less expansive cold pool.

However, examining the rate of the cold pool expansion (essentially the slope of the curves shown in Fig. 6), which can be representative of the propagation speed of the cold pool, does shown some dependency upon particular latent cooling terms. Cold pool area is used in calculating the propagation speed through the equation:

\[ v = \frac{\Delta \sqrt{A}}{\Delta t} \]  

(37)

where \( v \) represents the propagation speed (in m/s), and \( A \) is the area of the cold pool at time \( t \).

The storm-average propagation speed is calculated by averaging (37) from the start time of each cold pool until 51 minutes afterward at 2- minute intervals. Average propagation speeds range from 3.9 to 9.8 m/s (Fig. 13), with no significant relationships among the different values for
simulations where the warm rain process or ice processes were altered. A slight increase in average propagation speed is seen with a narrow hail distribution (versus a broad distribution), and with more IN than less IN, however these differences are not significant. There is very little variation in the 2-minute maximum propagation speed during the first 51 minutes of the cold pools between the different simulations (Fig. 13), except for the LessIN run having a maximum speed approximately 4 m/s greater than all other runs.

4.5.2 Depth

Cold pool depth is calculated using a bottom-up approach, start at the lowest model grid level and searching upward to find the height where the potential temperature perturbation first falls below -2K. Its accuracy is thus limited by the grid resolution to increments of 250 meters, and could be overestimated by nearly this amount. Air that is being latently-cooled by -2K or more within downdrafts will be included by this approach, and thus produce a positive bias in cold pool depth values. While it is possible to exclude grid boxes where the -2K \( \theta' \) surface co-exists within a downdraft, doing so would exclude portions of the cold pool that have direct interactions with low-level downdrafts. Thus, the determination of cold pool depth is not straightforward, and these limitations must be kept in mind.

A time series of the spatially-averaged cold pool depth for the first 50 minutes is shown in Figure 14. It can be seen that while the trend is for average depth to increase in time, significant fluctuations can exist. Some runs build up to a single maximum depth and then decrease in time, while others encounter multiple peaks throughout this time period. The peaks in the average depth may represent times where cold pool depth was increased by more (or larger) cool downdrafts contributing to the cold pool, but this has not been examined. However,
the small variability among the simulations of the temporally and spatially-averaged depth for
the 51-minute period after the start of the cold pool (Fig. 15), and the maximum depth that
occurs at any time during this period (Fig. 16), would appear to indicate that changes in
hydrometeor and microphysical processes do not significantly affect cold pool depth. Thus,
storm dynamics are likely playing a dominant role in determining the cold pool depth.

4.5.3 Strength

“Cold pool strength” is an ambiguous variable, when looking through the literature where
the term is often used, but no definition of what it constitutes is provided. Usually, stronger cold
pools are those that have colder surface values of potential temperature perturbation (or the
chosen variable for representing a cold pool, refer to section 1.1). This convention is not useful
for this study, however, as all model runs reach a minimum cold pool temperature of -10K. As
such, a new metric was developed for representing cold pool strength. Here, the fraction
(expressed as a percent) of the -2K cold pool area occupied by the -6K area is used:

\[
\frac{Area_{CP,-6K}}{Area_{CP,-2K}} \cdot 100
\]

and can be thought of as representing the percentage of the total cold pool area that is “strong”.
Various considerations influenced the decision to use the -6K definition as the “strong” cold
pool. While all runs reach minimum cold pool temperatures of -10K and -6K, most do not do so
in the 51 minutes after formation for the former, and the percentages are often less than 1% for
the latter. All runs reach a minimum \( \theta' \) value of -6K for the majority of the 51-minute time
frame analyzed here, and thus is useful as the threshold for which to calculate strength.

The maximum strength achieved by each model run in the 51-minute time frame after the
cold pool form is quite variable among the 10 simulations (Fig. 17). Both warm rain and ice
processes appear to have an influence, here. Interestingly, the speed of the warm rain process, which was found to have a positive relationship to the speed of cold pool formation, here appears to be detrimental to its strength. Even greater differences in strength is seem among the simulations with varying amounts of ice, however. More IN appears to correspond to a stronger cold pool, whereas when ice nucleation is severely limited (such as the LessIN_IFoff and LessIN_IFHMoff cases) the cold pool fails to reach appreciable strength. Although not quantified here, these trends may be due to effects of having more numerous and smaller ice particles in the high IN case (yielding more surface area of particles for greater latent cooling rates) and the converse for the when ice nucleation is limited. This effect too might explain the trend for the simulations where the rate of the warm rain process was changed, as the slower-warm rain process may thus favor graupel production and correspond to a stronger cold pool. A narrow hail distribution also yields more, smaller hail, increasing the surface area and thus potentially latent cooling rates. Therefore, for cold pool strength as it is defined here, the amount of ice appears to play a dominant role in determining the maximum cold pool strength.

4.6 THE INFLUENCE OF HYDROMETEORS ON COLD POOL CHARACTERISTICS

This section will analyze the relationship between latent cooling in the downdrafts (either due to a particular hydrometeor phase change, or to the total latent cooling in the downdrafts) and the cold pool properties discussed in sections 4.5.1 through 4.5.3. Here one latent cooling term is used for each hydrometeor type, i.e., melting and sublimation are combined for graupel or hail. Total latent cooling is the latent cooling due to phase changes of all three hydrometeors (graupel, hail, and rain). Integrated latent cooling within the downdrafts sustaining the cold pool is calculated over the first 10, 20, 30, 40, and 50 minutes after the start of the cold pool using the
method described in section 4.3. The integrated latent cooling is then compared to various cold pool properties occurring within the same time period. For all properties, the relationship gradually improves for the longer time scales. As such, only plots showing the integrated latent cooling and cold pool properties in the 50 minutes since the start of the cold pool will be shown and discussed.

The correlation between the average propagation speed and the integrated total latent cooling (Fig. 18) is a strong one (r-value of 0.87); clearly more latent cooling in the downdrafts would be associated with a faster propagating (expanding) cold pool. Looking at the relationship between the average propagation speed and the individual hydrometeor latent cooling terms (Fig. 19), this strong relationship results from the fact that the latent cooling due to graupel is dominating the latent cooling in the downdrafts (Fig. 10), and thus is itself a good predictor of the average propagation speed (r-value of 0.79). However, the correlation is slight higher (r-value of 0.94) between rain evaporation and average propagation speed. While the latent cooling of graupel dominates the latent cooling in the downdrafts throughout the storm, the amount of rain evaporation is the dominant process that influences average propagation speed. This is hypothesized to be due to the fact that the latent cooling due to the evaporation of rain in the downdrafts occurs nearest in altitude to the cold pool (or even within), and thus its effects are more pronounced. There is not enough variation in the maximum propagation speeds between the simulations (Fig. 13) to discern its relationship with the latent cooling.

Similar to the average propagation speed, there is a strong relationship (r-value of 0.88) between total latent cooling and the average cold pool depth (Fig. 20), indicating that more latent cooling results in cold pools that are, on average, deeper. There are also relatively high correlations between the individual hydrometeor latent cooling terms (Fig. 21), with graupel
having the best relationship (r-value of 0.88) followed by rain (r-value of 0.79) and hail (r-value of 0.72). The similarity in these relationships indicates that no single hydrometeor type stands out as more influential in determining cold pool depth.

Total latent cooling versus 50-minute average cold pool strength (Fig. 22) only has a moderate relationship (r-value of 0.56) between total latent cooling and cold pool strength, due to many of the runs having very weak cold pools. Similar r-values are seen relating the average strength to individual hydrometeor latent cooling terms (Fig. 23). A comparison between maximum strength and total latent cooling is seen in Figure 24 and similarly shows a modest relationship. It can be concluded that the total integrated latent cooling is not a good of an indicator of cold pool strength as formulated here; additional study is required to ascertain what variables are.

The relationship between latent cooling within the downdraft forming the cold pool (described in section 4.2) and cold pool properties at various time-scales was also evaluated. There was no relationship between the two, regardless of the chosen time scale or cold pool property. This demonstrates that the properties of the downdraft that form the cold pool do not influence the subsequent cold pool properties.
5.1 CONCLUSIONS

Due to the complexities surrounding precipitation processes, an understanding of how cold pools are coupled to the storm dynamics and microphysics has remained elusive. Previous studies have been inconclusive in their findings and have primarily focused on differences in the large-scale microphysical processes as they relate to different microphysics schemes. As such, questions remain regarding the role of microphysical processes in determining cold pool characteristics. Finding the answers to these questions could improve the representation of cold pools and cold pool-induced convection in numerical models.

Using a set of 10 model runs with varying hydrometeor characteristics, the dominant microphysical processes within downdrafts forming and sustaining the cold pool are identified as well as the relationship between latent cooling and cold pool properties. Within downdrafts forming the cold pool, variance exists among the dominant hydrometeor phase changes governing the cooling. In some simulations, graupel dominates and in others, rain. In all runs, there is minimal contribution from hail. None of this variance is seen within downdrafts sustaining the cold pool, however. Here, graupel sublimation is the dominant process, followed by rain evaporation and graupel melting, with minimal contribution from melting or sublimation of hail.

The speed of the warm-rain process significantly influences the start time of the cold pool, with a slower-warm rain process delaying its onset. The average cold pool propagation speed and depth show a strong, positive relationship to total latent cooling, especially at longer time scales. Rain evaporation also correlates extremely well with the average propagation speed,
whereas no single hydrometeor phase change alone appears to control the cold pool depth.

Using a novel method to define cold pool strength, only a moderate relationship exists with any type of latent cooling occurring in the downdrafts.

5.2 STUDY LIMITATIONS AND FUTURE WORK

There are numerous limitations to the current study. A single environment was used for all of the simulations, and thus the dominant latent cooling terms found here may not be applicable to other, or even most, convection. In addition, because of the large number of simulations that was required to produce the variability needed to find relationships among the variables, performing the entire set of simulations again for other microphysical schemes was impractical. However, such a study needs to be performed to ensure that the findings here are not specific to the microphysical scheme employed. In addition, other cases also need to be simulated with different dynamics (say in environments with different CAPE or vertical wind shear), to understand the relative importance of the storm dynamics (affecting the downdrafts) and the microphysics themselves. Finally, since the proposed method of quantifying cold pool strength did not yield very clear results, it would also be beneficial to explore other variables and ways in which to create a useful metric of cold pool strength.
CHAPTER 6: FIGURES AND TABLES

Figure 1: Observed surface temperature from the Oklahoma Mesonet overlaid with radar reflectivity at 00 UTC 24 May 2011
Figure 2: Derived Skew-T from the observed sounding data from Purcell, OK on 23 May 2011 that was used to initiate the model
Table 1: CM1 model configuration used in this study

<table>
<thead>
<tr>
<th>Attribute</th>
<th>Namelist variable</th>
<th>Value</th>
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<td><strong>Time:</strong></td>
<td></td>
<td></td>
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<tr>
<td>Run Time</td>
<td>run_time</td>
<td>205 minutes (extended to 250 minutes and 305 minutes for certain runs; see section 3.3)</td>
</tr>
<tr>
<td>Time Step</td>
<td>dtl</td>
<td>0.5 minutes (30 seconds)</td>
</tr>
<tr>
<td>Acoustic Time Step</td>
<td>nsound</td>
<td>0.05 minutes (3 seconds)</td>
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<tr>
<td><strong>Domain:</strong></td>
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<tr>
<td>Domain Size</td>
<td>nx, ny, nz</td>
<td>250 x 250 x 20 km</td>
</tr>
<tr>
<td>Grid Spacing</td>
<td>dx, dy, dz</td>
<td>250 meters (horizontal and vertical)</td>
</tr>
<tr>
<td>Grid Design</td>
<td>--</td>
<td>Arakawa C-grid</td>
</tr>
<tr>
<td><strong>Terrain:</strong></td>
<td></td>
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<td>Terrain</td>
<td>itern</td>
<td>None</td>
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<tr>
<td><strong>Turbulence:</strong></td>
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<tr>
<td>Subgrid Turbulence</td>
<td>iturb</td>
<td>TKE</td>
</tr>
<tr>
<td>** Radiation:**</td>
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<td></td>
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<td><strong>Boundary Conditions:</strong></td>
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<td>3 km deep starting at 17 km</td>
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<tr>
<td>Vertical Boundary Conditions</td>
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<td>No slip</td>
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<td><strong>Microphysics:</strong></td>
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<td>Microphysics scheme</td>
<td>ptype</td>
<td>NSSL 2-moment scheme with graupel and hail (see section 3.2.2)</td>
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<td>Variable, see section 3.3</td>
</tr>
<tr>
<td>Graupel Shape Parameter</td>
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<tr>
<td>Cloud Condensation Nuclei</td>
<td>ccn</td>
<td>Variable, see section 3.3</td>
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<td></td>
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<td>Initialization type</td>
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<td>Bubble Height (center)</td>
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<td>2 km AGL</td>
</tr>
<tr>
<td>Bubble Depth</td>
<td>var5 (bvrad)*</td>
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</tr>
<tr>
<td>Bubble Radius</td>
<td>var9 (bhrad)*</td>
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</tr>
<tr>
<td>Maximum Perturbation</td>
<td>var10 (bptpert)*</td>
<td>3°C</td>
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**Table 2: Description of differences between model runs**

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<thead>
<tr>
<th>Model Name</th>
<th>CCN (cm(^3))</th>
<th>INP Factor</th>
<th>Additional Changes</th>
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<tbody>
<tr>
<td>Control</td>
<td>700</td>
<td>1</td>
<td>Hail shape parameter (alpha_h) = 0.5</td>
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<tr>
<td>FastWR</td>
<td>350</td>
<td>1</td>
<td></td>
</tr>
<tr>
<td>ModerateWR</td>
<td>1000</td>
<td>1</td>
<td>Run time extended to 250 minutes</td>
</tr>
<tr>
<td>SlowWR</td>
<td>1400</td>
<td>1</td>
<td>Run time extended to 360 minutes</td>
</tr>
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<td>BroadHail</td>
<td>700</td>
<td>1</td>
<td>Hail shape parameter (alpha_h) = 0.2</td>
</tr>
<tr>
<td>NarrowHail</td>
<td>700</td>
<td>1</td>
<td>Hail shape parameter (alpha_h) = 0.9</td>
</tr>
<tr>
<td>MoreIN</td>
<td>700</td>
<td>10(^{-1})</td>
<td></td>
</tr>
<tr>
<td>LessIN</td>
<td>700</td>
<td>10(^{-1})</td>
<td></td>
</tr>
<tr>
<td>LessIN_IFoff</td>
<td>700</td>
<td>10(^{-1})</td>
<td>Immersion freezing turned off</td>
</tr>
<tr>
<td>LessIN_IFHoff</td>
<td>700</td>
<td>10(^{-1})</td>
<td>Immersion freezing and Hallet-Mossop process turned off</td>
</tr>
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</table>
**Figure 3:** Hail size distribution for $\alpha$-values of 0 (black), 0.2 (red), 0.5 (blue), and 0.9 (pink) representing a negative-exponential, broad, central, and narrow hail distribution out to the largest diameters (left) and focused on the smallest diameters (right).
Figure 4: Simulated reflectivity (shaded) and cold pool (contoured) for the Control (a), FastWR (b), ModerateWR (c), SlowWR (d), MoreIN (e), BroadHail (f), NarrowHail (g), LessIN (h), LessIN_IFoff (i), and LessIN_IFHMoff (j) towards the end of the respective runs.
Figure 5: Time series of maximum updraft (red) and downdraft (blue) velocities below 5 km for all 10 simulations
Figure 6: Time series of cold pool area from the start to the end of the 10 simulations
Figure 7: The starting time of the cold pool for each model run
Figure 8: Integrated latent cooling within the 5 m/s downdrafts, 10 minutes prior to the start of each cold pool
Figure 9: Contribution of graupel sublimation (a), graupel melting (b), hail melting and sublimation (c), and rain evaporation (d) to the total latent cooling within the downdraft forming the cold pool.
Figure 10: Integrated latent cooling within the 1 m/s downdrafts that directly touch the surface cold pool from the start of the cold pool to 51 minutes later.
Figure 11: Comparison of the cold pool area 51 minutes after its formation
Figure 12: The average propagation speed for the 51-minute period beginning at the start time of each run’s cold pool
Figure 13: Maximum 2-minute average propagation speed occurring in the first 51 minutes after cold pool formation for all model runs

![Graph showing maximum 2-minute average propagation speed of cold pool in first 51 minutes]
Figure 14: Time series of average cold pool depth for each model run
Figure 15: Average cold pool depth in the first 51 minutes after cold pool formation for all model runs
**Figure 16**: Maximum cold pool depth achieved in each model run during the first 51 minutes after the formation of the cold pool
Figure 17: Maximum cold pool strength for each model run in the first 51 minutes after cold pool formation.
Figure 18: Relationship between average propagation speed and total integrated latent cooling for all model runs
Figure 19: Relationship between average propagation speed and the latent cooling of rain (a), graupel (b), and hail (c) for all model runs
Figure 20: Relationship between average depth and total integrated latent cooling for all model runs
Figure 21: Relationship between average depth and the latent cooling of rain (a), graupel (b), and hail (c) for all model runs.

(a) 50 Minute Average Depth vs. Integrated Latent Cooling of Rain

(b) 50 Minute Average Depth vs. Integrated Latent Cooling of Graupel

(c) 50 Minute Average Depth vs. Integrated Latent Cooling of Hail
Figure 22: Relationship between average strength and total integrated latent cooling for all model runs

![Graph showing the relationship between 50 minute average strength and total integrated latent cooling. The graph displays a positive correlation with an r-value of 0.55. Different colored markers represent different model runs: Control, LessIN, MoreIN, FastWR, ModerateWR, SlowWR, BroadHail, NarrowHail, LessIN_IFH Moff, and LessIN_IF off.](image-url)
Figure 23: Relationship between average strength and the latent cooling of rain (a), graupel (b), and hail (c) for all model runs.

50 Minute Average Strength vs. Integrated Latent Cooling of Rain

r-value: 0.62

50 Minute Average Strength vs. Integrated Latent Cooling of Graupel

r-value: 0.47

50 Minute Average Strength vs. Integrated Latent Cooling of Hail

r-value: 0.36
**Figure 24:** Relationship between maximum strength and total integrated latent cooling for all model runs

![Graph showing the relationship between maximum strength and total integrated latent cooling for different model runs. The graph includes a regression line with an r-value of 0.54.]
REFERENCES


Robert Jeffrey Trapp, Mesoscale-Convective Processes in the Atmosphere. Cambridge University Press,


