Cold Pool Responses to Changes in Soil Moisture

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Abstract This study examines the role of soil moisture in modulating convective cold pool properties in an idealized modeling framework that uses a cloud-resolving model coupled to an interactive land surface model. Five high-resolution simulations of tropical continental convection are conducted in which the initial soil moisture is varied. The hundreds of cold pools forming within each simulation are identified and composited across space and time using an objective cold pool identification algorithm. Several important findings emerge from this analysis. Lower soil moisture results in greater daytime heating of the surface, which produces a deeper, drier subcloud layer. As a result, latent cooling through the evaporation of precipitation is enhanced, and cold pools are stronger and deeper. Increased propagation speed, combined with wider rain shafts, results in wider cold pools. Finally, the rings of enhanced water vapor that surround each cold pool when soil is wet disappear when the soil moisture is reduced, due to the suppression of surface latent heat fluxes. Instead, short-lived “puddles” of enhanced water vapor permeate the cold pools. The results are nonlinear in that the properties of the cold pools in the two driest-soil simulations depart substantially from the cold pool properties in the three simulations initialized with wetter soil. The dividing line between the resulting wet-soil and dry-soil regimes is the permanent wilting point. Below the permanent wilting point, transpiration is subdued due to a sharp increase in water stress. These results emphasize the role of land surface-boundary layer-cloud interactions in modulating cold pool properties.

Plain Language Summary When rain falls from storm clouds, some of the rain evaporates. In order to evaporate, the rain absorbs energy from the air around it, cooling the surrounding air. As the air cools, it becomes denser and accelerates toward the ground, forming a region of wind that blows downward (a “downdraft”). Then, upon reaching the surface, this cool, dense air collects and spreads out to form a cold pool. Cold pools are important because, as the cold pool spreads out, it pushes the environmental air in its path out of the way, forcing it upward. When this surrounding air is pushed upward, it can create a new storm cloud. In this study, we use computer model simulations to examine the effects of changing the wetness of the soil (the “soil moisture”) on the cold pools. We first identify and track the cold pools forming in each simulation, and we then measure their sizes and strengths. The results indicate that the simulations with the driest soil have the largest and strongest (coldest and densest) cold pools. Previous cold pool modeling studies have documented “rings” of very humid air that surround cold pools forming over an ocean surface. We find that these rings occur in the three simulations with the highest soil moisture but not in the two driest-soil simulations. Instead, the cold pools in the two driest-soil simulations have “puddles” of humid air in their interiors and low humidity at their peripheries. Many previous modeling studies examining cold pools over land have not included any sort of land surface in their simulations. That is, there are no plants or soil layers, and the ground does not interact, that is, exchange moisture and heat, with the air above. Our results indicate that these sorts of interactions can make a substantial difference in cold pool properties, and we therefore recommend that simulations take these factors into account when simulating and forecasting storm clouds.

1. Introduction

Convective cold pools help to modulate convective processes by suppressing convection in some locations while promoting it in other regions. Subsidence and enhanced boundary layer static stability suppress the formation of new convection in the interiors of existing cold pools (Tompkins, 2001). Meanwhile, cold pools trigger convection mechanically via the propagation of gust fronts, which lift surrounding boundary layer air to its level of free convection (e.g., Moncrieff & Liu, 1999; Torri et al., 2015). Through some combination of rain evaporation, enhanced latent heat fluxes, and advection of preexisting moisture anomalies, cold pools may also cause the accumulation of water vapor into “rings” near cold pool boundaries (Langhans &
Romps, 2015; Schlemmer & Hohenegger, 2016; Tompkins, 2001; Torri & Kuang, 2016). These water vapor rings have the potential to make cold pool peripheries thermodynamically favorable for the formation of new convection.

Several recent studies have performed detailed analyses of processes governing the emergence of water vapor rings in models. Langhans and Romps (2015) conclude that water vapor rings are generated primarily by latent heat fluxes. Schlemmer and Hohenegger (2016) confirm that latent heat fluxes are more important than evaporation of precipitation for providing water vapor to the water vapor rings (their “moist patches”) but also note that a substantial fraction (one third to one half) of the water vapor comes from moisture present in the subcloud layer prior to cold pool onset. In agreement with Schlemmer and Hohenegger (2016), Torri and Kuang (2016) suggest that latent heat fluxes are more important than evaporation of precipitation but that the latter cannot be neglected, particularly near the surface. Torri and Kuang (2016) and Langhans and Romps (2015) consider only cold pools over ocean, whereas Schlemmer and Hohenegger (2016) consider cold pools both over land and over ocean.

The precise location of water vapor rings relative to cold pools’ boundaries, that is, whether they are inside, outside, or colocated with cold pools’ gust fronts, has been the subject of some debate. In previous modeling studies, these rings have generally formed inside of the gust fronts bounding cold pools. However, observations of cold pools over tropical oceans suggest that these rings, if they exist at all, are colocated with or outside of the gust fronts (Chandra et al., 2018; de Szoeke et al., 2017; Zuidema et al., 2017).

In addition to modifying the location of convection, cold pools influence the nature of convection. Cold pools result in the formation of wider and deeper clouds, especially when cold pools collide with one another (Boing et al., 2012; Feng et al., 2015; Purdom, 1982; Schlemmer & Hohenegger, 2014; Wilson & Schreiber, 1986). In fact, the diurnal transition from shallow to deep convection can be suppressed in numerical models by eliminating the processes that generate cold pools (Khairoutdinov & Randall, 2006; Kurowski et al., 2018).

Despite their importance for convective initiation and organization, cold pools have only recently begun to be incorporated into climate models’ convective parameterization schemes (Del Genio et al., 2015; Park, 2014; Rio et al., 2013; Suselj et al., 2019; Zhao et al., 2018). Several aspects of cold pool science, such as microphysical controls on cold pool development (e.g., Dawson et al., 2010; Falk et al., 2019; Grant & van den Heever, 2015; Li et al., 2015; Mallinson & Lasher-Trapp, 2019; Morrison, 2012; van den Heever & Cotton, 2004) and interactions between cold pools and the land or ocean surfaces below (Drager & van den Heever, 2017; Fast et al., 2019; Gentine et al., 2016; Grant & van den Heever, 2016, 2018; Huang et al., 2018; Kurowski et al., 2018; Pei et al., 2018), remain active areas of research. It has also been recognized that boundary layer properties influence cold pool properties. Deeper boundary layers result in stronger cold pools, as do drier boundary layers (McCaul & Cohen, 2002). However, interactions between the surface and boundary layer as they pertain to cold pool development have not been examined. This is the area to which the present work aims to contribute.

We investigate the hypothesis that changes in soil moisture result in interactions between the surface, boundary layer, and clouds that feed back onto various cold pool properties such as size, strength, structure, and longevity. Soil moisture helps to govern the Bowen ratio, the ratio of sensible to latent heat fluxes at the surface (Bowen, 1926), by determining the amount of soil water available for evaporation directly off of the soil surface, as well as the vigor of transpiration by vegetation (e.g., Lee, 1992). If the evapotranspiration is limited and if all else is equal, then in order to achieve surface energy balance, the sensible heat flux must increase. Therefore, if all else is equal, then decreased soil moisture results in enhanced surface sensible heat fluxes and suppressed surface latent heat fluxes and thus an increased Bowen ratio (e.g., Garratt, 1992). Surface fluxes influence cold pool longevity both directly, through the injection and removal of sensible heat and moisture into the cold pool, and indirectly, by fueling boundary layer circulations and turbulence in the near-cold pool environment that interact with the cold pool via entrainment (Grant & van den Heever, 2016, 2018). The cold pool longevity and structure, in turn, impact the likelihood and location of cold pool triggering of new convection.

The present work was conducted in synergy with the United States Office of Naval Research’s Propagation of Intra-Seasonal Tropical Oscillations (PISTON) field campaign that took place during 2018 and 2019. The PISTON field campaign aims to improve our understanding of several factors, including land-atmosphere
interactions, that affect the propagation of the Boreal Summer Intraseasonal Oscillation (BSISO) through the Maritime Continent, with a particular emphasis on the Philippine Archipelago (Office of Naval Research, 2016). Cold pools are of particular interest given the potential importance of scale interactions between individual convective cells and larger features of the atmospheric circulation (Toms et al., 2020). Furthermore, it is hypothesized that cold pools in this region, in concert with other processes such as mountain circulations and land/sea breezes, impact the diurnal cycle of convection (e.g., Riley Dellaripa et al., 2020). The diurnal cycle in this region is not represented well in most climate models (e.g., Baranowski et al., 2018; Dirmeyer et al., 2012) or regional models (e.g., Riley Dellaripa et al., 2020). Cold pool parameterizations in convection-parameterizing models can potentially improve the simulated diurnal cycle in continental environments (RIO et al., 2009). With an eventual goal of improving these parameterizations, and thereby improving the representation of the diurnal cycle, the present study seeks to clarify whether, and to what extent, soil moisture affects cold pool properties.

This work investigates the effects of soil moisture on cold pool development in a set of numerical model simulations. Five idealized simulations of tropical continental convection are performed, each with a different initial soil moisture content. A cold pool identification algorithm is applied in order to generate a composite of the cold pools in each simulation. The differences between the composites are then explored, and physical mechanisms are assessed. Finally, the results are discussed in the context of convective initiation.

2. Model Setup, Sensitivity Experiments, and Analysis Methods

2.1. Model Setup and Sensitivity Experiments

The simulations used in this study are performed using the open-source Regional Atmospheric Modeling System (RAMS) (vandenheever.atmos.colostate.edu/vdhpage/rams.php) (Cotton et al., 2003), release 6.2.08. RAMS is a regional, nonhydrostatic atmospheric model that is coupled to the Land-Ecosystem-Air Chemistry Feedback version 3 (LEAF-3) soil-vegetation-atmosphere transfer model (Lee, 1992; Walko et al., 2000) and contains a sophisticated double-moment, bin-emulating bulk microphysics scheme with eight hydrometeor classes. The full model settings are provided in Table 1; selected aspects specific to these simulations are discussed below.

The idealized simulations are performed on a nonrotating (f = 0 s⁻¹) domain with doubly periodic lateral boundaries and no topography. The domain size is 150 km × 150 km × ~21 km, the horizontal grid spacing is 125 m, and the vertical grid spacing is stretched from 40 to 250 m (127 vertical levels). At this resolution, it is expected that the largest atmospheric turbulent eddies are resolved (i.e., large-eddy simulation); the effects of smaller eddies are parameterized using a modified form of the Smagorinsky (1963) scheme (Table 1).

Based on sensitivity tests at various resolutions, Grant and van den Heever (2016) recommend that horizontal (vertical) grid spacing of 100 m (50 m) be used in order to represent turbulent cold pool dissipation processes accurately. The simulations performed here approach this resolution. The coupled LEAF-3 model contains 11 soil levels that extend to a depth of 0.5 m following Grant and van den Heever (2014). The simulations are initialized at 07:00 LT and run for 14 hr, so as to capture a single day’s diurnal cycle of convection, with output files saved every 5 min.

The initial conditions are horizontally homogeneous except for pseudorandom thermal perturbations in the lowest ~500 m of the atmosphere. The initial atmospheric conditions are based on the conditions on the Philippine island of Luzon, during boreal summer. The initial winds are calm, and the initial thermodynamic and moisture profiles are obtained from the Laoag site in Luzon (weather.uwyo.edu/upperair/sounding.html). The profiles used in this study are adapted from the 0 UTC (~8 LT) sounding on 2 August 2010. This particular morning sounding is chosen because it corresponds to the active phase of the BSISO over Luzon, it is nearly free of apparent cloud layers, it is not influenced by tropical cyclone activity, the winds are weak, and convection is observed to develop later in the day. The red curves in Figure 1 show the raw sounding data. Before the sounding data are used to initialize the model, the “spike” in water vapor content at 152 hPa is removed, and then the thermodynamic and water vapor profiles are smoothed using a running mean filter with a span of 2 km. The resulting smoothed sounding is shown by the black curves in Figure 1. Although we are using a 0 UTC (8 LT) sounding, the model is initialized with this sounding at 23 UTC (7 LT) in order to allow for more model spin-up time.
The soil texture used in the LEAF-3 model is silty clay loam, the vegetation class is wooded grassland, and the prescribed normalized difference vegetation index (NDVI) is 0.6. These settings are selected to be roughly representative of Luzon during early August based on the RAMS global data sets of soil texture, vegetation class, and monthly NDVI.

Five sensitivity simulations, the DRENCHED-SOIL, WET-SOIL, MID-SOIL, PWP-SOIL (named for the permanent wilting point, which is defined and further discussed in section 2.2), and DRY-SOIL simulations, are performed. The initial soil moisture values in these simulations are 95%, 75%, 50%, 45%, and 25% of the saturation volumetric soil moisture (0.477 m$^3$ m$^{-3}$), respectively, and the initial soil temperatures (identical in all simulations) are slightly warmer than those of the air above, following Grant and van den Heever (2014). The 95% value is roughly typical of the mean conditions over Luzon in ERA5 reanalyses (Copernicus Climate Change Service [C3S], 2017) during the boreal summers of 2010 and 2016, representing weak and strong BSISO seasons, respectively, and the 45% and 50% values are approximately representative of the dry season (based on January through May of the same years). The 75% value is representative of the driest conditions over Luzon during boreal summer in reanalysis, and the 25% value is indicative of drought.

| Table 1 |
| --- |
| Details of the RAMS Model Setup |
| **Model aspect** | **Settings** |
| Model version | RAMS version 6.2.08 with LEAF-3 land surface model |
| Grid | Δx = Δy = 125 m; 150 km × 150 km domain size |
| | Δz stretched from 40 m at the surface to 250 m aloft; stretch ratio = 1.025 |
| | 127 vertical levels; model top at z ~ 21 km |
| Time integration | Δt = 1.5 s |
| Initial conditions | 14 hr simulation duration, from 23 UTC (7 LT) to 13 UTC (~21 LT) |
| | Horizontally homogeneous profiles of temperature and moisture (see smoothed model sounding in Figure 1), except for pseudorandom potential temperature perturbations that decrease in amplitude linearly with height from 0.1 K at z ~ 20 m to 0 K at z ~ 520 m. No initial winds. |
| Land surface | 11 soil levels |
| | Silty clay loam soil type |
| | Wooded grassland vegetation type |
| | Normalized difference vegetation index (NDVI) = 0.6 |
| Initial soil moisture | Spatially uniform value that varies according to the simulation: |
| | DRENCHED-SOIL: 95% of saturation volumetric soil moisture (100% = 0.477 m$^3$ m$^{-3}$) |
| | WET-SOIL: 75% |
| | MID-SOIL: 50% |
| | PWP-SOIL: 45% |
| | DRY-SOIL: 25% |
| Boundary conditions | Periodic lateral boundaries |
| | Rigid lid with Rayleigh friction absorbing layer over the top 2 km |
| Microphysics | RAMS two-moment bin-emulating bulk microphysics with eight hydrometeor classes, coupled to aerosol module (Meyers et al., 1997; Saleeby & Cotton, 2004, 2008; Saleeby & van den Heever, 2013) |
| | Gamma distribution shape parameter equals 4 for cloud and drizzle and 2 for all other hydrometeor classes |
| Aerosols | Sulfate aerosols (Saleeby & van den Heever, 2013), no sources or sinks, radiatively inactive |
| | DeMott et al. (2010) ice nucleation formulation |
| | Initial surface concentrations of cloud condensation nuclei (CCN), giant CCN, and ice nuclei (IN) are 400, 1 × 10$^{-6}$, and 0.01 mg$^{-1}$, respectively, decreasing exponentially with altitude with a scale height of 7,000 m. |
| Radiation | Harrington (1997) two-stream radiation scheme, updated every 5 min |
| Turbulence | Insolation spatially uniform based on 1–2 August 2010 at Laoag, Philippines. |
| | Smagorinsky (1963) scheme with vertical diffusion modifications based on the formulation of Hill (1974). Vertical diffusion of ice liquid potential temperature and water vapor mixing ratio is calculated based on perturbations relative to the initial sounding. |
| Topography | none |
| Coriolis | none |
2.2. Overview of the Simulations

Figures 2a–2f show domain mean vertical profiles of various quantities, averaged over a time period extending from 11:00 LT to 18:00 LT, for all five simulations. As is discussed later, this time window corresponds to our cold pool analysis period. In general, there are substantial differences between the two driest-soil simulations and the three wetter-soil simulations, which are quite similar within each subgroup for all of the fields compared here. Figures 2a, 2b, and 2e indicate that the boundary layer becomes warmer, drier, and deeper with decreasing soil moisture. The thermodynamic variable plotted in Figure 2a is the density potential-temperature \( \vartheta^0 \), which is defined following Emanuel (1994) as

\[
\vartheta^0 = \frac{R_v}{R_d} \left( 1 + \frac{r_v + r_{\text{cond}}}{r_v} \right) \approx \vartheta (1 + 0.608 r_v - r_{\text{cond}}),
\]

where \( \vartheta \) is the potential temperature, \( R_v \) is the gas constant of water vapor, \( R_d \) is the gas constant of dry air, \( r_v \) is the water vapor mixing ratio, and \( r_{\text{cond}} \) is the total condensate mixing ratio. The density potential temperature is similar to the virtual potential temperature except that it contains an extra correction for density increases due to condensate loading. It has been used in previous studies (e.g., Drager & van den Heever, 2017; Feng et al., 2015; Tompkins, 2001) to define cold pool regions and boundaries. Although all five simulations are initialized with the same atmospheric profile, the partitioning between surface sensible and latent heat fluxes, which is governed in part by the amount of soil moisture, dictates the amount of near-surface heating and moistening. Decreased soil moisture leads to decreased evapotranspiration and increased sensible heating. Increased sensible heating, in turn, leads to increased near-surface static instability and thus more vigorous and deeper boundary layer mixing. The water vapor mixing ratio \( (r_v) \) within the boundary layer decreases with drier soil because surface latent heat fluxes are diminished and more dry air from above is entrained into the boundary layer by the boundary layer turbulence. Within the boundary layer, increasing temperature and decreasing \( r_v \) combine to yield decreased relative humidity and increased saturation deficit (defined here as the difference between the saturation \( r_v \) and the actual \( r_v \)) with decreasing soil moisture, as shown by Figures 2c and 2d, respectively.

The cloud base height increases with decreasing soil moisture (Figure 2e), in concert with the increasing boundary layer depth. A perhaps less intuitive result is that the maximum domain mean cloud water mixing ratio more than doubles in the PWP-SOIL and DRY-SOIL simulations compared to the other three simulations. When this average is taken only over cloudy points, most of this discrepancy disappears (not shown), which indicates that the increase in domain mean cloud water mixing ratio is dominated by an increase in the cloud fraction (likely a combination of increased number of clouds and increased cloud size) rather than the in-cloud mixing ratio.

We now discuss precipitation production and accumulation, which will help us later to understand the differences in the cold pool properties. The domain mean precipitation mixing ratio (Figure 2f) reaches a greater value aloft in the PWP-SOIL and DRY-SOIL simulations than in the other three simulations. A similar result was obtained by Hu et al. (2017), who compared regions whose soils had stronger sensible heat fluxes to regions whose soils had lower sensible heat fluxes. However, they attributed this trend to a “soil-type breeze” similar in character to a land-sea breeze circulation, which is necessarily quite different from the mechanism acting here due to the present study’s relative spatial homogeneity in any given simulation. However, this difference in domain mean precipitation mixing ratio is smaller at the surface than aloft, which indicates that much more precipitation is evaporating aloft in the PWP-SOIL and DRY-SOIL simulations. Curiously, the MID-SOIL simulation exhibits the least precipitation throughout the column. This intermediate-soil-wetness disadvantage with respect to precipitation will be explored in future work.
In order to understand the nonlinear trends with respect to soil moisture in Figures 2a–2f, it is important to recognize that soil moisture is allowed to evolve in these simulations. Figure 2g shows the evolution of soil moisture saturation fraction at the bottom and top soil levels, as well as the field capacity and permanent wilting point (for reference), over the course of the entire simulation.

Also plotted in Figure 2g are the field capacity (the amount of soil moisture retained by soil following gravitational draining; 67.5% for the soil type used in the present study) and the permanent wilting point.
The PWP, which is defined as the soil moisture threshold below which plants’ roots are unable to absorb enough water to offset losses via transpiration, has been hypothesized to play an important role in the development of atmospheric circulations over land (Hohenegger & Stevens, 2018) because of its role in modulating latent heat fluxes and thus the Bowen ratio. Unlike the simplified soil model used by Hohenegger and Stevens (2018), the LEAF-3 model includes both evaporation and transpiration. Evaporation from the top soil layer into the atmosphere depends on the soil surface specific humidity, which is parameterized following the RAMS Technical Manual (vandenheever.atmos.colostate.edu/vdhpage/rams/docs/RAMS-TechnicalManual.pdf) to be a function of the field capacity, not the PWP. In this formulation, the ground water vapor mixing ratio is multiplied by a wetness factor that equals 1 when the top layer of soil is wetter than the field capacity, 0 when soil is completely dry, and increases gradually with increasing soil moisture between these two extremes. By contrast, transpiration from the root zone soil layers into the atmosphere is a function of the PWP via its dependence on soil water potential (Lee, 1992). Transpiration is controlled by multiple factors, including soil moisture, and the stomatal conductance’s dependence on soil moisture is nearly a step function whose threshold is the PWP. Readers are referred to the aforementioned RAMS Technical Manual, as well as Philip (1957) and Lee and Pielke (1992), for more information about the model’s treatment of surface evaporation, and readers are directed to Lee (1992) for additional details about transpiration in LEAF-3.

We may consider there to be three relevant soil moisture regimes within this modeling system. In the wettest regime, with soil moisture greater than the field capacity, there is sufficient soil moisture for both evaporation and transpiration. Within this regime, neither evaporation nor transpiration is a strong function of soil moisture. In the intermediate regime, with soil moisture between the field capacity and PWP, evaporation decreases with decreasing soil moisture, while transpiration remains approximately constant. Finally, in the driest regime, with soil moisture below the PWP, evaporation continues to decrease with decreasing soil moisture, and transpiration is essentially nonexistent. The DRENCHED-SOIL and WET-SOIL simulations represent the wettest regime, the MID-SOIL simulation falls within the intermediate regime, and the PWP-SOIL and DRY-SOIL simulations are part of the driest regime. The nonlinearities apparent across these simulations can be viewed through this lens: The DRENCHED-SOIL and WET-SOIL simulations are nearly identical, and the PWP-SOIL and DRY-SOIL simulations diverge sharply from the other three. The fact that the MID-SOIL simulation more closely resembles its moister counterparts than it does the PWP-SOIL simulation (despite 50% being much closer numerically to 45% than to 75%) provides evidence that transpiration is more important than evaporation in these highly vegetated simulations.

Figures 3a and 3b show representative snapshots of a subregion of the WET-SOIL and DRY-SOIL simulations, respectively. In both simulations, the convection is isolated, as might be expected given the calm initial winds, and reaches depths characteristic of cumulus congestus clouds. Many cold pools (apparent as blue blotches in the near-surface density potential temperature [θρ] field) form, and the cold pools are approximately stationary (i.e., not advected by the mean wind; not shown) and nearly circular in both cases. Note, however, that the color scales for θρ are quite different between the two simulations, in agreement with the variation shown in Figure 2a. The DRY-SOIL simulation exhibits higher mean values of θρ, along with greater variability in θρ values.
3. Analysis Approach

An updated version of the cold pool identification and tracking algorithm of Drager and van den Heever (2017) is used to analyze the ensemble of cold pools developing within each simulation. Although the spirit of the algorithm is unchanged, new features have been added, and the implementation has shifted from pixel-based cold pools to polygon-based cold pool boundaries for greater precision. The updates made to the Drager and van den Heever (2017) algorithm are discussed in supporting information Text S1.

Once the cold pool tracking is complete, the cold pools are aligned in space, according to the centers of their respective cylindrical polar coordinate systems, and in time, according to their individual respective reference times (t = 0 min, defined for each cold pool to be the time the cold pool is first identified; see Text S1 for more information). Composites are then generated. To create the composites, each field ψ(x, y, z, t), where t is defined relative to t = 0 min and ψ could represent any variable such as ρ or vertical velocity w, is linearly interpolated to two types of cylindrical polar grids. In the first type of cylindrical polar grid, the radial coordinate is normalized by the cold pool radius at each azimuth (see Figure 4), in a manner similar to that of Langhans and Romps (2015). In the second type, the radial coordinate is not normalized. Then, azimutal averaging is performed to yield ψ(r\_norm, z, t) (normalized radial coordinate) and ψ(r, z, t) (nonnormalized) for each variable for every cold pool, with radial spacing of Δr\_norm = \frac{1}{15} and Δr = 0.25 km. Finally, composite fields are computed by taking the arithmetic means of ψ(r\_norm, z, t) and ψ(r, z, t) across the full set of cold pools in each simulation.

An important distinction between the normalized and nonnormalized composites is that, although every cold pool is included in the t = 0 min normalized composites, fewer cold pools are included before and after t = 0 min based on whether, and to what extent, each cold pool can be tracked backward or forward in time. By contrast, nonnormalized composites can be constructed in the absence of cold pool boundary contours, and therefore all cold pools are included in the composites at each time relative to t = 0 min. The nonnormalized composites are computed as far backward as t = −60 min, in order to capture the environments in which cold pools form, and as far forward as t = 120 min, so that cold pool dissipation may be assessed. The disadvantage of using nonnormalized composites is that “smearing” of the cold pool features occurs due to differing sizes within each ensemble of cold pools, or even at different azimuths within a given cold pool.

4. Cold Pool Characteristics

4.1. Analysis Period and Numbers of Cold Pools

The remaining analyses presented in this paper consider only those cold pools whose t = 0 min reference times fall between 11 LT and 16 LT, inclusive. Few cold pools form before 11 LT. Cold pools forming after 16 LT—whose contributions to the composites would extend until after 18 LT—are excluded from the analyses in order to prevent processes that occur only during the evening and nighttime hours, such as dew formation, from influencing the composites. Therefore, only daytime cold pool dissipation processes are considered in the present study. The total numbers of unique cold pools in each simulation that fall within our analysis period are 1,869 (DRENCHED-SOIL), 1,996 (WET-SOIL), 1,930 (MID-SOIL), 3,311 (PWP-SOIL), and 3,282 (DRY-SOIL).
4.2. Cold Pool Area Statistics

Figure 5 displays information about cold pool area. It is immediately apparent from Figure 5a that the cold pools in the PWP-SOIL and DRY-SOIL simulations have about twice the area of those in the other three simulations, in a mean sense. This trend holds throughout the plotted time interval. Figures 5c and 5d corroborate the trend of larger cold pools in the PWP-SOIL and DRY-SOIL simulations: The PWP-SOIL and DRY-SOIL simulations exhibit a smaller fraction of small cold pools and a larger fraction of large cold pools.

The cold pool identification and tracking algorithm operates by looking for well-defined density potential temperature boundaries. If no such boundary exists, or if the boundary is blurred or irregular at a given time, then the algorithm will not identify a cold pool. Therefore, the length of time over which a cold pool is tracked in the final tracking stage can be roughly interpreted as a cold pool lifetime. Under this interpretation, it is apparent from Figure 5b that cold pools in the PWP-SOIL and DRY-SOIL simulations are shorter lived than those in the other three simulations, in agreement with the results shown in Figure S2. The topic of cold pool longevity is revisited in section 5.3.

4.3. Cold Pool Strength

In the discussion that follows, perturbation quantities are compared in order to allow cold pool properties to be assessed independently of the differences between the simulations’ horizontal mean states. All perturbation quantities are calculated as departures from a simulation-dependent, height-dependent, time-varying horizontal mean.

Figure 6 shows probability density functions (PDFs) and mean values of various metrics of cold pool strength (Drager & van den Heever, 2017). For reference, several of these quantities are illustrated in Figure 4b. The equivalent potential temperature perturbation, $\tilde{\theta}_e$, is included because it has previously been used to identify
cold pools, although we will argue that this metric can be misleading (see section 4.5). These PDFs are generated using the nonnormalized, azimuthally averaged values at the lowest aboveground model level and within 4 km of the cold pool center across times $-60 \min \leq t \leq 120 \min$, except for the maximum updraft: The maximum updraft is obtained for radii between 0.75 and 4 km and only for $t \geq 0 \min$, in order to ensure that the value corresponds to the maximum uplift along the gust front rather than the earlier updraft that generates cold pool’s parent cloud. Using minimum or maximum azimuthally averaged values rather than pointwise extrema prevents contamination from nearby cold pools.

Examination of Figures 6a–6e reveals that cold pools in the PWP‐SOIL and DRY‐SOIL simulations are, on the whole, substantially stronger than those in the DRENCHED‐SOIL, WET‐SOIL, and MID‐SOIL simulations. The mean values of minimum $\theta'_\rho$, maximum $v_r$, $w_{\text{min}}$, and $w_{\text{max}}$ are approximately twice as large (in an absolute value sense) in the PWP‐SOIL and DRY‐SOIL simulations as in the other three simulations, and the mean $p'$ value is about 3 times as large in the PWP‐SOIL and DRY‐SOIL simulations as in the other three. The larger relative change in $p'$ than in $\theta'_\rho$ suggests that the cold pools in the PWP‐SOIL and DRY‐SOIL simulations are deeper than those in the other three simulations (see section 4.6).

In order to elucidate cold pool structure, Figure 7 shows normalized by‐radius composites of the same variables as in Figures 6a–6e, plotted at $t = 0 \min$. These composites confirm the direction of the trend toward increased cold pool strength with decreasing soil moisture (particularly in the case of the PWP‐SOIL and DRY‐SOIL simulations). In Figure 7a, the values of $\theta'_\rho$ are minimized near the cold pool center and increase outward. The largest negative $\theta'_\rho$ perturbation occurs in the DRY‐SOIL simulation, and the trend holds throughout the interior of the cold pool ($0 \leq r_{\text{norm}} \leq 1$). In the PWP‐SOIL and DRY‐SOIL simulations, there is a peak in $\theta'_\rho$ at $r_{\text{norm}} \approx 1.5$, indicating a ring of enhanced warmth outside the cold pool boundary; this ring is revisited in section 5.2. In Figure 7b, $v_r$ peaks inside of the cold pool boundary ($r_{\text{norm}} < 1$) in all five simulations, and the magnitude of this peak is greatest in the DRY‐SOIL simulation. Far outside the cold pool, $v_r < 0 \text{ m s}^{-1}$ due to the residual circulation that generated the parent cloud. The vertical velocities (Figure 7c) are negative within the cold pool and positive outside the cold pool, indicative of a parent downdraft and uplift ahead of the gust front, respectively. Downdraft and updraft strengths are both greatest in the DRY‐SOIL simulation. Finally, values of $p'$ (Figure 7d) are maximized near cold pool center and decay...
outward, and they are much greater in the PWP-SOIL and DRY-SOIL simulations than in the other three. The extent to which the PWP-SOIL and DRY-SOIL simulations’ cold pools are stronger than those of the moister three simulations is exaggerated in Figure 7 compared to Figures 6a–6e. This is because the cold pools in the moister three simulations do not reach their peak intensity until \( t = 5 \) min (not shown). Figures 6a–6e account for this by considering a range of times relative to \( t = 0 \) min, while Figure 7 only shows the \( t = 0 \) min snapshots.

Although the cold pools are stronger (the difference from the mean background state is greater) in the PWP-SOIL and DRY-SOIL simulations than those in the DRENCHED-SOIL, WET-SOIL, and MID-SOIL simulations, the former cold pools are still generally warmer, in an absolute sense, than the latter (Figure 8a). As will be discussed further in section 4.4, the PWP-SOIL and DRY-SOIL cold pools exhibit moist perturbations near their centers, whereas the DRENCHED-SOIL, WET-SOIL, and MID-SOIL cold pools exhibit dry perturbations in their centers. However, despite their dry perturbations, the interiors of the DRENCHED-SOIL, WET-SOIL, and MID-SOIL cold pools are moister than the interiors of the PWP-SOIL and DRY-SOIL cold pools when total (as opposed to perturbation) water vapor mixing ratio is considered (Figure 8b).

### 4.4. Water Vapor Structure

Figure 9a shows the composite near-surface water vapor structure, as a function of \( r_{\text{norm}} \), for each of the four simulations at \( t = 0 \) min. The DRENCHED-SOIL, WET-SOIL, and MID-SOIL simulations exhibit the water vapor rings, sometimes referred to as moist patches, that were discussed in section 1. These rings manifest as negative values of perturbation water vapor mixing ratio, \( r'_v \), within the cold pool interiors, followed by positive values of \( r'_v \) outside of the cold pool (i.e., at \( r_{\text{norm}} > 1 \)). The PWP-SOIL and DRY-SOIL simulations’
composites do not contain a water vapor ring. Instead, each of these composites exhibits a small (~0.2 g kg$^{-1}$) positive water vapor mixing ratio perturbation—a water vapor “puddle”—that is approximately uniform within the cold pool and decays with increasing distance outside of the cold pool.

As was discussed in section 1, there has been some disagreement as to whether water vapor rings, such as those exhibited by the DRENCHED-SOIL, WET-SOIL, and MID-SOIL simulations, reside inside, along the edges, or outside of cold pools. As Chandra et al. (2018) point out, the location of water vapor rings not only provides clues regarding how the water vapor rings are formed but also has implications for the triggering of subsequent convection. For example, if water vapor rings are located outside of cold pools, then the thermodynamic and mechanical mechanisms for cold pool triggering of convection may act synergistically. This is potentially the case in the DRENCHED-SOIL, WET-SOIL, and MID-SOIL simulations analyzed here: The water vapor rings (Figure 9a) and cold pool-induced updrafts (Figure 7c) are both located outside of the cold pools. By contrast, if the water vapor rings are located inside of the cold pools, then the local moisture enhancements provided by water vapor rings will not as easily assist in generating new convection because they reside within negatively buoyant air.

Attempts to determine the location of water vapor rings relative to cold pools have been complicated by the use of disparate methods for defining cold pools. Several recent studies (e.g., Dawson et al., 2010; Schiro & Neelin, 2018; Schlemmer & Hohenegger, 2014, 2016) define cold pools in terms of equivalent potential temperature, $\theta_e$, whereas others use temperature- or buoyancy-based metrics (Drager, 2016). Since $\theta_e$ has such a strong dependence on moisture, a water vapor ring that is located along or inside the edge of a cold pool will cause the region of low $\theta_e$ associated with the cold pool to be smaller than the corresponding region of low temperature or negative buoyancy (Drager & van den Heever, 2017). It follows that if the water vapor ring’s positive influence on $\theta_e$ due to enhanced moisture is greater in magnitude than the cold pool’s negative influence on $\theta_e$ due to low temperatures, then the water vapor ring will always occur outside of the $\theta_e$-based cold pool, as is the case in Schlemmer and Hohenegger (2016). This is why both Zuidema et al. (2017) and Chandra et al. (2018) characterize the Schlemmer and Hohenegger (2016) study as exhibiting water vapor rings outside the cold pools when in fact the rings appear to be located inside of the gust front (see, e.g., Figure 6 of Schlemmer & Hohenegger, 2016, in which the updrafts—which via continuity are indicative of the gust front—are located along or just outside the edge of the water vapor ring). More generally, it is difficult to compare water vapor ring results across various studies due to the different metrics used to define cold pool boundaries, and we recommend that comparisons across studies take into account the disparate cold pool definitions.

4.5. Equivalent Potential Temperature: A Measure of Cold Pool Strength?

Schlemmer and Hohenegger (2014) propose a cold pool parameterization framework in which the buoyancy term in the equation for cold pool propagation speed (sometimes referred to as cold pool intensity) is replaced by a $\theta_e$ deficit term. That is, $-g \frac{\theta_e}{\theta_e}$, where $g$ denotes acceleration due to gravity, the overbar represents the base state, and the prime represents a deviation from the base state, is replaced by, essentially, $-g \frac{\theta'_e}{\theta_e}$ (see Equation 6 of Schlemmer & Hohenegger, 2014). Pucillo et al. (2020) take a similar approach.

The equivalent potential temperature can be an attractive variable to use because it is conserved under moist vapor-liquid pseudoadiabatic processes and therefore contains information about the height of the source region(s) of cold pool air. There are, of course, uncertainties due to entrainment of environmental air into
the downdraft, ice processes, other diabatic processes, and the potential for \( \theta_e \) to vary nonmonotonically with height. Nevertheless, \( \theta_e \) has often been used to obtain estimates of where downdraft air originates. (Schiro and Neelin, 2018, and Zuidema et al., 2017, are recent examples of studies that do this.) But is it an appropriate variable to use to define cold pool strength, in a conceptual and/or parameterization framework?

Our results suggest that it may not be. The mean values of minimum \( \theta_e \) (Figure 6f) exhibit a trend opposite to those of the other five metrics in Figure 6: The PWP-SOIL and DRY-SOIL simulations’ cold pools exhibit

Figure 9. Surface and near-surface normalized composites at \( t = 0 \) min (a–d) and nonnormalized composites at \( t = 15 \) min (e, f). The plotted variables are (a) \( r_v \), (b) \( \theta_e \), (c) surface latent heat flux, (d) surface sensible heat flux, (e) vegetation water (rain intercepted by and collecting on leaves), and (f) stomatal resistance (note logarithmic vertical axis). Panels (a) and (b) are plotted for \( z \sim 20 \) m and are computed relative to a time-varying horizontal domain mean, while all other variables are surface quantities.
approximately half the strength (in a $\varrho_\theta$ sense) of those in the DRENCHED-SOIL and WET-SOIL simulations. The MID-SOIL simulation lies between these two extremes. Recall that the PWP-SOIL and DRY-SOIL simulations’ cold pools are roughly twice as strong as those in the other three simulations according to the other five, more dynamically based metrics. In short, dynamical strength does not necessarily translate into $\varrho_\theta$ strength, and $\varrho_\theta$ trends can be misleading.

The equivalent potential temperature also appears to be a problematic metric of cold pool strength from a parameterization perspective. In the DRENCHED-SOIL case, $-g\frac{\varrho_\theta}{\varrho}$ is approximately 3 times as large as $-g\frac{\varrho_\theta}{\varrho}$ for the average cold pool. By contrast, in the DRY-SOIL case, $-g\frac{\varrho_\theta}{\varrho}$ is about 20% smaller than $-g\frac{\varrho_\theta}{\varrho}$. Therefore, not only do the trends reverse, but the magnitude of the error also differs across the soil moisture regimes examined here.

### 4.6. Cold Pool Depth

Figures 10a and 10b show composites of $\varrho_\theta$ and the transverse circulation at $t = 30$ min for the WET-SOIL and DRY-SOIL simulations, respectively. For brevity, the following discussion and corresponding plots consider only the WET-SOIL and DRY-SOIL simulations. It is clear that the cold pools in the DRY-SOIL simulation are, in a composite sense, deeper than those in the WET-SOIL simulation. The ground-based region of negative $\varrho_\theta$ extends to a greater height in the DRY-SOIL simulation (~1.75 km) than in the WET-SOIL simulation (~1 km). In addition, the surface-based region of radially outward directed winds, that is, the outflow, is deeper in the DRY-SOIL simulation (~1 km) than in the WET-SOIL simulation (~0.5 km). It is not immediately clear why the DRY-SOIL cold pools are deeper than the WET-SOIL cold pools, and indeed, the precise mechanisms governing cold pool depth in these soil moisture sensitivity tests are not fully understood. However, cold pool depth does not typically exceed the depth of the subcloud layer, and since the subcloud layer is deeper in the DRY-SOIL simulation, cold pools are permitted to become deeper as well.

### 5. Cold Pool Processes

#### 5.1. Mechanisms Governing Cold Pool Strength and Area

As discussed in section 4.3, the cold pools in the PWP-SOIL and DRY-SOIL simulations are stronger than those in the other three simulations. In order to explain this trend, we now explore the latent cooling that gives rise to the cold pools in the first place. Figures 10c and 10d show composites of model-derived latent heating and cooling from condensation and evaporation processes for the WET-SOIL and DRY-SOIL simulations, respectively. The maximum latent cooling in the subcloud layer occurs at different times (relative to $t = 0$ min) in the WET-SOIL and DRY-SOIL simulations. Therefore, instead of plotting the composites at a particular $t$ value, we construct these plots by taking the minimum composite value across all $t$ values from $t = -30$ min to $t = 30$ min at each point in $r$-$z$ space.
Figure 11, which shows composites of cloud mixing ratio $r_{\text{cloud}}$ (including cloud water and cloud ice hydrometeor species) in the top row and rain mixing ratio $r_{\text{rain}}$ in the bottom row, is generated similarly.

In each of the two bottom panels of Figure 10, there are two semiconnected regions of latent cooling, one above cloud base, which corresponds to evaporation of cloud droplets as the cold pools’ parent clouds dissipate, and one below cloud base, which corresponds to evaporation of rain drops. The lower regions are of particular interest because evaporation within the rain shaft helps to drive cold pool formation. The lower region of latent cooling is both deeper and greater in magnitude in the DRY-SOIL simulation (Figure 10d) than in the WET-SOIL simulation (Figure 10c). Therefore, if differences in downdraft vertical velocity and rain shaft lifetime can be neglected, then parcels of air descending below cloud base have a greater residence time in the subcloud layer and undergo greater rates of latent cooling during descent in the DRY-SOIL simulation than in the WET-SOIL simulation.

The rates of latent cooling are governed by both the dryness of the subcloud layer and the properties of the falling rain drops. The dryness of the subcloud layer can be quantified according to the saturation deficit (see Figure 2d), which is greater in the PWP-SOIL and DRY-SOIL simulations than in the other three simulations. Increased saturation deficit leads to more evaporation and thus enhanced latent cooling. It is clear from the vertical gradients in $r_{\text{precip}}$ in Figure 2f and $r_{\text{rain}}$ in Figures 11c and 11d that more evaporation is occurring in the DRY-SOIL simulation, in agreement with the latent cooling differences (Figures 10c and 10d).

As was discussed in section 4.2, the cold pools in the PWP-SOIL and DRY-SOIL simulations are larger than those in the other three simulations at the initial time of detection. This result appears to be due to the greater width of the rain shafts at cloud base in the DRY-SOIL simulation (compare Figure 11d at $z \sim 2.5$ km to Figure 11c at $z \sim 1.25$ km), which in turn is due to the wider clouds (compare Figure 11b to Figure 11a). In other words, wider clouds yield wider rain shafts, which in turn generate initially larger cold pools. Since these larger cold pools are also deeper, with more vertically integrated negative buoyancy through the subcloud layer, they expand more rapidly and thus remain wider. The mechanisms leading to wider clouds in the DRY-SOIL simulation are not obvious. We speculate that this result is due to the increased depth of the subcloud layer (i.e., increased cloud base height), which can result in wider updraft plumes at cloud base (e.g., Williams & Stanfill, 2002) and thus wider clouds. It was noted in section 2.2 that the cloud fraction is much greater in the PWP-SOIL and DRY-SOIL simulations than in the other three simulations. We conclude that this is the result of both greater numbers of clouds (see Figure S2a for a sense of this) and greater cloud width.

5.2. Mechanisms Governing Cold Pool Structure

As discussed in section 4.4, the cold pools in the DRENCHED-SOIL, WET-SOIL, and MID-SOIL simulations exhibit water vapor rings outside of the cold pools, whereas the cold pools in the PWP-SOIL and DRY-SOIL simulations exhibit water vapor “puddles” within the boundary of the cold pools. It should be noted that some ephemeral water vapor rings are apparent in horizontal cross sections of $r$, in the DRY-SOIL simulation, usually in association with the strongest cold pools (not shown). In contrast to the water vapor rings in
the three wettest-soil simulations, these DRY-SOIL water vapor rings are located inside of the gust front and appear to disappear quickly via mixing.

The relative lack of water vapor rings in the PWP-SOIL and DRY-SOIL simulations sheds additional light on the mechanisms leading to the formation of water vapor rings discussed in section 1. As was discussed in section 5.1, there is more evaporation of precipitation in the parent rain shafts of cold pools in the DRY-SOIL simulation than in the WET-SOIL simulation. Since water vapor rings only rarely and fleetingly emerge in the DRY-SOIL simulation, we conclude that rain shaft evaporation alone is not sufficient to generate water vapor rings, as other studies have also suggested (Langhans & Romps, 2015; Schlemmer & Hohenegger, 2016; Torri & Kuang, 2016).

The two other main potential sources of moisture for water vapor rings are surface latent heat fluxes and advection of preexisting moisture perturbations (see section 1). Figures 9c and 9d show normalized-by-radius composites of surface latent heat flux and sensible heat flux, respectively, at \( t = 0 \) min. It is immediately apparent that the Bowen ratio differs dramatically between the two driest-soil simulations and the other three simulations, with more sensible heat flux and less latent heat flux in the PWP-SOIL and DRY-SOIL simulations than in the other three simulations. With the exception of the latent heat flux in the PWP-SOIL and DRY-SOIL simulations, all surface fluxes are suppressed at cold pool center due to the enhanced static stability (not shown) and the suppressed winds (see Figure 7b), as was the case in Grant and van den Heever (2018). Cooler land surface and vegetation temperatures (not shown) generated by cloud shading and interception of cool precipitation also help to suppress the sensible heat fluxes near cold pool center (Drager & van den Heever, 2017). The sensible heat flux composites in the PWP-SOIL and DRY-SOIL simulations, along with the latent heat flux composites in the other three simulations, exhibit local maxima near \( r_{\text{norm}} = 1 \) (Figures 9c and 9d) in association with enhanced winds (the gust front; Figure 7b). These local maxima become much more pronounced shortly after \( t = 0 \) min as winds increase in strength (not shown).

As was mentioned in section 4.3, there is a ring of enhanced \( \theta_\rho \) that maximizes in strength near \( r_{\text{norm}} = 1.5 \) in the PWP-SOIL and DRY-SOIL simulations (Figure 7a), and in the other three simulations, there are rings of enhanced \( r_c \) near \( r_{\text{norm}} = 1.5 \) (Figure 9a). Prior studies are generally in agreement that latent heat fluxes are important, and possibly of primary importance, for the development of water vapor rings (Langhans & Romps, 2015; Schlemmer & Hohenegger, 2016; Torri & Kuang, 2016). If this is indeed the case, then it is possible that the local latent heat flux maxima at \( r_{\text{norm}} = 1 \) in the three wettest-soil simulations are associated with the corresponding water vapor rings near \( r_{\text{norm}} = 1.5 \). It is also possible that the local sensible heat flux maximum at \( r_{\text{norm}} = 1 \) in the PWP-SOIL and DRY-SOIL simulations is associated with the corresponding ring of enhanced \( \theta_\rho \) (which can be interpreted as a ring of enhanced temperature given the lack of \( r_c \) perturbation) near \( r_{\text{norm}} = 1.5 \).

Complicating matters are the possible roles of preexisting temperature and \( r_c \) perturbations. Figures 12a and 12b show nonnormalized-by-radius composites of \( T' \) (perturbation temperature) and \( r_c' \), respectively, at \( t = -30 \) min. These correspond to half an hour before each cold pool is detected and are therefore representative of the preexisting perturbations. There is a small but significant positive temperature perturbation near \( r = 0 \) km in the PWP-SOIL and DRY-SOIL simulations, and there is also a large positive \( r_c \) perturbation in the other three simulations. We expect that the influence of the preexisting perturbations may be more fleeting than that of the surface heat fluxes given that the air in which the preexisting perturbations reside should be lifted fairly early in the cold pool lifecycle via cold pool mechanics. Nevertheless, since we have not used tracers or Lagrangian particles in this analysis, we cannot make any definitive statements regarding whether surface heat fluxes or preexisting perturbations are more important.

We now discuss the moisture perturbations near cold pool center, which are positive (moist) in the two driest-soil simulations and negative (dry) in the other three simulations (Figure 9a). It is evident from Figure 2b that moisture decreases more precipitously with height as soil moisture increases. This implies that, if we consider a hydrometeor-free downdraft originating from cloud base, then such a downdraft would generate a larger negative surface moisture perturbation in the three wettest-soil simulations than in the two drier-soil simulations. However, in reality, downdraft air contains a source of water vapor in the form of evaporating hydrometeors, and there is more evaporation in the
PWP-SOIL and DRY-SOIL simulations (see section 5.1). This evaporation is evidently sufficient to generate a positive surface moisture perturbation in the PWP-SOIL and DRY-soil simulations but not in the other three simulations.

5.3. Cold Pool Dissipation

Figures 12c and 12d ($t = 5$ min) corroborate many of the results discussed in section 4 regarding cold pool size and water vapor structure, and they provide a spatial reference for comparison with the bottom row of panels (Figures 12e and 12f, $t = 90$ min). Figure 12e shows that the cold pools in the PWP-SOIL and DRY-SOIL simulations are shorter lived than those in the other three simulations. In the PWP-SOIL and DRY-SOIL simulations, $T' > 0$ K near $r = 0$ km, indicating that the cold pools in this simulation have dissipated by this time. By contrast, the composite $T'$ is negative in the other three simulations, indicating that the cold pools in these simulations have dissipated more slowly. The composites of $r'_v$ at $t = 90$ min (Figure 12f) indicate that water vapor puddles in the two driest-soil simulations dissipate along with their

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**Figure 12.** Time evolution of nonnormalized composites of temperature perturbation $T'$ (panels a, c, and e) and $r'_v$ (panels b, d, and f). The times plotted are $t = -30$ min (panels a and b), $t = 5$ min (panels c and d), and $t = 90$ min (panels e and f).
associated cold pools, whereas water vapor rings persist in the three wettest-soil simulations beyond the lifetimes of their corresponding cold pools.

Cold pool dissipation need not occur uniformly throughout a cold pool (Grant & van den Heever, 2018). Next, we describe two cold pool processes that have been discussed in prior literature that appear to be active in our simulations. First, the wet-patch effect primarily affects the evolution of the central parts of the cold pools. Second, the turbulent mixing effect erodes cold pools from their edges inward.

The wet-patch effect (Drager & van den Heever, 2017), in which fallen precipitation cools the ground and vegetation and suppresses (or even reverses the sign of) sensible heat fluxes, prolongs the cold pool dissipation process within the part of the cold pool that has experienced precipitation. Figure 9e shows that less rain water is intercepted by vegetation in the PWP-SOIL and DRY-SOIL simulations than in the other three simulations. This is because the time-integrated rainfall, in a per-cold pool composite sense, decreases with decreasing soil moisture (not shown). It is likely that the increased vegetation-intercepted rain water in the three wettest-soil simulations helps to increase the cold pool longevity in these simulations relative to that in the PWP-SOIL and DRY-SOIL simulations (Drager & van den Heever, 2017).

In a large-eddy simulation of an idealized cold pool in a dry continental environment, Grant and van den Heever (2018) documented cold pool dissipation from the outside inward due to turbulent entrainment of warm environmental air and enhanced (suppressed) sensible heat fluxes near the edge (center) of the cold pool. The PWP-SOIL and DRY-SOIL simulations' cold pools exhibit stronger flows, so they should exhibit greater turbulence at their boundaries. Furthermore, given the greater surface heating in the PWP-SOIL and DRY-SOIL simulations, we can expect that turbulent boundary layer motions are more vigorous overall in this simulation than in the other three simulations. This implies that the outer edges of the PWP-SOIL and DRY-SOIL simulations' cold pools are more likely to be eroded by turbulent entrainment. The identification and tracking algorithm is less successful at locating the PWP-SOIL and DRY-SOIL cold pools' edges than at locating those of the cold pools in the other three simulations at times after \( t = 0 \text{ min} \) (see Figure 5b). This is likely due to a combination of ill-defined boundaries—indeed, anecdotally they are “fuzzier” in the DRY-SOIL (see Figure S1) and PWP-SOIL simulations—and decreased cold pool longevity.

5.4. Role of the PWP

As was discussed in section 2.2, there appear to be three main soil moisture regimes. Each regime has a different partitioning of surface sensible and latent heat fluxes, as determined by the amount of evaporation from the top soil level (a function of the field capacity) and the amount of transpiration by plants (a function of the PWP). The partitioning of surface heat fluxes affects the depth, temperature, moisture, and vigor of circulations in the boundary layer. These affect convection and, by extension, the formation and evolution of cold pools.

Although there are some subtle differences between the MID-SOIL simulation’s cold pools and those in the moistest two simulations, which are likely controlled by changes in direct surface evaporation, there appear to only be two main regimes for cold pool properties: moister than the PWP and drier than the PWP. The fact that the PWP-SOIL simulation (initialized at 45% of saturation, just below the PWP of 45.7%) behaves much more like the DRY-SOIL simulation (25%) than the MID-SOIL (50%) simulation serves to emphasize that there are two well-defined regimes bounded by the PWP.

The existence of two well-defined regimes appears to depend on the presence of vegetation, as it is the water stress experienced by plants—specifically, the modulation of this water stress by soil moisture—that enables the PWP’s relevance. Physically, in soils drier than the PWP, roots are unable to take up water, and plants therefore close their stomata (Figure 9f). Reducing the vegetation coverage would increase the importance of direct surface evaporation relative to transpiration. Direct surface evaporation is not a function of the PWP and instead increases gradually as a function of soil moisture until the field capacity is exceeded (see section 2.2). We therefore speculate that the total evapotranspiration over a sparsely vegetated surface would transition more gradually between dry-soil and wet-soil regimes, and the cold pool properties would undergo a correspondingly more gradual transition.

For short-rooted plants, we expect that the importance of the initial soil moisture relative to the PWP will be reduced, as short-rooted plants will draw water only from the top soil levels. If the initial soil moisture is above the PWP, and if large amounts of transpiration reduce the root zone soil moisture to a level below...
the PWP, then a corresponding transition in boundary layer, cloud, and cold pool properties could occur over the course of a single day. Over bare soils, we do not expect the PWP to play any role in determining cold pool properties. Instead, evaporation should be determined entirely by the moisture of the top soil level. As discussed in section 2.2, the top soil level in both the MID-SOIL and PWP-SOIL simulations decreases to well below the PWP over the course of the day. Therefore, the transition between dry-soil and wet-soil regimes may depend on the efficacy with which water is replenished from below into the top soil level. More replenishment appears to occur in the WET-SOIL simulation than in the MID-SOIL simulation (section 2.2; Figure 2g). Therefore, there may be a replenishment threshold somewhere between ~50% initial soil moisture (MID-SOIL) and ~75% initial soil moisture (WET-SOIL), and this replenishment threshold could control the transition between dry-soil and wet-soil regimes over bare soils.

6. Discussion and Conclusion

This study has examined the role of soil moisture in governing tropical continental convective cold pool properties using idealized, high-resolution, cloud-resolving model simulations coupled to an interactive land surface model. Cold pool analyses were performed for five simulations: DRENCHED-SOIL (initialized at 95% of soil saturation), WET-SOIL (75%), MID-SOIL (50%), PWP-SOIL (45%), and DRY-SOIL (25%). Our hypothesis was that soil moisture has important effects on cold pool size, strength, structure, and longevity. The main findings in this regard, which are summarized in the schematic presented in Figure 13, are as follows:

1. Decreasing the soil moisture to levels below the permanent wilting point (PWP) yields cold pools that are approximately twice as strong and twice as large, both in area and in depth. Even though direct surface evaporation is present in these simulations, transpiration, which is governed in part by the PWP, appears to be the dominant factor controlling this response. Transpiration is suppressed below the PWP due to the increase in water stress experienced by plants whose roots are unable to extract any water from sufficiently dry soils. In the PWP-SOIL and DRY-SOIL simulations, the stronger surface sensible heat fluxes create a deeper, drier boundary layer with a higher cloud base. The resulting clouds generate wider rain shafts with greater amounts of rain aloft. Due to the relative dryness of the subcloud layer, more evaporative cooling occurs. Larger, stronger cold pools are therefore able to form.

2. Soils moister than the PWP yield a water vapor structure in which (1) near-surface air in the center of the cold pool is relatively dry, and (2) the ring of near-surface air surrounding the cold pool is anomalously moist. The water vapor rings are located outside of the cold pools, ahead of the corresponding gust fronts.

3. When soil is drier than the PWP, short-lived “puddles” of moist air fill the cold pools. The air inside the cold pools is approximately uniformly moist, and the air surrounding the cold pools is dry. It is speculated that these features are the result of hydrometeor evaporation within the cold pools’ parent rain shafts.

4. Cold pools are shorter lived in the PWP-SOIL and DRY-SOIL simulations than in the other three wetter-soil experiments due to a weaker wet-patch effect (Drager & van den Heever, 2017) and stronger turbulent mixing (Grant & van den Heever, 2018) in the PWP-SOIL and DRY-SOIL simulations.

These results are all modulated by soil moisture-induced changes in the partitioning of surface sensible and latent heat fluxes, which in turn affect boundary layer and cloud properties.

We speculate that the results obtained here are applicable to low-shear continental settings, particularly situations in which the following criteria are satisfied: (1) The soil moisture fluctuates above and below the PWP; (2) vegetation is present, such that the PWP becomes relevant; and (3) cumulus congestus clouds form and precipitate regardless of soil moisture content.
The results obtained here regarding cold pool structure and longevity suggest different roles for cold pools in regions with different soil moistures that should be considered in parameterizations of cold pools and convective initiation. Based on the decreased cold pool lifetime in the PWP-SOIL and DRY-SOIL simulations relative to the other three wetter-soil simulations, we can infer that cold pools may suppress convection in their interiors for shorter periods of time in low-soil moisture environments compared to high-soil moisture environments. Furthermore, the colocation of water vapor rings and updrafts in the DRENCHED-SOIL, WET-SOIL, and MID-SOIL simulations, combined with the lack of water vapor rings in the PWP-SOIL and DRY-SOIL simulations, suggests that the thermodynamic mechanism of cold pool-induced triggering of new convection becomes more important as soil moisture increases. Evidently, the relative lack of importance of the thermodynamic triggering mechanism in the PWP-SOIL and DRY-SOIL simulations does not prevent new convection from forming. Indeed, convection continues later into the evening hours in the PWP-SOIL and DRY-SOIL simulations than in the other three simulations (Figure S2). Further analysis is required to determine the mechanism(s) responsible for this extension of the diurnal cycle of convection.

Future research should further explore the environmental parameter space, including different initial temperature, moisture, and wind profiles and varying amounts of larger-scale forcing. It would also be useful to consider a variety of vegetation types (including no vegetation) and soil textures. Different environmental conditions may be expected to support different types of convection that could serve as a testbed for the generalizability of the results obtained here. We note that other land surface schemes treat transpiration and direct surface evaporation in a manner that differs from the one used here. For example, the scheme by Chen and Dudhia (2001) suppresses all direct surface evaporation below the PWP, whereas the LEAF-3 scheme used here does not. Therefore, intermodel comparisons in this regard would be appropriate.

The present study contributes to a growing body of literature on land surface effects on continental cold pools and highlights the interactions between surface heat fluxes, the boundary layer, and clouds that control cold pool properties. More practically, our results also emphasize the importance of accurately representing initial soil moisture conditions for forecasting cold pools, particularly in regions whose soil moisture content regularly transitions across the PWP. Finally, it is recommended that future cold pool parameterization efforts incorporate the effects of the land surface, such as soil moisture, on cold pools in order to better represent the cold pool dissipation process and time scale.

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