The Sensitivity of Simulated Shallow Cumulus Convection and Cold Pools to Microphysics

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ABSTRACT

The sensitivity of nested WRF simulations of precipitating shallow marine cumuli and cold pools to microphysical parameterization is examined. The simulations differ only in their use of two widely used double-moment rain microphysical schemes: the Thompson and Morrison schemes. Both simulations produce similar mesoscale variability, with the Thompson scheme producing more weak cold pools and the Morrison scheme producing more strong cold pools, which are associated with more intense shallow convection. The most robust difference is that the cloud cover and LWP are significantly larger in the Morrison simulation than in the Thompson simulation. One-dimensional kinematic simulations confirm that dynamical feedbacks do not mask the impact of microphysics. These also help elucidate that a slower autoconversion process along with a stronger accretion process explains the Morrison scheme’s higher cloud fraction for a similar rain mixing ratio. Differences in the raindrop terminal fall speed parameters explain the higher evaporation rate of the Thompson scheme at moderate surface rain rates. Given the implications of the cloud-cover differences for the radiative forcing of the expansive trade wind regime, the microphysical scheme should be considered carefully when simulating precipitating shallow marine cumulus.

1. Introduction

Significant precipitation with surface rain rates exceeding 1 mm h\(^{-1}\) is frequently observed from shallow cumuli in the Caribbean trade wind region (Short and Nakamura 2000; Nuijens et al. 2009; Snodgrass et al. 2009). Rain evaporation cools the surrounding air and generates convective downdrafts that form cold pools at the surface (Zuidema et al. 2012). The suppression of convection within the stable surface cold pool region

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downwind boundary (Li et al. 2014), the mesoscale cloud organization due to cold pools (Seifert and Heus 2013), and the aerosol effects on the organization of shallow cumulus convection through cold pools (Xue et al. 2008; Jiang et al. 2009). Li et al. (2014) discuss the ability of cold pool propagation to increase the humidity, coverage, and strength of subcloud updrafts near the cold pool downwind boundary, with a similar finding for tropical cold pools reported in Schlemmer and Hohenegger (2014). Xue et al. (2008) and Seifert and Heus (2013) report a sensitivity of the shallow cumulus cold pool mesoscale organization to imposed cloud drop number or aerosol number concentration, as well as domain size and grid spacing.

The parameterization of rain microphysics may influence the trade wind cumulus convection independent of aerosol effects. Stevens and Seifert (2008) examine the sensitivity of simulations of shallow cumulus convection to bulk microphysical schemes but do not examine rain evaporation and cold pools. Abel and Shipway (2007) improve the Met Office model’s representation of trade wind cumulus convection, primarily by reducing the specified cloud droplet number concentration $N_d$, thereby facilitating the autoconversion process; they do not focus on rain evaporation or cold pools. Shipway and Hill (2012a) use a one-dimensional (1D) kinematic framework to examine differences in warm-rain microphysics schemes, including rain evaporation, but the microphysical–dynamical feedbacks are explicitly neglected by construction. The relationship between microphysical parameterizations and cold pools has been assessed for deeper convection, such as midlatitude squall lines [Morrison et al. (2012); van Weverberg et al. (2012), and references therein], but not yet for cold pools produced by only warm rain. The development of shallow convection and precipitation is also subject to environmental conditions. Li et al. (2014) discuss how the interaction of the environmental wind shear in the subcloud layer with the cold pool circulation facilitates the access of the updrafts at the cold pool boundaries to the ambient humidity. Large-eddy simulation (LES) studies indicate that, as surface winds strengthen, the boundary layer and clouds deepen (Nuijens and Stevens 2012).

The goal of this study is to understand the shallow cumulus precipitation and cold pool properties in response to cloud and rain microphysics by comparing model simulations that are configured in the exact same setup, except for the microphysical scheme. The analyzed day of 19 January 2005 contains cold pools embedded within the more synoptically defined cloud lines of a dissipated cold front (Caesar 2005). Although this case does not fit the archetypical trade wind regime description, with cloud tops that remained below 4 km, it does represent one of the typical regimes of the wintertime Caribbean. This day has also been selected by other studies for examining trade wind regime cloud properties; for instance, the well-sampled cloud microphysics from this day has been studied and implemented to improve the Met Office’s Large Eddy Model depiction of precipitating shallow cumuli (Abel and Shipway 2007).

This study builds on the previous study (Li et al. 2014) of a nested Weather Research and Forecasting (WRF) Model simulation, with the exact same simulation configuration, except for the microphysical scheme. In addition to the double-moment rain microphysical scheme developed by Greg Thompson [an extension of Thompson et al. (2008)] used in Li et al. (2014), this study also employ the Morrison double-moment scheme (Morrison et al. 2005). With these two simulations, we explore the systematic differences in precipitating convection and their associated cold pools. These two schemes are chosen because both are commonly used available options within WRF, and Li et al. (2014) have already demonstrated that nested WRF simulations invoking the Thompson microphysical scheme produce cold pools comparable to those observed, if weaker and smaller. This study augments that of Van Zanten et al. (2011), who assessed the behavior of 12 LES models against a composite of the less-disturbed 16 December 2004–8 January 2005 RICO time period.

The clouds and precipitation produced by the Thompson and Morrison schemes are also evaluated within the one-dimensional Kinematic Driver model described by Shipway and Hill (2012a), with prescribed updraft velocities. This serves to assess the rain production process rates independently from interactions with other dynamical processes in WRF and allows for a more thorough investigation of the sensitivity of evaporation to the fall speed parameterization.

2. WRF simulations with Thompson and Morrison schemes

a. Description of the microphysical schemes

Aerosol effects are not examined, and the cloud droplet number concentration is specified at 100 cm$^{-3}$ in both simulations. This is the default value in the Thompson scheme and is slightly higher than the prescription of $N_d = 70$ cm$^{-3}$ in Van Zanten et al. (2011), but it falls within the $N_d$ range measured during RICO (Hudson and Noble 2014). Some additional comparisons are also done to a simulation using the Morrison scheme $N_d$ default of 250 cm$^{-3}$. Both schemes are
double moment in rain (prognosed rain mass and number) and single moment in cloud (prognosed mass only).

Autoconversion (the conversion of water from cloud to rain through collision and coalescence among cloud drops) and accretion (the further growth of the raindrops through collecting cloud drops) are parameterized collectively in the Thompson scheme following a modified version of Berry and Reinhardt (1974, hereafter BR). The Thompson double-moment rain scheme evaluated here is an undocumented evolution of a previous single-moment rain scheme (Thompson et al. 2008). BR schemes are explicitly reviewed by Gilmore and Straka (2008), who note that BR schemes are effective because they include information on both the mean cloud droplet size and cloud dispersion, but large rain mixing ratio growth-rate differences are possible between different implementations of the same scheme. The Morrison scheme is based on the separate autoconversion and accretion schemes of Khairoutdinov and Kogan (2000, hereafter KK). The KK scheme is parameterized from explicit microphysics produced by LESs of marine stratocumulus. The KK autoconversion and accretion parameterizations have straightforward dependencies upon the cloud $q_c$ and rain $q_r$ mixing ratios and $N_d$: the autoconversion rate is $1350q_c^{0.47}N_d^{-1.79}$, and the accretion rate is $67(q_c,q_r)^{1.15}$ (both in units of kilograms per cubic centimeter per second). The KK autoconversion rates, compared to those derived from observed stratocumulus drop size distributions (Wood 2005), are biased high at low rates, while the BR rates are biased low (R. Wood, 2014, personal communication) noted that Fig. 1 in Wood (2005) incorrectly depicts the Thompson–BR rates to be too low.

Rain self-collection and breakup can have a significant impact on surface precipitation rates for shallow cumulus (Stevens and Seifert 2008). Both the Morrison and Thompson schemes include rain self-collection, following Beheng (1994). This is an extension of the original KK parameterization within the Morrison scheme. Rain breakup is implemented within both schemes using slightly different variations of Verlinde and Cotton (1993) [see also Morrison and Milbrandt (2011)].

The KK scheme is employed within two different LES models participating in Van Zanten et al. (2011) and is also used within the Met Office Large Eddy Model simulation for this day (Abel and Shipway 2007). Shipway and Hill (2012a) compare the Thompson (as Thompson09) and Morrison schemes within a one-dimensional kinematic framework that we also employ and report more accumulated water due to more autoconversion, less accretion, and more evaporation for the Thompson scheme. A newer version of the KK scheme, which is developed from explicit microphysics simulations driven by the Van Zanten et al. (2011) composite case, has also been developed by (Kogan 2013) to optimize the scheme for trade wind cumulus. This has an autoconversion rate of $7.98 \times 10^{10}q_c^{4.22}N_d^{-3.01}$ and an accretion rate of $8.53q_c^{1.08}q_r^{0.98}$, both in units of kilograms per cubic meter per second. However, this scheme is not yet operationally available within WRF and is not evaluated here.

For the same environmental conditions, more evaporation should result in colder cold pools. Small drops evaporate more readily, and a parameterized rain size distribution that has more small drops with a higher total drop number concentration for a given rain rate will evaporate more efficiently. In both the Thompson and Morrison schemes, the raindrop size distribution (RSD) follows an inverse exponential distribution: $N(D) = N_0e^{-\lambda D}$, where $N(D)$ (m$^{-4}$) represents the number of raindrops of diameter $D$ (m), $N_0$ (m$^{-4}$) is the intercept parameter, and $\lambda$ (m$^{-1}$) is the slope parameter. The values of $N_0$ and $\lambda$ are diagnosed from the prognosed total number of raindrops along with the rain mixing ratio. The evaporation rate is then solved analytically. The Morrison scheme follows the steady-state convective diffusion solution of Rutledge and Hobbs (1983), and the Thompson scheme includes several higher-order terms of the supersaturation to minimize the error of the solution (Srivastava and Coen 1992).

b. Simulation setup

The simulations use the updated version 3.2 of the Weather Research and Forecasting Model with multiple nested domains (Li et al. 2014). The parent domain and four two-way nested inner domains are centered at 18°N, 61.7°W within the operational domain of the Research Vessel (R/V) Seward Johnson during the RICO experiment (Zuidema et al. 2012). The simulations span from 0000 UTC 19 January to 0600 UTC 20 January, with only the 1-min output from the last 24 h analyzed unless otherwise indicated. The R/V Seward Johnson radiosondes are assimilated to improve the depiction of the boundary layer vertical structure, and the NCEP Final Analyses (NCEP FNL; NOAA/National Centers for Environmental Prediction 2000) supply the initial and lateral boundary conditions. Only the simulation results from the innermost domain are analyzed. This has a domain size of 24 km × 24 km from the surface to 10 hPa at a constant horizontal grid spacing of 100 m and vertical grid spacing varying from 6 to 200 m between the surface and the 4-km level. The use of realistic large-scale forcing and the nudging to radiosondes enables the cold pools to be produced in a more realistic environment than.
that of LESs with doubly periodic lateral boundary conditions. In addition, it may overcome some of the deficiencies of doubly periodic LESs; for example, Van Zanten et al. (2011) find that their doubly periodic LESs tend to produce a uniform rain field and a lack of strong rain events. Other specifications of the simulations are described in Li et al. (2014).

In contrast to the one-way input mode in the more traditional LES, the two-way interactions in our simulations allow feedbacks from small to large scales. This means that the differences in microphysics are given more freedom to interact with mesoscale dynamics as well as the convective organization.

The simulated potential temperature $\theta$, water vapor mixing ratio $q_v$, and wind profiles averaged over the entire day and over three 4-h segments through the day are shown in Fig. 1. The averages over the three 4-h segments make clear that the boundary layer winds change significantly over the course of the day, with the near-surface winds becoming stronger and more northerly with time and northwesterly winds at 3 km transitioning to calmer southwesterlies. In addition, the atmosphere becomes more stably stratified at about 2 km as the day evolves. As would be expected from simulations nudged to the prevailing environmental conditions, the simulations produce rain similar to (if slightly exceeding) that observed. The area-averaged rainfall rate derived from a scanning precipitation radar for this day is 1.87 mm day$^{-1}$ (Snodgrass et al. 2009), equivalent to 53 W m$^{-2}$. The Thompson and Morrison simulations produce 1-day domain-averaged surface rain rates of 2.1 and 2.4 mm day$^{-1}$, respectively, equivalent to 60–70 W m$^{-2}$ of latent heating. For comparison, the Van Zanten et al. (2011) simulations corresponded to a mean radar-derived latent heating estimate of 21 W m$^{-2}$.

c. Cloud and rain in WRF simulations

The evolution of the domain maximum surface rain rate ($RR_{\text{max}}$), domain-averaged surface rain rate ($RR_{\text{davg}}$), domain-averaged accumulated surface rain, wind (speed) shear between 3 and 1 km, domain maximum vertical velocity $w_{\text{max}}$ and updraft mass flux at the cloud-base level ($M = \rho_u f_{up} w_{up}$, where $\rho_u$ is the air density, $f_{up}$ is the fractional area of grid points containing $w > 0.5$ m s$^{-1}$ at the cloud-base level, and $w_{up}$ is the mean updraft vertical velocity at the cloud-base level), cloud cover (defined as the vertical maximum fraction of grid points containing $q_c > 0.1$ g kg$^{-1}$ at each level for each minute), and domain-averaged liquid water path (LWP) of one Thompson simulation and two Morrison simulations with $N_d = 100$ and 250 cm$^{-3}$ are shown in Fig. 2. The at-times high domain maximum surface rain rates do find a counterpart in the observations, with one of the three rain events documented on the ship producing a surface rain rate of 45 mm h$^{-1}$ (Zuidema et al. 2012; Li et al. 2014). A notable difference is that the Thompson simulation produces an earlier significant rain event, followed by a quiescent period from 0800 to 1400 UTC, while the Morrison simulation produces the first significant rain event about 2 h later, followed by a shorter quiescent period from 1200 to 1400 UTC (Figs. 2a–c). After 1400 UTC, the cloud-base-level vertical velocity and updraft mass flux in the Thompson simulation rises to match that in the Morrison simulation (Figs. 2e,f), associated with the similar cloud-core fraction below 3 km in the two simulations (Fig. 3e). Nevertheless, the Morrison simulation LWP
and cloud fraction continues to exceed that from the Thompson simulation (Figs. 2g,h).

The cloud cover of the Thompson simulation is lower than that of the Morrison simulation for most of the integration period (Fig. 2g), with mean cloud cover over the last 24 h of 12.7% for Thompson versus 19.8% for Morrison and mean cloud cover over the last 16 h (after 1400 UTC) of 14.8% for Thompson versus 22.4% for Morrison. The difference in mean cloud cover between the two simulations is significant above the 99% confidence level. The microphysics-caused difference is also evident in LWP as well as the accumulated surface rain, with a mean LWP of 60 g m\(^{-2}\) for Morrison and 48 g m\(^{-2}\) for Thompson and a mean accumulated surface rain of 1.29 mm for Morrison and 0.95 mm for Thompson.

The domain maximum cloud-base vertical velocities slightly exceed those from aircraft observations on this day (Abel and Shipway 2007) as well as from a shipboard Doppler lidar (Zuidema et al. 2012), but this may represent sampling differences. In the observations, the shipboard rain measurements and scanning radar images show precipitation occurring most frequently between 1200 and 1800 UTC (Zuidema et al. 2012). This is better captured by the Morrison simulation with a later rain initiation and shorter quiescent period. In the latter half of the simulation, when the vertical wind shear becomes larger, the Morrison simulation using the higher \(N_d\) has more difficulty supporting precipitating convection with vertical velocities similar to other simulations, in keeping with known sensitivities to aerosol increases (e.g., Stevens and Seifert 2008; Xue et al. 2008).

Further insight is provided through examining the vertical structure of the cloud and rain fractions and mixing ratios shown in Fig. 3, averaged over the same 4-h segments as in Fig. 1. Cloudy regions are defined by a cloud water mixing ratio \(q_c > 0.1\) g kg\(^{-1}\), the convective core through cloudy regions with positive vertical velocity \((w > 0)\), and a positive buoyancy relative to the domain mean \((b = g \{[(\theta - \bar{\theta})/\bar{\theta}] + 0.61(q_v - \bar{q}_v) - q_L\} > 0\), where \(g\) is the gravitational acceleration, and \(q_L\) is the liquid water mixing ratio). Rainy regions are defined by rainwater mixing ratios \(q_r > 0.1\) g kg\(^{-1}\). The clouds extend up to 3 km in both simulations (Fig. 3a), with the Morrison simulation often extending higher and clearly possessing more cloud water than the Thompson simulation. The \(q_r\) maxima occurs within cloud layer, which is also consistent with the aircraft observations for this day (Abel and Shipway 2007), although it is not necessarily typical of observations spanning the whole campaign (Van Zanten et al. 2011). The aircraft rain measurements relied on 2D cloud (2D-C) and 2D precipitation (2D-P) probes, for which the smallest 2D-C size channel occurred at 40–50 μm, though the 50–100-μm size range is typically underestimated.

From 1400 to 1800 UTC, wind shear at 3 km (Fig. 2d) is evident as enhanced cloud detrainment, more pronounced for the higher-extending clouds in the Morrison simulation. As the simulation progresses, the cloud tops lower in both simulations (Figs. 3a,e,i). Given the strengthened surface winds, one might expect a deeper boundary layer (Nuijens and Stevens 2012). The lowered cloud tops instead appear to reflect a new inversion
advected in at 2 km of about 4 g kg\(^{-1}\) for specific humidity and 5 K for potential temperature within 300-m vertical depth, apparent in the individual radiosondes (Zuidema et al. 2012) and to some extent in Fig. 1. The Morrison simulation with \(N_d = 250 \text{ cm}^{-3}\) shows a higher cloud and rain mixing ratio reaching higher altitudes earlier in time (Figs. 3b,d) but less rain produced at all vertical levels as the day progresses. This may reflect the influence of the increasing wind shear.

The relationship of the cloud cover to the domain-averaged surface rain rate is shown statistically in Fig. 4. Cloud cover is clearly related to the average surface rain rate within both simulations, but more clearly within the Thompson simulation. The cloud cover within the Morrison simulation is generally greater, reaching 40% at times. At low rain rates, the Morrison cloud cover is almost independent of the surface rain rate. Van Zanten et al. (2011) did not find a strong correspondence between the cloud cover and precipitation for their overall weaker precipitation rates. This could be consistent with the Morrison results at low rain rates shown here. The clear increase in cloud cover with surface rain at higher rain rates may reflect the shearing off of upper cloud through wind shear, as seen in observations (Zuidema et al. 2012).

A more difficult aspect of the two simulations to assess is their fidelity to the observed mesoscale organization. Two-dimensional fast-Fourier transforms (2D FFT) provide one objective approach to comparing the two simulations and, in addition, allow for an assessment of which spatial scales provide more of the overall variance (e.g., Seifert and Heus 2013). Time series of power spectra of the vertical velocity and total water at 2-km height reveal more variance for the Morrison simulation than for the Thompson simulation in total water, but not in the vertical velocity (except between 0800 and 1200 UTC; figures not shown), consistent with Figs. 2 and 4. The greater variance in total water is not linked to a particular eddy size, consistent with an overall slower autoconversion rate.

We include two snapshots from when RR\(_{\text{max}}\) is 55 mm h\(^{-1}\) for Thompson and 45 mm h\(^{-1}\) for Morrison at 0300 UTC 20 January (Fig. 5). The mean power spectra of vertical velocity within the cloud layer do not differ
significantly between the two simulations, yet the snapshots do reveal more subtle differences. Horizontal snapshots of the second innermost domain at 0300 UTC show that both simulations capture the synoptic cloud line and are clearly producing cold pools (Fig. 6). Examination of cloud field and surface temperature movies indicates these features are commonly present in both simulations through the model integration time. The Thompson simulation contains two smaller cloud lines within the innermost domain, as compared to the longer cloud line produced by the Morrison simulation. The clouds within the Morrison simulation are more congregated than those in the Thompson simulation (Figs. 5a,d), and the surface cold pool corresponding to the domain maximum surface rain rate is colder and larger, with updrafts located at the cold pool boundaries (Figs. 5b,e). A cross section is taken along the 3-m-level domain-mean wind direction across the point of domain maximum surface rain rate for each simulation (Figs. 5c,f). The cloud associated with the heaviest rain in the Thompson simulation is tilted by the vertical wind shear, with the heaviest rain occurring in the middle of the cloud layer and the downdraft dislocated from the cloud-base-level updraft (Fig. 5e). This geometry provides a positive feedback for furthering secondary convection, explored more deeply in Li et al. (2014). The Morrison simulation is much cloudier, consistent with Fig. 3i. Cloud overlies much of the cold pool, in contrast to observations. Different from the Thompson simulation, the rain in the Morrison simulation occurs near the cloud top, which is sheared off by the change of wind direction above 2 km (Figs. 1c,d). This explains the reason that the vertical cross section across the maximum surface rainrate grid point for the Morrison simulation does not capture the cloud mass at cloud-base level (Fig. 5f).

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

The surface air buoyancy is modified by the cold pools present within the domain. More larger cold pools would result in greater coverage of negative buoyancy anomaly. The surface air buoyancy anomaly $\Delta \theta_s(x, y, z) = \theta_s(x, y, z) - \overline{\theta_s}(z)$, where $\theta_s$ is calculated as $\theta(1 + 0.61q_v - q_L)$ and $\overline{\theta_s}(z)$ represents the domain-averaged value. The Thompson simulation has a larger area of negatively buoyant air with $-0.5 < \Delta \theta_s < 0$ K than the Morrison simulation (Fig. 7a), implying more weak cold pools in the Thompson simulation. However, for the infrequent cases that are associated with high rain rates (e.g., Figs. 5, 6), the Morrison simulation produces colder cold pools, resulting in the greater fraction of strong negative buoyancy with $\Delta \theta_s < -0.5$ K (Fig. 7b).

The 3-m-level cold pool properties are examined according to their associated domain maximum surface rain rate (Fig. 8). The domain maximum surface rain rate is used as a reference for the convective intensity, as it correlates better with the maximum surface changes within cold pools than the domain-averaged surface rain rate, except for the surface change in $q_v$ that shows little linear correlation with either $RR_{max}$ or $RR_{da}$. These dependencies of cold pool properties on $RR_{max}$ also facilitate comparison to Li et al. (2014), where these relationships are compared to observations. For the majority of cases with $RR_{max} < 20$ mm h$^{-1}$, the Thompson simulation produces, on average, larger surface $\theta$ decreases ($-0.44$ vs $-0.36$ K for the Morrison) and, hence, buoyancy. For the less frequent instances with intense rain ($RR_{max} > 40$ mm h$^{-1}$; e.g., Fig. 5), the surface air temperature, buoyancy and wind responses are stronger in the Morrison simulation (e.g., average $\theta$ decrease $-0.84$ K for Thompson vs $-1.04$ K for Morrison). The linear best-fit lines in Fig. 8 help visualize how the aforementioned differences produced by the two microphysical schemes may shape the different correlation relationship of the cold pool properties with $RR_{max}$ in the two simulations.

The properties of cold pools are first and most dominantly determined by the convective intensity, as previously shown in observations (Zuidema et al. 2012) and captured by both the simulations (Fig. 8). The differences in cold pool production by the two schemes are magnified under close examinations that focus on moderate and intense convection, respectively. The
consistencies between Figs. 7 and 8 point to the conclusion that the Thompson scheme is more likely to produce colder cold pools under the more frequently occurring moderate rain condition, resulting in larger area with weak negatively buoyant air. However, the few intense rain cases should not be ignored, because they are responsible for a large portion of surface precipitation and often further convection (Zuidema et al. 2012). For these intense rain cases, the Morrison scheme produces colder and stronger cold pools.

The evaporation process should explain the different cold pool temperature depressions produced by the two

0300 UTC 20 January

Thompson

Morrison

FIG. 6. The 3-m-level air temperature (shaded) and nonzero cloud water path (green contours) for the (left) Thompson and (right) Morrison simulations at 0300 UTC 20 Jan in the outer simulation domain of the innermost domain. The black square indicates the boundaries of the innermost domain.
simulations as a function of rain rate. The column-integrated evaporation rate $E_{\text{col}}$ and the vertically integrated evaporation rate from surface to 1-km altitude $E_{1\text{km}}$ are shown as functions of surface rain rate for the grid point containing the domain maximum surface rain rate from each output minute (Fig. 9). Both $E_{\text{col}}$ and $E_{1\text{km}}$ are higher in the Morrison simulation than in the Thompson simulation for high rain rates ($RR_{\text{max}} > 40 \text{ mm h}^{-1}$). For columns with $RR_{\text{max}} < 20 \text{ mm h}^{-1}$, only $E_{1\text{km}}$ is slightly higher in the Thompson simulation than in the Morrison simulation (average 0.36 g m$^{-2}$ s$^{-1}$ for Thompson vs 0.32 g m$^{-2}$ s$^{-1}$ for Morrison), corresponding to slightly stronger temperature depressions (Fig. 8). Therefore, the surface changes within cold pools are likely more directly related to the rain evaporation within the lower boundary layer (below 1 km).

In Fig. 10, the vertical profiles of the rain mixing ratio $q_r$, evaporation rate $E_r$, and evaporation efficiency, defined as $E_r/q_r$, are shown for the composite over columns containing $5 < RR_{\text{max}} < 10 \text{ mm h}^{-1}$ (moderate) and $40 < RR_{\text{max}} < 80 \text{ mm h}^{-1}$ (intense). At the moderate rain rates, the Thompson scheme clearly produces a higher evaporation rate below 1 km, as well as a greater evaporation efficiency, as is consistent with Fig. 9. At the high rain rates, the rain mixing ratios and evaporation rates are less in the Thompson simulation than in the Morrison simulation, and the evaporation efficiency is much diminished. Since the larger rain rates occur less frequently (Fig. 8), the moderate cases should dominate the characteristics of rain evaporation in both simulations and lead to greater rain evaporation at low altitudes in the Thompson simulation, on average. The moderate rain results are consistent with Shipway and Hill (2012a), but the larger rain rates are more likely to be associated with cold-pool-organized cumulus.

3. 1D column model experiments

Experiments with the one-dimensional Kinematic Driver model (KiD; Shipway and Hill 2012a) using the Thompson and Morrison microphysical schemes confirm that the dynamical feedbacks are not masking the differences in rain and evaporation that can be attributed to the schemes themselves. The analysis also allows for a cleaner comparison on the parameterizations of the processes contributing to the rain mass water budget: autoconversion, accretion, sedimentation, and rain evaporation.

As rain falls, the RSD is altered by the processes of rain self-collection and raindrop breakup, and the raindrop fall speed associated with different drop sizes. Size sorting is handled somewhat differently by the two schemes, in part through different parameterizations of the fall speeds, and in part by different treatments of sedimentation. The Morrison scheme determines the sedimentation of the number and mass contents by applying analytic formulations to the number- and mass-weighted fall speeds. The Thompson scheme introduces an ad hoc increase in the number-weighted fall speed to reduce the difference from the mass-weighted fall speed and thereby reduce the size sorting. In this study, we experiment with modifying the fall speed parameterization only to examine how differences in parameterized fall speed affect precipitation and associated evaporation. Condensation onto the raindrops is negligible (Shipway and Hill 2012a) and is not considered in either scheme.

The microphysical processes in KiD are driven by prescribed vertical velocity profiles. All experiments are run for 5400 s, with a 1-s time step, and outputting prognostic variables every 10 s. The column is set to be 7 km high, with 120 full-$\text{z}$ levels and 119 half-$\text{z}$ levels at a vertical grid spacing of approximately 58 m.
The initial thermodynamic profiles resemble the average profiles from WRF simulations (Fig. 1). In the default setting of this model, as suggested in Shipway and Hill (2012a), the $\theta$ profile is held constant throughout the simulation to minimize the feedbacks from the two different schemes, and only the water vapor mixing ratio is updated. The imposed vertical velocity profiles resemble an updraft core modeled on data from Abel and Shipway (2007) [test case 5 in Shipway and Hill (2012b)], varying exponentially with height and time. The profiles reach their peak value $w_{\text{max}}$ at approximately the same altitude as the peak of averaged updraft profiles simulated by both the WRF simulations. For each microphysical scheme, experiments are carried out by varying $w_{\text{max}}$ from 0.2 to 2 m s$^{-1}$, encompassing the range observed with Doppler lidar (Zuidema et al. 2012).

a. Cloud and rain comparisons in KiD

Two KiD simulations are performed using the same updraft profiles and configured with the two microphysical schemes, respectively. The averaged updraft profile reaches its maximum $w_{\text{max}}$ of 0.25 m s$^{-1}$ at 1283 m. The surface rain rates produced by these two 1D model simulations are less than 10 mm h$^{-1}$, corresponding to the more moderate surface rain rates in the WRF simulations. The time-averaged profiles of the cloud and rain mixing ratios over cloudy ($q_c > 0.1 \text{ g kg}^{-1}$) and rainy ($q_r > 0.1 \text{ g kg}^{-1}$) points (Figs. 11a,b) are similar to the average profiles from the 3D WRF simulations (Fig. 3).

Consistent with the WRF simulations, the Morrison scheme clearly produces more cloud than the Thompson scheme (Fig. 11a). A breakdown by the rain budget processes reveals, for a similar rain mixing ratio (Fig. 11b), the autoconversion rate is higher for the Thompson scheme (Fig. 11c), and the accretion rate is higher for the Morrison scheme (Fig. 11d), along with a higher evaporation rate (Fig. 11e) corresponding to higher surface rain rate.

The significantly smaller autoconversion rate of the Morrison scheme (Fig. 11c) is capable of explaining the higher Morrison cloud fractions in the WRF simulations. This is especially true for the lower cloud layer (below 1.5 km in Fig. 11c). The autoconversion process of these two schemes is discussed explicitly in Shipway and Hill (2012a).

The accretion contributes to much more rain production than the autoconversion for both schemes (Fig. 12), as expected for cumulus clouds. The Morrison scheme’s accretion rate is more active than that of the Thompson scheme (Fig. 12), especially at higher altitude (Fig. 11d). This may explain why rain is present at higher altitudes in the Morrison WRF simulation than in the Thompson simulation (Fig. 3).

The Thompson evaporation rate is significantly greater than that of the Morrison scheme for a similar averaged rain mixing ratio (Fig. 11), which is also seen in the WRF simulation results (Fig. 10). The evaporation all occurs below 1.1 km in the KiD simulations, maximized near cloud base, in contrast to the near-surface maximum evaporation rate in the WRF simulations.
The rain and evaporation in WRF simulations occur in downdrafts that are not represented in the KiD simulations, likely causing the difference in vertical structure of the evaporation.

The time series comparison of the column-integrated rain production rate and evaporation rate for the two schemes are shown in Fig. 12. For the same dynamical forcing, the Morrison scheme produces larger surface rain rates, and less rain evaporation, than the Thompson scheme, consistent with Shipway and Hill (2012a).

b. KiD rain evaporation comparison and fall speed dependence

The KiD simulations driven by updraft profiles reaching $w_{\text{max}}$ of 0.25 m s$^{-1}$ are more akin to the moderate cases in the WRF simulations than the more intense rain cases. The two pairs of KiD simulations selected in this section represent moderate and intense rain cases, respectively. The updraft profiles of these four KiD simulations are constructed in the same way as the previously discussed KiD simulations, but with different $w_{\text{max}}$ to facilitate similar surface rain rates between Thompson and Morrison simulations for either the moderate or the intense rain cases. The distinctions in KiD $q_r$, $E_r$, and $E_r/q_r$ for moderate and intense rain cases are explored in Fig. 13, where each profile is a composite of profiles of 1000 s that correspond to surface rain rate close to 5 mm h$^{-1}$ (for the pair of moderate rain simulations) or 60 mm h$^{-1}$ (for the pair of intense rain simulations). The composite profiles are arranged similarly, relative to each other, as seen in the WRF simulations (Fig. 10). Consistent with the WRF simulations, when compared on the basis of surface rain rates, the moderate Thompson case has the greatest rain evaporation efficiency among all four cases, and the Morrison scheme produces more rain, more rain evaporation, and greater evaporation efficiency than the Thompson scheme for intense rain rates.

One factor that can influence the modeled rain evaporation is the parameterized number- and mass-weighted fall speed. The Thompson scheme parameterizes a higher fall speed than the Morrison scheme for raindrop sizes up to about 3 mm, with the maximum difference occurring around 2 mm (Fig. 14; note that the fall speed is truncated at 9.1 m s$^{-1}$). The effects of the fall speed difference are investigated by replacing the fall speed parameters in the Thompson scheme with the ones used in the Morrison scheme (thus, only the results from Thompson simulations may be affected). Figure 13c shows that the evaporation efficiency of the modified moderate-rainrate Thompson simulation is significantly reduced, matching that of the Morrison simulation at altitudes below 800 m. For the intense rain rates, the replacement of the fall speed parameterization only slightly reduces the rain mixing ratio and rain evaporation of the Thompson simulation, with no discernible impact on the evaporation efficiency (Fig. 13).

These changes are likely explained through the treatment of size sorting. The number-weighted raindrop fall speeds are applied to the sedimentation of $N_d$, while the mass-weighted fall speeds are applied to the sedimentation of $q_r$. The reduction of the fall speeds...
may reduce the ratio of mass- and number-weighted fall speeds and, in turn, reduce raindrop size sorting. This change in size sorting may explain the higher concentration of larger drops and lower concentration of smaller drops for the RSD at the cloud-base level in the moderate Thompson simulation (Fig. 15a; compare the black and green lines). The change of the RSD expresses itself as an increase in the rain mixing ratio and a decrease in the rain evaporation rate, since larger drops evaporate less efficiently. When the rain rates are high, the Morrison KiD simulation possesses a greater number of both small and total raindrops at cloud base than the Thompson simulation (Fig. 15b), consistent with greater rain mixing ratio, greater evaporation rate, and greater rain evaporation efficiency (Fig. 13). In the WRF simulations, although the Morrison simulation has a greater mean raindrop size than the Thompson simulation, the total concentration is also greater, consistent with the enhanced evaporation for intense rain rates but a smaller difference in evaporation efficiency (Fig. 15c). The lack of sensitivity to changes in the fall speed parameterization at the higher rain rates may also reflect the limits introduced in the Thompson scheme to reduce size sorting.

4. Conclusions and discussion

In this study, the impact of microphysics on precipitating shallow convection and cold pools is assessed by examining the characteristics of rain production and evaporation within two popular microphysical schemes available in WRF—namely, the Thompson and Morrison schemes—using nested WRF simulations of one particular day. The Thompson scheme is applied within Li et al. (2014) to examine secondary convection at cold pool boundaries, with this study providing a measure of...
Li et al. (2014) findings to the microphysical scheme employed.

The two schemes have different influence on cold pool characteristics for moderate and intense rain cases. For moderate rain rates, the Thompson simulation contains more colder and stronger cold pools, consistent with more rain evaporation and smaller mean drop sizes. The situation is reversed for the intense rain rates that occur much less often but contribute importantly to the overall rain rates. The Morrison scheme generates more rain evaporation for the very intense rain cases. One-dimensional kinematic model experiments show the same relationships among the Thompson and Morrison simulations for moderate and intense rain cases. The consistency between the 1D model and WRF simulations indicates that the dynamical feedbacks do not mask the impact of microphysics on precipitation and evaporation.

The sensitivity of the evaporation to the fall speed parameterization is assessed in the 1D model by replacing the raindrop terminal fall speed parameters in the Thompson scheme with the ones used in the Morrison scheme. The Thompson scheme parameterizes faster raindrop fall speeds than does the Morrison scheme. With a lowered fall speed, the modified moderate Thompson rain mixing ratio increases, and the number of small raindrops decreases, decreasing evaporation and the evaporation efficiency to match that of the Morrison simulation. For the very intense rain cases that occur much less frequently in the 3D simulations, however, fall speed differences are not responsible for the differences in rain mixing ratio, evaporation rate, and evaporation efficiency between the two schemes.

Cold pools are related to observed cloud mesoscale organization; however, the relative ability of the two microphysical schemes to produce a realistic cloud and rain mesoscale organization is more difficult to assess. Although the detailed appearance and structure of
Mesoscale cloud organizations are different in the two simulations, the power spectra of both dynamic (winds) and thermodynamic ($\theta$, $q_t$) variables in terms of scales, particularly in the mesoscale range ($10^3$–$10^4$ m) obtained from 2D FFT analyses do not show a substantial difference between the two simulations. A visual examination of the simulation outputs suggests that the Thompson depiction of vertical cloud distribution is closer to radar observation, while the larger and stronger cold pools generated by the Morrison scheme are closer to the ranges evident in satellite imagery (Zuidema et al. 2012).

The most robust difference in cloud production between the two simulations is the higher average cloud fraction within the Morrison simulation (19.8% vs 12.7% for the Thompson scheme). Clouds in the Morrison simulation also reach higher altitudes more often, where they are subsequently sheared. The cloud cover is correlated to the surface rain rates in both simulations but more clearly in the Thompson simulation. At low rain rates, the Morrison cloud cover is almost independent of the surface rain rate. The differences in cloud cover and LWP are significantly larger than the difference caused by random initial conditions discussed in Grabowski (2014); thus, it should not be simply considered as a sort of uncertainty, but, rather, the outcome resulted from differently parameterized microphysical processes. The higher cloud fraction in the Morrison simulation is consistent with its slower autoconversion rate. Recently, Kogan (2013) has modified the KK scheme to better represent shallow cumulus clouds. The newer autoconversion parameterization depends more strongly on $N_d$ and the cloud water mixing ratio than does the current KK scheme. The autoconversion efficiency is reduced, and significantly less cloud is

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**FIG. 14.** The parameterized terminal fall speed of raindrops over the size spectrum 100 $\mu$m–3 mm for the Thompson (black) and the Morrison (red) scheme.

**FIG. 15.** The cloud-base-level (460 m) RSDs corresponding to (a) moderate KiD simulations by the Thompson scheme (black solid), Morrison scheme (red solid), and Thompson scheme with fall speed parameter replaced (green solid) and (b) intense KiD simulations by the Thompson scheme (black dotted), Morrison scheme (red dotted), and Thompson scheme using the Morrison fall speed parameterization (green dotted). The RSDs are averaged over the 1000 s corresponding to Fig. 13. (c) The cloud-base-level (450 m) RSDs in the WRF Thompson (black dotted) and Morrison (red dotted) simulations, averaged over grid columns with $RR_{max}$ between 50 and 70 mm h$^{-1}$.
produced overall, particularly near cloud top (Kogan 2013). If this parameterization were implemented in the Morrison scheme, the cloud fraction should be less and likely more similar to the Thompson-simulated cloud distributions. The Kogan (2013) accretion efficiency is also larger than in the earlier scheme. This newer parameterization is not investigated as part of this study, but we do suggest it for further simulations of precipitating shallow cumulus.

A further consideration is the breadth of the cloud and raindrop size distributions. The shape of the cloud drop spectrum is important for establishing the autoconversion rate within the Thompson scheme (the Morrison autoconversion depends on the number and mass of the cloud droplets but not the spectral width). Both schemes prescribe cloud droplets according to the observational analysis of Martin et al. (1994) for marine stratocumulus (dispersion is 0.278 for a gamma shape parameter $\mu = 12$). This adequately captures the dispersion of the RICO shallow cumulus drop size distributions at cloud base, but the dispersion increases significantly at higher altitudes within the cloud (Arabas et al. 2009). As such, the Thompson scheme should underestimate the autoconversion rates above cloud base, all else being equal. Neither scheme considers a raindrop size distribution that incorporates a nonzero $\mu$, but Geoffroy et al. (2014) confirm that an inverse exponential distribution fits in situ RSDs from RICO well. Nevertheless, exponential raindrop size distributions encourage excessive size sorting that can produce peak surface precipitation rates exceeding those from other schemes when employed in double-moment schemes (Wacker and Seifert 2001; Shipway and Hill 2012a). Simulations in which $\mu$ is allowed to differ from zero produce vertical rain mass mixing ratio profiles that are more weighted to the cloud base rather than to the surface (Stevens and Seifert 2008; Kogan 2013).

The assessment provided here adds to the growing literature evaluating the ability of high-resolution models to emulate the mesoscale organization of precipitating shallow cumulus. Neither scheme evaluated here is ideally suited for precipitating shallow marine cumulus, and the substantial difference in cloud cover and LWP has profound implications for the model radiative balance of the trade wind regime, suggesting more work remains to be done. A useful initial further step would be to evaluate microphysical parameterizations developed specifically for shallow marine cumulus, such as Kogan (2013), with the in situ microphysical datasets gathered during RICO and other shallow marine cumulus deployments, similar to the marine stratocumulus assessments of Wood (2005). A BR-based scheme would likely benefit from a revisitation of dedicated bin-model experiments (Gilmore and Straka 2008), such as those of Kogan (2013). Further dedicated simulations based on such assessed microphysics would contribute to a more confident understanding of the microphysical–dynamical feedbacks for the strongly precipitating shallow marine cumulus that influence a large portion of our planet. In particular, the influence of wind shear on cold pool mesoscale organization in the climatological trade wind region would benefit from further assessment through ensembles of simulations, building on the simulations evaluated here.

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