

Concurrent Sensitivities of an Idealized Deep Convective Storm to Parameterization of Microphysics, Horizontal Grid Resolution, and Environmental Static Stability

HUGH MORRISON

National Center for Atmospheric Research, Boulder, Colorado*

ANNARELI MORALES⁺

Department of Atmospheric Science, Colorado State University, Fort Collins, Colorado

CECILLE VILLANUEVA-BIRRIEL[#]

Department of Earth, Atmospheric, and Planetary Sciences, Purdue University, West Lafayette, Indiana

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ABSTRACT

This study investigated the sensitivity of idealized deep convective storm simulations to microphysics parameterization, horizontal grid spacing (Δx), and environmental static stability. Three different bulk microphysics schemes in the Weather Research and Forecasting Model were tested for Δx between 0.125 and 2 km and three different environmental soundings, modified by altering static stability above 5 km. Horizontally and temporally averaged condensation and surface precipitation rates and convective updraft mass flux were sensitive to microphysics scheme and Δx for all environmental soundings. Microphysical sensitivities were similar for $0.125 < \Delta x < 1$ km, but they varied for different soundings. Sensitivities of these quantities to Δx were less robust and varied with microphysics scheme. Other statistical convective characteristics, such as the mean updraft width and strength, exhibited similar sensitivities to Δx for all of the microphysics schemes. Microphysical sensitivities were primarily attributed to interactions between microphysics, cold pools, and dynamics that affected the spatial coverage of convective updrafts and hence the horizontally averaged convective mass flux, condensation rate, and surface precipitation. However, these linkages were less clear for the lowest convective available potential energy (CAPE) sounding, and in this case other mechanisms compensated to give a similar spatial coverage of convective updrafts even in simulations without a cold pool. For higher CAPE, there was considerable production of rimed ice from all of the microphysics schemes and its assumed characteristics, especially the fall speed, were important in explaining sensitivity via microphysical impacts on the cold pool. These results highlight the need for continued improvement in representing the production of rimed ice and its characteristics in microphysics schemes.

* The National Center for Atmospheric Research is sponsored by the National Science Foundation.

⁺ Current affiliation: Department of Atmospheric, Oceanic, and Space Sciences, University of Michigan, Ann Arbor, Ann Arbor, Michigan.

[#] Current affiliation: Georges Lemaître Centre for Earth and Climate Research, Université Catholique de Louvain, Louvain-la-Neuve, Belgium.

Corresponding author address: Hugh Morrison, National Center for Atmospheric Research, 3090 Center Green Dr., Boulder, CO 80301.

E-mail: morrison@ucar.edu

1. Introduction

Numerous studies have investigated the sensitivity of simulated deep convective storms to the parameterization of microphysics, horizontal grid spacing (Δx), or environmental thermodynamic conditions using non-hydrostatic models at “convection permitting” scales (i.e., Δx on the order of a few kilometers or less). Documenting and understanding these sensitivities is important since convection-permitting models are now used routinely for numerical weather prediction (NWP) at many research and operational centers (Kain et al. 2008; Lean et al. 2008; Weisman et al. 2008; Clark et al. 2012).

Climate modeling is also now performed widely at these scales, using global cloud-resolving models (CRMs) (e.g., [Miura et al. 2007](#)) or by embedding CRMs in each grid cell of traditional general circulation models (i.e., “superparameterization”) (e.g., [Grabowski 2001](#); [Khairoutdinov and Randall 2001](#); [Tao et al. 2009](#)).

The parameterization of microphysics is critical in such models because it drives latent heating and condensate loading, which are directly coupled with the buoyancy and model dynamics. Thus, studies have shown considerable sensitivity of deep convective storm simulations to microphysics (e.g., [Tao and Simpson 1989](#); [Liu et al. 1997](#); [Tao et al. 2003](#); [Gilmore et al. 2004](#); [van den Heever and Cotton 2004](#); [Liu and Moncrieff 2007](#); [Rajeevan et al. 2010](#); [Morrison and Milbrandt 2011](#); [Bryan and Morrison 2012](#), hereafter [BM12](#); [van Weverberg 2013](#)). The representation of both ice-phase and liquid-phase processes is important for deep convection. Ice microphysics is complicated by the wide variety of ice particle types, and the representation of these ice particles has a large impact on model simulations (e.g., [McCumber et al. 1991](#); [Gilmore et al. 2004](#); [Morrison and Milbrandt 2011](#); [van Weverberg et al. 2012](#)). Previous studies have also documented sensitivity to parameterization of liquid microphysical processes, such as raindrop breakup and evaporation, primarily by their effect on cold pools (e.g., [Ferrier et al. 1995](#); [Gilmore and Wicker 1998](#); [Morrison and Milbrandt 2011](#); [van Weverberg et al. 2012](#); [Morrison et al. 2012](#)).

Horizontal grid resolution can have a strong effect on convective storm development and structure in models (e.g., [Weisman et al. 1997](#); [Grabowski et al. 1998](#); [Bélair and Mailhot 2001](#); [Petch and Gray 2001](#); [Petch et al. 2002](#); [Adlerman and Droegeleier 2002](#); [Bryan et al. 2003](#); [Gentry and Lackmann 2010](#); [Fiori et al. 2010](#); [BM12](#); [Verrelle et al. 2015](#)). For example, [Weisman et al. \(1997\)](#) showed that $\Delta x = 4$ km simulations were able to capture the mesoscale structure and evolution of 1-km simulations, but coarser grid resolution simulations produced slower-evolving systems because of an inability to resolve the nonhydrostatic dynamics. [Bryan et al. \(2003\)](#) analyzed idealized squall-line simulations with Δx between 0.125 and 1 km and found that the lower-resolution simulations had a similar overall structure compared to the 0.125-km simulation, but the ability of the model to resolve large turbulent eddies at $\Delta x = 0.125$ km led to substantial quantitative differences in dynamical and microphysical quantities. They argued that Δx of order 100 m is needed for traditional subgrid-scale closure in large-eddy simulation (LES) to perform according to their design (by resolving part of the inertial subrange). Other studies have shown evidence for the beginning of statistical convergence in bulk quantities

(e.g., horizontally averaged vertical fluxes) as Δx was reduced to ~ 200 – 500 m ([Fiori et al. 2010, 2011](#); [Langhans et al. 2012](#); [Verrelle et al. 2015](#)). (Note that the definition of convergence varied somewhat in these studies.) In idealized numerical tests of a rising thermal in a statically stable environment, [Craig and Dornbrack \(2008\)](#) found statistical convergence for $\Delta x \sim 0.2L_{\text{buoy}}$, where L_{buoy} is a length scale related to the ratio of the thermal buoyancy and stratification (a few hundred meters under typical conditions), when L_{buoy} is smaller than the initial thermal size. In their study, convergence was defined as being achieved when differences in statistical quantities with different resolutions were comparable to differences between runs with small perturbations to initial conditions. However, resolution requirements for statistical convergence in simulating moist deep convection under less idealized conditions remain uncertain.

While several previous studies have investigated sensitivity to microphysics or Δx , there have been relatively few studies examining how changes to the microphysics and grid resolution together may affect simulations. For example, it is unclear if sensitivities to microphysics at relatively coarse resolution are similar to those at higher resolution. Similarly, the sensitivity to Δx may differ with the parameterization of microphysics. Furthermore, many sensitivity studies of moist deep convection have focused on a single case, often represented by a homogeneous initial environmental thermodynamic profile [we note that some studies of microphysical sensitivities have investigated the impact of thermodynamic sounding, e.g., [van Weverberg \(2013\)](#)]. It is well known that environmental conditions exert a dominant control on convective characteristics. Given close coupling of the convective dynamics and microphysics, it is reasonable to suspect that sensitivities of simulated deep convective storms to microphysics or Δx might vary in different environments.

A few studies have investigated concurrent sensitivities to microphysics and grid resolution. [Fiori et al. \(2011\)](#) studied the sensitivity of a supercell storm to Δx between 0.2 and 1 km and two different settings of a one-moment microphysics scheme that represented the rimed ice category with characteristics of either graupel or hail (as well as different turbulence closures). They found that the microphysics setting had some impact on the convergence properties with regard to Δx , with the hail-driven simulations showing a reduced spread as Δx was reduced compared to the graupel-driven simulations. [BM12](#) explored the sensitivity of a midlatitude squall line in a high CAPE environment (4200 J kg^{-1}) to different microphysics settings within a two-moment bulk scheme and Δx between 0.25 and 4 km. They found that surface precipitation was sensitive to both Δx and microphysics. For example, cloud water evaporation increased and surface

precipitation decreased with increasing resolution, which was attributed to smaller updraft sizes and greater lateral entrainment in cloudy updraft cores. They also found that sensitivity to microphysics was qualitatively similar across the range of Δx , while sensitivity to Δx was similar for the different microphysics settings.

In the current study we also investigate concurrent sensitivities to microphysics parameterization and Δx . However, this study is unique because it examines these sensitivities over a range of thermodynamic soundings for the initial environmental conditions and provides detailed analysis of microphysical and dynamical mechanisms driving these sensitivities. This study also differs from [Fiori et al. \(2011\)](#) and [BM12](#) in that it examines sensitivity to different microphysics schemes instead of parameters within a single scheme, and simulates a different storm type (an isolated storm in an unsheared environment compared to a supercell or squall line). Idealized storms are simulated here since this allows us to vary the environmental conditions systematically, and to clearly identify the key driving mechanisms and interactions in a simplified setting.

The overall goals of this study are to quantify and understand sensitivities to parameterization of microphysics and Δx for simulations of moist deep convection, and to assess generality when these sensitivities are examined concurrently and under environments with differing CAPE. Specifically, this research compares three bulk microphysics schemes (WSM6, Thompson, and Morrison) available in the Weather Research and Forecasting (WRF) Model run at five different horizontal grid spacings: $\Delta x = 0.125, 0.25, 0.5, 1$, and 2 km. All combinations of microphysics scheme and Δx are tested for three environmental soundings with CAPE of $1534, 2800$, and 4229 J kg^{-1} . The analysis focuses on two main themes. First, sensitivity of the horizontally and temporally averaged surface precipitation rate to Δx , microphysics, and thermodynamic sounding is examined by analysis of the bulk-condensed water budget and precipitation efficiency. Microphysical–dynamical interactions, especially linkages between the domain-mean condensation rate and convective updraft mass flux, are in turn investigated to explain changes in the condensed water budget. This analysis shows a connection between the convective updraft mass flux and cold pools, whose characteristics vary with different microphysics schemes. In this way we explain changes in surface precipitation through a chain of microphysical and dynamical process interactions. Second, we focus on the sensitivity of statistical convective updraft characteristics to Δx , microphysics scheme, and sounding, including the mean diameter, strength, and number of updraft cores.

The paper is organized as follows. [Section 2](#) describes the experimental design, including the model description and setup and a brief description of the microphysics parameterizations. [Section 3](#) presents results, centered on the chain of process interactions driving the sensitivity of surface precipitation through analysis of the condensed water budget, precipitation efficiency, convective mass flux, and cold pools, as well as sensitivity of other statistical convective updraft properties. A summary and conclusions are given in [section 4](#).

2. Experimental design

a. Model description and setup

The Advanced Research WRF version 3.3.1 is used for the simulations in this study. WRF is a compressible, three-dimensional (3D) nonhydrostatic atmospheric model ([Skamarock et al. 2008](#)). Here, radiative transfer, surface energy fluxes, and surface drag are neglected for simplicity. Lateral boundary conditions are open, and the surface and model top are free slip. Note that because of open lateral boundaries there is significant domain-mean vertical motion (of order 10 cm s^{-1}) that develops, and its magnitude varies among simulations. While domain-mean vertical motion can be limited by using periodic lateral boundaries, open boundary conditions have the advantage of allowing gravity waves to propagate out of the domain. WRF simulations are performed with Δx of $0.125, 0.25, 0.5, 1$, or 2 km. The vertical grid spacing is the same for all simulations, allowing us to focus on sensitivities to Δx : 75 vertical levels with approximately $260\text{--}300$ m spacing in the troposphere stretched to $300\text{--}450$ m in the stratosphere. This vertical grid spacing was chosen to try and resolve turbulent flow for the higher grid spacing simulations (Δx of 250 and 125 m), based on the finding from [Bryan et al. \(2003\)](#) that horizontal and vertical grid spacings of ~ 250 m or less are needed to resolve an inertial subrange, using a model with similar numerics. This vertical grid spacing may limit the ability to resolve isotropic turbulent eddies in the $\Delta x = 125$ -m simulation, but a detailed investigation of sensitivity to both horizontal and vertical grid spacing is beyond the scope of this study. The horizontal domain size is $80 \times 80 \text{ km}^2$ with a vertical domain height of 22 km. A Rayleigh damper with damping coefficient of 0.003 s^{-1} is applied to the top 5 km. All simulations are run for 2 h with a time step of 0.75 s. Statistical results are calculated from either 1 - or 5 -min output. Fifth- and third-order advection is applied horizontally and vertically, respectively, with limiters to ensure monotonicity ([Wang et al. 2009](#)). Subgrid-scale horizontal and vertical mixing is calculated using a

1.5-order prognostic turbulent kinetic energy scheme (Skamarock et al. 2008). This is inconsistent with the design of such mixing schemes for traditional LES closures for grid spacings greater than ~ 250 m (cf. Bryan et al. 2003), but is applied here to allow for a straightforward comparison as Δx is modified, and for consistency with previous studies that have applied similar subgrid mixing schemes to test microphysics and grid spacing sensitivities at Δx of $O(1)$ km (e.g., Bryan et al. 2003; Morrison and Milbrandt 2011; BM12). More traditional planetary boundary layer (PBL) vertical mixing parameterizations may be needed to properly represent subgrid-scale mixing for Δx at scales larger than the PBL depth, typically of $O(1)$ km, but detailed investigation of this issue is beyond the scope of this study.

The initial environmental conditions are horizontally homogeneous and based upon the widely employed thermodynamic profile of Weisman and Klemp (1982), with a CAPE of approximately 2715 J kg^{-1} based on the most energetic parcel (Fig. 1b). While this sounding is moist and with limited convective inhibition compared to typical midlatitude continental soundings, we use it for a few reasons. First, it provides continuity with several previous studies investigating sensitivity to Δx or microphysics that have employed this sounding (e.g., Weisman et al. 1997; Bryan et al. 2003; Morrison et al. 2009). Second, it has the advantage of allowing initiation of secondary convection in an unsheared environment, which is the focus of this study, rather than convection driven by the initial thermal, which is highly constrained by the specified thermal characteristics. Since this sounding is fairly moist throughout the troposphere, the role of entrainment and evaporation may be limited, which could affect sensitivity to Δx in particular (BM12). This issue is discussed further in section 4.

Two more soundings are generated by modifying the control sounding, giving either lower (1534 J kg^{-1}) or higher (4229 J kg^{-1}) CAPE (Figs. 1a,c). To ensure similar convective initiation in the lower troposphere, the low-level (surface to 5 km) potential temperature θ and water vapor mixing ratio q_v profiles are identical for the three initial soundings; CAPE is varied by modifying static stability above 5 km. For the lower CAPE sounding, $d\theta/dz$ is increased by 2.143 K km^{-1} from 5 to 12 km. For the higher CAPE sounding, $d\theta/dz$ is decreased by 1.429 K km^{-1} from 5 to 12 km. These modifications correspond to differences in θ , relative to the control sounding, that reach a maximum of $+15 \text{ K}$ (low CAPE) or -10 K (high CAPE) at 12 km. Here $d\theta/dz$ is unmodified above 12 km. Note that the q_v profiles and, hence, column-integrated precipitable water of the environment are identical for all soundings. The initial wind shear is zero, and the domain is initially motionless.

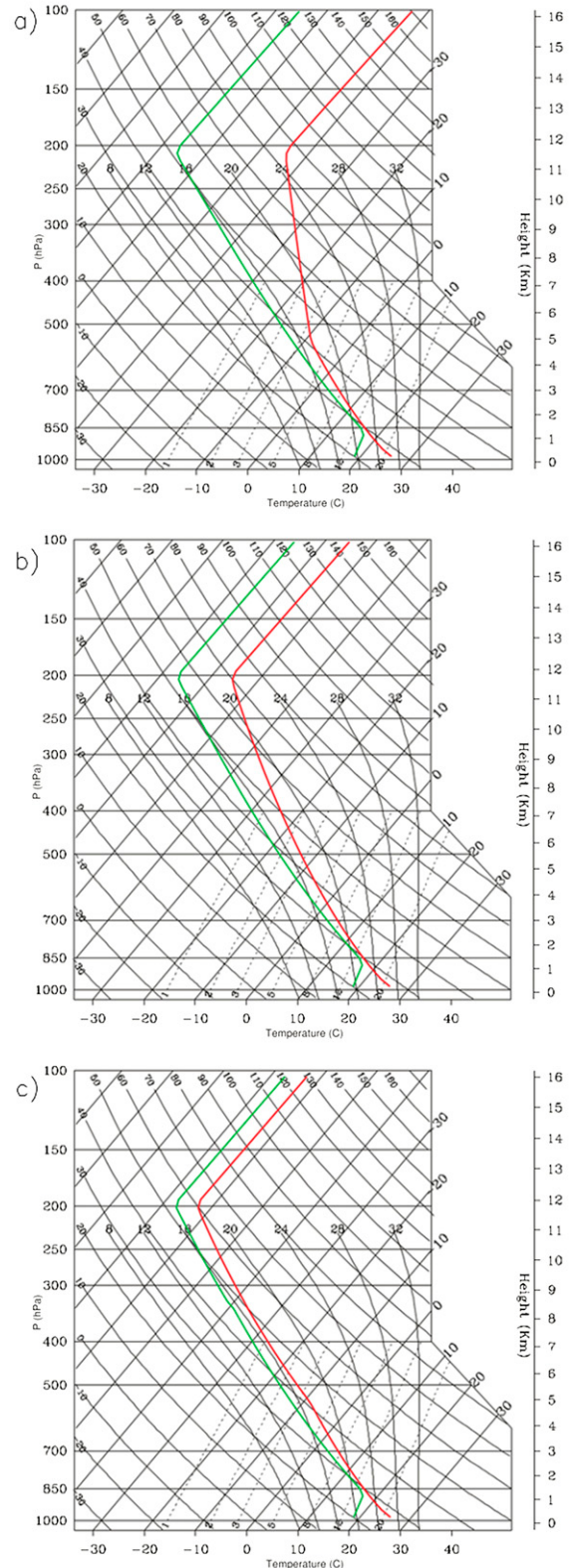


FIG. 1. Skew T -log p diagrams for the three soundings: (a) low CAPE, (b) moderate CAPE, and (c) high CAPE.

TABLE 1. List of microphysical schemes and their attributes.

Scheme	Reference	Mass variables*	Number variables*	Mixed-phase processes
WSM6	Hong and Lim (2006)	Q_c, Q_r, Q_i, Q_s, Q_g		Y
THO	Thompson et al. (2008)	Q_c, Q_r, Q_i, Q_s, Q_g	N_i, N_r	Y
MORR	Morrison et al. (2009)	Q_c, Q_r, Q_i, Q_s, Q_g	N_r, N_i, N_s, N_g	Y

* Mass mixing ratio (Q_x) and number mixing ratio (N_x) are considered for the hydrometeor species, where $x = c, r, i, s$, or g (cloud water, rain, cloud ice, snow, and graupel/hail, respectively).

Convection in the model is first initiated using a thermal bubble with small, random perturbations added to the theta field (ranging between ± 0.05 K) to initiate smaller-scale motions and storm asymmetry. The initial thermal is spheroidal with a horizontal radius of 10 km and vertical radius of 1.5 km centered at a height of 1.5 km and maximum perturbation potential temperature of 3 K. Convective characteristics (mean updraft width and strength) vary rapidly in time during the early phase dominated by the initial thermal. Thereafter convective updrafts initiate “naturally” without a direct influence from the initial thermal, and statistical properties of the convective updrafts are steadier over time. Our analysis, therefore, focuses primarily on the period from 60 to 120 min to minimize the effects of convection driven directly by the initial thermal. Additional sensitivity tests using the moderate CAPE sounding but with a reduced maximum perturbation potential temperature of 2 K for the initial thermal give similar results in terms of sensitivity to microphysics scheme (not shown). Moreover, these simulations produce a similar trend with regard to slower overall storm evolution for low- compared to high-resolution ($\Delta x = 2$ km vs 125 m). However, the initial convection takes longer to decay in the simulations with a weaker initial thermal, and, hence, it is difficult to compare results directly in terms of sensitivity to Δx to those with a stronger initial thermal over the same averaging period.

b. Microphysics parameterizations

A range of parameterizations is considered for investigating sensitivity to microphysics. For this study, the WRF single-moment 6-class (WSM6), Thompson (THO), and Morrison (MORR) schemes are used. These schemes differ widely in complexity; namely, WSM6 is one moment for all hydrometeor species (predicting mass mixing ratios only); Thompson is one moment for cloud water, snow, and graupel and two-moment for cloud ice and rain; and Morrison is two-moment for all species except cloud water (here we specify hail instead of graupel characteristics for the rimed ice species in MORR, except for sensitivity tests as noted). The corresponding mass mixing ratios Q_x and number mixing ratios N_x for the water and ice species

that are prognosed by each scheme are listed in [Table 1](#). Despite these differences, the schemes are similar in terms of the number of hydrometeor species and inclusion of microphysical processes for warm, mixed-phase, and ice conditions (e.g., vapor diffusion growth/shrinkage, autoconversion, accretion, riming, freezing, and melting), although the formulations for these processes generally differ among the schemes. The cloud droplet concentration is set to 400 cm^{-3} in the two schemes that specify this parameter (THO and MORR). Radar reflectivity is calculated assuming Rayleigh scattering following the approach of [Smith \(1984\)](#) using the specified or predicted size distribution and particle density parameters consistent with each scheme. For completeness the schemes are each described briefly in the [appendix](#).

3. Results

Moist deep convection is initiated in all simulations by the warm bubble within the first few minutes of integration. Precipitation reaches the surface within 25 min. Evaporation and melting of precipitation contribute to the development of a weak cold pool and downward motion in the domain center at low and midlevels. New convective cells are initiated and organize in a circular pattern along the outflow boundary after about 50–60 min. [Figures 2 and 3](#) illustrate the evolution of lowest model reflectivity and vertical velocity at a height of 6.1 km using the MORR microphysics scheme, moderate CAPE sounding, and Δx of 0.125 and 2 km, respectively. Overall storm evolution and structure is broadly similar among the simulations. However, there are differences in the detailed structure of the reflectivity and dynamical fields with different Δx , microphysics schemes, and thermodynamic soundings. For example, convective cells are initiated away from the main storm region in the $\Delta x = 125$ -m simulation shown in [Fig. 2](#) by $t = 120$ min (e.g., the cell near $X = 42$ and $Y = 18$ km), while these cells are absent in the corresponding $\Delta x = 2$ -km simulation.

Temporal evolution is further illustrated in [Fig. 4](#) by time series of surface precipitation rate (PRE), vertically integrated total condensation (COND; including conversion from vapor to condensate for all hydrometeor

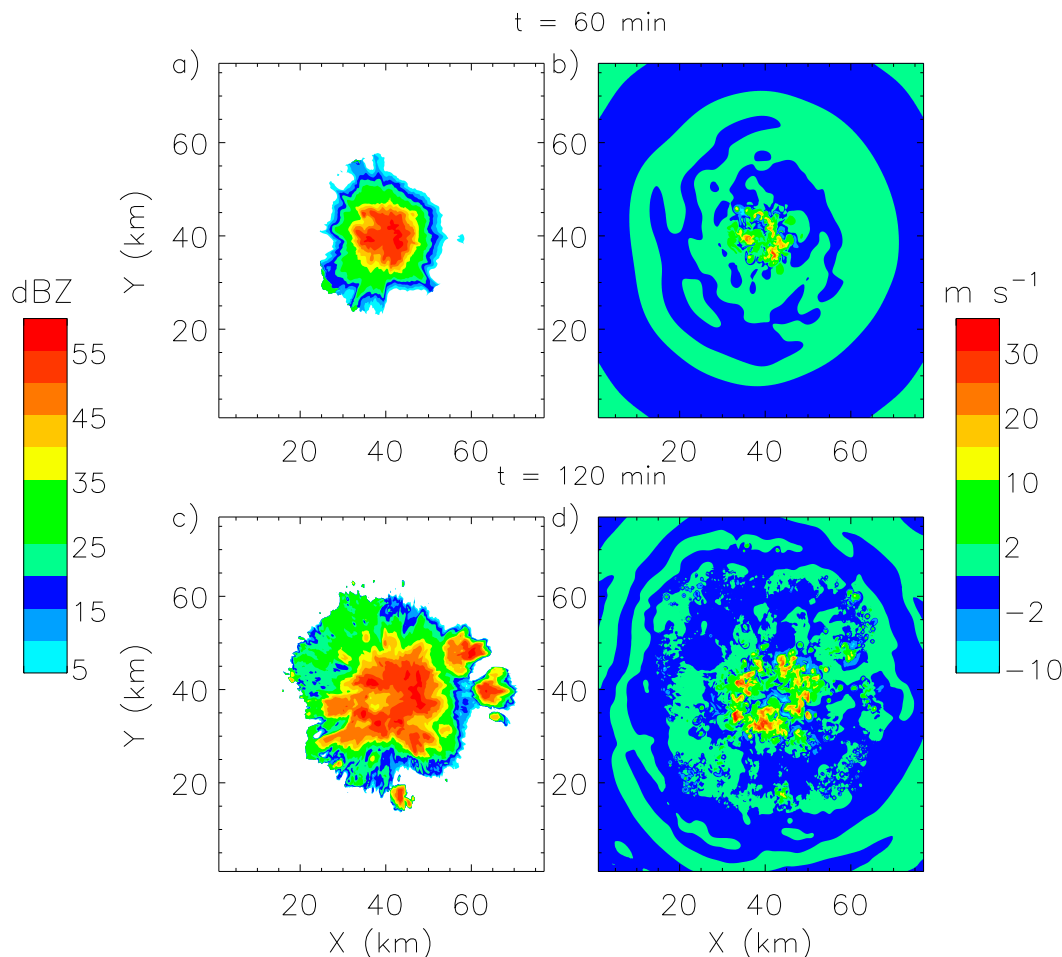


FIG. 2. Horizontal cross sections of the lowest (a),(c) model level radar reflectivity and (b),(d) vertical velocity at a height of 6.1 km, at (a),(b) 60 and (c),(d) 120 min using MORR microphysics, moderate CAPE sounding, and Δx of 0.125 km.

species), convective updraft mass flux (M_f), and fraction of the domain with convective updrafts (F_c) for MORR microphysics, the moderate CAPE sounding, and Δx of either 0.125 or 2 km. PRE, COND, and M_f are horizontal averages. The term M_f is calculated by summing the vertical velocity w times the air density for all grid points with $w \geq 2 \text{ m s}^{-1}$, and dividing by the total number of horizontal grid points in the domain. The term F_c is similarly calculated by summing the number of grid points with $w \geq 2 \text{ m s}^{-1}$ and dividing by the total number of horizontal grid points. Both M_f and F_c at a height of 6.1 km are shown. Note that this simplified definition of convective updraft may include unsaturated and/or negatively buoyant grid points (the latter associated with overshooting convection/gravity waves) that exceed the vertical velocity threshold. However, the particular choice for the threshold vertical velocity used to define convective updraft does not affect the conclusions of this work;

using a threshold between 1 and 5 m s^{-1} gives similar results in terms of differences between simulations.

Three primary phases of storm evolution are evident from Fig. 4. Over the first 25 min there is a rapid increase in M_f and F_c associated with convection driven by the initial thermal. There is a subsequent rapid decrease in M_f and F_c between 25 and 45 min as this initial convection decays. Since condensation is primarily driven by ascent in saturated conditions, COND shows a similar evolution. PRE also shows a similar evolution, but the peak is shifted to later times by 20–25 min. This reflects the time it takes for conversion of cloud to precipitation condensate and for precipitation hydrometeors to fall to the surface. A third phase is evident from 40 to 60 min until the end of the simulations with a slow, steady increase in M_f , F_c , COND, and PRE. This phase is associated with secondary convective initiation organized along the low-level outflow boundary, with sporadic

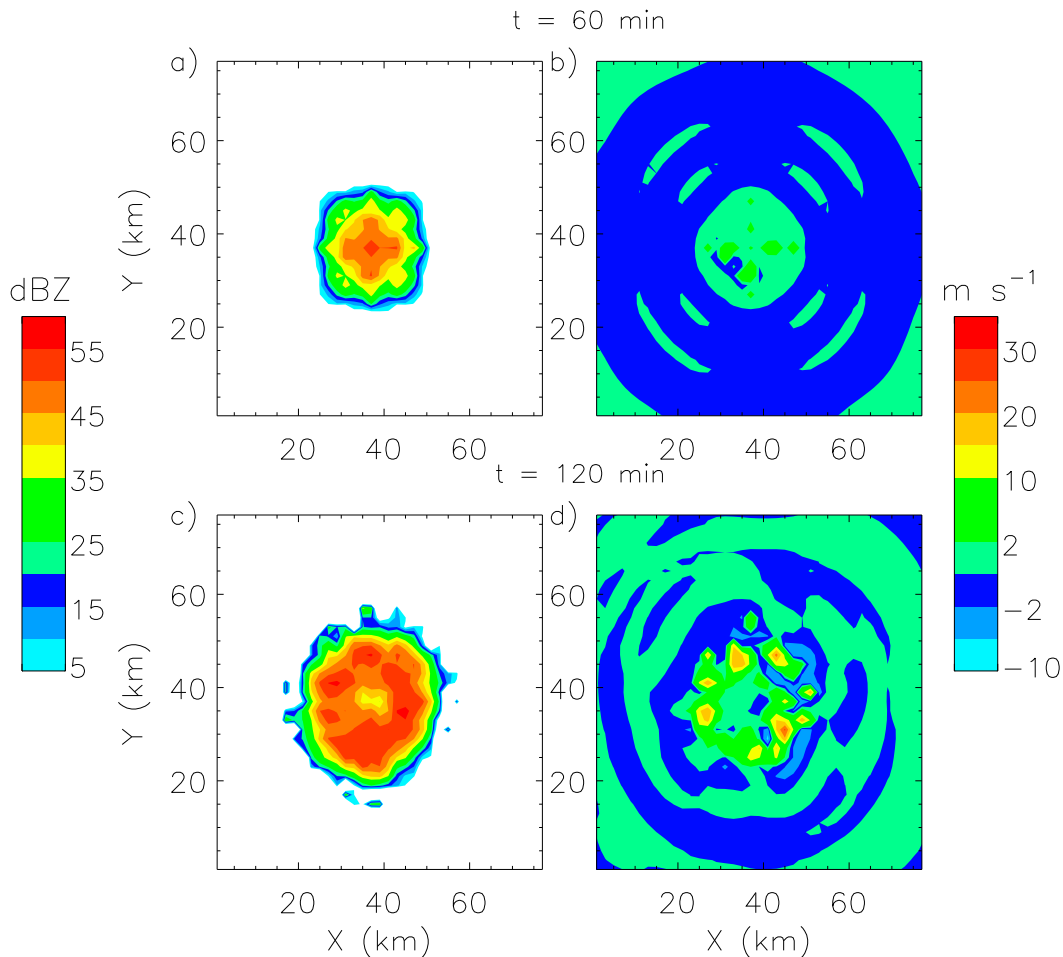


FIG. 3. As in Fig. 2, but for $\Delta x = 2$ km.

generation of new convective cells occurring until the end of the simulations. In general, the $\Delta x = 125$ -m simulations appear to evolve faster than the 2-km simulations, consistent with previous studies (Weisman et al. 1997; Petch et al. 2002; Bryan et al. 2003; BM12). Thus, M_f , F_c , COND, and PRE tend to be larger earlier in the simulations with Δx of 125 m, but larger in the 2-km simulations later. Thus, because the storms evolve differently, the temporally averaged (from 1 to 2 h) M_f , F_c , COND, and PRE do not show a consistent sensitivity to Δx . If the simulations were continued beyond 2 h in a larger domain (to ensure convection remained away from the lateral boundaries), it is possible that these quantities would be more sensitive to Δx . This should be kept in mind when interpreting the results presented below.

a. Surface precipitation

As stated in the introduction, a focus of this study is on changes in surface precipitation with microphysics, Δx , and thermodynamic sounding. The horizontally and

temporally averaged (from 60 to 120 min) surface precipitation rate, PRE, shows considerable scatter among the simulations (Fig. 5). A few main findings emerge from these results. First, while sensitivity to Δx is evident, with up to about a factor of 2 change across Δx for a given microphysics scheme and sounding, it is not consistent among simulations with different microphysics schemes or soundings. (Note there is some compensation of differences in PRE with Δx earlier and later in the simulations due to slower storm evolution at lower resolution as explained above.) Second, while there is sensitivity of PRE to CAPE for a given Δx and microphysics scheme (with differences again up to about a factor of 2), it is also not robust. For example, at small Δx (less than 500 m) there is a 20%–30% decrease for MORR but up to about a factor of 2 increase for WSM6 between the low and high CAPE simulations. One might anticipate an increase of PRE with CAPE, and this has been found in some observational analyses and is assumed in many convection parameterizations

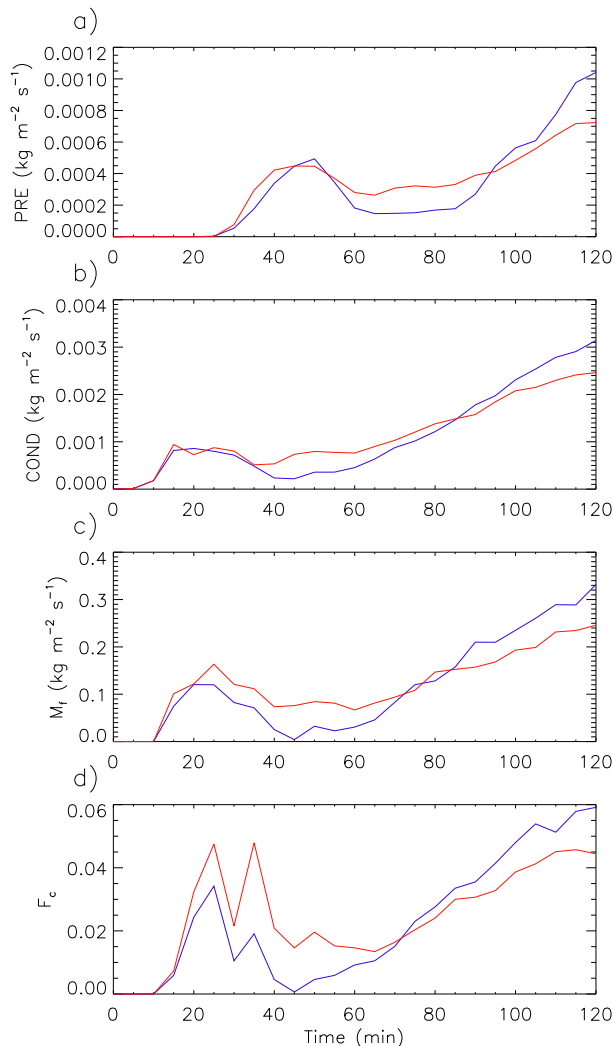


FIG. 4. Time series of horizontally averaged (a) surface precipitation rate (PRE), (b) vertically integrated total condensation (including conversion from vapor to condensate for all hydrometeor species, COND), (c) convective updraft mass flux M_f , and (d) fraction of the domain with convective updrafts, F_c , for MORR microphysics, the moderate CAPE sounding, and Δx of 2 (blue) or 0.125 km (red). Both M_f and F_c are shown at a height of 6.1 km.

that employ CAPE-based closures (e.g., Bechtold et al. 2004). The lack of a consistent relationship between PRE and CAPE here, despite stronger convection with higher CAPE, is explained by analysis of the condensed water budget in section 3b.

There is a more consistent sensitivity of PRE to microphysics scheme, with relative differences up to a factor of 3.5 for a given Δx and sounding. For the moderate and high CAPE soundings, MORR produces greater PRE than THO or WSM6 for all Δx , except the 2-km-high CAPE simulation in which all the schemes produce similar PRE. For the low CAPE sounding THO

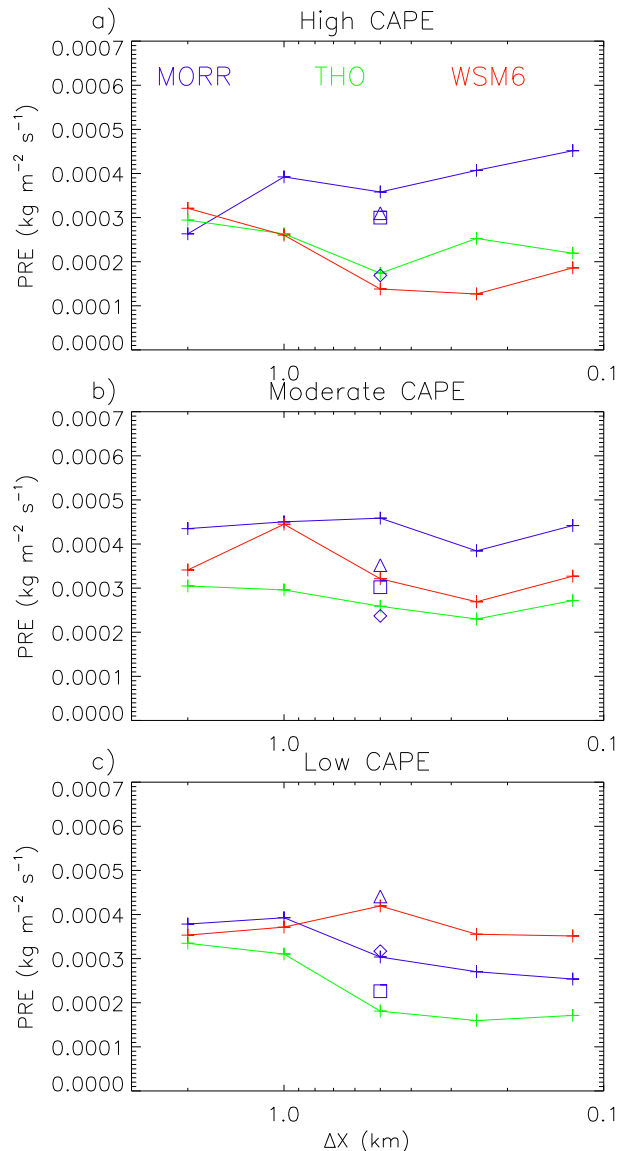


FIG. 5. Horizontally and temporally averaged (from 60 to 120 min) surface precipitation rate (PRE) as a function of Δx for the (a) high CAPE, (b) moderate CAPE, and (c) low CAPE soundings. Results for each microphysics scheme are indicated by the colored symbols and lines. Results are also shown for the sensitivity tests using MORR microphysics and $\Delta x = 500$ m but with no cold pool (blue triangles), graupel instead of hail for the rimed ice species (blue diamonds), and N_0 specified as a constant $8 \times 10^6 \text{ m}^{-4}$ (blue squares).

produces the smallest PRE for all Δx , while WSM6 produces the largest PRE for Δx of 0.5, 0.25, and 0.125 km. Thus, for a given sounding, sensitivity to microphysics is qualitatively similar for Δx between about 0.125 and 1 km, but with a reduced magnitude and in some instances change in the sign of differences among simulations for Δx of 1–2 km.

b. Bulk water budget and precipitation efficiency

To explain the sensitivities of horizontally and temporally averaged surface precipitation rate described above, we next analyze the bulk condensed water budget. The bulk condensed water budget is expressed as (e.g., Ferrier et al. 1996)

$$\text{PRE} = \langle \text{COND} \rangle - \langle \text{EVAP} \rangle - \text{RES}, \quad (1)$$

where all terms are horizontally and temporally averaged and $\langle \rangle$ indicates vertical integration from the surface to model top:

$$\langle P \rangle = \int_{\text{surface}}^{\text{top}} P \rho dz. \quad (2)$$

Here P indicates the relevant process (condensation or evaporation) and ρ is the air density. The first term on the right-hand side of (1), $\langle \text{COND} \rangle$, is the total condensation as defined earlier, but temporally averaged. $\langle \text{EVAP} \rangle$ is the evaporation and sublimation rate that includes contributions from all hydrometeor species. RES is a residual that is obtained from $\text{RES} = \langle \text{COND} \rangle - \langle \text{EVAP} \rangle - \text{PRE}$ (hereafter $\langle \rangle$ is dropped from relevant terms for convenience, unless otherwise noted). The residual primarily represents the difference in condensed water in the atmosphere between 60 and 120 min. RES includes a loss of condensate because of open lateral boundary conditions, but this is insignificant (generally less than 10% of RES).

COND for each simulation is shown in Fig. 6. Differences in COND with microphysics scheme, Δx , and sounding are similar to the differences in PRE. This suggests that COND is important for driving the sensitivity of PRE, and reflects the fact that COND is the source of condensed water for surface precipitation; this water must fall to the surface as precipitation, evaporate (or sublimate), or contribute to RES. Condensation of cloud droplets¹ dominates COND, with deposition onto ice hydrometeors accounting for only 10%–20% of the total. On the other hand, EVAP and RES seem to be less important in directly explaining sensitivity of PRE, especially for the sensitivity to microphysics scheme. Overall, simulations with larger EVAP do not produce smaller PRE (Fig. 7), as would be expected if differences in evaporation were the primary driver of changes in precipitation. Instead, greater COND in MORR for the moderate and high CAPE soundings and in WSM6 for

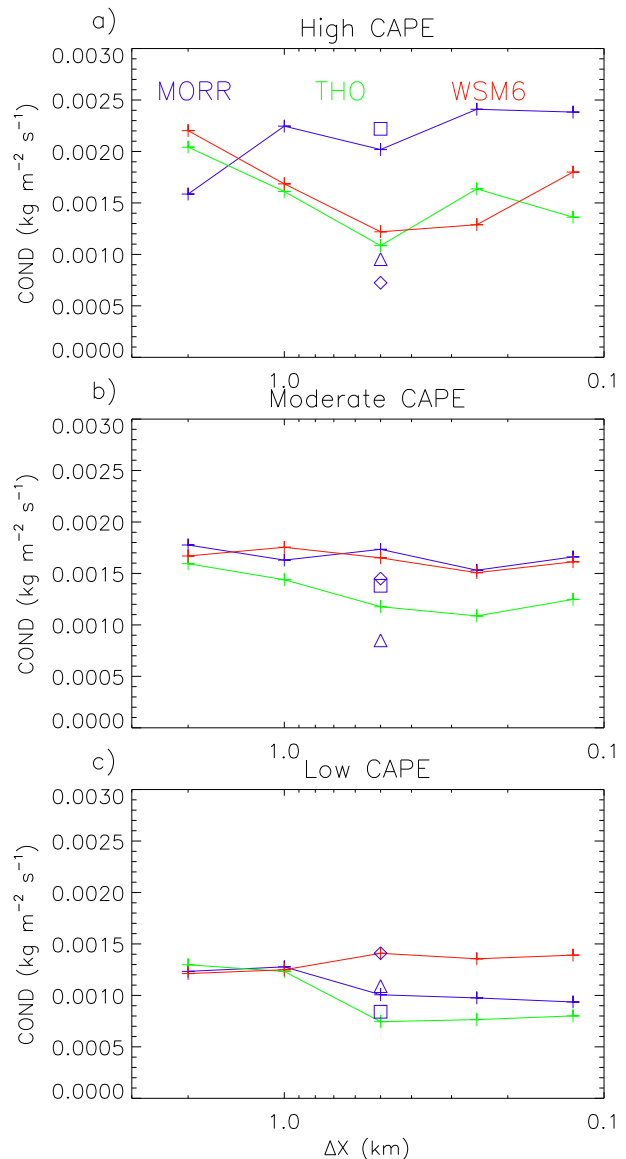


FIG. 6. As in Fig. 5, but for the horizontally and temporally averaged and vertically integrated total condensation rate, COND (including all hydrometeor categories).

the low CAPE sounding means there is more condensate available to evaporate. Thus, these simulations produce the largest EVAP but also the largest PRE. An exception is that WSM6 produces relatively large EVAP for the moderate and high CAPE soundings despite having fairly small COND. In other words, WSM6 produces a relatively large ratio EVAP/COND. For the moderate CAPE sounding, this explains why WSM6 and MORR have similar COND but MORR produces 25%–40% larger PRE. This is attributed primarily to enhanced rain evaporation in WSM6, which is partly explained by its use of a constant rain size distribution intercept parameter,

¹ Condensation of rain is neglected in MORR and THO, but included in WSM6. However, in WSM6 it has a limited contribution to total condensation relative to cloud droplets.

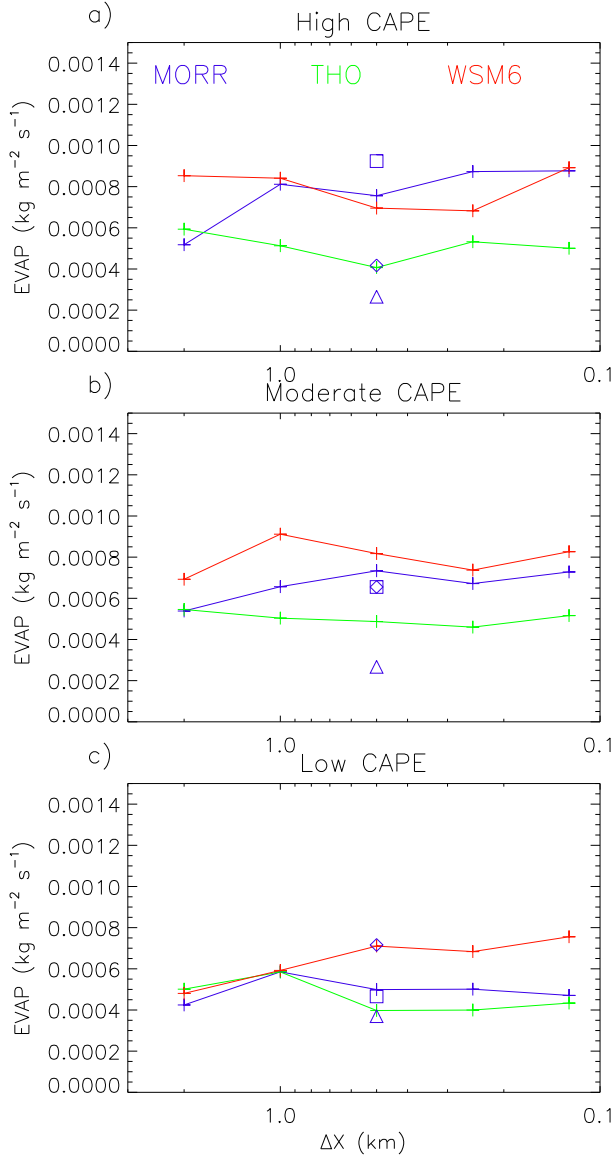


FIG. 7. As in Fig. 5, but for the horizontally averaged and temporally averaged and vertically integrated total evaporation rate, EVAP (including all hydrometeor categories).

N_0 , of $8 \times 10^6 \text{ m}^{-4}$ as described later in this subsection. Larger COND also means there is more condensate left in the atmosphere at the end of the simulations, and hence larger RES (Fig. 8). Since an increase in RES leads to a reduction of PRE in the condensed water budget as expressed by (1), differences in RES compensate somewhat for differences in COND in terms of the impact on PRE.

The lack of a consistent sensitivity of PRE to CAPE, despite stronger convection with higher CAPE, is explained by the condensed water budget. COND increases with CAPE (Fig. 6), especially for MORR

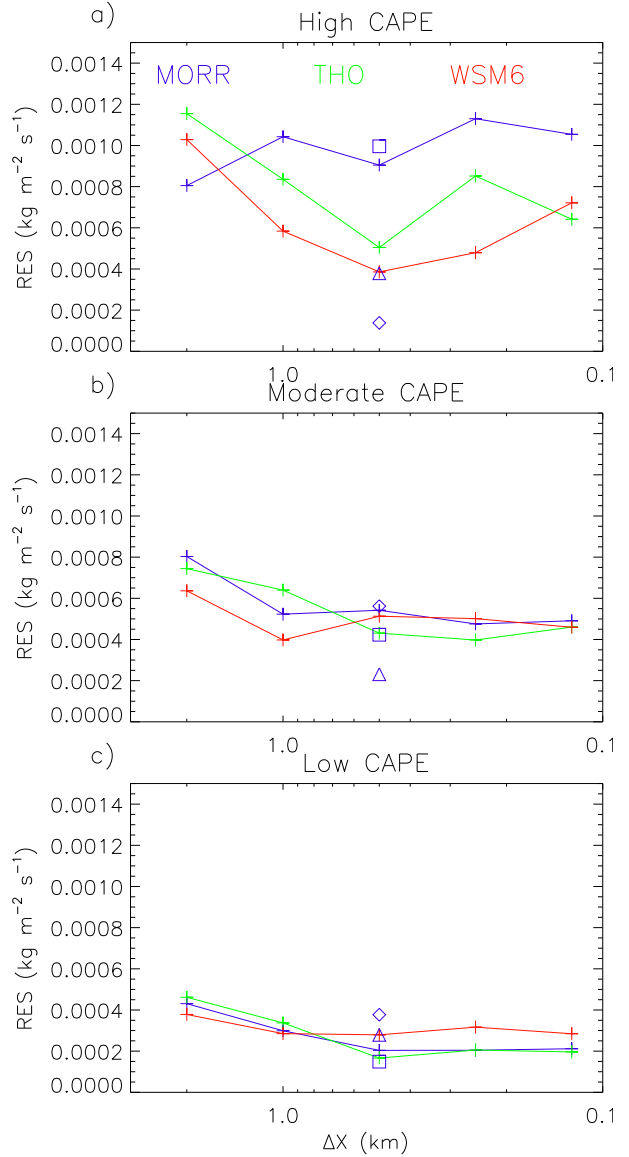


FIG. 8. As in Fig. 5, but for the residual term, RES, in the bulk condensed water budget [(1)].

and THO, which is expected given an increase in the updraft mass flux with greater conditional instability as discussed below. However, the increase of COND with higher CAPE is compensated by an increase in RES (Fig. 8), meaning that there is more condensate mass remaining in the atmosphere. There is also a small increase in EVAP with CAPE because of the increased amount of condensate available to evaporate. These differences in RES are consistent with horizontally and temporally averaged profiles of condensed water mass mixing ratios (Fig. 9). Higher CAPE leads to stronger and deeper updrafts that detrain substantial amounts of condensed water into the environment in

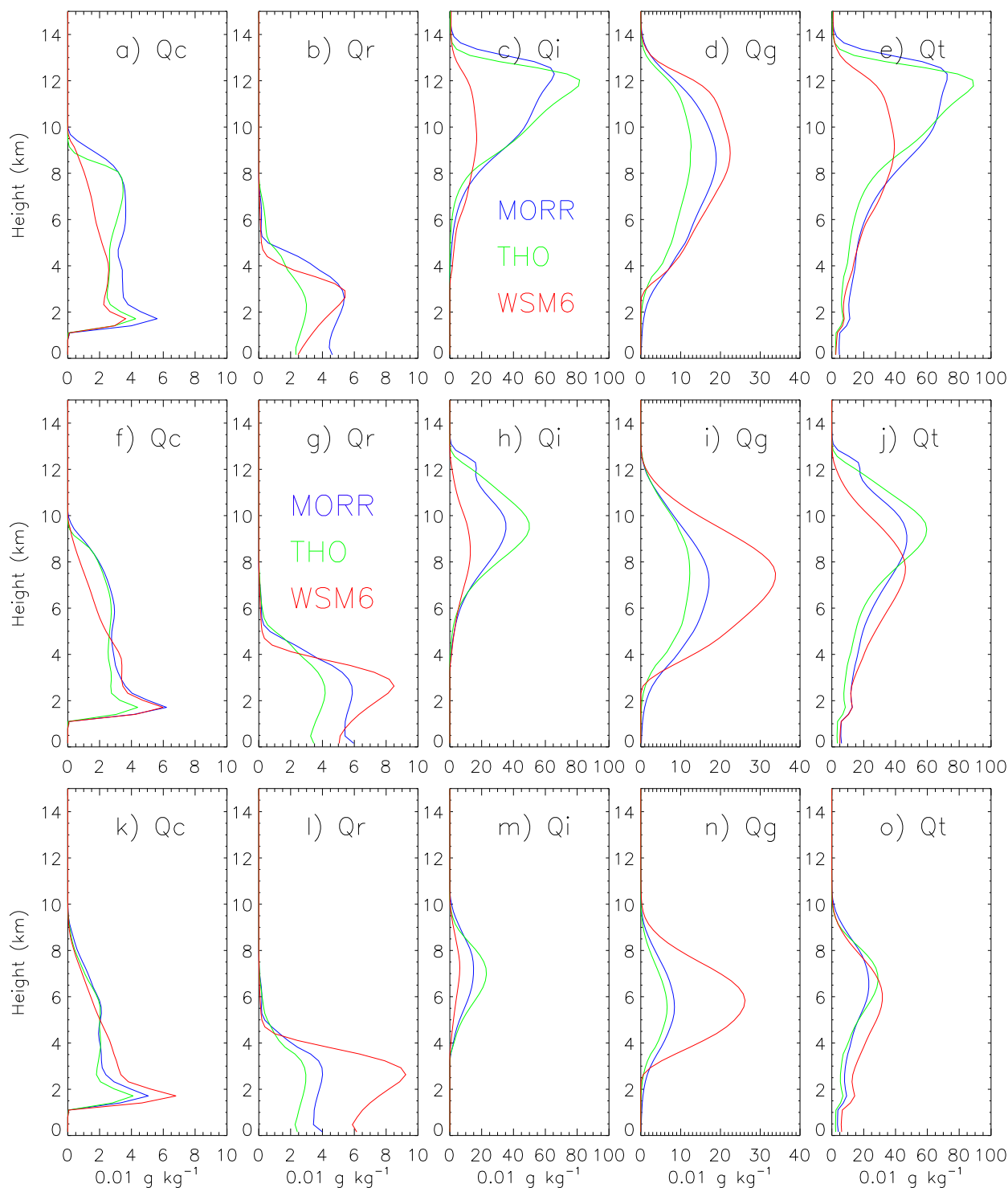


FIG. 9. Horizontally and temporally averaged profiles of hydrometeor mass mixing ratios for the $\Delta x = 500$ m simulations: cloud water (Q_c), rain (Q_r), cloud ice plus snow (Q_i), graupel/hail (Q_g), and the sum of all hydrometeors (Q_t) for the (a)–(e) high CAPE, (f)–(j) moderate CAPE, and (k)–(o) low CAPE soundings.

the upper troposphere and, hence, large cloud ice and snow mixing ratios aloft (Figs. 9c,h,m). This is a transient effect; eventually all condensed water aloft has to evaporate or reach the surface as precipitation. Nonetheless, the detrainment of substantial condensate in anvils with higher CAPE would generally be expected to reduce precipitation efficiency over the storm lifetime; much of this condensate would likely evaporate given its high altitude and the small terminal particle fall speeds of cloud ice and snow that dominate the anvil mass content. Differences in condensate profiles among the microphysics schemes for the various water species are generally consistent across the range of Δx , while sensitivity of condensate profiles to Δx is not consistent among the microphysics schemes. For example, there is roughly a factor of 2 increase in the amount of horizontally and temporally averaged graupel/hail in the MORR high CAPE simulation from Δx of 125 m to 2 km, but a small decrease in WSM6 (not shown).

These results can also be interpreted in the context of precipitation efficiency (PE), defined here in the usual way by the ratio of surface precipitation rate (PRE) to the total condensation (COND). As shown here PE is related to the condensed water budget by dividing (1) by COND, giving

$$PE = \frac{PRE}{COND} = 1 - \frac{EVAP}{COND} - \frac{RES}{COND}. \quad (3)$$

The lack of a consistent increase of surface precipitation with CAPE, despite the increase in COND, is reflected by a substantial decrease in PE with higher CAPE (Fig. 10). This decrease in PE is associated mainly with increased RES as discussed above, and compensates for the larger COND as CAPE increases. Interestingly, MORR produces the largest PE for all soundings and Δx . For the moderate and high CAPE soundings this occurs despite MORR producing large EVAP and RES, both of which reduce PE.

The relatively small PE in WSM6 compared to MORR is partly attributed to the representation of rain microphysics, and specifically the N_0 parameter, in the one-moment WSM6 scheme compared to the two-moment MORR scheme. This is supported by additional sensitivity tests using $\Delta x = 500$ m and MORR microphysics but with N_0 specified as a constant $8 \times 10^6 \text{ m}^{-4}$, following the one-moment treatment in WSM6. This leads to a decrease in PE for all soundings consistent with an increase in the ratio EVAP/COND (see Figs. 5, 7, and 10). This sensitivity of rain evaporation to N_0 has been documented in several previous papers (Morrison et al. 2009; Luo et al. 2010; BM12).

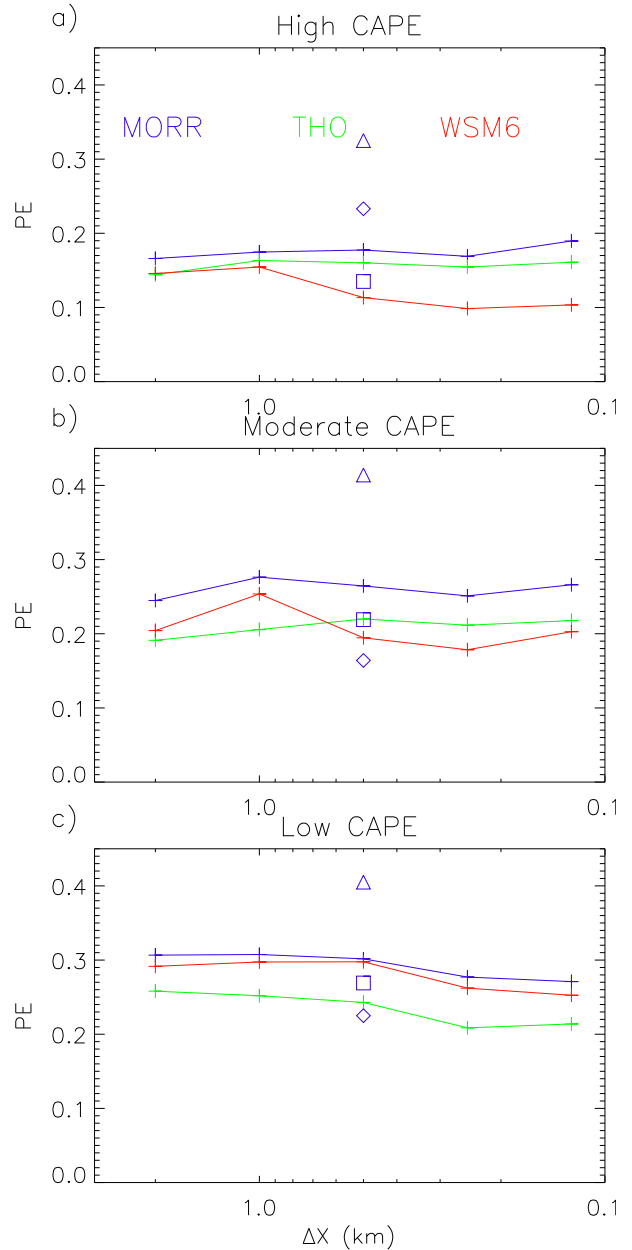


FIG. 10. As in Fig. 5, but for precipitation efficiency, PE.

c. Condensation rate and updraft mass flux

In the analysis of the condensed water budget above, it was shown that differences in COND are mainly responsible for driving differences in PRE among the microphysics schemes, while EVAP and RES were less important. Condensation in turn is primarily driven by upward motion in saturated convective updrafts. Thus, we next analyze the relationship between COND and dynamical quantities, specifically the convective updraft mass flux, to understand the dynamical influences driving the sensitivity of PRE.

Differences in the horizontally and temporally averaged, vertically integrated convective updraft mass flux, $\langle M_f \rangle$, with microphysics scheme generally correspond with differences in COND, as expected, and hence PRE (cf. Fig. 11 with Figs. 5 and 6). The large sensitivity of $\langle M_f \rangle$ to sounding is driven directly by differences in CAPE. Note, however, that the sensitivity of COND to CAPE is much smaller than the sensitivity of $\langle M_f \rangle$ to CAPE. This is because differences in the soundings directly drive differences in the convective mass flux only at mid- and upper levels, since the soundings are identical below 5 km (see section 2). There is decreasing availability of vapor for condensation with increasing height so that differences in updrafts above 5 km produce relatively small differences in COND. Similar to COND and PRE, sensitivity of $\langle M_f \rangle$ to Δx is evident but it is not consistent across the microphysics schemes.

What explains the sensitivity of $\langle M_f \rangle$ to microphysics, that in turn drives the sensitivity of COND and hence PRE? Differences in M_f at low levels, which are most critical for driving differences in COND, are consistent with differences in $\langle M_f \rangle$ although smaller in magnitude. Plots of temporally averaged convective fraction, F_c , and mass flux averaged within convective updrafts, M_c , at a height of 3.2 km (Figs. 12 and 13) further indicate that these differences in M_f with microphysics scheme are mainly due to differences in the spatial coverage of convective updrafts (F_c) rather than differences in the strength of the updrafts themselves (M_c). If anything, differences in M_c with microphysics scheme tend to oppose the differences in M_f , mainly for low and moderate CAPE, and, hence, partially compensate for the differences in F_c . MORR produces a much larger low-level F_c than THO or WSM6 for the high CAPE sounding, while WSM6 produces the largest F_c for the low CAPE sounding. These differences in F_c are in turn linked to microphysics through cold pools as explained next.

d. Cold pools and convective organization

In this subsection, we analyze linkages between microphysics, cold pools, and convective initiation that appear to drive differences in the fraction of the domain with convective updrafts, F_c , with the microphysics scheme and, hence, sensitivity of M_f , COND, and PRE through the interactions described above. Convective updrafts are organized primarily in a circular pattern around the outflow boundary corresponding to the cold pool edge after about 60 min (see Figs. 2 and 3). This suggests the importance of low-level convergence along the outflow boundary in generating new convective cells, as well as the cold pool circulation leading to horizontal transport and tilting of cells. An increase in cold pool size leads to a greater circumference around

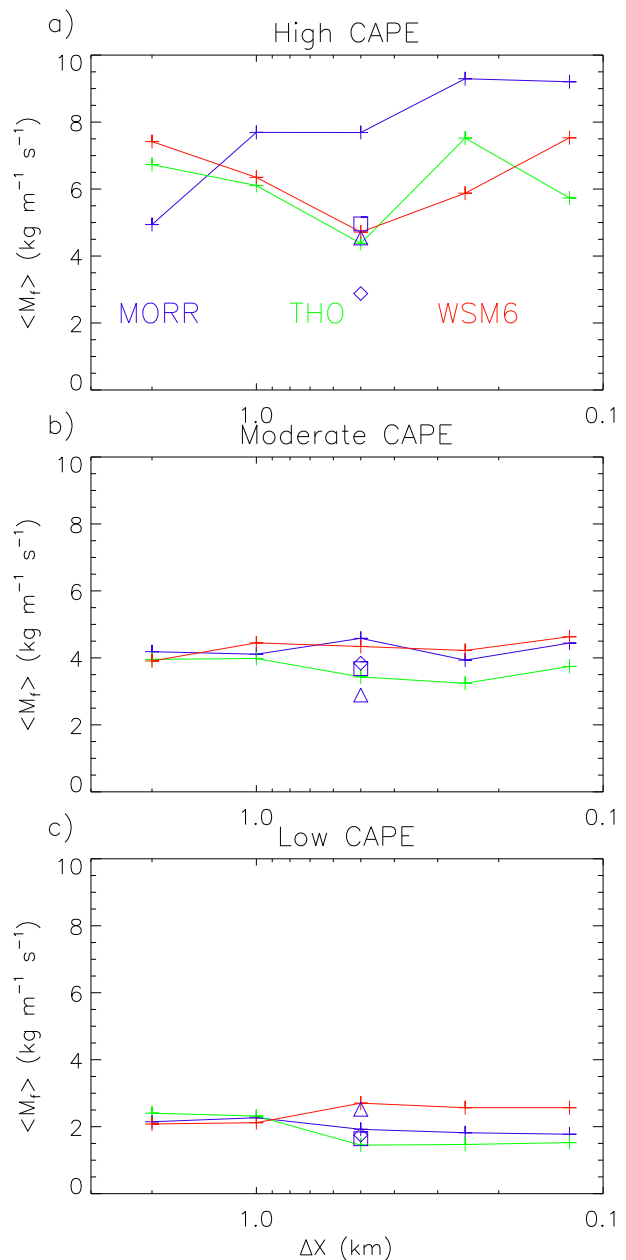


FIG. 11. As in Fig. 5, but for the horizontally and temporally averaged, vertically integrated convective updraft mass flux, $\langle M_f \rangle$.

the cold pool edge and, hence, a larger area for initiating secondary convection. This is reflected by differences in cold pool area among simulations with different microphysics schemes (Fig. 14) that generally, though not in all instances, correspond to differences in F_c (Fig. 12). Note that cold pool characteristics may play an even more important role in environments with greater convective inhibition, depending upon the strength and depth of the low-level environmental shear.

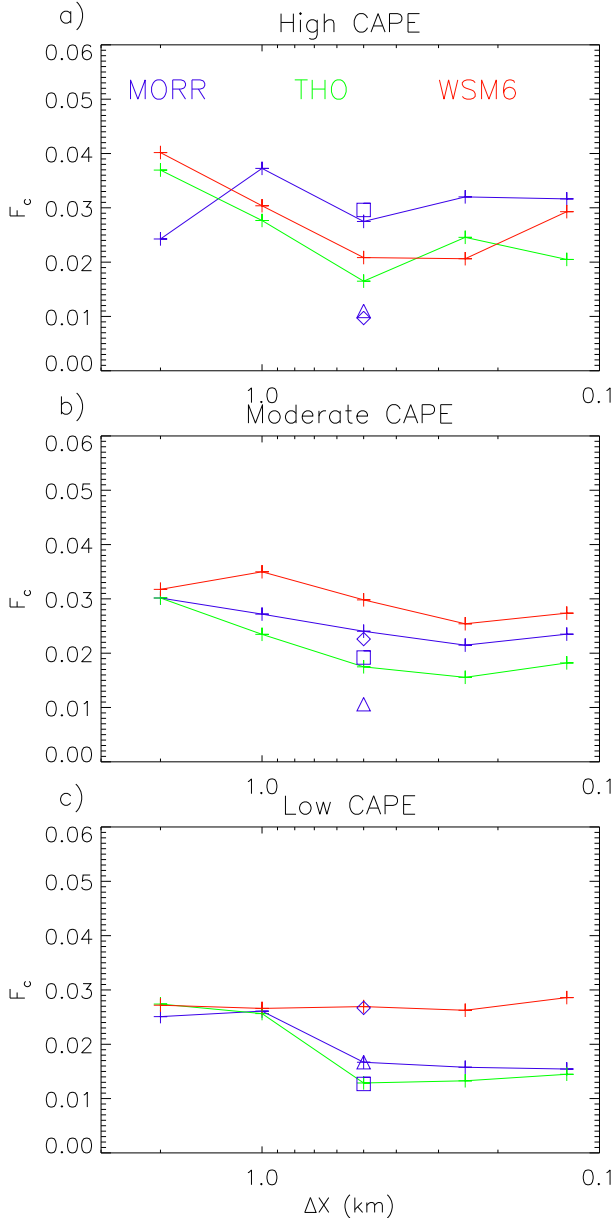


FIG. 12. As in Fig. 5, but for the fraction of the domain with convective updrafts, F_c , at a height of 3.2 km.

The importance of cold pools in initiating and organizing secondary convection is further demonstrated with additional sensitivity tests. We performed simulations using $\Delta x = 500$ m and MORR microphysics but with cooling due to evaporation, melting, and sublimation neglected so that the simulations do not generate a cold pool. Secondary convection is still initiated, but it is disorganized, and for the moderate and high CAPE soundings covers a smaller area than in the corresponding baseline simulations that include latent cooling (Fig. 15). We propose that in the absence of a

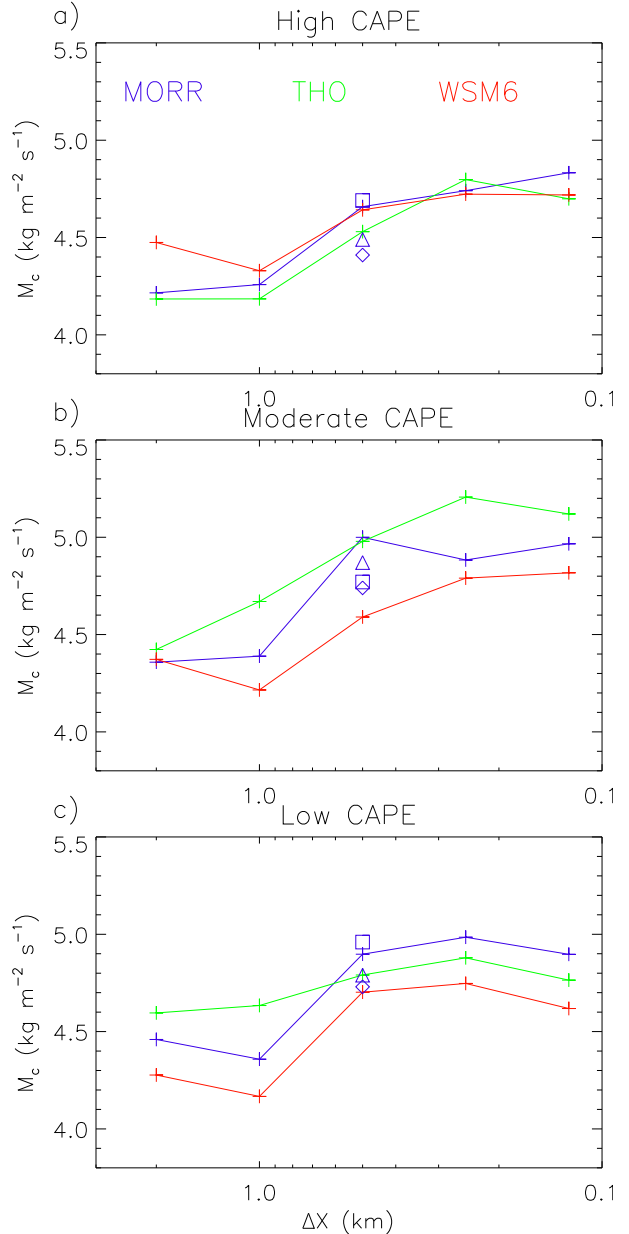


FIG. 13. As in Fig. 5, but for the mass flux averaged within convective updrafts, M_c , at a height of 3.2 km.

cold pool, gravity waves are still able to initiate secondary convection. The environment is moist with limited convective inhibition, and, hence, does not require much upward motion for parcels to reach the level of free convection. Figure 16 shows vertical cross sections of vertical velocity, the -2-K perturbation potential temperature isotherm, the boundary of the region with hydrometeor mixing ratio greater than 1 g kg^{-1} , and wind vectors, for the $\Delta x = 500$ m, high CAPE MORR simulations with and without cold pools. Figure 16a

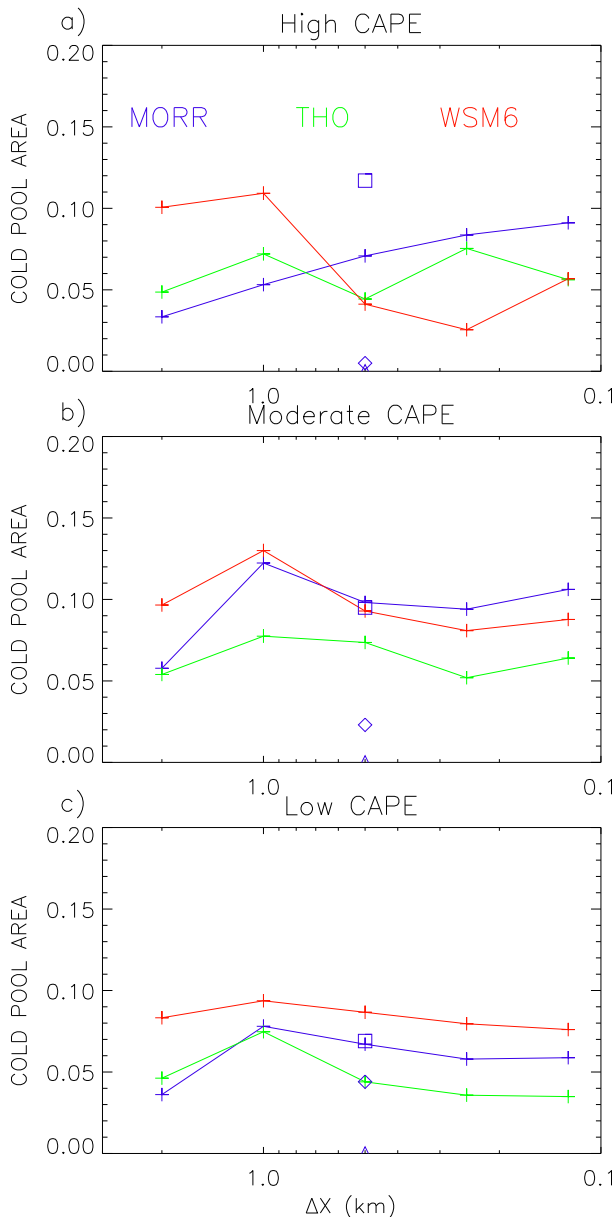


FIG. 14. As in Fig. 5, but for the cold pool area, defined as the fraction of the domain at the lowest model level with perturbation potential temperature (relative to the horizontal mean) less than -2 K.

clearly illustrates the role of the cold pool, downdraft, and low-level convergence in anchoring convection along the outflow boundary in the baseline run. In contrast, the role of downdrafts and low-level convergence appears to be minimal in the simulation without a cold pool, with the base of the updrafts anchored in wider regions of weak ascent (Fig. 16b) that encircle the storm, consistent with gravity wave features. Although they may not be a primary driver of convection in the baseline

simulations, gravity waves may still be important by modulating the cold pool circulation and driving weak ascent away from the cold pool edge. For example, the isolated convection that occurs outside of the main area of convection in the high CAPE baseline MORR simulation (see Figs. 2c,d) also appears to be unconnected to any well-defined low-level convergence feature, and we, therefore, propose that this convection is also likely initiated by gravity waves.

The F_c at a height of 3.2 km is similar among the low, moderate, and high CAPE simulations without cold pools (~ 0.1 – 0.15 , see Fig. 12). For the low CAPE sounding convective initiation in the absence of a cold pool, presumably by gravity waves, leads to a similar F_c as the baseline simulation with a cold pool. However, for the moderate and high CAPE soundings this mode of initiation is not able to compensate and F_c is smaller in the no cold pool runs than in the corresponding baseline simulations. These differences in F_c between the simulations with and without cold pools are in turn consistent with differences in $\langle M_f \rangle$, COND, and PRE through the mechanistic linkages described in sections 3a–c.

Another interesting effect of cold pools is their impact on precipitation efficiency, PE. The simulations without a cold pool have a much higher PE than the corresponding baseline MORR simulations with cold pools (Fig. 10). There are competing effects of cold pools and their ability to organize convection on PE. Previous studies have discussed how organization of convection can protect updrafts from entrainment by limiting the contact area between updrafts and their environment, and, hence, increase PE (e.g., Tobin et al. 2012). On the other hand, cold pools also affect the tilting of updrafts, horizontal transport of hydrometeors, and the location of secondary convective initiation relative to previous convection. It is this latter effect of cold pools on PE that appears to be dominant here. The cold pool circulation in the baseline simulations leads to more tilted updrafts, horizontal transport of hydrometeors away from convective cores, and initiation of secondary convection away from previous convection, all of which likely contribute to a decrease in PE relative to the simulations without a cold pool. The tilting of updrafts below about 4-km altitude in the baseline high CAPE MORR simulation is clearly seen in Fig. 16a, and contrasts with the nearly upright updraft in the corresponding simulation without a cold pool (Fig. 16b). Note that these results are expected to depend on the vertical wind shear of the environment (specified as zero here), which affects updraft tilting and structure through interaction with the cold pool as explained by “RKW” theory (Rotunno et al. 1988).

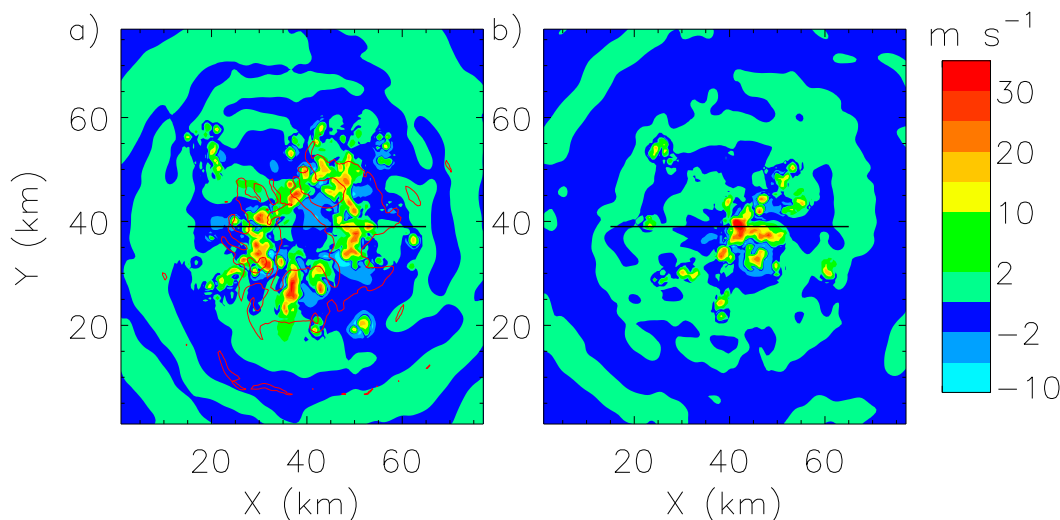


FIG. 15. Horizontal cross section of vertical velocity at a height of 6.1 km (color contours) and lowest model level perturbation potential temperature isotherm of -2 K relative to the horizontal mean (red contour lines) for 120 min and $\Delta x = 500$ m, MORR microphysics, and the high CAPE sounding: (a) baseline and (b) no cold pool simulation with latent cooling from evaporation, sublimation, and melting neglected. The black horizontal lines indicate locations of the vertical cross sections shown in Fig. 16.

That microphysics schemes produce large differences in cold pools is expected and has been well documented by previous studies (e.g., Morrison et al. 2009; Dawson et al. 2010; van Weverberg et al. 2012; BM12; Adams-Selin et al. 2013). To explore these differences further, additional sensitivity tests were performed to try and identify specific mechanisms driving cold pool differences

among the schemes. Differences in rain microphysics have been identified as an important driver of differences in cold pool characteristics (Gilmore and Wicker 1998; Morrison and Milbrandt 2011; BM12). The sensitivity tests using MORR but with a constant rain N_0 parameter following WSM6 (discussed in section 3b) show an increase in cold pool area for the high CAPE

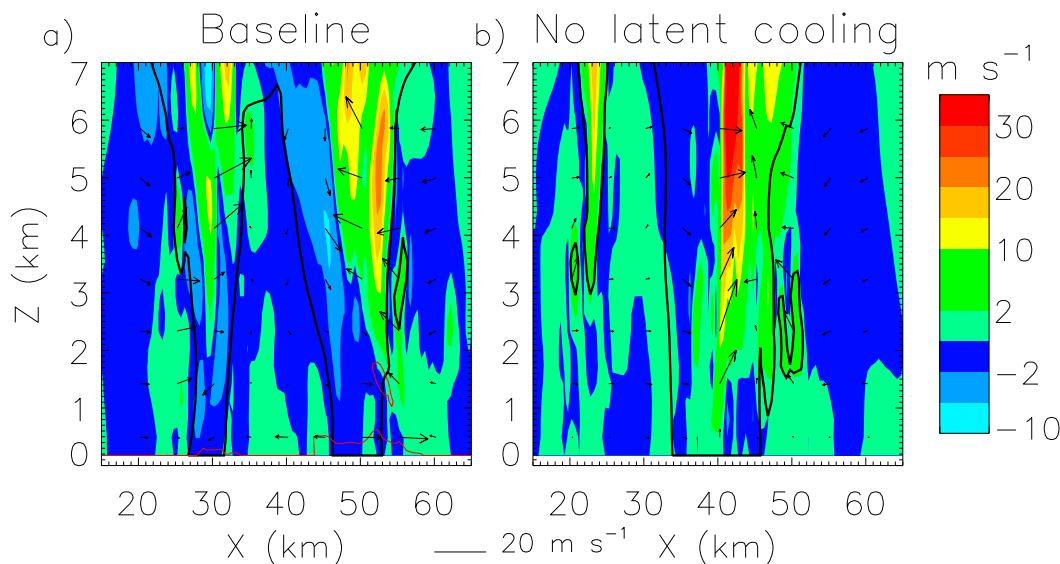


FIG. 16. Vertical cross sections of vertical velocity (color contours), hydrometeor mass mixing ratio of 1 g kg^{-1} (black contour lines), -2 K perturbation potential temperature isotherm relative to the horizontal mean (red contour lines), and 2D wind vectors for 120 min and the $\Delta x = 500$ m, MORR microphysics, and the high CAPE sounding: (a) baseline and (b) no cold pool simulation with latent cooling from evaporation, sublimation, and melting neglected. The locations of the vertical cross sections are indicated by the black lines in Fig. 15.

sounding, but little change for moderate and low CAPE. Thus, differences in representation of the rain N_0 alone do not appear to be a dominant driver of cold pool differences here, contrasting with results from some previous studies (Morrison et al. 2009; BM12).

There are also large differences in the treatment of rimed ice (graupel and hail) among the schemes, which previous studies have shown can be important in driving cold pool differences (Morrison and Milbrandt 2011; van Weverberg et al. 2012; BM12; Adams-Selin et al. 2013). We performed tests with $\Delta x = 500$ m and MORR, but with the bulk density and fall speed parameters of the rimed ice category specified to represent slower-falling, less dense graupel instead of hail. (Note that THO and WSM6 have more graupel-like characteristics for rimed ice; see the appendix.) In general, sensitivity to the specified rimed ice characteristics in MORR is larger for the moderate and high CAPE soundings compared to the low CAPE sounding, which is expected given the greater amounts of rimed ice in the higher CAPE simulations (see Figs. 9d,n). For the moderate and high CAPE soundings, assuming graupel characteristics leads to a reduced precipitation rate, less evaporation, and much weaker cold pools, compared to the corresponding baseline MORR simulations with hail (see Figs. 5, 7, and 14). Additional tests show that most of the impact is due to differences in fall speed between graupel and hail, rather than bulk density (keeping in mind that, like most schemes, bulk density and fall speed are not directly coupled in MORR). These changes are consistent with some differences between the baseline MORR (with hail) and THO or WSM6 schemes, for example a weaker cold pool in THO and WSM6 for the moderate and high CAPE soundings for $\Delta x = 500$ m. However, these differences are not consistent across the range of Δx ; THO and WSM6 produce a larger cold pool area than MORR for Δx of 1 and 2 km. This highlights the difficulty of isolating particular processes driving microphysical sensitivities, reflecting overall complexity of the schemes. For low CAPE, WSM6 produces a large ratio of graupel to cloud ice and snow mass compared to MORR and THO (Fig. 9n), which may help to explain the relatively strong, large cold pool for WSM6 for this sounding (Fig. 14c), in addition to the large rain evaporation rates in this simulation (see section 3b).

e. Convective characteristics

While there is not a robust sensitivity to Δx of the temporally averaged fraction of the domain with convective updrafts, F_c , or the horizontally averaged, vertically integrated convective updraft mass flux, $\langle M_{fj} \rangle$, sensitivities of other convective characteristics to Δx are more consistent across the microphysics schemes. As

expected, mean updraft diameter increases with Δx as convective overturning is forced to occur on larger scales, and the number of updraft cores in the domain decreases. This is seen by comparing Figs. 2b,d and 3b,d, and quantified by an analysis of mean updraft core diameter and number (calculated by finding the area of each horizontally contiguous region with vertically averaged $w \geq 2 \text{ m s}^{-1}$ and cloud water mixing ratio $\geq 0.1 \text{ g kg}^{-1}$, where the vertical averaging is between 5.5- and 7-km altitude, obtaining the mean area for all regions A and calculating an equivalent mean updraft diameter $D = \sqrt{4A/\pi}$). Note that reducing the threshold vertical velocity to 1 m s^{-1} leads to a slight increase in mean D but, otherwise, has little impact on results. Although there is some scatter among the microphysics schemes and soundings, especially for large Δx , in general D shows a consistent increase from about 2 km for $\Delta x = 125$ m, to 6 km or greater for $\Delta x = 2$ km (Fig. 17). The increase in D with larger Δx is compensated by a decrease in cell number N , explaining the lack of a consistent sensitivity of F_c to Δx .

There is some evidence for statistical convergence of D and N for Δx between 250 and 125 m. If we take the calculated D for each simulation at a given Δx as independent samples, then the p value using the Student's t test comparing the 125- and 250-m simulations is 0.1101, suggesting that we cannot reject the null hypothesis that the two datasets are from the same population at a 10% significance level. On the other hand, the p value comparing the 250- and 500-m simulations is 0.0002, suggesting with a very high degree of confidence that the two datasets are statistically different. A similar result holds for N . Note this analysis pertains to vertically coherent updraft structures, since vertical averaging was applied in defining the updraft core area. Averaging over a wider layer between 4.7- and 7.5-km altitude produces nearly the same mean D and N for a given Δx . However, if we use a single vertical level (6.1 km) instead of vertical averaging to define the updraft core area, then there is little evidence for convergence of D or N between Δx of 250 and 125 m since the p values are very small (less than 0.015). Thus, in this instance there is a statistically significant difference in D and N for Δx between 125 and 250 m. This occurs primarily because D is smaller (and N is larger) using a single vertical level instead of vertical averaging for $\Delta x = 125$ m, presumably because finer-scale horizontal structures are filtered by the vertical averaging. For $\Delta x = 250$ m and 500 m the two methods produce similar values of D and N .

Figures 13, 18, and 19 show the convective mass flux averaged within convective updrafts across the domain (i.e., the mean strength of convective updrafts), M_c , at

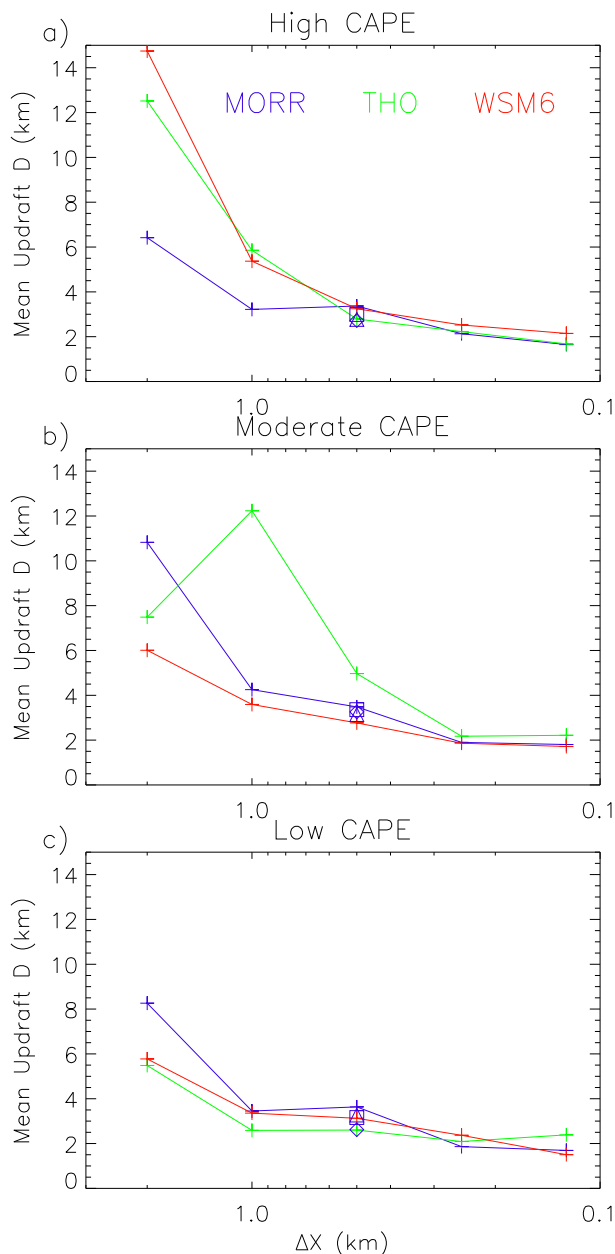


FIG. 17. As in Fig. 5, but for the mean updraft core diameter, D , defined from 5.5- to 7-km altitude at 120 min.

three different heights (3.2, 6.1, and 9.7 km) for each sounding. There is considerable sensitivity of M_c to Δx , and, unlike F_c , COND, and PRE, it is generally consistent across the microphysics schemes tested. At low levels (3.2 km), the largest M_c occurs at Δx of 125–250 m (Fig. 13). The term M_c is similar for all soundings at this height, consistent with the soundings being identical below 5 km. At mid- and upper levels there is an increase of M_c with CAPE as expected. Moving to higher altitude (6.1 km), the largest M_c occurs at Δx of 250–500 m

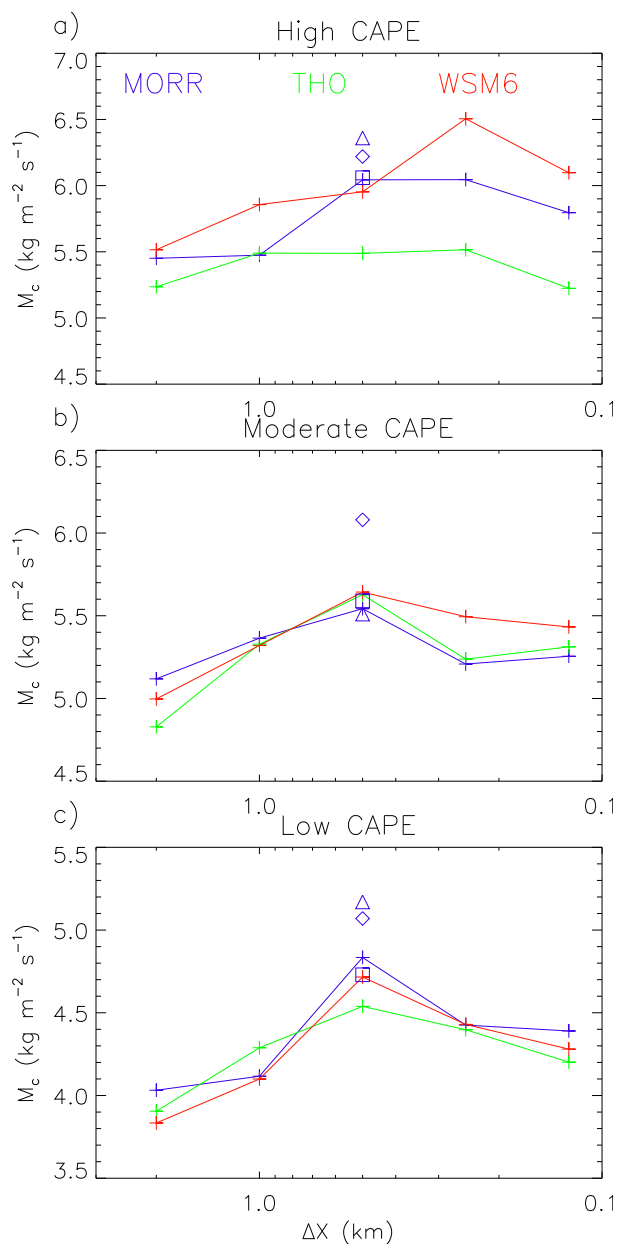


FIG. 18. As in Fig. 5, but the mass flux averaged within convective updrafts, M_c , at a height of 6.1 km. Note the different y-axis scales in the plots.

(Fig. 18). At upper levels (9.7 km), the largest M_c occurs at Δx of 500–2000 m for the high and moderate CAPE soundings (Fig. 19), although there is more scatter with the different microphysics schemes than at lower altitudes. There is also an overall decrease of M_c at this level compared to 6.1 km for all soundings, which cannot be explained by the decrease in air density alone and is, therefore, likely due to entrainment. Note this is above the level of neutral buoyancy for the low CAPE sounding and, hence, M_c likely represents overshooting

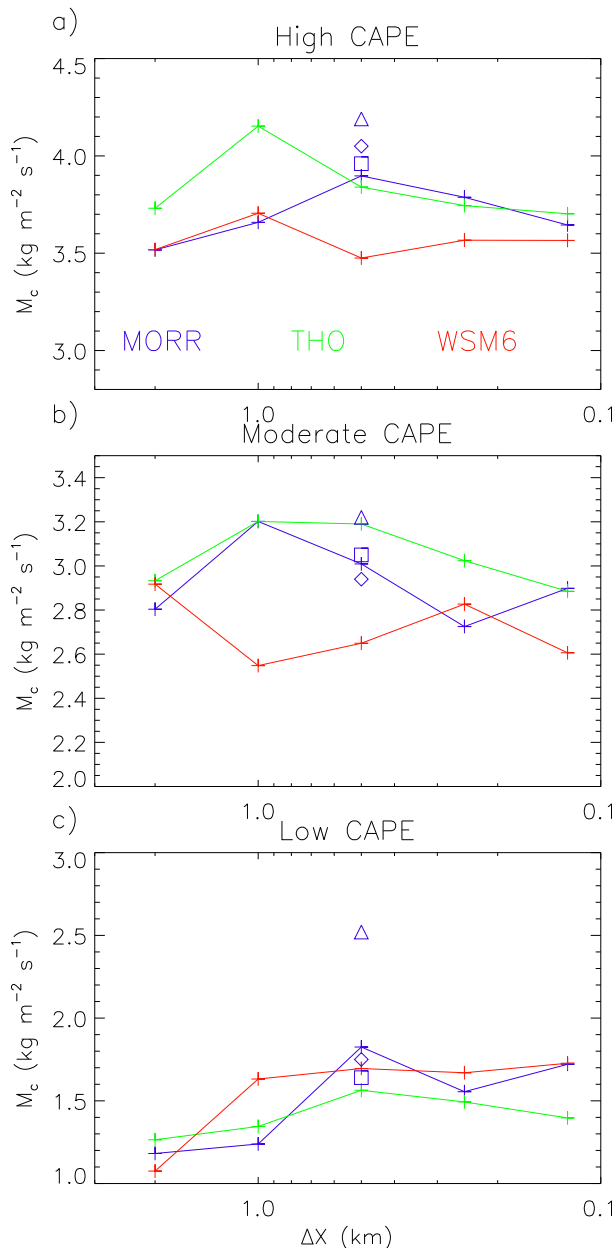


FIG. 19. As in Fig. 5, but the mass flux averaged within convective updrafts, M_c , at a height of 9.7 km. Note the different y-axis scales in the plots.

convection/gravity waves for this case. In contrast to mid- and upper-level M_c , the domain-maximum vertical velocity monotonically increases with decreasing Δx (not shown). This is expected since the model is able to better resolve smaller-scale fluctuations of vertical velocity at progressively higher resolution.

These results are generally consistent with previous studies, although there are differences with regard to particular measures of convective strength. BM12 found

that several measures of storm strength, including the temporally averaged, vertically integrated updraft mass flux and condensation rate, were at a maximum for Δx of 1 km. However, M_f and COND do not exhibit clear sensitivities to Δx here, while M_c was not analyzed in BM12. Nonetheless, the largest values of mid- and upper-level M_c occurring at intermediate Δx is consistent with the hypothesis put forth in BM12, explained briefly as follows. For large Δx , updrafts are wide and, hence, adverse perturbation pressure effects dominate, leading to relatively weak updrafts (Weisman et al. 1997; Bryan et al. 2003; Markowski and Richardson 2010). On the other hand, for small Δx , updrafts are narrow and turbulent eddies and entrainment are better resolved (as seen by vertical velocity fields that appear turbulent, cf. Fig. 2), which should lead to greater dilution and, hence, weakening of updrafts (BM12). The combination of these effects is expected to be smallest at intermediate Δx . Testing of this hypothesis by detailed analysis of pressure perturbation and entrainment effects is left for future work.

4. Summary and conclusions

In this study, we investigated the sensitivity of idealized deep convection simulations to parameterization of microphysics, horizontal grid spacing (Δx), and environmental static stability. Three different microphysics schemes with varying degree of complexity were tested: WSM6, THO, and MORR. The initial sounding was modified by altering the static stability above 5 km, with CAPE varying from 1534 to 4229 J kg^{-1} .

Overall, the horizontally and temporally averaged surface precipitation rate (PRE), total condensation and evaporation, and convective mass flux were sensitive to microphysics and Δx . Sensitivity to microphysics was qualitatively similar for Δx between 0.125 and 1 km, broadly consistent with the results of BM12. However, there was a tendency for a reduced magnitude of sensitivity and in some instances changes in the sign of differences among simulations for Δx of 1–2 km. These results suggest that sensitivity to microphysics can be tested at model resolutions somewhat coarser than eddy-resolving scales, although additional work is needed to assess the generality of this finding.

While a decrease in Δx led to faster storm development and differences in the evolution of PRE, COND, EVAP, and convective mass flux consistent with previous studies (e.g., Weisman et al. 1997; Petch et al. 2002), these differences were reduced when averaged over time. Moreover, sensitivities of temporally averaged PRE, COND, and EVAP to Δx were not robust, since they varied with different microphysics schemes.

The lack of a consistent sensitivity of these quantities to Δx , despite large changes in the small-scale dynamics and convective updraft properties (mean width, strength, and number) as Δx was modified, is consistent with Langhans et al. (2012), who investigated bulk convergence of moist convection simulations over complex terrain. However, because the storms evolved differently as Δx was modified, it is possible that the temporally averaged PRE, COND, and EVAP would be more sensitive to Δx if the simulations were extended beyond 2 h using a larger domain; additional work is needed to test this hypothesis.

These results are in contrast to BM12, who found a sharp increase of the cloud water evaporation and a reduction of temporally averaged surface precipitation rate as Δx was decreased from 1 to 0.25 km regardless of microphysical settings. In their study this was attributed to enhanced entrainment along the lateral cloud edges with more numerous and narrower updrafts, leading to decreased precipitation efficiency. Specific reasons for our contrasting results are unclear, but may be due to our use of a fairly moist sounding (Weisman and Klemp 1982) that likely reduced the role of cloud water evaporation from entrainment, and, hence, may have reduced the sensitivity of precipitation rate to Δx . It is hypothesized that using a drier sounding more typical of situations with steep tropospheric lapse rates may lead to greater sensitivity of the precipitation rate to Δx as entrainment and evaporation are enhanced at higher resolutions. The impact of tropospheric humidity on sensitivities of entrainment, evaporation, and precipitation to Δx provides motivation for further study. It would also be useful to examine these sensitivities in the context of thermodynamic soundings more typical of the moist tropics, with high humidity and low convective inhibition that favors initiation of secondary convection, but with a less steep free-tropospheric lapse rate compared to the soundings used here. The unsheared environment also differed from the moderate low-level vertical wind shear in BM12, leading to different modes of convective organization that may help to explain these differences (BM12 simulated a squall line with trailing stratiform precipitation). Exploring the role of environmental shear and its impact on convective organization on sensitivities to microphysics and Δx should also be a subject of further investigation.

An analysis of the bulk condensed water budget showed that sensitivity of horizontally and temporally averaged surface precipitation rate to microphysics scheme was mostly driven by differences in total condensation resulting from differences in the domain-average low-level convective updraft mass flux. Differences in convective mass flux were in turn associated with

differences in the spatial coverage of convective updrafts. The linkage to microphysics was primarily attributed to the effect on cold pools, which strongly affected the organization and initiation of secondary convection and, hence, the coverage of convective updrafts. A larger cold pool area was typically associated with more secondary convection, a larger convective fraction, greater mean updraft mass flux, more condensation, and more precipitation.

Although the mechanistic linkages described above were evident for all soundings, the sensitivity to microphysics itself was strongly dependent upon the sounding. For moderate and high CAPE, the representation of rimed ice as either fast-falling hail or slower-falling graupel had a large impact on simulations. Simulations with MORR using graupel had a larger precipitation rate, reduced latent cooling via evaporation and melting, and a weaker, smaller cold pool. On the other hand, there was less sensitivity to the representation of graupel or hail for the low CAPE sounding since there was less production of rimed ice given the relatively weak updrafts. The importance of CAPE to the sensitivity of treating graupel versus hail is consistent with van Weverberg (2013). We also note that sensitivity to rimed ice fall speed may be reduced for longer simulations, since lower fall speed graupel may be expected to eventually reach the surface (van Weverberg 2013).

While sensitivity of the temporally averaged fraction of the domain with convective updrafts and horizontally averaged, vertically integrated convective mass flux to Δx was not robust and varied using different microphysics schemes, other statistical convective characteristics (mean updraft diameter, strength, and number) showed a robust sensitivity to Δx that was more consistent across the schemes. Results suggested statistical convergence of mean diameter and number for Δx between 250 and 125 m for vertically coherent updraft core structures. However, there was no evidence for statistical convergence of mean updraft diameter and number when considering only a single vertical level, which likely reflects the inability to resolve smaller-scale turbulent motions even at $\Delta x = 125$ m. The impact of not resolving these finer-scale features is unclear, but should be investigated using simulations with Δx of tens of meters.

This study focused on sensitivities to Δx , representation of microphysics, and environmental thermodynamic conditions for *idealized* simulations of moist deep convection. This allowed us to investigate these sensitivities in a simplified framework, so that specific mechanisms and process interactions driving these sensitivities could be clearly identified. Isolating processes driving sensitivities can be very difficult for “real case” simulations with initial and boundary conditions from observations, and

full model physics coupled with dynamics. However, real case studies are needed to ascertain the realism of simulations, and to determine how the accuracy of simulations relative to observations is affected by changing Δx and the representation of microphysics for specific events. Future work should, therefore, extend these kinds of sensitivity studies to real cases.

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APPENDIX

Overview of the Microphysics Schemes

The WSM6 scheme (Hong and Lim 2006) is based on the work of Lin et al. (1983) with changes to ice-phase behavior by Hong et al. (2004) and improved mixed-phase particle fall speeds by Dudhia et al. (2008). Prognostic variables calculated in WSM6 include mass mixing ratio for five classes of hydrometeors (cloud water, rain, cloud ice, snow, and graupel). Rain and graupel particles are assumed to follow an inverse exponential size distribution [Eq. (1) in Hong and Lim (2006)] with intercept values of $8 \times 10^6 \text{ m}^{-4}$ and $4 \times 10^6 \text{ m}^{-4}$, respectively.

The Thompson scheme (Thompson et al. 2008) includes five species of hydrometeors (cloud water, rain, cloud ice, snow, and graupel). It prognoses number and mass mixing ratios for cloud ice and rain, and mass mixing ratios only for the other species. Rimed ice is represented by a hybrid graupel–hail category with a two-parameter diagnostic dependence of its size distribution intercept parameter based on the mass mixing ratio and amount of supercooled liquid water coincident in a grid volume. Each hydrometeor species, except for

snow, is assumed to follow a gamma size distribution [Eq. (A1) in appendix A of Thompson et al. (2008)]. The size distribution of snow is represented by a sum of exponential and gamma distributions [Eq. (1) in Thompson et al. (2008)], with parameters depending on snow water content and temperature. Snow is also assumed to be nonspherical with a bulk density varying inversely with diameter. Most other bulk microphysical parameterizations describe snow as spherical with constant density, including WSM6 and MORR.

The Morrison scheme (Morrison et al. 2005, 2009) includes cloud water, rain, cloud ice, snow, and graupel/hail. It has a user-specified switch to include either graupel or hail for rimed ice species. This study uses hail, except as noted, which was found by BM12 to produce a more realistic reflectivity and thermodynamic structure compared to graupel for a midlatitude squall-line case study. Prognostic quantities include number and mass mixing ratios for all species except cloud water, for which only mass mixing ratio is prognosed. The size distribution for these particles is represented by a gamma distribution [Eq. (1) in Morrison et al. (2009)]. For rain, snow, hail, and cloud ice, the shape parameter is set to zero (inverse exponential distributions), but it is a function of the specified number concentration for cloud water following the observations of Martin et al. (1994).

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