



1 **Benefits of a 4th Ice Class in the Simulated Radar Reflectivities of Convective Systems using**
2 **a Bulk Microphysics Scheme**

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47 ABSTRACT

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Current cloud microphysical schemes used in cloud and mesoscale models range from simple one-moment to multi-moment, multi-class to explicit bin schemes. This study details the benefits of adding a 4th ice class (frozen drops/hail) to an already improved single-moment 3-class ice (cloud ice, snow, graupel) bulk microphysics scheme developed for the Goddard Cumulus Ensemble model. Besides the addition and modification of several hail processes from a bulk 3-class hail scheme, further modifications were made to the 3-ice processes, including allowing greater ice supersaturation and mitigating spurious evaporation/sublimation in the saturation adjustment scheme, allowing graupel/hail to transition to snow via vapor growth and hail to transition to graupel via riming, wet graupel to become hail and the inclusion of a rain evaporation correction and vapor diffusivity factor. The improved 3-ice snow/graupel size-mapping schemes were adjusted to be more stable at higher mixing ratios and to increase the aggregation effect for snow. A snow density mapping was also added.

The new scheme was applied to an intense continental squall line and a moderate, loosely-organized continental case using three different hail intercepts. Peak simulated reflectivities agree well with radar for both the intense and moderate case and were superior to earlier 3-ice versions when using a moderate and large intercept for hail, respectively. Simulated reflectivity distributions versus height were also improved versus radar in both cases compared to earlier 3-ice versions. The bin-based rain evaporation correction affected the squall line more but overall agreement in reflectivity distributions was unchanged. The new scheme also improved the simulated surface rain rate histograms.

71 1. Introduction

72

73 Atmospheric cloud modeling has benefited immensely from the continued improvement in
74 computational power. Simulations using explicit spectral bin microphysics (SBM) with large 3D
75 domains in mesoscale models like WRF (the Weather Research and Forecasting model,
76 Michalakes et al. 2004; Skamarock et al. 2008) can now be performed (Iguchi et al. 2012a, b).
77 In addition to higher resolution (e.g., Khairoutdinov and Randall 2006) and the advent of MMFs
78 (multi-scale modeling frameworks, Randall et al. 2003, Tao et al. 2009), cloud-resolving
79 simulations have also benefited in the form of ever more sophisticated microphysics. Simple
80 bulk liquid (e.g., Kessler 1969) and ice schemes (e.g., Wisner et al. 1972) with only a few
81 categories have grown into multiple ice categories (e.g., Straka and Mansell 2005), two moments
82 (e.g., Ferrier 1994; Reisner et al. 1998; Morrison et al. 2009) and higher (Milbrandt and Yau
83 2005b), and highly detailed SBM (Ovtchinnikov and Kogan 2000; Khain et al. 1999; 2000;
84 2004). Detailed bin forms originated a while ago but are only now becoming practical, having
85 previously been limited to either 1D (Young 1974; Scott and Hobbs 1977), 2D (Takahashi 1976;
86 Hall 1980; Reisin et al. 1996; Khain and Sednev 1996) or without ice (Kogan 1991).

87 Though the ability to use SBM with ever increasing 3D domains is becoming a reality, these
88 types of simulations are still not common, and there is a trade-off versus bulk microphysics
89 schemes (BMSs), which assume hydrometeor distributions follow a prescribed form (typically
90 exponential or gamma). BMSs are much faster and require a lot less memory. For certain
91 applications (e.g., very large domains or long-term simulations), computational resources are
92 often not sufficient for SBM, which themselves are still not perfect (Li et al. 2010); these
93 resources could also be applied to better resolution, which is another important consideration
94 when it comes to realistically simulating convective entrainment and overturning (Bryan et al.

95 2003). BMSs are typically invoked using either one-moment (1M, only mass is predicted) or
96 two-moments (2M, both mass and number concentration are predicted); these schemes have also
97 seen numerous advancements and improvements in recent years. Numerous modeling studies
98 and BMSs were made or based on the 1M 3-class ice (3ICE) schemes of Lin et al. (1983) and
99 Rutledge and Hobbs (1983, 1984) developed in the early 1980's. These schemes were the
100 workhorses of cloud microphysics for many years and are still used in some form by many
101 schemes today, especially for NWP. However, they do have their biases (Lang et al. 2007, 2011;
102 hereafter L2007, L2011) and are susceptible to thresholding phenomena (Rutledge and Hobbs
103 1984). They use *a priori* hydrometeor classes of cloud ice, snow and either graupel or hail and
104 transfer hydrometeors from one class to another conditional upon specified thresholds; this can
105 result in abrupt and unnatural behavior and diverging solutions depending on if conditions are
106 met. An innovative approach was recently developed by Morrison and Grabowski (2008) based
107 on the concepts of Heymsfield (1982) and Hashino and Tripoli (2007) whereby the amounts of
108 mass acquired by riming and deposition are predicted separately. This allows for the history of
109 the riming fraction to be accounted for and results in a spectrum of particle densities with
110 smooth, natural transitions from ice to snow and snow to graupel. Lin and Colle (2011) included
111 the effects of partially rimed particles using a diagnostic riming intensity as well as functional
112 forms of the mass-, area- and velocity-diameter relationships. In general, the representation of
113 cloud microphysical processes is constantly improving as more and more schemes are including
114 aerosols and the representation of ice processes continues to improve (Muhlbauer et al. 2013).

115 Two-moment BMSs have become increasingly popular (Ziegler et al. 1985; Ferrier 1994;
116 Reisner et al. 1998; Meyers et al. 1997; Milbrandt and Yau 2005a; Morrison et al. 2005, 2009;
117 Seifert and Beheng 2006; Thompson et al. 2008; Lim and Hong 2010) and offer a good

118 compromise between the extreme cost of SBM and the restriction of 1M BMSs. They allow an
119 extra degree of freedom in defining the hydrometeor distributions compared to 1M schemes and
120 can account for size-sorting (Milbrandt and Yau 2005a) and aggregation as well as aerosol
121 effects (Lim and Hong 2010). Two-moment schemes are also superior to 1M in terms of rain
122 evaporation. A fixed rain intercept in 1M schemes tends to produce excessive rain evaporation,
123 namely in the stratiform region; this can be alleviated with 2Ms, which can lead to better
124 convective and stratiform rainfall structures (Morrison et al. 2009).

125 However, 2M schemes as well as SBM require observed cloud condensation nuclei (CCN)
126 and/or ice nuclei (IN) profiles to activate cloud water and/or ice particles, and despite their
127 potential (Ferrier et al. 1995; Milbrandt and Yau 2005a; Morrison et al. 2009), there are enough
128 uncertainties and nonlinearities that more advanced/sophisticated schemes do not always perform
129 better than simpler 1M bulk schemes (Wang et al. 2009; Varble et al. 2011). Besides the extra
130 degrees of freedom that are required to behave in a realistic manner, larger errors in a more
131 dominant process can overwhelm potential gains elsewhere. WRF contains a variety of
132 microphysics packages, including 1M, 2M and schemes that are a mixture of both, easily
133 allowing comparisons between the various schemes. Van Weverberg et al. (2013) used the
134 Advanced Research WRF to evaluate MCS simulations using three different BMSs over the
135 tropical western Pacific and found that although different, the results from the 2M schemes were
136 not superior to those using 1M; they found the most crucial element was the fall speeds of frozen
137 particles. Using WRF, Han et al. (2013) found that the 2M Thompson scheme (Thompson et al.
138 2008) had the best radar reflectivities for a winter storm but the fall velocities from the Goddard
139 scheme (Tao and Simpson 1993; Tao et al. 2003; L2007) agreed best with vertical profiler
140 observations. Powell et al. (2012) used vertically-pointing millimeter radar observations to

141 evaluate WRF simulations of MCS anvils near Niamey, Niger; they found the 1M Goddard and
142 WDM6 (Lim and Hong 2010, which has 1M ice and 2M liquid) schemes actually produced more
143 realistic anvils than did the 2M ice schemes. Systematic biases in cloud-resolving model (CRM)
144 BMSs were first identified when CRM simulations were used for satellite retrievals. Radiative
145 transfer models applied to CRM-simulated cloud fields revealed distributions that contained
146 excessive scattering signatures that were not representative of actual observed distributions
147 (Panegrossi et al. 1998; Bauer 2001; Olson et al. 2006). In addition to the excessive scattering,
148 several studies found excessively high reflectivities in the upper troposphere of CRM simulations
149 with excessive amounts and/or sizes of precipitation-sized ice produced by the BMSs as the
150 primary reason (L2007, L2011; Blossey et al. 2007; Zhou et al. 2007; Li et al. 2008; Matsui et al.
151 2009). Typically, the problem is associated with graupel (L2007; Li et al. 2008; Matsui et al.
152 2009). Though somewhat confirmed by Varble et al. (2011), they noted that excessively high
153 reflectivities can also be due to snow in a 2M scheme (Morrison et al. 2009) and that reasonable
154 reflectivities can be obtained using nonspherical graupel particles with variable density and
155 mass-diameter relationships with 1M. At any rate, there are enough inherent biases and room for
156 improvement that microphysics schemes in general require and continue to undergo refinement.

157 This study is a follow on to L2007 and L2011 and details the continued improvement and
158 enhancement of the Goddard 1M BMS used in the Goddard Cumulus Ensemble model (GCE), a
159 version of which (L2007) is one of the microphysics packages available in WRF. This scheme
160 has been evaluated in WRF and despite some continued biases (L2007, L2011) found to compare
161 quite well relative to other WRF schemes with regard to snow fallspeeds (Han et al. 2013) and
162 MCS anvils. The lineage is built upon the Rutledge and Hobbs (1983, 1984) 3ICE-graupel
163 version. Despite the improvements already made to the scheme (L2007, L2011), it still contains

164 some unrealistic aspects with regard to reflectivity structure and, without hail, lacks the ability to
165 simulate more intense radar echoes. Hail allows for the simulation of much more intense radar
166 echoes and much higher fall velocities than graupel and can thus to first order expand the scheme
167 to cover a far wider range of conditions than would be possible with say improving the graupel
168 category to 2M. This study details the addition of a 4th ice category, which encompasses the
169 spectrum of particles from smaller frozen drops to larger hail stones that have a high density
170 ($\sim 0.9 \text{ g cm}^{-3}$) as a result of being or having a coating at or near liquid at some point in their
171 history, as well as further refinements to the scheme that result in an improved 4-class ice (4ICE)
172 version of the Goddard 1M BMS. Despite the benefits of a smooth transition in particle
173 characteristics (e.g., density) with a single prognostic rimed-ice category, higher and lower-
174 density particles cannot coexist in such a scheme without a separate hail category, which may be
175 necessary to simulate for example a narrow hail shaft (Milbrandt and Morrison 2013). The new
176 Goddard 4ICE scheme is tested for two cases, an intense midlatitude squall line observed during
177 MC3E and a more moderate convective case from TRMM LBA (the Tropical Rainfall
178 Measuring Mission Large-Scale Biosphere-Atmosphere Experiment in Amazonia). The model
179 results are evaluated using radar reflectivity contoured frequency with altitude diagrams
180 (CFADs, Yuter and Houze 1995). Validation via comparison with *in situ* aircraft data can
181 provide a very detailed look at the performance of microphysical schemes (e.g., Molthan and
182 Colle 2012); however, such data are limited and difficult to compare against (if even available)
183 when it comes to convective cores and are essentially unavailable when it comes to intense
184 convective cores. Another approach has been to compare modeled versus observed radiances,
185 often radar reflectivity, statistically in the form of CFADs, the primary approach adopted in
186 L2007 and L2011. CFADs, which are essentially PDFs sampled at discrete levels through the

187 depth of a storm stacked in the vertical and then contoured, were first used primarily to
188 characterize observations (e.g., Yuter and Houze 1995). Lin (1999) and Lang et al. (2003)
189 constructed CFADs of model data, but Smedsmo et al. (2005), Eitzen and Xu (2005) and Braun
190 (2006) were the first to use radar CFADs to actually evaluate CRM performance. These were
191 quickly followed by several other studies (Blossey et al. 2007; L2007; Rogers et al. 2007; Zhou
192 et al. 2007; Li et al. 2008), establishing this method (or a close variation thereof) as a standard
193 way of evaluating CRM-type simulations (L2011; Varble et al. 2011; Iguchi et al. 2012a; Powell
194 et al. 2012; Guy et al. 2013; Han et al. 2013; Van Weverberg et al. 2013; Wu et al. 2013).
195 Though there are other ways to evaluate the performance of CRMs and their microphysics
196 schemes, having to match the observed radar reflectivity distributions is a more stringent test
197 than mean quantities (Powell et al. 2012). Radar observations are more readily available than *in*
198 *situ* observations, especially when it comes to convection, and the resolution of the data is
199 comparable to CRM grids and much better than satellite observations.

200 The main objectives of this study are to allow the improved 1M Goddard BMS (L2007; L2011)
201 to simulate intense convection via the addition of hail and to further improve upon the model's
202 overall performance via the reduction of biases in the synthetic radar structure and reflectivity
203 distributions. The paper is organized as follows. Section 2 describes the Goddard CRM, the
204 changes to the Goddard microphysics, the two case studies, and the numerical experiments.
205 Section 3 presents the simulation results and their validation using radar observations as well as
206 surface rain intensities. The summary and conclusions are given in section 4.

207

208 **2. Simulation setup and cases**

209

210 *a. The Goddard Cumulus Ensemble model (GCE)*

211

212 The new Goddard 1M 4ICE BMS is evaluated using the 3D GCE. The GCE was described in
213 Tao and Simpson (1993) and Tao *et al.* (2003) and more recently in Tao *et al.* (2013). The
214 model configuration closely follows that used in L2011. The GCE has a one-and-a-half order
215 sub-grid scale turbulence scheme (Soong and Ogura 1980), parameterizations for shortwave
216 (Chou and Suarez 1999), longwave (Chou and Kouvaris 1991; Chou *et al.* 1995, 1999; Kratz *et al.*
217 1998) and cloud optical properties (Sui *et al.* 1998; Fu and Liou 1993), positive definite
218 advection (Smolarkiewicz 1983, 1984; Smolarkiewicz and Grabowski 1990), and options for
219 anelastic (Ogura and Phillips 1962) or compressible flow (Klemp and Wilhelmson 1978). The
220 GCE has several microphysics options, but the primary BMS is the Rutledge and Hobbs (1983,
221 1984)-based 3ICE-graupel scheme (i.e., cloud water, rainwater, cloud ice, snow and graupel),
222 which has been improved to reduce unrealistically large amounts of graupel (L2007) and 40 dBZ
223 echoes above the freezing level (L2011) and modified to introduce ice nuclei concentrations into
224 the Bergeron parameterization (Zeng *et al.* 2008, 2009).

225

226 *b. Addition of hail processes and other microphysics improvements*

227

228 Prior improvements reduced excessive amounts of graupel (L2007) and excessive penetrations
229 of 40-dBZ echoes above the freezing level (L2011), alleviating some of the biases in the original
230 Goddard 1M 3ICE-graupel scheme. The improved versions, however, have two artifacts in their
231 simulated reflectivity structure. Time-height cross sections of peak reflectivities show a band of
232 elevated reflectivity maxima above the freezing level separated by corresponding local minima

233 just above the freezing level (e.g., Figs. 2c-d, 6d) as opposed to observations, which typically
234 show a steady monotonic decrease with height (e.g., Figs. 2a, 6a). The snow/graupel size
235 mapping implemented in L2011 tried to compensate for this and still produce reasonable peak
236 reflectivities by rapidly increasing the particle sizes (especially snow) at moderate to high mixing
237 ratios (Figs. 1c-d). This can lead to spurious artifacts in the reflectivity distributions near the
238 melting layer for stronger cases (Fig. 4d). To address these issues and allow the scheme the
239 ability to simulate more intense convection, a 4th ice class (frozen drops/hail) was added and the
240 scheme further refined to produce an improved Goddard 4ICE scheme.

241 Table 1 lists the changes in relation to previous versions of the scheme, which are detailed
242 below. Hail processes based on Lin et al. (1983) were added with some modification. Hail
243 riming, accretion of rain, deposition/sublimation, melting, shedding and wet growth processes
244 were left unchanged. Analogous to L2007 for graupel, the collection of other ice particles under
245 dry growth conditions (dry collection) was eliminated for hail to prevent an excessive buildup of
246 hail as collection efficiencies should be minimal; but, hail near wet growth conditions is
247 expected to be close to water coated and thus efficiently collect other ice particles. Hail within
248 95% of wet growth is thus allowed to collect other ice particles. Graupel is medium density
249 (~ 0.3 to 0.5 g cm^{-3}) and mainly a riming product; frozen drops, however, are high-density ($\sim 0.9 \text{ g}$
250 cm^{-3}) like that of solid ice or hail. Therefore, the new scheme differentiates hail initiation from
251 graupel by treating the product of any process that freezes rain as hail. Five new hail processes
252 were also added: wet hail accretion of graupel, hail rime splintering, wet graupel conversion to
253 hail, hail conversion to snow due to depositional growth (also added for graupel) to allow
254 hail/graupel particles that undergo significant deposition at cold temperatures in the absence of
255 liquid water to transition to snow, and hail conversion to graupel via riming, which tries to

256 account for lower hail bulk densities due to rime accumulation under non wet growth conditions.
257 Milbrandt and Morrison (2013) demonstrated how graupel densities can sharply decrease at
258 colder temperatures aloft when accounting for variable rime density. A similar analogy is
259 applied here with regard to the hail category, which too can be rimed. The latter two processes
260 are “apparent” and try to overcome some of the limitations associated with having fixed particle
261 categories by providing additional pathways for particles to change from one category to another.

262 In addition to the hail processes, further modifications were made to the improved 3ICE
263 processes. Snow autoconversion was strengthened by adjusting the timescale, threshold and
264 efficiency in the Kessler-type formulation. The original formulation is quite weak, but aircraft
265 observations of small ice particle distributions suggest that although diffusion dominates at
266 colder temperatures, there is evidence of aggregation (Field 1999). Though autoconversion was
267 strengthened, diffusion still remains the dominant process. The maximum snow collection of
268 cloud ice efficiency was increased to reflect the fact that ice particles with diameters greater than
269 about 200 microns are efficient collectors. A water vapor diffusivity correction factor (Byers
270 1965) was applied to all related processes, effectively replacing a more simplified function of
271 pressure and temperature. The formulation for the depositional growth of cloud ice into snow
272 (i.e., Psfi) invokes the time step over a time scale, so that tiny amounts of snow can form even
273 when cloud ice is quite small (i.e., the time scale to become snow is large). Therefore, an
274 arbitrary small threshold was introduced before activating this term¹. With ice supersaturations
275 commonly observed on the order of tens of percent (Jensen et al. 2001; Stith et al. 2002; Garrett
276 et al. 2005), the sequential saturation adjustment scheme (Tao et al. 2003) was further relaxed
277 from the 10% constraint in L2011 to a maximum of 20%. The deleterious effects of cloud

¹ A value of $1.e-5$ g/g scaled by the surface air density over the level air density was used to inhibit this artificial snow production mainly because it has a noticeable effect on longer-term (i.e., MMF) simulations.

278 boundaries advecting through an Eulerian grid (e.g., spurious evaporation) have been previously
279 noted [Klaassen and Clark 1985; Grabowski 1989; Grabowski and Smolarkiewicz 1990; Stevens
280 et al. 1996; Grabowski and Morrison 2008; Reisner (personal communication)]. To reduce these
281 effects, the saturation adjustment was further modified to restrict cloud evaporation and
282 sublimation to subsidence areas. The time-scale for ice sublimation can be appreciable, allowing
283 even smaller ice particles to exist in subsaturated conditions (e.g., Garrett et al. 2005); so, the
284 saturation adjustment was relaxed to allow cloud ice to persist under ice subsaturated conditions.

285 With the addition of hail, the size-mapping scheme for snow and graupel introduced in L2011
286 was adjusted. The rate at which the characteristic size (i.e., inverse of the slope parameter)
287 increases with mixing ratio, especially of snow, was lowered (see Fig. 1). As will be shown,
288 peak reflectivities are now determined largely by hail and do not require large graupel or snow
289 particles to generate these higher values. In addition, the aggregation effect, especially for snow,
290 was increased to allow particle sizes to grow more rapidly as temperatures rise from -25 C, and
291 an associated snow density mapping (Fig. 1g) was introduced as a function of snow size (i.e.,
292 Brandes et al. 2007). Graupel density was also divided into low and moderate based on a simple
293 mixing ratio threshold. Finally, to address the problem of excessive rain evaporation due to a
294 fixed rain intercept, a rain evaporation correction was adopted based on the results from an
295 explicit bin microphysics scheme (Li et al. 2009). This correction was made “physical” by
296 lowering the rain intercept (i.e., increasing the mean raindrop diameter) locally for each rain grid
297 cell until the new evaporation rate matched the correction factor. The scaling was capped such
298 that rain size never increases with decreasing rain mixing ratio.

299 For a more detailed description of the new scheme, including the new hail budget equation,
300 rain evaporation correction and snow mapping, please see the Appendix.

301

302 *c. Convective case studies*

303

304 1) AN INTENSE MIDLATITUDE SQUALL LINE: MC3E

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306 The Midlatitude Continental Convective Clouds Experiment (MC3E) was conducted in and
307 around central Oklahoma (OK) from April to May 2011 as a collaborative effort between the
308 DOE (Department of Energy) ARM (Atmospheric Radiation Measurement) and NASA GPM
309 (Global Precipitation Measurement) programs. The 20 May 2011 case featured a deep, upper-
310 level low over the central Great Basin moving into the central and southern Rockies before
311 lifting into the central and northern Plains. At the surface, low pressure in southeastern Colorado
312 drew warm, moist air up through the southern Plains to a warm front oriented E-W across
313 Kansas, while a dry line extended from the Texas/Oklahoma Panhandle down through the
314 Concho Valley. The result was a series of convective lines that formed over the Great Plains and
315 propagated eastward toward the Mississippi Valley. The most intense squall line to pass through
316 the MC3E sounding network was the result of convection that had developed over south-central
317 Kansas (KS) and north-central OK within the network merging with the northern end of a long
318 convective line that had formed along the dry line and extended from southwestern OK down to
319 the Big Bend. The northern portion of this longer line entered the MC3E sounding network
320 around 07 UTC 20 May and by 09 UTC had consolidated with the convection near the KS-OK
321 border to form a more intense convective segment with a well-defined trailing stratiform region
322 that then propagated through the network between 09 and 12 UTC. By 13 UTC, the convective
323 leading edge had exited the network, leaving the network dominated by a large area of stratiform

324 rain. Peak reflectivities within the network exceeded 60 dBZ over a depth of several km from 2
325 km above and to 3 km below the freezing level, a strong sign of significant hail (May and
326 Keenan 2005; Lerach et al. 2010), while 40 and 50 dBZ echoes reached upwards of 16 and 10
327 km, respectively, placing this case within the top 0.005% in intensity (i.e., 1 feature in 20,000 is
328 as strong) with regard to 40 dBZ echo penetrations based on a TRMM observed precipitation
329 feature climatology (Zipser et al. 2006). With 45 dBZ echoes reaching more than 10 km above
330 the freezing level, a further indication of the presence of hail (Waldvogel et al. 1979); this case is
331 well suited to test and evaluate the new Goddard 1M 4ICE scheme with hail.

332 As in previous GCE modeling studies (e.g., Zeng et al. 2007, 2008; L2011), the 3D GCE was
333 driven by large-scale forcing data obtained from a variational analysis approach (Zhang *et al.*
334 2001), in this case, from the MC3E sounding network. Model experiments were run for 4 days
335 starting at 0000 UTC 17 May and ending at 0000 UTC 21 May using 1-km horizontal grid
336 spacing and a 256 x 256 km horizontal domain (similar in size to the sounding network) with a
337 stretched vertical grid having 76 levels and a top near 27 km. A 2-second time step was used,
338 and surface sensible and latent heat fluxes were imposed from the variational analysis.

339

340 2) A MODERATE TROPICAL CONTINENTAL CASE: TRMM LBA

341

342 The 23 February 1999 case was previously presented in L2007 and L2011. It was
343 characteristic of the widespread, weaker monsoon-like convection observed within the westerly
344 wind regime (Cifelli et al. 2002; Rickenbach et al. 2002) during TRMM LBA. It falls within
345 about the top 1% of TRMM precipitation features (i.e., 1 feature in 100 is as strong) in terms of
346 40 dBZ echo height intensity (Zipser et al. 2006) and provides a good contrast to the more

347 intense 20 May 2011 MC3E case. On this day, daytime heating triggered widespread convection
348 that loosely organized into southeast-northwest bands. A long thin convective band developed
349 by 2000 UTC and by 2100 UTC was already decaying. Dual-Doppler observations were
350 collected for the northern portion of this line where 40 dBZ echoes reached to about 7 km.
351 Please see L2007 and L2011 for more details.

352 The current model setup closely follows L2011 and uses the same horizontally homogenous
353 initial conditions based on the 1200 UTC morning sounding taken at Rebio Jaru having weak
354 low-level northwesterly flow and a 500 m mixed-layer CAPE of 934 J kg^{-1} , cyclic lateral
355 boundary conditions and convection initiated by imposing time-varying (diurnal) surface fluxes
356 based on surface observations collected at two different sites (ABRACOS Hill and Ji Parana).
357 The horizontal domain was kept at 128 x 128 km as in L2011, but the horizontal grid resolution
358 was improved to 200 m in both directions and the time step reduced to 2 seconds. The stretched
359 vertical grid was kept at 70 levels with a top near 23 km. The north-south oriented rectangular
360 patch of higher sensible/lower latent heat fluxes (Ji Parana) imposed in L2011 was lengthened to
361 18 x 80 km (solid rectangle in Fig. 8b) and replaced by higher latent/lower sensible heat fluxes
362 (ABRACOS) if accumulated rainfall exceeded 3 mm over the patch to allow for some cloud
363 feedback. Simulations were run for 6 hours.

364

365 *d. Numerical experiments*

366

367 For each case, seven numerical experiments are conducted. Three experiments are made using
368 previous versions of the 3ICE-graupel scheme: the original (3ice0), L2007 (3ice1), and L2011
369 (3ice3). Four variations of the new 4ICE scheme are tested: with smaller-, medium- and larger-

370 sized hail with the bin rain evaporation correction (4iceb sml, 4iceb med, and 4iceb lrg,
371 respectively) and smaller-sized hail without the evaporation correction (4ice sml). Smaller-,
372 medium-, and large-sized hail use fixed hail distribution intercepts of 0.0200, 0.0020, and 0.0002
373 cm^{-4} , respectively. See Table 2 for a list of the numerical experiments performed for these cases.

374

375 3. Simulation results and validation

376

377 a. MC3E

378

379 Figure 2b shows the 4ICE control simulation (4iceb sml) compared to the observed convective
380 line (Fig. 2a) as it passed through north-central OK. Despite the restrictive double cyclic
381 boundary conditions, the model captures the general organization and intensity of the system
382 with an eastward propagating north-south oriented intense convective leading edge with a
383 trailing stratiform region off to the northwest. A vertical east-west cross section through the
384 center of the model domain (Fig. 2c) does show an erect intense uni-cellular convective
385 structure, but the simulated trailing stratiform region appears to be noticeably narrower.

386 Time-height cross sections (Fig. 3) of NEXRAD and simulated peak radar reflectivity² within
387 the MC3E sounding array and model domain are shown from 00 UTC 20 to 00 UTC 21 May
388 2011; the observed period 06 to 12 UTC (Fig. 3a) covers the formation of the main convective
389 line within the sounding array until the leading edge propagated out of the array. Peak
390 reflectivities within this line exceeded 60 dBZ with 50 dBZ echoes reaching 9 km and 40 dBZ
391 echoes 15 km. At 06 UTC, the simulations are still quite weak. The model imposed large-scale

² Simulated radar reflectivities were calculated from model rain, snow, graupel and hail contents assuming inverse exponential size distributions and accounting for all size mappings using the formulation of Smith et al. (1975) and Smith (1984).

392 forcing is horizontally uniform and first results in a patchwork of smaller convective cells over
393 the domain that require 2-3 hours to respond to the organizing shear and form into a squall line
394 (~09 UTC). The 3ICE graupel runs (Figs. 3b-d) significantly underestimate the peak intensity of
395 the observed squall line above the freezing level. The 3ice1 and 3ice3 runs produce a band of
396 elevated reflectivity maxima 3 km above the freezing level whereas the observed reflectivities
397 monotonically decrease with height above the freezing level. Graupel with its relatively slow
398 fallspeeds (Fig. 4b) is carried aloft by the strong updrafts in the convective cores (Fig. 4a) where
399 its mass is maximized well above the freezing level. In contrast, the 4ICE simulations (Figs. 3e-
400 h) produce higher reflectivity values just above the freezing level due to higher fall speeds (Fig.
401 4c) that keep the peak hail mass nearer the freezing level as well as peak values that decrease
402 monotonically with height: both in good agreement with the observations. Profiles of peak
403 reflectivity within the model domain over the period 09 to 15 UTC (sampled to match the
404 observed squall line structure from 06 to 12 UTC, Fig. 5) show all three 3ICE simulations have a
405 pronounced low bias that ranges from about 5 dBZ below the freezing level to as much as 15
406 dBZ above the freezing level. The 4ICE simulations show a marked improvement in the bias at
407 almost all levels except for 4iceb lrg, which produces excessively large reflectivities (~15 dBZ)
408 near the melting level. The medium hail profile has the smallest overall bias and agrees best
409 with the observed. Though not quite as good, the smaller hail runs are significantly improved
410 over the 3ICE with a consistent low bias of just 5 dBZ at all levels.

411 Figure 6 shows statistical CFADs constructed over these same observed (06 to 12 UTC) and
412 model (09 to 15 UTC) time periods and domains. The observed CFAD shows higher
413 concentrations broadly ranging from 0 to 40 dBZ below the melting level; aloft a coherent core
414 of higher probabilities increases from 10 dBZ near 200 mb to 25 dBZ just above the freezing

415 level. Concentrations of infrequent but more intense echoes extend out to near 65 dBZ at and
416 below the freezing level, 50 dBZ at 12 km, and 40 dBZ at 16 km. None of the simulated 3ICE
417 CFADs (Figs. 6b-d) produce reflectivities over 50 dBZ above the freezing level and thus miss
418 the stronger echoes in the tail of the observed distribution consistent with Fig. 5. Furthermore,
419 the core of highest probabilities is either too broad with too many 40 dBZ echoes due to graupel
420 (3ice0), shifted too high with too many 30 dBZ echoes due to snow (3ice1), or shifted too low
421 with too many weak echoes (3ice3) compared to the NEXRAD data. In contrast, all of the 4ICE
422 simulations produce much better concentrations of both the moderate core echoes and the
423 infrequent but intense echoes that arise from hail relative to the observations (Figs. 6e-h). The
424 4ICE core distributions are still too broad as a result of too many weak echoes, but their overall
425 slope along on the right edge is fairly well aligned with the observed core probabilities.

426 The improvements in the 4ICE distributions are reflected in the normalized overlap score
427 between the observed and simulated PDFs at each level in the CFADs where unity represents
428 perfect overlap and zero no overlap. Figure 7 shows the 4ICE PDFs are consistently better than
429 the 3ICE above the freezing level from 6 to 10 km. Below the melting layer, all of the
430 simulations are similar and better than in the mixed and ice phase regions. The agreement
431 between the simulations and observations drops off sharply near storm top where simulated radar
432 echoes become too weak. This discrepancy was noted in L2011 and could be due to entrainment
433 effects wherein dry air disproportionately sublimates small particles while preserving relatively
434 large particles. Cloudsat CFADs show a distinct difference between convective and anvil
435 regions (Luo et al. 2009) with the highest probabilities in the anvil concentrated at the lowest
436 reflectivities, indicating mostly smaller particles, before shifting to higher values lower in the
437 cloud consistent with accretion and aggregation. In contrast, convective cloud top PDFs are

438 broader, indicating a mix of small and large particles, with a much greater proportion of larger
439 particles. The current model CFADs resemble the Cloudsat anvil distributions at upper levels.

440

441 *b. LBA*

442

443 Hail is associated with intense convection and as shown needed to produce stronger radar
444 echoes. However, a key objective is for the 4ICE scheme to respond appropriately to the
445 environment without the need for excessive tuning. The moderate intensity 23 February case is
446 suitable for evaluating the new scheme for a weaker convective environment. A radar CAPPI for
447 this case (Fig. 8a) shows the northern end of the transient convective line as it was starting to
448 decay. Individual convective cells are loosely aligned with a small stratiform area extending
449 northwestwards. The simulated convective leading edge using the 4ICE scheme (4iceb sml, Fig.
450 8b) is also cellular in nature and loosely organized into a north-south line with a small stratiform
451 area extending northwestward, consistent with the weak southeast-northwest oriented shear on
452 this day. A vertical east-west cross section (Fig. 8c) through the center of the domain shows the
453 leading edge is somewhat multi-cellular (also see Fig. 10a) with a small, undeveloped stratiform
454 area. Following L2011, a 64 x 64 km subdomain over the northern portion of the simulated line
455 (dashed box in Fig. 8b) was used in the analyses with model data averaged to the 1-km resolution
456 of the radar analyses. The mean convective fraction within this subdomain over the simulation
457 period from 300 to 360 minutes using the 4ICE scheme ranged from 0.46 to 0.49 in close
458 proximity to the radar value of around 0.4³.

³ As in L2007 and L2011, convective fractions were computed based on Rickenbach and Rutledge (1998), a texture algorithm applied to radar reflectivity data that largely follows Steiner et al. (1995) to match the radar observations.

459 Figure 9 shows time-height cross sections of peak reflectivity observed by the NCAR S-pol
460 radar within the dual-Doppler analysis domain and simulated within the model subdomains. As
461 previously noted (L2007; L2011), the original scheme (3ice0) with its excessive graupel
462 produces 40-dBZ echoes that penetrate much higher (over 12 km, Fig. 9b) than was observed (~7
463 km, Fig. 9a). Though better, 3ice1 (Fig. 9c), which eliminates the dry collection of ice/snow by
464 graupel (L2007), still produces excessive 40-dBZ echoes penetrations. The 40-dBZ echoes in
465 3ice3 (L2011) are greatly reduced aloft and closer to the observed, but 3ice3 results in an
466 elevated reflectivity maximum above the freezing level (Fig. 3d) that was not observed. None of
467 the 3ICE graupel simulations can reproduce the 45-dBZ echoes immediately above the freezing
468 level (~4.9 km) as was observed (Fig. 9a). In contrast, despite the addition of the higher density
469 frozen drops-hail ice class, none of the 4ICE simulations (Figs. 9e-h) produces the excessive 40-
470 dBZ echo penetrations in 3ice0 and 3ice1 and all can replicate the observed monotonically
471 decreasing reflectivity structure above the melting level though clearly the medium (4iceb med)
472 and larger (4iceb lrg) hail peak reflectivities that are too strong around the freezing level. These
473 results suggest that hail or frozen drops with their higher fall speeds (Fig. 10c) relative to graupel
474 (Fig. 10b) are crucial not detrimental for reproducing the observed core reflectivity structure of
475 even moderate convection. Peak reflectivity profiles from the 64 x 64 km model subdomains
476 over the final 60 minutes (Fig. 11) confirms the strong (nearly 15 dBZ) to moderate (~5 dBZ)
477 over bias in the 3ice0 and 3ice1 simulations aloft, respectively, as well as the under bias (~8
478 dBZ) in 3ice3 near the melting level, which contributes to the elevated maxima. The 4ICE
479 simulations with smaller hail clearly perform the best and show almost no bias (less than ~4
480 dBZ) through nearly the entire depth of the storm. Remarkably, none of the 4ICE runs produce
481 the over bias evident in runs 3ice0 and 3ice1 in the upper part of the storm, and all produce

482 monotonically decreasing profiles with height in agreement with the observations. However,
483 quite obviously the medium to larger hail sizes in runs 4iceb med and 4iceb lrg are much too
484 large, producing over biases of up to ~10 to 15 dBZ around the melting level. These results
485 suggest the new 4ICE scheme is quite capable of responding appropriately to the intensity of the
486 convective environment and can outperform the 3ICE-graupel scheme in terms of peak
487 reflectivities even in a moderate intensity environment. This actually is consistent with
488 polarimetric radar and wind profiler evidence that frozen drops or hail quite often are present in
489 tropical convection (Jameson et al. 1996; May et al. 2001); *in situ* aircraft data also confirmed
490 the presence of frozen drops in this case (Stith et al. 2002)

491 Simulated and observed radar CFADs (Fig. 12) also show the 4ICE scheme equals or
492 outperforms the 3ICE-graupel scheme in terms of overall reflectivity distributions for the
493 moderate 23 February case. The highest observed concentrations (Fig. 12a) gradually decrease
494 from between ~5 to 20 dBZ at and below the melting level (~4.9 km) to ~ -5 to 15 dBZ near
495 storm top with less apparent aggregation than MC3E (Fig. 6a). Low probabilities of moderately
496 strong echoes reach ~50 dBZ at and below the freezing level, 40 dBZ at ~7 km, and 30 dBZ at
497 12 km. As in MC3E only more pronounced, the original scheme (3ice0, Fig. 12b) produces
498 excessive concentrations of 20 to 40 dBZ echoes between the melting level and 10 km, and
499 again, though the number of excessive 30 to 40 dBZ echoes is reduced in 3ice1 (Fig. 12c), the
500 core of peak probabilities is ~10 dBZ too high. Core probabilities for 3ice3 are noticeably better
501 (Fig. 12d), but the amount of echoes too weak is higher as is the penetration of 20 to 30 dBZ
502 echoes near storm top. There is also an unphysical notch in the higher echo distribution just
503 above the freezing level. Though hail amounts are small in the 4ICE simulations, they dominate
504 the higher reflectivity values; the peak concentrations are nearly invariant (Figs. 12e-h) but are as

505 good as or better than the 3ICE at every level, aligning well with the observed concentrations
506 along 20 dBZ from the melting level to 12 km. In terms of echoes stronger than 20 dBZ in the
507 distribution tail, the smaller hail results (Figs. 12e-f) match the observed frequencies extremely
508 well while the medium (Fig. 12g) and larger hail (Fig. 12h) runs are overly intense especially
509 near the melting level. And as with MC3E, simulated probabilities collapse below 0 dBZ at
510 storm top and are ~10 dBZ weaker than observed.

511 The overall level-by-level performance is confirmed by the profiles of normalized PDF
512 matching scores (Fig. 13). From the freezing level to 10 km, the 3ice3 and 4ICE PDFs are all in
513 excellent agreement with the observed with matching scores on the order of 0.8, much better
514 than 3ice0 and 3ice1. Above 10 km, the performance of 3ice3 drops off quickly while the 4ICE
515 simulations continue to perform well up to 12 km before they begin to deviate from the observed
516 echo distributions near storm top. Below the melting level, the 3ICE and 4ICE schemes all
517 perform about the same and reasonably well with scores of around 0.75.

518

519 *c. Rainfall comparisons and validation*

520

521 In addition to radar reflectivity, the 4ICE scheme is validated with regard to surface rain rates.
522 Figures 14a,b show the instantaneous surface rain rates associated with the 4iceb sml simulations
523 shown in Figs. 2b and 8b for MC3E and LBA, respectively. Surface rain features in the
524 convective region are larger, better organized and more intense in MC3E; the stratiform region is
525 also larger and more coherent. Simulated surface rain rate histograms (Figs. 14c and d) can be
526 compared to observed rain rate histograms derived from the national Doppler radar network for
527 MC3E (Fig. 14c) and ground-based radar deployed for the LBA field campaign (Fig. 14d). The

528 3ICE and 4ICE histograms tend to fall into two distinct clusters, which is more apparent in the
529 MC3E case. For MC3E, the 3ICE simulations significantly underestimate the occurrence of
530 more intense rain rates in the tail of the distribution (Fig. 14c); the 4ICE histograms also
531 underestimate probabilities of extreme rain rates but are distinctly better than the 3ICE. For
532 LBA, the results are noisier due to a smaller sample size, but the overall results similar: both sets
533 of simulations underestimate the proportion of strongest rain rates but with the 4ICE simulations
534 closer to the observed histogram than the 3ICE. These results suggest the 4ICE simulations are
535 producing more realistic surface rain rate distributions than the 3ICE in each environment.

536

537 **4. Summary and conclusions**

538

539 The improved Goddard 3ICE 1M BMS based on Rutledge and Hobbs (L2011, cloud ice, snow
540 and graupel) was modified and hail processes added to produce an improved 1M 4ICE BMS
541 (cloud ice, snow, graupel, and frozen drops-hail) capable of more realistically simulating the
542 radar reflectivity patterns of intense and moderate convection better than previous 3ICE versions.
543 Hail processes taken from the 3ICE-hail scheme based on Lin et al. (1983) include hail riming,
544 accretion of rain, deposition/sublimation, melting, shedding and wet growth. Hail collection of
545 other ice species under dry growth conditions was eliminated to prevent the same excessive
546 buildup as had occurred with graupel (L2007); however, hail near wet growth is permitted to
547 collect other ice particles. Processes that freeze rain now initiate hail not graupel. Five new hail
548 processes were added: wet hail accretion of graupel, hail rime splintering, hail (also graupel)
549 conversion to snow via depositional growth at colder temperatures, hail conversion to graupel via
550 riming under non wet growth conditions, and wet graupel conversion to hail.

551 Besides adding the frozen drops-hail category, snow autoconversion was strengthened based
552 on the evidence of aggregation at colder temperatures (Field 1999), the collection efficiency of
553 cloud ice by snow increased, a water vapor diffusivity correction factor added, and a small
554 threshold introduced to prevent ice deposition growth into snow when ice is small. Maximum
555 ice supersaturation was increased to 20%, and cloud evaporation restricted to areas of subsidence
556 to mitigate spurious evaporation effects at cloud boundaries. Cloud ice can now persist in ice
557 sub-saturated conditions as commonly observed. Snow and graupel size mappings from L2011
558 were adjusted, including an increased aggregation effect for snow. A corresponding snow
559 density mapping was added (Brandes et al. 2007), and graupel divided into low and moderate
560 densities. Lastly, an SBM-based rain evaporation correction factor (Li et al. 2009) was added.

561 The new Goddard 4ICE scheme was used to simulate an intense continental squall line
562 observed during MC3E to evaluate its ability to simulate intense convection with significant hail
563 as well as a loosely-organized transient line of moderate convection from TRMM LBA to ensure
564 the scheme does not over predict less intense convection. For the intense squall line, the 4ICE
565 scheme with smaller and medium-sized hail, outperformed prior versions of the 3ICE-graupel
566 scheme in terms of peak reflectivities; larger hail produced excessively high values. Without
567 hail, the 3ICE-graupel versions could not generate reflectivities over 50 dBZ below freezing,
568 vastly underestimating the peak observed reflectivities throughout the depth of the storm with a
569 low bias near 15 dBZ at the freezing level, 10 dBZ at midlevels, and 5 to 15 dBZ at upper levels.
570 In contrast, the bias was significantly reduced in the 4ICE runs above the freezing level, except
571 for the strong over bias of 15 dBZ near the melting level for larger hail. For medium hail the
572 bias is less than 5 dBZ at almost all levels except for a positive bias of ~5 dBZ at the freezing
573 level. The smaller hail simulations produced a consistent low bias of just 5 dBZ at nearly every

574 level, still a noticeable improvement over the 3ICE graupel simulations. The 4ICE simulations
575 produced radar reflectivity CFADs in better agreement with observations (as reflected in their
576 normalized PDF overlapping scores) from 5 to 10 km with more realistic extremes in the
577 distribution tails and more realistic reflectivity structures above the freezing level with peak
578 reflectivities monotonically decreasing with height as observed versus the 3ICE graupel
579 simulations, which often produced elevated reflectivity maxima. Below the melting level, the
580 4ICE runs with smaller hail had peak reflectivities similar to the 3ICE. Those for medium and
581 larger hail were greater due to contributions from melting hail and were closer to the
582 observations for medium hail but not for larger where values were excessive. The agreement
583 between simulated and observed CFADs below the melting level was similar for all runs.
584 Overall, the 4ICE simulation with medium hail performed the best for the intense MC3E case.

585 For TRMM LBA, adding frozen drops/hail *per se* did not necessarily cause unrealistically
586 large reflectivity values⁴. While medium and larger hail did result in excessive peak
587 reflectivities by as much as 10 to 15 dBZ near the melting level, smaller hail had very small
588 biases (less than 5 dBZ) that were on average as good or better than the 3ICE versions at every
589 level. Small hail eliminated the over bias at middle and upper levels in the original and L2007
590 3ICE runs and outperformed the L2011 run above the freezing level by reproducing the observed
591 monotonic decrease with height and eliminating the unrealistic elevated reflectivity maxima.
592 The 4ICE simulations also produced radar CFADs whose normalized PDF scores were equal to
593 or superior to the 3ICE at all levels. The 4ICE simulations with smaller hail performed the best
594 overall for the moderate intensity LBA case.

⁴ Trace amounts of mass first appear in the 4th ice class as frozen rain at ~10 m/s for the LBA cases and ~5 m/s for MC3E, but the sizes are very small and combined with the fact that reflectivity values take a while to grow strong shows that the 4ICE scheme is not predisposed to generating larger hail particles nor strong dBZs for every case.

595 Surface rainfall histograms were also used to evaluate the schemes for both cases with similar
596 results. In each case the 3ICE simulations significantly underestimated the occurrence of higher
597 rain rates compared to observed histograms while the 4ICE histograms had a noticeably higher
598 occurrence of stronger rain rates which were closer to but still less than the observed.

599 Though the bin rain evaporation correction (Li et al. 2009) did alter the temporal variation of
600 peak reflectivities, it had very little effect on either the peak reflectivity profiles or the model
601 CFADs in either case despite evidence that it reduced the intensity of the cold pool distribution
602 (see Fig. 15). The double cyclic lateral boundaries likely dampened its impact as initial
603 excessive evaporation without the correction could over moisten the sub-cloud layer and inhibit
604 successive over evaporation thus masking the effects of the correction. The LBA case was in a
605 moist environment and of short duration, which could also reduce the impact of the correction.

606 The 4ICE scheme with a frozen-drops hail category can simulate more intense radar echoes,
607 though choosing *a priori* the hail intercept for intense or moderate convection is not optimal. A
608 size-mapping scheme may alleviate the issue, but ultimately a multi-moment scheme is likely
609 needed. 4ICE also replicates the observed monotonic decrease in peak reflectivities with height
610 as a result of increasing the range of particle fall speeds to include the higher values associated
611 with frozen drops/hail, allowing a greater portion of their mass to remain near the freezing level.
612 The scheme also adds pathways (apparent processes) by which particles can move to other
613 categories due to their growth mechanisms. Hail can be rimed. If cloud freezes quickly, it
614 creates air pockets, which should lower hail density. Just as graupel is assumed to increase in
615 density and become hail when reaching wet growth, so too should hail density go down when
616 riming rates fall below wet growth. This is the basis for the new Primh hail riming to graupel
617 process. For the MC3E control case, Primh increases (decreases) the peak average graupel (hail)

618 content by 25% (5%), decreases the peak echoes by 3 dBZ at and above 12 km, and improves the
619 CFAD score by 2 to 4% from 4 to 13 km. The other new set of pathways relates to trace
620 amounts of graupel and hail that often persist aloft in the model long after convection ends.
621 Initially, sedimentation lowers their mass, but as the mass get small, with a fixed intercept, so
622 too do their fallspeeds. The result is tiny amounts of graupel and hail (mean volume diameters of
623 a few hundred microns or less) suspended over a broad area where they continue to grow from
624 deposition. The new P_{vapg} and P_{vaph} vapor conversion to snow processes reduce the area and
625 trace amounts of suspended graupel and hail but have minimal impact on the hail cores. For the
626 MC3E control case, peak reflectivity profiles are ~unchanged by P_{vapg}/P_{vaph} below 12.5 km
627 (the same is true for Primh below 9 km) as the collection of large amounts of super cooled water
628 dominates growth. Peak average snow, graupel and hail amounts increase ~5%, decrease 5%,
629 and increase 8%, respectively, when P_{vapg}/P_{vaph} are activated; P_{vapg}/P_{vaph} improve the
630 CFAD score by 2 to 5% from 4 to 11 km. The new apparent processes have a small effect on the
631 hail cores but allow the model to better address variations in particle density while slightly
632 improving the overall echo pattern.

633 Allowing a smooth transition in ice particle density (Morrison and Grabowski 2008) and
634 fallspeed coefficients (Lin and Colle 2011) is a recent and realistic addition to BMSs, though
635 comparisons with radar observations for intense cases are needed to evaluate their performance
636 and in contrast to having a separate hail category.

637 Ultimately, cases should be tested at higher resolution (100 m) to ensure the dynamics are well
638 resolved and not contributing to any biases. Further study is also needed to address the
639 abundance of weaker model dBZs near storm top to determine if this is an artifact of the radar
640 observations or the microphysics. Testing the 4ICE scheme in other environments using remote

641 sensing data for validation (Matsui et al. 2009) is important for systematically identifying and
642 eliminating any remaining biases. The ability to match the distribution and peak values of radar
643 reflectivities at all levels of a convective system is a fairly stringent test, but radar intensities are
644 not a unique solution and can arise from a variety of particle combinations. This is where *in situ*
645 aircraft observations could be very valuable in helping to further constrain the particle
646 characteristics. The next iteration of the Goddard scheme is planned to be 2M to ingest and
647 include the effects of aerosols and improve limitations inherent in 1M.

648
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663

APPENDIX

664
665

666 *a. The hail budget equation*

667

668 The hail budget equation for the new 4ICE scheme is given as:

$$669 \frac{\partial q_h}{\partial t} = -V \cdot \nabla q_h - \frac{1}{\rho} \frac{\partial}{\partial z} ((w - V_h) q_h \rho) + D_{qh} + Phfr + (1 - \delta_3) Piacr + (1 - \delta_3) Pr aci + (1 - \delta_2) Psacr$$

670

$$671 + (1 - \delta_2) Pr acs + Dgacr + Pr acg + Dhacw + Dhacr + Whaci + Whacs + Whacg + Pg2h$$

$$672 + Phdep - Phsub - Pvaph - Pr imh - Phmlt \quad (A1)$$

673 where the first three terms on the RHS of Eq. A1 are the horizontal advection, vertical advection

674 and diffusion of hail, respectively, while $Phfr$ is the freezing of rain to hail, $Piacr$ cloud ice

675 accretion of rain, $Pr aci$ rain accretion of cloud ice, $Psacr$ snow accretion of rain, $Pr acs$ rain

676 accretion of snow, $Dgacr$ graupel accretion of rain, $Pr acg$ rain accretion of graupel, $Dhacw$

677 hail riming, $Dhacr$ hail accretion of rain, $Whaci$, $Whacs$, and $Whacg$ wet hail accretion of

678 cloud ice, snow and graupel, respectively, $Pg2h$ the conversion of graupel to hail, $Phdep$ hail

679 deposition, $Phsub$ hail sublimation, $Pvaph$ the conversion of hail to snow via deposition,

680 $Pr imh$ hail riming to graupel and $Phmlt$ hail melting. $Phfr$, $Dhacw$, $Dhacr$, $Whaci$, $Whacs$,

681 $Phdep$, $Phsub$, and $Phmlt$ follow the formulations of Lin et al. (1983), while $Piacr$, $Pr aci$,

682 $Psacr$, $Pr acs$, $Dgacr$, and $Pr acg$ follow the formulations of Rutledge and Hobbs (1984).

683 $Whacg$ follows the Lin et al. (1983) formulation for $Whacs$ but using graupel instead of snow

684 parameters. Graupel is assumed to increase in density and become hail upon reaching wet

685 growth such that

686 $Pg2h = \frac{q_g}{dt}$ at the moment when $Dgacw + Dgacr > Pgwet$ (A2)

687 where the wet growth of graupel $Pgwet$ is computed using the formula for hail wet growth
 688 from Lin et al. (1983) but with graupel parameters⁵.

689 Just as lower density particles can transition to a higher density class of particles, in the new
 690 4ICE scheme, the reverse can occur. As such, when hail particles experience riming or
 691 deposition at colder temperatures, they are transitioned towards graupel and snow, respectively.
 692 For both processes, an increasing proportion of the mass of hail acquired via riming and
 693 deposition along with an equal portion of the previous hail mass is transferred to graupel and
 694 snow, respectively, using a sliding temperature scale with the proportion increasing with
 695 decreasing temperature. The conversion of hail to graupel via riming is thus formulated as:

696 $Pr imh = F_{rime} \times Dhacw$ (A6)

697 where F_{rime} is given by $F_{rime} = 2.0 * \left(\frac{Tairc}{(t00-t0)} \right)^2$ (A7)

698 where $Tairc$ is the air temperature in degrees Celcius, $t00$ is 238.16 degrees K, and $t0$ is 273.16
 699 degrees K. Similarly the conversion of hail to snow via deposition is formulated as:

700 $Pvaph = F_{vap} \times Phdep$ when the cloud water mixing ratio $q_c < 1.e-5$ g/g (A8)

701 where F_{vap} uses the same temperature scaling as F_{rime} . The same form of the relation is used for
 702 the conversion of graupel to snow via deposition $Pvapg$. Although Hallet-Mossop rime
 703 splintering (Hallet and Mossop 1974) is not directly part of the hail equation, it does affect the
 704 hail riming term $Dhacw$ and is computed as:

⁵ In nature, larger graupel particles would reach wet growth first. Converting only the mass above the mean volume diameter (0.5004) for the MC3E control case reduced the maximum average hail content by over 15%. However, peak reflectivities below 9 km were nearly unchanged due likely to high riming rates in the updraft cores but were decreased by 3 dBZ or more above 12 km.

705 $Pihmh = T_{fact} * Dhacw * 1000 * Xnsp \ln t * Xmsp \ln t$

706 where

707 $T_{fact} = 0.5$ when $-8^{\circ}C < Tairc < -2^{\circ}C$

708 $T_{fact} = 1.0$ when $-6^{\circ}C < Tairc < -4^{\circ}C$

709 where the peak number of ice splinters generated per milligram of rime $Xnsp \ln t = 370$ and the

710 mass of each splinter $Xmsp \ln t = 4.4e^{-8}$ grams. $Pihmh$ is first subtracted directly from $Dhacw$.

711 The water vapor diffusivity (D_v) in air was assumed to be a constant ($2.26e^{-5} \text{ m}^2\text{s}^{-1}$) in Rutledge

712 and Hobbs (1984); it is parameterized as a function of temperature and pressure in this study

713 following Massman (1998):

714
$$D_v = D_0 \left(\frac{T}{T_0} \right)^{1.81} \left(\frac{P_0}{P} \right),$$

715 where D_0 is the water vapor diffusivity ($2.18e^{-5} \text{ m}^2\text{s}^{-1}$) at $T_0=273.15 \text{ K}$ and $P_0=1013.25 \text{ hPa}$. The

716 importance of water vapor diffusivity on the diffusional growth rate of ice crystals as a function

717 of pressure and temperature is illustrated in Fig. 9.4 of Rogers and Yau (1989). An adjustment

718 factor (F_{dwv}) is used to adjust the diffusional growth rate of ice crystals based on a constant

719 water vapor diffusivity.

720

721 *b. Rain evaporation correction*

722

723 The rain evaporation correction uses the empirical formula of Li et al. (2009):

724 $r(q_r) = 0.11q_r^{0.27} + 0.98$ (A9)

725 where r is the ratio of the rain evaporation rate between bulk and bin microphysics and q_r (g/kg)
 726 is the rain mixing ratio; r is based on cloud-resolving model simulations using both bulk and bin
 727 microphysics and can be used to scale down the bulk rain evaporation rate. In the new 4ICE
 728 scheme, the correction factor is made “physical” by scaling the rain intercept (i.e., increasing the
 729 grid local raindrop size) until the bulk rain evaporation rate matches the bin such that:

$$730 \quad F_{tnw} = \left(\frac{1}{r}\right)^{3.35} \quad (A10)$$

731 where F_{tnw} is the scaling factor for the rain intercept parameter.

732

733 *c. Snow mapping*

734

735 The snow mapping scheme maps the snow intercept parameter as a combination of variations in
 736 temperature and mixing ratio. Variations in mixing ratio are set for two distinct conditions: (1)
 737 at cold temperatures where aggregation effects are small, sizes are small and only slowly
 738 increase with increasing mixing ratio and (2) near the melting layer where aggregations effects
 739 are large, sizes are larger and size increases significantly with increasing mixing ratio. Another
 740 set of parameters controls how quickly the cold setting variations transform to the warm setting
 741 variations through the aggregation zone. For both cold and warm regions, an exponent is used to
 742 control the snow intercept; as this exponent approaches zero, snow sizes relax to those for a fixed
 743 snow intercept (i.e., larger sizes), and when the exponent approaches one, snow sizes collapse to
 744 that of a small base size. The formula for the snow size exponent is given by:

$$745 \quad F_{exp} = X_{sml} - X_{sml} \times \min\left(S_{lim}, \max\left(0., \frac{(qs1-sno1)^{sexp1}}{dsno1}\right)\right) \quad (A11)$$

746 which is then used create the mixing ratio component of the snow intercept scaling factor:

747
$$Ftnsq = \left(\frac{qs1}{S_{base}} \right)^{Fexp} \quad (A12)$$

748 where $qs1$ is the snow water content in g/m^3 , and $Sno1$ is a snow water content threshold in
 749 g/m^3 above which snow sizes begin to increase. Snow sizes then continue to increase at an ever
 750 increasing rate over the next $dsno1$ g/m^3 until reaching the limit S_{lim} . X_{sml} is arbitrarily given
 751 a number close to but less than one. This allows snow sizes to vary ever so slightly (i.e., not be a
 752 constant size) between snow contents of 0.0 and $Sno1$. The parameter settings for cold
 753 conditions transform through the snow aggregation zone (~ -20 to 0° C) to those near 0° C as:

754
$$P = Pwarm - (Pwarm - Pcold) \left(\frac{Tairc}{Tcold} \right)^{Sexp} \quad (A13)$$

755 where $Pwarm$ is the parameter value near the melting level, $Pcold$ the parameter value for cold
 756 conditions, and $Tcold$ the air temperature in degrees C for the cold parameter settings. An air
 757 temperature component for the snow intercept scaling factor, given by

758
$$FtnsT = \left(\exp(-1. \times Tslopes \times Tairc) \right)^{Fexp} \quad (A14)$$

759 where $Tslopes$ is the rate of snow intercept change with temperature on a natural logarithm scale
 760 and $Tairc$ is capped by $Tcold$, is then combined with the mixing ratio component of the snow
 761 intercept scaling factor to obtain the total snow intercept scaling factor:

762
$$Ftns = Ftnsq \times FtnsT \quad (A15)$$

763 with the condition that the snow size cannot go below a minimum value $D_{snowmin}$. The snow
 764 intercept mapping is combined with the Brandes et al. (2007) relation between snow density and

765 median snow volume to get: $\rho_s = 0.001996 * \left(\frac{Ftns \times tns}{(qs \times \rho)} \right)^{0.2995} \quad (A16)$

766 where ρ_s is snow bulk density and ms the snow intercept. Table A1 lists the specific values
767 used for the snow mapping.

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1217

1218 Table 1. Microphysical processes modified or added to the original (i.e., Tao and Simpson 1993;
 1219 Tao et al. 2003) Goddard 1M Rutledge and Hobbs-based 3ICE-graupel bulk microphysics
 1220 scheme (updated from L2011). Current changes associated with the new single-moment 4ICE
 1221 scheme are shown in *italics*. New hail processes are in ***bold italics***. “f()” indicates “function
 1222 of”. Esi, Egc, and Esc are the collection efficiencies of cloud ice by snow, cloud by graupel and
 1223 cloud by snow, respectively. Qc0 is the cloud water threshold for snow riming, Qi0 the cloud ice
 1224 threshold for snow autoconversion, ssi the supersaturation percentage with respect to ice,
 1225 RH/RHice the relative humidity for water/ice, Dwv the water vapor diffusivity, Vs/g the
 1226 snow/graupel fall velocity, Bh,i the immersion mode ice nucleating efficiency, IN the ice number
 1227 concentration, and Tair the air temperature. Qr, Qi, Qs, Qg, and Qh are the rain, cloud ice, snow,
 1228 graupel, and hail mixing ratios, respectively. The process nomenclature essentially follows Lin
 1229 et al. (1983) and Rutledge and Hobbs (1983, 1984). ** =0 if $Dhacw + Dhacr < 0.95 * Phwet$.

1230

1231	Process	Original	Modifications	Reference(s)/Notes
1232	-----			
1233	Psaut	Efficiency	Efficiency=0.15, Qi0 changed from	
1234		f(Tair)	<i>g/g to g/m⁻³, time scale reduced from</i>	
1235			<i>1000 to 300 s, Qi0 lowered to</i>	
1236			<i>0.4 g m⁻³, efficiency=0.25</i>	
1237				
1238	Psaci	Esi = 0.1	Esi f(snow diameter), maximum	See snow size mapping in
1239			Esi = 0.25, <i>maximum Esi=0.70,</i>	Fig. 1
1240			<i>=0 when Qs=0</i>	

1241				
1242	Praci/Piacr		Accounts for addition of cloud	Cloud ice fall speed follows
1243			ice fall speed, $=0$ when $Qr=0$,	Hong et al. (2004)
1244			<i>becomes hail not graupel</i>	
1245				
1246	Pracw		$=0$ when $Qr=0$	
1247				
1248	Psfi	Independent	Depends on RH, accounts for	Meyers et al. (1992);
1249		of RH	cloud ice size via Meyers IN,	Krueger et al. (1995)
1250			which is a f(ssi), <i>added Dwv</i>	
1251			<i>correction factor & Qi threshold</i>	
1252				
1253	Psfw		<i>Added Dwv correction factor</i>	
1254				
1255	Dgacs/Dgaci		Turned off	See Lang et al. (2007)
1256				
1257	Dgacw	Egc=1.0	Egc is f(graupel diameter),	See graupel size mapping in
1258			maximum Egc = 0.65,	Fig. 1; Khain et al. (2001)
1259			$=0$ when $Qg=0$	
1260				
1261	Dgacr/Pracg		$=0$ when $Qg=0$ or $Qr=0$	
1262				
1263				

1264	Psacw/Pwacs	Esc=1.0,	Esc=0.45, Qc0=1.0 g/kg,	Lang et al. (2007);
1265		Qc0=0.5 g/kg	<i>Qc0=0.5 g/kg, =0 if Qs=0</i>	Morrison and Grabowski
1266				(2008)
1267				
1268				
1269	Pracs/Psacr		<i>=0 if Qr or Qs=0, =hail not graupel</i>	
1270				
1271	Rime	None	Added and applied to	Hallet and Mossop (1974);
1272	Splintering		Psacw/Pgacw, not f(Vs/g)	f(Tair) and splinter mass
1273			or f(cloud size), <i>added for</i>	follow Ferrier (1994)
1274			<i>Dhacw</i>	
1275				
1276	Pidw/Pidep	Based on	Based on Meyers IN, which is	Fletcher (1962);
1277		Fletcher	a f(ssi), <i>added Dvw</i>	Meyers et al. (1992)
1278			<i>correction factor to Pidep</i>	
1279				
1280	Pint	Based on	Based on Meyers IN, which is	Fletcher (1962);
1281		Fletcher	a f(ssi), previous ice	Meyers et al. (1992)
1282			concentration checked	
1283				
1284	Immersion	None	Added based on Diehl	Diehl and Wurzler (2004);
1285	Freezing			Diehl et al. (2006), assumes
1286				Bh,i = 1.01 e-2 for pollen

1287				
1288	Contact	None	Added based on Cotton	Cotton et al. (1986);
1289	Nucleation		and Pruppacher for	Pruppacher and Klett (1980),
1290			Brownian diffusion only	500 active nuclei per cc
1291				with radii of 0.1 microns
1292				
1293	Saturation	Sequential	Modified sequential, iterative,	Tao et al. (2003)
1294	Adjustment	based on Tao	ssi up to 10%, <i>ssi up to 15%</i>	<i>Reisner (personal)</i>
1295			<i>(T < -44°C) and 20% (T > -38°C),</i>	
1296			<i>no evaporation if W > -0.1 m/s,</i>	
1297			<i>no sublimation if RHice > 70%</i>	
1298			<i>or W > 0 m/s</i>	
1299				
1300	Psdep/Pgdep		=0 if Qs/Qg=0	
1301				
1302	Snow/Graupel	None	Allowed if outside cloud and air	
1303	Sublimation		subsaturated, <i>allowed if air subsaturated</i>	
1304				
1305	<i>Pvapg/Pvaph</i>		<i>convert Qg/Qh to Qs via deposition</i>	
1306			<i>if Qc < 1.e-5 g/g, f(Tair)</i>	
1307				
1308	Snow/	Based on	Based on intercepts mapped according	
1309	Graupel	fixed	to snow/graupel mass and Tair, <i>revised,</i>	

1310	size	intercepts	<i>greater aggregation effect for snow</i>	
1311				
1312	<i>Snow density</i>	$=0.1 \text{ g cm}^{-3}$	$=0.05 \text{ g cm}^{-3}, f(\text{snow size})$	<i>Brandes et al. (2007)</i>
1313				
1314	<i>Graupel</i>	$=0.4 \text{ g cm}^{-3}$	$=0.3 \text{ g cm}^{-3}$ if $Qg < 2.0 \text{ g m}^{-3}$	<i>Brown and Swann (1997);</i>
1315	<i>density</i>		$=0.5 \text{ g cm}^{-3}$ if $Qg > 2.0 \text{ g m}^{-3}$	<i>Straka and Mansell 2005</i>
1316				
1317	Cloud ice	None or	Based on Hong, <i>included in</i>	Hong et al. (2004);
1318	fall Speed	Starr and Cox	<i>all sweep volumes</i>	Starr and Cox (1985)
1319				
1320	Ern		<i>Added rain evaporation correction</i>	<i>Li et al. (2009)</i>
1321			<i>via tnw, max correction=1.30,</i>	
1322			<i>Added Dwv correction factor</i>	
1323				
1324	Pmlts/Pmltg		<i>Added Dwv correction factor</i>	
1325				
1326	Whaci**		<i>Added from Lin, =0 if Qh=0,</i>	<i>Lin et al. (1983)</i>
1327				
1328	Whacs**		<i>Added from Lin, =0 if Qh/Qs=0</i>	<i>Lin et al. (1983)</i>
1329				
1330	Whacg**		<i>Follows Whacs, =0 if Qh/Qg=0</i>	
1331				
1332	Dhacw		<i>Added from Lin, =0 if Qh=0</i>	<i>Lin et al. (1983)</i>

1333			
1334	Primh	<i>convert Qh to Qg via riming,</i>	<i>see text on Milbrandt</i>
1335		<i>f(Phwet, Tair)</i>	<i>and Morrison (2013)</i>
1336			
1337	Dhacr	<i>Added from Lin, =0 if Qh/QR=0</i>	<i>Lin et al. (1983)</i>
1338			
1339	Phwet	<i>Added from Lin, =0 if Qh=0</i>	<i>Lin et al. (1983)</i>
1340			
1341	Phfr	<i>Follows Pgfr but frozen rain</i>	<i>Rutledge and Hobbs (1984)</i>
1342		<i>= hail not graupel, =0 if QR=0</i>	
1343			
1344	Phdep	<i>Added from Lin, =0 if Qh=0</i>	<i>Lin et al. (1983)</i>
1345			
1346	Phmlt/ Whacr	<i>Added from Lin</i>	<i>Lin et al. (1983)</i>
1347	-----		
1348			

1349 Table 2. Numerical experiments performed for both the 20 May 2012 MC3E and 23 February
 1350 1999 TRMM LBA cases using various options of the Goddard 1M bulk microphysics scheme
 1351 where N_{0h} is the hail intercept.

1352

1353	Experiment	Description	Reference(s)/Notes
1354	-----		
1355	3ice0	3ICE graupel (original)	Rutledge and Hobbs (1983, 1984)
1356			
1357	3ice1	Original + no graupel dry collection	Lang et al. (2007)
1358			
1359	3ice3	Original + no graupel dry collection	Lang et al. (2011)
1360		+ snow/graupel size mapping	
1361			
1362	4ice sml	New 4ICE with smaller hail	$N_{0h} = 0.0200 \text{ cm}^{-4}$
1363			
1364	4iceb sml	New 4ICE with smaller hail	$N_{0h} = 0.0200 \text{ cm}^{-4}$
1365		+ rain evaporation correction	
1366			
1367	4iceb med	New 4ICE with medium hail	$N_{0h} = 0.0020 \text{ cm}^{-4}$
1368		+ rain evaporation correction	
1369			
1370	4iceb lrg	New 4ICE with larger hail	$N_{0h} = 0.0002 \text{ cm}^{-4}$
1371		+ rain evaporation correction	

1372 Table A1. Values of snow intercept mapping parameters used to obtain the characteristic snow
 1373 size mapping shown in Fig 1e.

1374

1375	Snow Mapping Parameter	Cold Value	Warm Value	Value
1376	-----			
1377	<i>Tcold</i>			-25 °C
1378	<i>Twarm</i>			0 °C
1379	<i>sno1</i>	1.0 g m ⁻³	0.0 g m ⁻³	
1380	<i>dsno1</i>	4.0 g m ⁻³	1.0 g m ⁻³	
1381	<i>s exp1</i>	1.1	0.6	
1382	<i>S_{lim}</i>			0.8
1383	<i>X_{sml}</i>			0.97
1384	<i>S_{base}</i>			0.040 g m ⁻³
1385	<i>st exp1</i>			0.5
1386	<i>tns</i>			0.10 cm ⁻⁴
1387	<i>Tslopes</i> (cm ⁻⁴ C ⁻¹)			0.1842

1388
1389

FIGURE CAPTIONS

1390

1391 Figure 1. Characteristic sizes (inverse of the slope parameter) of precipitation ice particle
1392 distributions (inverse exponential) as a function of precipitation ice content and temperature for
1393 (a) snow in the original Rutledge and Hobbs (1983,1984)-based Goddard scheme, (b) graupel in
1394 the original Goddard scheme, (c) snow in the modified Goddard 3ICE scheme, (d) graupel in the
1395 modified Goddard 3ICE scheme, (e) snow in the new Goddard 4ICE scheme, (f) graupel in the
1396 new Goddard 4ICE scheme, and (g) snow density in the new Goddard 4ICE scheme. (a) – (d)
1397 adapted from L2011.

1398

1399 Figure 2. Horizontal cross sections of radar reflectivity for the 20 May 2011 MC3E case (a)
1400 observed by the NEXRAD Doppler radar network at 10:30 UTC over north-central Oklahoma
1401 (figure obtained from the National Mosaic and Next Generation Quantitative Precipitation
1402 Estimation) and (b) simulated using the new 4ICE scheme with smaller hail and bin rain
1403 evaporation correction (4iceb sml) at a simulation time of 85.5 h (13:30 UTC). The vertical east-
1404 west cross section of radar reflectivity shown in (c) was taken through the center of the domain
1405 from the same simulation and time as (b).

1406

1407 Figure 3. Time-height cross sections of maximum radar reflectivity for the 20 May 2011 MC3E
1408 case (a) observed by NEXRAD Doppler radar and simulated using the (b) original 3ICE, (c)
1409 level 1 improved 3ICE, (d) level 3 improved 3ICE, (e) new 4ICE with smaller hail, (f) new 4ICE
1410 with smaller hail and bin rain evaporation correction, (g) new 4ICE with moderate hail and bin
1411 rain evaporation correction, and (h) new 4ICE with larger hail and bin rain evaporation

1412 correction Goddard microphysics scheme. Right axes are heights in km, while horizontal dashed
1413 lines show the level of indicated environmental temperatures in degrees C. Times range from 00
1414 UTC 20 May to 00 UTC 21 May 2011.

1415

1416 Figure 4. Same as Fig. 2c except showing vertical cross sections of simulated (a) vertical
1417 velocities (b) graupel fall speeds, and (c) hail fall speeds.

1418

1419 Figure 5. Vertical profiles of maximum radar reflectivity for the 20 May 2011 MC3E case
1420 extracted between 06 and 12 UTC from Doppler radar observations and between 09 and 15 UTC
1421 from the three Goddard 3ICE simulations and four Goddard 4ICE simulations shown in Figure 2.

1422

1423 Figure 6. Radar reflectivity CFADs for the 20 May 2011 MC3E case constructed from (a)
1424 NEXRAD Doppler radar observations and simulations using the (b) original 3ICE, (c) level 1
1425 improved 3ICE, (d) level 3 improved 3ICE, (e) new 4ICE with smaller hail, (f) new 4ICE with
1426 smaller hail and bin rain evaporation correction, (g) new 4ICE with moderate hail and bin rain
1427 evaporation correction, and (h) new 4ICE with larger hail and bin rain evaporation correction

1428 Goddard microphysics scheme. Heavy thick lines in (b) - (h) show the edges of the core
1429 observed frequency probabilities [i.e., the 5 % contours shown in (a)] and the outer limits of the
1430 observed frequency distributions [i.e., the 0 % contours shown in (a)]. Right axes are heights in

1431 km, while horizontal dashed lines show the level of indicated environmental temperatures in
1432 degrees C.

1433

1434 Figure 7. Vertical profiles of PDF matching scores (i.e., the amount of overlap between the
1435 simulated and observed PDF at each level) for the 20 May 2011 MC3E simulations using the (b)
1436 original 3ICE, (c) level 1 improved 3ICE, (d) level 3 improved 3ICE, (e) new 4ICE with smaller
1437 hail, (f) new 4ICE with smaller hail and bin rain evaporation correction, (g) new 4ICE with
1438 moderate hail and bin rain evaporation correction, and (h) new 4ICE with larger hail and bin rain
1439 evaporation correction Goddard microphysics scheme.

1440

1441 Figure 8. Horizontal cross sections of radar reflectivity for the 23 February 1999 LBA case (a)
1442 observed by the S-Pol radar at 20:50 UTC over Amazonia overlaid with storm-relative winds
1443 from a dual-Doppler wind analysis and the track of the University of North Dakota Citation
1444 aircraft (figure adapted from http://radarmet.atmos.colostate.edu/lba_trmm/23feblba_cappi.html
1445 and Lang et al. 2007) and (b) simulated using the new 4ICE scheme with smaller hail and bin
1446 rain evaporation correction (4iceb sml) at a simulation time of 330 minutes (21:00 UTC). The
1447 vertical east-west cross section of radar reflectivity shown in (c) was taken through the center of
1448 the domain from the same simulation and time as (b). The solid rectangle and dashed box shown
1449 in (b) denote the north-south oriented rectangular patch of higher sensible/lower latent heat
1450 fluxes (Ji Parana) imposed to initiate convection and the analysis domain, respectively.

1451

1452 Figure 9. Time-height cross sections of maximum radar reflectivity for the 23 February 1999
1453 LBA case (a) observed by the S-pol ground-based radar and simulated using the (b) original
1454 3ICE, (c) level 1 improved 3ICE, (d) level 3 improved 3ICE, (e) new 4ICE with smaller hail, (f)
1455 new 4ICE with smaller hail and bin rain evaporation correction, (g) new 4ICE with moderate hail
1456 and bin rain evaporation correction, and (h) new 4ICE with larger hail and bin rain evaporation

1457 correction Goddard microphysics scheme. Right axes are heights in km, while horizontal dashed
1458 lines show the level of indicated environmental temperatures in degrees C. Model data were
1459 taken from a 64 km x 64 km subdomain. Black and gray labels at the bottom of (a) are the UTC
1460 and approximate matching times, respectively.

1461
1462 Figure 10. Same as Fig. 8c except showing vertical cross sections of simulated (a) vertical
1463 velocities (b) graupel fall speeds, and (c) hail fall speeds.

1464
1465 Figure 11. Vertical profiles of the maximum radar reflectivity for the 23 February 1999 LBA
1466 case extracted from the S-pol radar observations and the last 60 minutes of the three Goddard
1467 3ICE simulations and four Goddard 4ICE simulations shown in Figure 6. Model data were taken
1468 from a 64 km x 64 km subdomain.

1469
1470 Figure 12. Radar reflectivity CFADs for the 23 February 1999 LBA case constructed from (a) S-
1471 pol radar observations and the final 60 minutes of the simulations using the (b) original 3ICE, (c)
1472 level 1 improved 3ICE, (d) level 3 improved 3ICE, (e) new 4ICE with smaller hail, (f) new 4ICE
1473 with smaller hail and bin rain evaporation correction, (g) new 4ICE with moderate hail and bin
1474 rain evaporation correction, and (h) new 4ICE with larger hail and bin rain evaporation
1475 correction Goddard microphysics scheme. Heavy thick lines in (b) - (h) show the edges of the
1476 core observed frequency probabilities [i.e., the 5 % contours shown in (a)] and the outer limits of
1477 the observed frequency distributions [i.e., the 0 % contours shown in (a)]. Right axes are heights
1478 in km, while horizontal dashed lines show the level of indicated environmental temperatures in
1479 degrees C. Model data were taken from a 64 km x 64 km subdomain.

1480

1481 Figure 13. Vertical profiles of PDF matching scores for the 23 February 1999 LBA simulations
1482 over the final 60 minutes using the (b) original 3ICE, (c) level 1 improved 3ICE, (d) level 3
1483 improved 3ICE, (e) new 4ICE with smaller hail, (f) new 4ICE with smaller hail and bin rain
1484 evaporation correction, (g) new 4ICE with moderate hail and bin rain evaporation correction, and
1485 (h) new 4ICE with larger hail and bin rain evaporation correction Goddard microphysics scheme.
1486 Model data were taken from a 64 km x 64 km subdomain.

1487

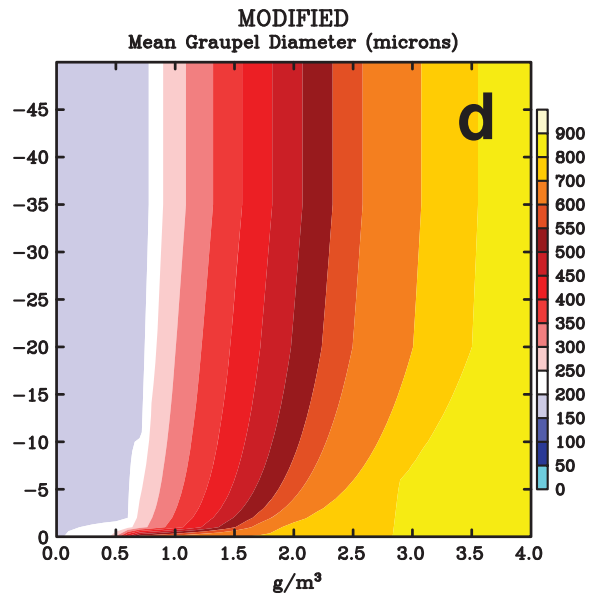
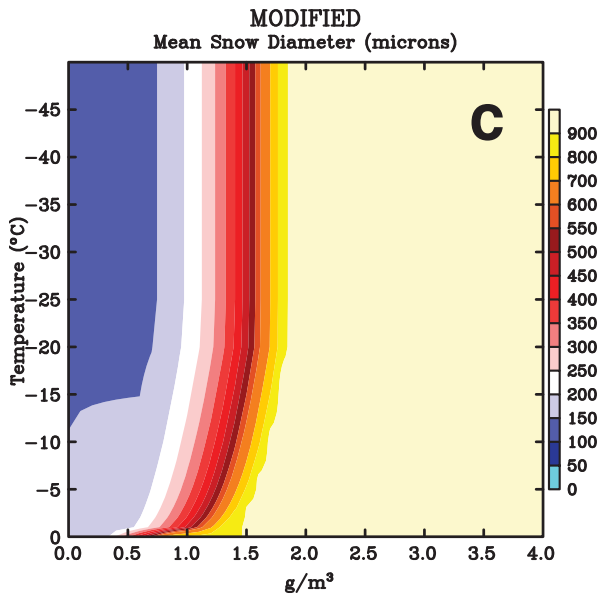
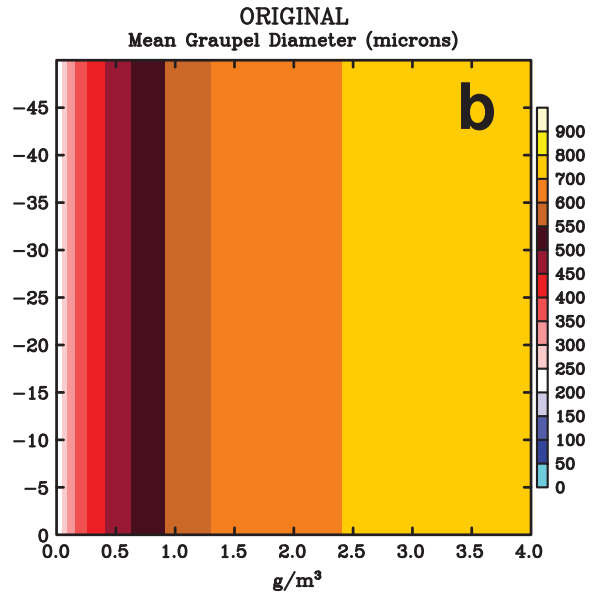
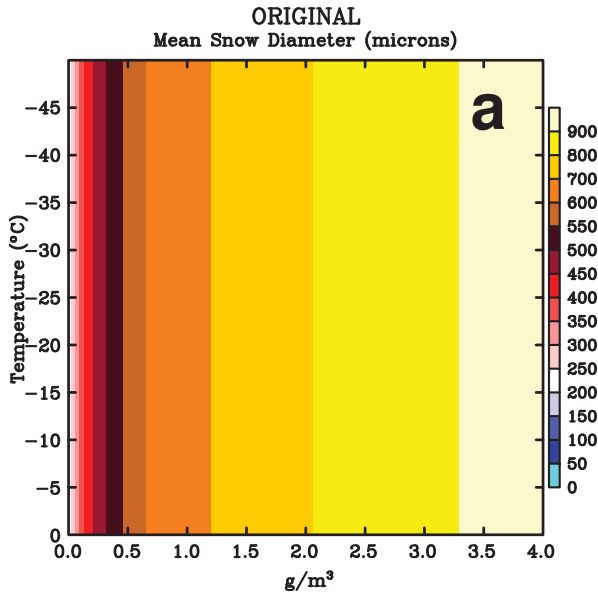
1488 Figure 14. Instantaneous surface rainfall rates corresponding to the horizontal radar reflectivity
1489 cross sections shown for the 20 May MC3E case in Fig. 2b (a) and the 23 February LBA case in
1490 Fig. 8b (b). (c) surface rainfall histograms observed by the Doppler radar network around the
1491 MC3E sounding array from 06-12 UTC and simulated with Goddard microphysics from 09-15
1492 UTC for the 20 May MC3E case. (d) surface rainfall histograms derived from ground-based
1493 radar observations collected from 2002-2130 UTC and simulated over the final 60 minutes of
1494 simulation time over a 64 km x 64 km subdomain (shown by the dashed square in panel b) for
1495 the 23 February 1999 case using the Goddard microphysics schemes.

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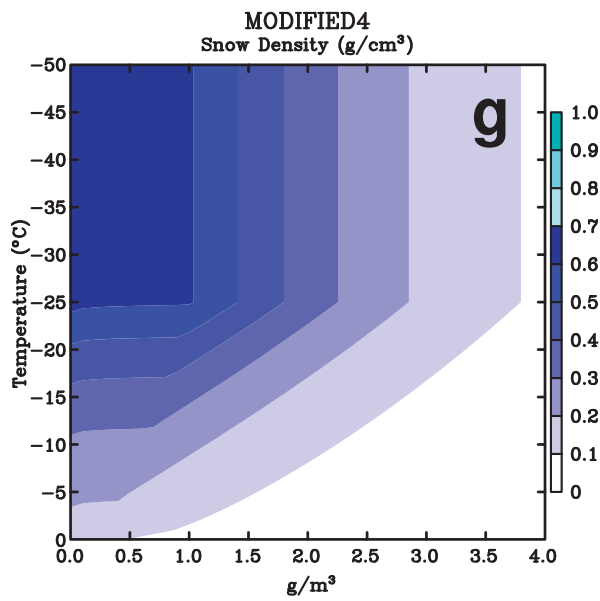
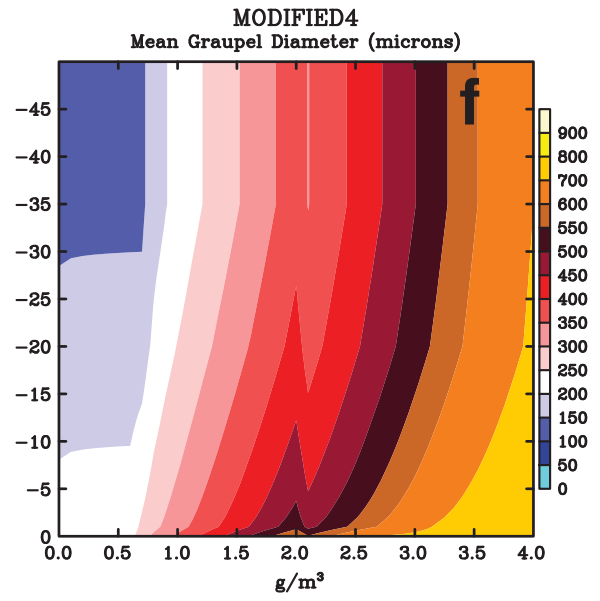
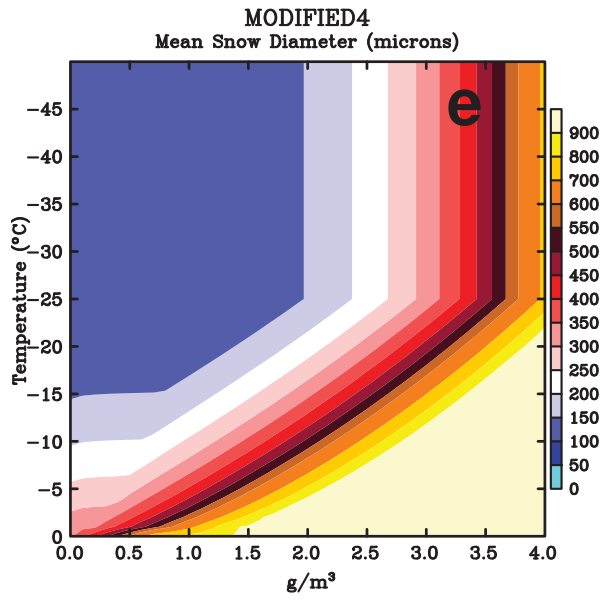
1497 Figure 15. Distribution of surface cold pool intensities for the 20 May MC3E case for the
1498 smaller hail runs with and without the bin rain evaporation correction, 4iceb sml and 4ice sml,
1499 respectively. Intensities are shown in terms of the surface potential temperature deviations (K)
1500 over the 09 to 15 UTC analysis period for regions where the lowest level rain mixing ratio
1501 exceeds 0.1 g m^{-3} .

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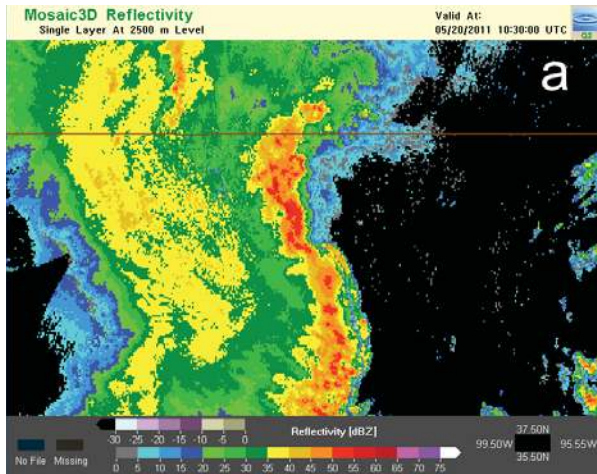
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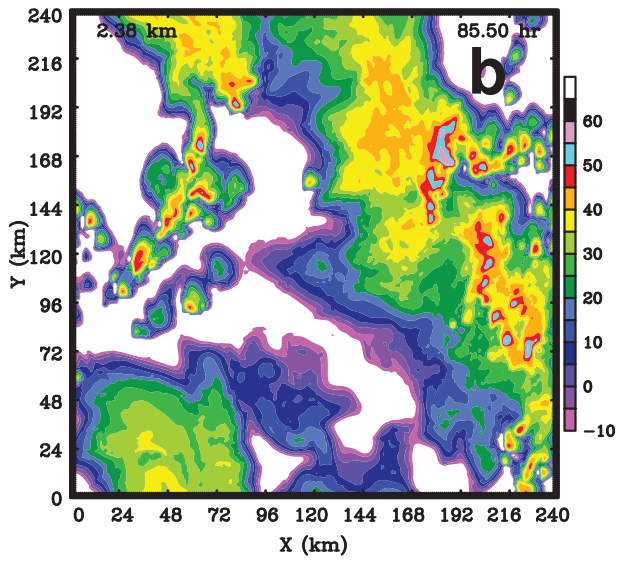


1546 Figure 1. Characteristic sizes (inverse of the slope parameter) of precipitation ice particle
1547 distributions (inverse exponential) as a function of precipitation ice content and temperature for
1548 (a) snow in the original Rutledge and Hobbs (1983,1984)-based Goddard scheme, (b) graupel in
1549 the original Goddard scheme, (c) snow in the modified Goddard 3ICE scheme, (d) graupel in the
1550 modified Goddard 3ICE scheme, (e) snow in the new Goddard 4ICE scheme, (f) graupel in the
1551 new Goddard 4ICE scheme, and (g) snow density in the new Goddard 4ICE scheme. (a) – (d)
1552 adapted from L2011.



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MC3E 20 May 2011 4iceb sml
Reflectivity (dBZ)



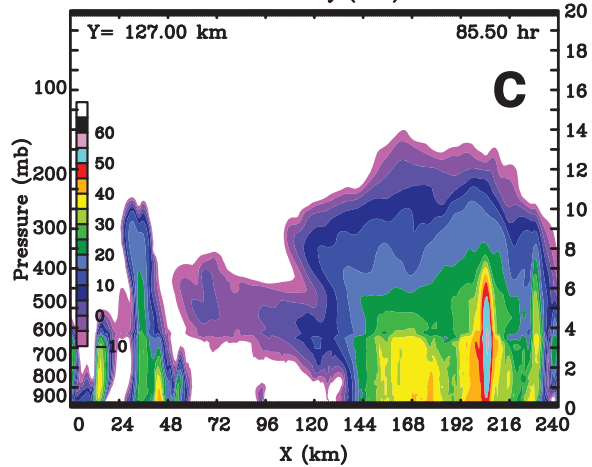
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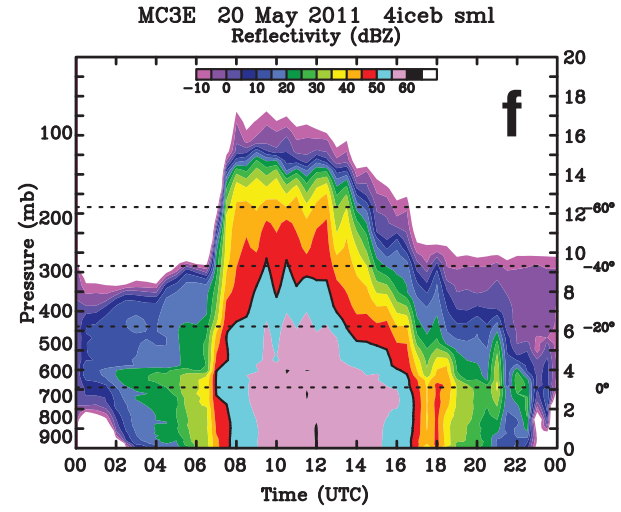
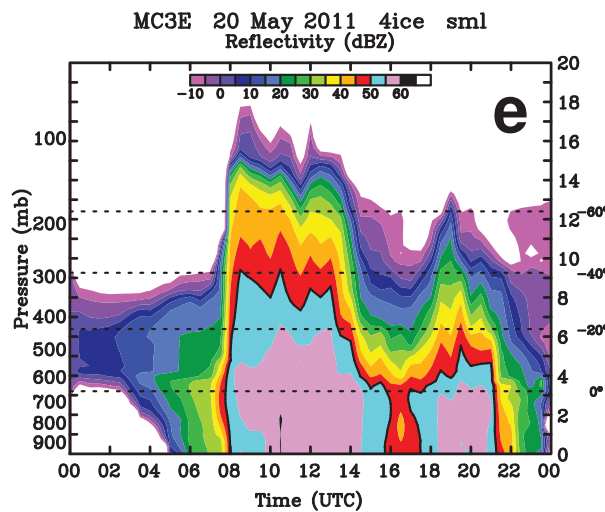
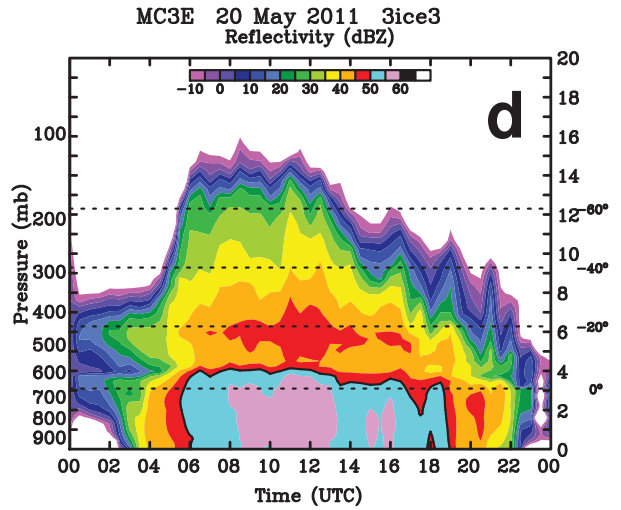
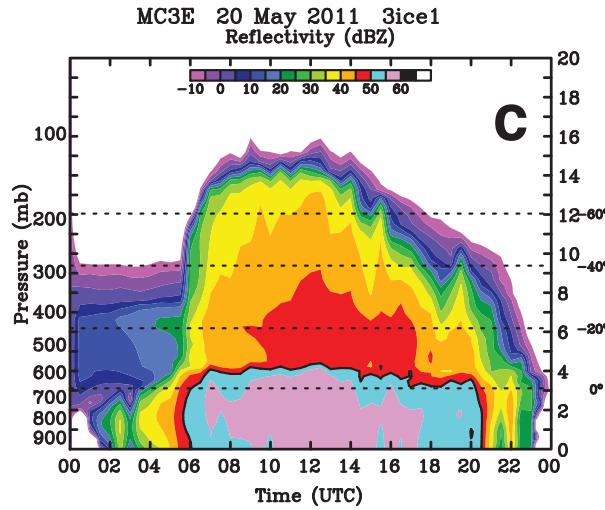
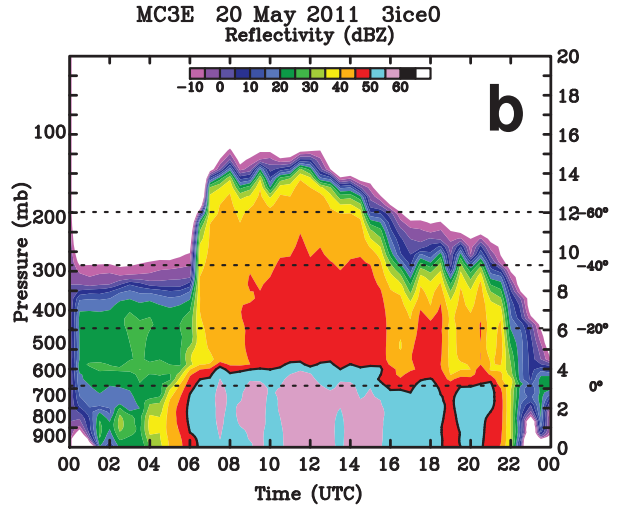
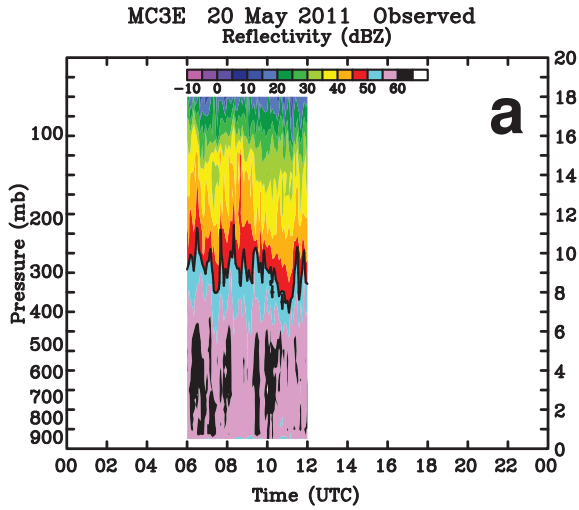
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MC3E 20 May 2011 4iceb sml
Reflectivity (dBZ)



1559 Figure 2. Horizontal cross sections of radar reflectivity for the 20 May 2011 MC3E case (a)
1560 observed by the NEXRAD Doppler radar network at 10:30 UTC over north-central Oklahoma
1561 (figure obtained from the National Mosaic and Next Generation Quantitative Precipitation
1562 Estimation) and (b) simulated using the new 4ICE scheme with smaller hail and bin rain
1563 evaporation correction (4iceb sml) at a simulation time of 85.5 h (13:30 UTC). The vertical east-
1564 west cross section of radar reflectivity shown in (c) was taken through the center of the domain
1565 from the same simulation and time as (b).
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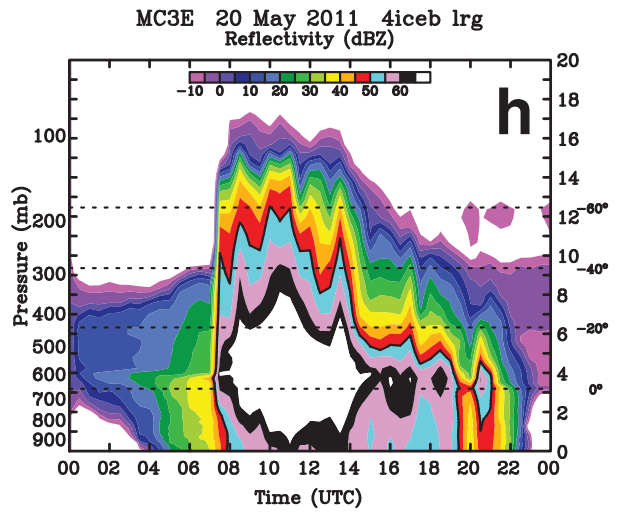
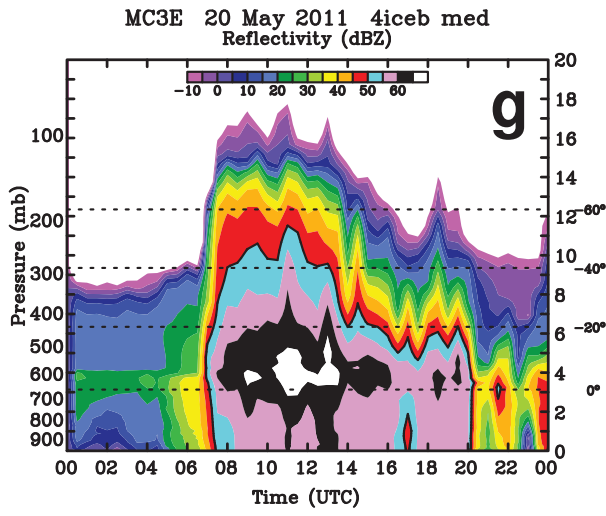


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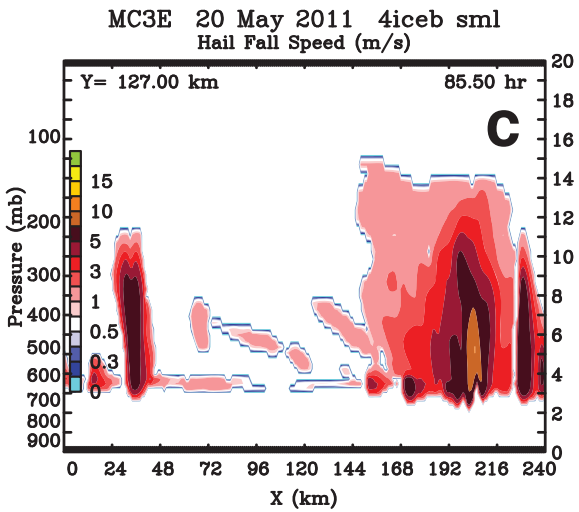
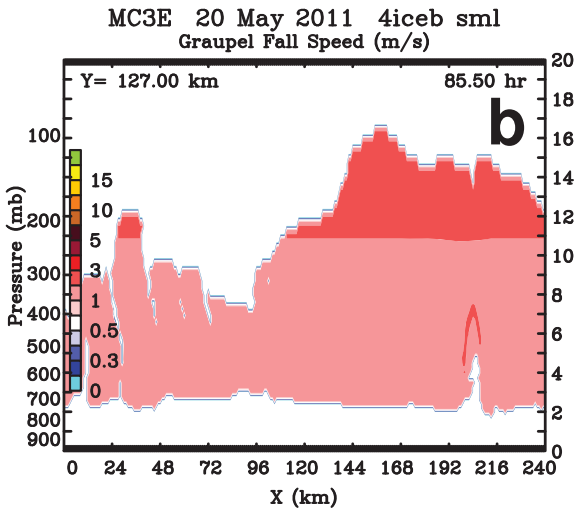
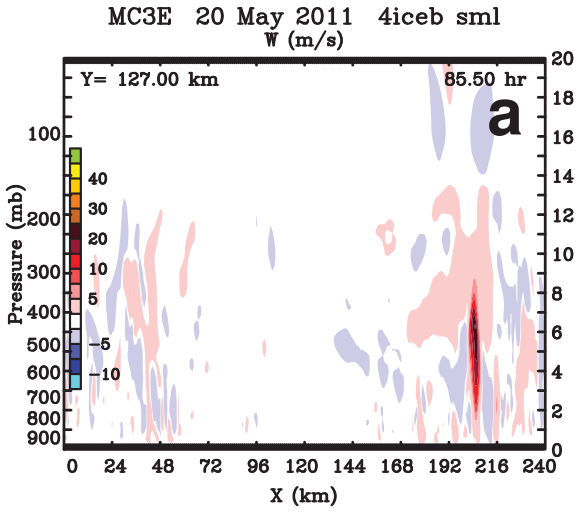
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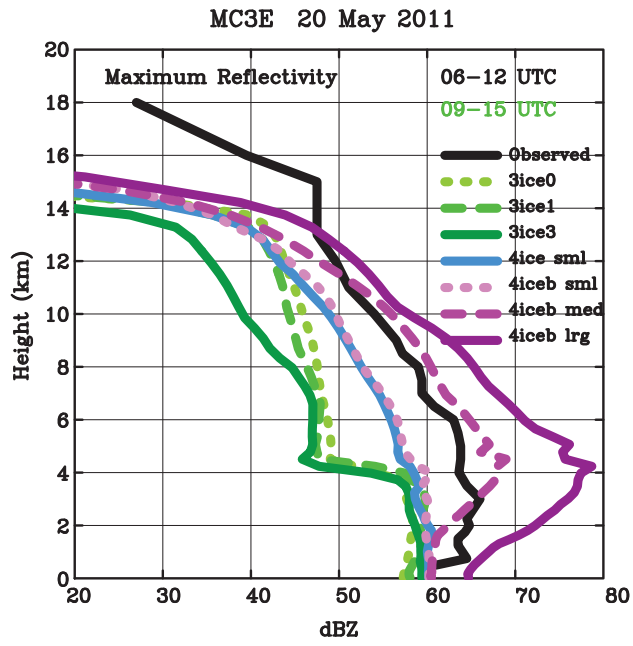
1576 Figure 3. Time-height cross sections of maximum radar reflectivity for the 20 May 2011 MC3E
1577 case (a) observed by NEXRAD Doppler radar and simulated using the (b) original 3ICE, (c)
1578 level 1 improved 3ICE, (d) level 3 improved 3ICE, (e) new 4ICE with smaller hail, (f) new 4ICE
1579 with smaller hail and bin rain evaporation correction, (g) new 4ICE with moderate hail and bin
1580 rain evaporation correction, and (h) new 4ICE with larger hail and bin rain evaporation
1581 correction Goddard microphysics scheme. Right axes are heights in km, while horizontal dashed
1582 lines show the level of indicated environmental temperatures in degrees C. Times range from 00
1583 UTC 20 May to 00 UTC 21 May 2011.
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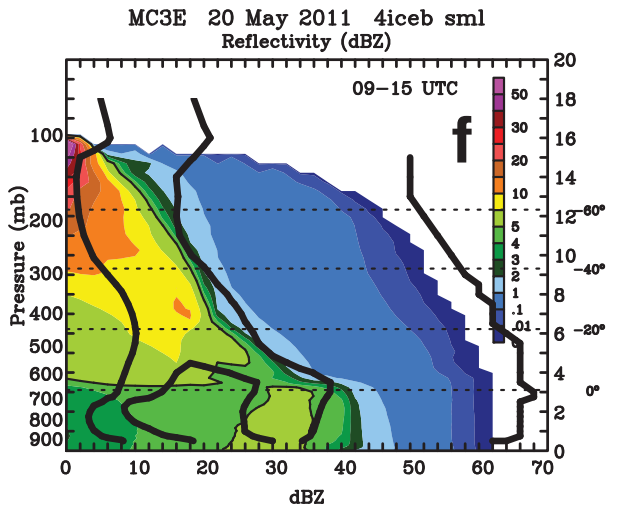
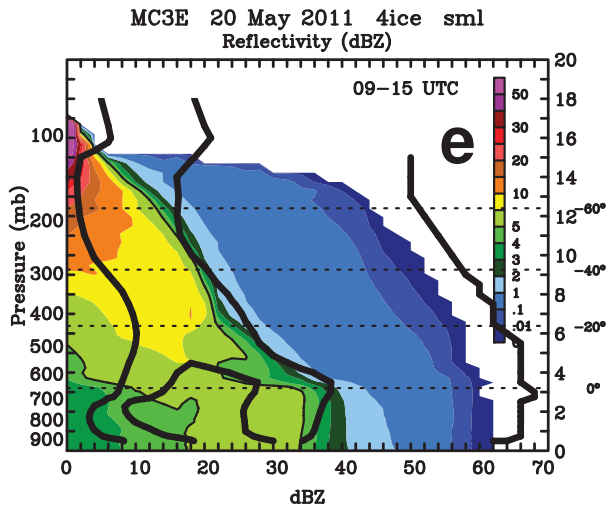
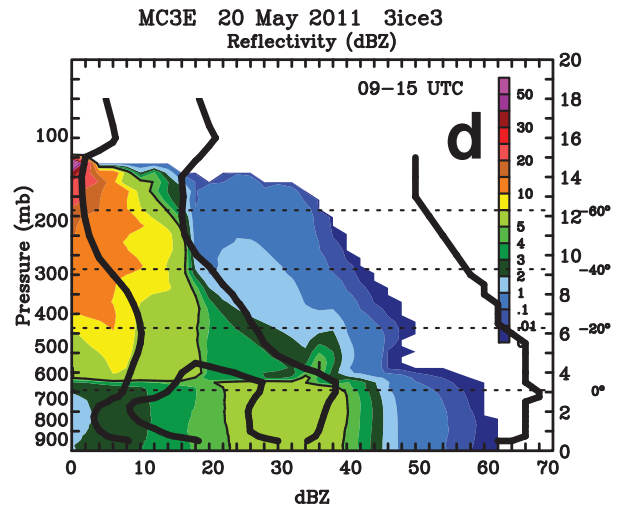
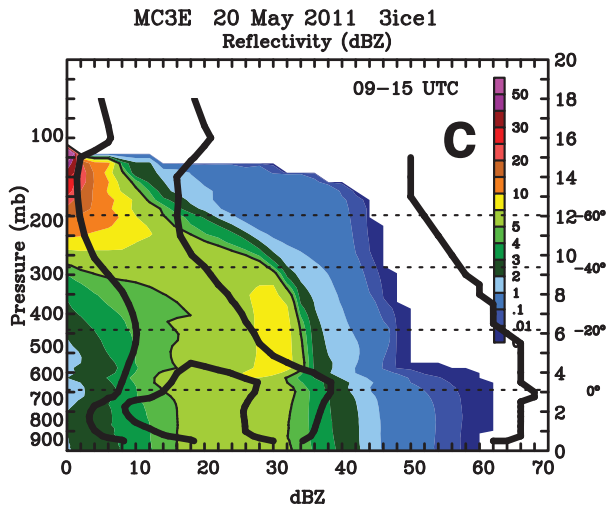
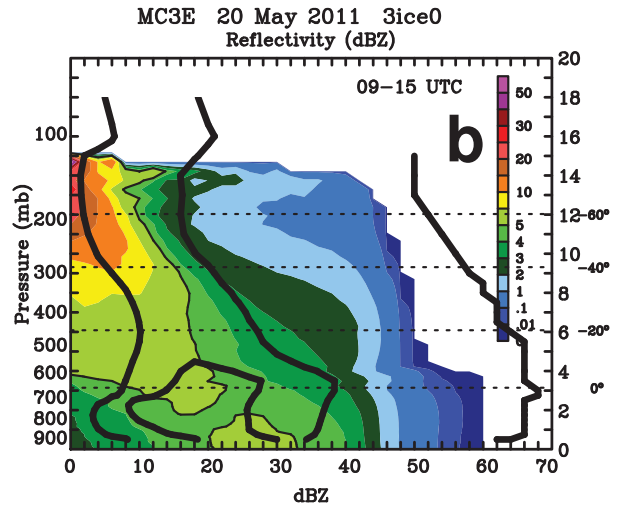
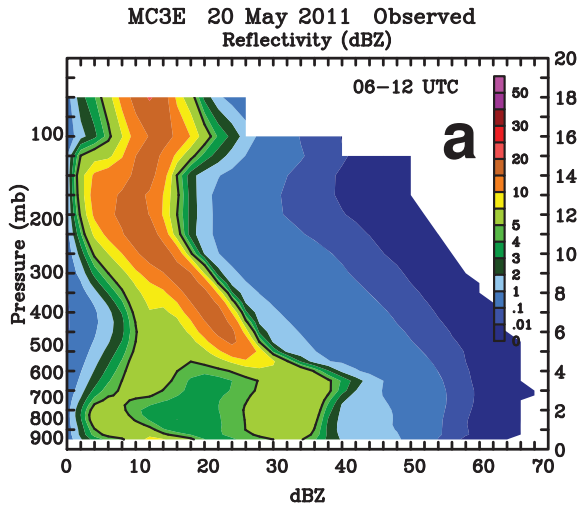
1591
1592 Figure 4. Same as Fig. 2c except showing vertical cross sections of simulated (a) vertical
1593 velocities (b) graupel fall speeds, and (c) hail fall speeds.
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1599 Figure 5. Vertical profiles of maximum radar reflectivity for the 20 May 2011 MC3E case
1600 extracted between 06 and 12 UTC from Doppler radar observations and between 09 and 15 UTC
1601 from the three Goddard 3ICE simulations and four Goddard 4ICE simulations shown in Figure 2.
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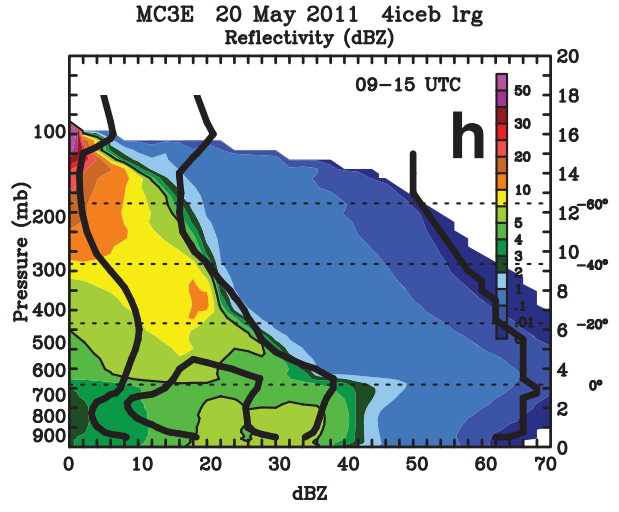
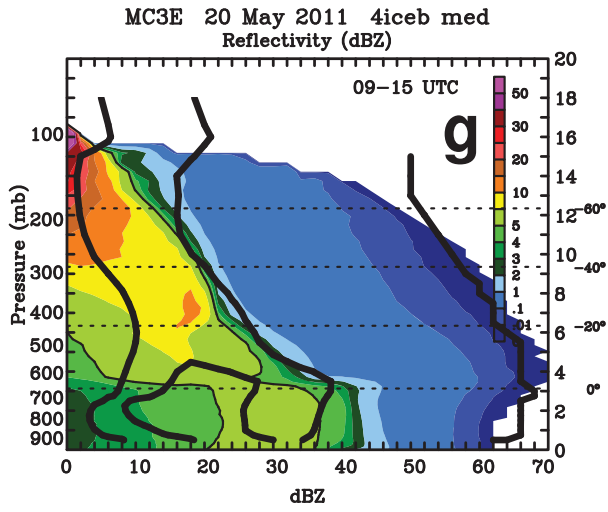


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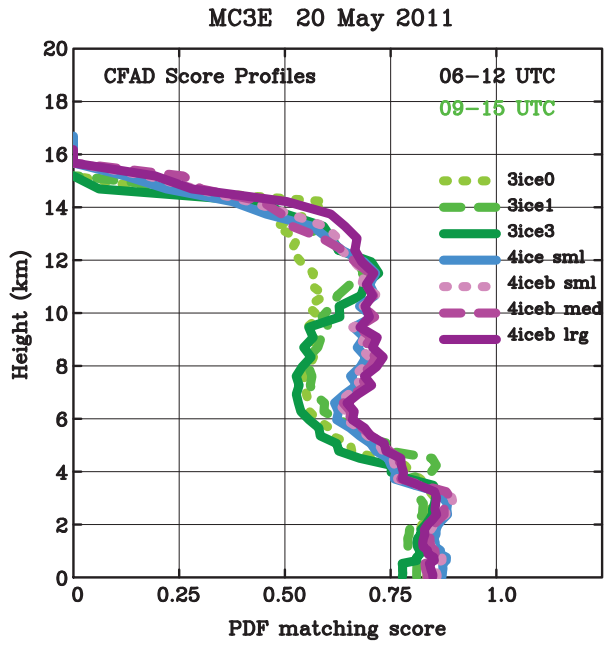
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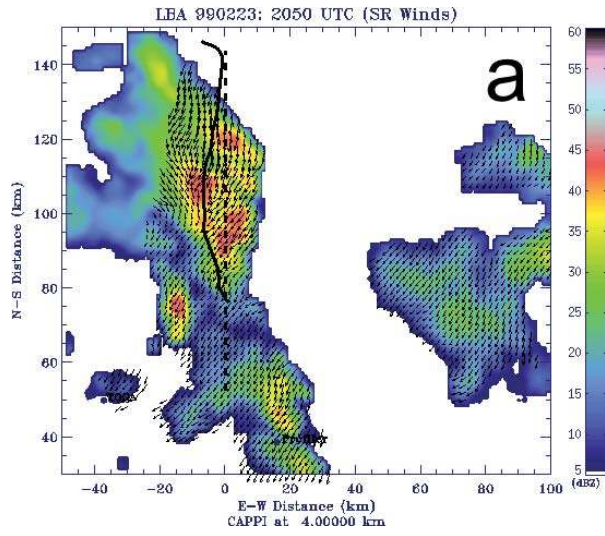
1613 Figure 6. Radar reflectivity CFADs for the 20 May 2011 MC3E case constructed from (a)
1614 NEXRAD Doppler radar observations and simulations using the (b) original 3ICE, (c) level 1
1615 improved 3ICE, (d) level 3 improved 3ICE, (e) new 4ICE with smaller hail, (f) new 4ICE with
1616 smaller hail and bin rain evaporation correction, (g) new 4ICE with moderate hail and bin rain
1617 evaporation correction, and (h) new 4ICE with larger hail and bin rain evaporation correction
1618 Goddard microphysics scheme. Heavy thick lines in (b) - (h) show the edges of the core
1619 observed frequency probabilities [i.e., the 5 % contours shown in (a)] and the outer limits of the
1620 observed frequency distributions [i.e., the 0 % contours shown in (a)]. Right axes are heights in
1621 km, while horizontal dashed lines show the level of indicated environmental temperatures in
1622 degrees C.

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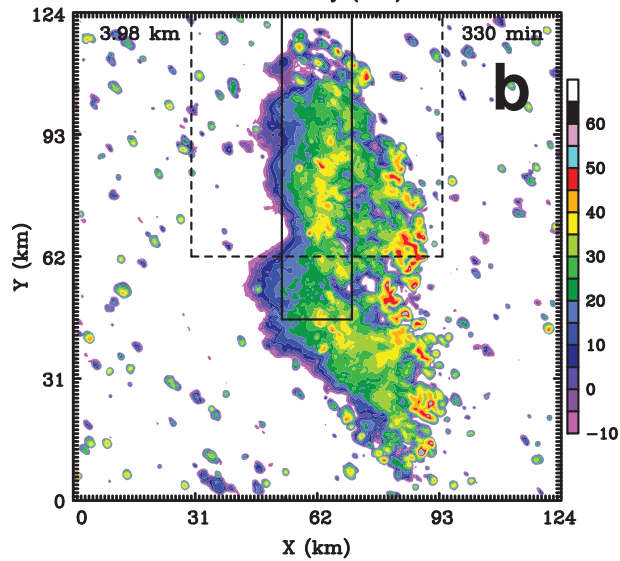
1625 Figure 7. Vertical profiles of PDF matching scores (i.e., the amount of overlap between the
1626 simulated and observed PDF at each level) for the 20 May 2011 MC3E simulations using the (b)
1627 original 3ICE, (c) level 1 improved 3ICE, (d) level 3 improved 3ICE, (e) new 4ICE with smaller
1628 hail, (f) new 4ICE with smaller hail and bin rain evaporation correction, (g) new 4ICE with
1629 moderate hail and bin rain evaporation correction, and (h) new 4ICE with larger hail and bin rain
1630 evaporation correction Goddard microphysics scheme.

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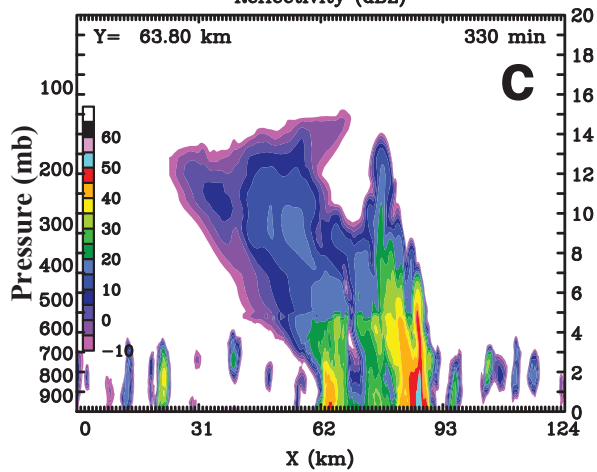
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LBA Feb 23, 1999 4iceb sml
Reflectivity (dBZ)



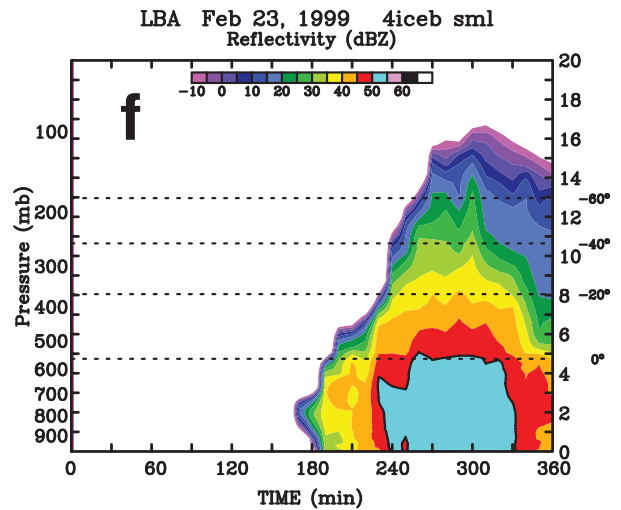
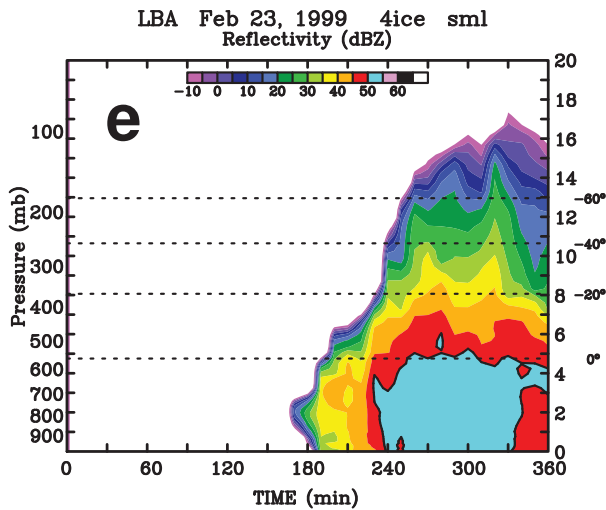
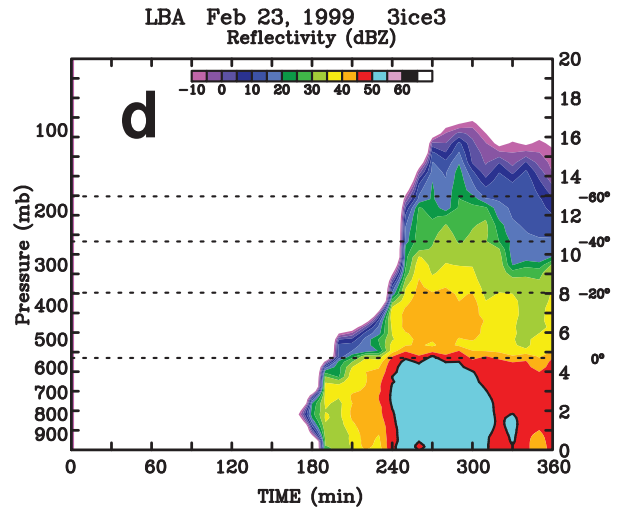
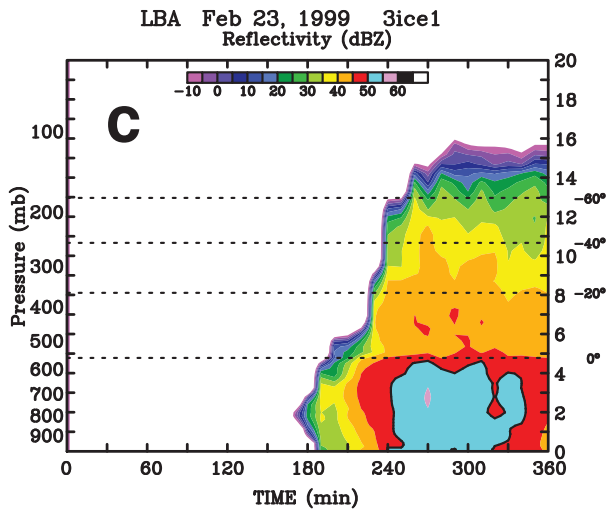
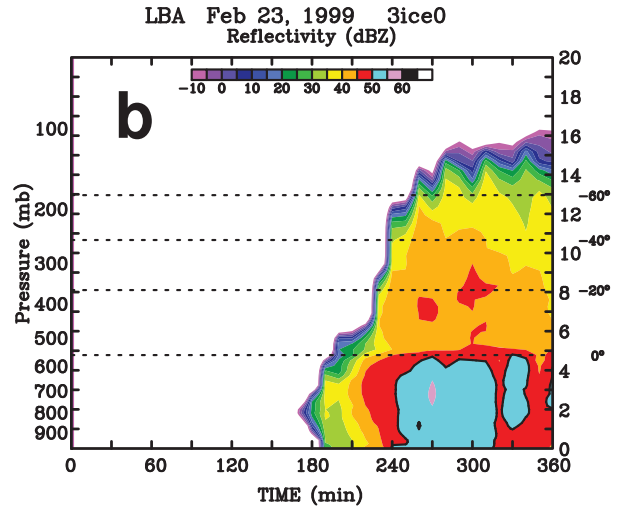
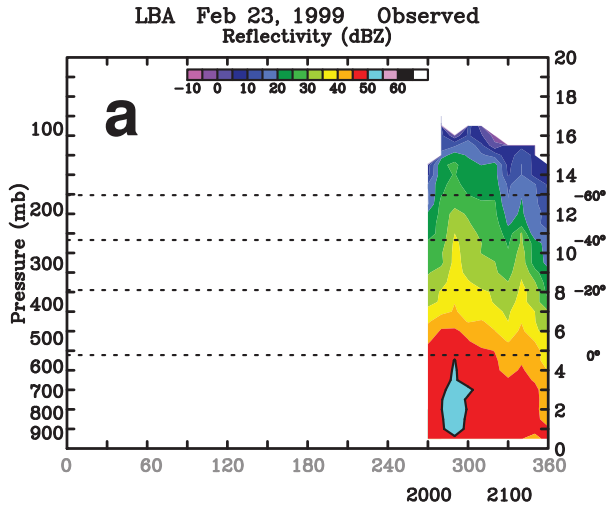
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LBA Feb 23, 1999 4iceb sml
Reflectivity (dBZ)

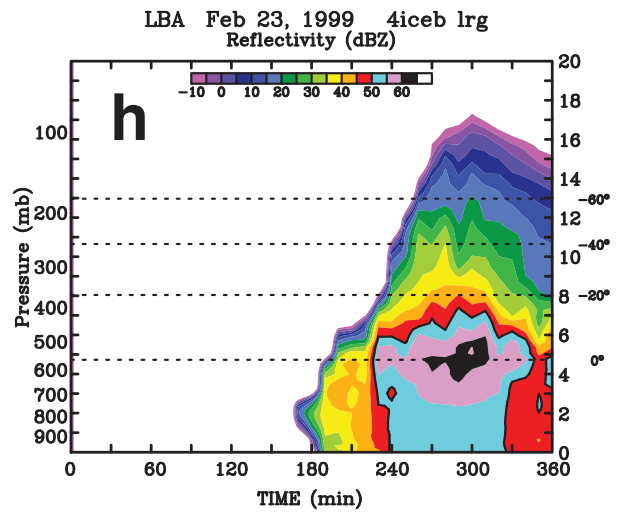
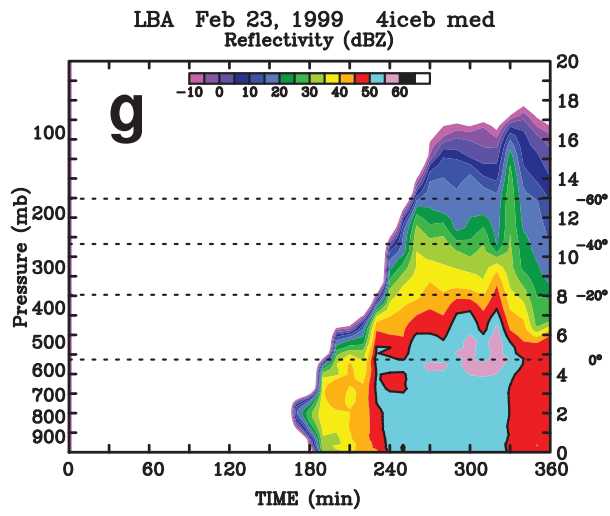


1637 Figure 8. Horizontal cross sections of radar reflectivity for the 23 February 1999 LBA case (a)
1638 observed by the S-Pol radar at 20:50 UTC over Amazonia overlaid with storm-relative winds
1639 from a dual-Doppler wind analysis and the track of the University of North Dakota Citation
1640 aircraft (figure adapted from http://radarmet.atmos.colostate.edu/lba_trmm/23feblba_cappi.html
1641 and Lang et al. 2007) and (b) simulated using the new 4ICE scheme with smaller hail and bin
1642 rain evaporation correction (4iceb sml) at a simulation time of 330 minutes (21:00 UTC). The
1643 vertical east-west cross section of radar reflectivity shown in (c) was taken through the center of
1644 the domain from the same simulation and time as (b). The solid rectangle and dashed box shown
1645 in (b) denote the north-south oriented rectangular patch of higher sensible/lower latent heat
1646 fluxes (Ji Parana) imposed to initiate convection and the analysis domain, respectively.
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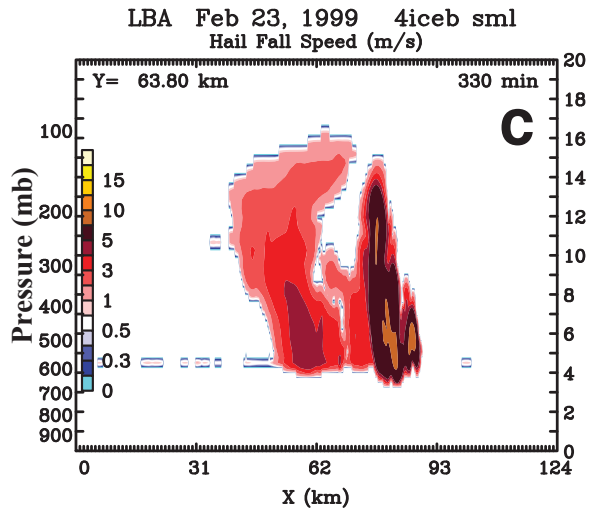
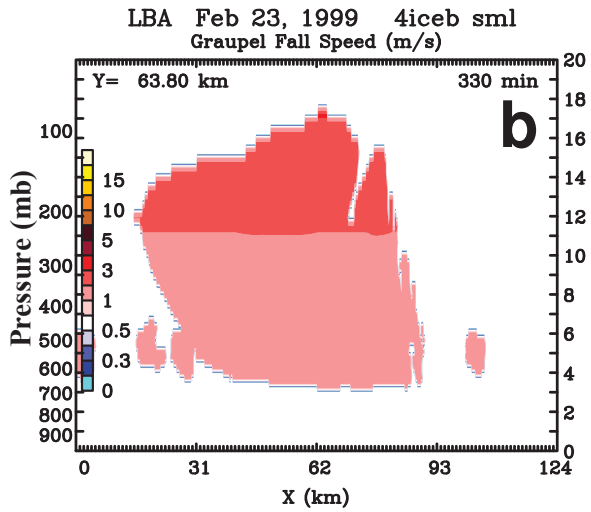
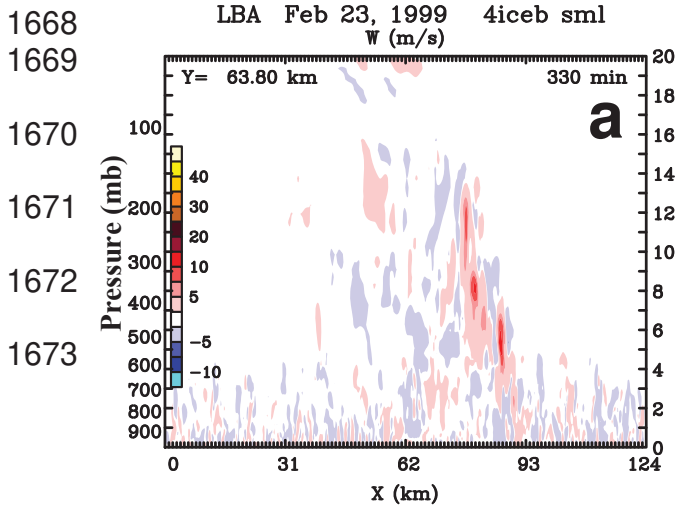
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1658 Figure 9. Time-height cross sections of maximum radar reflectivity for the 23 February 1999
1659 LBA case (a) observed by the S-pol ground-based radar and simulated using the (b) original
1660 3ICE, (c) level 1 improved 3ICE, (d) level 3 improved 3ICE, (e) new 4ICE with smaller hail, (f)
1661 new 4ICE with smaller hail and bin rain evaporation correction, (g) new 4ICE with moderate hail
1662 and bin rain evaporation correction, and (h) new 4ICE with larger hail and bin rain evaporation
1663 correction Goddard microphysics scheme. Right axes are heights in km, while horizontal dashed
1664 lines show the level of indicated environmental temperatures in degrees C. Model data were
1665 taken from a 64 km x 64 km subdomain. Black and gray labels at the bottom of (a) are the UTC
1666 and approximate matching times, respectively.
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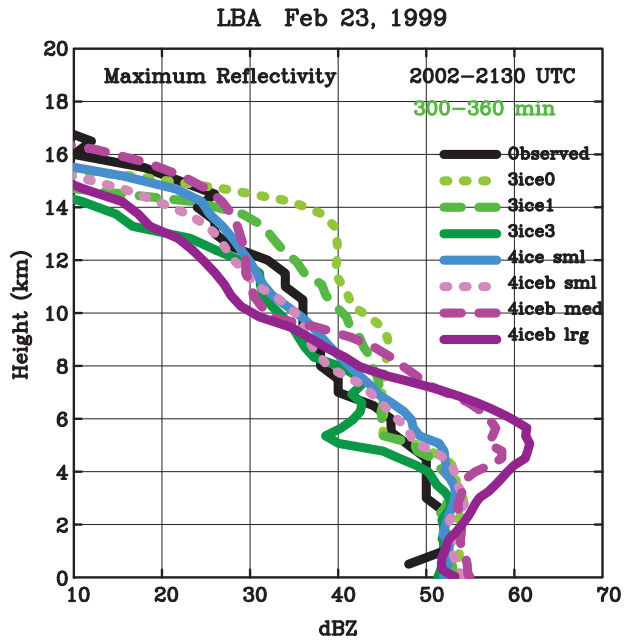


1674 Figure 10. Same as Fig. 8c except showing vertical cross sections of simulated (a) vertical
1675 velocities (b) graupel fall speeds, and (c) hail fall speeds.

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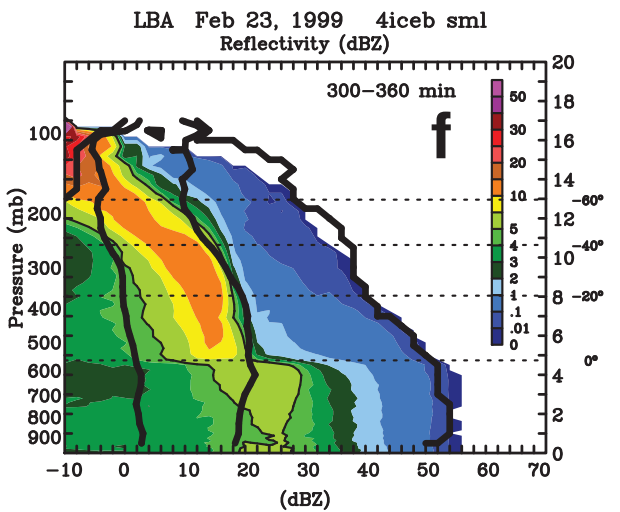
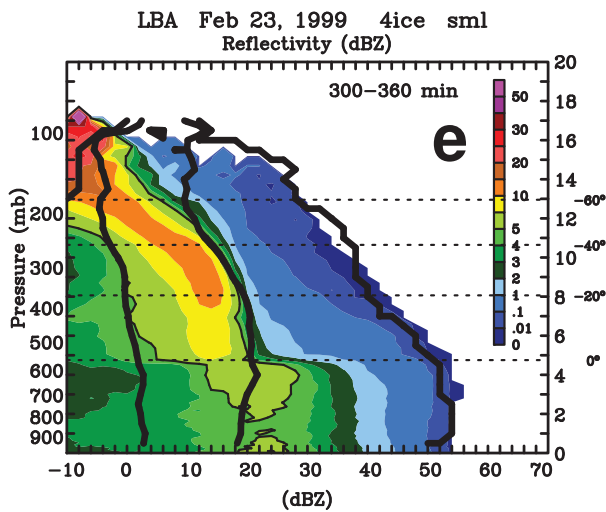
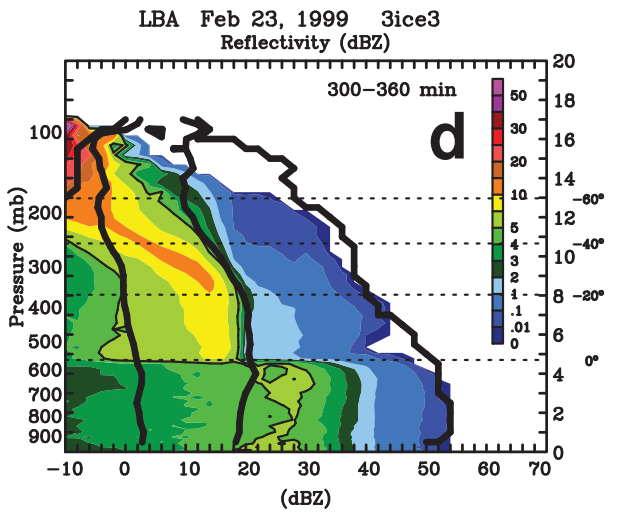
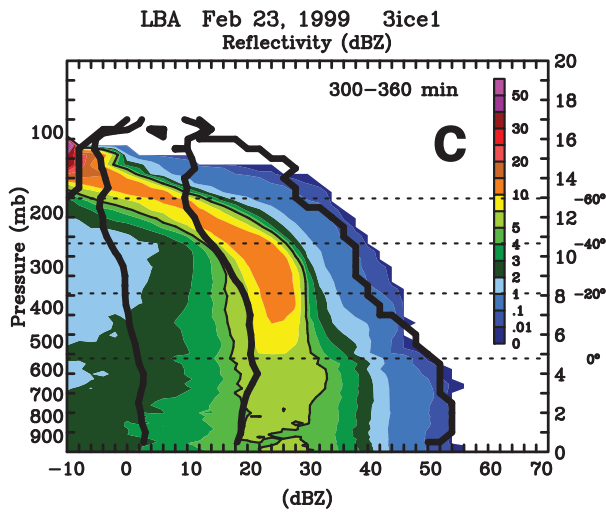
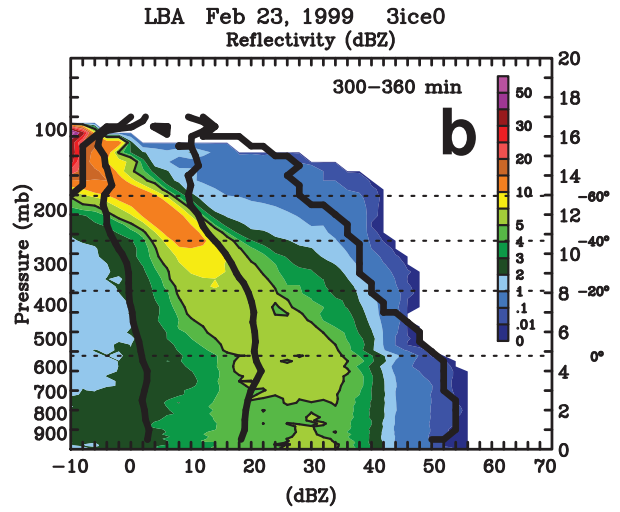
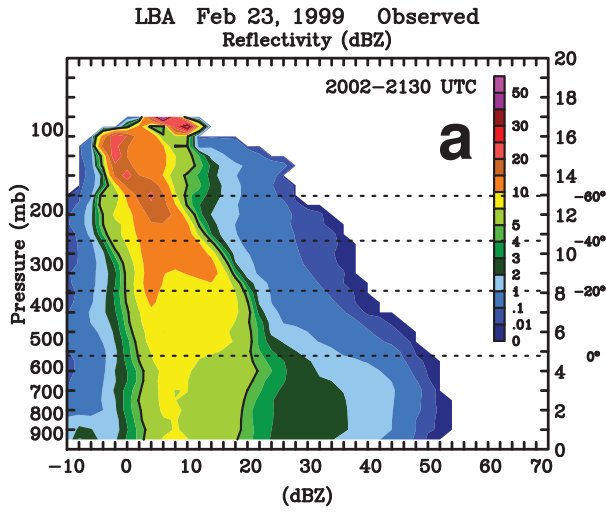
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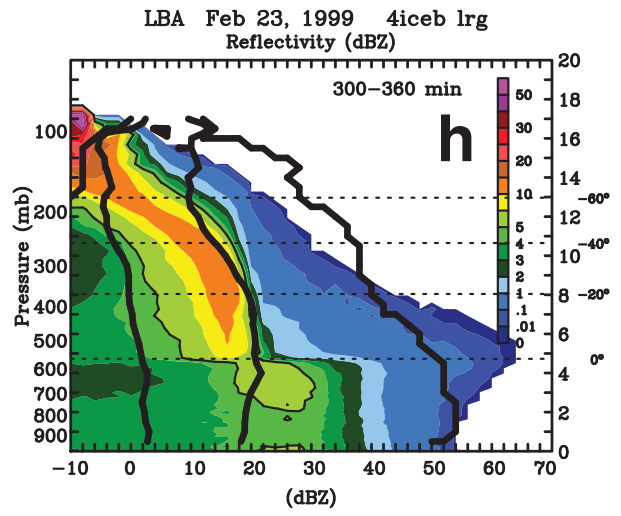
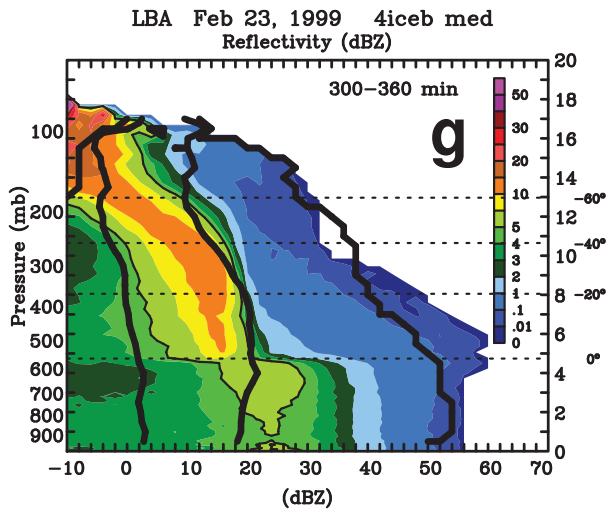


1681 Figure 11. Vertical profiles of the maximum radar reflectivity for the 23 February 1999 LBA
1682 case extracted from the S-pol radar observations and the last 60 minutes of the three Goddard
1683 3ICE simulations and four Goddard 4ICE simulations shown in Figure 6. Model data were taken
1684 from a 64 km x 64 km subdomain.
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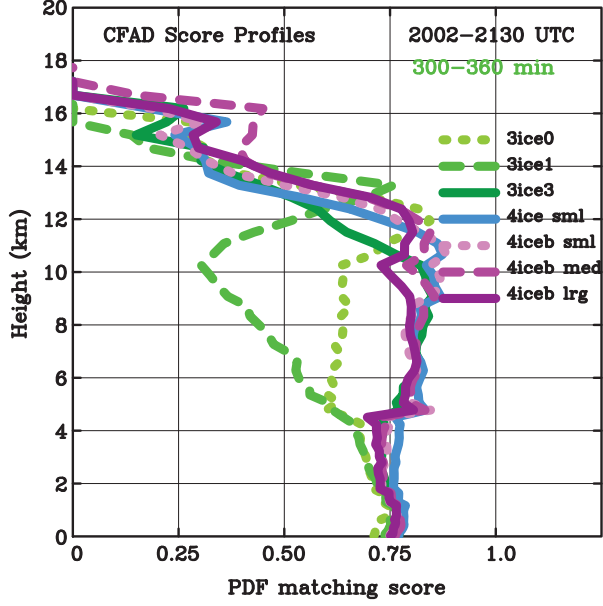
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1734 Figure 12. Radar reflectivity CFADs for the 23 February 1999 LBA case constructed from (a) S-
1735 pol radar observations and the final 60 minutes of the simulations using the (b) original 3ICE, (c)
1736 level 1 improved 3ICE, (d) level 3 improved 3ICE, (e) new 4ICE with smaller hail, (f) new 4ICE
1737 with smaller hail and bin rain evaporation correction, (g) new 4ICE with moderate hail and bin
1738 rain evaporation correction, and (h) new 4ICE with larger hail and bin rain evaporation
1739 correction Goddard microphysics scheme. Heavy thick lines in (b) - (h) show the edges of the
1740 core observed frequency probabilities [i.e., the 5 % contours shown in (a)] and the outer limits of
1741 the observed frequency distributions [i.e., the 0 % contours shown in (a)]. Right axes are heights
1742 in km, while horizontal dashed lines show the level of indicated environmental temperatures in
1743 degrees C. Model data were taken from a 64 km x 64 km subdomain.
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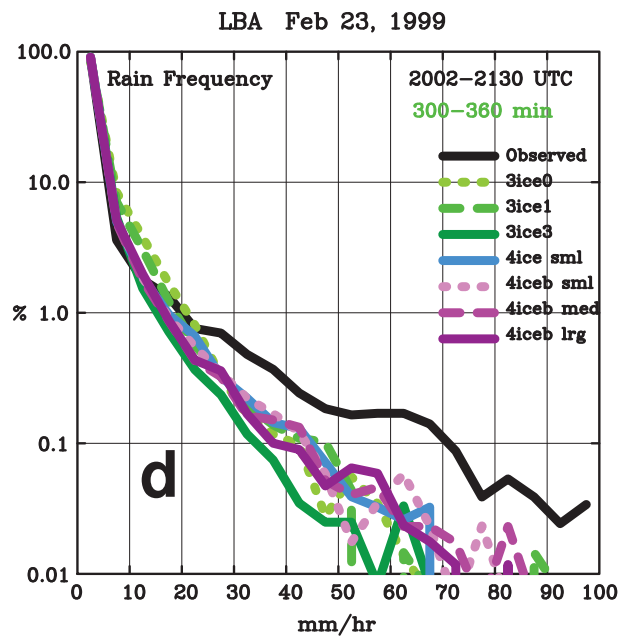
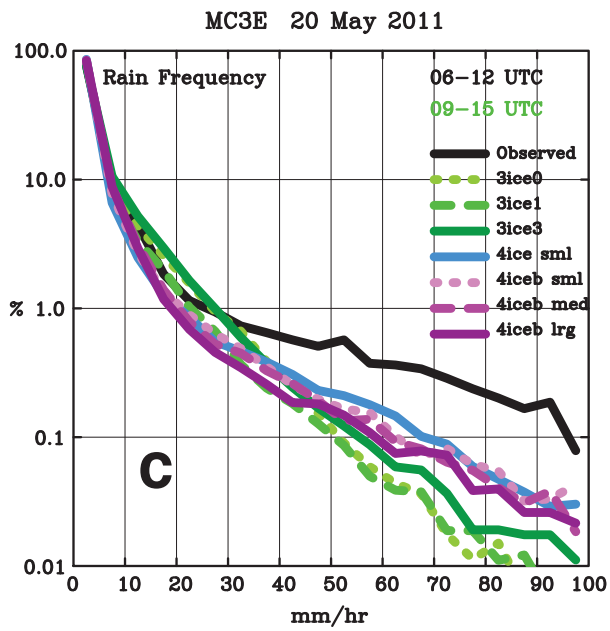
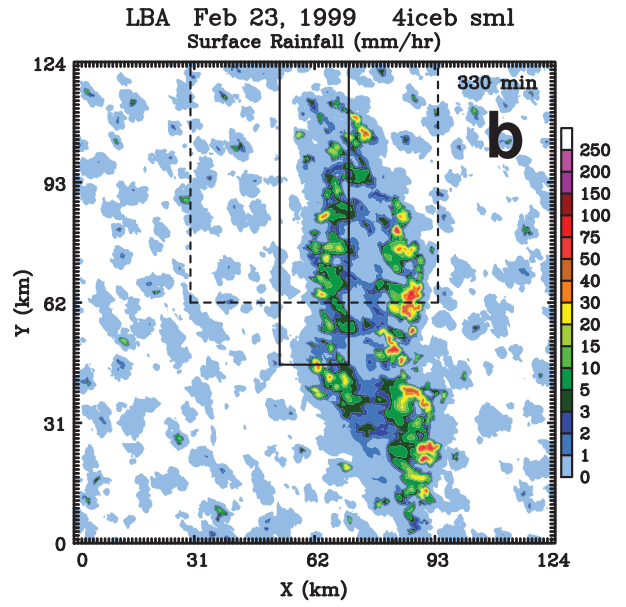
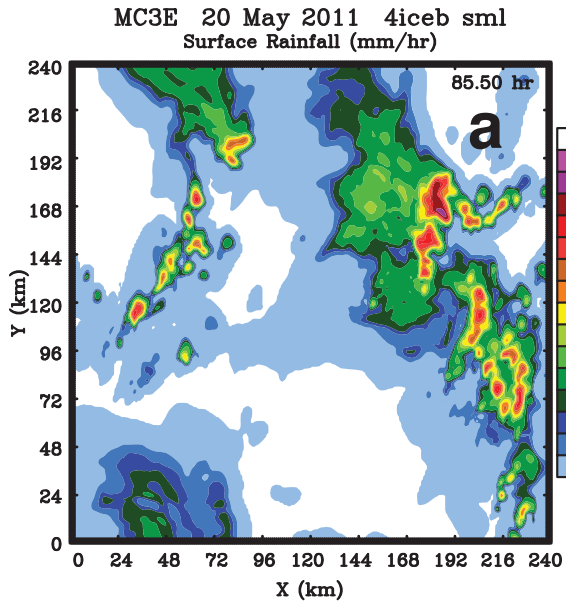
LBA Feb 23, 1999



1747 Figure 13. Vertical profiles of PDF matching scores for the 23 February 1999 LBA simulations
1748 over the final 60 minutes using the (b) original 3ICE, (c) level 1 improved 3ICE, (d) level 3
1749 improved 3ICE, (e) new 4ICE with smaller hail, (f) new 4ICE with smaller hail and bin rain
1750 evaporation correction, (g) new 4ICE with moderate hail and bin rain evaporation correction, and
1751 (h) new 4ICE with larger hail and bin rain evaporation correction Goddard microphysics scheme.
1752 Model data were taken from a 64 km x 64 km subdomain.

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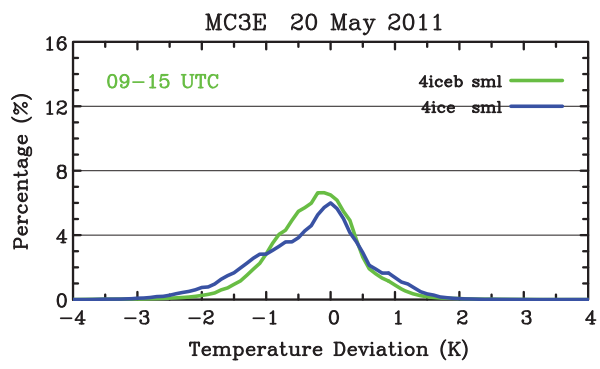
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1756 Figure 14. Instantaneous surface rainfall rates corresponding to the horizontal radar reflectivity
1757 cross sections shown for the 20 May MC3E case in Fig. 2b (a) and the 23 February LBA case in
1758 Fig. 8b (b). (c) surface rainfall histograms observed by the Doppler radar network around the
1759 MC3E sounding array from 06-12 UTC and simulated with Goddard microphysics from 09-15
1760 UTC for the 20 May MC3E case. (d) surface rainfall histograms derived from ground-based
1761 radar observations collected from 2002-2130 UTC and simulated over the final 60 minutes of
1762 simulation time over a 64 km x 64 km subdomain (shown by the dashed square in panel b) for
1763 the 23 February 1999 case using the Goddard microphysics schemes.

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1767 Figure 15. Distribution of surface cold pool intensities for the 20 May MC3E case for the
1768 smaller hail runs with and without the bin rain evaporation correction, 4iceb sml and 4ice sml,
1769 respectively. Intensities are shown in terms of the surface potential temperature deviations (K)
1770 over the 09 to 15 UTC analysis period for regions where the lowest level rain mixing ratio
1771 exceeds 0.1 g m^{-3} .
