Response to interactive comment on "Use of an observation-based aerosol profile in simulations of a mid-latitude squall line during MC3E: Similarity of stratiform ice microphysics to tropical conditions" by Ann M. Fridlind et al. by Anonymous Referee #1

General comments

This article constructs hygroscopic aerosol size distribution profiles from MC3E aircraft and ground-based data over six days. These profiles are used to run 4 NU-WRF simulations of a squall line case study. Observed and simulated cloud ice microphysical properties in the stratiform outflow region are then compared. The work is very comprehensive and cites the existing literature thoroughly. The results about similarity between continental and tropical ice microphysics are quite interesting.

We very much appreciate the helpful questions and comments. Point-by-point responses below have greatly improved the manuscript by reducing figures, adding section numbers, and making corrections and clarifications throughout.

Although factors like "fall speeds, aggregation and vapor growth rates, [etc]" are listed in the results, I would have appreciated more discussion on how the modeled ice microphysics might be improved to bring something like the number and mass size distributions into better agreement with observations.

Clarification added to Section 5: "The NU-WRF biases relative to observations shown here are consistent with the hypothesis that microphysics schemes are missing a key aspect of an updraft microphysics pathway that can largely determine outflow size, most likely associated with warm-temperature ice multiplication (e.g., Ackerman et al., 2015; Lawson et al., 2015; Ladino et al., 2017). Here we show that NU-WRF biases in stratiform ice mass size distribution are worsened when warm-temperature contributions to ice formation are decreased; Ackerman et al. (2015) find the same in parcel simulations and also demonstrate how biases can be decreased when warm-temperature contributions are substantially increased. In the simulations shown here, we also speculate that gravitational collection of stratiform ice may be too efficient, at least in the mid-troposphere, as evidenced by reflectivity increasing and number concentration decreasing substantially more rapidly than observed between 8 and 6 km (cf. Figs. 10 and 17)."

I missed also a discussion of the one-hour offset between the simulated and observed rain event initiation. Is there a hypothesis for this?

Clarification added to section 4.1: "The simulated squall line passes roughly an hour earlier than observed, which could be attributable to two general causes: (i) uncertainties in the initial and boundary conditions, including those influential to surface heat fluxes, and (ii) errors in model parameterization components, including microphysics scheme elements, which can independently influence the rainfall structure in NU-WRF simulations in this case (cf. Tao et al., 2016, their Fig. 11)."

There were two other points on which I would have appreciated clarification. I was surprised by the result that modeling only homogeneous freezing (HOMF) results in "substantially larger and

fewer outflow ice crystals". Normally homogeneous freezing yields many more and smaller crystals (e.g. DeMott et al. 1998 GRL). Why does the opposite occur here?

Clarification added to section 4.2.2: "Whereas favoring homogeneous freezing of droplets generally yields more ice particles in an updraft parcel (e.g., DeMott et al., 1998), here we find the opposite in aged stratiform outflow, where snow is the dominant hydrometeor class. Snow number concentration maxima intermittently reach ~500 L⁻¹ in all simulations except HOMF, where they reach only ~30 L⁻¹. Since 500 L⁻¹ is the limit imposed on the Cooper (1986) parameterization contributions to total ice number concentration (see Section 4.1), we conclude that removing that source is likely chiefly responsible for larger ice in HOMF outflow. We note that ice number concentrations are not conserved by design in order to enforce limits on size distribution slope parameters (Morrison et al., 2009), which complicates drawing firm conclusions about the contributions of specific processes."

Then I found the results for the size distributions in Figures 14 to 17 and radar reflectivity in Figure 22 incongruous: the distribution comparisons indicate that the simulated ice crystals are far too big, while it is suggested from the reflectivity comparison that the simulated ice is too small. Am I missing something? Clarification in both cases would be helpful.

Clarification added to section 4.3.1: "Thus, specifically at the elevations where the aircraft sampled (Fig. 16, white bars in observed reflectivity), simulated reflectivity is substantially greater than observed, consistent with ice particles substantially larger than observed (Figs. 11–13), but that is not the case at all elevations."

Otherwise my comments are related to readability. I find the article rather figure-heavy, and I think the results would be made be more accessible if the figures were condensed in some places and simplified in others. For example, Figure 2 is only referred to once, and since only the 20 May panel is particularly relevant, this panel could be combined with Figure 12. In Figure 13, only the rain gauge-corrected QPE measurements and BASE simulation are discussed, so panels a and c could be removed. Or Figures 17 and 18 could be moved to Supplemental Information, since the altitudinal dependence of Ni and mass distributions is already seen between Figures 15 and 16 and the discussion of 2DC images is quite brief.

We combined Figs. 4 and 5 and removed 6, 10–11, and 19–20. We retained 2 (for reader to quickly assess other case study conditions), 13 (emphasizes substantial uncertainty in rainfall products), 17 (15-17 are main focus), and 18 (for modelers to know what ice looks like).

I think breaking down the "Evaluation of hydrometeor size distributions in 20 May case study simulations" section into subsections, e.g. "Precipitation intensity", "Mass and number concentration distributions", and "Radar retrievals", would also ease readability.

We now use two levels of subsections in Sections 3 and 4.

Specific comments

Page 4, Line 6 – Please be consistent in the instrument acronyms. What is called the "DMA" here is later called the "HTDMA" in Figure 3 and introduced as the "TDMA" in Section 2. Again for the CPC, it is not always clear whether the measurements to which you refer are form the ground-

based or aircraft CPC; it is inferred from the other instruments you mention. You could make this more explicit.

HTDMA now used throughout. CPC now always preceded by "ground-based" or "airborne."

Page 6, Line 31 – The statement "unknown aerosol source terms are neglected" is unclear to me. The airport and power plants are mentioned in the section of Aerosol input data, but there is no discussion of back trajectories or systematic confirmation of hypothesized sources.

By unknown we meant that aerosol source terms cannot be readily observed and specified. Simplification and clarification made: "Aerosol source terms beyond advection across outer domain boundaries are neglected (e.g., primary emission and gas-to-particle conversion)."

Page 7, Lines 1-2 - 1s there also a quantitative basis (other than "similarity to April case studies") for the 8000 cm-3 and 0.005 um values chosen in the NUCL simulation? If so, this should be mentioned.

Clarification added also in response to referee 2: "Based on the April and 1 May nucleation-mode fits listed in Fig. 6, this represents the most commonly fit mode diameter and mode standard deviation, and a modest number concentration (maximum on 1 May) that is lower than typically observed in the 10–30-nm diameter range during intense new particle formation events (e.g., Crippa and Pryor, 2013)."

Page 7, Line 3 – The statement "simulations use a preliminary version of the 20 May aerosol input data" is unclear to me. The Aerosol input data section does not mention multiple processings or versions.

Clarification added: "During the course of this study, minor changes were made to aerosol observation processing concurrently with the simulations being run; simulations therefore use a preliminary version of the 20 May aerosol input data, which is negligibly different from the final version for our purposes. AERO and NUCL aerosol input files are included in Supplement 1 for completeness."

Page 8, Line 1 and Page 47, Table 2 – Could you please include the standard deviation in the "top three elevations", e.g. $7.6 \pm x$ m, and associated temperatures?

We prefer not to complicate the table because the elevations and temperatures are in a narrow range based on level legs within horizontally homogeneous conditions and the table is already complicated by showing a range of minimum and maximum values from two observational data sets.

Page 11, Line 16 – It is not clear what "similarly coherent" means here. Could you word this more substantively?

Reworded to "both predicted and observed stratiform ice size distributions exhibit relatively welldefined properties that do not vary rapidly in time."

Page 28, Figure 7 – The caption indicates that the CPC profiles on the left and UHSAS profiles on the right are in red and blue respectively as in Figures 4 and 5, but this is not the case. The thick black line for layer-wise median ratio is not so easily distinguished from the thinner black lines; perhaps the UHSAS/CPC traces can also be changed from black in the rightmost subplot. Finally

it is not clear what the "layer-wise" ratio means; are these values also calculated for km-deep layers?

Clarification added to figure and caption: "The median of airborne CPC and UHSAS aerosol number concentrations within 1-km-deep layers for each MC3E flight, and the ratio of those median values for the seven flights with both instruments (black lines). The median of profile values at each elevation (red lines) are archived as Supplement 2."

Page 29, Figure 8 – The numbers in the subpanels of this figure need to be moved to a table. This will significantly ease comparing the values between days and allow the y-axis to be readjusted for better comparison of the different traces. It is also unclear to me what the various colors (red, green, blue, purple, black) represent. The caption refers to "measurement time", but this should be clarified. A brief discussion of why the 2-mode fit is better than the 3-mode and vice versa at certain times might also be included in the third paragraph of page 5.

We used a fixed vertical axis to emphasize case study differences. The black values are archived with Supplement 1 and we disagree that the underlying values deserve a dedicated table. Clarifications added to caption also in response to referee 2: "Aerosol dry number size distributions ($dN_a/dlogD_a$) reported from HTDMA during the two-hour pre-rain period (colored solid lines; legend indicates Julian date in UTC), lognormal fits to HTDMA (colored dashed lines; text indicates fitted number concentrations in cm⁻³, geometric mean dry diameter in μ m and standard deviation), and the final case study distribution derived from the mode-wise linear mean of contributing parameters and its hygroscopicity parameter (κ) derived as the number-weighted mean of contributing HTDMA values (black dashed lines and black text; archived with Supplement 1). In the 20 May case, zero and 8000 cm⁻³ particles in the nucleation mode illustrate BASE and NUCL simulation inputs (dotted black lines)."

Reworded "It is found that two to three modes provide the best fit" to "The Vogelmann et al. (2015) algorithm optimizes a fit of two or three modes" to emphasize that we relied entirely on that algorithm since results appeared consistently satisfactory.

Page 35, Figure 14 – In my opinion, this figure could be removed, and the simulated values added to Table 2.

We have retained it because this figure conveys information that is difficult to fully capture in a table and we removed six other figures.

Page 41, Figure 20 – I am confused by the black BASE trace for number concentration. Doesn't this simulation have a fixed droplet concentration of 250 mg-1, as stated on page 6, line 26?

Correction made to text: "250 cm⁻³."

Page 42, Figure 21 – It is unclear whether only the top left panel is an integrated reflectivity; it seems so given its different scale, but this should be clarified in the caption. A definition of ZHH (as the horizontally-polarized radar reflectivity, right?), along with definitions for the pink, white, and red circles in various subpanels, would help in the interpretation of this figure.

Clarifications added to caption: "Horizontally polarized radar reflectivity (Z_{HH} in dBZ) from KVNX radar (left, dotted red circle): (top) example updraft object at ~12 UTC (solid red) among others identified in units of dBZ km (red-enclosed, see text), (middle) movement of example updraft

from initial location (solid red) towards intersection with the aircraft sampling location (whiteenclosed, see text) projected onto 2-km Z_{HH} at ~14 UTC, and (bottom) Z_{HH} curtain obtained from column-wise averages over tracked regions from ~12–15 UTC with Citation ascent legs in time and height (white bars) and averaging time used in Fig. 22 (white lines). From the AERO simulation (right): (top) identification of a typical updraft object projected onto simulated Z_{HH} at ~11 UTC (solid red) among others identified (red enclosed, see text), (middle) its movement from the identified location (solid red) to intersection with the aircraft sampling location (whiteenclosed, see text) projected onto simulated 2-km Z_{HH} at ~13 UTC, and (bottom) Z_{HH} curtain obtained from column-wise averages over tracked regions from ~11–14 UTC with mid-point of hour-long averages used in Fig. 22 (white lines)."

Page 43, Figure 22 – "Time 1", "Time 2", etc. have not been defined for the simulations. It would be clearer to label the gray traces 'AERO, Time1' etc. so that the reader knows these are only from that simulation.

Clarification added to caption: "AERO simulation times 1, 2, 3 and 4 indicated in Fig. 21 (light to dark grey lines)."

Technical comments / suggestions Page 3, Line 25 – A term like "droplet activation" or "ice nucleation" or "new particle formation" would more clearly indicate the process(es) meant by "aerosol consumption" here.

Clarification added: "via droplet activation".

Page 3, Line 26 – Remove the second "be".

Removed, thank you.

Page 6, Lines 9 - 10 - Add a 'to': "appears to be variably biased relative to the groundbased measurements".

Added, thank you.

Page 8, Lines 16-22 – Reword through here for clarity, e.g. "Consistent with underestimated Ni, the Dmax at which BASE mass distributions peak is roughly 3-5 times larger than that at which the observed distribution peaks. The Dmax at which the BASE mass distributions peak increases monotonically with increasing mass concentration, whereas the observed mass distributions tend to..

Reworded, thank you.

Page 8, Line 28 – There is an unfinished sentence beginning with "At 6.7 and 7.6 km".

"At 6.7 and 7.6 km. However," corrected to "At 6.7 and 7.6 km, however,"

Page 12, Line 18 – "Updrfts" to "updrafts"

Corrected, thank you.

Page 14, Line 4 – "have" to "has"

Corrected, thank you.

Page 14, Line 6 – "are" to "is"

Corrected, thank you.

Page 22, Figure 1 - It would ease readability if the ARM central facility were marked with a color other than yellow, since the pentagon, bull's eyes, and thumbtacks are all yellow as well.

Agreed. Since this is a stock figure that we did not generate, we did not attempt to adjust it.

Page 24, Figure 3, panel d – Is there a red trace for 0.013 um here? If so, it is not visible.

It is strongly intermittent. Clarification added to caption: "(intermittent at smallest cut)."

Page 33, Figure 12 – It would ease readability if Q2 were expanded to National Mosaic and Multi-Sensor QPE system in this caption, as well as in the text, and again if QPE were expanded here and in the text.

We have now spelled out "National Mosaic and Multi-Sensor Quantitative Precipitation Estimate" in the caption to Figure 12 and in the text.

Pages 36-38, Figures 15-17 – The red and blue traces should be labeled PMS 2DC and HVPS rather than obs1 and obs2.

Both are merged PSDs from the same raw data, adopted here as an estimate of poorly established uncertainty. Clarification added to caption: "Size distributions of ice mass (left) and number (right) in four ranges of ice water content (IWC, ranges in parentheses in g m⁻³) derived from merger of 2DC and HVPS raw data independently by Wang et al. (2015a, 'obs1' in red) and Wu and McFarquhar (2016, 'obs2' in blue). Both are shown as an estimate of poorly established uncertainty."

Page 41, Figure 20 – The y-axis should be '[km]' not '[m]'.

Figure removed.

Response to interactive comment on "Use of an observation-based aerosol profile in simulations of a mid-latitude squall line during MC3E: Similarity of stratiform ice microphysics to tropical conditions" by Ann M. Fridlind et al. by Anonymous Referee #2

This study examines and reports aerosol size distribution profiles for six convection case studies observed during the MC3E field campaign, intended for use in model simulation of those cases. The authors demonstrate use of the aerosol size distribution profiles in NU-WRF simulations of the 20 May case study with Morrison twomoment microphysics focusing on examining the stratiform cloud microphysical properties. There are some interesting findings such as ice crystal number concentrations are consistently dominated by a single mode near Dmax of 400 μ m, and a mass mode near Dmax of 1000 μ m becomes dominant with decreasing elevation to the –10 0C. Therefore, the study is worthy being published. However, this reviewer does have some concerns about the current form as listed below,

We very much appreciate the helpful questions and comments. Point-by-point responses below have greatly improved the manuscript by reducing figures, adding section numbers, and making corrections and clarifications throughout.

(1) I am a little confused about the objectives for the second half of the paper that demonstrates the use of the derived aerosol size distribution. The Introduction does not have a clear statement about the goal of this part. Their results show that simulation using the aerosol size distribution derived does not much affect ice microphysics and stratiform microphysical properties including particle size distribution. These results kind of dispute the importance of aerosol size distribution used in model simulations. Logically, to show the importance of the developed product (i.e., aerosol size distribution), the paper should present results that are significantly changed by aerosol size distribution such as precipitation rate, convection, etc. But the authors did not go to this direction and kind of ignored the point about the importance of the derived aerosol size distributions. This is ok only if the authors clearly state the reasons for doing what they chose to do and the relevant objectives.

Clarification added to Section 5: "If a warm-temperature ice multiplication mechanism is dominating outflow ice distributions in a manner that cannot be generally reproduced in simulations and is not well understood, it is difficult to confidently assess how or to what degree hygroscopic and ice-nucleating aerosols can be expected to modulate outflow ice properties. For instance, in this study we cannot be confident of the relevance of our sensitivity tests for understanding natural convective outflow owing to inadequate baseline fidelity compared with observations."

(2) Section 3 does not have a clear structure. This part is very important to the entire paper, and the authors need to be clear about (a) the methodology of how the aerosol size distributions are derived, (b) the final products provided to the community, and (c) the discussion about caveats and uncertainties. However, the current writing in this section makes readers difficult to get those. The authors are still talking that the methodology in the last 4 paragraphs of this section.

We now use two levels of subsections in Sections 3 and 4. Some additional text is added for clarification.

(3) The contribution of small CCN to droplet nucleation and ice particle concentration at upperlevels needs some further examination. The conclusion is premature. See comment #20.

Our activation treatment does not omit secondary droplet nucleation above cloud base (see response to comment #20 below). We also now clearly state that the value of our sensitivity tests is limited (see response to comment #1 above).

(4) About Section 5, although I enjoyed reading the discussion, much of the discussion should be moved to the Introduction since they are the very relevant literature studies providing the background for this work.

We consider results unexpected based on past literature, and therefore do not present discussion of results before presenting the results themselves. In the introduction we do mention Ackerman et al. (2015) as a motivating factor.

In addition, some of the things discussed here are not even mentioned in the main text or not much related (for example, the lack of the positive differential radar reflectivity and the importance of the tropical convection in global circulation).

Clarification added to Section 5: "Case studies are generally better for model development if they are relatively typical rather than unusual or rare. ... Analyses of dual-polarimetric radar observations could be further systematically employed to identify the environmental conditions associated with stratiform microphysics regimes ..."

Reference to global circulation now refers back to introduction.

(5) There are many inconsistencies between Figure, Figure captions, and the corresponding text, and also a few figure captions do not clearly describe the figures. There are quite a few sentences what do not make sense or are wrongly stated. Please refer to the specific comments below for the details.

Please see responses below and those to referee 1.

(6) Too many figures: some figures can be combined such as Fig 4 and 5, and some are not key to the main points such as Fig. 9-11, and Fig. 16-17, which could be the options for the supplemental materials since there is already a supplemental file.

We combined Figs. 4 and 5 and removed 6, 10–11, and 19–20. We retained Fig. 9 to show one comparison of derived PSD aloft with observations and 15–17 (main focus).

Detailed comments,

1. P1 Line 14-15, not sure what you want to say here, especially about the specific meaning of "the microphysics pathways associated with deep tropical convection outflow".

Reworded for clarification: "Based on several lines of evidence, we speculate that updraft microphysical pathways determining outflow properties in the 20 May case are similar to a tropical regime, likely associated with warm-temperature ice multiplication that is not well understood or well represented in models."

2. P2 Line 2, aerosol should be plural here.

Changed.

3. P2 Line 10-14, this is a very long sentence. Suggest to break into two sentences to make it easier to read.

Done.

4. P2, last paragraph, the last a few sentences of this paragraph need to be revised to clearly state the objectives of this study. If the objective is to achieve more accurate simulations, then is the goal achieved?

With respect to the last four sentences in this paragraph, we achieve the goals stated in the first to third, which respectively begin "Here we" and "We also". The last sentence begins "Enabling accurate simulation" because we intend the derived aerosol PSDs for that purpose. Since the latter is better discussed in Section 5, we removed the last sentence.

5. P3 Line29, aerosol should be plural here.

Changed.

6. P4, Line 5-11: the description here about Figure 3 suggests Na is from DMA or CPC and kappa is from HTDMA. However, the Figure 3 caption said only HTDMA, and no DMA data is shown. Please clarify the inconsistency. In addition, description about instrumental uncertainty for each instrument would be helpful here.

HTDMA now used consistently throughout. Clarification added to Section 3.2: "Based on the discrepancy between ground-based CPC and HTMDA measurements, we estimate that overall uncertainty in derived total aerosol number concentrations is roughly a factor of two throughout this work."

7. P4, Line 15-16, something is missing in the later half of the sentence. Otherwise, it does not make sense.

Latter half simplified to "nucleation mode aerosols were commonly present in large concentrations but were also commonly absent."

8. P4, Line 15-19, the description here would be clearer if the ratios of CCN to CPC aerosol concentrations are shown.

Agreed, but since we only show CCN data for completeness (not used in our fits) and we list values in Fig. 3a, we prefer to briefly state the range of ratios rather than adding another figure panel.

9. P5, Line 8 and Lin 17: what are non-case-study dates and case study dates?

Figure and sentence removed (Section 2 describes case study selection).

10. P5, I do not understand what is said in the sentence "UHSAS/CPC again sometimes decrease, not because UHSAS decreases but because CPC increases, consistent with evidence that the surface is also a source of fine particles". CPC increases suggested more small particles, which could be from particle nucleation at the elevated altitudes. This is observed quite often. So, I do not understand why we can infer that surface is the source.

Sentence clarified: "However, the local minimum in the ratio of UHSAS to CPC seen at the surface is consistent with a surface source also for fine partices (e.g., Wang et al., 2006, their Fig. 7), which could be both spatiotemporally variable and regional in nature (e.g., Crippa et al., 2013)."

11. Figure 7, there are no red and blue lines.

Figure corrected and caption revised also in response to referee 1: "The median of airborne CPC and UHSAS aerosol number concentrations within 1-km-deep layers for each MC3E flight, and the ratio of those median values for the seven flights with both instruments (black lines). The median of profile values at each elevation (red lines) are archived as Supplement 2."

12. Figure 8, why are there two colored solid lines for the measurement from HTDMA? It is really confusing with so many numbers on each panel and the description is not clear for some numbers such as the numbers at the right bottom part of each panel. Strongly suggest to use a table to show the parameters for the three modes. Also, need to explain the purpose of showing the 0 and 8000 cm-3 in the nucleation mode for May 20 case.

The black values are archived with Supplement 1 and we disagree that the underlying values deserve a dedicated table. Clarifications added to caption also in response to referee 2: "Aerosol dry number size distributions ($dN_a/dlogD_a$) reported from HTDMA during the two-hour pre-rain period (colored solid lines; legend indicates Julian date in UTC), lognormal fits to HTDMA (colored dashed lines; text indicates fitted number concentrations in cm⁻³, geometric mean dry diameter in μ m and standard deviation), and the final case study distribution derived from the mode-wise linear mean of contributing parameters and its hygroscopicity parameter (κ) derived as the number-weighted mean of contributing HTDMA values (black dashed lines and black text; archived with Supplement 1). In the 20 May case, zero and 8000 cm⁻³ particles in the nucleation mode illustrate BASE and NUCL simulation inputs (dotted black lines)."

13. Fig. 8, there are such large differences in the measurements of HTDMA for 4/25 and 5/24 in the smallest mode (although it is not clear each colored solid line represent), then any fit should have very large uncertainty. Is it meaningful for such a fit?

Clarification added to Section 4.1: "Since nucleation-mode aerosol (in the smallest fitted mode) are present very non-uniformly in time and space during some MC3E case studies (cf. Fig. 6), we finally test whether that is likely to be important."

14. Fig. 9, what is N? What is total aerosol number size distribution?

Clarification added to figure and caption: "Derived modes and aerosol number size distribution over 1-km-deep layers (black dotted and dashed lines, respectively) compared with bin-wise mean and median out-of-cloud UHSAS size distributions (red and blue lines, respectively) for the 25 April case study, with sample size (cf. Fig. 4) and total aerosol number concentration (N_a) in cm⁻³."

15. P6 Line 27-32, the text here is confusing: first, need to be specific about aerosol configurations in AERO. It is not enough to just say "initialized with the aerosol profile described above" since it is not clear "above". To me, Fig. 8 is above but there are many different aerosol parameters listed on the panel for 5/20.

Clarification added: "Aerosol are initialized within all domains to the 20 May aerosol input profile derived as described in Section 3.4 (see Supplement 1), and are fixed to it at the outermost domain boundaries."

Second, since AERO has prognostic droplet number concentrations, I do not understand why need to fix droplet number concentrations at the boundary? Shouldn't fixing aerosol be enough?

Clarification added per response to comment #16.

Third, I do not understand "Unknown aerosol source terms are neglected", thus I am confused with the later part if the sentence "how all else being equal, this increases the difference between BASE and AERO results".

By unknown we meant that aerosol source terms cannot be readily observed and specified. Simplification and clarification made also in response to referee 1: "Aerosol source terms beyond advection across outer domain boundaries are neglected (e.g., primary emission and gas-to-particle conversion)."

Lastly, it is not clear what cloud microphysics scheme is used for other simulations besides BASE.

Clarification added: "We compare observed hydrometeor size distribution properties with those simulated using Morrison et al. (2009) two-moment microphysics with hail." Additional detail is then added on the ice nucleation parameterizations used throughout (mostly off in HOMF).

16. P6 Line 33, BASE should have no aerosol since droplet number is not prognostic as shown in Table 1.

Clarification added: "In the baseline simulation (BASE), we use a fixed droplet number concentration of 250 cm⁻³. In the AERO simulation, droplet number concentration is treated prognostically as follows."

17. P7 Line 1-2, why 8000 cm-3? This sounds a very large aerosol number concentration.

Reference added and clarification also in response to referee 1: "Based on the April and 1 May nucleation-mode fits listed in Fig. 6, this represents the most commonly fit mode diameter and rounded mode standard deviation, and a modest number concentration (maximum on 1 May) that is lower than typically observed in the 10–30-nm diameter range during intense new particle formation events (e.g., Crippa and Pryor, 2013)."

18. P7, the third paragraph and Fig. 12: Q2 and Q2corr cover the entire domain, why not compare the precipitation over the entire domain? Suggest to add such a plot to Fig. 12 (after all, it would be a more robust comparison compared with that over a small domain of 100x100 km2).

We illustrate observed and simulated precipitation rates over the entire domain in Figs. 9 and 16 for context, but the objective of Fig. 8 is to show the observed and simulated time series specifically within the aircraft sampling domain that is also used for the comparisons of stratiform ice and rain properties. Clarification added to caption: "averaged over the region sampled by aircraft after 13 UTC indicated by a red rectangle in Fig. 9."

19. Figure 14, There is only one observation dataset shown in the figure, why are there two sources (Wang etal. 2015a and Wu and McFarquhar 2016)? The related discussion about the two measurements is on P8 Line 9 but the figure does not show both.

The box and whisker plots contain both observational data sets. Caption simplified" "from aircraft observations (left, see text) and from the BASE simulation (right)". Clarification added Section 4.2.1: "Fig. 10 shows ice water content (IWC) and ice number concentration (Ni) from both independently derived observational data sets."

20. P9 Line 12-14, If Morrison scheme is used, do you consider second droplet nucleation or only cloud-base nucleation is considered? I would expect secondary nucleation at higher altitudes could make significant differences if small CCN is present. Therefore, I would suggest to do another test with the secondary nucleation considered if it is not considered in the NUCL.

Clarification added to Section 4.1: "Aerosol activation follows the treatment of Abdul-Razzak and Ghan (2000), in which the supersaturation is taken as the minimum value over the time step following Morrison and Grabowski (2008, their Eqn. A10), as in Vogelmann et al. (2015, see their Sect. 5.1)." This approach does not limit droplet activation to cloud base.

21. P9 Line 18-20, I think the point is mainly supported by much smaller ice particle number concentration simulated by the model.

We consider uncertainty in observed particle number concentration far greater, as emphasized in the last sentence of the following paragraph.

22. Figure 21, please define Zm and ZHH. Also, I do not understand why each panel is plotted for a different time? And the figure order does not reflect a time evolution, and the color legend is different for the same type of figures between observation and model simulation such as Panels 2 and 3. What does the red color denote in the first four panels?

Clarifications added to caption also in response to referee 1: "Horizontally polarized radar reflectivity (Z_{HH} in dBZ) from KVNX radar (left, dotted red circle): (top) example updraft object at ~12 UTC (solid red) among others identified in units of dBZ km (red-enclosed, see text), (middle) movement of example updraft from initial location (solid red) towards intersection with the aircraft sampling location (white-enclosed, see text) projected onto 2-km Z_{HH} at ~14 UTC, and (bottom) Z_{HH} curtain obtained from column-wise averages over tracked regions from ~12–15 UTC with Citation ascent legs in time and height (white bars) and averaging time used in Fig. 22 (white lines). From the AERO simulation (right): (top) identification of a typical updraft object projected onto simulated Z_{HH} at ~11 UTC (solid red) among others identified (red enclosed, see text), (middle) its movement from the identified location (solid red) to intersection with the aircraft sampling location (white-enclosed, see text) projected onto simulated 2-km Z_{HH} at ~13 UTC, and (bottom) Z_{HH} curtain obtained from column-wise averages over tracked regions from ~11–14 UTC with mid-point of hour-long averages used in Fig. 22 (white lines)."

23. P10 Line 5-6, why suddenly talking about BASE since only AERO is compare with observations in both Figures 21 and 22.

Corrected, thank you.

24. P10 Line 30-31, suggest to reword the sentence. It is not easy to understand currently.

Agreed, reworded: "We note that breakup equilibrium is thought to require rain rates on the order of 50 mm h^{-1} , substantially greater than typical of stratiform regions (e.g., less than 15 mm h^{-1} in Fig. 8), but its existence, size distribution characteristics, and prevalence in nature have been elusive (e.g., McFarquhar, 2010; D'Adderio et al., 2015)."

25. P11 Line 15-16, "we find that predicted and observed stratiform ice size distributions are similarly coherent within the stratiform region": I am not sure what this sentence really means since simulated and observed size distributions are totally different as shown in Figs. 14-17.

Clarification added also in response to referee 1: Reworded to "both predicted and observed stratiform ice size distributions exhibit relatively well-defined properties that do not vary rapidly in time."

26. The third paragraph in Section 5: this paragraph summarizes observed results. It is natural to comparatively describe how model does here, and this information is missing from the summary currently.

Simulated number concentration and peak of ice mass size distribution are summarized in the last sentence of the second paragraph. Added there re sensitivity tests: "Results are insensitive to prognosing droplet number concentration using an observation-based profiles with or without nucleation-mode aerosol (in place of fixed droplet number concentration). Additionally turning off all ice nucleation and multiplication parameterizations except homogeneous cloud droplet and raindrop freezing leads to less and larger ice."

Added to the end of the third paragraph : "In simulations, unlike in observations, the D_{max} where the mass size distribution peak increases substantially with mass concentration at each elevation (where there is more ice mass, it is also systematically larger) and the number concentration decreases rapidly with elevation. Beneath the aircraft-sampled region, simulated mass-weighted mean diameter of rain is roughly 0.7 mm larger than retrieved, consistent with overlying ice size bias; collocated reflectivity within the range observed is consistent with a corresponding low bias in precipitation rate (Fig. 8)."

Use of an observation-based aerosol profile in simulations of a mid-latitude squall line during MC3E: Similarity of stratiform ice microphysics to tropical conditions

Ann M. Fridlind¹, Xiaowen Li^{2,3}, Di Wu^{3,4}, Marcus van Lier-Walqui^{1,5}, Andrew S. Ackerman¹, Wei-Kuo Tao³, Greg M. McFarquhar⁶, Wei Wu⁶, Xiquan Dong⁷, Jingyu Wang⁷, Alexander Ryzhkov⁸, Pengfei Zhang⁸, Michael R. Poellot⁹, Andrea Neumann⁹, and Jason M. Tomlinson¹⁰

¹NASA Goddard Institute for Space Studies, 2880 Broadway, New York, NY, USA

²Morgan State University, Baltimore, MD, USA

³NASA Goddard Space Flight Center, Greenbelt, MD, USA

⁴Science Systems and Applications, Inc., Lanham, MD, USA

⁵Columbia University, New York, NY, USA

⁶University of Illinois, Urbana-Champaign, IL, USA

⁸Cooperative Institute for Mesoscale Meteorological Studies, University of Oklahoma, and National Severe Storms Laboratory, Norman, OK, USA

⁹University of North Dakota, Grand Forks, ND, USA

¹⁰Pacific Northwest National Laboratory, Richland, WA, USA

Correspondence to: Ann Fridlind (ann.fridlind@nasa.gov)

Abstract.

Advancing understanding of deep convection microphysics via mesoscale modeling studies of well-observed case studies requires observation-based aerosol inputs. Here we derive hygroscopic aerosol size distribution input profiles from groundbased and airborne measurements for six convection case studies observed during the Midlatitude Continental Convective

- 5 Cloud Experiment (MC3E) over Oklahoma. We demonstrate use of the aerosol inputs in mesoscale model simulations of the only well-observed case study that produced extensive stratiform outflow, on 20 May 2011. At well-sampled elevations between -10 and -23° C over widespread stratiform rain, ice crystal number concentrations are consistently dominated by a single mode near ~400 μ m in randomly oriented maximum dimension (D_{max}). The ice mass at -23° C is primarily in a closely collocated mode, whereas a mass mode near $D_{max} \sim 1000 \,\mu$ m becomes dominant with decreasing elevation to the -10° C level,
- 10 consistent with possible aggregation during sedimentation. However, simulations with and without observation-based aerosol inputs systematically overpredict mass peak D_{max} by a factor of 3–5 and underpredict ice number concentration by a factor of 4–10. Previously reported simulations with both two-moment and size-resolved microphysics have shown biases of a similar nature. The observed ice properties are notably similar to those reported from recent tropical measurements. Based on several lines of evidence, we speculate that the microphysics pathways associated with deep tropical convection outflow also occurred
- 15 <u>updraft microphysical pathways determining outflow properties</u> in the 20 May MC3E case case are similar to a tropical regime, likely associated with warm-temperature ice multiplication that is not well understood or well represented in models.

⁷University of Arizona, Tucson, AZ, USA

1 Introduction

The impacts of hygroscopic, absorbing, and ice-nucleating <u>aerosol aerosols</u> on deep convection have been the subject of intensive study using both observational and modeling approaches, as summarized in several recent reviews (e.g., Tao et al., 2012; Wang, 2013; Altaratz et al., 2014). A hindrance for the modeling studies is the widely reported finding that different

- 5 advanced microphysics schemes, given the same environmental conditions and setup, often predict substantially differing results in terms of ice mass mixing ratios and other cloud properties (e.g., Zhu et al., 2012; Van Weverberg et al., 2013; Fan et al., 2015; Wang et al., 2015b; Tao et al., 2016). Microphysics schemes give such diverse results at least in part owing to the complexity of updraft microphysics and a paucity of existing field and laboratory data adequate to constrain all of the relevant physical processes and parameters (e.g., Zeng et al., 2011; Varble et al., 2014a).
- One important An objective for the representation of convective updraft microphysics in climate models, and by extension in the higher-resolution simulations commonly used to help develop climate model parameterizations (e.g., ??Wong et al., 2015), is confidence in the simulation of convective updraft outflow ice properties, which is realistic representation of the relatively long-lived convective outflow that may substantially impact global radiative budgetsand circulation(e.g., Houze, 2004; Schumacher et al., 2016; Elsaes , circulation, and climate sensitivity (e.g., Houze, 2004; Schumacher et al., 2004; Mauritsen and Stevens, 2015; Donner et al., 2016; Elsaes
- 15 Using the Cloud-Associated Parameterization Testbed (CAPT) approach to study simulation of deep convection by nine global models in various configurations, Lin et al. (2012) found that models produced grossly differing stratiform heating profiles. In a comparison of cloud-resolving simulations using a range of dynamic cores and microphysics schemes under similar conditions, Fridlind et al. (2012) found similarly pronounced differences in predicted stratiform outflow, with substantial associated impacts on radiative fluxes. Based on comparison of large-domain larger-domain convection-permitting simulations with tropical satel-
- 20 lite data, Van Weverberg et al. (2013) concluded that such simulations are sensitive to microphysics parameterizations and that more complex schemes do not necessarily perform better. Evidence from recent tropical field measurements has indicated that microphysics schemes could be failing to represent efficient ice multiplication that may strongly impact tropical updraft glaciation rate, outflow ice size, and precipitation efficiency (Ackerman et al., 2015), providing further motivation to advance fundamental knowledge of updraft microphysical pathways. Owing to the challenging complexity of coupled dynamical and
- 25 microphysical processes within outflow-generating updrafts and the increasing ability of computational approaches to resolve such coupling (e.g., Lebo and Morrison, 2015), the goal of improving understanding of deep convection processes through high-resolution simulation of well-observed case studies is increasingly attractive (e.g., Yang et al., 2015).

Establishing reliability of high-resolution simulations to advance fundamental knowledge of convective microphysics depends on observational constraint of initial conditions as well as simulation results. Whereas thermodynamic conditions may

30 be well characterized by routine observations or reanalysis fields (e.g., Zhu et al., 2012), aerosol initial conditions for any observed case study generally require detailed observational inputs (e.g., Yang et al., 2015). Here we develop hygroscopic aerosol input data sets for six convection events that were well observed during the Midlatitude Continental Convective Cloud Experiment (MC3E), a joint field program of the U.S. Department of Energy Atmospheric Radiation Measurement (ARM) program and the NASA Global Precipitation Measurement Mission (Jensen et al., 2016). Aerosol input profiles are archived as Supplement 1. We also demonstrate use of derived aerosol input size distributions in simulations of the only event with extensive stratiform outflow that was well-sampled by aircraft, on 20 May 2011 (Wang et al., 2015a; Wu and McFarquhar, 2016), with an emphasis on comparing simulated hydrometeor size distributions with observations. Enabling accurate simulation of such long-lived and radiatively important stratiform outflow is a valuable goal for global models (e.g., ?).

In the following sections, we describe selection of six convection case studies from the MC3E campaign (Section 2), derivation of aerosol specifications for each case from ground-based and aircraft measurements (Section 3), and comparison of simulated hydrometeor size distributions with observations for the 20 May case study (Section 4). Results are summarized and discussed in the context of other recent measurement campaigns and modeling studies (Section 5).

2 Case studiesstudy selection

- 10 The MC3E domain (Fig. 1) is defined by a sounding array containing a triangular X-band radar array and a central facility with additional instruments, including a Ka-band ARM Zenith Radar (KAZR), a NOAA S-band (2.8-GHz) profiling radar, a TSI model 3010 Condensation Particle Counter (CPC), a DMT model 1 Cloud Condensation Nuclei (CCN) counter, and a <u>Humidified</u> Tandem Differential Mobility Analyzer (TDMAHTDMA). We begin by focusing on the 22 April–25 May 2011 time period of MC3E for which a large-scale forcing data set was initially derived using a variational analysis approach (Jensen
- 15 et al., 2015). During this time period, ten flights of the University of North Dakota Citation aircraft provide profiles of aerosol properties to elevations of 8 km or higher (on 22, 25, and 27 April, and 1, 10, 11, 18, 20, 23, and 24 May). Aerosol number size distribution in the 0.06–1-μm diameter size range was measured on the Citation with an Ultra High Sensitivity Aerosol Spectrometer (UHSAS) and the number concentration of aerosols with diameter larger than 10 nm was measured with a TSI 3771 Condensation Particle Counter (CPC)CPC.
- 20 Owing to the importance of identifying fine-scale convection structural features in simulations, we first select case studies when the C-band Scanning ARM Precipitation Radar (C-SAPR) was fully or partly operational, which eliminates two flight dates (10 and 11 May). In order to allow simultaneous use of profiling instruments, we focus on cases in which substantial convection features passed directly over the KAZR and other nearby instruments at the central facility, which eliminates two more flight dates (22 April and 18 May). This leaves six flight dates for which aerosol property profiles are derived here for use
- 25 in convection simulations: 25 and 27 April and 1, 20, 23, and 24 May. Figure 2 illustrates the varying convection that passed over the central facility on each date, including the long duration of stratiform rain following deep convection in the 20 May case.

3 Aerosol input data

3.1 Objective

Based on evidence that aerosol consumption <u>via droplet activation</u> may be efficient in strong updrafts (e.g., Fridlind et al., 2004; Yang et al., 2015) and nanometer-sized particles could be <u>be</u>-nucleated (e.g., Ekman et al., 2006; Khain et al., 2012),

emphasis is placed on deriving size distribution profiles that include aerosol of all available sizes for each case. Owing to lack of measurements, we unfortunately omit coarse-mode (supermicron) aerosol, which may constitute $\sim 1-10$ cm⁻³ aerosol that are a small fraction of relevant hygroscopic aerosol aerosol but may be especially relevant to heterogeneous ice nucleation (DeMott et al., 2010; Corr et al., 2016). To make simulation specifications relatively simple, it is also assumed that a single size

distribution profile will be used in each case (no time dependence of specified environmental aerosol conditions), as in past deep convection studies that have specified observation-based aerosol profiles (e.g., Barth et al., 2007; Fridlind et al., 2012; Yang et al., 2015).

3.2 Ground-based measurements

As shown in Fig. 3, the ground-based aerosol instrumentation operated continuously with few interruptions throughout the 10 campaign. For each case study a two-hour time period prior to the detection of surface precipitation at the central facility is first identified on each date (dotted vertical lines in Fig. 3). When averaging measurements over these pre-convection periods, we find that the total aerosol number concentration reported by the <u>DMA-HTDMA</u> (0.012–0.74 μ m dry diameter) agrees with that reported by the ground-based CPC (0.01–3 μ m) to within 30% in all cases except 25 April and 1 May, when the <u>DMA-HTDMA</u> (DMA concentration is 80% higher and 50% lower, respectively. The reasons for disagreement are unclear; here we will

15 rely on the DMA-HTDMA data for size distribution information, while noting the discrepancy. Based on the discrepancy between ground-based CPC and HTMDA measurements, we estimate that overall uncertainty in derived total aerosol number concentrations is roughly a factor of two throughout this work.

Also shown in Fig. 3 is aerosol hygroscopicity parameter (κ) derived from HTDMA measurements, linearly averaged in six reported size ranges. Commonly low κ values of ~ 0.1 are consistent with those derived from airborne aerosol size distribution

and CCN measurements at a similar time of year over the Southern Great Plains site (Vogelmann et al., 2015). Similar to long-term measurements from the organics-rich Amazon rain forest (Pöhlker et al., 2016), there appears to be a common trend of increasing κ with size between the Aitken and accumulation mode size ranges.

In general, the variability of spread between CCN, HTDMA, and ground-based CPC indicate that nucleation mode aerosols smaller than 0.1 μ m in dry diameter commonly appeared were commonly present in large concentrations - but were also

commonly absent. CCN data reported at the highest supersaturation measured (slightly above 1%) variably account for roughly 15–80% of the aerosol reported by the ground-based CPC, and range over nearly an order of magnitude (\sim 400–3000 cm⁻³) across the six case studies, with an intermediate value of \sim 2000 cm⁻³ on 20 May.

3.3 Aircraft measurements

The MC3E aircraft measurements were commonly taken during precipitation at the ground site in order to sample cloud and
 precipitation conditions (cf. Fig. 3). We filter all aircraft aerosol measurements to remove in-cloud samples by imposing the stringent requirement that hydrometeors in the 2–50-μm diameter range measured by a Cloud Droplet Probe (CDP) remain below the detection limit (cf. McFarquhar and Cober, 2004), which is roughly 0.03 cm⁻³ given the CDP sample area of 0.3 mm² (Lance et al., 2010) and a typical Citation aircraft speed of 100 m s⁻¹. Unfortunately, out of the six convection case

studies considered here, UHSAS data were available only for the first three and <u>airborne</u> CPC data only for the latter three. After surveying the available <u>airborne</u> CPC and UHSAS data for the six case studies, we therefore analyze aerosol measurements from all flight dates to provide estimates of missing information.

Out-of-cloud airborne CPC profiles measured on 20, 23 and 24 May indicate that nucleation mode aerosols could be present

- 5 in the region even when they were not detected at the ground site during the pre-convection period (Fig. ??.4). Based on the <u>airborne</u> CPC data available from twelve flights during MC3E, freshly nucleated particles were commonly associated with condensation nuclei concentrations in excess of 10⁴ cm⁻³, typically limited to or most concentrated below 1–3 km in altitude, and encountered during every flight except that on 23 May. Thus, even when not present at the ground site, as on 20 May, nucleation mode particles were virtually always present somewhere nearby. However, aircraft data consistently indicated a high
- 10 degree of variability in the distribution of nucleation mode particles in the boundary layer. Maps of <u>airborne CPC</u> concentration as a function of latitude and longitude on each flight indicated that the nucleation mode was commonly limited to a short flight segment (not shown), indicative of transects through plumes likely generated by emissions from multiple nearby power plants that may not broaden efficiently downwind (e.g., Wang et al., 2006; Stevens et al., 2012). The airport could also be a source affecting the airborne samples (e.g., Westerdahl et al., 2008).
- 15 Out-of-cloud UHSAS profiles measured on 25 and 27 April and 1 May indicate median concentrations of 100–1000 cm⁻³ commonly decreasing with increasing elevation (Fig. ??4). On the latter two dates, long flight legs at a single elevation indicate horizontal variability commonly exceeding an order of magnitude both greater and lesser than relatively well-defined mean profiles.

Fig. ?? shows the Figure 5 shows profiles of median UHSAS and CPC concentrations on three non-case-study dates for

- 20 airborne CPC concentrations during all MC3E flights, as well as their ratio for the seven flights on which both instruments simultaneously functioned. UHSAS is shown to represent a fraction of airborne CPC that generally decreases with height above ~2 km, consistent with the expectation that the surface is a source of the larger aerosol. Below 2 km, however, UHSAS /CPC again sometimes decrease, not because UHSAS decreases but because CPC increases, consistent with evidence that However, the local minimum in the ratio of UHSAS to CPC seen at the surface is also a source of fine particles consistent with a surface
- source also for fine particles (e.g., Wang et al., 2006, their Fig. 7), which could be both spatiotemporally variable and regional in nature (e.g., Crippa and Pryor, 2013). Considering the general vertical trend of number concentrationduring the seven MC3E flights for which both UHSAS and CPC data are simultaneously available, median out-of-cloud UHSAS number concentration summed over 0.06–1.0 µm dry diameter accounts for 20–60% of collocated median airborne CPC number concentration when taken over 1-km vertical layers during each flight(Fig. 5)... The campaign-wide median profile of the ratio of UHSAS to CPC profiles and their ratio is archived as Supplement 2.
 - Since each case study date offered only one of the two instruments UHSAS or CPC but not both, the median ratio of UHSAS to airborne CPC number concentration shown in Fig. 5 is used as a guide for scaling ground-based measurements, which are derived as follows.

3.4 Derivation of hygroscopic aerosol input data

3.4.1 Below 1 km

Input profiles for each case study are derived at 1-km vertical resolution owing to the commonality of relatively sparse aircraft data over some elevations. Ground-based measurements are used to define aerosol number size distribution and hygroscopicity

- 5 in the 1-km layer at the surface. First, all DMA-HTDMA size distributions measured during each pre-convection period are fit with lognormal modes using the approach described by Vogelmann et al. (2015). It is found that two to three modes provide the best fit The Vogelmann et al. (2015) algorithm optimizes a fit of two or three modes for each size distribution (Fig. 6), where the nucleation-mode fits are intended for truncation at a user-determined size guided by model representation of ambient molecular clusters (cf. Jiang et al., 2011). The mode properties are then averaged in time. In the case that three modes provided a best fit,
- 10 those are each averaged by mode. In the case that only two modes provide a best fit, then if the mode geometric mean diameter of the smaller mode is smaller than or equal to 0.01 μ m, that mode is considered the smallest of three; otherwise, that mode is considered the middle of three. A simple linear average of the modal properties (number concentration, geometric mean diameter and standard deviation) is then adopted in each of the three modes. A hygroscopicity parameter (κ) is then derived for each mode as follows. First a κ value is calculated from the mean growth factor measured by the TDMA-HTDMA during each
- 15 pre-convective period, available in six size cuts over 0.013–0.40 μ m in dry diameter. Then a κ value is assigned to each DMA HTDMA size bin, using linear interpolation. Finally, a κ value is calculated for each mode as a number-weighted average over occupied bins (cf. Fig. 6).

The ambient aerosol modal properties are treated as constant in space and time for each case study. However, the

3.4.2 Above 1 km

20 The number concentration in each mode is assumed to vary with height above the surface layer, and its variation is derived from aircraft measurements as follows. First, we adopt the ground-based three-mode fit for each case study as representative of the bottom kilometer of the atmosphere. Owing to an absence of fine-mode aerosol size distribution information aloft, we then assume that aerosol in the smallest mode are confined to the surface layer, consistent with the occurrence of increased concentrations primarily near the surface (Fig. ??4). Numbers in the larger two modes above the surface layer are then determined as

25 follows.

If When only UHSAS data are available , for (first three case studies, cf. Fig. 4), in each 1-km layer above the surface layer, the number concentration in the larger mode is set to the total concentration measured by UHSAS, and the number concentration in the smaller mode is set such that the ratio of UHSAS to total assumed aerosol matches the experiment-wide median ratio at that altitude. Number concentrations in any of the smallest few UHSAS bins that exceeded five times the

30 concentration in bins with diameter larger than 0.1 μ m (in terms of dN/dlogD) appeared spurious, and these are neglected when present (e.g., as in lowest two levels in Fig. 7). The resulting number-wise scaling of the ground-based DMA-size distributions to the total UHSAS concentration numbers obtained by aircraft often gives remarkably close fits to bin-wise median-the UHSAS size distributionsabove 1 km, although aircraft data, as demonstrated in Fig. 7. However, nearest the surface, aircraft measurements appear to be variably biased relative to the ground-based measurements within 1 km of the surface: : substantially biased low on 25 April(, but perhaps not significantly owing to very small sample size (cf. Fig. 7) and versus modestly high on 27 April (Fig. ??), with closest and close agreement on 1 May (Fig. ??not shown).

Otherwise only When only airborne CPC data are available (for the remaining the second three case studies), and then 5 in each 1-km layer above the surface layer, the ratio of the number concentration of the larger aerosol mode to the smaller aerosol mode is set to the experiment-wide median as a function of height, and the total of both modes is set to the median concentration measured by airborne CPC in the corresponding 1-km layers.

For layers above the maximum measurement altitude all case studies, at elevations above the aircraft measurements (5–10 km, cf. Fig. 4), the aerosol are taken to be fixed to that in the highest layer for which measurements are available(from either UHSAS or CPC, depending on the case study). Resulting aerosol specifications for simulations of each case are archived as

Supplement 1.

Owing to the relatively simple modal structure of the input aerosol profiles derived here, an estimated coarse mode could be appended using climatological data or other field measurements (e.g., Corr et al., 2016), but we do not attempt that here.

4 Evaluation of hydrometeor size distributions in 20 May case study simulations

15 4.1 Simulations

We demonstrate use of derived aerosol input data in the 20 May case study. Our simulations of the case (Table 1) use the NASA Unified Weather Research and Forecasting (NU-WRF) model (Peters-Lidard et al., 2015), set up as described by Tao et al. (2016), with an innermost domain of 1-km horizontal grid spacing (Tao et al., 2016, their Fig. 2). We compare observed hydrometeor size distribution properties with those simulated using Morrison et al. (2009) two-moment microphysics with hailand

- 20 fixed droplet number concentration of 250 mg⁻¹ (. Except in one sensitivity test described below, we use all ice formation parameterizations that are standard in the NU-WRF implementation. Heterogeneous freezing of cloud droplets and raindrops is limited to temperatures colder than -4° C. Immersion freezing of cloud droplets and raindrop follows Bigg (1953). Contact nuclei available to freeze cloud droplets is calculated as a function of supercooling following Meyers et al. (1992). Deposition and condensation freezing follows Thompson et al. (2004, their Eqn. 2) in adopting the Cooper (1986, their Fig. 4.3) fit to ice
- 25 concentrations measured in moderately supercooled clouds, except implemented where ice supersaturation exceeds 8% or where liquid saturation is exceeded and temperature is colder than -8° C; when those conditions are met, the number of cloud ice crystals nucleated is also not permitted to drive the total ice number concentration (including cloud ice, snow and hail) over 500 L⁻¹. Secondary ice formation via rime-splintering between -3 and -8° follows Hallett and Mossop (1974). Homogeneous freezing of cloud droplets and raindrops is instantaneous at -40° C. A maximum cloud ice number concentration
- 30 of $300 L^{-1}$ is also imposed. Cloud ice number concentrations consistently exhibit $\sim 300 L^{-1}$ maxima in all simulations reported here, but we found stratiform outflow size distributions compared with observations (see Section 4.2) were insensitive to removing that limit.

In the baseline simulation (BASE) and with prognostic droplet number concentration using semi-prognostic aerosol initialized with the aerosol profile described above (AERO), we use a fixed droplet number concentration of 250 cm⁻³. In the AERO simulation, mode-wise aerosol plus droplet number concentrations are everywhere initialized droplet number concentration is treated prognostically as follows. Aerosol are initialized within all domains to the aerosol input profiles and fixed to them

- 5 profile derived as described in Section 3.4 (see Supplement 1), and are fixed to it at the outermost domain boundaries. Aerosol activation follows the treatment of Abdul-Razzak and Ghan (2000), in which the supersaturation is taken as the minimum value over the time step following Morrison and Grabowski (2008, their Eqn. A10), as in Vogelmann et al. (2015, see their Sect. 5.1). During simulation time, mode-wise aerosol are aerosol are treated semi-prognostically as the sum of unactivated aerosol and droplets present, consumed by droplet collision-coalescence, and transported as in Fridlind et al. (2012, their 'DHARMA-2M'
- 10 simulation) and Endo et al. (2015, their 'DHARMA BIN' simulations). Unknown aerosol source terms are neglected; all else being equal, this increases the difference between BASE and AERO results (based on droplet number concentration differences discussed below). Aerosol source terms beyond advection across outer domain boundaries are neglected (e.g., primary emission and gas-to-particle conversion). The HOMF simulation is identical to the AERO simulation except that all ice nucleation and multiplication mechanisms are turned off except for homogeneous freezing of cloud droplets and raindrops (see Table 1).
- Since nucleation-mode aerosol (in the smallest fitted mode) are present very non-uniformly in the time and space during some MC3E region, their concentration is set to zero in the BASE and AERO simulations case studies (cf. Fig. 6), we finally test whether that is likely to be important. In a sensitivity test simulation (NUCL), 8000 cm⁻³ nucleation-mode particles are added to the bottom 2 km in a mode with geometric mean diameter of 0.005 μ m and geometric standard deviation of 3, similar to the April case studies. We note that simulations use a preliminary version of the 20 May aerosol input data, which is
- 20 negligibly different from the final version for the purposes of this study (the preliminary version is included in Supplement 3. Based on the April and 1 May nucleation-mode fits listed in Fig. 6, this represents the most commonly fit mode diameter and mode standard deviation, and a modest number concentration (maximum on 1for reference).

As noted above, the May) that is lower than typically observed in the 10–30-nm diameter range during intense new particle formation events (e.g, Crippa and Pryor, 2013). To clarify the contrast between NUCL and AERO simulations, the nucleation

- 25 mode number actually fitted in the 20 May case is unique during MC3E owing to robust in situ sampling of extensive stratiform outflow from deep convection by the Citation aircraft (cf. Wang et al., 2015a; Wu and McFarquhar, 2016). Here we use ice number and mass size distributions derived from a PMS two-dimensional cloud (2DC) probe and a SPEC Inc.high-volume precipitation spectrometer (HVPS)probe (cf. Baumgardner et al., 2011, and references therein) on the aircraft. Since the derivation of number and mass size distributions and their integrals from such probes introduces substantial sources of uncertainty
- 30 that are often not well quantified to date (e.g., Baumgardner et al., 2011), we use two independent derivations described in Wang et al. (2015a) and Wu and McFarquhar (2016), which differ in details of the methods used to process and estimate mass from the raw image probe data. For computation of mass median area-equivalent diameter (MMD_{eq}), we follow the Wu and McFarquhar (2016) approach for a first estimate, substituting the Baker and Lawson (2006) habit-independent mass-area dimensional relation for a second estimate (Table 2)set to zero in AERO (dotted lines in Fig. 6). During the course of this study.
- 35 minor changes were made to aerosol observation processing concurrently with the simulations being run; simulations therefore

used a preliminary version of the 20 May aerosol input data that is negligibly different from the final version for our purposes. AERO and NUCL aerosol input files are included in Supplement 1 for completeness.

Over a 100x100-km domain centered on the C-SAPR radar, Fig. 8 shows the time series of surface precipitation derived from C-SAPR, from the National Mosaic and Multi-Sensor <u>QPE system Quantitative Precipitation Estimate</u> (Q2, Zhang et al.,

- 5 2011) with and without rain gauge correction (Tang et al., 2014), and from the BASE simulation with fixed droplet number concentration in the region sampled by the Citation aircraft (region bounded by red rectangle in Fig. 9). Although the The simulated squall line passes roughly an hour earlier than observed, we find qualitative which could be attributable to two general causes: (i) uncertainties in the initial and boundary conditions, including those influential to surface heat fluxes, and (ii) errors in model parameterization components, including microphysics scheme elements, which can independently influence
- 10 the rainfall structure in NU-WRF simulations in this case (cf. Tao et al., 2016, their Fig. 11). Nonetheless, we find relatively good agreement between predicted and retrieved maximum precipitation rates (about 20–30 mm h⁻¹) and the duration of rates greater than 50% of maximum (about 1 h). At a time representative of Citation aircraft sampling of the stratiform outflow (13:40 UTC), Fig. 9 shows a map of Q2 precipitation over the inner domain; the region sampled by the Citation aircraft is bounded by a red rectangle. Also shown is surface precipitation from the BASE simulation at the time of heavy stratiform
- 15 precipitation (13:00 UTC, cf. Fig. 8). The BASE simulation shows a precipitation structure oriented in a band from southwest to northeast, similar morphologically to that observed (, as do all subsequent simulations; sensitivity tests discussed below (not shown).

4.2 Comparison with in situ observations

4.2.1 Baseline simulation

- 20 As noted above, the 20 May case is unique during MC3E owing to robust in situ sampling of extensive stratiform outflow from deep convection by the Citation aircraft (cf. Wang et al., 2015a; Wu and McFarquhar, 2016). Here we use ice number and mass size distributions derived from a PMS two-dimensional cloud (2DC) probe and a SPEC Inc. high-volume precipitation spectrometer (HVPS) probe (cf. Baumgardner et al., 2011, and references therein) on the aircraft. Since the derivation of number and mass size distributions and their integrals from such probes introduces substantial sources of uncertainty that are often not
- 25 well quantified to date (e.g., Baumgardner et al., 2011), we use two independent derivations described in Wang et al. (2015a) and Wu and McFarquhar (2016), which differ in details of the methods used to process and estimate mass from the raw image probe data. For computation of mass median area-equivalent diameter (MMD_{eq}), we follow the Wu and McFarquhar (2016) approach for a first estimate, substituting the Baker and Lawson (2006) habit-independent mass-area dimensional relation for a second estimate (Table 2).
- Over the red-enclosed regions shown in Fig. 9, which bound the aircraft in situ sampling of stratiform conditions, Fig. 10 shows ice water content (IWC) and ice number concentration (N_i) derived from merged aircraft instruments from both independently derived observational data sets alongside simulated values. Observed ice number concentrations at three wellsampled elevations (Table 2) are within the range of those reported from nine storms measured over Colorado and Oklahoma in

May and June 2012 during the Deep Convective Clouds and Chemistry Experiment (Corr et al., 2016, $10-120 L^{-1}$). We have omitted analysis of observations at lower elevations (temperatures warmer than -10° C) owing to initially suspected encounters with supercooled water, which can be difficult to confidently rule out (Wang et al., 2015a). Conditions at or near ice saturation are generally expected over heavy stratiform rain (e.g., Grim et al., 2009), but conditions ranging from liquid saturation to

- 5 ice subsaturation above the stratiform melting layer could be associated with differing midlevel inflow positions or embedded convective-scale perturbations (e.g., Barnes and Houze, 2016). Later analyses of the 20 May case provide evidence of local ice subsaturation above the melting level associated with bright band variability observed in C-SAPR fields (Kumjian et al., 2016). Here we focus on the top three elevations that were well-sampled and consistently more than 1 km above the variable bright band zone.
- The aircraft observations shown in Table 2 are taken from five level legs flown between 13.9 and 14.9 UTC, except roughly one-third of the observations at -23° C that are taken from an isolated level leg later in the same flight (cf. Wang et al., 2015a, their Fig. 5). Since results are not sensitive to excluding the later samples, we consider the observations statistically representative of the 13.9–14.9 UTC time period. For our comparisons, simulations are sampled roughly one hour earlier, consistent with earlier squall line passage, using 10-minute output fields between 13 and 14 UTC.
- 15 With increasing elevation in the BASE run, summing all model ice classes, simulated IWC generally decreases while N_i increases; both observational analyses show similar patterns in some respects, although the trend in N_i across the three best sampled elevations is not discernible. Overall, the largest apparent deviation of simulations from observations in Fig. 10 is roughly 4–10 times fewer ice crystals, although sampling remains relatively sparse and observational uncertainty could be very large. We do not attempt to quantify uncertainty in N_i here owing to the difficulty of doing so and the relative lack of importance to analyses below, which are primarily focused on the size distributions of mass rather than number. In similar
- simulations of the 20 May case, Fan et al. (2015) show a similar order of magnitude underestimate of measured N_i .

Figures 11–13 show the underlying mass and number size distributions at the three well-sampled elevations (5.8, 6.7 and 7.6 km) as a function of ice crystal randomly oriented maximum dimension (D_{max} , cf. Wu and McFarquhar, 2016). At 5.8 km (-10° C), simulated and observed mass and number size distributions are compared in four mass concentration intervals span-

- 25 ning 0.2–1 g m⁻³ (Fig. 11). The Consistent with underpredicted N_i , the D_{max} where the BASE mass size at which BASE mass distributions peak is roughly 3–5 times larger than observed, consistent with underestimated N_i within each mass concentration rangethat at which the observed size distributions peak. The D_{max} where at which the BASE mass size distribution peaks distributions peak increases monotonically with increasing mass whereas the observed mass size distributions tend to peak consistently at D_{max} of roughly 1–2 mm, generally independent of IWC range; other recent deep convection observations have
- 30 found notably weak dependence of convective outflow ice size on mass concentration at fixed elevations (Fridlind et al., 2015; Leroy et al., 2015). We note that the D_{max} at which observed and simulated size distribution lines cross one another (are equal) is greater for number than for mass because the effective density of the relevant ice particles (namely, snow) is less in the observations than in the model microphysics scheme (0.1 g cm⁻³, Morrison et al., 2009). Overall, there is a marked absence of particles with $D_{\text{max}} < 1000 \,\mu\text{m}$ in the BASE simulation, suggesting that they are not produced or are lost via a process such 35 as aggregation.

10

Observed number size distributions peak at D_{max} of roughly 400 μ m, which does not significantly change with either mass mixing ratio or elevation (cf. Figs. 12–13). At 6.7 and 7.6 km (–16 and –23°C). However, however, mass size distributions appear to fall into two modes: one peaking at $D_{\text{max}} \sim 500 \ \mu$ m (most apparent at the lowest mass mixing ratios) and a second peaking at $D_{\text{max}} \sim 1$ mm. The D_{max} where observed number size distributions peak (at all elevations and mass concentrations)

- 5 is similar to that where the smaller-mode mass size distribution peaks. In the observations, evolution with decreasing height from alignment to non-alignment of the mass and number size distribution peaks (namely, a shift of the mass size distribution peak to larger sizes that is not accompanied by a shift of the number size distribution peak) is suggestive of aggregation that is adequate to increase mass median D_{max} but insufficient to increase number median D_{max} , conceivably owing in part to greater sticking efficiency among larger colliding particles.
- Subjective inspection of ice crystal images generally shows that aggregates are more common at larger sizes and lower elevations, consistent with the possibility that aggregation may be largely responsible for the coherent trend in observed particle size with elevation. However, the general irregularity of the ice particles (Fig. 14) makes confidently distinguishing aggregates from non-aggregates far more difficult than in a case where dendrites are the dominant habit, for instance, and aggregate fraction can be readily estimated for simulation evaluation (e.g., Avramov et al., 2011). In this case, aggregates appear present at the
- 15 highest elevation sampled (-23° C), but it has been suggested that aggregation may be a negligible process at temperatures warmer than -20° C (e.g., Barnes and Houze, 2016) and we cannot rule out the possibility that aggregation is not a dominant determinant of size distribution trends seen here in observations between -10 and -23° C.

4.2.2 Sensitivity tests

Figure 15 demonstrates the effect of replacing fixed droplet number concentration in the BASE simulation with the aerosol input data derived in Section 3 and prognostic droplet number concentration. The AERO ice size distributions are found to be largely unaffected compared with the BASE simulation. If nucleation-mode aerosols are added to the aerosol input file (NUCL simulation), results are similarly little affected. Inner-domain averages of cloud water mixing ratio and number concentration indicate that AERO droplet number concentrations are substantially smaller than fixed BASE values, especially aloft, and nucleationmode aerosols are scarcely activated in the NUCL simulation (Fig. ??not shown). A sensitivity test in which all heterogeneous

- freezing parameterizations and ice multiplication mechanisms are turned off (HOMF), by contrast, results in substantially larger and fewer outflow ice crystals, worsening agreement with observations (cf. Fig. 15). Thus, we find that the combined effect of ice crystal formation parameterizations. Whereas favoring homogeneous freezing of droplets generally yields more ice particles in an updraft parcel (e.g., DeMott et al., 1998), here we find the opposite in aged stratiform outflow, where snow is the dominant hydrometeor class. Snow number concentration maxima intermittently reach ~500 L⁻¹ in all simulations except HOMF, where
- 30 they reach only ~30 L⁻¹. Since 500 L⁻¹ is the limit imposed on the Cooper (1986) parameterization contributions to total ice number concentration (see Section 4.1), we conclude that removing that source is likely chiefly responsible for larger ice in HOMF outflow. We note that species number concentrations are not conserved by design in order to enforce limits on size distribution slope parameters (Morrison et al., 2009), which complicates drawing firm conclusions about the contributions of specific processes. In summary, we find that the combined effects of heterogeneous ice nucleation and ice multiplication have a

greater effect on outflow ice size than droplet spectra changes over the range in BASE versus AERO simulations. The fact that all of the simulations also substantially overestimate outflow ice size (where directly observed) is consistent with the possibility that the microphysics scheme could be missing some critical aspects of ice nucleation or ice multiplication.

In all simulations N_i decreases by roughly a factor of 8 between 7.6 and 5.8 km (as in Fig. 10). Observed N_i does not show
a discernible trend over the well-sampled elevations examined here (Table 2). These results suggest that simulated aggregation is more aggressive than observed in this case. In Fan et al. (2015) simulations of the same case with another two-moment scheme and a size-resolved microphysics scheme, N_i decreases by roughly a factor of 20 over a similar altitude range (cf. their Fig. 11b). Profiles of stratiform N_i measured during the Bow Echo and Mesoscale Convective Vortex Experiment (BAMEX) exhibited 25% decline per degree C between 0 and -10°C, but were not reported at colder temperatures (McFarquhar et al., 2007; Smith et al., 2009). Because measurement uncertainty in N_i remains essentially unquantified to date (e.g., Fridlind et al.,

10 2007; Smith et al., 2009). Because measurement uncertainty in N_i remains essentially unquantified to date (e.g., Fridlind et al. 2007, uncertainty estimated at a factor of five), we do not attempt to draw conclusions at this point.

4.3 Comparison with radar observations

4.3.1 Radar reflectivity

Radar reflectivity time series from the NEXRAD KVNX radar can place the aircraft-sampled elevations and locations into

- 15 greater context. By identifying columns of enhanced specific differential phase above the melting level in KVNX data, which can be taken as an indication of updraft location (Van Lier-Walqui et al., 2016), and using the nearest radiosonde to represent horizontal winds, we estimate that roughly two hours passed between ice detrainment from updrafts at roughly 35.5°N and Citation sampling at roughly 36.5°N (Fig. 16, left panels). A similar analysis of supercooled liquid water above the melting level and horizontal winds in the BASE simulation indicates a slightly shorter time period (Fig. 16, right panels); we have not
- 20 attempted to objectively optimize this analysis since results are not strongly sensitive to choice of time and location owing to the fact that conditions are quite horizontally uniform in both observed and simulated stratiform outflow.

Figure 17, derived from the fields shown in Fig. 16, illustrates that simulated reflectivity profiles below roughly 9 km are biased high in the AERO simulation(consistent with stratiform ice that is too large), but simulated reflectivity above roughly 10 km is biased low. Thus, specifically at the elevations where the aircraft sampled (Fig. 16, white bars in observed reflectivity),

- 25 simulated reflectivity is substantially greater than observed, consistent with ice particles substantially larger than observed (Figs. 11–13), but that is not the case at all elevations. Referring back to Fig. 16 (bottom panels) it is apparent that ice detrained from updraft tops in the BASE AERO simulation may be either insufficiently concentrated or too small, but we have no direct measurements to confirm either possibility. At 8 km, Fan et al. (2015) simulations using both two-moment and size-resolved microphysics schemes show similar significant overestimates of 8-km reflectivity within stratiform outflow (cf. their Fig. 3b),
- 30 indicative of similar biases in ice size (systematically larger than observed).

4.3.2 Polarimetric retrievals

The radar fields used to place aircraft sampling into context are also sensitive to rain size distribution beneath the Citation sampling location. Figure 18 shows the median and inner half of raindrop mass-weighted mean diameter (D_m ; the fourth moment of the drop number size distribution divided by the third moment) as retrieved from KVNX data following Ryzhkov et al. (2014), with an estimated uncertainty of roughly 5–10% (Thurai et al., 2012). The retrievals shown are made along the

- 5 lowest-elevation KVNX beam, which varies in height with distance, but simulated values vary relatively little over that height range for the subregion selected to match the Citation sampling location. In that stratiform region (rectangular regions in Fig. 9) at the onset of the heaviest stratiform precipitation (13 UTC observed, 12 UTC simulated, cf. Fig. 8), simulated median D_m is roughly 40% (0.7 mm) larger than observed, consistent with simulated stratiform ice size larger than observed at 5.8 km (roughly 2 km above the melting level).
- Retrieved D_m of 1.5–2 mm in the stratiform regime is on the high end of climatological values for various locations (cf. Thurai et al., 2010, their Fig. 2), but quite similar to stratiform values measured by disdrometer and retrieved from profiling radar in the same storm (cf. Williams, 2016, their Fig. 5b) and also in a tropical mesoscale convective system (cf. Varble et al., 2014b, their Fig. 17). Simulated D_m values are larger than the upper end of stratiform values climatologically and show a high bias also found in similar simulations under tropical conditions using the same scheme (cf. Varble et al., 2014b, 'WRF-2M' in the same scheme (cf. 17).
- 15 their Fig. 17).

Figure 19 shows simulated (BASE and AERO) and retrieved D_m values as a function of collocated precipitation rate. Simulated stratiform rain D_m values shown in Fig. 18 (selected to match the Citation location during aircraft sampling) are roughly equal to the microphysics scheme's breakup equilibrium value of 2.4 mm (cf. Morrison and Milbrandt, 2015), which is seen throughout the high precipitation rate limit in simulations. Observed D_m asymptotes more monotonically to a relatively

- ²⁰ broader range in the high precipitation rate limit, where many retrieved values lie within retrieval uncertainty of 2.4 mm. We note that the breakup equilibrium is thought to require rain rates on the order of 50 mm h⁻¹, substantially greater than typical of stratiform conditions (e.g., less than 15 mm h⁻¹ in Fig. 8), but its existence, size distribution characteristics, and prevalence of breakup equilibrium has been clusive in nature , despite the idea that it might be guaranteed at much higher rain rates than found under stratiform conditions in nature have been elusive (e.g., McFarquhar, 2010; D'Adderio et al., 2015).
- We also note that a mass-weighted mean diameter of 2.4 mm corresponds to a mean volume diameter of 1.1 mm for an exponential size distribution in the microphysics literature (e.g., Morrison and Milbrandt, 2015, their Appendix C), whereas the two diameters with the definition of the latter are used interchangeably in the radar literature (e.g., Testud et al., 2001). Considering raindrop size in general terms, the reduced droplet number concentrations in the AERO versus BASE simulation are associated here with a reduction in the frequency of D_m values below 2.4 mm at convective rain rates of 20–40 μ m
- 30 (cf. Fig. 19). This reduction is consistent with a pattern of increasing raindrop size with increasing aerosol or droplet number concentration shown in past modeling studies over a wide range of thermodynamic conditions (e.g., Storer et al., 2010) and also found over multi-day statistics using similar retrievals of raindrop size alongside ground-based aerosol observations under tropical conditions (May et al., 2011).

5 Summary and discussion

We report hygroscopic aerosol size distribution profiles for six convection case studies observed during the MC3E field campaign over Oklahoma. Each profile is derived by merging ground- and aircraft-based measurements. Missing aircraft data owing to instrument failures are filled by using experiment-wide analysis of flights where all instruments functioned well. The

5 aerosol profiles, archived as Supplement 1, are intended for use in modeling studies of convection microphysics, where both aerosol and hydrometeor size distribution data are required to evaluate fidelity of model physics.

We demonstrate use of the aerosol size distribution profiles in NU-WRF simulations of the 20 May case study, where widespread stratiform outflow was also well sampled by aircraft. Using Morrison et al. (2009) two-moment microphysics with hail in NU-WRF as an illustrative example, we compare simulated ice size distributions with measurements made in the outflow

10 region. Across several sensitivity tests (Table 1), we find that both predicted and observed stratiform ice size distributions are similarly coherent within the stratiform region whibit relatively well-defined properties that do not vary rapidly in time. However, simulated ice number concentrations (N_i) are roughly 5–10 times lesser than observed and the peak of ice mass size distributions roughly 3–5 times larger, correspondingly. Results are insensitive to prognosing droplet number concentration using an observation-based profile with or without nucleation-mode aerosol (which is found to be spatiotemporally variable

15 across case studies). Additionally turning off all ice nucleation and multiplication parameterizations except homogeneous cloud droplet and raindrop freezing leads to less and larger ice.

Across three well-sampled elevations between 5 and 8 km (at -10, -17, and -23° C), observed ice number size distributions peak at a randomly oriented maximum dimension (D_{max}) of roughly 400 μ m at all elevations, and lack a discernible vertical trend in total N_i (Table 1). At the highest elevation sampled, the derived mass size distribution appears to peak at a D_{max} only

- slightly larger than 400 μ m. At lower elevations, the peak D_{max} of the observed mass size distribution is shifted to a size twice as large, at roughly 1 mm, perhaps owing to aggregation that is apparent in ice crystal images. However, some mass remains in the smaller size range where numbers are concentrated. always concentrated. In simulations, unlike in observations, the D_{max} where the mass size distribution peaks increases substantially with mass concentration at each elevation (where there is more ice mass, it is also systematically larger) and the number concentration decreases rapidly with elevation. Beneath the
- 25 aircraft-sampled region, simulated mass-weighted mean diameter of rain is roughly 0.7 mm larger than retrieved, consistent with overlying ice size bias; collocated reflectivity within the range observed is consistent with a corresponding low bias in precipitation rate (Fig. 8).

In general, stratiform microphysics features seen in this 20 May mid-latitude squall line case appear notably similar to those observed in the tropics, as during the recent High Altitude Ice Crystals/High Ice Water Content (HAIC/HIWC) campaign that

sought to robustly characterize ice properties that might be encountered by commercial aircraft transiting mesoscale convective systems around Darwin, Australia (Dezitter et al., 2013; Leroy et al., 2015; ?)(Dezitter et al., 2013; Leroy et al., 2015, 2017). Perhaps most prominently, ice mass median area-equivalent diameter (MMD_{eq}) values of 500–700 μ m between -15 and -25°C (Table 2) are close to those found around Darwin in the same temperature range, and MMD_{eq} maxima of 900–1200 μ m are also within the range found there (?, their Figs. 6 and 9)(Leroy et al., 2017, their Figs. 6 and 9). Figs. 12 and 13, where the mass size distributions shown are visually integrable, show that the majority of mass in the 20 May case is generally found in a size range roughly bounded by half and twice the mass median size. Despite quite a bit of scatter, this condition found during HAIC/HIWC (?, their Fig. 9) (Leroy et al., 2017, their Fig. 9) is indicative of a relatively narrow mode of ice mass around its median size, similar to that previously reported by Heymsfield (2003) from a combination of tropical and

5 mid-latitude measurements. We leave more detailed comparison of MC3E and HIWC/HAIC size distributions to future work, but here briefly note several other general similarities.

Although we have not identified the capped column habit that is common among convective ice crystal habits in the tropics (e.g., Grandin et al., 2014; Ackerman et al., 2015), there is a predominance of irregular, compact crystals on 20 May (cf. Fig. 14), similar to those seen in tropical convective outflow during HAIC/HIWC (Leroy et al., 2015) and during the Tropical

- 10 Composition, Cloud and Climate Coupling and NASA African Monsoon Multidisciplinary Analyses field campaigns (Lawson et al., 2010). A less prominent similarity that can be generally gleaned from Figs. 11 and 12 is that the ice size distributions on 20 May show relatively weak correlation of ice mass median D_{max} with IWC at fixed elevations aloft, especially in constrast contrast to simulations here; a similar observation-simulation contrast has been reported under tropical conditions (Ackerman et al., 2015, their Fig. 3). Over ten-degree temperature intervals colder than -5° C (analogous to level flight legs)
- 15 here), HAIC/HIWC Darwin observations show a pattern of MMD_{eq} increasing or decreasing by less than 100–200 μ m over a wider range of IWC sampled during HAIC/HIWC (up to ~3 g m⁻³ in ?) (up to ~3 g m⁻³ in Leroy et al., 2017) than sampled here (up to 1 g m⁻³, Table 2). Profiles of Rayleigh reflectivity and Doppler velocity from a widespread tropical stratiform rain sampled during the Tropical Warm Pool—International Cloud Experiment (TWP-ICE) (Fridlind et al., 2015, their Fig. 11) also appear similar to the 20 May observations (Fig. 17, Doppler velocity not shown here), consistent with generally similar
- 20 stratiform ice size distributions over tropical and 20 May conditions.

In parcel simulations designed to study how relatively narrow mass size distributions of substantial outflow ice could develop within tropical updrfts detraining at roughly -40° C, Ackerman et al. (2015) concluded that copious crystal production at temperatures warmer than roughly -10° C is required. In that study, copious mass concentrated in a relatively narrow mass size distribution centered on an area-equivalent diameter of ~300 μ m required an ice growth time period much longer than

- that associated with homogeneous droplet freezing at 5 or so degrees warmer. Given an updraft profile, increasing number concentrations of ice at temperatures circa -10°C increased the IWC earried to -40°C; any microphysical processes that competed with vapor growth of the ice crystals nucleated near -10°C served to reduce the IWC available for detrainment at -40°C. Conversely, an absence of ice production near -10°C favored microphysical pathways that produced larger hydrometeors that sedimented from updrafts rather than detraining, consistent with simulations of tropical deep convection generally producing too little IWC over stratiform rain areas (e.g., Varble et al., 2014b).-
 - We speculate that similar *ice-updraft* microphysical pathways that determine stratiform outflow ice properties are active in this 20 May case as in the tropical convection observed in many flights during HAIC/HIWC. This can be considered quite surprising since mid-latitude continental convection updrafts are well known to be much stronger than their tropical oceanic counterparts (e.g., Liu and Zipser, 2015). However, it appears that deep convection updrafts may be direct source
- 35 regions for individual outflow ice crystals (especially at upper elevations), consistent with the standard conceptual model of

stratiform ice generation (e.g., Biggerstaff and Houze, 1991), and that ice which becomes stratiform rain may also exhibit rather narrow mass size distributions of relatively small crystals, consistent with an earlier and less complete data set gathered by Airbus (Grandin et al., 2014; Fridlind et al., 2015). The outflow ice size distributions, especially at lower elevations) are are also modified at least in part by aggregation, consistent with layered patterns of ice crystal morphology obtained from dual-

- 5 polarimetric radar particle identification within tropical stratiform precipitation decks (Barnes and Houze, 2016); however. However, contributions to the structure of aged anvil ice size from differences in detrained size with elevation are not clear at temperatures between circa -10 and -20° C in the 20 May case, where signatures of dendritic growth are absent but reflectivity and mean Doppler velocity are generally increasing towards the melting level. In other words, the relative roles of detrained size, differential sedimentation, and aggregation in shaping vertical trends in stratiform ice size distribution are not clear.
- 10 The aircraft engine issues that motivated the HAIC/HIWC campaign are thought to be associated with unexpectedly high IWC for given radar reflectivities (Lawson et al., 1998; Mason and Grzych, 2011; Leroy et al., 2015). Such conditions, which require mass concentrated relatively narrowly in relatively small ice crystals, have been documented at mid-latitudes (Mason and Grzych, 2011). Whether or not they occurred in the 20 May case, it appears likely that a similar set of microphysical processes were active. Furthermore, it appears likely that such processes are not well represented in bin or bulk microphysics
- 15 schemes generally (e.g., Ackerman et al., 2015; Fan et al., 2015; Varble et al., 2014b; Barnes and Houze, 2016). In one observation-driven modeling study, for instance, Zeng et al. (2011) propose an ad hoc "ice enhancement factor in the tropics" to bring simulations into statistical agreement with space-borne radar measurements. Developing tropical cumulus updrafts have also exhibited rapid ice production via ice multiplication that could depend on splinters formed during drop freezing rather than riming, which is not well understood to date and not represented in any commonly used microphysics scheme, and which may have a dominant impact on observed and simulated updraft glaciation rates (Lawson et al., 2015).

In parcel simulations designed to study how relatively narrow mass size distributions of substantial outflow ice could develop within tropical updrafts detraining at roughly -40° C, Ackerman et al. (2015) concluded that copious crystal production at temperatures warmer than roughly -10° C is required. In that study, copious mass concentrated in a relatively narrow size distribution centered on an area-equivalent diameter of ~300 μ m required an ice growth time period much longer than that

- 25 available after homogeneous droplet freezing occurring less than ~ 1 km lower. Given an updraft speed profile, increasing number concentrations of ice at temperatures circa -10° C increased the IWC carried to -40° C; any microphysical processes that competed with vapor growth of the ice crystals nucleated near -10° C served to reduce the IWC available for detrainment at -40° C. Conversely, an absence of ice production near -10° C favored microphysical pathways that produced larger hydrometeors that sedimented from updrafts rather than detraining, consistent with simulations of tropical deep convection generally producing
- 30 too little IWC over stratiform rain areas (e.g., Varble et al., 2014b).

Differences between the simulated 20 May stratiform ice microphysics and observations shown here could arise variously from differences between model and natural ice crystal physical properties (density or structure of crystals), their associated fall speeds, aggregation and vapor growth rates, and the coupling of processes within outflow-generating updrafts, in addition to the ice crystal production rates via primary nucleation and ice multiplication. These factors The NU-WRF biases

35 relative to observations shown here are consistent with the hypothesis that microphysics schemes are missing a key aspect

of an updraft microphysics pathway that can largely determine outflow size, most likely associated with warm-temperature ice multiplication (e.g., Ackerman et al., 2015; Lawson et al., 2015; Ladino et al., 2017). Here we show that NU-WRF biases in stratiform ice mass size distribution are worsened when warm-temperature contributions to ice formation are decreased; Ackerman et al. (2015) find the same in parcel simulations and also demonstrate how biases can be decreased when warm-temperature

5 contributions are substantially increased. In the simulations shown here, we also speculate that gravitational collection of stratiform ice may be too efficient, at least in the mid-troposphere, as evidenced by reflectivity increasing and number concentration decreasing substantially more rapidly than observed between 8 and 6 km (cf. Figs. 10 and 17).

Simulation biases require dedicated efforts to examine, but appear amenable to progress. For instance, in a follow-on study of this 20 May case (van Lier-Walqui et al., 2016), we examine the stratiform column processes in isolation using a one-

10 dimensional modeling approach to make a statistical determination of ice crystal properties and aggregation sticking efficiencies; for that work, the KAZR Doppler spectra are found to be essentially free from turbulence broadening in the quiescent stratiform environment, yielding copious information on ice size distribution variation over large regions of stratiform outflow. If outflow ice size distributions aloft are as similar to those present in detraining updrafts as suggested by HAIC/HIWC data from Darwin (at least for ice not sedimented rapidly within updrafts and prior to any substantial aggregation in the outflow), 15 then the 20 May case study is also well suited to study of updraft microphysics.

Case studies are generally better for model development if they are relatively typical rather than unusual or rare. Based on combined analysis of S-band (NEXRAD) and C-band dual-polarimetric radar signatures over several sites and seasons, it has been noted that the 20 May stratiform ice precipitation lacked the positive differential radar reflectivity commonly found in midlatitude stratiform precipitation containing plate-like and oriented crystals (Williams et al., 2015). Williams et al. (2015) report

- a general absence of robust positive differential reflectivity in the trailing stratiform regions of "vigorous summer squall lines" and attribute that speculatively to the combined effects of irregular ice crystals and stronger electric fields. Strong electric fields have been associated with chain aggregates (e.g., Connolly et al., 2005), which to our knowledge were not profuse over the heavy stratiform rain region in the 20 May case. However, compact and irregular crystals and aggregates are consistent with the available particle images, suggesting that lack of differential reflectivity signature may be indicative of a common stratiform
- 25 microphysics regime across tropical mesoscale convective systems and mid-latitude summer squall lines. <u>The 20 May case</u> therefore could be relatively typical of mid-latitude conditions within such a regime.

Analysis Analyses of dual-polarimetric radar observations could be further systematically employed to identify the environmental conditions associated with stratiform microphysics regimes, assuming some variety exists, as has been suggested by Leroy et al. (2017). In this 20 May case, analysis of dual-polarimetric radar signatures from C-SAPR and KVNX using the quasi-vertical profile

- 30 technique during stratiform rain on 20 May (Kumjian et al., 2016; Ryzhkov et al., 2016) have (Kumjian et al., 2016; Ryzhkov et al., 2016) yielded conclusions generally consistent with the ice properties and microphysical pathways discussed. High specific differential phase in the absence of differential reflectivity enhancements in the elevation range examined here are is consistent with relatively high ice number concentrations and the associated propensity for an active aggregation process despite an absence of dendritic growth. A strong negative gradient in differential reflectivity with elevation above the melting layer is indicative of
- 35 efficient aggregation; we note that this is most intense approaching the melting level. However, the gradient changes sign near

the uppermost elevations sampled by aircraft and examined here (cf. Kumjian et al., 2016, their Fig. 4), so we do not interpret this as conclusive evidence that aggregation is the primary process dominating the ice size distribution shape evolution colder than -10° C. Finally, within the melting layer, very high differential reflectivity and anomalously high backscatter differential phase are another indication of efficient aggregation above the melting layer (Trömel et al., 2014; Ryzhkov et al., 2016), confirmed by in situ observation of aggregates with D_{max} greater than 17 mm just above it (not shown).

Such analyses of dual-polarimetric radar observations could be further systematically employed to identify the environmental conditions associated with stratiform microphysics regimes, assuming some variety exists, as has been suggested by (?). Owing to the leading importance of tropical convection stratiform ice to global circulationand climate phenomena (e.g., Moncrieff et al., 2012), as discussed in Section 1, the dominant microphysics regime seen so far in HAIC/HIWC and some past measurements

- 10 (?)(Leroy et al., 2017), similar to that in the 20 May case, could be among those most important to properly represent in climate models. Aerosol interactions with convection could also be strongly dependent on the microphysics pathways active within a regime. If a warm-temperature ice multiplication mechanism is dominating outflow ice distributions in a manner that cannot be generally reproduced in simulations and is not well understood, it is difficult to confidently assess how or to what degree hygroscopic and ice-nucleating aerosols can be expected to modulate outflow ice properties. For instance, in this
- 15 study we cannot be confident of the relevance of our sensitivity tests for understanding natural convective outflow owing to inadequate baseline fidelity compared with observations. Establishing how typical the 20 May case study is may clarify what other case study conditions could be complementary or more relevant for the purposes of model development. With respect to hygroscopic aerosol, from the perspective of CCN concentrations (2000 cm⁻³ at ~1% supersaturation), the 20 May case is relatively polluted (cf. Fig. 3).
- This is not the first MC3E convection modeling study to conclude that ice microphysics is not yet well-represented across microphysics schemes (e.g., Pu and Lin, 2015). Stratiform outflow from deep convection has also been previously identified as an area where different microphysics schemes in cloud-resolving or convection-permitting simulations produce particularly diverse results (e.g., Morrison et al., 2012; Varble et al., 2014b), with substantial associated impacts on simulated radiative fluxes (e.g., Fridlind et al., 2012; Wang et al., 2015b). Soundly advancing understanding of aerosol effects on deep convection
- 25 requires better establishing and successfully reproducing in simulations the primary microphysical pathways operating under various environmental conditions. Identifying regimes where similar and distinct microphysical conditions can be identified in observations could usefully advance understanding and model development.

6 Code availability

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Aerosol analysis codes are available in Interactive Data Language on request. The ice size distributions reported by Wu and McFarquhar (2016) were processed using the University of Illinois Optical Array Probe Processing Software (UIOPS), which is

open source software available from https://github.com/weiwu5/UIOPS. NU-WRF software is available from http://nuwrf.gsfc.nasa.gov.

7 Data availability

Reported aircraft data are available from the DOE ARM program field campaign archive (http://www.arm.gov/campaigns/mc3e) and the NASA Precipitation Measurement Mission Ground Validation program archive (https://pmm.nasa.gov/science/ground-validation). Based on the raw aircraft microphysical measurements (Delene and Poellot, 2013), the ice number and mass size

- distributions derived as reported by Wang et al. (2015a) and Wu and McFarquhar (2016) are available on request. Ground-based aerosol data are available from the DOE ARM program instrument data stream archive (https://www.arm.gov/data/datastreams).
 The C-SAPR quantitative precipitation estimate is available from the DOE ARM program as an evaluation product (http://www.arm.gov/data/datastreams).
 NEXRAD measurements are available from the US government archive (http://catalog.data.gov/dataset). Specific differential phase and drop size distribution parameters calculated from NEXRAD measurements are available on request. NU-WRF sim-
- 10 ulations are also available on request.

Author contributions. A. Fridlind prepared the aerosol data analysis. X. Li and D. Wu modified NU-WRF to use derived input aerosol specifications, with the assistance of A. Ackerman. D. Wu ran NU-WRF simulations and compared results with in situ microphysics and precipitation observations. M. van Lier-Walqui prepared analysis of observed and simulated radar reflectivity, wind fields, and rain size distribution parameters. G. McFarquhar, W. Wu, X. Dong, and J. Wang provided airborne cloud microphysics measurements. A. Ryzhkov
 and P. Zhang provided rain size distribution parameter retrievals. M. Poellot, A. Neumann, and J. Tomlinson provided airborne aerosol measurements. A. Fridlind and W.-K. Tao coordinated this project with companion studies.

Competing interests. The authors declare that they have no conflict of interest.

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References

- Abdul-Razzak, H. and Ghan, S. J.: A parameterization of aerosol activation: 2. Multiple aerosol types, J. Geophys. Res., 105, 6837–6844, 2000.
- Ackerman, A. S., Fridlind, A. M., Grandin, A., Dezitter, F., Weber, M., Strapp, J. W., and Korolev, A. V.: High ice water content at low radar reflectivity near deep convection-Part 2: Evaluation of microphysical pathways in updraft parcel simulations, Atmos. Chem. Phys., 15,

5

- Altaratz, O., Koren, I., Remer, L. A., and Hirsch, E.: Review: Cloud invigoration by aerosols-Coupling between microphysics and dynamics, Atmos. Res., 140-141, 38-60, doi:10.1016/j.atmosres.2014.01.009, http://dx.doi.org/10.1016/j.atmosres.2014.01.009, 2014.
- Avramov, A., Ackerman, A. S., Fridlind, A. M., van Diedenhoven, B., Botta, G., Aydin, K., Verlinde, J., Korolev, A. V., Strapp, J. W.,
- McFarquhar, G. M., Jackson, R., Brooks, S. D., Glen, A., and Wolde, M.: Toward ice formation closure in Arctic mixed-phase boundary 10 layer clouds during ISDAC, J. Geophys. Res., 116, doi:10.1029/2011JD015910, 2011.
 - Baker, B. and Lawson, R. P.: Improvement in determination of ice water content from two-dimensional particle imagery. Part I: Image-tomass relationships, J. Appl. Meteorol. Clim., 45, 1282-1290, 2006.
- Barnes, H. C. and Houze, Jr., R. A.: Comparison of observed and simulated spatial patterns of ice microphysical processes in tropical oceanic 15 mesoscale convective systems, J. Geophys. Res., 121, 8269-8296, doi:10.1002/(ISSN)2169-8996, http://onlinelibrary.wiley.com/doi/10.
- 1002/2016JD025074/abstract; jsessionid=F5A4904C85FF88672D094C14FD161E3C.f01t01, 2016.
 - Barth, M., Kim, S., Wang, C., Pickering, K., Ott, L., Stenchikov, G., Leriche, M., Cautenet, S., Pinty, J. P., and Barthe, C.: Cloud-scale model intercomparison of chemical constituent transport in deep convection, Atmos. Chem. Phys., 7, 4709–4731, 2007.
- Baumgardner, D., Brenguier, J.-L., Bucholtz, A., Coe, H., DeMott, P., Garrett, T. J., Gayet, J. F., Hermann, M., Heymsfield, A., Korolev, A., Krämer, M., Petzold, A., Strapp, W., Pilewskie, P., Taylor, J., Twohy, C., Wendisch, M., Bachalo, W., and Chuang, P.: Airborne 20 instruments to measure atmospheric aerosol particles, clouds and radiation: A cook's tour of mature and emerging technology, Atmos. Res., 102, 10-29, doi:10.1016/j.atmosres.2011.06.021, 2011.
 - Bigg, E. K.: The supercooling of water, Proc. Phys. Soc. London, B66, 688-694, 1953.
 - Biggerstaff, M. and Houze, Jr., R. A.: Kinematic and precipitation structure of the 10-11 June 1985 squall line, Mon. Weath. Rev., 119, 3034-3065, 1991.
- 25

30

- Connolly, P. J., Saunders, C. P. R., Gallagher, M. W., Bower, K. N., Flynn, M. J., Choularton, T. W., Whiteway, J., and Lawson, R. P.: Aircraft observations of the influence of electric fields on the aggregation of ice crystals, Quart. J. R. Meteorol. Soc., 131, 1695–1712, doi:10.1256/qj.03.217, 2005.
- Cooper, W. A.: Ice initiation in natural clouds, in: Precipitation Enhancement—A Scientific Challenge, edited by Braham Jr, R., p. 171, American Meteorological Society, Boston, MA, 1986.
- Corr, C. A., Ziemba, L. D., and Scheuer, E.: Observational evidence for the convective transport of dust over the central United States, J. Geophys. Res., 121, 1306-1319, doi:10.1002/2015JD023789, 2016.

Crippa, P. and Pryor, S. C.: Spatial and temporal scales of new particle formation events in eastern North America, Atmos. Environ., 75, 257-264, doi:10.1016/j.atmosenv.2013.04.051, 2013.

35 D'Adderio, L. P., Porcù, F., and Tokay, A.: Identification and analysis of collisional breakup in natural rain, J. Atmos. Sci., 72, 3404–3416, doi:10.1175/JAS-D-14-0304.1, 2015.

^{11729-11751, 2015.}

Delene, D. and Poellot, M.: GPM Ground Validation UND Citation Cloud Microphysics MC3E, NASA EOSDIS Global Hydrology Resource Center Distributed Active Archive Center, Huntsville, AL, USA, http://ghrc.nsstc.nasa.gov, doi:10.5067/GPMGV/MC3E/MULTIPLE/DATA201, 2013.

DeMott, P. J., Rogers, D. C., Kreidenweis, S. M., Chen, Y., Twohy, C., Baumgardner, D., Heymsfield, A. J., and Chan, K. R.: The role of heterogeneous freezing nucleation in upper troposphereic clouds: Inferences from SUCCESS, Geophys. Res. Lett., 25, 1387–1390, 1998.

- heterogeneous freezing nucleation in upper troposphereic clouds: Inferences from SUCCESS, Geophys. Res. Lett., 25, 1387–1390, 1998.
 DeMott, P. J., Prenni, A. J., Liu, X., Kreidenweis, S. M., Petters, M. D., Twohy, C. H., Richardson, M. S., Eidhammer, T., and Rogers, D. C.: Predicting global atmospheric ice nuclei distributions and their impacts on climate, Proc. Nat. Acad. Sci., 107, 11217–11222, doi:10.1073/pnas.0910818107, 2010.
 - Dezitter, F., Grandin, A., Brenguier, J.-L., Hervy, F., Schlager, H., Villedieu, P., and Zalamansky, G.: HAIC High Altitude Ice Crystals,
- 10 in: 5th AIAA Atmospheric and Space Environments Conference, American Institute of Aeronautics and Astronautics, Reston, Virginia, doi:10.2514/6.2013-2674, 2013.
 - Donner, L. J., O amp apos Brien, T. A., Rieger, D., Vogel, B., and Cooke, W. F.: Are atmospheric updrafts a key to unlocking climate forcing and sensitivity?, Atmos. Chem. Phys., 16, 12 983–12 992, doi:10.5194/acp-16-12983-2016-supplement, 2016.
- Ekman, A., Wang, C., Ström, J., and Krejci, R.: Explicit simulation of aerosol physics in a cloud-resolving model: Aerosol transport and
 processing in the free troposphere, J. Atmos. Sci., 63, 682–696, 2006.
- Elsaesser, G. S., Del Genio, A. D., Jiang, J. H., and Van Lier-Walqui, M.: An improved convective ice parameterization for the NASA GISS global climate model and impacts on cloud ice simulation, J. Clim., 30, 317–336, doi:10.1175/JCLI-D-16-0346.1, 2017.
- Endo, S., Fridlind, A. M., Lin, W., Vogelmann, A. M., Toto, T., Ackerman, A. S., McFarquhar, G. M., Jackson, R. C., Jonsson, H. H., and Liu, Y.: RACORO continental boundary layer cloud investigations: 2. Large-eddy simulations of cumulus clouds and evaluation with in situ and ground-based observations, J. Geophys. Res., 120, 5993–6014, 2015.
- Fan, J., Liu, Y.-C., Xu, K.-M., North, K., Collis, S., Dong, X., Zhang, G. J., Chen, Q., Kollias, P., and Ghan, S. J.: Improving representation of convective transport for scale-aware parameterization: 1. Convection and cloud properties simulated with spectral bin and bulk microphysics, J. Geophys. Res., 120, 3485–3509, doi:10.1002/(ISSN)2169-8996, 2015.

Fridlind, A. M., Ackerman, A. S., Jensen, E. J., Heymsfield, A. J., Poellot, M. R., Stevens, D. E., Wang, D., Miloshevich, L., Baumgardner,

- 25 D., and Lawson, R. P.: Evidence for the predominance of mid-tropospheric aerosols as subtropical anvil cloud nuclei, Science, 304, 718–722, 2004.
 - Fridlind, A. M., Ackerman, A. S., McFarquhar, G. M., Zhang, G., Poellot, M. R., DeMott, P. J., Prenni, A. J., and Heymsfield, A. J.: Ice properties of single-layer stratocumulus during the Mixed-Phase Arctic Cloud Experiment: 2. Model results, J. Geophys. Res., 112, D24 202, doi:10.1029/2007JD008646, 2007.
- 30 Fridlind, A. M., Ackerman, A. S., Chaboureau, J.-P., Fan, J., Grabowski, W. W., Hill, A. A., Jones, T. R., Khaiyer, M. M., Liu, G., Minnis, P., Morrison, H., Nguyen, L., Park, S., Petch, J. C., Pinty, J. P., Schumacher, C., Shipway, B. J., Varble, A. C., Wu, X., Xie, S., and Zhang, M.: A comparison of TWP-ICE observational data with cloud-resolving model results, J. Geophys. Res., 117, D05 204, doi:10.1029/2011JD016595, 2012.
- Fridlind, A. M., Ackerman, A. S., Grandin, A., Dezitter, F., Weber, M., Strapp, J. W., Korolev, A. V., and Williams, C. R.: High ice water
- 35 content at low radar reflectivity near deep convection Part 1: Consistency of in situ and remote-sensing observations with stratiform rain column simulations, Atmos. Chem. Phys., 15, 11713–11728, doi:10.5194/acpd-15-11713-2015, 2015.

Grandin, A., Merle, J.-M., Weber, M., Strapp, J., Protat, A., and King, P.: AIRBUS flight tests in high total water content regions, in: 6th AIAA Atmospheric and Space Environments Conference, American Institute of Aeronautics and Astronautics, Reston, Virginia, doi:10.2514/6.2014-2753, 2014.

Grim, J. A., McFarquhar, G. M., Rauber, R. M., Smith, A. M., and Jewett, B. F.: Microphysical and thermodynamic structure and evolution

5 of the trailing stratiform regions of mesoscale convective systems during BAMEX. Part II: Column model simulations, Mon. Weath. Rev., 137, 1186–1205, doi:10.1175/2008MWR2505.1, 2009.

Hallett, J. and Mossop, S. C.: Production of secondary ice particles during the riming process, Nature, 249, 26–28, 1974.

Heymsfield, A. J.: Properties of tropical and midlatitude ice cloud particle ensembles. Part I: Median mass diameters and terminal velocities, J. Atmos. Sci., 60, 2573–2591, 2003.

- 10 Houze, Jr., R. A.: Mesoscale convective systems, Rev. Geophys., 42, 1–43, doi:10.1029/2004RG000150, 2004.
- Jensen, M. P., Toto, T., Troyan, D., Ciesielski, P., Holdridge, D., Kyrouac, J., Schatz, J., Zhang, Y., and Xie, S.: The Midlatitude Continental Convective Clouds Experiment (MC3E) sounding network: Operations, processing and analysis, Atmos. Meas. Tech., 8, 421–434, doi:10.5194/amt-8-421-2015, 2015.

Jensen, M. P., Petersen, W. A., Bansemer, A., Bharadwaj, N., Carey, L. D., Cecil, D. J., Collis, S. M., Del Genio, A. D., Dolan, B., Gerlach,

- 15 J., Giangrande, S. E., Heymsfield, A., Heymsfield, G., Kollias, P., Lang, T. J., Nesbitt, S. W., Neumann, A., Poellot, M., Rutledge, S. A., Schwaller, M., Tokay, A., Williams, C. R., Wolff, D. B., Xie, S., and Zipser, E. J.: The Midlatitude Continental Convective Clouds Experiment (MC3E), Bull. Amer. Meteorol. Soc., in press, doi:10.1175/BAMS-D-14-00228.1, 2016.
 - Jiang, J., Zhao, J., Chen, M., and Eisele, F. L.: First measurements of neutral atmospheric cluster and 1–2 nm particle number size distributions during nucleation events, Aerosol Sci. Tech., 45, ii–v, 2011.
- 20 Khain, A. P., Phillips, V., Benmoshe, N., and Pokrovsky, A.: The role of small soluble aerosols in the microphysics of deep maritime clouds, J. Atmos. Sci., 69, 2787–2807, doi:10.1175/2011JAS3649.1, 2012.

Kumjian, M. R., Mishra, S., Giangrande, S. E., Toto, T., Ryzhkov, A. V., and Bansemer, A.: Polarimetric radar and aircraft observations of saggy bright bands during MC3E, J. Geophys. Res., 121, 3584–3607, doi:10.1002/2015JD024446, 2016.

Ladino, L. A., Korolev, A., Heckman, I., Wolde, M., Fridlind, A. M., and Ackerman, A. S.: On the role of ice-nucleating aerosol in the

- 25 formation of ice particles in tropical mesoscale convective systems, Geophys. Res. Lett., 44, in press, doi:10.1002/2016GL072455, 2017. Lance, S., Brock, C. A., Rogers, D., and Gordon, J. A.: Water droplet calibration of the Cloud Droplet Probe (CDP) and in-flight performance in liquid, ice and mixed-phase clouds during ARCPAC, Atmos. Meas. Tech., 3, 1683–1706, doi:10.5194/amt-3-1683-2010, 2010.
 - Lawson, R. P., Angus, L. J., and Heymsfield, A. J.: Cloud particle measurements in thunderstorm anvils and possible weather threat to aviation, J. Aircraft, 35, 113–121, 1998.
- 30 Lawson, R. P., Jensen, E., Mitchell, D. L., Baker, B., Mo, Q., and Pilson, B.: Microphysical and radiative properties of tropical clouds investigated in TC4 and NAMMA, J. Geophys. Res., 115, D00J08, doi:10.1029/2009JD013017, 2010.
 - Lawson, R. P., Woods, S., and Morrison, H.: The microphysics of ice and precipitation development in tropical cumulus clouds, J. Atmos. Sci., 72, 2429–2445, doi:10.1175/JAS-D-14-0274.1, 2015.
- Lebo, Z. J. and Morrison, H.: Effects of horizontal and vertical grid spacing on mixing in simulated squall lines and implications for
 convective strength and structure, Mon. Weath. Rev., 143, 4355–4375, doi:10.1175/MWR-D-15-0154.1, 2015.
- Leroy, D., Fontaine, E., Schwarzenboeck, A., Strapp, J. W., Lilie, L., Delanoë, J., Protat, A., Dezitter, F., and Grandin, A.: HAIC/HIWC field campaign-specific findings on PSD microphysics in high IWC regions from in situ measurements: Median mass diameters, particle size distribution characteristics and ice crystal shapes, Tech. Rep. 2015-01-2087, SAE International, 2015.

- Leroy, D., Fontaine, E., Schwarzenboeck, A., Strapp, J. W., Korolev, A., McFarquhar, G. M., Dupuy, R., Gourbeyre, C., Lilie, L., Protat, A., Delanoë, J., Dezitter, F., and Grandin, A.: Ice crystal sizes in high ice water content clouds. Part II: Statistics of mass diameter percentiles in tropical convection observed during the HAIC/HIWC project, J. Atmos. Oceanic Technol., 34, 118–136, doi:10.1175/JTECH-D-15-0246.1, 2017.
- 5 Lin, Y., Donner, L. J., Petch, J., Bechtold, P., Boyle, J., Klein, S. A., Komori, T., Wapler, K., Willett, M., Xie, X., Zhao, M., Xie, S., Mcfarlane, S. A., and Schumacher, C.: TWP-ICE global atmospheric model intercomparison: Convection responsiveness and resolution impact, J. Geophys. Res., 117, D09 111, doi:10.1029/2011JD017018, 2012.
 - Liu, C. and Zipser, E. J.: The global distribution of largest, deepest, and most intense precipitation systems, Geophys. Res. Lett., 42, 3591–3595, doi:10.1002/(ISSN)1944-8007, 2015.
- 10 Mason, J. G. and Grzych, M.: The challenges identifying weather associated with jet engine ice crystal icing, Tech. Rep. 2011-38-0094, SAE International, Warrendale, PA, doi:10.4271/2011-38-0094, 2011.

Mauritsen, T. and Stevens, B.: Missing iris effect as a possible cause of muted hydrological change and high climate sensitivity in models, Nature Geosci., 8, 346–351, doi:10.1038/ngeo2414, 2015.

May, P. T., Bringi, V. N., and Thurai, M.: Do we observe aerosol impacts on DSDs in strongly forced tropical thunderstorms?, J. Atmos. Sci., 68, 1902–1910, doi:10.1175/2011JAS3617.1, 2011.

- McFarquhar, G. M.: Raindrop size distribution and evolution, Rainfall: State of the Science, pp. 49–60, doi:10.1029/2010GM000971, 2010.
 McFarquhar, G. M. and Cober, S.: Single-scattering properties of mixed-phase Arctic clouds at solar wavelengths: Impacts on radiative transfer, J. Clim., 17, 3799–3813, 2004.
 - McFarquhar, G. M., Timlin, M. S., Rauber, R. M., Jewett, B. F., Grim, J. A., and Jorgensen, D. P.: Vertical variability of cloud hydrometeors
- 20 in the stratiform region of mesoscale convective systems and bow echoes, Mon. Weath. Rev., 135, 3405–3428, doi:10.1175/MWR3444.1, 2007.
 - Meyers, M. P., DeMott, P. J., and Cotton, W. R.: New primary ice-nucleation parameterizations in an explicit cloud model, J. Appl. Meteorol., 31, 708–721, 1992.

- Morrison, H. and Grabowski, W. W.: Modeling supersaturation and subgrid-scale mixing with two-moment bulk warm microphysics, J. Atmos. Sci., 65, 792–812, doi:10.1175/2007JAS2374.1, 2008.
 - Morrison, H. and Milbrandt, J. A.: Parameterization of cloud microphysics based on the prediction of bulk ice particle properties. Part I: Scheme description and idealized tests, J. Atmos. Sci., 72, 287–311, doi:10.1175/JAS-D-14-0065.1, 2015.
- 30 Morrison, H., Thompson, G., and Tatarskii, V.: Impact of cloud microphysics on the development of trailing stratiform precipitation in a simulated squall line: Comparison of one- and two-moment schemes, Mon. Weath. Rev., 137, 991–1007, doi:10.1175/2008MWR2556.1, 2009.

Morrison, H., Tessendorf, S. A., Ikeda, K., and Thompson, G.: Sensitivity of a simulated midlatitude squall line to parameterization of raindrop breakup, Mon. Weath. Rev., 140, 2437–2460, doi:10.1175/MWR-D-11-00283.1, http://journals.ametsoc.org/doi/abs/10.1175/

35 MWR-D-11-00283.1, 2012.

15

25

Peters-Lidard, C. D., Kemp, E. M., Matsui, T., Santanello, Jr., J. A., Kumar, S. V., Jacob, J. P., Clune, T., Tao, W.-K., Chin, M., Hou, A., Case, J. L., Kim, D., Kim, K.-M., Lau, W., Liu, Y., Shi, J., Starr, D., Tan, Q., Tao, Z., Zaitchik, B. F., Zavodsky, B., Zhang, S. Q., and

Moncrieff, M. W., Waliser, D. E., Miller, M. J., Shapiro, M. A., Asrar, G. R., and caughey, J.: Multiscale convective organization and the YOTC virtual global field campaign, Bull. Amer. Meteorol. Soc., 93, 1171–1187, doi:10.1175/BAMS-D-11-00233.1, 2012.

Zupanski, M.: Integrated modeling of aerosol, cloud, precipitation and land processes at satellite-resolved scales, Environmental Modelling & Software, 67, 149-159, doi:10.1016/j.envsoft.2015.01.007, 2015.

- Pöhlker, M. L., Pöhlker, C., Klimach, T., Hrabe de Angelis, I., Barbosa, H. M. J., Brito, J., Carbone, S., Cheng, Y., Chi, X., Ditas, F., Ditz, R., Gunthe, S. S., Kesselmeier, J., Könemann, T., Lavrič, J. V., Martin, S. T., Moran-Zuloaga, D., Rose, D., Saturno, J., Su, H., Thalman,
- R., Walter, D., Wang, J., Wolff, S., Artaxo, P., Andreae, M. O., and Pöschl, U.: Long-term observations of atmospheric aerosol, cloud 5 condensation nuclei concentration and hygroscopicity in the Amazon rain forest – Part 1: Size-resolved characterization and new model parameterizations for CCN prediction, Atmos. Chem. Phys. Disc., pp. 1-54, doi:10.5194/acp-2016-519, 2016.
 - Pu, Z. and Lin, C.: Evaluation of double-moment representation of ice hydrometeors in bulk microphysical parameterization: Comparison between WRF numerical simulations and UND-Citation data during MC3E, Geosci. Lett., pp. 1–9, doi:10.1186/s40562-015-0028-x, 2015.
- 10 Ryzhkov, A., Diederich, M., Zhang, P., and Simmer, C.: Potential utilization of specific attenuation for rainfall estimation, mitigation of partial beam blockage, and radar networking, J. Atmos. Oceanic Technol., 31, 599-619, doi:10.1175/JTECH-D-13-00038.1, 2014. Ryzhkov, A., Zhang, P., Reeves, H., Kumjian, M., Tschallener, T., Trömel, S., and Simmer, C.: Quasi-vertical profiles—A new way to look at polarimetric radar data, J. Atmos. Oceanic Technol., 33, 551-562, doi:10.1175/JTECH-D-15-0020.1, 2016.
- Schumacher, C., Houze, Jr., R. A., and Kraucunas, I.: The tropical dynamical response to latent heating estimates derived from the TRMM 15 precipitation radar, J. Atmos. Sci., 61, 1341-1358, 2004.
- Smith, A. M., McFarquhar, G. M., Rauber, R. M., Grim, J. A., Timlin, M. S., Jewett, B. F., and Jorgensen, D. P.: Microphysical and thermodynamic structure and evolution of the trailing stratiform regions of mesoscale convective systems during BAMEX. Part I: Observations, Mon. Weath. Rev., 137, 1165-1185, doi:10.1175/2008MWR2504.1, 2009.
 - Stevens, R. G., Pierce, J. R., Brock, C. A., Reed, M. K., Crawford, J. H., Holloway, J. S., Ryerson, T. B., Huey, L. G., and Nowak, J. B.:
- 20 Nucleation and growth of sulfate aerosol in coal-fired power plant plumes: Sensitivity to background aerosol and meteorology, Atmos. Chem. Phys., 12, 189-206, doi:10.5194/acp-12-189-2012, 2012.
 - Storer, R. L., Van Den Heever, S. C., and Stephens, G. L.: Modeling aerosol impacts on convective storms in different environments, J. Atmos. Sci., 67, 3904-3915, doi:10.1175/2010JAS3363.1, 2010.
 - Tang, L., Tian, Y., and Lin, X.: Validation of precipitation retrievals over land from satellite-based passive microwave sensors, J. Geophys.
- Res., 119, 4546-4567, doi:10.1002/2013JD020933, 2014. 25

30

- Tao, W.-K., Chen, J.-P., Li, Z., Wang, C., and Zhang, C.: Impact of aerosols on convective clouds and precipitation, Rev. Geophys., 50, RG2001, doi:10.1029/2011RG000369, 2012.
 - Tao, W.-K., Wu, D., Lang, S., Chern, J., Peters-Lidard, C., Fridlind, A., and Matsui, T.: High-resolution NU-WRF simulations of a deep convective-precipitation system during MC3E: Further Improvements and Comparisons between Goddard microphysics schemes and
- observations, J. Geophys. Res., 121, 1278-1305, doi:10.1002/(ISSN)2169-8996, 2016. Testud, J., Oury, S., Black, R., Amayenc, P., and Dou, X.: The concept of "normalized" distribution to describe raindrop spectra: A tool for cloud physics and cloud remote sensing, J. Appl. Meteorol., 40, 1118-1140, 2001.

Thompson, G., Rasmussen, R. M., and Manning, K.: Explicit forecasts of winter precipitation using an improved bulk microphysics scheme. Part I: Description and sensitivity analysis, Mon. Weath. Rev., 132, 519-542, 2004.

Thurai, M., Bringi, V. N., and May, P. T.: CPOL radar-derived drop size distribution statistics of stratiform and convective rain for two 35 regimes in Darwin, Australia, J. Atmos. Oceanic Technol., 27, 932–942, doi:10.1175/2010JTECHA1349.1, 2010.

- Thurai, M., Bringi, V. N., Carey, L. D., Gatlin, P., Schultz, E., and Petersen, W. A.: Estimating the accuracy of polarimetric radar–based retrievals of drop-size distribution parameters and rain rate: An application of error variance separation using radar-derived spatial correlations, J. Hydromet., 13, 1066–1079, doi:10.1175/JHM-D-11-070.1, 2012.
- Trömel, S., Ryzhkov, A. V., Zhang, P., and Simmer, C.: Investigations of backscatter differential phase in the melting layer, J. Appl. Meteorol. Clim., 53, 2344–2359, doi:10.1175/JAMC-D-14-0050.1, 2014.

5

20

- Van Lier-Walqui, M., Fridlind, A. M., Ackerman, A. S., Collis, S., Helmus, J. J., MacGorman, D. R., North, K., Kollias, P., and Posselt, D.: Polarimetric radar signatures of deep convection: Columns of specific differential phase observed during MC3E, Mon. Weath. Rev., 144, 737–758, doi:10.1175/MWR-D-15-0100.1, 2016.
 - van Lier-Walqui, M., Fridlind, A. M., Ackerman, A. S., Williams, C., McFarquhar, G., Wu, W., Li, X., and Korolev, A.: In situ and radar
- 10 Doppler spectrum constraints of ice sticking efficiency and ice properties in a mid-latitude squall line, International Conference on Clouds and Precipitation, Manchester, UK; 29 July, 2016.
 - Van Weverberg, K., Vogelmann, A. M., Lin, W., Luke, E. P., Cialella, A., Minnis, P., Khaiyer, M., Boer, E. R., and Jensen, M. P.: The role of cloud microphysics parameterization in the simulation of mesoscale convective system clouds and precipitation in the Tropical Western Pacific, J. Atmos. Sci., 70, 1104–1128, doi:10.1175/JAS-D-12-0104.1, 2013.
- 15 Varble, A., Zipser, E. J., Fridlind, A. M., Zhu, P., Ackerman, A. S., Chaboureau, J.-P., Collis, S., Fan, J., Hill, A., and Shipway, B.: Evaluation of cloud-resolving and limited area model intercomparison simulations using TWP-ICE observations: 1. Deep convective updraft properties, J. Geophys. Res., 119, 13 891–13 918, doi:10.1002/2013JD021371, 2014a.
 - Varble, A., Zipser, E. J., Fridlind, A. M., Zhu, P., Ackerman, A. S., Chaboureau, J.-P., Fan, J., Hill, A., Shipway, B., and Williams, C.: Evaluation of cloud-resolving and limited area model intercomparison simulations using TWP-ICE observations: 2. Precipitation microphysics, J. Geophys. Res., 119, 13–919–13–945, doi:10.1002/2013JD021372, 2014b.
- Vogelmann, A. M., Fridlind, A. M., Toto, T., Endo, S., Lin, W., Wang, J., Feng, S., Zhang, Y., Turner, D. D., Liu, Y., Li, Z., Xie, S., Ackerman, A. S., Zhang, M., and Khairoutdinov, M.: RACORO continental boundary layer cloud investigations: 1. Case study development and ensemble large-scale forcings, J. Geophys. Res., 120, 5962–5992, 2015.
- Wang, C.: Impact of anthropogenic absorbing aerosols on clouds and precipitation: A review of recent progresses, Atmos. Res., 122, 237–249,
 doi:10.1016/j.atmosres.2012.11.005, 2013.
- Wang, J., Collins, D., Covert, D., Elleman, R., Ferrare, R. A., Gasparini, R., Jonsson, H., Ogren, J., Sheridan, P., and Tsay, S.-C.: Temporal variation of aerosol properties at a rural continental site and study of aerosol evolution through growth law analysis, J. Geophys. Res., 111, D18 203, doi:10.1029/2005JD006704, 2006.

Wang, J., Dong, X., and Xi, B.: Investigation of ice cloud microphysical properties of DCSs using aircraft in situ measurements during MC3E

- 30 over the ARM SGP site, J. Geophys. Res., 120, 3533–3552, doi:10.1002/2014JD022795, 2015a.
- Wang, S., Sobel, A. H., Fridlind, A., Feng, Z., Comstock, J. M., Minnis, P., and Nordeen, M. L.: Simulations of cloud-radiation interaction using large-scale forcing derived from the CINDY/DYNAMO northern sounding array, J. Adv. Model. Earth Syst., 7, 1472–1498, doi:10.1002/2015MS000461, 2015b.
- Westerdahl, D., Fruin, S. A., Fine, P. L., and Sioutas, C.: The Los Angeles International Airport as a source of ultrafine particles and other
 pollutants to nearby communities, Atmos. Environ., 42, 3143–3155, doi:10.1016/j.atmosenv.2007.09.006, 2008.
- Williams, C. R.: Reflectivity and liquid water content vertical decomposition diagrams to diagnose vertical evolution of raindrop size distributions, J. Atmos. Oceanic Technol., 33, 579–595, doi:10.1175/JTECH-D-15-0208.1, 2016.

- Williams, E. R., Smalley, D. J., Donovan, M. F., Hallowell, R. G., Hood, K. T., Bennett, B. J., Evaristo, R., Stepanek, A., Bals-Elsholz, T., Cobb, J., Ritzman, J., Korolev, A., and Wolde, M.: Measurements of differential reflectivity in snowstorms and warm season stratiform systems, J. Appl. Meteorol. Clim., 54, 573–595, doi:10.1175/JAMC-D-14-0020.1, 2015.
- Wong, M., Ovchinnikov, M., and Wang, M.: Evaluation of subgrid-scale hydrometeor transport schemes using a high-resolution cloudresolving model, J. Atmos. Sci., 72, 3715–3731, doi:10.1175/JAS-D-15-0060.1, 2015.

5

20

- Wu, W. and McFarquhar, G. M.: On the impacts of different definitions of maximum dimension for nonspherical particles recorded by 2D imaging probes, J. Atmos. Oceanic Technol., 33, 1057–1072, doi:10.1175/JTECH-D-15-0177.1, 2016.
 - Yang, Q., Easter, R. C., Campuzano-Jost, P., Jimenez, J. L., Fast, J. D., Ghan, S. J., Wang, H., Berg, L. K., Barth, M. C., Liu, Y., Shrivastava, M. B., Singh, B., Morrison, H., Fan, J., Ziegler, C. L., Bela, M., Apel, E., Diskin, G. S., Mikoviny, T., and Wisthaler, A.:
- 10 Aerosol transport and wet scavenging in deep convective clouds: A case study and model evaluation using a multiple passive tracer analysis approach, J. Geophys. Res., 120, 8448–8468, http://onlinelibrary.wiley.com/doi/10.1002/2015JD023647/abstract;jsessionid= 42E3B742AF932D2602260FB45FA767B5.f02t01, 2015.
 - Zeng, X., Tao, W., Matsui, T., Xie, S., Lang, S., Zhang, M., Starr, D., and Li, X.: Estimating the ice crystal enhancement factor in the tropics, J. Atmos. Sci., 68, 1424–1434, doi:10.1175/2011JAS3550.1, 2011.
- 15 Zhang, J., Howard, K., Langston, C., Vasiloff, S., Kaney, B., Arthur, A., Van Cooten, S., Kelleher, K., Kitzmiller, D., Ding, F., Seo, D.-J., Wells, E., and Dempsey, C.: National Mosaic and Multi-Sensor QPE (NMQ) system: Description, results, and future plans, Bull. Amer. Meteorol. Soc., 92, 1321–1338, doi:10.1175/2011BAMS-D-11-00047.1, 2011.
 - Zhu, P., Dudhia, J., Field, P. R., Wapler, K., Fridlind, A., Varble, A., Zipser, E., Petch, J., Chen, M., and Zhu, Z.: A limited area model (LAM) intercomparison study of a TWP-ICE active monsoon mesocale convective event, J. Geophys. Res., 117, D11 208, doi:10.1029/2011JD016447, 2012.



Figure 1. MC3E operations map: around the ARM central facility (small yellow square) are arrayed the X-band radars (white triangle), the sounding array (large yellow pentagon), and the C-SAPR radar (yellow bull's eye symbol). Figure courtesy of Michael Jensen.



Figure 2. KAZR radar reflectivity at the central facility during six case studies that begin on 25 and 27 April, and 1, 20, 23, and 24 May 2011.



Figure 3. Ground-based aerosol properties at the central facility during MC3E: (a) total number concentration (N_a) measured by the HT-DMA; (b) N_a measured by the HTDMA, ground-based CPC, and CCN at maximum supersaturation circa 1%; (c) ratio of N_a measured by HTDMA to that measured by ground-based CPC; and (d) hygroscopicity parameter (κ) measured by the HTDMA at six sizes (intermittent at smallest cut). For each case study, pairs of dashed and dotted vertical lines bound the Citation flight duration and a two-hour pre-rain period, respectively; values above plots are pre-rain period averages, linearly averaged over available sizes for κ .



Figure 4. Airborne out-of-cloud CPC Out-of-cloud measurements of aerosol number concentration available on case study days (black symbols), from UHSAS (top) and average profile airborne CPC (bottom) and median values over km-deep layers (red lines).



Figure 5. Airborne out-of-cloud UHSAS measurements The median of airborne CPC and UHSAS aerosol number concentration available on case study days concentrations within 1-km-deep layers for each MC3E flight, and the ratio of those median values for the seven flights with both instruments (black symbols lines), and average. The median of profile over km-deep layers values at each elevation (blue red lines) - are archived as Supplement 2.

Airborne out-of-cloud UHSAS measurements of aerosol number concentration (black symbols) on three flight days with CPC measurements, and average UHSAS and CPC profiles over km-deep layers (blue and red lines, respectively).

Average CPC and UHSAS aerosol number concentration profiles over km-deep layers (red and blue lines, respectively), the ratio of UHSAS to CPC concentrations on flights with both instruments (thin black lines), and a profile of layer-wise median ratio (thick black line).



Figure 6. Aerosol dry number size distributions $(dN_a/dlogD_a)$ during a two-hour pre-rain period for each case study: measurements reported from the HTDMA during the two-hour pre-rain period (colored solid lines; legend indicates Julian date in UTC), lognormal fits to measurements HTDMA (colored dashed lines), and a number-weighted average (black dashed line, see-; text). For each measurement time are listed the indicates fitted mode-wise number eoncentration concentrations in cm⁻³, geometric mean dry diameter in μ m and standard deviation. Also listed for), and the pre-rain period average is final case study distribution derived from the mode-wise number-weighted average hygrscopicity linear mean of contributing parameters and its hygroscopicity parameter (κ) derived as the number-weighted mean of contributing HTDMA values (black dashed lines and black text; archived with Supplement 1). For In the 20 May casealso shown are distributions with, zero and 8000 cm⁻³ particles in the nucleation mode illustrate BASE and NUCL simulation inputs (dotted black lines).



Figure 7. Derived mode-wise modes and total aerosol number size distributions distribution over km-deep layers (black dotted and dashed lines, respectively) compared with bin-wise mean and median out-of-cloud UHSAS size distributions (red and blue lines, respectively) for the 25 April case study, with sample size (cf. Fig. 4) and total aerosol number concentration (N_a) in cm⁻³.



Figure 8. Time series of precipitation from the BASE simulation, <u>National Mosaic and Multi-Sensor Quantitative Precipitation Estimate</u> (Q2), C-SAPR, and gauge-corrected Q2 averaged over the region <u>bounded sampled</u> by <u>aircraft ~13.9–14.9 UTC indicated by</u> a red rectangle in Fig. 9.

As in Fig. 7 for the 27 April case study. As in Fig. 7 for the 1 May case study.



Figure 9. Surface precipitation rate $(mm h^{-1})$ from Q2 at 14:00 UTC (upper left), gauge-corrected Q2 (upper right, see text), from C-SAPR at 13:40 UTC (lower left), and in the BASE simulation at 13:00 UTC (lower right). Red rectangles bound the Citation aircraft flight legs examined here.



Figure 10. Total ice water content (IWC, top) and ice number concentration (N_i , bottom) derived from aircraft observations (left), as reported by Wang et al. (2015a) see text) and Wu and McFarquhar (2016), and from the BASE simulation (right), within the respective red-bounded geographic regions shown in Fig. 9. Simulated ice is the sum of all ice classes. Observed ice is the sum of all size bins shown in Fig. 11. Box and whisker symbols represent the median, inner half, and 5th and 95th percentiles.



Figure 11. Lee mass size Size distributions of ice mass (left) and number size distributions (right) in four ranges of ice water content (IWC, ranges given in parentheses in units of g m⁻³) as derived from Wang et al. (2015a, red) merger of 2DC and Wu and McFarquhar (2016, blue) HVPS raw data independently by Wang et al. (2015a, 'obs1' in red) and Wu and McFarquhar (2016, 'obs2' in blue). Both are shown as an restrict of poorly established uncertainty. Also shown are size distributions from the BASE simulation (black) at 5.8 km (-10° C) within the respective red-bounded geographic regions shown in Fig. 9. Error bars indicate one standard deviation of values sampled at each size. Simulated ice is the sum of all ice classes at each size. The



Figure 12. As in Fig. 11 except at 6.7 km (-16° C). No IWC greater than 0.8 g m⁻³ was measured at 6.7 km.



Figure 13. As in Fig. 11 except at 7.6 km (-23° C). No IWC greater than 0.4 g m⁻³ was measured at 7.6 km.



Figure 14. 2DC image collage from flight legs above the melting level in the 20 May stratiform outflow region. Three time series examples are given at each elevation. The vertical dimension of each time series is 960 μ m. Here we focus on the top three elevations that are greater than ~1 km above the variable melting level height of ~3.9 km (see text).



Figure 15. Simulated ice number and mass size distributions averaged over the red-bounded geographic region shown in Fig. 9 over 13–14 UTC. Each panel shows the results from four simulations as labeled (see Table 1).



Figure 16. Simulated cloud droplet mixing ratios Horizontally polarized radar reflectivity (Z_{HH} in dBZ) from KVNX radar (left, dotted red circle): (top) example updraft object at ~12 UTC (solid red) among others identified in units of dBZ km (red-enclosed, see text), (middle) movement of example updraft from initial location (solid red) towards intersection with the aircraft sampling location (white-enclosed, see text) projected onto 2-km Z_{HH} at ~14 UTC, and number concentrations (rightbottom) averaged Z_{HH} curtain obtained from column-wise averages over the full domain shown tracked regions from ~12–15 UTC with Citation ascent legs in time and height (white bars) and averaging time used in Fig. 9-17 (white lines). From the AERO simulation (right): (top) identification of a typical updraft object projected onto simulated Z_{HH} at ~11 UTC (solid red) among others identified (red enclosed, see text), (middle) its movement from the identified location (solid red) to intersection with the aircraft sampling location (white-enclosed, see text), (middle) its movement from the identified location (solid red) to intersection with the aircraft sampling location (white-enclosed, see text) projected onto simulated 2-km Z_{HH} at ~11 UTC (solid red) among others identified (red enclosed, see text), (middle) its movement from the identified location (solid red) to intersection with the aircraft sampling location (white-enclosed, see text) projected onto simulated 2-km Z_{HH} at ~13 UTC, and (bottom) Z_{HH} curtain obtained from column-wise averages over 13–14 tracked regions from ~11–14 UTC with mid-point of hour-long averages used in Fig. 17 (white lines).



Figure 17. Reflectivity profiles obtained from one-hour average of reflectivity time series shown in Fig. 16 from KVNX (red line) and AERO simulation times 1, 2, 3 and 4 indicated in Fig. 16 (light to dark grey lines).



Figure 18. Mass-weighted mean diameter (D_m) as a function of time in the AERO simulation and in retrievals averaged over the respective red-bounded geographic regions shown in Fig. 9. Lines indicate median values (see legend). Shaded regions indicate inner half of retrieved values and simulated values at the radar beam mean height. Note offset in time axes (top and bottom) to align approximate timing in observations versus simulations.



Figure 19. Joint histogram of mass-weighted mean diameter (D_m) and collocated precipitation rate in BASE and AERO simulations and retrievals averaged over the red-bounded geographic region shown in Fig. 9 over 8–12 UTC (simulated) or 9–13 UTC (retrieved).

	Prognostic	Nucleation-mode	Homogeneous	
Experiment	N_d	aerosol	freezing only	
BASE	_	_	_	
AERO	\checkmark	_		
NUCL	\checkmark	\checkmark		
HOMF	\checkmark	_	\checkmark	

Table 1. Summary of NU-WRF simulations without and with aerosol input profile and prognostic droplet number concentration (N_d) .

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Table 2. Aircraft-observed ice water content (IWC), ice crystal number concentration (N_i) , and mass median area-equivalent diameter (MMD_{eq}) statistics by elevation, with range given over two derivation methods (see text).

Elevation (km)	Temperature (C)	Mean IWC (g m ⁻³)	Max. IWC (g m ⁻³)	$\frac{\text{Mean } N_i}{(L^{-1})}$	Max. N_i (L ⁻¹)	Mean MMD _{eq} (µm)	Max. MMD _{eq} (µm)
7.6 km	-23	0.21-0.28	0.38-0.43	39–47	78-87	515-530	900-1025
6.7 km	-17	0.44-0.50	0.94–0.96	51–54	84–100	701–704	1025-1200
5.8 km	-10	0.52-0.56	0.89–1.0	45-46	72-80	948–993	1850-2200