

Comparison of ice cloud properties simulated by the Community Atmosphere Model (CAM5) with in-situ observations

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Abstract

Detailed measurements of ice crystals in cirrus clouds were used to compare with results from the Community Atmospheric Model Version 5 (CAM5) global climate model. The observations are from two different field campaigns with contrasting conditions: Atmospheric Radiation Measurements Spring Cloud Intensive Operational Period in 2000 (ARM-IOP), which was characterized primarily by midlatitude frontal clouds and cirrus, and Tropical Composition, Cloud and Climate Coupling (TC4), which was dominated by anvil cirrus. Results show that the model typically overestimates the slope parameter of the exponential size distributions of cloud ice and snow, while the variation with temperature (height) is comparable. The model also overestimates the ice/snow number concentration (0th moment of the size distribution) and underestimates higher moments (2nd through 5th), but compares well with observations for the 1st moment. Overall the model shows better agreement with observations for TC4 than for ARM-IOP in regards to the moments. The mass-weighted terminal fallspeed is lower in the model compared to observations for both ARM-IOP and TC4, which is partly due to the overestimation of the size distribution slope parameter. Sensitivity tests with modification of the threshold size for cloud ice to snow autoconversion (D_{cs}) do not show noticeable improvement in modeled moments, slope parameter and mass weighed fallspeed compared to observations. Further, there is considerable sensitivity of the cloud radiative forcing to D_{cs} , consistent with previous studies, but no value of D_{cs} improves modeled cloud radiative forcing compared to measurements. Since the

25 autoconversion of cloud ice to snow using the threshold size D_{cs} has little physical
26 basis, future improvement to combine cloud ice and snow into a single category,
27 eliminating the need for autoconversion, is suggested.

28

29 **1. Introduction**

30 The parameterization of cloud microphysics plays a critical role in general
31 circulation model (GCM) simulations of climate (e.g., Stephens, 2005). Ice
32 microphysics in particular plays an important role in the global radiative balance
33 (e.g., Mitchell et al., 2008; Zhao et al., 2013), since its parameterization strongly
34 impacts the microphysical and hence radiative properties of ice clouds. It also
35 strongly affects mixed-phase cloud properties, with impacts on precipitation
36 formation and conversion of liquid to ice.

37 Because traditional GCMs are unable to resolve smaller-scale features that drive
38 cloud processes, and because of the need for computational efficiency for climate
39 simulations, the parameterization of microphysics in these models has historically
40 been highly simplified. The first GCMs specified cloud properties diagnostically (e.g.,
41 see review in Stephens (2005)). In later decades GCMs treated one or more species
42 of cloud water, with precipitation water treated diagnostically (Ghan and Easter,
43 1992; Rotstayn, 1997; Rasch and Kristjansson, 1998) or prognostically (Fowler et
44 al., 1996; Posselt and Lohmann, 2008). Several earlier schemes partitioned the total
45 condensate into liquid and ice diagnostically as a function of temperature (Del
46 Genio, 1996). More recently schemes have begun to separately prognose liquid and
47 ice, with an explicit representation of various processes converting water mass

48 between liquid and ice such as freezing, riming, and the Bergeron-Findeisen-
49 Wegener process (Fowler et al., 1996; Lohmann and Roeckner, 1996; Rotstayn et al.,
50 2000; Morrison and Gettelman, 2008; Gettelman et al., 2010). To represent cloud-
51 aerosol interactions and impacts on droplet and ice crystal sizes and hence radiative
52 properties, additional complexity has been added to GCM microphysics schemes to
53 prognose both mass and number mixing ratios of cloud droplets and ice (Ghan et al.,
54 1997; Lohmann et al., 1999; Liu and Penner, 2005; Ming et al., 2007; Morrison and
55 Gettelman, 2008). Thus, there has been a steady march toward increasing
56 complexity of microphysics schemes in GCMs.

57 Nonetheless, several aspects of microphysics remain uncertain. In addition to
58 important issues related to the inability of GCMs to resolve cloud-scale processes,
59 there are underlying uncertainties in the microphysical processes themselves,
60 especially for the ice phase. These uncertainties present challenges, not only for
61 GCMs but also for models of all scales. Much of this uncertainty is rooted in the wide
62 variety of ice particle shapes and types that occur in the atmosphere, leading to a
63 large range of particle fallspeeds, vapor diffusional growth rates, and aggregation
64 efficiencies, to name a few key parameters and processes. Moreover, the
65 parameterization of critical processes like ice nucleation remains uncertain. These
66 uncertainties have important implications for cloud radiative forcing in particular.
67 For example, changes in ice particle fallspeed based on observed ice particle size
68 distributions were found to have a large impact on cirrus coverage and ice water
69 path, with large changes in cloud forcing up to -5 W m^{-2} in the tropics (Mitchell et al.,
70 2008).

71 The representation of ice particle properties in most current microphysics
72 schemes is highly simplified. For example, in the Community Atmosphere Model
73 Version 5 (CAM5, Neale et al., 2010), ice particles are represented as spheres
74 (Morrison and Gettelman, 2008). As in nearly all bulk schemes, ice in CAM5 is
75 separated into different categories representing small ice (cloud ice) and larger ice
76 (snow), each with different bulk densities and fallspeed-size relationships.
77 Conversion between cloud ice and snow is parameterized by “autoconversion” that
78 represents the growth of ice particles through vapor diffusion, aggregation, and
79 riming. However, autoconversion has little physical basis since it does not
80 correspond with a specific microphysical process and results in discrete transition
81 of particle properties from cloud ice to snow. The conversion of cloud ice to snow is
82 tuned in CAM5 by modifying the size threshold for autoconversion, D_{cs} .

83 Another issue is that there is often a lack of self-consistency in ice particle
84 properties in schemes. For example, nearly all bulk schemes (not only in GCMs but
85 in finer-scale models as well) have fallspeed-size relationships that are not directly
86 coupled to particle densities or mass-size relationships, leading to unphysical
87 behavior. For example, increasing particle density can lead to a *decrease* in mass-
88 weighted mean fallspeed because this leads to a smaller mean particle size, while
89 the fallspeed size relationship depends on mean particle size but not density. As
90 pointed out by Mitchell et al., (2011), self-consistency among these relationships is
91 important because of the physical coupling of these parameters. For example, the
92 effective radius and mass-weighted mean fallspeed are both dependent upon mass-

93 size and projected area-size relationships, so that a change in these relationships
94 should be reflected in both the fallspeed and effective radius (Mitchell et al. 2011).

95 Aircraft in-situ observations of ice particles provide an opportunity for detailed
96 testing of assumptions concerning ice particle properties in microphysics schemes.
97 While in situ observations are limited in time and space, statistical comparison with
98 model output, especially in terms of relationships among variables, provides some
99 constraint on microphysics schemes. Here we will investigate how well specific ice
100 microphysical parameters are predicted and diagnosed in CAM5 as compared to in
101 situ observations. While previous work has evaluated ice microphysics in CAM5
102 using aircraft observations (Zhang et al., 2013), we provide a more detailed
103 comparison including several size distribution moments as well as mass-weighted
104 fallspeed for two different field campaigns. Focusing on several parameters is
105 important because these quantities are closely inter-related. We then evaluate
106 results, including cloud radiative forcing, in the context of sensitivity to the
107 autoconversion size threshold D_{cs} – a key tuning parameter for radiative forcing in
108 CAM5. A unique aspect of this study is that we compare several ice microphysical
109 parameters with the same quantities estimated from observations. To our
110 knowledge this has not been done previously for climate models, but is important
111 because it allows us to dig deeper into reasons for biases in key quantities like mass-
112 weighted fallspeed.

113 The paper is organized as follows. In Section 2, the methodology of this study is
114 presented. In section 2.1 the two aircraft campaigns and associated observations
115 that are used in this study are described, while Section 2.2 deals with the model

116 setup. The microphysical parameters that are used for model – observation
117 comparison are detailed in Section 2.3. The comparison results are presented in
118 Section 3. Here, the results using default CAM5 parameters are first discussed in
119 Section 3.1 while a sensitivity study of the ice – snow autoconversion impact on
120 microphysical parameters is included in Section 3.2. Section 4 deals with cloud
121 radiative forcing effects from the autoconversion sensitivity study. Finally, in Section
122 5, a summary and conclusions are presented.

123

124 **2. Methodology**

125 **2.1. Aircraft Measurements**

126 Aircraft measurements of ice crystal size distributions from two different field
127 campaigns are used here for the comparison with model results. These observations
128 are from the Tropical Composition, Cloud and Climate Coupling (TC4) (Toon et al.,
129 2010) mission in 2007 and the Atmospheric Radiation Measurements (ARM) Spring
130 Cloud Intensive Operational Period (IOP) (e.g. Dong et al., 2002) in 2000 (hereafter
131 called “ARM-IOP”).

132 The TC4 campaign was based in the tropics (Costa Rica and Panama, see Fig. 1)
133 and one of the main science goals of TC4 was to improve knowledge of how anvil
134 cirrus form and evolve (Toon et al., 2010). The mostly convectively-generated anvil
135 cirrus were sampled by the NASA DC8 aircraft and the subfreezing periods had a
136 low cloud temperature of ~ -60 °C. Particle size distributions were acquired with a
137 Droplet Measurement Technologies (DMT) Cloud Imaging Probe (CIP) sizing from
138 about 50-1000 μm and a 2D DMT Precipitation Imaging Probe (PIP) sizing from

139 about 200 μm - 1 cm. Averaging was done over 5-second intervals, with a total in-
140 cloud period of about 20 hours ($\sim 15,600$ km). Total condensed water content
141 (TWC, ice plus liquid when present) was measured with a Counterflow Virtual
142 Impactor (CVI) for $\text{TWC} > 0.01 \text{ gm}^{-3}$. Because of the ice shattering issue, we do not
143 use the small particle probe data ($< 75 \mu\text{m}$) and modify the CIP data to account for
144 ice shattering using particle interarrival times (see Field et al., 2006). Liquid water
145 was detected and its content estimated from a Rosemount Icing Probe (RICE).
146 Liquid water encounters were infrequent and have been filtered out of the data set.
147 Further, data were filtered to eliminate updrafts and downdrafts above 1 m/s, and
148 data containing round particles larger than one millimeter in diameter, indicating
149 rain or graupel, were also eliminated.

150 During the TC4 campaign, a 2D-S (Stereo) probe was also flown on the NASA
151 DC8 aircraft (Toon et al., 2010). This probe has a lower size detection limit and
152 better resolution compared to the CIP. Heymsfield et al. (2014) used volume
153 extinction coefficients (σ) to compare 2D-S and CIP+PIP observations against a
154 diode laser hygrometer (DLH) probe, and found that σ from CIP+PIP compared well,
155 while the 2D-S σ were about 50% higher than the DLH σ . They suggested that the
156 reason for the overestimation of 2D-S σ was due to occasional small particles from
157 shattering that were not removed during the post processing procedures. We
158 therefore only use the CIP + PIP observations here.

159 The ARM-IOP was based in the mid-latitudes (Oklahoma, USA, see Fig. 1) and
160 measured a variety of cloud types associated with frontal passages, convection, and
161 synoptically-generated cirrus clouds. Particle size distributions were acquired with

162 a 2D Cloud (2DC) probe sizing from about 50—1000 μm and a 2D Precipitation
163 (2DP) probe. The data were acquired with the University of North Dakota Citation
164 Aircraft. Processing was done as noted above, with averaging over 5-seconds
165 intervals. The total in-cloud time was about 7 hours ($\sim 3,400$ km). TWC
166 measurements were also made with the CVI and liquid water was detected with the
167 RICE probe. All periods of liquid water were removed from the data set, and the
168 same filtering technique mentioned above was used.

169 Images from the two-dimensional probes were analyzed using D_{max} , where D_{max}
170 is the diameter of the smallest circle that completely encloses the projected image.
171 Area ratio, given by the area of the imaged particle divided by the area of the
172 smallest enclosing circle, was used to filter poorly imaged particles from the analysis
173 following the criteria given in Field et al. (2006). A complete discussion of these two
174 data sets, probe evaluations, and processing methods are given in Heymsfield et al.,
175 (2014).

176

177 **2.2. Model setup**

178 The global model from the National Center for Atmospheric Research (NCAR)
179 CAM5 is used in this study. The treatment of clouds in GCMs is typically divided into
180 parameterization of convective clouds and a more detailed microphysics treatment
181 of stratiform clouds. CAM5 includes aerosol effects and detailed microphysics only
182 for stratiform clouds, which includes detrained mass from convective anvils. The
183 stratiform microphysics scheme is an updated version (v1.5) of the 2-moment cloud
184 microphysical scheme of Morrison and Gettelman (2008) and Gettelman et al.

185 (2010). Cloud liquid and ice mass and number mixing ratios are prognosed, while
186 rain and snow mass and number mixing ratios are diagnosed. Particle size
187 distributions are assumed to follow gamma functions. Aerosols affect both cloud
188 droplet and ice crystal number concentrations. The version here is noted as MG1.5,
189 where the major change to the microphysics compared to Gettelman et al. (2010)
190 and relevant to this study is an improvement in how nucleation of ice is applied to
191 increase crystal number: this is now done consistently with the addition of mass
192 from nucleation before microphysical processes are calculated within the time step.

193 For this study, CAM5 was run for six years (from 2000 through 2005), using the
194 first year as spin up time and analyzing the last five years. We used the Atmosphere
195 Model Intercomparison Program (AMIP) style configuration, with prescribed sea
196 surface temperature (annual cycle of the sea surface temperature which repeats
197 every year) and fixed CO₂ concentrations. The resolution was 1.9x2.5°, with 30
198 vertical layers, and global results were output as monthly means. However, over the
199 model grid boxes that overlap the regions from where observations were gathered
200 (Fig. 1), we output instantaneous microphysical parameters and state variables
201 every 3 hours. Note that the grid boxes over the TC4 area are chosen such that they
202 cover mainly ocean due to differences in tuning of the convective microphysics over
203 ocean and land, which can affect radiation and detrained condensate mass feeding
204 into the cloud microphysics. However, including grid boxes over land has a minimal
205 impact and does not change our conclusions (not shown).

206

207 **2.3. Microphysical parameter description**

208 The in situ measurements give detailed information about the size distributions,
209 masses, and projected areas of ice particles, from which mass-weighted terminal
210 fallspeeds and other parameters can be estimated. The mass-weighted terminal
211 fallspeed is an important factor in controlling lifetime of clouds, as well as
212 controlling many other cloud parameters, since this quantity is relevant for
213 sedimentation of ice and snow mass. For comparing the model and measurements,
214 we will introduce a description of the size distribution parameters used here, and
215 then describe the calculation of mass-weighted terminal fallspeeds from the model.

216 **2.3.1. Size distribution parameters**

217 First we note that in CAM5, several output microphysical parameters are given
218 as grid-box means rather than in-cloud values. The grid-box mean takes into
219 account of the fraction of the grid-box that contains condensate (snow and cloud
220 ice). Here, all parameters and equations described are for in-cloud values, unless
221 otherwise stated. In MG1.5 (as in nearly all bulk microphysics schemes), snow and
222 cloud ice are divided into two separate categories, with both size distributions (ϕ)
223 assumed to be represented by gamma functions:

$$224 \quad \phi(D) = N_0 D^\mu e^{-\lambda D}, \quad (1)$$

225 where D is the particle diameter, N_0 is the intercept parameter, μ is the shape
226 parameter and λ is the slope parameter. Currently, the shape parameter is set to
227 zero for both snow and cloud ice, meaning that the distributions are represented by
228 inverse exponential functions.

229 We focus the comparison of modeled and observed size distribution
230 parameters on λ and various size distribution moments (M). Herein we analyze the

231 0th to 5th moments. While number and mass concentrations are proportional to the
 232 0th and 3rd moments in the model, other relevant parameters such as bulk projected
 233 area (relevant for collection of cloud water) and mass-weighted fallspeed depend on
 234 other moments. Thus, we investigate a range of moments for comparison with
 235 observations. The k^{th} moment of the size distribution (M_k^*), where $k > -1$, is found by
 236 integrating the distribution in this form:

$$237 \quad M_k^* = \int_0^\infty N_0 D^k e^{-\lambda D} dD = \frac{N_0 \Gamma(k+1)}{\lambda^{k+1}}, \quad (2)$$

238 where Γ is the Euler gamma function. Here the * indicates moments that are
 239 calculated from integration of the size distribution from 0 to infinity. Thus the 0th
 240 moment, which is equal to the number concentration (N), can be expressed as

$$241 \quad M_0^* = \frac{N_0}{\lambda} = N. \quad (3)$$

242 Snow and cloud ice particles are assumed to be spherical in the model, thus the
 243 mass concentration, q , is proportional to the 3rd moment:

$$244 \quad q = \frac{\pi \rho_p}{6} M_3^* = \frac{\pi \rho_p N_0 \Gamma(4)}{6 \lambda^4} = \frac{\pi \rho_p N_0}{\lambda^4} = \frac{\pi \rho_p N}{\lambda^3}. \quad (4)$$

245 where (3) is used to relate N_0 to N . Here, ρ_p is the bulk density of the particles. Note,
 246 however, that in situ measurements indicate that in reality the mass is closer to the
 247 2nd moment than the 3rd since the particles in nature are generally not spherical. An
 248 expression for λ can be found by rearranging terms in (4):

$$249 \quad \lambda = \left(\frac{\pi \rho_p N}{q} \right)^{1/3}, \quad (5)$$

250 or by using moments:

$$251 \quad \lambda = \left(\frac{6 M_0^*}{M_3^*} \right)^{1/3}. \quad (6)$$

252 Note that the size distribution parameters and moments are derived from the q and
 253 N after they are updated from the microphysical processes, consistent with the
 254 quantities used for the radiation calculations.

255 A key point is that even though cloud ice and snow are divided into separate
 256 categories in MG1.5, the size distributions for each extend from sizes of zero to
 257 infinity (i.e., a complete distribution), as in nearly all bulk microphysics schemes.
 258 Thus, we must combine the cloud ice and snow distributions to derive parameters
 259 for comparing with observations, which do not differentiate between cloud ice and
 260 snow. For λ , this is done by using $N_{si} = N_s + N_i$ and $q_{si} = q_s + q_i$ in (5), where the
 261 subscripts s and i stands for snow and cloud ice, respectively. For ρ_p , we use a mass-
 262 weighted density ($\rho_{p,si}$) that combines the snow ($\rho_{p,s}$) and cloud ice ($\rho_{p,i}$) particle
 263 densities, specified as 250 and 500 kgm^{-3} , respectively. However, there is an
 264 additional complication when calculating mass-weighted quantities because cloud
 265 ice and snow may cover different fractions of the model grid-box. We therefore also
 266 take into account the grid-box snow and cloud ice fractions when mass-weighting
 267 the density. Note that in MG1.5, the fraction of snow (F_s) is, by design, always equal
 268 or greater than the fraction of cloud ice (F_i) because it is assumed that the cloud ice
 269 is a source of snow, while snow can also fall into non-cloudy parts of the grid-box
 270 from above (i.e., the maximum overlap assumption). Furthermore, this is done
 271 regardless of the snow mass mixing ratio, which could in fact be zero. The mass-
 272 weighted snow/ice particle density is therefore given by:

$$273 \quad \rho_{p,si} = \frac{F_i \frac{\rho_{p,i} q_i + \rho_{p,s} q_s}{q_i + q_s} + (F_s - F_i) \rho_{p,s}}{F_s}, \quad (7)$$

274 where the left term in the numerator represents the part of the grid-box that
275 contains cloud ice and snow, while the right term represent the part that only
276 contains snow. The entire expression is then weighed by the fraction of the grid-box
277 that snow and cloud ice covers (which, as stated above is equal to the snow
278 fraction).

279 The λ and N_0 derived from observations were calculated by linear fit in log-
280 linear space to the measured size distributions. The fits were performed using a
281 principal component analysis to minimize the error normal to the fit line. Only size
282 spectra that provided at least 5 size bins with non-zero concentration were
283 considered in order to maintain a reasonable fit. This threshold was generally met
284 in this study when a measurable size distribution existed from 75 μm to at least
285 275 μm in length. When larger particles were present up to 30 bins were included
286 in the fits. The potential fitting errors, and resulting λ and N_0 errors, depend on the
287 number of bins used for the fit, the number of particles measured in each size bin,
288 and the accuracy of the instruments in a particular size range. These conditions are
289 most favorable in broad size distributions with low λ . Due to probe inaccuracies
290 (Strapp et al., 2001) and smaller sample volume for small particles, the errors will
291 be larger for high λ .

292 For determining the moments in (2), the integration over D is from zero to
293 infinity. However, the minimum size of ice crystals considered from the
294 observations is 75 μm . Therefore, for consistency the integration of the modeled
295 moments must be done from 75 μm to infinity to directly compare with the
296 measurements:

297
$$M_k = \int_{D_{min}}^{\infty} N_0 D^k e^{-\lambda D} dD = \frac{N_0 \Gamma(k+1) \Gamma(k+1, D_{min})}{\lambda^{k+1}} . \quad (8)$$

298 Here, $\Gamma(k+1, D_{min})$ is the incomplete gamma function. Note that in the model
 299 calculations, we still use the q and N consistent with integration across the entire
 300 size distribution from zero to infinity instead of from D_{min} to infinity to calculate λ
 301 using (5). This is consistent with the λ derived from observations, which were
 302 calculated by linear fit in log-linear space to the measured size distributions.

303 The measured moments ($M_{obs,k}$) are calculated using

304
$$M_{obs,k} = \sum_{D_{min}}^{D_{max}} N(D) D^k . \quad (9)$$

305 Only integer moments were computed, and physical quantities may not correspond
 306 to the same moment for both the observations and model (for example, ice water
 307 content is proportional to M_3 in the model following the assumption of spherical
 308 particles but is closer to M_2 in the observations). The idea is that each moment
 309 weights a certain portion of the size distribution differently (low moments for small
 310 particles, and high moments for large ones), to allow a simple comparison with the
 311 modeled distributions. Since the measured moments are in a pure form, the
 312 observed and modeled moments can be compared directly.

313 **2.3.2. Mass weighted terminal fallspeed**

314 The mass-weighted terminal fallspeed is another parameter derived from
 315 observations that we will compare with model results. In CAM5, the size dependent
 316 terminal fallspeed (V) is expressed as a power law relation:

317
$$V = aD^b, \quad (10)$$

318 where a and b are empirical constants. In MG1.5, a and b have different values for
 319 ice and snow ($a_i=700 \text{ m}^{1-b}\text{s}^{-1}$, $b_i=1$ following Ikawa and Saito (1991) and $a_s=11.72$
 320 $\text{m}^{1-b}\text{s}^{-1}$, $b_s=0.41$ following Locatelli and Hobbs (1974)). For the comparison, we use
 321 the mass-weighted terminal fallspeed (V_m), which is obtained by integrating the size
 322 distribution in (1), multiplied by V in (10) and weighting by the mass mixing ratio.
 323 The mass-weighted terminal fallspeed can be expressed as:

$$\begin{aligned}
 324 \quad V_m &= \frac{\int_{D_{min}}^{\infty} \left(\frac{\rho_{a0}}{\rho_a}\right)^{\kappa} \frac{\pi \rho_p}{6} a D^{b+3} \phi(D) dD}{\int_{D_{min}}^{\infty} \frac{\pi \rho_p}{6} D^3 \phi(D) dD} = \frac{\left(\frac{\rho_{a0}}{\rho_a}\right)^{\kappa} \frac{a \Gamma(b+4) \Gamma(b+4, D_{min})}{\lambda^{b+4}}}{\frac{\Gamma(4) \Gamma(4, D_{min})}{\lambda^4}} \\
 325 \quad &= \left(\frac{\rho_{a0}}{\rho_a}\right)^{\kappa} \frac{a \Gamma(b+4) \Gamma(b+4, D_{min})}{6 \lambda^b \Gamma(4, D_{min})}. \quad (11)
 \end{aligned}$$

326 Here, ρ_a is the air density, and ρ_{a0} is typical air density at 850mb, which is an air
 327 density factor based on Heymsfield et al. (2007). For ice, $\kappa = 0.35$ (Ikawa and Saito,
 328 1991) and for snow, $\kappa = 0.54$ (Heymsfield et al., 2007). Relating V_m to the size
 329 distribution moments, for cloud ice, V_m is proportional to M_4/M_3 while for snow V_m
 330 is proportional to $M_{3.41}/M_3$.

331 Since the snow and cloud ice categories are not distinguished in the
 332 observations, the modeled snow and cloud ice V_m need to be combined into $V_{m,si}$ in
 333 order to compare with observations. We follow the same formulation as for the
 334 mass-weighted particle density:

$$335 \quad V_{m,si} = \frac{F_i \frac{V_{m,i} q_i + V_{m,s} q_s}{q_i + q_s} + (F_s - F_i) V_{m,s}}{F_s}, \quad (12)$$

336 where $V_{m,s}$ and $V_{m,i}$ are the snow and cloud ice mass-weighted terminal fallspeed
 337 respectively.

338 The mass-weighted fallspeeds from the in-situ observations were computed
 339 using the Best/Reynolds number approach described in Heymsfield and Westbrook
 340 (2010). They included the area ratio of the particles (area of the particle’s projected
 341 area to the area of a circumscribing disk) when determining the mass-weighted
 342 fallspeeds. The projected area is measured directly with the CIP (25 μm resolution)
 343 in TC4 and the 2DC (30 μm resolution) in the ARM-IOP project. Mass is computed
 344 from the power-law relationship $m = 0.00528D^{2.1}$ given in Heymsfield et al. (2010),
 345 which when integrated gave generally good agreement with the total mass
 346 measured by the CVI.

347 2.3.3. Critical Diameter for ice snow autoconversion

348 In MG1.5, the conversion of cloud ice to snow via “autoconversion” is treated by
 349 transferring mass and number mixing ratio from condensate (cloud ice) to
 350 precipitation (snow) based on the critical size threshold, D_{cs} and an assumed
 351 conversion timescale (Morrison and Gettelman, 2008). Expressions for the grid-
 352 scale tendencies are:

$$\begin{aligned}
 353 \quad \left(\frac{\partial q'_i}{\partial t}\right)_{auto} &= -F_i \frac{\pi \rho_i N_{oi}}{6\tau_{auto}} \left[\frac{D_{cs}^3}{\lambda_i} + \frac{3D_{cs}^2}{\lambda_i^2} + \frac{6D_{cs}}{\lambda_i^3} + \frac{6}{\lambda_i^4} \right] e^{-\lambda_i D_{cs}} \\
 354 \quad \left(\frac{\partial N'_i}{\partial t}\right)_{auto} &= -F_i \frac{N_{oi}}{\lambda_i \tau_{auto}} e^{-\lambda_i D_{cs}} \quad (13)
 \end{aligned}$$

355 (Morrison and Gettelman, 2008). Here $\tau_{auto} = 3$ min is the assumed autoconversion
 356 time scale. The quantities with a prime denote the grid-box average values. Since
 357 cloud ice and snow have much different particle densities and terminal fallspeed
 358 parameters (as described in Sections 2.3.1 and 2.3.2), there is a discontinuity of bulk
 359 ice properties after conversion from cloud ice to snow. Although D_{cs} is a size

360 parameter for conversion of cloud ice to snow, not all particles larger than D_{cs} are
361 classified as snow since the cloud ice distribution is complete (meaning that it
362 extends from zero to infinity with significant concentrations larger than D_{cs}). The
363 parameter D_{cs} is chosen rather arbitrary and is one of the main tuning parameters in
364 CAM5: for a given N_i , a larger value for D_{cs} allows higher cloud ice water content
365 before conversion to snow. The default value for D_{cs} in MG1.5 is 250 μm but we will
366 also show results with $D_{cs} = 80, 100, 150, 400$ and 500 μm in Section 3.2, which is
367 similar to the range of D_{cs} tested by Zhao et al. (2013). However, we first describe
368 comparison of the model and observations using the default value of D_{cs} in Section
369 3.1.

370

371 **3. Results**

372 **3.1. Control model – observations comparison ($D_{cs} = 250 \mu\text{m}$)**

373 The measurements were collected mainly in cirrus clouds, but the formation
374 mechanisms generally differed between the TC4 and ARM-IOP cases (Heymsfield et
375 al., 2014). The cirrus in TC4 were mainly anvils associated with deep convection
376 while the cirrus from the ARM-IOP were in situ-generated. We therefore expect to
377 see some differences in the modeled parameters between the two locations, as also
378 seen in the observations (Heymsfield et al., 2014). First we compare the slope
379 parameter λ between model and measurements.

380 **3.1.1. Slope parameter**

381 Figure 2 shows the modeled (red) and measured (black) λ as a function of
382 temperature (which is nearly analogous to height). The solid lines are the geometric

383 mean of the measured or modeled λ . The modeled λ is about a factor of 2 higher
384 than the observed across the entire range of temperatures analyzed. As shown
385 below, this difference between the model results and observations is consistent with
386 both an over-prediction of number concentration of particles larger than 75 μm (N_{75}
387 or M_0) and under-prediction of M_3 .

388 The change in λ as a function of temperature, however, is fairly similar
389 between model and observations. By fitting the data to the exponential equation
390 $\lambda=Ae^{-BT}$, the B coefficient for modeled and measured fitted data for ARM-IOP are,
391 respectively, -0.028 and -0.025, while for TC4 they are -0.03 and -0.032. Note that in
392 Heymsfield et al. (2014), the B coefficient determined for TC4 is -0.0868. In their
393 paper, the size distribution shape parameter (μ) is not assumed to be zero, as we
394 assume in this study. A non-zero μ results in a steeper λ - T relationship and hence B
395 decreases (becomes more negative). For the ARM-IOP case, Heymsfield et al. (2014)
396 found the B coefficient to be -0.0292, which is comparable with our model results.

397 The reason that λ decreases with increasing temperature in the model is
398 mainly due to the change in the ratio of snow to cloud ice mass as temperature
399 increases (or as height decreases). Figure 3 shows that when the modeled λ is
400 calculated individually for snow and cloud ice, λ for snow is fairly constant over all
401 temperatures. Further, the cloud ice category has larger λ values than snow, and
402 larger λ shifts the size distribution to smaller sizes. When considering Figs. 2 and 3,
403 it is clear that cloud ice mass dominates at low temperatures (< -50 $^{\circ}\text{C}$), while snow
404 mass dominates at relatively higher temperatures (> -20 $^{\circ}\text{C}$); the combined λ is
405 closer to λ_i at low temperatures and closer to λ_s at warmer temperatures. This is

406 partly explained by the limited amount of vapor available for growing ice particles
407 at lower temperatures. In addition, more ice particles are typically nucleated at low
408 temperatures, and there is more competition for the available vapor. Thus, mean
409 particle size tends to be smaller at low temperatures, and conversion from cloud ice
410 to snow is limited.

411 **3.1.2. Moments**

412 Figures 4 and 5 show the moments for ARM-IOP and TC4, respectively. Recall
413 that the zero moment (M_0) is the same as the number concentration of particles
414 larger than 75 μm , N_{75} . For ARM-IOP (Fig. 4), M_0 is overestimated by about a factor
415 of 2 between -35 $^{\circ}\text{C}$ and -10 $^{\circ}\text{C}$, while at temperatures lower than -40 $^{\circ}\text{C}$ the model
416 underestimates compared to the measurements. For deposition ice nucleation in
417 CAM5, the parameterization by Meyers et al. (1992) is used at temperatures $> -37^{\circ}\text{C}$
418 (but with constant freezing rate at temperatures $< -20^{\circ}\text{C}$). It has been shown in
419 several papers that this parameterization will typically over-predict ice nucleation
420 by at least an order of magnitude (e.g., Prenni et al., 2007; DeMott et al., 2010). Here
421 the differences in number concentration are much smaller and the assumption of
422 holding the freezing rate constant for deposition nucleation at temperatures $< -20^{\circ}\text{C}$
423 seems to improve prediction of ice nucleation at temperatures warmer than -40 $^{\circ}\text{C}$.
424 At lower temperatures ($< -40^{\circ}\text{C}$), the ice nucleation scheme in CAM5 allows for
425 competition between heterogeneous and homogeneous freezing of deliquescence
426 aerosols (Liu and Penner, 2005). In this scheme, heterogeneous ice nucleation
427 occurs in the form of immersion freezing of dust, and is based upon classical
428 nucleation theory. In certain cases, for in situ generated cirrus, heterogeneous ice

429 nucleation on a few aerosols will start at lower ice saturation than for homogeneous
430 freezing of deliquescence aerosols (e.g. DeMott et al., 1997; Gierens, 2003). These
431 newly formed ice crystals can rapidly deplete the vapor by vapor diffusion, limiting
432 homogeneous aerosol freezing and leading to small ice crystal concentration. If, on
433 the other hand, the number of heterogeneous frozen ice crystals is small enough,
434 homogeneous freezing can still occur and the resulting ice crystal concentration can
435 be fairly high (e.g. Barahona et al., 2009; Eidhammer et al., 2009). It is possible that
436 the prediction of ice crystals from heterogeneous nucleation is too high at lower
437 temperatures, where the classical nucleation theory for immersion freezing is used
438 (e.g. Zhang et al., 2013; Eidhammer et al., 2009). This may be why we see an
439 underestimation of M_0 at temperatures below -40 °C because the competition
440 between heterogeneous and homogeneous nucleation leads to suppression of
441 homogeneous freezing of deliquescence aerosols. Zhang et al. (2013) came to a
442 similar conclusion in their study with CAM5.

443 The measurements only go down to -55 °C, thus we cannot say how well the
444 model performs at lower temperatures. For M_0 at temperatures between -10 and -
445 35°C, both the model and observations show a decrease in M_0 as a function of
446 temperature. The modeled M_0 show a slightly smaller decrease with increasing
447 temperature compared to the observations. The aggregation efficiency specified in
448 the model is rather low (0.1), compared to some estimates at warmer temperatures
449 (near freezing, in conditions with a quasi-liquid layer), or in the dendritic growth
450 regime near -13 to -15° C (Pruppacher and Klett 1997). This could result in a smaller
451 decrease in M_0 with temperature. However, the ice nucleation rate in CAM could also

452 be a source of the large modeled M_0 values. It is not possible based on current
453 observational data to isolate the cause of this bias.

454 The first moment (M_1), which represents the total integrated particle size of
455 the snow and cloud ice population of particles larger than 75 μm , has similar trends
456 to M_0 for ARM-IOP (Fig. 4), with overestimation at higher temperatures ($T > -30$ $^{\circ}\text{C}$)
457 and underestimation at lower temperatures. For the higher moments, M_2 shows a
458 reasonable agreement at temperatures between -25 and -10 $^{\circ}\text{C}$, while there is still an
459 underestimation at lower temperatures. For M_3 , M_4 and M_5 , the model
460 underestimates values over almost the entire temperature regime, while the trend
461 with temperature is in slightly better agreement than for the smaller moments. An
462 underestimation of the higher moments by the model indicates that the
463 concentration of large particles is too low. This could be due to uncertainties in
464 several microphysical processes and parameters including the rather low
465 aggregation efficiency or too slow diffusional growth.

466 When considering the TC4 moments (Fig. 5), the modeled M_0 in general
467 compares better with observations than for ARM-IOP. However, the model still
468 overestimates M_0 , with about a factor of 1.5 over-prediction for temperatures less
469 than -10 $^{\circ}\text{C}$. Note that although the observations and model results for TC4
470 considered here are of stratiform cloud types (anvil cirrus), detrainment plays an
471 important role. The source of the ice crystal number concentration of the detrained
472 condensate comes from an assumed particle radius (25 μm for deep convection and
473 50 μm for shallow convection) and therefore the model does not explicitly calculate
474 ice nucleation from the detrained ice. The slope of M_0 with temperature is again

475 fairly similar between the model and observations. The first moment (M_1) shows a
476 remarkably close agreement between observations and model. However, when
477 considering the higher moments (M_2 , M_3 , M_4 and M_5), the model tends to have lower
478 values compared to observations. Again, the rate of change of the moments with
479 temperature is about the same between the model and observations at
480 temperatures less than $-10\text{ }^\circ\text{C}$. Interestingly, both the model and observations show
481 a slight increase in M_4 and M_5 at around $-30\text{ }^\circ\text{C}$. Overall, the TC4 model results are in
482 better agreements with observations than for the ARM-IOP case.

483 For the moments, we have only considered particles larger than $75\text{ }\mu\text{m}$. For
484 comparison Figs. 4 and 5 also show the moments for the ARM-IOP and TC4 cases
485 from the model when integrating the moments from either $0\text{ }\mu\text{m}$ or $75\text{ }\mu\text{m}$. Clearly
486 the lower moments increase when including all sizes, while the higher moments are
487 not as sensitive to inclusion of small sizes in the integration.

488 The moment comparison gives an illustration of the behavior of the modeled
489 and observed size distributions. However, this comparison does not reveal
490 differences in ice (+snow) water content (IWC) since IWC in the model is
491 proportional to M_3 (assumed spherical shape) while the observed IWC is
492 proportional closer to M_2 . Therefore we also show a comparison of the IWC (Fig. 6).
493 The observed IWC from ARM-IOP is rather insensitive to temperature, while the
494 modeled IWC has a sharp increase with temperature, with smaller than observed
495 values at low temperatures and larger values at relatively high temperatures. For
496 the TC4 IWC, the model and observation have a similar temperature trend but the
497 modeled IWC is slightly lower than the observed IWC.

498 **3.1.3. Mass weighted terminal fallspeed**

499 Fig. 7 shows the mass-weighted terminal fallspeeds (V_m). Fig. 7a compares V_m
500 from the model and observations for both TC4 and ARM-IOP. Figs. 7b (TC4) and 7c
501 (ARM-IOP) are included to show the spread of V_m for the model and observations. In
502 general, V_m determined from the model are somewhat lower than the V_m derived
503 from the measurements. Furthermore, TC4 tends to have higher V_m than ARM-IOP,
504 and this is seen in both the model and observations. The V_m at temperatures above -
505 25 °C (-20 °C) increase sharply in the TC4 (ARM-IOP) observations, while the
506 modeled V_m show less variation with temperature in this region. However, note that
507 there are very few measurements at temperatures above about -20°C for ARM-IOP
508 and TC4. At lower temperatures (<-25 °C), the V_m derived from observations are
509 about a factor of 1.2 higher in the TC4 case compared to the model, but the trend of
510 modeled V_m with temperature is in reasonable agreement with observations. There
511 is less variation of V_m with temperature for the ARM-IOP observations compared to
512 TC4, which is not captured by the model. The increase of V_m with temperature in the
513 model mostly reflects an increase in the ratio of snow to cloud ice, since V_m is
514 inversely proportional to λ while λ does not vary much with temperature for cloud
515 ice and snow individually (see Fig. 3). Thus, the trend of V_m with temperature in the
516 model is mostly controlled by conversion of cloud ice to snow, which influences the
517 mass densities and fallspeeds. As described in Section 4, this conversion has a
518 limited physical basis. Further, the physical reason for the general increase of V_m
519 with temperature in the model is the increase of mean particle size (combined cloud
520 ice and snow) with temperature, consistent with the change in λ with temperature

521 (see Fig. 2). As can be seen in the model, V_m at temperatures less than -60 °C is
522 smaller than 0.3 m/s and small ice dominates in this region.

523 In general, smaller modeled V_m compared to observations is expected since
524 V_m is inversely proportional to λ (see Eq. (11)). Since the modeled λ is larger than
525 measured (see Fig. 2), the modeled V_m should be smaller than those derived from
526 measurements. To illustrate the effect that the factor of 2 in bias for λ has on V_m , we
527 calculated V_m , assuming snow and cloud ice $\lambda = \lambda/2$ (Fig. 8, blue curves). Where
528 snow dominates the total ice mass results are now closer to observations, but where
529 cloud ice is prevalent the V_m are still lower in the model than the observations.

530 The modeled V_m are not only dependent on λ , but also on the assumed power
531 law fallspeed-size parameters for cloud ice and snow in Eq. (10). To test the
532 sensitivity to these parameters, we ran a simulation with a_i and a_s increased by 50%.
533 These results are also shown in Fig. 8 (green curves). At lower temperatures, where
534 cloud ice dominates the total ice mass, V_m does not change much. However, at higher
535 temperatures where snow contributes more significantly to the total mass, V_m
536 increases by about 50%. This is seen in both the ARM-IOP and TC4 cases. For the
537 ARM-IOP case, the increase in a is clearly too large compared to observations, but
538 for the TC4 case, the comparison between model and observations improves (but
539 still has values somewhat larger than those from observations). This may reflect
540 differences in fallspeed parameters between in situ and anvil cirrus as suggested by
541 observations (Heymsfield et al., 2014). However, the increased a parameter in the
542 simulations probably compensates for the over-prediction of λ . Thus, this result
543 does not suggest that a should be increased by up to 50% to obtain better

544 agreement with observations. Rather, it suggests the importance of accurately
545 predicting λ as well as specifying realistic values of the fallspeed parameters.

546

547 **3.2. Cloud ice to snow autoconversion sensitivity tests**

548 As shown in Section 3.1, the model does a reasonable job in predicting some
549 of the size distribution parameters and aspects of the mass-weighted terminal
550 fallspeed. However, there are still clear discrepancies between model results and
551 observations. Moreover, the trends of λ , V_m , and the size distribution moments with
552 temperature in the model are mainly controlled by the partitioning of cloud ice and
553 snow, which is primarily determined by cloud ice to snow autoconversion but has
554 limited physical basis as described below.

555 The critical size for autoconversion of cloud ice to snow, D_{cs} , is one of the
556 major tuning parameters in CAM5. For example, Zhao et al (2013) found that among
557 16 parameters in CAM5, the top of atmosphere radiative forcing responded most
558 efficiently to the tuning of D_{cs} (changes in cloud ice and snow fallspeed parameters
559 and the lower limit on cloud droplet number had smaller impact). When cloud ice is
560 converted to snow, mass and number mixing ratios are moved from one category to
561 another, with discrete changes to particle density and the fallspeed parameters.
562 Cloud ice to snow autoconversion has a limited physical basis since it does not
563 represent a specific microphysical process, and hence the “best” value for D_{cs} is not
564 well established empirically or theoretically. If it is tuned to make the model results
565 comparable with observed cloud radiative forcing, the calculation of other
566 important microphysical parameters might be degraded (Zhang et al., 2013). For

567 example, Zhang et al. (2013) found that using $D_{cs} = 250 \mu\text{m}$ led to close agreement
568 with observations from the SPARTICUS (Small Particles in Cirrus) campaign for the
569 effective particle size, while the total cloud radiative forcing (shortwave +
570 longwave) at the top of the atmosphere was closer to observations when using
571 higher D_{cs} values. However, as shown in Section 3.1, several microphysical
572 parameters that we compared showed rather poor agreement using $D_{cs} = 250 \mu\text{m}$.
573 Here we compare the same parameters as above, but across a range of settings for
574 D_{cs} .

575 We conducted 5 additional simulations with $D_{cs} = 80, 100, 150, 400$ and
576 $500 \mu\text{m}$. We chose a rather wide span of D_{cs} settings since this parameter is not
577 constrained physically. The range of values tested here is similar to Zhao et
578 al. (2013) ($100 - 500 \mu\text{m}$) and larger than in Zhang et al. (2013) ($175 - 325 \mu\text{m}$) and
579 Gettelman et al. (2010) ($150 - 250 \mu\text{m}$). Figure 9 shows λ for all the different D_{cs}
580 values. Overall, none of the values of D_{cs} tested improves the comparison with
581 observation, and hence λ is still too large in the model. The differences between the
582 various runs are not monotonic with changes in D_{cs} and do not show a clear trend
583 with temperature (at some temperatures they are higher than the control run, at
584 some temperatures they are lower, regardless if D_{cs} is higher or lower than in the
585 control run).

586 Figures 10 and 11 show the moments for ARM-IOP and TC4, respectively. For
587 M_0 in the ARM-IOP case there is a clear increase with smaller D_{cs} values. When D_{cs} is
588 increased, there is only a change in M_0 at the highest temperatures (above $-20 \text{ }^\circ\text{C}$).
589 None of the various D_{cs} simulations significantly improve M_0 compared to

590 measurements. For M_1 , the higher values of D_{cs} improve the comparison slightly at
591 temperatures above about $-30\text{ }^\circ\text{C}$. For larger moments the simulations are similar at
592 higher temperatures, but there are some differences at lower temperatures. $D_{cs} = 80$
593 μm compares slightly better at low temperatures for M_1 , M_2 and M_3 , but overall, the
594 moment comparison with observations does not notably improve by varying D_{cs} for
595 the ARM-IOP case.

596 When considering the moments for TC4, the trend of M_0 with temperature
597 shows a slightly different picture than in the ARM-IOP case. Simulations with large
598 D_{cs} produce the largest M_0 at low temperatures. However, this trend reverses at
599 higher temperatures, so that simulations with small D_{cs} have the largest M_0 .
600 Nonetheless, the trend in M_0 with temperature still compares best with
601 measurements when using $D_{cs} = 250\text{ }\mu\text{m}$. For M_1 , the $D_{cs} = 250\text{ }\mu\text{m}$ simulation also
602 compares best with measurements, while for the higher moments, the sensitivity to
603 D_{cs} cases is smaller, with all simulations exhibiting bias compared to observations.

604 It is clear that changes in D_{cs} have a large impact on the mass-weighted
605 terminal fallspeed V_m (Fig. 12). When cloud ice is converted to snow at relatively
606 small sizes ($D_{cs} = 80\text{ }\mu\text{m}$), V_m is almost the same at all temperatures. This is because
607 the particles are mainly snow, and the slope parameter λ for snow is almost
608 constant in this case (see Fig. 3, and note that the $D_{cs} = 80\text{ }\mu\text{m}$ case has a similar
609 temperature trend for snow, only with somewhat higher values).

610 When the conversion from cloud ice to snow occurs at larger sizes
611 ($D_{cs} > 400\text{ }\mu\text{m}$), V_m is small at low temperatures, and only increases to larger values
612 at temperatures above about $-50\text{ }^\circ\text{C}$. At higher temperatures V_m is largest with

613 $D_{cs} = 500 \mu\text{m}$. This occurs because conversion from cloud ice to snow is delayed
614 when D_{cs} is large, so that the mean particle size and hence V_m are relatively large
615 once cloud ice is converted to snow. The higher D_{cs} simulations have a comparable
616 temperature trend for TC4, but V_m are still too low compared to observations. In
617 summary, none of the values of D_{cs} gives a clearly improved comparison with
618 observations for the parameters analyzed here.

619

620 **4. Sensitivity of cloud radiative forcing to D_{cs}**

621 In the previous section we showed that changing D_{cs} has a large impact on the
622 mass-weighted terminal fallspeed and the smaller moments in the size distribution.
623 As changes in D_{cs} impact V_m and other processes (such as Bergeron-Findeisen
624 process, i.e. the conversion of liquid to ice through ice depositional growth), the
625 liquid and ice water paths change as well as the effective radii. These changes in
626 turn impact the cloud radiative forcing consistent with previous studies (Gettelman
627 et al., 2010; Zhang et al., 2013; Zhao et al., 2013). These studies used MG
628 microphysics in CAM5 and showed that, globally, it is the longwave cloud forcing
629 that is most influenced by changes to D_{cs} . Gettelman et al. (2010) and Zhao et al.
630 (2013) also showed that the changes in total cloud forcing (longwave plus
631 shortwave) varies in magnitude as a function of latitude, with the mid-latitudes
632 experiencing the largest changes in terms of sensitivity to D_{cs} . Moreover, as
633 previously stated, Zhang et al. (2013) found that among 16 different parameters,
634 changes to D_{cs} had the largest impact on top of the atmosphere radiation. In our
635 simulations, with regard to changes to D_{cs} , we come to some of the same

636 conclusions. Here we also show which microphysical variables have the most impact
637 on the cloud radiative forcing through changes in D_{cs} .

638 Figure 13 shows how the zonally-averaged shortwave and longwave
639 radiative cloud forcing (SWCF and LWCF respectively) is affected by changes to D_{cs}
640 as a function of latitude. The LWCF has an increase with increasing D_{cs} over all
641 latitudes, while the SWCF has opposite effects between mid-latitudes and tropics.
642 The cloud radiative forcing is dependent upon the ice and snow effective radii
643 (proportional to M_3/M_2) as well as ice and snow water contents (proportional to M_3
644 in the model), in addition to cloud droplet effective radius and cloud liquid water
645 content. To investigate which quantities are the major controlling factors in the
646 sensitivity of cloud radiative forcing to D_{cs} , we plot several key zonally-averaged
647 quantities in Fig. 14. Figure 14a, b, c and d shows the combined cloud ice plus snow
648 water path, cloud liquid water path, snow water path and cloud ice water path,
649 respectively (note that the water path is the vertical integral of the water content).
650 Figure 14e shows the effective radii of cloud ice and snow, while Fig. 14f shows the
651 effective radius of cloud droplets.

652 As D_{cs} increases, less cloud ice is converted to the snow category
653 monotonically as is shown in Figs 14c and 14d at mid-latitudes. There is limited
654 impact on the total cloud ice plus snow water path in the mid latitudes since
655 changes in the snow and cloud ice water paths have opposing effects (Fig. 14a). In
656 the tropics, on the other hand, there is some increase in the combined snow and
657 cloud ice water path, since there is a slight increase in snow water path along with
658 an increase in ice water path with increasing D_{cs} (see Figs 14c and 14d). If TC4 is

659 representative of the zonally-averaged snow water path in the tropics, based on the
660 analysis presented in Section 3, we suspect that the higher snow water path with
661 larger D_{cs} is due to increases in snow at relatively high temperatures, i.e., lower
662 altitudes (not shown). However, it is clear from all the parameters shown in Fig. 14
663 that the change in cloud ice water path is one of the main controlling factor in the
664 changes to LWCF (Fig. 13b). For example, details such as the clustering of cloud ice
665 water path for the simulations with D_{cs} less than 250 μm are closely mirrored in
666 LWCF.

667 SWCF is also a function of liquid, snow and cloud ice water paths and
668 effective radii. Figure 13a shows that the response of SWCF to changes in D_{cs} has
669 opposite effects in mid-latitudes compared to the tropics. By comparing Fig. 13a
670 with Fig. 14, it is clear that the cloud liquid water path is the primary controlling
671 factor in explaining the SWCF changes. Snow water path has some of the same
672 variations as cloud liquid water path with D_{cs} (higher water path in tropics with
673 increasing D_{cs} and lower in the mid-latitudes). However, overall changes in the cloud
674 liquid water path with D_{cs} mirror changes in SWCF closer than changes in snow
675 water path. Thus, the shortwave cloud forcing response appears to be mostly
676 explained by indirect impacts of D_{cs} on liquid water path rather than directly
677 through changes in the cloud ice and snow radiative properties. Furthermore, there
678 is little correspondence between changes in the effective radii of snow, cloud ice, or
679 liquid and SWCF with modification of D_{cs} . This is seen in Figs. 13-14, which show
680 little correspondence between changes in effective radii and SWCF, compared to
681 changes in liquid water path.

682 Finally, we show the zonally-averaged total cloud radiative forcing (TCF,
683 SWCF+LWCF) in Fig. 15. Overall, the magnitude of TCF decreases with increasing D_{cs} ,
684 moving the modeled TCF closer to CERES observations. However, the magnitude of
685 the modeled TCF is still over-estimated compared to the observations in the tropics
686 and into the mid-latitudes. Only in a small window in the southern hemisphere (-60
687 to -70°) do D_{cs} cases $\leq 250 \mu\text{m}$ compare well with the observations. In summary,
688 variations in D_{cs} impose a relatively large change in cloud radiative forcing, but none
689 of the values tested here notably improve the modeled cloud radiative forcing
690 compared to observations.

691

692 **5. Summary and conclusions**

693 We have presented a GCM – observational comparison of important ice
694 microphysical parameters, such as the size distribution slope parameter, moments
695 of the snow and ice particle size distributions, and mass-weighted fallspeed. These
696 parameters are closely linked to the direct radiative forcing of cloud ice and snow,
697 and also have important indirect effects by impacting cloud liquid. It is therefore
698 crucial to obtain a good agreement between model and observations of snow and ice
699 size distributions parameters in the model, in order to conduct climate impact
700 studies.

701 We used CAM5 with MG1.5 microphysics for this study. The aircraft
702 observations were collected during TC4 (tropical anvil cirrus) and ARM-IOP (mid-
703 latitude continental in-situ generated cirrus)

704 Our results with the control simulation ($D_{cs} = 250 \mu\text{m}$) indicate that the slope
705 parameter in MG1.5 is about a factor of two higher than that determined from
706 observations. This is true for both regions. However, the trend with temperature is
707 comparable. For the moments, the model generates about a factor of two larger ice
708 crystal number concentrations (ice plus snow, and for particles larger than $75 \mu\text{m}$)
709 at relatively high temperatures, while the ARM-IOP case indicate that the model
710 generates too few crystals at low temperatures. We hypothesize this results from
711 too many ice crystals formed heterogeneously at temperatures $< -37^\circ\text{C}$, so that the
712 competition between homogeneous and heterogeneous nucleation does not allow
713 for homogeneously formed ice crystals. This is consistent with Zhang et al. (2013),
714 who used SPARTICUS data in their evaluation of ice nucleation schemes in CAM5.
715 The first moment has the best comparison between model and observations, while
716 higher moments are generally under-predicted. The mass-weighted fallspeeds were
717 about a factor of 1.2 lower in the model compared to observations.

718 In MG1.5, as in nearly all bulk microphysics schemes, ice is separated into
719 cloud ice and snow categories with different particle densities and fallspeed
720 parameters. The size threshold for conversion of cloud ice to snow, D_{cs} , is one of the
721 main tuning parameters for cloud radiative forcing in CAM5. We conducted five
722 additional simulations covering a large range of D_{cs} values. However, none of these
723 simulations notably improved the comparison between the model and observations
724 of the size distribution parameters and mass-weighted fallspeed. We note that the
725 snow is determined diagnostically in MG1.5 and therefore is assumed to be in steady
726 state within a time step (i.e. the source and sink terms are equal to what is removed

727 due to fallout). In this case, snow still undergoes processes such as sublimation,
728 melting and riming. However, if snow was determined prognostically the steady
729 state assumption no longer applies and there is memory of snow mass and number
730 mixing ratios across time steps (work is underway to modify CAM5 microphysics to
731 include prognostic rain and snow). Thus, there could be differences in the sensitivity
732 to D_{cs} in a prognostic snow scheme compared to the diagnostic snow scheme
733 examined here.

734 The changes to D_{cs} also have large impacts on cloud radiative forcing. Changes
735 in the total ice water path (cloud ice plus snow) with D_{cs} were fairly small, especially
736 in mid-latitudes, because of opposing effects on the cloud ice and snow water paths.
737 However, the longwave cloud radiative forcing is primarily influenced by cloud ice
738 water path and hence the increase in cloud ice water path with increasing D_{cs} led to
739 an increase in longwave cloud forcing. On the other hand, changes in the shortwave
740 cloud forcing were mostly influenced by changes in cloud liquid water path
741 indirectly driven by changes in D_{cs} . Overall, there was a noticeable change in total
742 cloud forcing when increasing D_{cs} from 250 μm , especially in the mid-latitudes. For
743 example, there was a 10 Wm^{-2} increase in total cloud radiative forcing in the
744 southern mid-latitudes when D_{cs} was increased from 250 μm to 400 μm . The
745 changes were somewhat smaller in the mid-latitudes when decreasing D_{cs} . None of
746 the values of D_{cs} tested here led to notable improvement in the distribution of cloud
747 radiative forcing.

748 Large sensitivity of the size distribution parameters and moments and mass-
749 weighted fallspeed, as well as cloud radiative forcing, to D_{cs} motivates additional

750 work to improve how ice particle properties change with increasing particle size.
751 This is especially true given that no particular value of D_{cs} led to substantially better
752 overall results. Furthermore, the autoconversion of cloud ice to snow, using the
753 threshold size D_{cs} , has little physical basis. One possible approach is to combine
754 cloud ice and snow into a single category such as proposed by Morrison and
755 Grabowski (2008), entirely removing the need for autoconversion. Ice particle mass-
756 size and projected area-size relationships (from which fallspeed-size relationship
757 would be derived) would then vary across the particle size distribution to represent
758 the different properties of small and large ice particles specified from observations.
759 This would lead to some complication because simple analytic integrations, for
760 example for the mass-weighted fallspeed, are no longer possible. However,
761 numerical integration can be performed with values stored in a lookup table (as
762 used by Morrison and Grabowski [2008]), or with simplified expressions based on
763 curve-fitting. Future work will explore these ideas.

764

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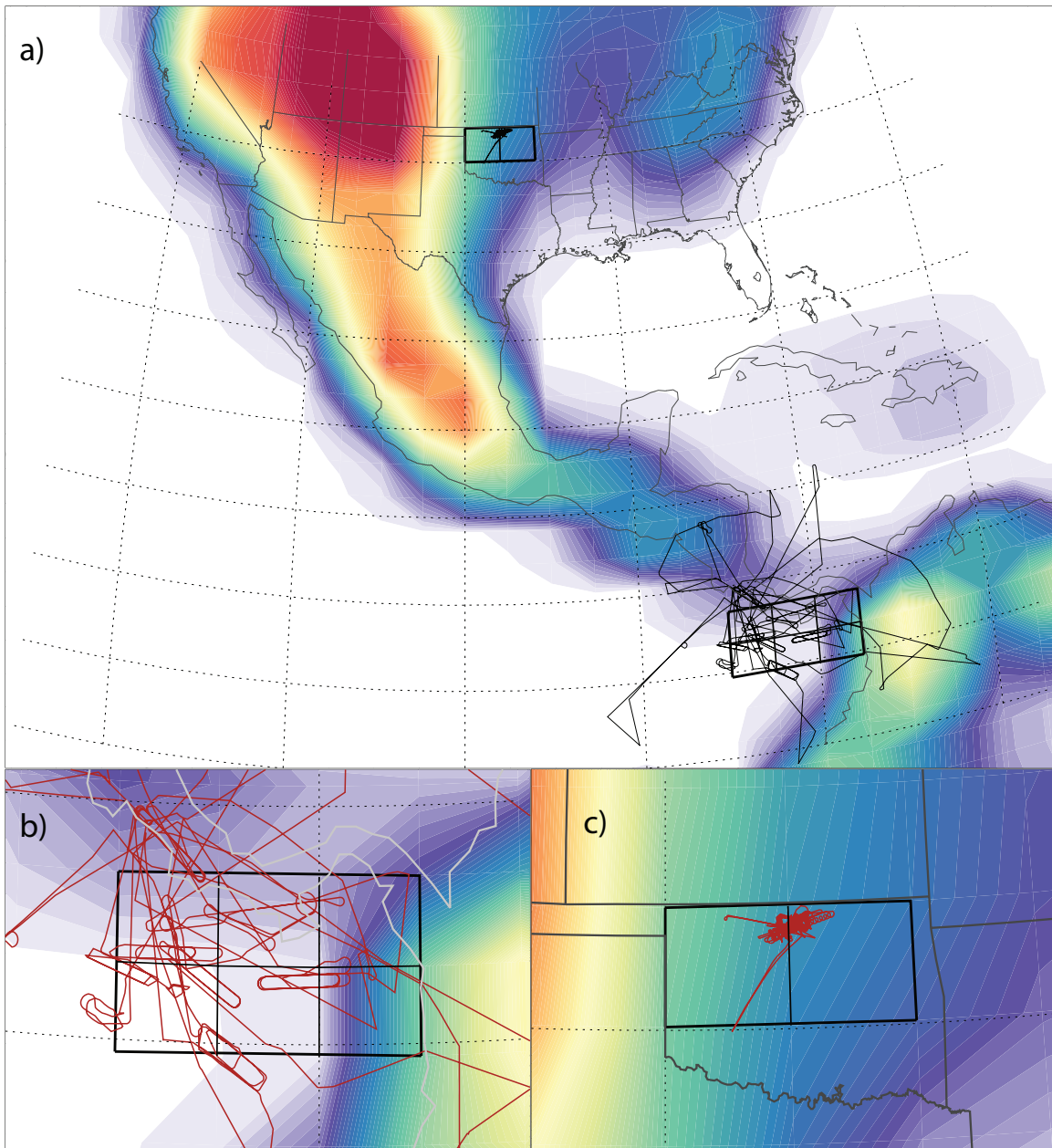
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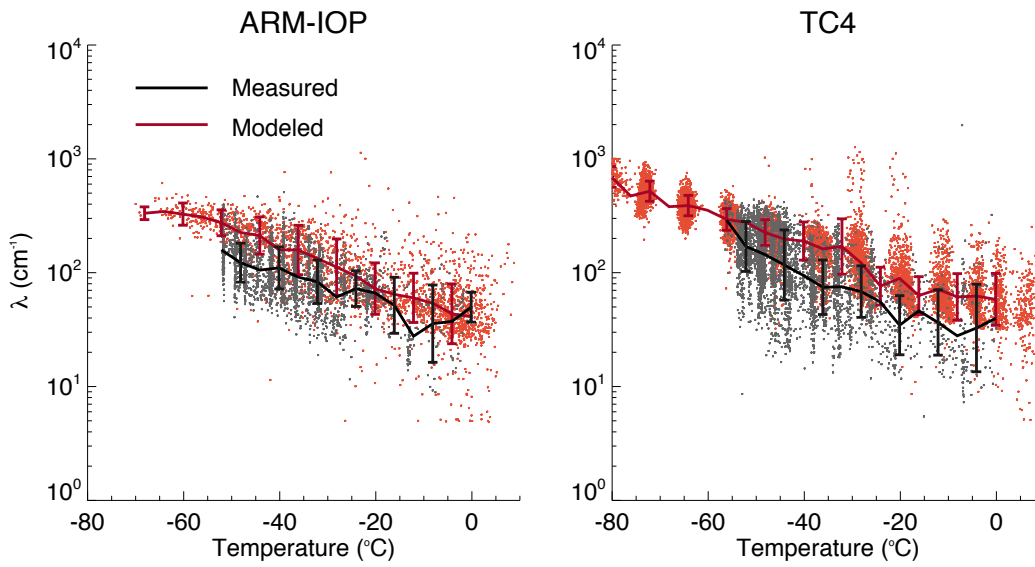
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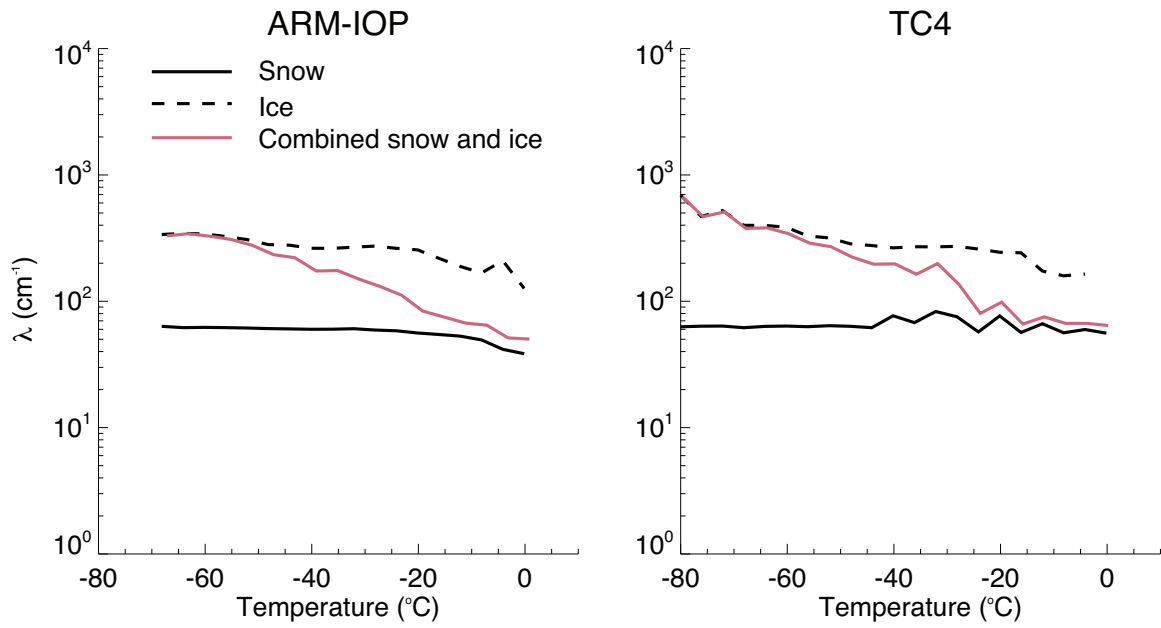
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Fig. 1. a) Location of ARM-IOP and TC4, along with model grid boxes. b) TC4 with a more detailed view of the flight tracks. c) Same as in b) but for ARM-IOP.



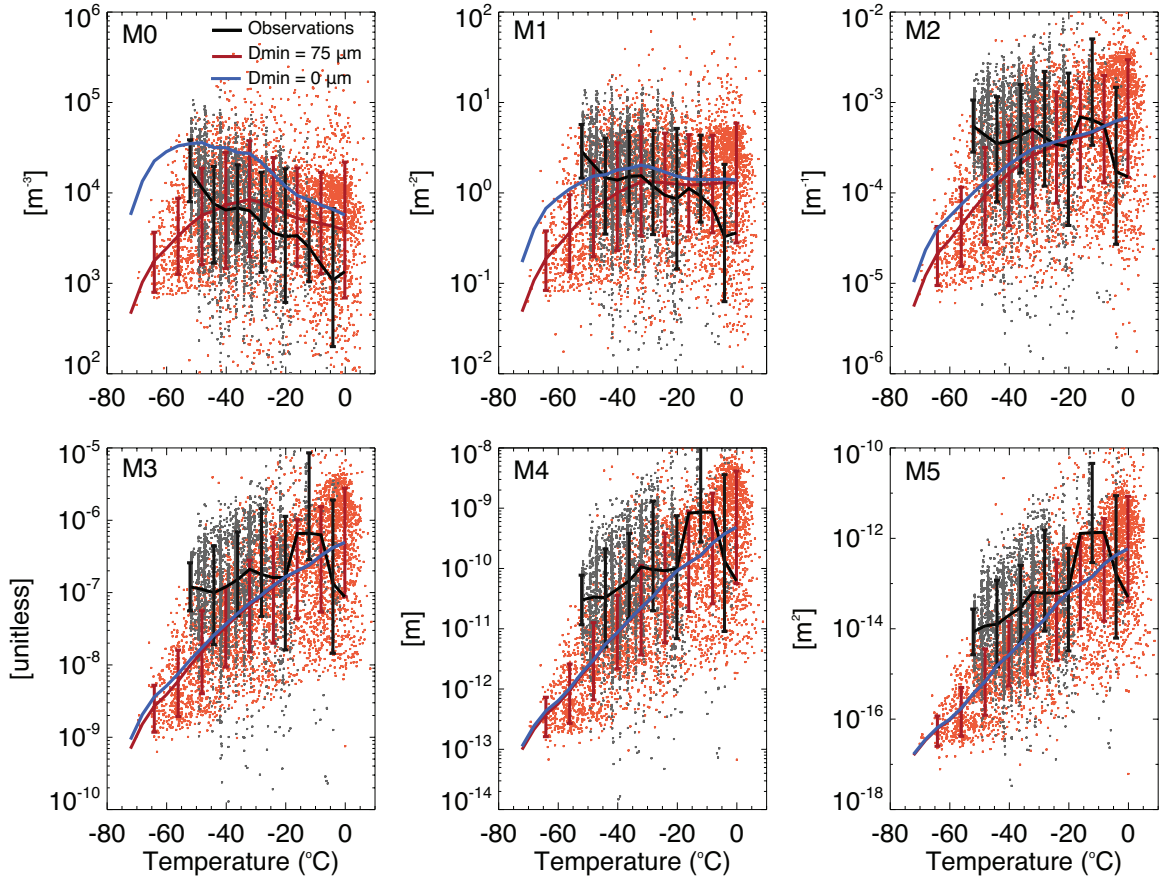
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Fig. 2. Slope parameter, modeled (red) and measured (black) for ARM-IOP and TC4. The lines are the geometric mean, the dots represents a fraction of the measurements and modeled values, while the vertical bars represents the geometric standard deviation.



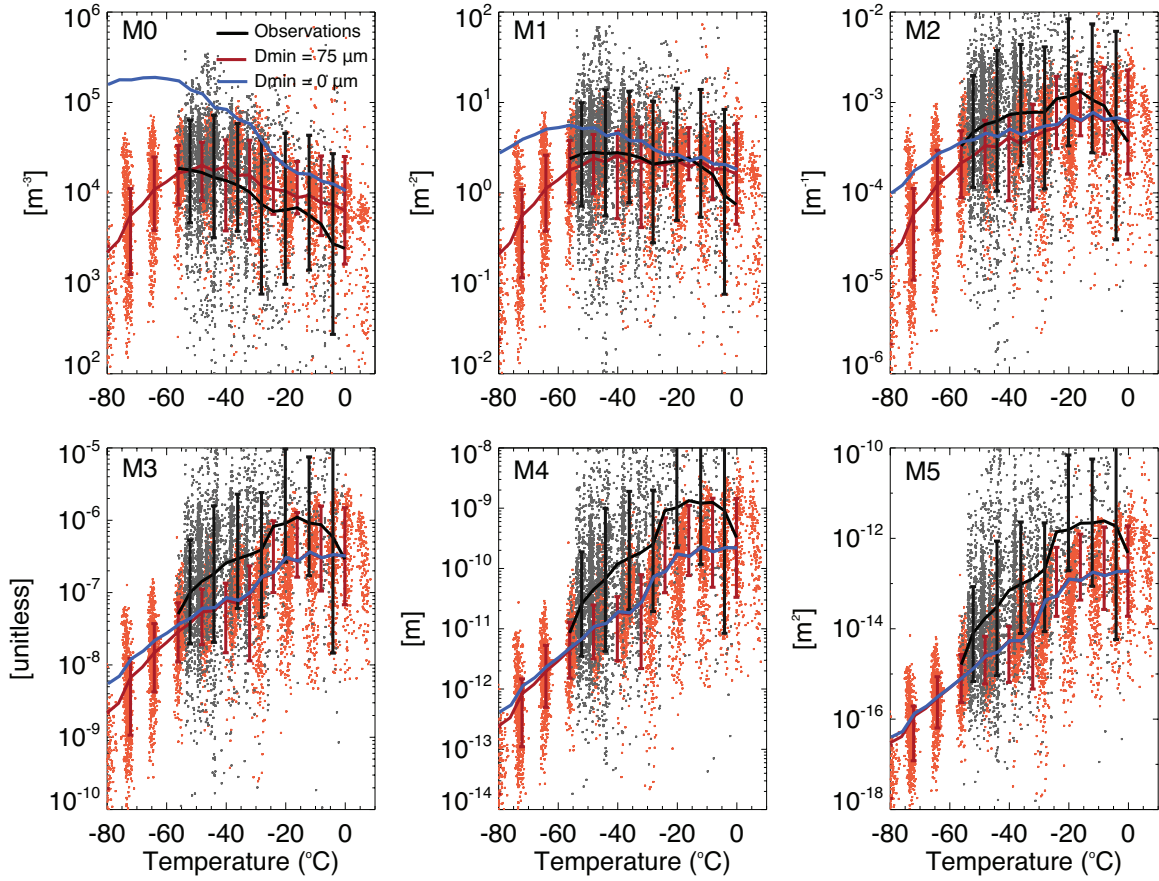
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Fig. 3. Modeled slope parameter, for ice and snow individually. Also shown is the combined snow and ice slope parameter, as shown in Fig. 2.



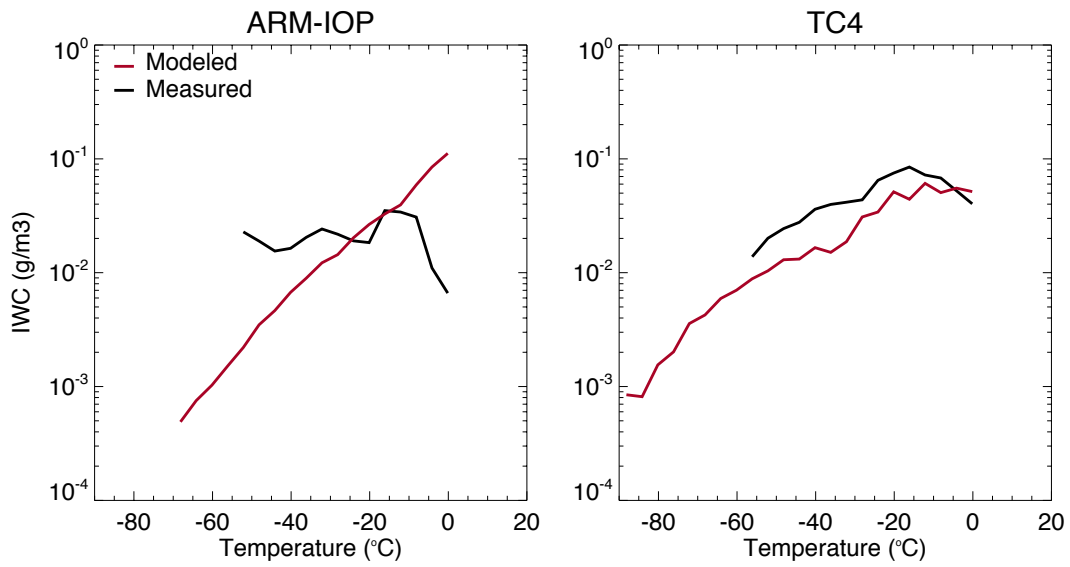
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Fig. 4. Moments from ARM-IOP (black: measurements. red: model integrated from 75 μm, blue: model integrated from 0 μm). Lines are geometric mean, dots represents a fraction of the measurements and model results, while vertical lines are the geometric standard deviation.



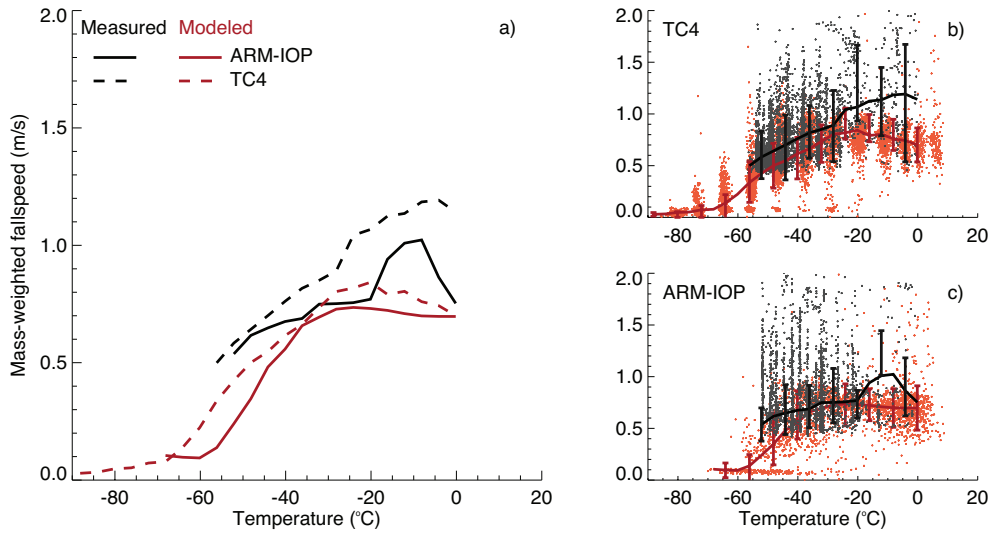
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Fig. 5. Same as Fig 4, but for TC4.



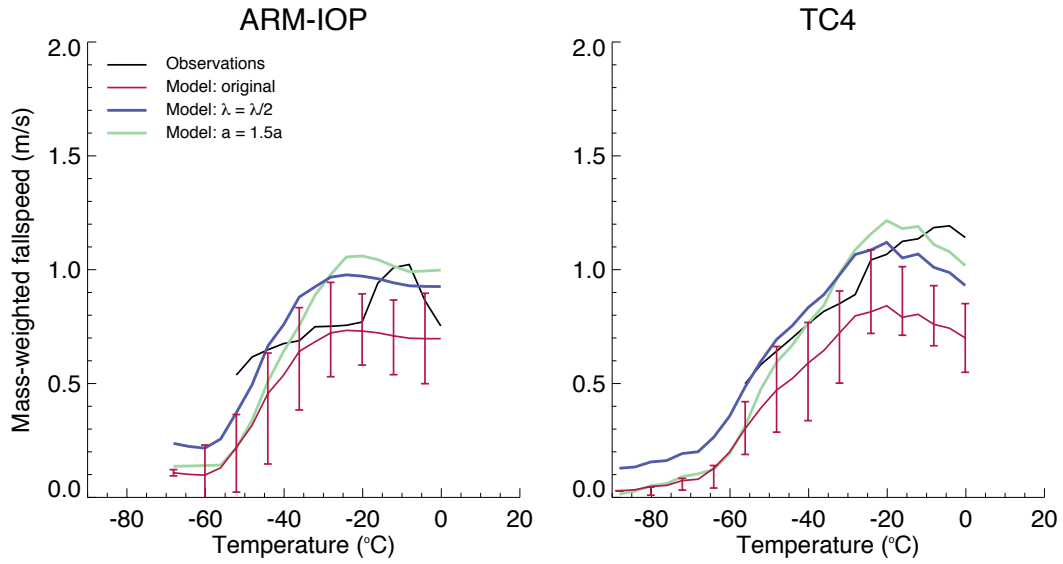
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 999 **Fig. 6.** IWC from model (red) and observations (black)

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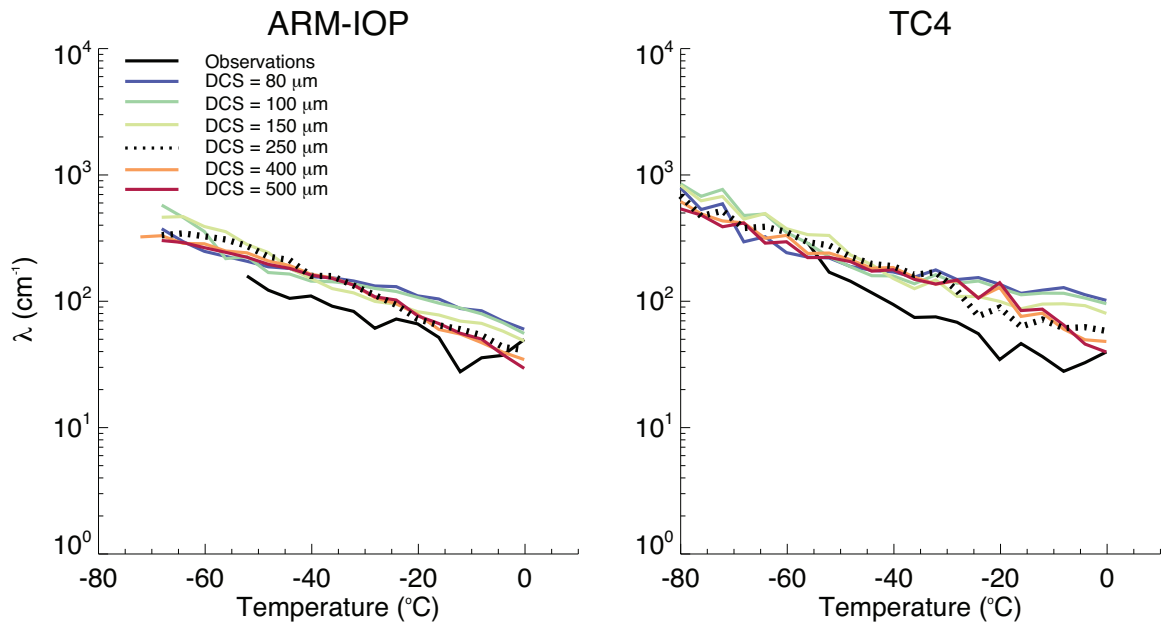
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Fig. 7. Mass weighted terminal fall speed. a) Measured and modeled V_m for ARM-IOP and TC4 for comparing fallspeeds between campaigns. b) and c) Mass weighted fall speeds for showing the measurement and modeling spread.



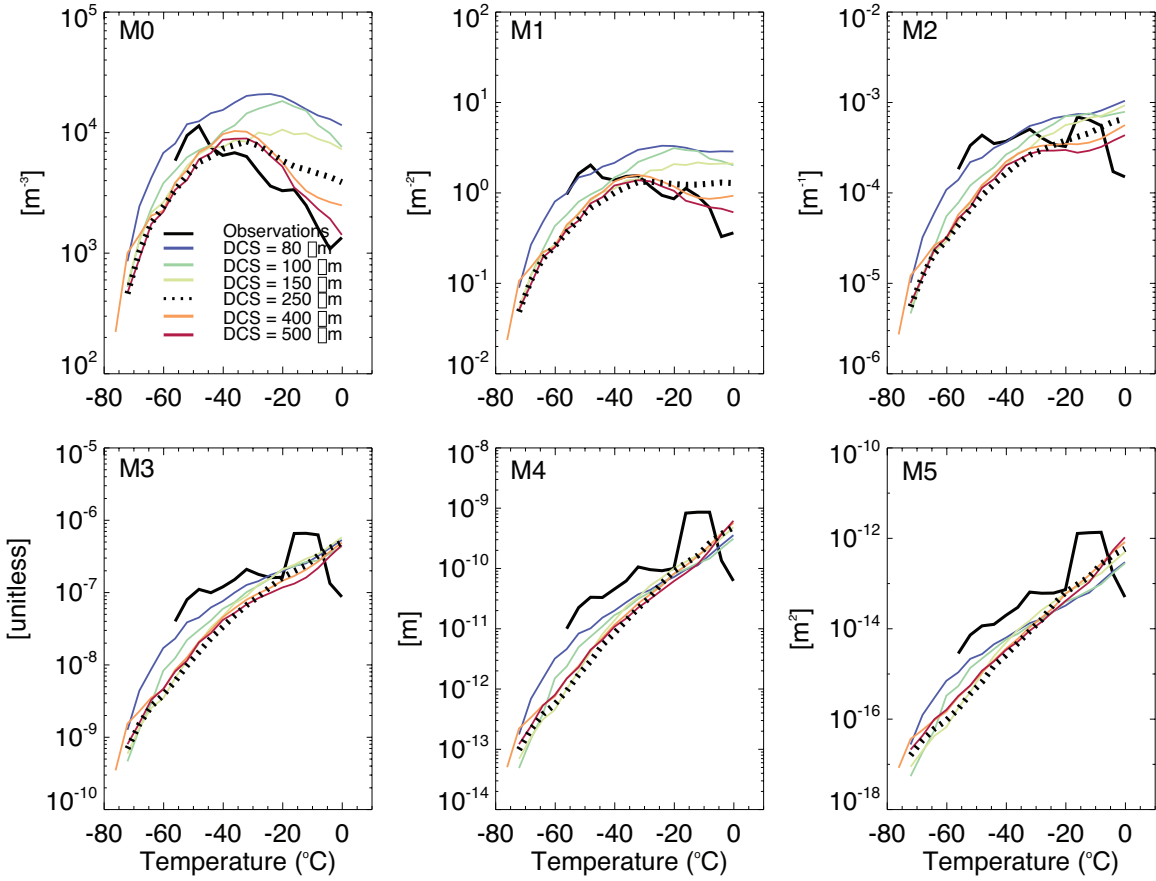
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Fig. 8. Mass weighted terminal fall speed with snow and cloud ice $\lambda = \lambda/2$ (blue), and a_i and a_s increased with 50% (green).



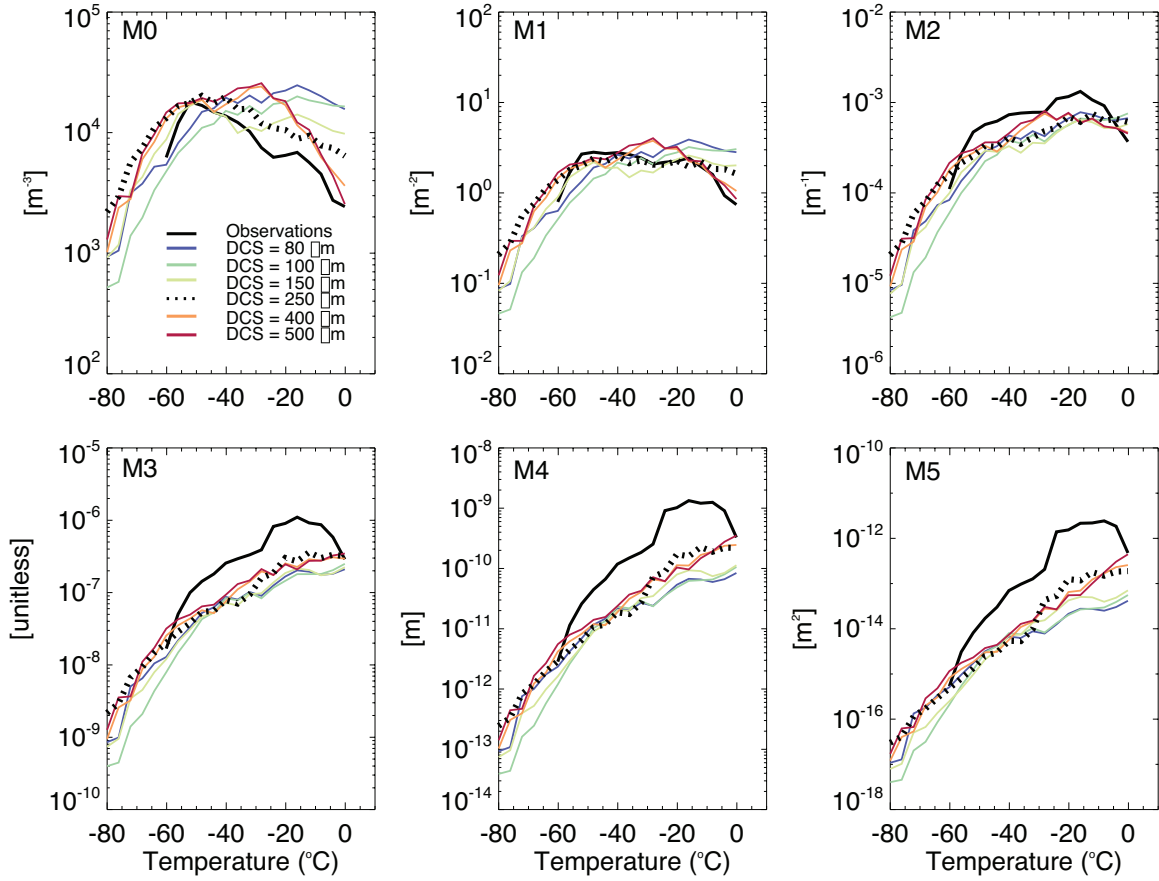
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Fig. 9. Same as Fig. 2, but with simulations using different D_{CS} values.



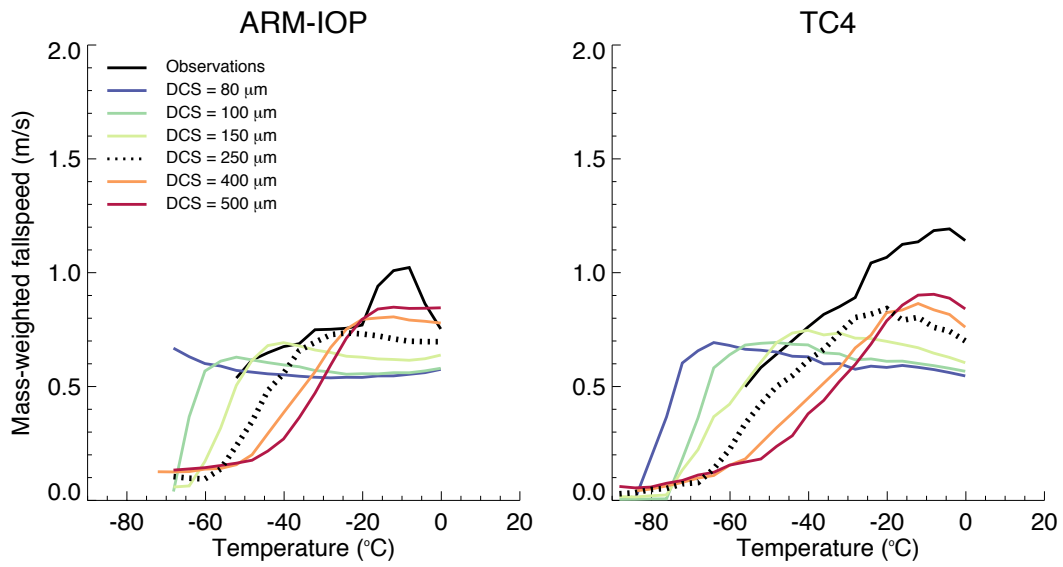
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Fig. 10. Same as Fig. 4, (ARM-IOP) but with various D_{cs} values



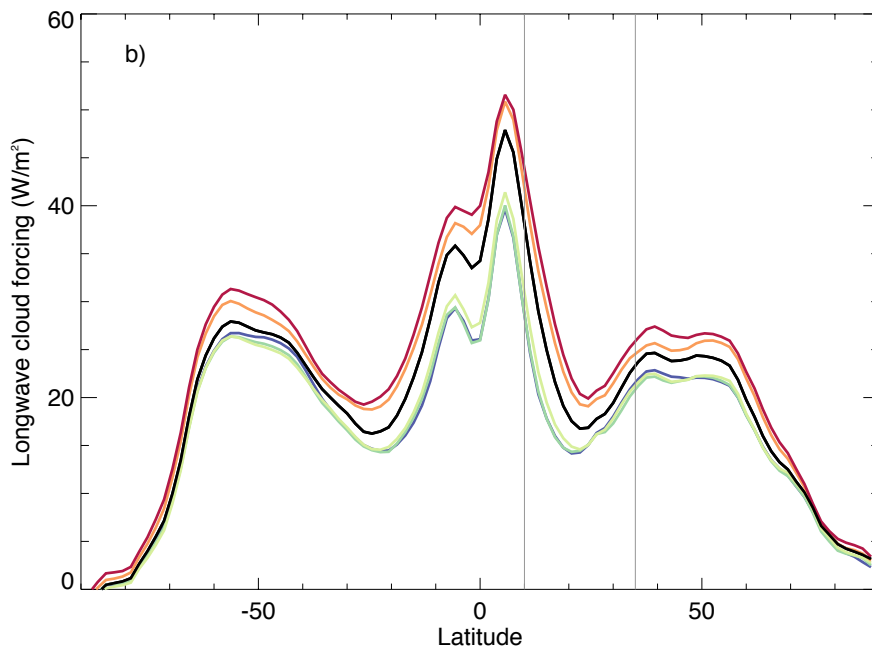
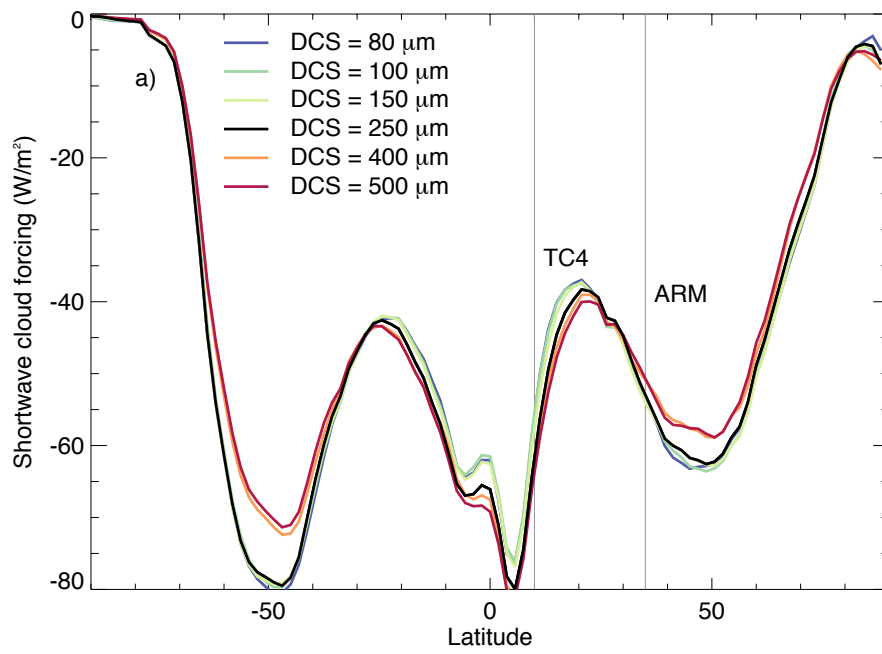
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Fig. 11. Same as Fig. 5 (moments, TC4), but using different D_{cs} values.



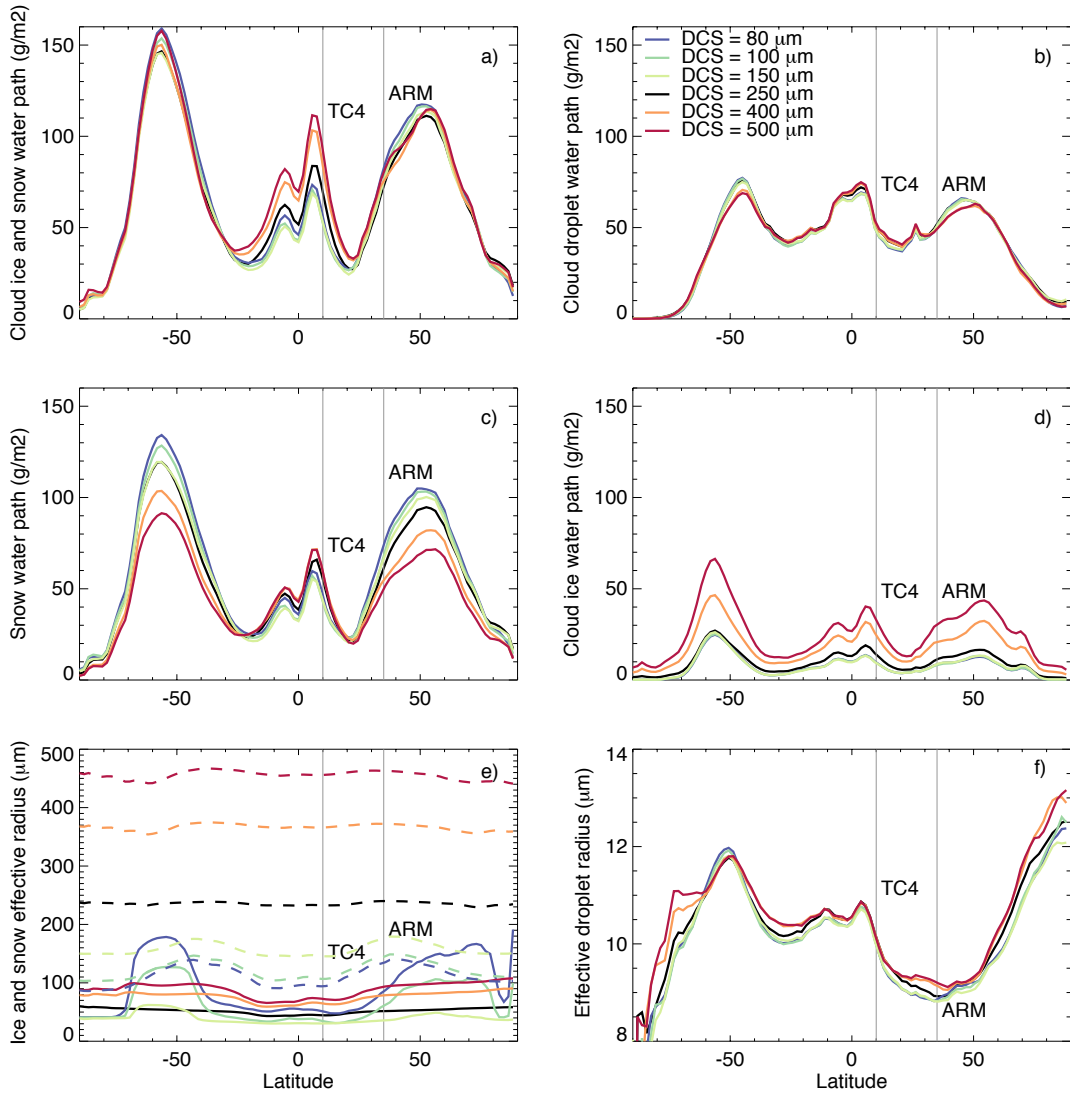
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Fig. 12. Same as Figs. 7 and 8, but using different D_{cs} values.



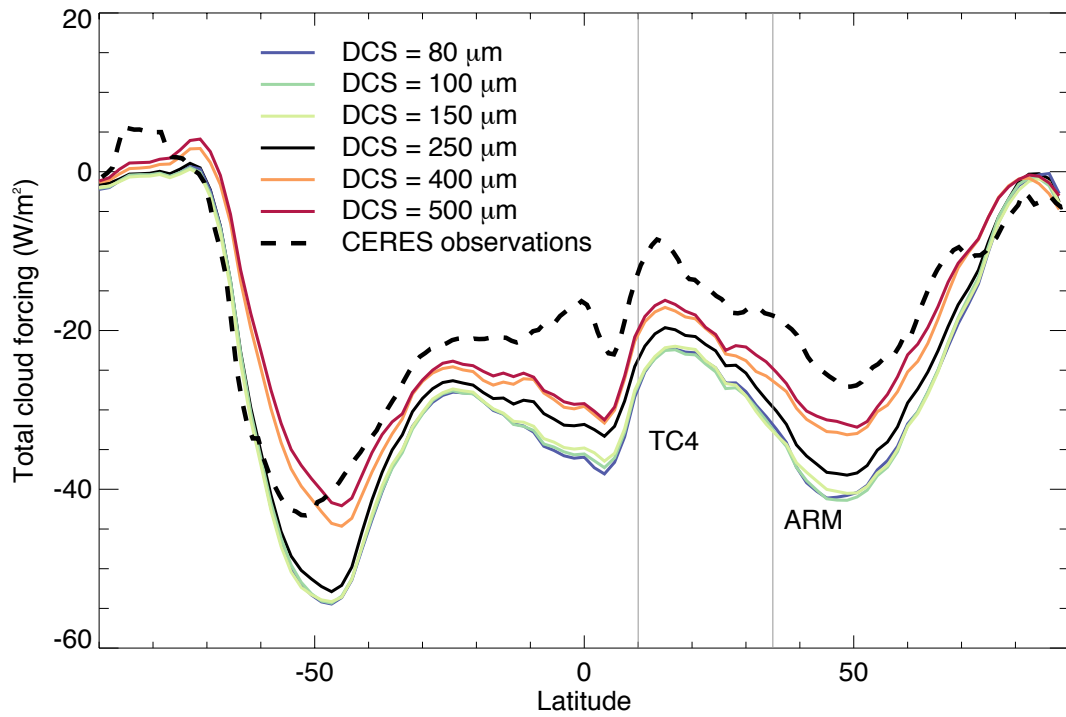
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Fig. 13. Zonal averaged shortwave and longwave radiative cloud forcing for the six runs, varying D_{cs} .



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Fig. 14. Zonal averaged cloud ice and snow water path (a), cloud droplet water path (b), snow water path (c), cloud ice water path (d), ice (solid) and snow (dashed) effective radius (e) and effective droplet radius (d) for the six different D_{CS} simulations.



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Fig. 15. Total radiative cloud forcing (LWCF+SWCF). Dashed line is observed cloud radiative forcing from CERES.