The Sensitivity of the Radiation Budget in a Climate Simulation to Neglecting the Effect of Small Ice Particles

Faisal S. Boudala

Department of Physics and Atmospheric Science, Dalhousie University, Halifax, Nova Scotia, Canada

George A. Isaac

Cloud Physics and Severe Weather Research Section, Environment Canada, Toronto, Ontario, Canada

N. A. McFarlane and J. Li

Canadian Centre for Climate Modelling and Analysis, Environment Canada, Victoria, British Columbia, Canada

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ABSTRACT

The sensitivity of the atmospheric radiation budget to ignoring small ice particles \((D \approx 100 \mu m)\) in parameterization of the mean effective size of ice particles was investigated by using the Canadian Centre for Climate Modelling and Analysis (CCCma) third-generation general atmospheric circulation model (AGCM3). The results indicate that small ice particles play two crucial roles in the radiative transfer that influence the simulated climate. First, they inhibit the IR radiation from escaping to space and, second, they enhance the scattering of solar radiation. On average, these two effects tend to partially cancel each other out. However, based on AGCM simulations, the small ice crystals make clouds more opaque to IR radiation. Generally, 5-yr seasonally averaged GCM results suggest that the strongest anomalies in outgoing longwave radiation (OLR) are found in the Tropics, reaching 15 to 25 W m\(^{-2}\) in areas where cold high cirrus anvil clouds are prevalent. The global average change in net cloud radiative forcing was 2.4 W m\(^{-2}\) in June–August (JJA) and 1.7 W m\(^{-2}\) in December–February (DJF). The change in globally averaged 5-yr mean cloud forcing was close to 1.9 W m\(^{-2}\). When the small particles were included, the globally averaged 5-yr mean precipitation decreased by about 8%, but cloudiness increased only slightly (by 2%). The 5-yr averaged global mean surface (screen) temperature also increased slightly (about 0.2°C) when the small ice particles were included.

1. Introduction

Atmospheric general circulation models (AGCMs) require solar and terrestrial infrared (IR) radiation to be calculated accurately in order to simulate climate. Ice clouds play a major role in the earth’s climate by absorbing the IR radiation and reflecting the solar radiation (Ramanathan et al. 1983; Ramaswamy and Ramanathan 1989; Stephens et al. 1990). The interactions of radiation with ice clouds are incorporated through parameterization of single scattering properties in terms of mean effective sizes \((D_{ge})\) of ice crystals and some other microphysical variable such as ice water content (IWC; e.g., Ebert and Curry 1992; Fu 1996; Fu et al. 1998) that are predicted by GCMs. However, \(D_{ge}\) is not predicted by most GCMs, but it is inferred from predicted microphysical variables. Therefore, accurate parameterization of effective sizes of ice crystals in terms of microphysical variables available in the AGCM is crucial for climate studies.

There are various definitions of the effective size of ice particles as summarized by McFarquhar and Heymsfield (1998) and Wyser (1998). In this paper, \(D_{ge}\) is defined following Fu (1996) as

\[
D_{ge} = \frac{2\sqrt{3} \text{IWC}}{3\rho_i A_c},
\]

where \(\rho_i\) is the density of pure ice, and \(A_c\) is the mean cross-sectional area of ice particles per unit volume. These microphysical variables are normally determined...
based on in situ measurements of the ice particle size distribution. There are parameterizations of $D_{ge}$ which are developed based on such in situ observations that consider small ice particles (e.g., Boudala et al. 2002; Ivanova et al. 2001; McFarquhar 2001; McFarquhar et al. 2003). The effective size of ice particles has also been parameterized based on direct in situ measurements of IWC and $A_s$ (or extinction coefficient; Garrett et al. 2003). There are also parameterizations of a bimodal size distribution of ice particles as a function of IWC and temperature, and thus $D_{ge}$ can be derived provided that IWC and temperature are known (e.g., Ivanova et al. 2001; McFarquhar and Heymsfield 1997). However, the determination of these quantities based on aircraft measurements using the current particle-measuring probes is still a challenge. The ice particle sizes and concentrations are normally measured using probes such as the Particle Measuring System (PMS) 2D optical array probes (2D-C and 2D-P) and the Forward Scattering Spectrometer Probe (FSSP) cloud droplet probe. The PMS 2D-C and 2D-P probes measure concentrations in the particle size ranges of 25–800 and 200–6400 μm, respectively.

The 2D optical array probe was developed almost 36 yr ago by Knollenberg (1970). The instrument has been used in the cloud physics community ever since. Although the actual imaging method remains the same, the algorithms for data extraction methods such as correctly sizing, counting, and identifying the shapes of particles are continuously evolving. It is now well known that the small particles ($D \leq 100$ μm) are not well measured using the traditional 2D-C probe due to many problems in identifying and sizing the particles (e.g., Korolev et al. 1998a). It has also been known for some time that there are problems with measuring the large ice particles because of things like particle shattering due to collisions with the probes themselves (Cooper 1977). This issue has been a subject of some discussions in the latest works of Korolev and Isaac (2005) and Field et al. (2006). Field et al. have identified that the particle shattering effects become significant when the mass weighted mean size exceeds 1 mm and the particle size distributions are relatively broad. However, because of the fact that there are many corrections to be made as mentioned earlier, the exact amount of the errors associated with particle shattering is difficult to quantify. One way to exclude the shattered particles is by removing all particles crossing the sampling volume in unusually short interarrival times or short distances between two successive images as discussed by Field et al. (2006) and earlier by Cooper (1977). Some of the old image processing software already have such algorithms. However, the precipitation probes such as 2D-P with a larger sample volume and course resolution tend to naturally filter out the shattered particles; thus it is the 2D-C probe that is affected the most in the presence of shattered particles (see Field et al. 2006).

An earlier study by Gardiner and Hallett (1985) indicated that the FSSP probe measured ice concentrations 2–3 orders of magnitude higher than those derived from a replicator. However, a recent study by Arnott et al. (2000) suggests that the replicator underestimates concentrations of particle ($D < 50$ μm) and thus cannot be compared with the FSSP data. A more recent study by Field et al. (2003) also suggests that the FSSP probe may overestimate ice concentration, but on average only by a factor of 2 due to shattering of ice particles.

There are also considerable uncertainties in determining the IWC from observed particle size distributions or direct measurements. As a result of these problems, developing an accurate and widely applicable parameterization of $D_{ge}$ has been a great challenge. Typically parameterizations are developed based on measurements conducted in a particular region or meteorological regime (e.g., tropical or extratropical regions) and thus may not be appropriate to be used for the entire globe. In this research, “small ice particles” refers to the ice particle sizes ($D \leq 100$ μm) that are not reliably measured with the current available instruments such as the PMS 2D-C and 2D-P probes, but this definition may differ for other researchers. The method for determining the size spectra from small to large ice particles, recognizing all the above problems, has been described by Boudala et al. (2002), and a brief summary is given in section 2.

Small ice crystals can be important for both solar and thermal IR radiative transfer (Platt et al. 1989; Kinne et al. 1992). In the geometric optics limit, the extinction of radiation is largely determined by the cross-sectional area ($A$) of ice particles projected normal to the propagation direction of the radiation. The absorption of solar radiation, however, is determined by both IWC and $A$. Therefore, the addition of the small particles enhances both scattering and absorption. The addition of small ice particles increases both $A_s$ and IWC leading to counteracting effects on $D_{ge}$ in accordance with Eq. (1). However, it has been shown using in situ measurements in extratropical stratiform clouds that the net effect of adding small ice particles is to decrease $D_{ge}$ on average close to 40% (Boudala et al. 2002).

The effect of adding small particles on IR radiation cannot be explained solely based on the geometric optics approximation as has been done for solar radiation,
particularly for particle sizes \((D < 40 \mu m)\) that are typically found in the natural atmosphere (e.g., Yang et al. 2001; Chýlek et al. 1992; Fu et al. 1998). When small ice particles are added to the larger ice crystals, it has been found that for some IR wavelengths, the scattering cross section is larger than the enhanced absorption, and the opposite occurs for the other IR wavelengths (Stackhouse and Stephens 1991). Generally in the IR wavelength that exhibit strong absorption, the absorption cross section is more closely related to \(A\). In the case of weak absorption, the absorption cross section is related to particle volume or mass. Since the addition of small ice particles more significantly affects \(A\) than the mass, the general tendency of adding small ice particles is to increase absorption (Foot 1988; Stackhouse and Stephens 1991). However, it should be noted that the effect of neglecting small ice particles on the radiation budget depends on the sizes and amount of these particles present in natural clouds (Arnott et al. 1994), and this is not well known at this time due to limitations of the measurements as mentioned earlier. Using observations in tropical cirrus clouds, Heymsfield and McFarquhar (1996) found that smaller ice particles \((D \leq 90 \mu m)\) make up more than 50% of the mass and are responsible for more than 50% of the extinction in the upper colder parts of cirrus. However, in the lower warmer region, they found that large ice crystals dominated the cloud and small particles were only responsible for 10% of the extinction. Nonetheless, they ignored the contribution of small ice particles in the parameterization of cloud optical properties. Using a 2-km-thick idealized cirrus cloud, Fu and Liou (1993) showed that decreasing \(D_{ge}\) from 50 to 25 \(\mu m\) increased IR heating (cooling) by a factor of 2 at the base (top) of cloud with a little change in solar heating. They also showed that the net cloud forcing at the top of the atmosphere (TOA) due to this change is positive and increases with decreasing \(D_{ge}\), except for \(D_{ge} \leq 25 \mu m\) and IWP > 20 g m\(^{-2}\) where the forcing may be negative. A study of the radiative effects of small ice particles, conducted by simulating stratiform anvils in the upper tropical troposphere using a one-dimensional cloud model, showed that neglecting small ice crystals \((D < 20 \mu m)\) could amount to an uncertainty of \(-40 W m^{-2}\) in the diurnally averaged net cloud forcing at the top of the atmosphere (Zender and Kiehl 1994) due to a significant increase in solar albedo. Also, the climate sensitivity study conducted by McFarquhar et al. (2003) shows that changing effective radius from 30 to 10 \(\mu m\) amounts to a 25% increase in shortwave forcing at TOA and at the surface. Although there have been some studies of the effect of small particles based on some idealized cirrus clouds or assumed variation in mean size of ice particles, the radiative impact of neglecting small ice particles in GCMs has not been considered explicitly. Although the actual contribution of small ice particles is not accurately known, it is worthwhile to test the sensitivity of neglecting these particles in parameterizations of \(D_{ge}\) based on the currently available measurement information.

As noted above, based on in situ observations, Boudala et al. (2002) developed parameterizations of the mean effective size of ice particles as a function of ice water content and temperature, and temperature alone, with and without small ice particles \((D \approx 100 \mu m)\). The purpose of the present work is to study the sensitivities of the radiation budget and climate simulations to these different parameterizations. The sensitivity of climate simulations will be studied using 5-yr simulations with the third-generation atmospheric general circulation model (AGCM3) of the Canadian Centre for Climate Modelling and Analysis (CCCma).

2. The CCCma AGCM

The CCCma AGCM3 is used to investigate the climate impact of small ice particles. This model, a successor to AGCM2 described by McFarlane et al. (1992), is documented by McFarlane et al. (2005). The description provided here summarizes features of relevance to the present work, particularly in regard to the representation of clouds and radiation.

The horizontal spatial structure of the main prognostic variables in AGCM3 are represented by the spectral transform method similar to AGCM2 while the vertical structure is represented by rectangular finite elements defined for hybrid vertical coordinates (Laprise and Girard 1990). The operational version of this model (used in the present work) has 32 layers extending approximately up to 50 km above the surface or 1 hPa and employs a triangular 47-wave spherical harmonics representation (T47). Sea surface temperatures and sea ice extent and concentration are specified using seasonally varying climatological data. Surface exchanges of heat, moisture, and momentum are parameterized following Abdella and McFarlane (1996). Atmospheric convection is parameterized by a cumulus parameterization as described by Zhang and McFarlane (1995). Cloud cover is specified as a function of relative humidity and potential temperature stratification. The cloud water content is assumed to be proportional to the adiabatic value found by vertically lifting a parcel of air through a specific depth (Betts and Harshvardhan 1987) similar to AGCM2. The zonal mean distribution of background aerosol (sulfate, dust, and sea salt) is implemented based on Shettle and Fenn (1979).
The solar radiation part of this model is based on Fouquart and Bonnel (1980) with an extension from two spectral bands to four spectral bands. For gaseous transmission, $O_2$, water vapor and $CO_2$ are considered in the solar radiation, but the absorption of $O_2$ is not considered. Rayleigh scattering is considered in all four bands from spectral range $0.25$ to $4 \mu m$. The infrared radiation part is based on Morcrette (1991) with six bands covering from spectral range $4$ to $1000 \mu m$. Water vapor, $CO_2$, $O_3$, $CH_4$, $N_2O$, $CFC_{12}$, and $CFC_{11}$ are contained in the gaseous transmission. The water vapor continuum is based on a parameterization by Zhong and Haigh (1995). This water vapor continuum scheme is based on the results of line-by-line model (LBLRTM) Clough–Kneizys–Davies Model version 2.2 (CKD2.2; Clough et al. 1989).

The water cloud optical properties are parameterized based on Dobbie et al. (1999) for solar radiation and Chylek et al. (1992) for the infrared. The ice water optical properties are parameterized following Fu (1996) for solar radiation and Fu et al. (1998) for the infrared. In the original radiation algorithm in GCM3, the mean effective size ($D_{ge}$) of nonspherical ice crystals was defined in terms of IWC based on Lohmann and Roeckner (1996). This parameterization will also be briefly discussed in the sections below.

The parameterization of the mean effective size of ice particles used for this study is based on Boudala et al. (2002) and given as

$$D_{ge} = 60.1 \exp(0.008T)$$  \hspace{1cm} (2)

and

$$D_{ge+s} = \begin{cases} 46.4 \exp(0.015T) \\ 53.01 \text{IWC}^{0.06} \exp(0.013T) \end{cases}$$  \hspace{1cm} (3)

without [Eq. (2)] and with small ice particles [Eq. (3)], respectively, where $T$ is temperature in °C and IWC is $g \text{ m}^{-3}$. The parameterizations given in Eqs. (2) and (3) were based on data collected during several field projects conducted in extratropical regions. The main instruments used for this work includes the PMS 2D-C, 2D-P, and FSSP probes discussed earlier, and the Nevzorov total water content (TWC)/liquid water content (LWC) probes (Korolev et al. 1998b). The shattered and elongated thin particle images ($D \geq 100 \mu m$) are mainly excluded from the 2D-P and 2D-C data using image processing software. The small particles ($D \leq 100 \mu m$) are estimated using the FSSP concentrations, and 2D-C spectra. The FSSP spectra are assumed to be described by a gamma distribution function and overlap with 2D-C spectra at 125 $\mu m$ where the 2D-C measurement is reliable. When the gamma distribution function describing the FSSP spectra is integrated, it was set to give the concentration measured with the FSSP probe corrected for shattering effects (Field et al. 2003). Only sizes between 125 and 575 $\mu m$ have been included from 2D-C measurements, and the rest of the sizes greater than 575 $\mu m$ are derived from 2D-P measurements. Thus based on the discussion presented earlier, the shattering effects on the parameterizations are not expected to be that significant. The total mass derived from the total spectra (small + large) agreed quite well with the IWC measured independently with the Nevzorov probe, particularly at higher IWCs, which validates the parameterizations. The recent work by Isaac et al. (2006) reveals that the Nevzorov probe “may” also suffer from similar shattering problems, but they were unable to quantify the problem; more studies are needed to evaluate the implications of this for the parameterizations used in this paper. A detailed discussion of the development of the parameterization is given in Boudala et al. (2002). As noted below, the $D_{ge}$ parameterization agrees quite well with that reported recently by Garrett et al. (2003) when small particles are included. It is noteworthy that Garrett et al. utilized direct measurements of IWC and extinction in the Tropics, which is different from the measurement methods used by Boudala et al. Thus, notwithstanding the measurement uncertainties and the fact that the two independently proposed parameterizations were developed based on measurements in different geographical regions, their similarity is encouraging. It justifies the use of a single parameterization for the entire globe for the present study since a central goal is to gain insight into the sensitivity to representations of $D_{ge}$ in climate simulations and in particular the possible importance of accounting for small particles. Considerable uncertainty remains as to the range of applicability of these parameterizations, and there are circumstances in which they would not be adequate. Other studies have shown that the mean effective sizes of ice particles in some tropical cirrus cloud regimes (e.g., McFarquhar 2001) are relatively larger than the extratropical regions (e.g., Boudala et al. 2002). In these particular cloud regimes, the effective size is likely to depend more strongly on variables (such as IWC and number concentration) that are typically either specified a priori or are less reliably simulated by current GCMs. For example, in AGCM3, IWC is generated through a simplified bulk parameterization of microphysical processes that does not depend explicitly on particle sizes. Therefore, the model-predicted IWC is assumed to represent the total IWC that includes the small ice particles.
3. Sensitivity experiments

The numerical experiments discussed below were designed to study aspects of the sensitivity of climate simulations to specification of the effective size of ice crystals in clouds. Equation (2) was used to represent the case without small ice particles, and \( D_{ge} \) depends only on \( T \). Equation (3) was used to represent the case with small ice particles and includes relationships where \( D_{ge} \) depends on \( T \) alone and on both \( T \) and IWC. In the operational version of AGCM3, the mean effective size parameterization of ice particles is based on that used by Lohmann and Roeckner (1996), which is given as \( D_{ge} = 129 \text{ IWC}^{0.216} \), but limited to be no smaller than 31 (\( \mu \text{m} \)) and not allowed to exceed 77 (\( \mu \text{m} \)). The effect of using this parameterization will be compared with the results from using the parameterizations of the mean effective size based on Boudala et al. (2002) that are given in Eqs. (2) and (3). As noted above, it is assumed that the model-predicted IWC represents the total IWC, no matter what parameterization is used.

Two different interactive simulations have been conducted. Interactive simulations were chosen in preference to offline radiation budget calculations in order to account for the roles of feedback effects associated with neglecting small ice particles in climate simulations. One of the simulations is to test the sensitivity to neglecting small crystals. For this test, the parameterization of mean effective ice crystal size was done without and with small particles, \( D_{ge}(T) \) and \( D_{ge+}(T) \), and is applied for the entire globe during a 5-yr GCM simulation. The second experiment is to evaluate some key parameters of the climate simulations using the proposed parameterization against observations and relative to a control simulation using the standard operational parameterization. For this experiment, one 5-yr simulation with \( D_{ge+}(\text{IWC}, T) \) that includes small particles has been conducted.

a. Sensitivity in the IR radiation and comparison with satellite observations

Figure 1 shows the mean seasonal difference \([F_\text{IR}(D_{ge+}(T)) - F_\text{IR}(D_{ge+}(T))]\) in simulated outgoing longwave radiation (OLR) at the TOA based on a 5-yr simulation. The three panels are for June–August (JJA), September–November (SON), and December–February (DJF). Generally, the largest differences in OLR are found in the Tropics reaching 15 to 25 W m\(^{-2}\) in areas where convective clouds that are associated with the intertropical convergence zone are commonly found. These anomalies seem to propagate northward in JJA and southward in DJF. The anomalies are also mostly positive, which implies that in the absence of small ice particles, the atmosphere is more transparent to IR radiation. Figure 2 shows the vertical cross section of anomaly in cloudiness for JJA (top panel) and DJF (bottom panel). It is interesting to note that the most pronounced cloudiness anomaly in the upper troposphere is negative and flanked by broader and generally much less pronounced positive anomalies, implying that the most pronounced effect of accounting for small ice particles is increasing cloudiness in these upper-tropospheric regions. The pronounced decreases in cloudiness flanking the increases in the tropical upper troposphere are indicative in part of an upward shift in cloudiness associated with accounting for the effects of small particles. As noted in the preceding section, the cloud cover is diagnosed based on model-predicted relative humidity and thermal stratification. Generally the changes in both temperature and relative humidity are small in the upper troposphere but in combination are responsible for the changes in cloudiness. For example, there are small but systematic increases in relative humidity (not shown) in the tropical upper troposphere associated with accounting for the contribution of small particles to the effective size parameter. Changes in cloudiness particularly in the lower troposphere are generally small. Although changes in OLR and cloudiness are generally consistent, the direct effect associated with the change in the effective size due to adding small ice particles seems to be mainly responsible for the IR changes depicted in Fig. 1. The inclusion of small particles can enhance both the cloud-top cooling and cloud-base warming because the decrease in effective size increases both absorption and emission of IR radiation. However, the cooling effect is relatively stronger, resulting in a net reduction in OLR when the small ice particles were included. Similar results have been found by Stackhouse and Stephens (1991).

As noted above, the OLR changes are most pronounced in the Tropics. This is consistent with the change in the effective ice crystals size due to adding small ice crystals in relation to temperature, shown in Fig. 3. The small ice particle contribution increases with decreasing temperature reaching up to 50% at very cold temperatures. The ice clouds in the tropical upper atmosphere are found at much higher levels and much colder temperatures than the clouds found in the high latitudes and are responsible for much of the IR forcing (Hartmann et al. 1992). This is consistent with observations during the International Cloud Climatology Project (ICCP) (Doutriaux-Boucher and Seze 1998) and satellite observations (Riedi et al. 2000), which indicate that the frequency of occurrence of cold ice clouds tends to be a maximum in the area of the inter-
Fig. 1. Global distributions of anomaly $\{F_{\text{IR}}(D_{\text{ge}}(T)) - F_{\text{IR}}[D_{\text{ge}+s}(T)]\}$ for outgoing longwave radiation at the top of the atmosphere based on a 5-yr simulation: (a) JJA, (b) SON, and (c) DJF.
As a result of this, the effect of adding small ice particles to the IR flux at the top of the atmosphere is likely to be much more pronounced in the Tropics as compared to the high latitudes, reaching 15 to 25 W m$^{-2}$ within the intertropical convergence zone. As noted above and illustrated in Fig. 3, the Garrett et al. (2003) tropical parameterization of $D_{ge}$ (G2003) is very similar to $D_{ge}$ that includes small particles. It is interesting to note that in the DJF season (Fig. 1c), the negative anomaly in the OLR over the tropical Pacific Ocean is associated with a relatively strong positive anomaly in planetary albedo shown in Fig. 4. This change is associated both with changes in the upper-tropospheric cloudiness and changes in effective size in those regions, and is consistent with the changes in shortwave cloud forcing discussed in the following section.

Figure 5 shows zonal and seasonally averaged IR cloud forcing at the TOA in DJF. The satellite observations based on Earth Radiation Budget Experiment (ERBE) data (see Barkstrom 1984) are marked as dashed lines. In broad accord with observations, the
simulated cloud forcing peaks in areas where significant clouds are formed in association with well-known features of the general circulation of the atmosphere. The IR cloud forcing due to adding small ice crystals is pronounced in these cloudy regions and consistent with the OLR anomalies depicted in Fig. 1. It must be noted that considerable effort was put into adjusting parameters that determine cloud properties in the standard model so as to achieve reasonably good agreement, in a zonal mean sense, with observed radiative forcing, particularly in the Tropics. The response to implementation of the new parameterizations must be viewed in this context. An operational implementation of one of the new parameterizations would in general require a review and retuning of other parameterizations to achieve a similarly acceptable agreement with observations. However, there is no evidence to suggest that the simulations based on the mean effective size of ice particles, which was parameterized as a function of both temperature and IWC, agree better with observations as compared to the $D_{ge}$ parameterized as a function of temperature alone. It is also noteworthy that in the Tropics the IR forcing for the CCCma standard simulation is closer to that for the simulation using the new parameterization without including small ice crystals. This is consistent with the curves in Fig. 6, which shows the mean effective ice crystal size parameterized as a function of IWC and temperature in this work and by McFarquhar (2001) based on measurements in tropical cirrus clouds formed in a region of convective outflow given as

$$D_{ge}(\text{IWC, } T) = \left[ a + b(Tn) + c(Tn)^2 + d(Tn)^3 \right],$$

where $n = \text{IWC}/\text{IWC}_0$, and the coefficients $a$, $b$, $c$, and $d$ are dependent on temperature (see McFarquhar 2001 for more details), and $\text{IWC}_0 = 1$. The CCCma GCM3
uses the effective size parameterization with IWC alone by Lohmann and Roeckner (1996), which is given as
\[ D_{ge} = 129.06 \text{ IWC}^{0.216} \] and later adapted with some tuning in a form
\[ D_{ge} = \max[\min(129.06 \text{ IWC}^{0.216}, 77), 31]. \] (5)

The Lohmann and Roeckner (1996) parameterization, with limitations as noted above, is also shown in Fig. 6. If the tropical \( D_{ge} \) is mainly determined by IWC, for high IWCs, which would be expected in a tropical cirrus cloud formed in a convective outflow region, the new parameterization with small ice crystals would underestimate the effective size and overestimate the optical depth. This would partly explain the more pronounced IR cloud forcing in the Tropics. Observations in mid- to high-latitude regions indicate that \( D_{ge} \) has a weak IWC dependence, but the \( D_{ge} \) in tropical clouds as reported by McFarquhar (2001) shows very strong IWC dependence. In contrast, the observations reported by Garrett et al. (2003), also in tropical clouds, suggest that the \( D_{ge} \) mainly depends on temperature in basic agreement with Boudala et al. (2002; Fig. 3). It is difficult to compare \( D_{ge} \) (IWC) directly with the parameterization given in Fig. 3, which depends only on temperature. However, in ice clouds, IWC generally increases with increasing temperature (Boudala et al. 2002; Field et al. 2004; Stephens et al. 1990; McFarquhar and Heymsfield 1997). Thus, one can expect larger \( D_{ge} \) with increasing

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**Fig. 5.** Seasonally and zonally averaged IR cloud forcing (W m\(^{-2}\)) in DJF. (top) Based on a 5-yr simulation without small crystals and with small ice crystals as indicated by symbols (left) \( D_{ge}(T) \) and (right) \( D_{ge+s}(T) \), respectively. (bottom) (left) The standard CCCma GCM3 simulations for the same season and (right) the simulations based on the mean effective size parameterized as a function of temperature and ice water content, \( D_{ge+s}(T, \text{IWC}) \). The satellite observation is marked by a dashed line and the model result is marked by a solid line.
Generally, the operational version of AGCM3 predicts less shortwave cloud forcing than observed in the extra Tropics of the summer hemisphere. The new parameterizations that include small ice particles seem to alleviate this deficiency slightly. It should be noted that zonal averaging eliminates the substantial spatial variability of cloud forcing associated with the different parameterization of $D_{ag}$. As discussed in the previous sections, localized effects of small particles on the radiation field are much larger than is evident in the zonally averaged fields.

The potential importance of accounting for small particles in parameterization of ice crystal effective sizes, particularly in the Tropics, has been demonstrated by these results. Results of the simulations for other seasons show very similar behavior although the strength of the forcing changes can differ.

c. Sensitivity in net radiative forcing

Adding small ice crystals in the radiation parameterization has two potential effects in the atmospheric radiation budget. It increases the IR absorption, which tends to warm the atmosphere, and it enhances the scattering of solar radiation, which tends to cool the atmosphere. It is the net energy balance ($\Delta E$) at the top of the atmosphere that determines whether the atmosphere is on average warming or cooling. This net energy balance at the TOA is defined as the difference between the energy (mainly solar) that is entering the atmosphere and the thermal (IR) energy that is leaving the atmosphere. Figure 8 shows a model-simulated global distribution of the difference in net energy balance ($\Delta [E[D_{ag}(T)] - \Delta E[D_{ag+1}(T)]$) at the TOA between the simulation without small ice particles $[\Delta E[D_{ag}(T)]$ and with small ice particles $[\Delta E[D_{ag+1}(T)]$ for an average of five simulated years. Comparison of JJAS with DJF shows that the anomaly is relatively larger in the Tropics, and in a winter season in both hemispheres, particularly in the Northern Hemisphere winter (DJF), shows mostly negative trends reaching up to $-10$ W m$^{-2}$ in some localized regions over the North Atlantic and Indian Oceans. The simulated and observed net energy balance at the TOA is negative in the winter hemisphere (net energy loss) and positive in the summer hemisphere (net energy gain). Therefore, the negative trend in the figure implies that $\Delta E[D_{ag}(T)]$ is more negative than $\Delta E[D_{ag+1}(T)]$ in the winter hemisphere and more energy is lost to space in the absence of small ice crystals. However, during the summer hemisphere the negative trend implies that $\Delta E[D_{ag+1}(T)]$ is more positive than $\Delta E[D_{ag}(T)]$, which also implies that the atmosphere gains more energy in the presence of small ice crystals. In winter, the incoming solar ra-

fig. 6. Comparisons of mean effective ice crystal size parameterized as function IWC and temperature in this work and by McFarquhar (2001) for tropical cirrus. The effective size parameterization with IWC alone used by Lohmann and Roeckner (1996) and later adapted in CCCma GCM3 with some tuning is also shown.

IWC or increasing temperature. The possible ranges of the $D_{ag}$ (30–77 μm) in simulations using the operational version of AGCM3 are similar to $D_{ag}(T)$ derived without small ice particles, as can be seen in Fig. 6. When small particles are included, however, for a given cloud with predicted IWCs (IWC < 0.002 g m$^{-3}$ or IWC > 0.1 g m$^{-3}$), Eq. (5) may predict $D_{ag}$ of 31 or 77 μm, respectively, and these may not be compatible with $D_{ag}$ predicted by $D_{ag+1}(T)$, which never exceeds 46 μm in subfreezing temperatures. Tropical ice clouds contain IWCs much higher than 0.1 g m$^{-3}$, and thus it is not surprising to see the significant differences between the two simulations in the Tropics that are depicted in Fig. 5.

b. Sensitivity in solar radiation

Figure 7 shows the mean DJF shortwave cloud forcing for the five simulated years as mentioned in the previous section. As in the case of the IR cloud forcing, the effect of small ice crystals in shortwave cloud forcing is larger in the Southern Hemisphere Tropics by about 10 W m$^{-2}$. This is consistent with changes in the planetary albedo noted in the preceding section and expected as a consequence of including small ice crystals, which on average reduces the mean effective sizes of ice crystals. Since optical depth increases with decreasing $D_{ag}$ for a given IWC, this causes more solar radiation to be reflected back to space.
Radiation is reduced in both hemispheres, but the IR flux is relatively unchanged. This may be one part of the reason why the effect is much stronger in the winter season as compared to the summer season. It should also be noted, however, that the strongest effects are very much localized, and it is perhaps noteworthy that the spatial patterns of energy balance anomalies in Fig. 8 have some similarities to those of OLR. In DJF, over the North Atlantic, Indian Oceans, and southern Africa, the strong reductions in outgoing IR fluxes are correlated to the minimum in energy loss shown in Fig. 8. In JJA (see Figs. 1 and 8), for example, over northwestern Africa, the eastern and western coasts of the South American continent, and some parts of the Indian Ocean a similar behavior is shown suggesting that the reduction in IR flux due to adding small particles plays a significant role in the seasonal radiative energy balance of the globe.

In DJF, the global mean energy balance at the TOA was calculated to be 10.6 W m$^{-2}$ with small ice crystals, but without small ice crystals it was 9.5 W m$^{-2}$. This means that on average the atmosphere gains a net energy of 1 W m$^{-2}$ in that period due to adding small ice crystals. The calculated energy balance difference at the ground was almost twice as large, implying a corresponding reduction in net energy lost to the ground by the atmosphere. In JJA, the net change in energy balance at the top of the atmosphere was about −1 W m$^{-2}$, similar in magnitude to that in DJF but with an opposite sign, but at the ground the net change is reduced to 1.3 W m$^{-2}$. However, the 5-yr globally averaged energy balance (for all seasons combined) at the TOA was −0.3 W m$^{-2}$ for the model simulations without small particles, which is similar to the CCCma standard control simulation (see section 3a), and 0.8 W m$^{-2}$ when the small particles are included. Both of these values are relatively small and well within the uncertainty of satellite measurements of the TOA radiation balance.

The net global mean cloud forcing at the TOA in JJA was −21.4 W m$^{-2}$ for the simulation without small ice crystals as compared to −19 W m$^{-2}$ when the small particles are added. The difference of about 2.4 W m$^{-2}$ is substantial relative to the simulated forcing. For DJF

![Fig. 7. Seasonal and zonally averaged shortwave cloud forcing (W m$^{-2}$) in DJF. Symbols are as in Fig. 5.](image)
the net cloud forcing is $-20.5$ and $-22.2$ W m$^{-2}$ with and without small ice crystals, respectively, with a slightly lower difference of 1.7 W m$^{-2}$. The globally averaged 5-yr mean values of cloud radiative forcing, cloudiness, precipitation, and temperature are given in Table 1. The change in globally averaged 5-yr mean cloud forcing due to adding small ice crystals is close to 1.9 W m$^{-2}$. The globally averaged precipitation decreased by about 8% when small ice crystals were added, resulting in slightly better agreement with observations. In contrast to precipitation, cloudiness increased slightly by about 2% when small particles were included. The global mean surface (screen) temperature also increased slightly (by about 0.2°C) when small ice particles were included.

4. Summary and conclusions

Parameterizations of effective sizes of ice crystals as a function of temperature, IWC, and temperature with and without small ice crystals ($D \leq 100$ μm) have been tested using the Canadian Climate Centre for Climate Modelling and Analysis (CCCma) third-generation general circulation model (GCM3). In this study, a 5-yr
Table 1. Five-year mean globally averaged temperature, precipitation, cloudiness, and net radiative forcing based on simulations without small particle \( (D_{ge}) \) and with small ice particles \( (D_{ge+s}) \). The observed (obs) values of cloudiness based on the ISCCP and precipitation based on the Global Precipitation Climatology Project (GPCP) are also given.

<table>
<thead>
<tr>
<th></th>
<th>( D_{ge} )</th>
<th>( D_{ge+s} )</th>
<th>( D_{ge+s} - D_{ge} )</th>
<th>Obs</th>
<th>( D_{ge+s} - \text{obs} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Precipitation (mm day(^{-1}))</td>
<td>2.84</td>
<td>2.76</td>
<td>-0.08</td>
<td>2.71 GPCP</td>
<td>0.05</td>
</tr>
<tr>
<td>Cloudiness</td>
<td>0.61</td>
<td>0.63</td>
<td>0.02</td>
<td>0.62 ISCCP</td>
<td>0.01</td>
</tr>
<tr>
<td>Temperature (°C)</td>
<td>14.2</td>
<td>14.4</td>
<td>0.2</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Cloud forcing (W m(^{-2}))</td>
<td>-20.6</td>
<td>-18.7</td>
<td>1.9</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

mean seasonally averaged dataset has been used to depict the effects of implementing these parameterizations.

The simulation results indicate that the addition of small ice particles in parameterizations of effective size gives rise to two notable effects in climate simulations. First they inhibit the loss of IR radiation to space and, second, they enhance the scattering of solar radiation. On average, these two effects counteract each other, but seasonally averaged global distribution of energy balance data shows that the small ice crystals make clouds more opaque to IR radiation. This effect is stronger in the winter hemisphere. The simulation with \( D_{ge} \) that contained small ice crystals gave higher cloud forcing (IR and solar) in the tropical cloudy regions as compared to both the operational simulation and the satellite observations. As noted in the foregoing discussion, this enhanced departure from observations is not unexpected because the operational parameterizations have been tuned to achieve reasonably good (zonal mean) agreement with the observed forcing in the Tropics. However the sensitivity of the simulated cloud radiative forcing to inclusion of small ice particles is demonstrated to be substantial. Although there have been some changes in cloudiness, precipitation, and temperature, these differences are relatively small. Thus much of the more pronounced differences in forcing in the Tropics are associated with the fact that the simulated ice clouds in the tropical upper atmosphere are found at higher levels and are colder than the clouds found at high latitudes. Therefore, since the contribution of small ice particles to the effective size of ice crystals increases with decreasing temperature, the parameterization that includes small particles predicts smaller effective sizes of ice crystals and thus higher cloud forcing in tropical cirrus as compared to midlatitude cirrus.

Model simulations based on the new parameterizations, which include small ice particles, overestimated the effects of both longwave and shortwave forcing at the top of the atmosphere in the Tropics where the significant high clouds are formed, as compared to observations. As noted above, this locally larger departure from observations is not unexpected in simulations where a particular parameterization is replaced without any corresponding changes in linked parameterizations. However, it is also possible that the effective sizes of ice crystals \( (D_{ge}) \) in some tropical cloud regimes are inherently larger, as has been demonstrated by McFarquhar (2001), and thus underestimated by parameterizations used in this work. This may be so notwithstanding that the tropical \( D_{ge} \) values reported by Garrett et al. (2003) are in relatively good agreement with the proposed parameterization, which includes the effects of small particles.

The results indicating that simulated cloud radiative forcing is substantially affected by including the effects of small particles underline the importance of taking these effects into account in future model development. The possibility of applying the proposed single \( D_{ge} \) parameterization for the entire globe, while appealing because of its simplicity and ease of implementation, requires further study to determine its range of applicability and reliability. However, introduction of a more complicated parameterization that attempts to account for different cloud regimes in a physically meaningful way requires a corresponding enhancement in the treatment of clouds in general and microphysical processes in particular. Such developmental activities are currently under way within CCCma in regard to construction of a newer version of the AGCM, which, among other things, employs a prognostic cloud scheme that is quite different in its formulation and implementation from that used in AGCM3 and explicitly predicts liquid and ice water contents. Preliminary tests using the proposed new parameterization for effective size with the current experimental version of this new model have qualitatively corroborated the AGCM3 simulations discussed in this paper.

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