

Cirrus clouds in a global climate model with a statistical cirrus cloud scheme

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Supplementary Material

A two-moment treatment of cloud microphysics for warm and mixed-phase clouds

In addition to the new cirrus cloud scheme described in Sect. 2.2, a prognostic cloud droplet number concentration (n_l) equation for liquid clouds is added in the version of CAM3 used in this study compared with the version used in Liu et al. (2007a). Together with the prognostic liquid water mass (q_l), ice water mass (q_i), and ice crystal number concentration (n_i), the complete set of equations for the two-moment treatment of cloud microphysics for liquid clouds and mixed-phase clouds are:

$$\frac{\partial q_l}{\partial t} = A(q_l) + DTR(q_l) + Q_{cond} - Q_{liq2pr} - Q_{liq2ice}, \quad (S1)$$

$$\frac{\partial q_i}{\partial t} = A(q_i) + DTR(q_i) + Q_{liq2ice} + Q_{vap2ice} - Q_{ice2pr}, \quad (S2)$$

$$\frac{\partial n_l}{\partial t} = A(n_l) + DTR(n_l) + N_{nucl} - N_{self} - N_{liq2pr} - N_{evapl} - N_{liq2ice} \quad (S3)$$

$$\frac{\partial n_i}{\partial t} = A(n_i) + DTR(n_i) + N_{nuci} + N_{frz} + N_{sec} - N_{ice2pr} - N_{evapi}, \quad (S4)$$

where the A operator on the left side of Eqs. (S1)-(S4) represents the advective, turbulent, and convective transports, as well as the gravitational settling, and the DTR operator represents the detrained cloud liquid and ice from both deep and shallow convection.

Q_{cond} represents the net condensation rate for liquid clouds in warm and mixed-phase clouds as diagnosed by the fractional cloud closure scheme of Zhang et al. (2003), and acts to remove any supersaturation with respect to water (Liu et al., 2007a). Q_{liq2pr} and Q_{ice2pr} are the loss rates of cloud liquid mass and ice mass from the conversion of cloud condensate into precipitation, respectively. They include the auto-conversion of liquid water to rain, the collection of cloud water by rain from above, the auto-conversion of ice to snow, the collection of ice by snow, and the collection of liquid by snow (Rasch and Kristjánsson, 1998). $Q_{vap2ice}$ represents the deposition on ice crystals from water vapor (Liu et al., 2007a). $Q_{liq2ice}$ represents the growth of ice water at the expense of liquid water due to the Bergeron-Findeisen process in mixed-phase clouds, and is calculated based on Rotstajn et al. (2000). The in-cloud saturation vapor pressure used to calculate ice supersaturation in mixed-phase clouds is the saturation vapor pressure weighted by the proportions of ice and liquid water mass. This direct conversion from liquid to ice was not allowed in Liu et al. (2007a) who assumed a conversion from water vapor to ice which resulted in a smaller conversion rate of liquid to ice in mixed-phase clouds. This direct conversion was used in simulating observed clouds in the Mixed-Phase Arctic Cloud Experiment (M-PACE) in a single column model (Liu et al., 2007b) and in a short-range weather forecasting approach (Xie et al. 2008).

Detrained cloud mass from convection is assumed to be in the liquid phase for clouds warmer than -35°C , and is assumed to be in the ice phase for cirrus clouds. The conversion of this detrained condensate from liquid phase into the ice phase in the mixed-phase clouds is completed through the Bergeron-Findeisen process. Detrained cloud droplet number concentration and ice crystal number concentration are calculated from detrained cloud liquid and ice mass by assuming a spherical particle with constant volume-mean radius for both cloud droplets and ice crystals. These volume-mean radii for liquid droplets and ice crystals are calculated from the effective radius used in the standard CAM3 (Boville et al., 2006), following the treatment in Liu et al. (2007a) for ice crystals.

Microphysical terms for the cloud droplet number concentration (n_l) include the droplet source from activation (N_{nuci}), and droplet sinks from precipitation (N_{liq2pr}), evaporation

(N_{evapl}), self-collection (N_{self}), and freezing ($N_{liq2ice}$). The loss rate of cloud droplet number concentration from precipitation is assumed to be proportional to the loss rate of liquid water mass from precipitation, and is calculated as

$$N_{liq2pr} = \frac{Q_{liq2pr}}{q_l} n_l. \quad (S5)$$

The same assumption is applied to the loss rate of cloud droplet number concentration from evaporation. Only the net evaporation rate of cloud liquid mass (Q_{cond}) is diagnosed in CAM3, using the scheme of Zhang et al. (2003). Here we assume that the evaporation of cloud droplets only occurs when there is net evaporation of cloud liquid mass, and is calculated as:

$$N_{evapl} = \frac{-\min(0, Q_{cond})}{q_l} n_l. \quad (S6)$$

The depletion of cloud droplets from the freezing process ($N_{liq2ice}$) includes the initial freezing of cloud droplets into ice crystals from contact freezing (N_{frz}), and the depletion of cloud droplets from the Bergeron-Findeisen process. The fraction of cloud liquid droplet number concentration depleted due to the Bergeron-Findeisen process is assumed to be half of that of the cloud liquid mass depleted due to the same process ($Q_{liq2ice}$), except when the cloud liquid mass is completely depleted, and then all cloud droplets are depleted. Thus, we assume that part of the cloud mass that is depleted from the Bergeron-Findeisen process comes from the shrinking of cloud droplets.

The droplet activation term (N_{nucl}) follows the treatment of Lohmann et al. (1999). New droplets are assumed to form only when the number concentration of cloud droplets that would be activated exceeds the number concentration of preexisting cloud droplets and when net condensation occurs:

$$N_{nucl} = \max\left[\frac{1}{\Delta t}(N_{ac}a - n_{old}), 0\right], \text{ when } Q_{cond} > 0, \quad (S7)$$

where N_{ac} is the number concentration of cloud droplets activated from aerosol particles, and n_{old} is the grid-mean cloud droplet number concentration predicted by Eq. (S3) at the previous time step. N_{ac} is calculated from aerosol fields using a parameterization based on Köhler theory (Abdul-Razzak and Ghan, 2000; 2002). This parameterization combines the treatment of multiple aerosol types and a sectional representation of size to deal with arbitrary

aerosol mixing states and arbitrary aerosol size distributions. Five categories of aerosols are externally mixed: sulfate, biomass burning OM/BC, fossil fuel OM/BC, sea salt, and dust. The bulk hygroscopicity parameter for each category of aerosol is the volume-weighted average of the parameters for each component taken from Ghan et al. (2001) (also see Table 2 in Wang and Penner (2009)). The size distributions of the five types of aerosols are prescribed as in Table S1. The vertical velocity (w) used in calculating N_{ac} is calculated from $w = \bar{w} + c \times \sigma_w$, where \bar{w} is the large-scale vertical velocity, σ_w is the subgrid variance of the vertical velocity diagnosed from the eddy diffusivity and the mixing length, and c is a coefficient that depends on σ_w (Wang and Penner, 2009).

The self-collection of cloud droplets follows the treatment of Beheng (1994) and is parameterized as

$$N_{self} = 1.29 \times 10^{10} \times b \left(\rho \frac{q_l}{a} \right)^2, \quad (\text{S8})$$

where ρ is the air density (kg/m^3).

The source and sink terms of ice crystal number concentration include deposition/condensation freezing (N_{nucl}), contact freezing of cloud droplets from dust particles (N_{frz}), the secondary ice production by ice splintering between -3 and -8 °C (N_{sec}), the loss from precipitation (N_{ice2pr}), and the loss from sublimation of ice crystal particles (N_{sublim}). These source and sink terms are described in Liu et al. (2007a), with modifications as follows. In Liu et al. (2007a), ice particles are assumed to evaporate completely only when cloud dissipates as the cloud fraction (a) decreases. In addition to the evaporation assumed in Liu et al. (2007a), we assumed in this study that ice crystal particles that are advected into the clear sky part of a grid box also evaporate, which is consistent with the treatment of cloud condensate from the scheme of Zhang et al. (2003).

The effective radii of ice crystals is parameterized from the ice crystal number concentration, ice water content, and temperature as described in Chen (2006) and Liu et al. (2007a). Cloud droplet effective radius (r_{el}) is calculated from the volume-mean cloud droplet radius (r_{vl}) based on the parameterization of Rotstajn and Liu (2003), which takes account of the change in the dispersion of the cloud droplet size distribution caused by the change in the cloud droplet number concentration. r_{el} is calculated as:

$$r_{el} = \beta r_{vl}, \quad (\text{S9})$$

where β is the droplet size spectral shape factor and is approximated as:

$$\beta = \frac{(1 + 2\varepsilon^2)^{2/3}}{(1 + \varepsilon^2)^{1/3}}, \quad (\text{S10})$$

where ε is the relative dispersion of the size distribution of cloud droplets and is approximated as:

$$\varepsilon = 1 - 0.7 \exp(-\alpha N_l), \quad (\text{S11})$$

where N_l is the in-cloud droplet number concentration, and α is a coefficient. An α of 0.003 is used in this study, which represents the middle curve that fits the observed ε and N_l in Fig. 1 of Rotstayn and Liu (2003).

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Table 1. Descriptions of simulations

Table S1. Size distribution parameters for aerosols

Aerosol component	Ni	Ri, μm	Sigma
Sulfate ^a	1.0	0.05	1.9
Fossil fuel OM/BC ^a	1.0	0.05	1.9
Biomass OM/BC and natural OM ^b	1.0	0.08	1.65
Sea Salt ^c	0.965	0.035	1.92
	0.035	0.41	1.70
Dust ^d	0.152	0.01	2.3
	0.727	0.045	1.6
	0.121	0.275	2.5

^aThe size distribution of sulfate and internal mixture fossil fuel OM/BC is the fossil fuel size distribution taken from Penner et al. (2001, Table 5.1).

^bThe size distribution of internal mixture biomass burning OM and BC is the biomass burning size distribution from Penner et al. (2001, Table 5.1).

^cThe size distribution of sea salt is taken from Quinn and Coffman (1998).

^dThe size distribution of dust is taken from de Reus et al. (2000).