Simulation of arctic low-level clouds observed during the FIRE Arctic Clouds Experiment using a new bulk microphysics scheme

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Abstract. A new bulk cloud microphysics scheme that accounts for aerosol microphysical properties and size distribution is implemented into the single-column version of the ARCSyM. This scheme is distinguished from other bulk microphysics schemes by its prognostic determination of cloud particle number concentration and saturation ratio. The new scheme is compared to a simpler bulk microphysics scheme and observations taken during the FIRE Arctic Clouds Experiment in May 1998. Qualitatively, the two microphysics schemes are generally in agreement with the observed cloud formation and evolution. Comparison with aircraft measurements at 3 times shows that the new scheme better discriminates cloud phase and reproduces reasonably well the observed liquid and ice water content for two cases. The better performance of the new scheme is attributed to its more elaborated treatment of the freezing process which is made possible by the prognostic determination of cloud particle number concentration and the assumption of a bimodal lognormal cloud size distribution. Sensitivity studies are performed to assess four aerosol microphysical properties on the evolution of cloud microphysical processes. Results show that the INN concentration, the aerosol number concentration, the slope of the aerosol size distribution, and the aerosol solubility may impact substantially on cloud phase and total water content. The liquid water path and ice water path can vary by as much as 100 g m\(^{-2}\) locally as a result of the variation of these parameters related to aerosols.

1. Introduction

The importance of arctic clouds and radiation to the regional and global climate is summarized by Curry et al. [1996]. The difficulty in simulating and remotely sensing these clouds motivated the FIRE Arctic Clouds Experiment [Curry et al., 2000], which was conducted in conjunction with the Surface Heat Budget of the Arctic Ocean (SHEBA) experiment [Perovich et al., 1999]. The difficulties in modeling arctic clouds have been hypothesized by Curry et al. [1996] to arise from the complex vertical structure of the stable arctic atmosphere, the presence of mixed-phase clouds, and the susceptibility of the arctic clouds to modification by aerosols.

Curry et al. [2000] describe an initial application of the SHEBA/FIRE data set to evaluating cloud parameterizations in the context of single-column model simulations. Results of simulations for May 1998 showed that the models tend to underpredict low cloud amount and the column liquid water path. The underestimation of the column liquid water path was in part attributed to inaccurately representing mixed-phase clouds as entirely crystalline.

Mixed-phase clouds occur at high frequency in the Arctic from November to June [Curry et al., 1990; Pinto, 1998; Perovich et al., 1999]. From observations obtained during autumn from the Beaufort Arctic Storms Experiment (BASE), Pinto [1998] hypothesized that mixed-phase clouds are maintained in the arctic boundary layer through a balance of condensation of liquid water through cooling and heterogeneous freezing of the cloud drops. Jiang et al. [2000] simulated the observed case study described by Pinto and found strong sensitivity of the phase of the cloud to the concentration of ice-forming nuclei (IFN). Simulations by Girard and Blanchet [2000a] (hereinafter referred to as GB) with an explicit aerosol-cloud microphysics model have shown that low-level clouds may remain in a mixed-phase state as long as 10 hours in the Arctic during winter.

Bulk microphysical parameterizations used in climate models [e.g., Fowler et al., 1996] typically distinguish the phase of cloud water by the cloud temperature. The discrimination between the liquid and the ice phases may be specified in terms of a single cutoff temperature, or as a function of temperature with the ratio of ice water content to liquid water content varying from zero at 0°C to 1 at -40°C. Curry et al. [1996] cite observations of liquid drops in arctic clouds at temperatures as low as -32°C and clouds that are completely glaciated at -14°C. Clearly, a simple temperature discriminator is not sufficient to diagnose the phase of supercooled clouds in the Arctic.

The complexity of the interactions between arctic cloud microphysics and aerosol are discussed by GB (2000a, b). Aerosol composition and concentration modify cloud microphysical properties through their ability to nucleate ice and water. Blanchet and Girard [1995] hypothesized that the microphysical properties associated to Arctic haze aerosols in the Arctic may alter the cloud particle mean diameter and, consequently, increase the precipitation efficiency. This would result in a larger dehydration rate of the air mass which in turn would lead to a decrease of the downward longwave radiation flux at the
surface. Other indirect effects of aerosols such as the Twomey and the Albrecht effects [Twomey, 1991; Albrecht, 1989] may strongly affect cloud microphysical properties.

Current bulk microphysics schemes used in general circulation models generally describe cloud with four prognostic variables: liquid, ice, rain, and snow water content [e.g., Fowler et al., 1996]. In these schemes the effective radius of cloud particles is parameterized as a function of temperature [e.g., On and Liu, 1995] or liquid or ice water content [e.g., McFarlane et al., 1992]. Current moisture schemes are therefore of limited utility in studying cloud microphysical property changes resulting from aerosol forcing. To account for the influence of aerosol on cloud microphysical properties, prognostic equations must be included for either particle concentration or size, in addition to condensed water content.

In this paper we describe a new bulk microphysics scheme that is capable of accounting for the impact of aerosol on cloud nucleation. This scheme is developed by merging the cloud schemes from GB (2000b) for cloud liquid and cloud ice and the production of precipitation following Grell et al. [1995]. The new microphysics scheme predicts the ice and liquid water content, rain and snow, number concentration of cloud droplets and ice crystals, and cloud saturation ratio. Aerosol concentration and composition are prescribed. This scheme is tested against observations taken during the FIRE Arctic Cloud Experiment in May 1998. The new microphysics scheme is also compared to the original version of the microphysics scheme developed by Grell et al. [1995] which does not include prognostic equations for particle concentration or saturation ratio. Finally, a sensitivity experiment is carried out to assess the effect on cloud formation of changing the IFN and aerosol concentration, the slope of the aerosol size distribution and the aerosol solubility.

2. Model Description

The single-column version of the Arctic Regional Climate System Model (ARCSyM) [Lynch et al., 1995] was used in our study. The ARCSyM is based on the NCAR regional climate model Version 2 (RegCM) [Giorgi et al., 1993a, b]. The model has 38 vertical levels with 13 levels between 900 hPa and the surface that allows for a high-resolution simulation of the vertical structure of the boundary layer. In this paper a brief description of ARCSyM physics package is given. The bulk microphysics schemes used in this study are described in details thereafter. One is referred to Lynch et al. [1995] for a more extensive description of the other components of the model. In the ARCSyM the planetary boundary layer scheme of Holtslag et al. [1990] is used. The model uses the CCM2 solar radiation scheme [Briegleb, 1992a, b] and the Rapid Radiative Transfer Model (RRTM) infrared radiation scheme [Mlawer et al., 1997]. The cloud scheme of Grell et al. [1995] is used for simulating cumulus cloud processes.

Two bulk microphysics schemes have been implemented in ARCSyM: (1) the NCAR/MM5 Penn State cloud scheme [Grell et al., 1995] (hereinafter referred to as the MM5 microphysics scheme), and (2) the new microphysics scheme developed in this research. The MM5 cloud scheme is based on the bulk microphysics scheme of Grell et al. [1995] and is an upgraded modified version of the microphysics scheme of Dudhia [1989]. Prognostic variables are the ice mixing ratio, liquid mixing ratio, snow mixing ratio, and rain mixing ratio. Microphysical processes simulated include nucleation of ice crystals, nucleation of water droplets, condensation, evaporation, deposition, sublimation, autoconversion of cloud droplet to rain, autoconversion of ice crystals to snow, melting/freezing of snow/rain and of cloud droplets/ice crystals, and collision processes.

The new scheme is distinguished from the MM5 cloud scheme by the addition of three prognostic variables: the saturation ratio, the number concentration of cloud droplets, and ice crystals. These prognostic variables added to the ice and liquid mixing ratios allow for determining the mean diameter of ice crystals and water droplets diagnostically. Cloud particle size distribution is assumed to be a superposition of two lognormals with a standard deviation of 1.3, each lognormal representing one phase, following GB (2000b). This choice of cloud particle size distribution is based on simulation of Arctic low cloud with an explicit microphysics model (GB, 2000a) and observations [Rogers and Yau, 1989]. Although observations indicate that the gamma size distribution can also be used to represent cloud spectra in models [Rogers and Yau, 1989], the lognormal distribution is preferred here since it is more representative of the cloud spectra obtained with the explicit microphysics model (GB, 2000a) which has been used to develop the bulk microphysics scheme.

The new scheme and the MM5 scheme are based upon 8 and 5 prognostic equations respectively. The five common prognostic equations are water vapor mixing ratio \( q_v \):

\[
\frac{\partial q_v}{\partial t} = -\frac{u}{\partial x} + \frac{\partial q_v}{\partial x} - v \frac{\partial q_v}{\partial y} - w \frac{\partial q_v}{\partial z} + \text{PCC + PCI + PII + PID + PREC + PRES + } \delta_v
\]

(1) cloud liquid mixing ratio \( q_l \):

\[
\frac{\partial q_l}{\partial t} = -\frac{u}{\partial x} + \frac{\partial q_l}{\partial x} - v \frac{\partial q_l}{\partial y} - w \frac{\partial q_l}{\partial z} + \text{PCI + PRH + PACC + PRSC + } \delta_v
\]

(2) cloud ice mixing ratio \( q_i \):

\[
\frac{\partial q_i}{\partial t} = -\frac{u}{\partial x} + \frac{\partial q_i}{\partial x} - v \frac{\partial q_i}{\partial y} - w \frac{\partial q_i}{\partial z} + \text{PII + PID + PRG + PRH + PRA + PACCI + PRSI + PRM + } \delta_v
\]

(3) rain mixing ratio \( q_r \):

\[
\frac{\partial q_r}{\partial t} = -\frac{u}{\partial x} + \frac{\partial q_r}{\partial x} - v \frac{\partial q_r}{\partial y} - w \frac{\partial q_r}{\partial z} + \text{PREC + PKR + PCA + } \text{PACC + } \delta_v
\]

(4) and snow mixing ratio \( q_s \):

\[
\frac{\partial q_s}{\partial t} = -\frac{u}{\partial x} + \frac{\partial q_s}{\partial x} - v \frac{\partial q_s}{\partial y} - w \frac{\partial q_s}{\partial z} + \text{PKS + PRA + PACCI + } \delta_v
\]

(5) The new scheme additionally includes the following prognostic equations: number concentration of cloud droplets \( N_c \);
\[
\frac{\partial N_s}{\partial t} = \frac{-u \partial N_s}{\partial x} + v \frac{\partial N_s}{\partial y} - w \frac{\partial N_s}{\partial z} - \frac{\rho}{m(D_s)} (\text{PCA + PRSC + PACCC - PRG - PRH - PCI + PRM}) + \delta_n
\] (6)

\[
\text{number concentration of ice crystals } N_s:
\]
\[
\frac{\partial N_s}{\partial t} = \frac{-u \partial N_s}{\partial x} + v \frac{\partial N_s}{\partial y} - w \frac{\partial N_s}{\partial z} - \frac{\rho}{m(D_s)} (\text{PRA + PRSI + PACCI - PRG - PRH - PII + PRM}) + \delta_n
\] (7)

\[
\text{and saturation ratio } S:
\]
\[
\frac{dS}{dt} = \frac{-L}{R_sT_s^2} \left[ \frac{T_s}{\partial T_s} - \frac{u}{\partial x} - \frac{v}{\partial y} - \frac{w}{\partial z} \right] + \frac{\rho}{\varepsilon_e} \left[ \text{PCC + PCI + PII + PID - PRES - PREC} \right]
\] (8)

where \(u, v, w, p, T, r, L, \varepsilon_e, R_s, \) and \(\delta\) are, respectively, the zonal velocity, longitudinal velocity, vertical velocity, pressure, temperature, air density, latent heat of fusion, vapor partial pressure at water saturation, gas constant for water vapor, and the turbulent vertical mixing; \(\varepsilon_e = R_s/R_e\) where \(R_e\) is the gas constant for dry air. The terms \(m(D_s)\) represent the mass of the ice crystal/cloud droplet of diameter \(D_s\), which is the mean diameter of the size distribution. The variables related to moist processes (PRES, PREC, PRA, PCA, PRM, PID, PCC, PCI, PII, PACCI, PACCC, PRSI, PRC, PRH, PRG) represent the sources and sinks of the microphysical fields, and they are defined in Appendix A. A detailed formulation of these microphysical processes is given by Dudhia [1989] and Grell et al. [1995]. The saturation budget equation follows the formulation of Rogers and Yau [1989] (p.119). \(S\) represents the mean value or large-scale value of the saturation ratio of the grid cell. Sub-scale variations of \(S\) are not considered. The first term in (8) represents the change of \(S\) due to air mass radiative and advective cooling and the change of \(S\) due to condensation and deposition. In our simulations with the single-column model, the advective terms in equations (2) to (7) are neglected due to the absence of observations. Adveotive tendencies in the equations for water vapor and saturation are provided by the European Centre for Medium Range Weather Forecasting (ECMWF) reanalyses as described below. The vertical eddy diffusion \(d\) is determined by the K theory. The eddy diffusion coefficient is a function of the local Richardson number as described by Grell et al. [1995]. Cloud entrainment is neglected.

Seven microphysical processes are treated differently in the new scheme relative to the MM5 scheme: the cloud water droplet nucleation (PCII), ice crystal nucleation (PII), the condensation/evaporation onto/of cloud water droplets (PCC), the deposition/sublimation onto/of ice crystals (PID), the heterogeneous freezing of water droplets (PRH), the homogeneous freezing of water droplets and interstitial aerosols (PRG), and the auto conversion of ice crystals to snow (PII). The treatment of these microphysical processes is described by GB [2000b] for the new scheme and by Grell et al. [1995] for the MM5 scheme. In this paper, because of their important role in the results in this research the treatment of the heterogeneous freezing of cloud droplets and autoconversion of cloud ice to snow is described below.

### 2.1. Heterogeneous Freezing

The freezing of cloud water droplets is an important process that determines the formation and longevity of mixed-phase clouds. Laboratory experiments [e.g., Pruppacher and Klett, 1997] and measurements in the Arctic [Pinto, 1998] have shown that cloud glaciation varies as a function of the temperature and the cloud droplet size, the largest having more chances to freeze due to their higher probability to encounter an IFN in the atmosphere.

In the new bulk microphysics scheme the heterogeneous freezing of cloud water droplets depends on cloud droplet size, temperature, and IFN concentration. The parameterization of Meyers et al. [1992] is used for the IFN concentration. The median freezing temperature \(T_f\), which is defined as the temperature at which half of the cloud droplets of a certain size freeze, is determined by the parameterization of Heverly [Pruppacher and Klett, 1997, p. 351] as follows:

\[
T_f = \frac{1}{a} \ln \left( \ln \frac{D_c^{2}}{BV} \right)
\] (9)

where \(a = 0.65 \ (C)^{-1}, B = 2.0 \times 10^8 \ cm^{-1} \ s^{-1} \) and \(V\) is the droplet volume. In the model, the temperature and the cooling rate are used in (9) to determine \(V\), the volume of water droplets that freeze heterogeneously. By integrating from the diameter \(D\) of water droplets of volume \(V\) to infinity over the assumed lognormal size distribution of cloud water droplets, one finds the potential number of water droplets per cubic meter which freeze heterogeneously \(N_H\):

\[
N_H = \frac{\int_{D_c}^{\infty} N_s \exp \left( -\frac{(\ln D - \ln D_c)^2}{2(\ln \sigma)^2} \right) d(ln D)}{(2\pi)^{1/2} \ln \sigma}
\] (10)

where the standard deviation of the lognormal size distribution of water droplets \(\sigma\) is assumed to be 1.3. The real number of water droplets experiencing heterogeneous freezing is limited by the number concentration of ice-forming nuclei, \(C_\sigma\) given by the parameterization of Meyers et al. Therefore the rate of heterogeneous freezing of water droplets (PRH) is

\[
(PRH)_{GBG} = \min \left[ \frac{N_H m[D_c]}{\rho d t}, \frac{C_\sigma m[D_c]}{\rho d t} \right]
\] (11)

In the MM5 bulk microphysics scheme the heterogeneous freezing of water droplets depends on temperature only and follows Bigg [Pruppacher and Klett, 1997]:

\[
(PRH)_{MM5} = B \exp \left[ A(T_0 - T) \right] \frac{D_c^2}{\rho d t N}
\] (12)

where \(B = 100 \ cm^3 \ s^{-1}, A = 0.66 \ K^{-1}, \rho_c\) is the water density, and the prescribed number concentration of water droplets \(N = 10^4 \ cm^{-3}\).
2.2. Autoconversion of Cloud Ice to Snow

In the MM5 bulk microphysics scheme, the autoconversion of cloud ice to snow is determined as follows:

\[
(PRA)_{MM5} = \max \left[ \frac{\rho q_i - M_{\text{max}} C_{iN}}{\rho \Delta t} \right]
\]

where \( M_{\text{max}} \) is the mass of an ice crystal of 500 mm diameter [Grell et al., 1995]. In the new scheme the same parameterization is used with the exception that the prognostic value of the number concentration of ice crystals (\( N_i \)) is used instead of \( C_{iN} \) as follows:

\[
(PRA)_{GBG} = \max \left[ \frac{\rho q_i - M_{\text{max}} N_i}{\rho \Delta t}, 0 \right]
\]

3. Results

The case chosen for this study was that of May 1-20, 1998, during the FIRE Arctic Cloud Experiment [Curry et al., 2000]. This period is ideal for evaluating the microphysics parameterizations since the period includes a range of ice, liquid, and mixed-phase clouds over a broad range of temperatures.

The parameterizations are evaluated using the single-column version of ARCSyM, following the strategy described by Randall et al. [1996] and Randall and Cripe [1999].

3.1. Observations

Cloud properties were obtained from surface-based and aircraft measurements, which have been described by Curry et al. [2000]. Observations used in this study are briefly described here. Surface-based remote measurements were made almost continuously from the following instruments: cloud radar (35 GHz), micropulse lidar (0.5235 mm), and microwave radiometer (23.8 and 31.8 GHz). The following insitu measurements were obtained from the NCAR C-130 aircraft: the King hot-wire probe, FSSP and Gerber PVM-100A which measure the liquid water content; the SPEC Cloud Particle Imager (CPI) which produces digital images of cloud particles [Lawson et al., this issue] from which ice water content is determined, and the Continuous Flow Diffusion Chamber (CFD) which measures the IFN concentration [Rogers et al., this issue].

These observations were used to initialize and evaluate the model. The observed IFN concentrations were prescribed for the model simulations. However, IFN measurements were obtained only during flights that took place between 2000 and

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**Figure 1.** Temporal evolution of the column liquid water path retrieved from the microwave radiometer (solid line), simulated with the new microphysics scheme (dotted), and simulated by the MM5 microphysics scheme (dashed) for the period May 1-20.
2400 UTC on May 4, 7, 11, and 18. In the simulations the IFN concentration is kept constant between each measurement; that is, IFN value is the same as that of the last measurement. For the first three days, the IFN concentration measured on May 4 is assumed.

Aerosol concentration is specified in the following way: The observed aerosol concentration is not used in the simulations since in-cloud observations are not necessarily indicative of the aerosol concentration prior to the cloud formation due to various cloud-aerosol interactions. We assume a constant aerosol concentration of 300 cm$^{-3}$ which is representative of what was measured by the PMS active scattering aerosol spectrometer probe (ASASP) in clear sky condition. A Junge size distribution is assumed with a slope of 2.5, following GB (2000b). A slope of 2.5 may appear somewhat surprising since slopes varying between 3.0 and 4.0 are usually reported in the literature (Pruppacher and Klett, 1997, p. 262). At 0.05-0.1 mm radius, which corresponds to the modal radius range of the aerosol size distribution, the slope decreases dramatically to reach zero at the maximum of the size distribution. The choice of a small aerosol slope made here is based on the results from explicit microphysics models of GB (2000a) and Khvorostyanov and Curry [1999] which show that the size of aerosols that contribute most to the formation of cloud particles is around 0.1 mm. The aerosol soluble fraction is specified to be 50%, with ammonium sulfate comprising the soluble portion of the aerosol in an internal mixture. The homogeneous freezing temperature of the unactivated solution droplets that form the interstitial aerosols is prescribed to be -40°C, which corresponds to the homogeneous freezing temperature of highly diluted ammonium sulfate aerosols [Cicco and Abbott, 1999]. The homogeneous freezing is assumed to be instantaneous; that is, all diluted solution droplets freeze within a time step at temperature below -40°C.

Surface temperature was determined from the SHEBA 20m tower data. The surface is assumed to be a mixture of multiyear sea ice and leads (open water). The observed open water fraction and surface albedo is used. The ocean surface temperature in leads is assumed to be -1.8°C. The surface pressure tendency, vertical velocity and large-scale horizontal advective tendencies of temperature, humidity, and momentum were obtained from initialized analysis of the European Center for Medium-Range Weather Forecasting (ECMWF) numerical weather prediction model. Horizontal-large scale tendencies have also been derived from aircraft observations in the case of May 18 due to the poor quality of ECMWF advective tendencies for this day. On May 18 the aircraft performed 20 km by 20 km box patterns flights at 540 m and 30 m. Measurements taken during
these two flights were used to calculate temperature and humidity advective tendencies at these heights. This technique was not used for other days since it gives only punctual values of the advective tendencies and may not be representative of the entire day. In the May 18 case, the observed advective tendencies were assumed for the entire day in order to see if the microphysics schemes were able to produce the observed clouds with the right advective tendencies.

The formation and dissipation of clouds are very sensitive to the specified boundary advection, which undoubtedly contains errors. Hence the focus of our analysis is not so much on the detailed timing of formation and dissipation of clouds but rather on cloud phase and production of hydrometeors. This focus allows us to evaluate the microphysical processes in the new and MM5 schemes and to assess the utility of the enhancements in the new scheme. So our analysis addresses, in particular, the comparison of the new and MM5 schemes and available observations on cloud microphysical properties.

3.2. Baseline Simulations

Twenty days of simulations were performed using versions of the ARCSYM that incorporated the new or MM5 microphysics scheme, using the initial conditions described above (here-
in after referred to as the baseline simulations). To reduce the impact of errors in advective tendencies [Pinto et al., 1999], the temperature and humidity profiles are reinitialized each 12 hours with the observed values provided by the aerological soundings. This method reduces the discrepancies between simulation and observation after many days of integration and allows a better evaluation of the microphysics schemes.

Figure 1 shows the liquid water path (LWP) for the 20 day simulation obtained with the new scheme and the MM5 scheme compared to the observed value determined from retrievals obtained from the microwave radiometer. Values from the microwave radiometer are hourly averaged calculated from 2 min samples. LWP values observed during precipitation events have been removed. Because of high sample frequency and few continuous precipitation events during this period, only seven hours of data are missing. The MM5 scheme underestimates the LWP for the first seven days and does not produce any significant liquid water afterward. The new scheme also appears to underestimate the LWP, but it is much closer to observations than the MM5 scheme. The large peak obtained with the new scheme on day 1 is likely to be due to the absence of model spin-up. Over the 20 days, the mean value of LWP simulated by the new and MM5 schemes are 42.3 and 1.6 g m\(^{-2}\), respectively, compared to 47.5 g m\(^{-2}\) for observations. Curry et al. [2000] simulate the same case with the ARCSyM model using a simple microphysical scheme of Dudhia [1989] and with the Colorado State University single column model (CSU) [Fowler et al., 1996]. They also obtained an underestimate of the LWP, both models producing LWP peaks of less than 50 g m\(^{-2}\) over the 20 days. According to Curry et al. [1990], comparison of aircraft measurements of LWP with the microwave radiometer measurements indicates that the microwave radiometer values are too high.

Figure 2 shows that the ice water path (IWP) is generally much higher in the simulation done with the new scheme when compared to the simulation with the MM5 scheme. Determinations of observed ice water content are preliminary and have been determined for only a few periods and are not shown here.

Three periods corresponding to different cloud configurations can be identified as follows: (1) May 1-10 is characterized by a persistent boundary layer with mixed-phase cloud having tops between 500 m and 1000 m. Intermittent upper level clouds are present, which are predominantly ice. (2) May 11 is characterized by a deep mixed-phase cloud with a top at 6 km which dissipates on May 12. A persistent boundary layer liquid cloud then formed on May 12 which persists until May 16. (3) May 18 is characterized by a complex multilayer cloud structure with a shallow boundary layer liquid cloud with a top at 500 m, an intermediate level cloud between 3 and 4 km, and a dissipating ice cloud between 6 and 8 km. At least one aircraft flight with in situ cloud measurements was made during each of these three periods.

During May 1-5, the liquid water path observed by the microwave radiometer is partially reproduced by the new scheme (see Figure 1). From May 1 to May 4, 1200 UTC, the new scheme reproduces the three observed peaks either earlier or later in time. At about 1200 UTC on May 4 the observed LWP peak is reproduced by the new scheme. However, it reaches its maximum at 0000 UTC on May 5 and quickly dissipates afterward as opposed to an observation that shows persistence of this peak until May 6. Figure 2 shows a sharp increase of the ice water content on May 5 which indicates the new scheme may overestimate the heterogeneous freezing of water droplets for this case. Furthermore, the amplitude of the negative largescale humidity advection provided by the ECMWF might be too high, which could lead to cloud evaporation. The MM5 scheme strongly underestimates the LWP for the first five days. Furthermore, the IWP is much lower than that of the new scheme.

Observations of the May 4 cloud are described in detail by Curry et al. [2000]. The time series of cloud radar returns shows a persistent boundary layer cloud surmounted by altostratus
clouds that had almost disappeared by 2200 UTC when the aircraft arrived on site. The boundary layer was characterized by a cloud-topped mixed layer, with a cloud top at 1080 m and a base at 660 m. Temperatures range from -16°C at the surface to -24°C at 1080 m. Figure 3 shows the vertical profile of liquid and ice water content predicted by the new and MM5 schemes compared to the aircraft CPI and King hot-wire probe measurements at 2300 UTC on May 4. The observed LWC peak at 1000 m is reproduced by the new scheme 150 m lower in the troposphere, while the MM5 scheme underestimates it substantially. A secondary smaller peak at 150 m is also produced by the new scheme. Observation shows an IWC peak at 150 m and no liquid water. It appears that the new scheme underestimates the heterogeneous freezing rate of water droplets in this case. Both schemes underestimate the ice water content between the surface and 1 km. While the observation shows a quasi-constant cloud droplet number concentration throughout the cloud layer, the new scheme produces a maximum of cloud droplet number concentration at 700 m. However, when averaged over the cloud layer, the simulated cloud droplet concentration is close to observation with a mean value of 225 and 180 cm⁻³ for the simulated and observed values, respectively. The difference in the vertical distribution of the number concentration of water droplets is likely to be due to mixing processes in the boundary layer.

On May 11, a deep mixed-phase cloud forms at 0300 UTC, associated with advection of cold moist air above 1 km. The upper portion of the cloud slowly dissipates on May 12, leaving a low-level cloud with a top around 1 km. Temperatures range from -6°C at cloud base to -25°C at cloud top. The lidar depolarization indicates multiple thin layers of liquid water. Figure 4 shows the vertical profile of liquid water content simulated with the MM5 and new schemes compared to aircraft observation at 2300 UTC on May 11. Observations show that the largest values of liquid water content are between 1000 m and 1700 m with a small peak at 400 m and another at 2250 m. The new scheme strongly overestimates the lowest peak and does not reproduce the upper peak at 2250 m. Simulated liquid water content in the middle region between 1000 and 1700 m is strongly underestimated by the model. Further, a large peak in liquid water content at 3 km is formed with the new scheme. Radar images indicate relatively strong echoes at this height between 1200 UTC and 2400 UTC. Lidar images also indicate that an intermittency of liquid and ice phase was likely to occur at this altitude. This is confirmed by the microwave radiometer retrieval that shows a LWP maximum at 1200 UTC on May 11. Therefore it appears that the new scheme reproduces the situation observed a few hours earlier. Once again, the MM5 moisture scheme strongly underestimates the liquid water content at all levels. The MM5 cloud scheme produces virtually no liquid water for this case, producing a cloud that is almost entirely crystalline.

On May 18, strong cold air advection near the surface occurs and a boundary layer cloud forms. Furthermore, a multilayer
upper ice cloud from 3-8 km slowly dissipates throughout the day. Comparison of the ECMWF advection with values estimated from aircraft observations shows that ECMWF substantially underestimates the cold air advection in this case; hence the aircraft advective tendencies are used for the entire day. When the aircraft advective tendencies are used, the new scheme reproduces quite well the liquid boundary layer cloud formation at 0200 UTC and its evolution until the end of the day. The MM5 cloud scheme, even with the aircraft advection, does not reproduce this cloud. Both schemes reproduce the upper layer clouds. Figure 5 shows the vertical profile of liquid water content as simulated by the new scheme compared to aircraft observations at 2300 UTC. The maximum at 425 m is well captured by the model, although the magnitude of the peak is strongly overestimated. Also, a second large maximum appears in the first few meters above the surface.

These discrepancies are likely to be due to the large scale tendencies assumed in the model simulation. Aircraft advective tendencies have been used in the simulation starting from May 18 at 0000 UTC to 2400 UTC. Measurements of these advective tendencies were made around 2200 UTC on May 18. Therefore the assumed constant advective tendencies during the whole day is likely to produce this error in the liquid water content vertical profile. Real advective tendencies probably fluctuated during the day. The strong cool advection calculated from the aircraft was probably weaker or of opposite sign earlier that day. On the basis of the vertical profile of the liquid water content at 1200 UTC (not shown here), one can assume that the aircraft advective tendencies were probably representative of the period 1100 UTC to 2400 UTC.

3.3. Discussion

Comparison of the two microphysics schemes over the 20 day period shows that the MM5 scheme regularly produces less cloud liquid and ice water than the new scheme. The aerosol activation scheme and the prognostic saturation ratio allow for a more physically based treatment of the heterogeneous freezing
of water droplets and the autoconversion rate of cloud ice to snow. The heterogeneous freezing rate of cloud water droplets is much larger in the MM5 scheme. In the new scheme, in addition to temperature the heterogeneous freezing rate depends on cloud droplet size and IFN concentration while in the MM5 scheme, the heterogeneous freezing rate depends only on temperature. Besides the treatment of the heterogeneous freezing rate in the MM5 scheme, the cold bias of the ECMWF advective tendencies in the low levels is likely to amplify the overestimation of the heterogeneous freezing rate by the MM5 microphysics scheme. The new scheme appears to be less influenced by the ECMWF cold bias due to the size dependence of the heterogeneous freezing of water droplets. Indeed, the ECMWF cold bias leads to larger cooling rates that contribute to increase supersaturation and produce smaller water droplets. In turn, smaller droplets freeze at lower temperature. Consequently, a cloud can remain in liquid phase even at a lower temperature.

Since autoconversion of ice crystal to snow depends on the ice crystal concentration, snow formation is more efficient in the MM5 scheme. Furthermore, in the MM5 scheme the critical value of ice water content above which the autoconversion of cloud ice to snow is initiated is a function of the IFN concentration. When the IFN concentration is low, as in the upper atmosphere or in the boundary layer after May 10, the autoconversion of cloud ice to snow is initiated quickly. Figure 6 shows that the percentage of snow mass to the total ice mass (cloud ice plus snow added) is much larger in the simulation with the MM5 scheme. This situation favors the increase of precipitation and the decrease of cloud droplet and cloud ice concentration by Bergeron effect and collision process in the simulation with the MM5 scheme. In the new scheme a combination of high ice water content and low ice crystal number concentration is necessary to initiate the autoconversion process. As the number concentration of ice crystals is prognostic and often higher than the IFN concentration due to sedimentation, the autoconversion of cloud ice to snow is less important in this case, as shown in Figure 6.

4. Sensitivity Experiments

To further interpret the problems and successes of the new microphysics scheme in simulating these clouds, we perform several sensitivity experiments. The effect on cloud formation of four parameters related to aerosol microphysical properties has been assessed: the ability of aerosols to nucleate ice (IFN concentration), the number concentration of aerosols, the aerosol solubility, and the aerosol size distribution. The baseline simulation with the new moisture scheme serves as a comparison basis.

![Figure 8](image_url) Figure 8. Time series of the liquid water path (solid line) and ice water path (dotted line) difference between the simulation performed with the Meyers et al. IFN concentration and baseline simulation (observed IFN concentration) (perturbed simulation minus baseline simulation).
Table 1. Mean Liquid Water Path (LWP) and Ice Water Path (IWP) (g m$^{-2}$) for Various Aerosol Assumptions

<table>
<thead>
<tr>
<th>Simulation Specifications</th>
<th>IWP, g m$^{-2}$</th>
<th>LWP, g m$^{-2}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Baseline simulation: observed IFN concentration, slope of the aerosol size distribution of 2.5, aerosol solubility of 50%, and aerosol concentration of 300 cm$^{-3}$</td>
<td>12.6</td>
<td>42.3</td>
</tr>
<tr>
<td>Baseline with Meyers et al. [1992] IFN concentration</td>
<td>14.3</td>
<td>37.3</td>
</tr>
<tr>
<td>Baseline with the slope of the aerosol size distribution = 3.0</td>
<td>12.6</td>
<td>42.1</td>
</tr>
<tr>
<td>Baseline with the slope of the aerosol size distribution = 2.0</td>
<td>12.8</td>
<td>42.4</td>
</tr>
<tr>
<td>Baseline with aerosol concentration = 600 cm$^{-3}$</td>
<td>12.0</td>
<td>42.5</td>
</tr>
<tr>
<td>Baseline with the aerosol solubility = 90%</td>
<td>12.5</td>
<td>42.4</td>
</tr>
</tbody>
</table>

See text. All simulations are performed with the same condition as the baseline simulation described in section 3.2 except for the specifications mentioned below.

4.1. IFN Concentration

To assess the model sensitivity to IFN concentration, we compare simulations using parameterizations from Fletcher [1962] and Meyers et al. [1992]. Figure 7 compares these two parameterizations of IFN concentrations for three cases, May 4, 7, and 18, against observed values of IFN concentration. In general, the observed IFN concentration is much higher than predicted in the lower troposphere, whereas it is smaller in the upper atmosphere. With regard to the lower troposphere, Rogers et al. [this issue] hypothesized that the large observed IFN concentration might be associated with open leads observed in May from which originate ocean bacteria that may act as IFN [Szyrner and Zawadzki, 1997].

To assess the effect of changing IFN concentration on cloud formation, the baseline simulation using the new parameterization was repeated with the Meyers et al. IFN concentration and by reducing the Meyers et al. IFN concentration by a factor of 10. Figure 8 compares the time series of the LWP and IWP difference between the baseline and the reduced IFN runs. Results show that both liquid and ice water path are affected by the IFN concentration change. On average, the LWP is reduced by 12%.

Figure 9. Time series of the liquid water path (solid line) and ice water path (dotted) difference between the simulation performed with an aerosol concentration of 600 cm$^{-3}$ compared to the baseline simulation (aerosol concentration of 300 cm$^{-3}$).
Figure 10. Time series of the liquid water path (solid line) and ice water path (dotted) difference between the simulation performed with (a) the slope of the aerosol size distribution of 3.0 and baseline simulation (slope of 2.5) and (b) the slope of the aerosol size distribution of 2.0 and baseline simulation (slope of 2.5) (perturbed simulation minus baseline simulation).

and the IWP is increased by 14% in the Meyers et al. IFN cases when compared to baseline run (see Table 1). Smaller IFN concentration leads to fewer ice crystal nucleation, causing saturation to remain high, which favors cloud droplet nucleation. Figure 8 shows that the impact of changing IFN concentration is variable for different cloud conditions. The largest impact occurs on May 11, which is caused by two factors: (1) advection contributes to a relative humidity increase, and (2) the relatively low in-cloud temperature combined with the very low observed IFN concentration above the boundary layer make the difference between observed IFN concentration and that obtained with Meyers et al. [1992] parameterization very large. A substantial increase of ice water content and decrease of liquid water content occurs in that case. Note that occasionally, when the observed IFN concentration is higher than that predicted by the Meyers et al. parameterization, the opposite effect occurs, i.e., a decrease of ice water and an increase of liquid water. This is the case in the boundary layer in the first few days and on May 18.

4.2. Aerosol Number Concentration
Besides seasonal variations the aerosol concentration in the Arctic experiences large variations in short periods of time [e.g., Barrie, 1986]. To assess the sensitivity of the mixed-phase cloud formation to the aerosol number concentration, we compared two simulations with an aerosol concentration of 600 and 900 cm$^{-3}$, respectively, to the baseline simulation (300 cm$^{-3}$). Figure 9 shows the LWP and IWP obtained from the baseline simulation compared to the simulation with the aerosol concentration of 600 cm$^{-3}$. On average, the LWP increases and the IWP decreases as shown in Table 1. The increase of the aerosol concentration provides additional cloud condensation nuclei that contribute to increase the cloud droplet number concentration and reduce their mean diameter. The mean cloud droplet number concentration increases by 25% and 136% when the aerosol number concentration is increased by factors of 2 and 3, respectively. Large cloud droplets freeze heterogeneously at a higher temperature than small cloud droplets. This explains the increase of the LWP and the decrease of the IWP when the aerosol concentration increases. Figure 9 shows that the effect of increasing aerosol concentration is larger during the first days of the simulation. This is related to temperature that ranges from -15°C to -30°C in the boundary layer during this period. This temperature range corresponds to the heterogeneous freezing temperature of water droplets smaller than 10 $\mu$m radius. Therefore a small change of the mean radius of water droplets has a substantial effect on the heterogeneous rate at these temperatures. Beyond May 10, temperature is warmer and
the change of water droplet radius has a small effect on the heterogeneous freezing rate of water droplets.

4.3. Slope of the Aerosol Size Distribution

A change of the aerosol size distribution may occur as a result of an aerosol population change. Indeed, the slope of the aerosol size distribution will be smaller if giant aerosols are in large concentration and inversely. A different aerosol size distribution leads to a change in condensation nuclei number concentration. GB (2000b) have shown that microphysical properties of arctic diamond dust, ice fog, and thin stratus can be sensitive to a change of the slope of the aerosol size distribution at low temperature. The IFN concentration at low temperature is high and may be higher than \( N_a \), which is the aerosol concentration with a diameter larger than the critical diameter of activation. In these situations, a change of \( N_a \), associated with a change in aerosol size distribution will directly affect the number of ice crystal nucleated. On the other hand, at higher temperature, the IFN concentration is often lower than \( N_a \) and a change of the latter does not affect the number of ice crystals nucleated since it is limited by the IFN concentration. With regard to water droplet nucleation, GB (2000b) found little effect of the slope of the aerosol size distribution since the air mass infrared cooling rate, which is the generator of these clouds, is too weak and does not allow for much water nucleation. Using a 3-D model, Ghan et al. [1997] found that the effect of aerosol number concentration on cloud water content is much smaller in high latitudes when compared to lower latitudes.

However, results from previous studies are not necessarily valid for conditions prevailing in the simulated case of this study, which include different formation mechanisms and a wide range of temperatures. Figure 10 shows the baseline simulation with the slope of the aerosol size distribution set to 2.5 compared to simulations with the slope set to 3.0 and 2.0. Results indicate that an increase of the slope of the aerosol size distribution results in a decrease of the mean LWP and an increase of the mean IWP (see Table 1). Although these changes are small on average, LWP and IWP changes can reach up to 20 g m\(^{-2}\) locally. Figure 10 shows that the IWP varies more than the LWP. As the slope of the aerosol size distribution increases, the number of activated condensation nuclei decreases for the same supersaturation. The number concentration of water droplets then decreases and their size increases. Since large cloud droplets freeze heterogeneously at a higher temperature than smaller cloud droplets, the ice crystal concentration increases and cloud droplet concentration decreases. At low temperatures this situation may lead to less ice crystal nucleation for the reasons described above. On the other hand, ice crystal number concentration may also decrease due to the fact that \( N_a \) decreases as the slope of the aerosol size distribution increases. Therefore the IWP variation appears to be related to whether water saturation is reached and cloud droplets nucleate.
4.4. Aerosol Solubility

The sensitivity of the microphysics scheme to the specification of aerosol soluble fraction was examined by performing a simulation with 90% soluble fraction (compared to 50% for the baseline simulation). Figure 11 shows the difference between these two simulations. The effect of increasing aerosol soluble fraction is small and is effective only when cloud droplet nucleation occurs. On average, liquid water increases and ice water decreases as the aerosol soluble fraction increases (see Table 1). As the soluble fraction increases, the size of aerosols increases by absorption of water vapor. This leads to an increase of $N_i$ and therefore an increase of the number of cloud droplet nucleated. On the other hand, this situation gives rise to the formation of smaller water droplets that freeze heterogeneously at lower temperature, resulting in a decrease of the ice water content.

5. Conclusions

A new bulk microphysics scheme is described and assessed against data from the FIRE Arctic Clouds Experiment during May 1998. This microphysics scheme is distinguished from other similar schemes by the addition of three prognostic variables: the saturation ratio and the number concentrations of water droplets and ice crystals. A bimodal lognormal cloud size distribution is assumed. Furthermore, prescribed aerosol size distribution and aerosol microphysical properties are used as input for the parameterization of cloud microphysical processes. These features allow for a more elaborate treatment of freezing processes and nucleation of water droplets and ice crystals.

The new scheme was compared with the MM5 bulk microphysics scheme and also with observations during the period May 1 to 20. The parameterizations were tested in the context of a single-column model, forced by advection determined from ECMWF analyses. Qualitatively, both the new and the MM5 schemes reproduce relatively well the formation and dissipation of observed clouds during this period. This study focused particularly on cloud phase and water content. Comparison with observations was hampered by uncertainties in the ECMWF large-scale tendencies, possible errors in the liquid water path retrievals from the microwave radiometer, and the absence of many cases where a determination of ice water content has been done. Nevertheless, a qualitative picture of the performance of the two microphysics schemes has emerged from this comparison. Both schemes generally underpredict the liquid water path when compared with observations, with the MM5 values being an order of magnitude smaller than those simulated using the new scheme. The two microphysics schemes were also compared to aircraft measurements taken on
May 4, 11, and 18, which included mixed-phase clouds at temperatures varying from -40° to -15°C. Results show that the new scheme reproduces the vertical profile of liquid water content and ice water content for the May 4 case (boundary layer mixed-phase cloud) and the May 18 (boundary layer liquid cloud) case. On May 11 a deep mixed-phase cloud, predominantly crystalline, was observed. The new scheme produces a predominantly liquid cloud at that time. The sensitivity study clearly showed that this error is caused by the underestimation of the IFN concentration. The MM5 scheme generally underestimates aircraft measurement of liquid and ice water content for all three cases. It appears that the MM5 scheme overestimates ice phase over liquid phase leading to an overestimation of the autoconversion rate of ice crystal to snow. Thus the MM5 scheme overestimates snow precipitation and collection processes. As a result, both ice water and liquid water content are smaller.

The relative success of the new scheme in predicting cloud phase is due to the prognostic determination of cloud water droplet and ice crystal number concentration which allows for a more elaborated treatment of the ice crystal nucleation rate by heterogeneous freezing of cloud droplets. This treatment allows for the persistence of mixed-phase clouds over a broad range of temperatures.

Sensitivity studies revealed that the IFN concentration is the most important parameter in a microphysics scheme with regard to the formation of ice and mixed-phase cloud. Changing IFN concentration not only modifies the cloud phase but also the total water content of a cloud. Replacing observed IFN concentration by the parameterization of Meyers et al. [1992] leads to a decrease of the mean LWP of 12% and an increase of the IWP by 14%. This test clearly demonstrates the importance of the IFN concentration and the need for a more realistic IFN representation in climate models.

Three other parameters related to aerosols also appear to substantially affect cloud microphysical properties under certain conditions. It was found that as the slope of the aerosol size distribution increases, the LWP decreases and IWP increases. This result is explained in part by the fact that increasing the slope of the aerosol size distribution leads to fewer and larger cloud water droplets which in turn favors their heterogeneous freezing. The number concentration of aerosols can also substantially modify cloud phase. As the aerosol concentration increases, the LWP increases and the IWP decreases. The additional cloud condensation nuclei provided by the increase of aerosol concentration favors the formation of more cloud droplets with smaller diameters. Consequently, the heterogeneous freezing rate of water droplets decreases. The aerosol solubility appears to have less effect on cloud phase and water content. The effect of increasing solubility is to increase LWP and decrease IWP. When aerosol solubility is higher, the number of cloud condensation nuclei increase and as a result, larger concentration of smaller droplets are nucleated.

Aerosol composition and size distribution can affect substantially cloud phase and cloud water content. Results with the new microphysics scheme have shown that cloud water content and phase are strongly related to aerosol microphysical properties. A better knowledge of aerosol microphysical properties as well as climate models with realistic prescribed aerosol climatology or prognostic aerosols are required to better simulate clouds, particularly in the Arctic where boundary layer mixed-phase clouds are common.

**Notation**

- **PACCC**: Collection of cloud water droplet by rain, /s.
- **PACC1**: Collection of ice crystals by snow, /s.
- **PCA**: Autoconversion of cloud water droplet to rain, /s.
- **PCC**: Condensation/evaporation of cloud water droplets, 1996.
- **PCI**: Nucleation of cloud water droplets, /s.
- **PID**: Deposition/sublimation onto/off ice crystals, /s.
- **PII**: Nucleation of ice crystals, /s.
- **PKR**: Rain precipitation, /s.
- **PKS**: Snow precipitation, /s.
- **PRA**: Autoconversion of cloud ice crystals to snow, /s.
- **PREC**: Evaporation of rain, /s.
- **PRES**: Sublimation of snow, /s.
- **PRG**: Homogeneous freezing of water droplets, /s.
- **PRH**: Heterogeneous freezing of water droplets, /s.
- **PRM**: Ice crystals melting, /s.
- **PRSC**: Cloud droplet sedimentation, /s.
- **PRSI**: Ice crystal sedimentation, /s.

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