Arctic Mixed-Phase Clouds Simulated by a Cloud-Resolving Model: Comparison with ARM Observations and Sensitivity to Microphysics Parameterizations

YALI LUO*
National Institute of Aerospace, and NASA Langley Research Center, Hampton, Virginia

KUAN-MAN XU
NASA Langley Research Center, Hampton, Virginia

HUGH MORRISON
National Center for Atmospheric Research, + Boulder, Colorado

GREG McFARQUHAR
University of Illinois at Urbana-Champaign, Urbana, Illinois

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ABSTRACT

Single-layer mixed-phase stratiform (MPS) Arctic clouds, which formed under conditions of large surface heat flux combined with general subsidence during a subperiod of the Atmospheric Radiation Measurement (ARM) Program’s Mixed-Phase Arctic Cloud Experiment (MPACE), are simulated with a cloud-resolving model (CRM). The CRM is implemented with either an advanced two-moment [Morrison et al. (MCK)] or a commonly used one-moment [Lin et al. (LFO)] bulk microphysics scheme and a state-of-the-art radiative transfer scheme.

The MCK simulation, which uses the two-moment scheme and observed aerosol size distribution and ice nuclei (IN) number concentration, reproduces the magnitudes and vertical structures of cloud liquid water content (LWC), total ice water content (IWC), and number concentration and effective radius of cloud droplets as suggested by the MPACE observations. The simulation underestimates ice crystal number concentrations by an order of magnitude and overestimates effective radius of ice crystals by a factor of 2–3. The LFO experiment, which uses the one-moment scheme, produces values of liquid water path (LWP) and ice plus snow water path (ISWP) that were about 30% and 4 times, respectively, those produced by MCK. The vertical profile of IWC exhibits a bimodal distribution in contrast to the constant distribution of IWC produced in MCK and observations.

A sensitivity test that uses the same ice–water saturation adjustment scheme as in LFO produces cloud properties that are more similar to the LFO simulation than MCK. The mean value of the intercept parameter of snow size spectra ($N_{0s}$) from MCK is one order of magnitude smaller than that assumed in LFO. A sensitivity test that prescribes the larger LFO $N_{0s}$ results in 20% less LWP and 5 times larger snow water path than that in MCK. When an exponential ice size distribution replaces the gamma size distribution in MCK, the ISWP decreases by 70% but the LWP increases by 7% versus that in the MCK. Increasing the IN number concentration from the observed value of 0.16 to 3.2 L$^{-1}$ forces the MPS clouds to become glaciated and dissipate, but the simulated ice number concentration agrees initially with the observations better. Physical explanations for these quantitative differences are provided. It is further shown that the differences between the LFO and MCK results are larger than those due to the estimated uncertainties in the prescribed surface fluxes. Additional observations and simulations of a variety of cases are required to further narrow down uncertainties in the microphysics schemes.

* Current affiliation: State Key Laboratory of Severe Weather, Chinese Academy of Meteorological Sciences, Beijing, China.

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Corresponding author address: Dr. Yali Luo, State Key Laboratory of Severe Weather, Chinese Academy of Meteorological Sciences, Beijing 100081, China.
E-mail: yali@cams.cma.gov.cn

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1. Introduction

Atmospheric numerical models with a horizontal grid spacing of 1–2 km are known as cloud-resolving models (CRMs). CRMs are able to resolve convective-scale and mesoscale circulations and, hence, can better represent the interactions between physical processes involving smaller scales than general circulation models (GCMs). Physical processes, such as those involving clouds and precipitation, cannot be explicitly resolved and have to be parameterized in GCMs because of the grid spacing of a GCM, which is typically on the order of 100 km in the horizontal and 1 km in the vertical. There are still large uncertainties in parameterizations of subgrid-scale processes, and improvement of parameterizations has been slow in spite of the enormous efforts made over the past decades (Randall et al. 2003). In GCMs, subgrid-scale processes interact mainly through the time-varying large-scale variables (and surface conditions), while in reality they directly interact with each other. A unified formulation of the entire spectrum of these interactions is necessary for more accurate climate and weather prediction, but it is difficult to achieve this with the large grid spacings used in GCMs (Arakawa 2004). With the rapid growth of computational capacity, continental-scale numerical weather prediction is currently performed at cloud-resolving scales [e.g., Weather Research and Forecasting (WRF); see Skamarock et al. (2005)]. For climate simulation, CRMs have been used as a “superparameterization” to replace most of the traditional parameterizations in each grid cell of GCMs (e.g., Grabowski 2003), and global versions of CRMs are emerging (Tomita et al. 2005).

Microphysical processes, as well as turbulent and radiative transfer processes, still need to be parameterized in CRMs. Most CRMs rely on bulk microphysics schemes to represent the complicated interactions between atmospheric thermodynamic states and hydrometeors and among various hydrometeor species. Bulk microphysics schemes typically divide the hydrometeor spectrum into cloud water, cloud ice, rain, and one or more ice-phase precipitation species (e.g., snow, graupel, and hail). Each hydrometeor class is represented by a specified size distribution function (e.g., gamma, exponential, and lognormal). The microphysics schemes that predict only hydrometeor mixing ratios are called the one-moment approach (e.g., Lin et al. 1983, hereafter L83). An improvement to the one-moment approach is to predict the rates of change for both mixing ratios and number concentrations of hydrometeors, that is, the two-moment approach (e.g., Ferrier 1994; Meyers et al. 1997; Morrison et al. 2005, hereafter M05; Phillips et al. 2007). An advantage of this approach is that the effective sizes of cloud particles, one of the most important parameters determining cloud radiative impacts, can be predicted, in contrast to the one-moment approach. Another advantage is that two-moment schemes potentially can represent the size distributions of hydrometeors more realistically and, thus, represent microphysical processes more accurately than one-moment schemes (e.g., Meyers et al. 1997; Morrison and Pinto 2006).

Arctic clouds have been identified as playing a central role in the Arctic climate system. However, the role of clouds is even less well understood in the Arctic than in other geographic regions owing to sparse observations. Arctic field programs, such as the Surface Heat Budget of the Arctic (SHEBA; Uttal et al. 2002) and the First International Satellite Cloud Climatology Project (ISCCP) Regional Experiment (FIRE) Arctic Cloud Experiment (ACE; Curry et al. 2000), revealed that mixed-phase stratiform (MPS) clouds appear to dominate the low-cloud population within the Arctic (Intrieri et al. 2002). The Arctic mixed-phase clouds are long-lived (5–10 days) liquid-topped clouds that precipitate ice (Hobbs and Rangno 1998; Curry et al. 2000; Intrieri et al. 2002). They occur in an environment that is generally depleted in ice-forming nuclei (IFN; e.g., Curry et al. 2000; Sirois and Barrie 1999) and in a radiative environment in which the net longwave radiation dominates the cloud-top radiative energy budget (e.g., Zuidema et al. 2005).

Adequate simulation of Arctic clouds is needed to address Arctic cloud–radiative–surface interactions that may impact global climate (e.g., Curry et al. 1996) and to predict weather related to the persistence and large horizontal extent of these cloud systems. However, there have been few simulations of Arctic MPS clouds with CRMs, primarily because the observations of cloud physical properties needed to evaluate model performance are sparse and there is a lack of large-scale forcing data available to drive CRMs. The Department of Energy (DOE) Atmospheric Radiation Measurement (ARM) Program (Stokes and Schwartz 1994; Ackerman and Stokes 2003) recently launched its Mixed-Phase Arctic Cloud Experiment (MPACE), 27 September to 22 October 2004, at the North Slope of Alaska (NSA) sites (Harrington and Verlinde 2004; Verlinde et al. 2007). During the field campaign, detailed information about Arctic clouds were measured using the ARM millimeter-wave cloud radar, multipulse lidar, laser ceilometers, microwave radiometer (MWR), and three instrumented aircraft. Furthermore, the large-scale forcing data were derived for a 17.5-day Intensive Observation Period in October 2004 (Xie et
al. 2006) by applying the method of Zhang and Lin (1997) and Zhang et al. (2001) to the available data. These forcing data can be used to drive models [CRMs, single-columns models (SCM; Randall et al. 1996), and large-eddy simulation (LES) models].

The objectives of this study are twofold. One is to evaluate CRM simulations of Arctic MPS clouds with a state-of-the-art dataset. The available MPACE data offer a promising opportunity for improving cloud microphysical parameterizations in CRMs. Here, single-layer MPS clouds observed during a subperiod of MPACE are simulated using a CRM, driven by the ARM-derived large-scale forcing. The CRM includes a state-of-the-art radiative transfer scheme and either a one- or two-moment microphysics scheme. The performance of the CRM is evaluated through comparing simulated cloud properties, such as the vertical profiles of cloud liquid water content (LWC), ice water content (IWC), droplet number concentration, ice number concentration, and effective sizes of droplets and ice crystals, with the MPACE aircraft observations (McFarquhar et al. 2007), as well as the retrievals of liquid water path (LWP; Turner et al. 2007) and observations of precipitation from ground-based instruments deployed at the NSA sites.

The second objective of this study is to explore the sensitivities of the simulated clouds to representation of various microphysical processes and parameters. To achieve this objective, a range of sensitivity tests are conducted. In particular, we attempt to answer the following questions: What differences in the simulated cloud properties are produced by using either a one- or two-moment microphysics approach? What microphysical processes and parameters may significantly influence the simulated MPS clouds?

Section 2 describes the CRM used in this study, with a focus on the prediction of hydrometeor number concentrations. Section 3 gives a description of the case and cloud property observations. Design of the numerical experiments is presented in section 4. Results from the CRM simulations utilizing either the one- or two-moment approach are compared with the aircraft measurements in section 5. Section 6 contains results from the sensitivity tests. A summary and conclusions are given in section 7.

2. The numerical model

The dynamic framework of the CRM used in this study is based on the anelastic forms of hydrostatic, momentum, and continuity equations in two dimensions (x and z) (Krueger 1988; Xu and Krueger 1991) with a third-moment turbulence closure (Krueger 1988). The CRM includes the Fu and Liou (1993) radiative transfer parameterization and either a one-moment or two-moment microphysics parameterization.

The two-moment bulk microphysics scheme of M05 has been implemented, which predicts the mixing ratios and number concentrations of cloud water, cloud ice, rain, and snow. The equation used to predict the hydrometeor number concentrations is

\[
\frac{dn_x}{dt} = -\frac{1}{\rho_0} \partial_x (\rho_0 \overline{n_x^2 w^y}) + A_x + S_x + M_x, \tag{1}
\]

where \(n_x\) is the number concentration with the subscript \(x\) being \(c, i, r,\) or \(s\) for cloud water, cloud ice, rain, and snow, respectively; \(\rho_0\) is the dry air density of the initial (reference) state; \(\overline{n_x^2 w^y}\) is the ensemble mean of the turbulent flux of \(n_x\) in the vertical direction; and \(A_x\) refers to activation (for cloud water) and nucleation (for cloud ice), \(S_x\) represents sedimentation, and \(M_x\) denotes all other microphysical processes. The effects of turbulent fluctuations on number concentrations of raindrops and snow are ignored in the current version of the CRM, and the effects of turbulence on their mixing ratios are also ignored (Krueger 1988). For number concentrations of cloud water and ice, \(K\) theory is applied to determine the turbulence terms; that is,

\[
\rho_0 \overline{n_x^2 w^y} = -\rho_0 K \frac{\partial n_x}{\partial z}. \tag{2}
\]

The exchange coefficient \(K\) is calculated using \(K = cl\sqrt{TKE}\), where \(c\) is a constant (0.24), \(l\) is the turbulence length scale, and TKE is the turbulent kinetic energy. Both \(l\) and TKE are determined by the third-moment turbulence closure (Krueger 1988).

Droplet activation is treated by a physically based scheme (Abdul-Razzak et al. 1998; Abdul-Razzak and Ghan 2000). This scheme not only relates droplet activation to aerosol characteristics, but also couples it with a local cooling rate determined by cloud-scale and subgrid turbulent vertical velocity as well as radiative cooling. The error of the parameterization is less than 10% under a wide variety of conditions (Abdul-Razzak et al. 1998; Abdul-Razzak and Ghan 2000). The turbulent upward motion for the droplet activation calculation is approximated as the square root of the vertical component of TKE per unit mass. Sedimentation of cloud particles is calculated with terminal fall velocities related to particle sizes and air density (Ikawa and Saito 1991). Parameterizations of all other microphysical processes follow M05, including deposition; condensation freezing of ice nuclei; contact- and immersion-freezing nucleation of cloud droplet and raindrops; autoconver-
sion of cloud water to rain and of cloud ice to snow; self-collection of cloud droplets and raindrops; snow aggregation; accretion of cloud droplets, rain, and cloud ice by snow; rime splintering from accreted droplets and raindrops by snow; accretion of cloud water by rain; deposition/sublimation of cloud ice and snow; melting of snow; evaporation of rain and melted snow; saturation adjustment of cloud water as well as the decrease in number concentrations during evaporation/sublimation.

In the M05 scheme, the gamma size distribution is assumed for cloud droplets and cloud ice crystals while the Marshall–Palmer (exponential) size distribution is used for raindrops and snowflakes. A gamma size distribution can be expressed as

\[ N(D) = N_0 D^\mu e^{-\lambda D}, \]

where \( D \) is diameter, \( N_0 \) is the “intercept” parameter, \( \mu \) is the spectral shape parameter, and \( \lambda \) is the slope parameter. The value \( \mu \) is determined by the relative radius dispersion (\( \eta \); defined as the ratio between the standard deviation and the mean radius):

\[ \mu = 1/\eta^2 - 1. \]

Practically, parameters \( N_0 \) and \( \lambda \) can be diagnosed from the specified \( \mu \) and predicted mixing ratio (\( q \)) and number concentration (\( n \)) of the species. That is, only \( \mu \) needs to be specified using the two-moment approach. For the one-moment approach, two of the three parameters (\( N_0 \), \( \mu \), or \( \lambda \)) need to be specified.

For cloud droplets, \( \eta \) is related to the number concentration \( n_c \) in the M05 scheme. However, the exact \( \eta-n_c \) relationships for Arctic clouds are not yet developed. There are currently only a few formulations relating \( \eta \) to \( n_c \), and these are based on observations at lower latitudes. For example, Rotstayn and Liu (2003, hereafter RL03) fitted three curves to measurements in polluted and unpolluted warm stratiform and shallow cumulus clouds. These curves are designed to represent the average variation of \( \eta \) with \( n_c \), as well as lower and upper bounds of this variation. These curves shown in Fig. 1a are defined by

\[ \eta = 1 - 0.7e^{-\alpha n_c}, \]

where \( \alpha \) equals 0.001 for the lower curve, 0.003 for the middle curve, and 0.008 for the upper curve. The corresponding \( \mu-n_c \) relationships are displayed in Fig. 1b. The relationship of Eq. (5) with \( \alpha \) of 0.003 is used in this study. Note that there was considerable scatter in the data used by RL03 to obtain the \( \eta-n_c \) relationship of Eq. (5). Miles et al. (2000) created a database of stratus cloud droplet size distribution parameters, derived from in situ data reported in the existing literature. The datasets included several parameters for 42 marine stratocumulus clouds and 52 continental stratocumulus clouds. These observations, however, do not show a systematic increase or decrease in \( \eta \) with increasing \( n_c \). For cloud ice, a constant \( \mu \) of 5 is used in M05, corresponding to an \( \eta \) of \( \sim 0.408 \). Note that the Marshall–Palmer distribution is a special case of Eq. (3) with \( \mu \) equal to zero.

The CRM also includes the commonly used one-moment bulk microphysics scheme of L83 with modifications to its ice-phase microphysics parameterization by Krueger et al. (1995). This scheme represents the rates of change of mixing ratios for five hydrometeor species (cloud water, cloud ice, rain, snow, and graupel) and is combined with an ice–water saturation adjustment (Lord et al. 1984) to determine the condensation/evaporation of cloud water and deposition/sublimation of cloud ice. Cloud water and cloud ice are assumed to be monodisperse. Precipitating hydrometeor species are assumed to have exponential size spectra. Number concentrations of the precipitating hydrometeor species can be diagnosed from the predicted mixing ratios and specified microphysical parameters describing the hydrometeor size spectra. However, aerosol characterization is not physically linked to the hydrometeor number concentrations.

The water–ice saturation adjustment scheme of Lord et al. (1984) requires assumptions about both the coexistence of cloud water and cloud ice at temperatures less than 0°C and the partitioning between condensation and deposition. Specifically, the Lord et al. scheme assumes that the saturation vapor mixing ratio \( q^* \) is a mass-weighted average of the respective saturation val-

![Fig. 1. (a) The \( \eta-n_c \) relationships represented by Eq. (5) in the text with \( \alpha \) being 0.003 (solid line), and 0.001 and 0.008, respectively (dot–dashed lines). (b) The corresponding \( \mu-n_c \) relationships. See text for further explanation.](image-url)
ues over liquid water and ice at $-40^\circ C \leq T \leq 0^\circ C$ when both cloud water and cloud ice are present. Under subsaturated conditions, cloud water is evaporated first so that the water vapor mixing ratio ($q_v$) would be equal to $q^*$. If subsaturated conditions are still present after all cloud water evaporates, enough cloud ice is sublimated such that $q_v \leq q^*$. On the other hand, production of either cloud water ($\Delta q_c$) or cloud ice ($\Delta q_i$) depends linearly on temperature under supersaturated conditions so that $\Delta q_c = q_v - q^*$ at $T = 0^\circ C$ and $\Delta q_i = q_v - q^*$ at $T = -40^\circ C$. A similar formulation was also developed by Tao et al. (1989) except for removing the iterative adjustment procedure used in Lord et al. (1984).

In the radiation calculation, optical properties associated with cloud water, cloud ice, rain, and snow are calculated following Fu and Liou (1993), Fu (1996), and Fu et al. (1998), which requires information on masses of the hydrometeors, a generalized effective diameter $D_{\text{ge}}$ (Fu 1996) of ice crystals, an effective radius of cloud droplets ($r_e$), and an effective radius of rain droplets. Both cloud ice and snow are treated as ice crystals in the Fu and Liou radiative transfer scheme. The $D_{\text{ge}}$ of cloud crystals is a mass-weighted average of those of cloud ice ($D_{\text{gei}}$) and snow ($D_{\text{ges}}$) obtained from

$$D_{\text{ge}} = \frac{\text{IWC} + \text{SWC}}{\text{IWC}/D_{\text{gei}} + \text{SWC}/D_{\text{ges}}}.$$  

With the M05 cloud microphysics, $r_e$, $D_{\text{gei}}$, and $D_{\text{ges}}$ are determined by the predicted size distributions. With the L83 cloud microphysics, both $r_e$ and $D_{\text{ges}}$ are specified (10 and 120 $\mu$m, respectively) and $D_{\text{gei}}$ is empirically determined from IWC, as in our earlier studies (Xu 2005; Luo et al. 2007). For rain droplets, a typical value of effective radius is used, following Manton and Cotton (1977).

3. Description of the case study

The east-northeast flow brought cold near-surface air from the sea ice located about 500 km north over the warm open ocean adjacent to the northern coast of Alaska (Fig. 2). The contrast between the cold air and warm open ocean resulted in large ocean sensible and latent heat fluxes that, combined with the conditions of large-scale subsidence associated with the high pressure system built over the pack ice to the northeast of the Alaskan coast, promoted a well-mixed cloudy boundary layer. Single layer mixed-phase clouds were formed under these conditions (Verlinde et al. 2007). These clouds were then advected over the Alaskan coast where they were observed at the ARM NSA sites at Barrow and Oliktok Point (Fig. 2). The Active Remote Sensing of Clouds (ARSCL) algorithm (Clothiaux et al. 2000) derived cloud distribution exhibits the presence of single layer stratocumulus in the period 9–14 October 2004 (not shown). The time–height distributions of radar reflectivity, lidar backscatter, and lidar depolarization (e.g., Fig. 6 in Verlinde et al. 2007) reveal the locations of cloud top and cloud liquid base, and the presence of shafts of ice precipitation and/or supercooled drizzle throughout the cloud layer and below cloud.

The bulk microphysical properties of the MPS clouds that occurred during MPACE [i.e., total condensed water content, LWC, IWC, effective radius of supercooled water droplets, effective radius of ice crystals (defined following Fu 1996), total water droplet number concentration, and total ice crystal number concentration] were derived by McFarquhar et al. (2007) from measurements obtained by instruments on the University of North Dakota Citation aircraft. The Citation was equipped with a range of probes for measuring the size, shapes, and phases of the complete range of hydrometeors that can be sampled within a cloud. There was one daily Citation flight on 9 and 12 October, and two flights on 10 October, which occurred in single layer MPS clouds that were similar in structure. The four flights covered a period of $\sim 6.5$ h, with about half of the period used for in-cloud observation. Here the cloud base is defined as the lidar-derived liquid cloud bottom. The cloud top is defined as the cloud-radar-derived cloud top or, when cloud radar data was not available, as the location where the total condensed water content became greater than 0.001 g m$^{-3}$ (McFarquhar et al. 2007). The bulk properties are available at 10-s resolu-

![Fig. 2. Composite visible satellite image from the NASA Terra satellite for 9 Oct 2004. Dots indicate locations of the ARM sites on the North Slope of Alaska: Barrow, Oliktok Point, and Atqasuk.](image-url)
tion, but represent a 30-s running average of the measured ice properties. There were 1131 in-cloud samples obtained from the four flights. The bulk cloud properties sampled by the four flights are used to validate model simulations in this study.

Other evaluation data include measurements of LWP provided by the microwave radiometer (Turner et al. 2007), and those of surface precipitation provided by the ARM Barrow, Atqasuk, and Oliktok Point sites corrected by those measured at the National Weather Service Barrow station. Large uncertainties, however, existed in the ARM surface precipitation measurements during MPACE because of both the blowing snow conditions and the lack of a dense observational network (Xie et al. 2006).

4. Design of CRM simulations

We conduct a set of simulations using the cloud-resolving model described in section 2 to explore the model’s ability to simulate the MPS clouds and its sensitivity to the microphysics scheme and parameter. All of these simulations start with the same initial profiles of the atmospheric state. They are prescribed with the same surface latent and sensible fluxes, large-scale subsidence, and horizontal advection of temperature and moisture. Details of forcing data are described in section 4a. For the sensitivity simulations, different treatments of some microphysical processes and parameters, described in section 4b, are used. The horizontal grid spacing is 2 km. The vertical grid spacing varies with height from 30 to 102 m at heights below 1.9 km and is constant (500 m) above 1.9 km. The domain width is 256 km in the horizontal and 20 km in the vertical. A time step of 5 s is used for all simulations.

a. Initial conditions, large-scale forcing, and aerosol specification

The initial and lower boundary conditions, large-scale forcing data, and aerosol properties provided by Klein et al. (2006) are used in all simulations. The period of our simulation is from 1700 UTC 9 October to 0500 UTC 10 October. The initial profiles of temperature and water vapor are based on the 1700 UTC 9 October sounding at Barrow (Figs. 3a,b) with the inversion height at ~1.4 km. The CRM is initialized with an adiabatic profile of liquid water (Fig. 3b). No ice is present at the initial time. The total water mixing ratio below inversion is 1.95 g kg\(^{-1}\). The CRM starts from horizontally homogeneous fields except for the added random perturbations with a maximum of 0.1 K to the potential temperature field at the lowest several levels.

The forcing data were based on an analysis of the European Centre for Medium-Range Weather Forecasts (ECMWF) model data for the oceanic region adjacent to the NSA sites (Xie et al. 2006) since the ob-
served clouds were formed over the open ocean before reaching the measurement sites (section 3). The magnitude of the large-scale subsidence ($\omega$) linearly increases with decreasing pressure from a zero value at the surface to a value of about 3.3 hPa h$^{-1}$ at and above the inversion (Fig. 3c). This is used to vertically advect all thermodynamic and microphysical variables in the model. The large-scale horizontal advective tendencies of temperature and moisture are also prescribed (Klein et al. 2006; also shown in Figs. 3d,e). Owing to the lack of observations, the large-scale horizontal advective tendency of the cloud variables are set to zero. The CRM’s horizontally averaged winds ($u$ and $v$) are also nudged toward the initial values ($-13$ m s$^{-1}$ for $u$ and $-3$ m s$^{-1}$ for $v$) with a time scale of 1 h (Xu and Randall 1996). Surface sensible and latent heat fluxes are specified as 136.5 and 107.7 W m$^{-2}$, respectively. For radiation purposes, the lower boundary is an open-ocean surface. A SST of 274.01 K, consistent with the measurements made with the Aerosondes (Curry et al. 2004), is used in the upward longwave radiation calculation. The spectral surface albedos for the six bands of the Fu and Liou (1993) radiation code are calculated using the parameterization of Jin et al. (2004).

The CRM’s droplet activation parameterization is physically linked to the characterization of aerosols. We use a bimodal lognormal size distribution of dry aerosol, obtained from a Met One Instruments, Inc., hand-held particle counter (HHPC-6) on board the Aerosonde unmanned aerial vehicle (UAV) and a condensation nuclei counter from the NOAA/Earth System Research Laboratory located near Barrow (Morrison et al. 2008). The size distribution for each mode of the lognormal distribution is represented by

$$\frac{dN}{d\ln r} = \frac{N_r}{\sqrt{2\pi}\ln \sigma} \exp\left(-\frac{\ln^2(r/r_m)}{2\ln^2\sigma}\right),$$

where the parameters $N_r$, $\sigma$, and $r_m$ are the total number concentration, standard deviation, and geometric mean radius of each mode, respectively. For the smaller mode, the values of these variables are 72.2 cm$^{-3}$, 2.04, and 0.052 $\mu$m. The corresponding values for the larger mode are 1.8 cm$^{-3}$, 2.5, and 1.3 $\mu$m. The aerosol composition is assumed to be ammonium bisulfate with an insoluble fraction of 30%, as recommended by Klein et al. (2006), based on observations of Bigg and Leck (2001) and Zhou et al. (2001).

In situ out-of-cloud observations for number concentration of active ice forming nuclei were obtained on 9 and 10 October from the Continuous Flow Diffusion Chamber (Rogers et al. 2001) aboard the Citation aircraft. These measurements represent the total number concentration of active IFN that have diameters less than 2 $\mu$m acting in deposition, condensation-freezing, and immersion-freezing modes. The measured mean concentration of these IFN is about 0.16 L$^{-1}$, which is used to represent the aforementioned nucleation modes in the CRM simulations.

### b. Sensitivity tests

To explore the possible impacts of microphysical processes and parameters on CRM-simulated MPS clouds, a range of sensitivity tests are performed (Table 1). The baseline simulation (hereafter referred to as CONTROL) is performed with a two-moment approach for both cloud particles and precipitating hydrometeor species using the M05 scheme. A sensitivity experiment (OneM) is performed with a one-moment approach for all hydrometeor species, (as described in section 2), to quantify the benefits of the two-moment approach. Note that graupel is allowed to occur in the OneM simulation, but it never does.

A sensitivity test (SAT), which is the same as the CONTROL except for using the water–ice saturation adjustment scheme of Lord et al. (1984), is designed to examine the role of the water–ice saturation adjustment used in the one-moment microphysics parametrization (Lord et al. 1984; Tao et al. 1989). The Lord et al.
adjustment scheme, described in section 2, is different than the M05 scheme used in the CONTROL, which determines deposition/sublimation of cloud ice (as well as snow and rain) using a nonsteady vapor diffusion approach, and applies a saturation adjustment approach only to cloud liquid water, which is reasonable because of short droplet phase relaxation time.

The rest of the microphysics experiments test several microphysical parameters used in the M05 scheme. Experiment IN20 is performed by increasing the IFN number concentration by a factor of 20 from the measured value, that is, from 0.16 to 3.2 L$^{-1}$. This experiment is motivated by previous numerical modeling studies of Arctic MPS clouds that showed large sensitivity of simulated MPS clouds to the availability of IFN (Harrington et al. 1999; Jiang et al. 2000; Morrison and Pinto 2006). Morrison and Pinto (2006) found that the prediction of $n_s$ could critically affect an MPS cloud simulated by a mesoscale numerical model. To examine this issue, experiment N0S is performed by setting the intercept parameter $N_0$, equal to a constant value of $3.0 \times 10^6$ m$^{-4}$ (Gunn and Marshall 1958; L83) so that the number concentration of snow particles, $n_s$, is diagnosed rather than predicted. The last sensitivity test, $\mu_0$, examines the spectral shape parameter $\mu$ in the gamma size distribution [Eq. (3)] of cloud ice in the two-moment approach. Experiment $\mu_0$ is performed with $\mu = 0$, instead of 5 in the CONTORL. That is, cloud ice is represented by an exponential (rather than a gamma) size distribution in the $\mu_0$ experiment.

Another set of sensitivity tests (Table 1) aims at examining the impact of estimated uncertainties in the surface fluxes, which are compared to the differences between the one-moment and two-moment schemes. These tests are the same as either the CONTROL or the OneM simulations except for simultaneously increasing or decreasing the surface sensible and latent heat fluxes by 10% or 25%, respectively. One reason for performing these tests is that the magnitudes of these fluxes were based on the ECMWF model data and, therefore, may contain model uncertainties. Another reason is that previous modeling studies indicate that surface turbulent flux could influence properties of simulated mixed-phase Arctic clouds (e.g., Harrington and Olsson 2001).

5. Comparison between CRM simulations and aircraft observations

We first examine the CONTROL and OneM simulations since they represent results using the two distinct (two versus one moment) microphysics schemes.

Fig. 4. Vertical profiles of (a), (c) liquid water content and (b), (d) total ice water content from the aircraft observations (solid lines: means and shading: plus and minus one standard deviation). Results from the (a), (b) CONTROL and (c), (d) OneM simulations. Vertical profiles of (e) ice water content and (f) snow water content from the CONTROL (lines without symbols) and OneM (lines with diamonds) simulations. Each of the simulations present results at different times: 3.25 (long-dashed line), 10.25 (short-dashed line), and 11.75 h (dot–dashed line).

a. Vertical profiles of hydrometeor mass

The vertical profiles of LWC and IWC plus snow water content (SWC) (hereafter ISWC) from the CONTROL and OneM simulations and observations (mean plus/minus standard deviation computed from the four flights) are compared to examine the vertical variations of cloud distributions. The simulated LWC and ISWC are horizontally and time averaged over 30-min blocks during the 12-h simulation period. Only those centered at 3.25, 10.25, and 11.75 h are shown in Fig. 4. The observations represent both spatial and temporal variability since many of the observations were obtained at different locations (Barrow, Oliktok Point, and in between). Following McFarquhar et al. (2007), the vertical axis of Fig. 4 is a normalized height ($H_n$), defined as $(H - H_b)/(H_t - H_b)$, where $H$ is the height, $H_b$ base
height, and \( H_t \) top height. The cloud top and cloud base are located at \( H_n = 1 \) and \( H_n = 0 \), respectively. A negative \( H_n \) represents a height below the liquid cloud base. Observations below liquid cloud base typically refer to the presence of precipitating ice and, on occasion, refer to an erroneously identified cloud base. The observations are categorized into 20 bins of \( H_n \) within the cloud layer. There are about 50 samples for each of the observed cloud properties within each \( H_n \) bin.

McFarquhar et al. (2007) analyzed the variation of the observed microphysical variables with height. To compare against the model simulations, the most notable features are summarized here. The observed, averaged LWCs increase with height within the cloud layer with a peak of \( \sim 0.32 \) g m\(^{-3}\) located near the cloud top. The standard deviations of the observed LWC range from 0.05 to 0.08 g m\(^{-3}\) below cloud top (\( H_n < 0.8 \)) and increase to \( \sim 0.14 \) g m\(^{-3}\) at the cloud top. The larger variation of the observed LWC near cloud top may be related to entrainment. It could also be due to horizontal variations in the top height with the plane flying in and out of the cloud. The observations also indicate that there is a small amount of ISWC \( (0.01 \) g m\(^{-3}\)) with a relatively constant vertical distribution within the cloud layer, but with large variations (up to 0.04 g m\(^{-3}\)) in the lower part of the cloud layer \( (H_n < 0.25) \). The large variations suggest that large ISWCs were only occasionally observed near cloud base. The observed fraction of ice to the total condensed water, however, increases toward the base of the cloud (McFarquhar et al. 2007).

The liquid and ice coexist throughout the entire period of the two simulations (Figs. 4a–d), consistent with observations that showed mixed-phase clouds occurred 71% of the time. The cloud top and base in the model are located at 1.33 and 0.65 km, respectively. In both simulations, ice crystals (including snow) occur throughout the cloud layer and fall below liquid cloud base to the surface, consistent with radar and lidar measurements shown in Verlinde et al. (2007). However, there are some obvious differences between the two simulations. In the CONTROL, both the LWC and the ISWC reach a steady state after \( \sim 3 \) h. The LWCs increase with height and the ISWCs are constant with height within the cloud layer. Both LWC and ISWC are located within the uncertainty range of the observations. In the OneM experiment, the liquid cloud layer decays with time and the ice mass increases with time. The amount of LWC is underestimated compared to the observations. The ISWCs from the OneM experiment exhibit larger variations with height as well as larger amounts at most heights within the cloud layer than those in the observations or the CONTROL results.

To further explore the differences in ice crystal mass between the CONTROL and OneM simulations, separate vertical profiles of IWC and SWC from the two simulations are compared (Figs. 4e,f). The IWCs from the CONTROL are nearly constant with height within the cloud layer. The IWCs from the OneM run exhibit two peaks: one located near the cloud top and the other at the lower part of the cloud layer during the majority of the 12-h simulation period. The only exception occurs at the last hour when there is a single peak at \( H_n \) of \( \sim 0.8 \). These differences are related to the cloud ice deposition process in the CONTROL and OneM simulations, as shown in the time–height distributions of cloud ice deposition rate in Figs. 5a,b. In the CONTROL, deposition (from water vapor to cloud ice) occurs smoothly in height and in time within the cloud layer at instantaneous rates less than 0.01 g kg\(^{-1}\) h\(^{-1}\). In the OneM experiment, deposition (positive values) or sublimation (negative values) rates exhibit significant variability within the cloud layer and are one order of magnitude larger than those seen in the CONTROL. The OneM simulation also produces deposition rates (in the lower part of the cloud layer) that oscillate with a period of about 30 min during the first 8 h of the simulation. After cloud water decreases significantly in the OneM simulation (i.e., after 10 h of the simulation), the cloud ice deposition process is enhanced significantly within the cloud layer because the saturation vapor mixing ratio converges to that with respect to ice when liquid water is small.

To examine the collective effects of cloud microphysical processes on dynamics, the domain-averaged subgrid turbulent kinetic energy is compared between the CONTROL and OneM simulations (Figs. 5c,d). While the TKE produced by the CONTROL is relatively constant below the cloud top except for near the surface and changes little with time, the TKE produced in the OneM simulation shows more significant variability with time and height. The OneM TKE in the interior of the cloud layer oscillates with the same period as its deposition process, although other microphysical processes also contribute to the buoyancy production term in the TKE equation. The OneM TKE near the cloud base has smaller values than those in the interior of the cloud layer and below the cloud base. These results suggest that different representations of cloud microphysical processes have distinct impacts on simulated turbulent-scale dynamics.

Snow exists from the cloud top to the surface in both simulations with maxima located near the cloud base.
However, the OneM experiment produced SWCs several times as large as those from the CONTROL. The larger SWCs in the OneM experiment are partly attributed to the greater deposition of cloud ice (Fig. 5b) that is subsequently converted to snow through the autoconversion process. As will be shown in section 6, this result can also be attributed to the intercept parameter of the snow size distribution ($N_0$) specified in the one-moment scheme, which is larger than that predicted in the CONTROL.

**b. Vertical profiles of hydrometeor number concentration and effective radius**

Number concentrations and effective radii of cloud liquid droplets and ice crystals are important cloud properties that significantly influence cloud optical properties and various microphysical processes. These variables are not predicted in the OneM experiment. Therefore, we compare those from the CONTROL simulation to the observations (Fig. 6). Averages of the observed droplet number concentrations ($n_d$) are relatively constant with height in the cloud layer with values of about 50 cm$^{-3}$ (Fig. 6a). The variation of the observed $n_d$ at each height bin ranges from 20 to 35 cm$^{-3}$. The simulated $n_d$ is constant with height and has a value of $\sim 60$ cm$^{-3}$, generally consistent with and within the variability of the observations. The observations suggest that the effective radii of cloud droplets ($r_e$) generally increase with height within the cloud layer (Fig. 6b). The standard deviation of $r_e$ ranges between 1 and 2 $\mu$m for most height bins except near the cloud top where it increased to $\sim 3$ $\mu$m. The simulation reproduced the observed increase of $r_e$ with height within the cloud layer. At most height bins, the simulated $r_e$ is within the uncertainty range of the observations except near the cloud base where the observations are greater than the simulated $r_e$. The underestimation of $r_e$ near cloud base is probably related to the smaller simulated LWC at that height, compared to the observations (Fig. 4a). On the other hand, on some of the ramped ascents and descents there may have been some uncertainties in the identification of cloud base from the measurements.

The vertical profile of total ice crystal number concentration ($n_i$) from the observations (Fig. 6c) shows a relatively constant distribution with height with a mean of 1–3 L$^{-1}$, significantly greater than the observed IFN number concentration. The standard deviation of $n_i$ is comparable to or greater than the mean value. The simulated $n_i$ (including both cloud ice and snow) is less than 0.5 L$^{-1}$, smaller than the observed mean. Note
that the observed \( n_i \) refers to concentration of ice particles with diameter greater than 53 \( \mu m \). Fridlind et al. (2007) estimated the possible error in the observed \( n_i \), as being at most about half of an order of magnitude. The discrepancy between the observed and simulated \( n_i \) would be even larger if ice particles with diameters smaller than 50 \( \mu m \) were excluded from the simulated results.

The observations show that the vertical profile of effective radius of ice crystals (\( r_{ei} \), defined following Fu (1996), is constant with height and the mean values of \( r_{ei} \) are \( \sim 25 \mu m \). In the CONTROL simulation, ice effective radius is calculated by mass weighting the inverse values for cloud ice and snow, which also follow the definition of Fu (1996) (see section 2). The vertical profile of simulated \( r_{ei} \) is constant with height, consistent with the observations. However, the \( r_{ei} \) are greater than those observed (60 versus 25 \( \mu m \)). This is partly due to simulated values of \( n_i \) that are smaller than those observed (Fig. 6c).

Underestimation of \( n_i \) was noticed in almost all models that participated in the ARM MPACE model intercomparison (e.g., Fridlind et al. 2007). Reasons for this are not clear yet. The major ice forming mechanism in the CONTROL is contact freezing of droplets. The 12-h-averaged contact-freezing rate increases with height within the cloud layer from almost zero to a value of \( 7 \times 10^{-6} \) g kg\(^{-1}\) h\(^{-1}\) at the cloud top. The formation of ice by deposition, condensation freezing, and immersion freezing occurs near the cloud top with an averaged rate of \( 2 \times 10^{-6} \) g kg\(^{-1}\) h\(^{-1}\). The best quantified mechanism for ice enhancement is probably the shedding of ice splinters during riming, that is, the Hallett and Mossop (1974, hereafter HM) mechanism. However, ice splinter production through the HM mechanism is not significant in the simulation because the cloud temperature ranges from \(-15^\circ \) (cloud top) to \(-10^\circ C \) (cloud base), colder than the temperature necessary for the HM mechanism to operate (\(-3^\circ \) to \(-8^\circ C \)). It is likely that other mechanisms for high ice particle concentration may be missing in the two-moment microphysics scheme. For example, Rangno and Hobbs (2001) argued that the fragmentation of delicate crystals (such as dendrites and aggregates) during crystal–crystal and crystal–droplet collisions, and the shattering of some drops during freezing in free fall, may play a role in the production of relatively high ice particle concentrations in the Arctic clouds. Fridlind et al. (2007) claimed that two other mechanisms—the formation of ice nuclei from drop evaporation residuals and drop freezing during evaporation—could be strong enough to account for the MPACE observations.

6. Results from sensitivity experiments

We have shown that the CONTROL simulation reproduced most of the aircraft-observed cloud properties except for its underestimation of ice crystal number concentration and overestimation of ice crystal effective size. These two quantities have the largest uncertainties in the observations (McFarquhar et al. 2007). The OneM experiment underestimated the observed LWC and produced a bimodal vertical structure of IWC that was not observed. Its simulated mixed-phase stratus glaciated earlier than in the CONTROL simulation. Four additional sensitivity experiments (Table 1), as described in section 4b, are presented in this section to further explore the impact of microphysical processes and parameters on the simulated MPS clouds. Another set of sensitivity tests is used to explore the impact of uncertainties in the surface fluxes, which are compared to the differences between the CONTROL and OneM simulations.
a. Vertically integrated hydrometeor amount

Figure 7 shows the time variability of the vertically integrated amount of each hydrometeor species, that is, LWP, rainwater path (RWP), ice water path, and snow water path (SWP) for the CONTROL, OneM, μi0, N0s, SAT, and IN20 simulations. The model results are averaged over the entire horizontal domain in space and 30 min in time. Time-averaged values and standard deviations of LWP, IWP, SWP, and RWP between 4 and 12 h from all simulations are given in Table 2. Comparison among these simulations reveals the following major findings: First, a persistent MPS cloud layer is produced by the CONTROL, μi0, and N0s simulations, which reaches a steady state after 3 h, although their steady-state LWP values differ (176.5 ± 2.8, 188.6 ± 4.4, 142.6 ± 6.9 g m⁻², respectively; see Table 2). Second, both the OneM and SAT experiments produce smaller LWPs (54.2 ± 6.4 and 97.9 ± 12.1 g m⁻², respectively; see Table 2) than the other simulations shown in Fig. 7a. The temporal evolutions of IWP and SWP are similar between OneM and SAT, for example, a large increase near the end of the simulations (Figs. 7b,c). This behavior differs markedly from the other simulations. Third, the IN20 experiment produces the smallest time-averaged LWP (8.0 ± 12.0 g m⁻²) but the largest IWP (20.9 ± 13.2 g m⁻²) and SWP (43.4 ± 5.6 g m⁻²) among the simulations (Table 2).

<table>
<thead>
<tr>
<th>Simulation</th>
<th>LWP</th>
<th>IWP</th>
<th>SWP</th>
<th>RWP</th>
</tr>
</thead>
<tbody>
<tr>
<td>CONTROL</td>
<td>176.5</td>
<td>4.4</td>
<td>5.2</td>
<td>8.5</td>
</tr>
<tr>
<td>OneM</td>
<td>54.2</td>
<td>4.7</td>
<td>28.7</td>
<td>0.0</td>
</tr>
<tr>
<td>SAT</td>
<td>97.9</td>
<td>4.6</td>
<td>16.0</td>
<td>0.3</td>
</tr>
<tr>
<td>IN20</td>
<td>8.0</td>
<td>20.9</td>
<td>43.4</td>
<td>0.6</td>
</tr>
<tr>
<td>μi0</td>
<td>188.6</td>
<td>1.5</td>
<td>1.5</td>
<td>11.0</td>
</tr>
<tr>
<td>N0s</td>
<td>142.6</td>
<td>2.8</td>
<td>25.8</td>
<td>0.7</td>
</tr>
<tr>
<td>CTR +10%</td>
<td>176.0</td>
<td>3.9</td>
<td>8.9</td>
<td>0.6</td>
</tr>
<tr>
<td>CTR −10%</td>
<td>168.5</td>
<td>4.5</td>
<td>4.5</td>
<td>0.5</td>
</tr>
<tr>
<td>CTR +25%</td>
<td>180.1</td>
<td>3.8</td>
<td>10.4</td>
<td>2.3</td>
</tr>
<tr>
<td>CTR −25%</td>
<td>153.3</td>
<td>4.8</td>
<td>4.0</td>
<td>0.4</td>
</tr>
<tr>
<td>1M +10%</td>
<td>58.6</td>
<td>5.5</td>
<td>32.8</td>
<td>4.7</td>
</tr>
<tr>
<td>1M −10%</td>
<td>54.2</td>
<td>3.2</td>
<td>23.6</td>
<td>1.2</td>
</tr>
<tr>
<td>1M +25%</td>
<td>46.4</td>
<td>7.7</td>
<td>39.2</td>
<td>4.0</td>
</tr>
<tr>
<td>1M −25%</td>
<td>37.9</td>
<td>3.4</td>
<td>19.7</td>
<td>1.2</td>
</tr>
</tbody>
</table>

FIG. 7. Time series of (a) LWP, (b) IWP, (c) SWP, and (d) RWP produced by CRM simulations: CONTROL (solid), N0S (short-dashed line), SAT (long-dashed line), OneM (dotted-dashed line), and μi0 (dots-dashed line). (e) Time series of LWP (solid line), IWP (long-dashed line), SWP (short-dashed line), IWP plus SWP (dotted-dashed line), and RWP (dots-dashed line) produced by the IN20 experiment.
with LWP decreasing monotonically with time until a complete dissipation of liquid clouds at 7 h (Fig. 7e). Detailed discussions of these findings are given below.

1) Effects of spectral shape parameter of cloud ice

The LWP derived from the MWR measurements averaged over the 12-h simulation period is 210 g m\(^{-2}\) at Barrow, which has an uncertainty of 20–30 g m\(^{-2}\) (e.g., Wang 2007). The steady-state LWP from the CONTROL, 176.5 g m\(^{-2}\), is 84% of the retrievals. The \(\mu_0\) experiment generated more LWP (188.6 versus 176.5 g m\(^{-2}\)) and less IWP (1.5 versus 4.4 g m\(^{-2}\)) and SWP (1.5 versus 5.2 kg m\(^{-2}\)) compared to the CONTROL simulation. Decreasing the spectral shape parameter (\(\mu_i\)) from 5 in the CONTROL simulation to 0 in the \(\mu_0\) experiment increases the phase relaxation time associated with cloud ice, that is, making cloud ice deposition occur more slowly. The slower deposition to cloud ice at the expense of cloud water causes less IWP and SWP to occur more slowly. The slower deposition to cloud ice associated with cloud ice, that is, making cloud ice deposition rate less, thus results in a smaller LWP (142.6 versus 176.5 g m\(^{-2}\)) averaged over the 12-h simulation period is 210 g m\(^{-2}\). The joint PDF (%) of \(N_{sw}\) and height predicted by the CONTROL simulation. Contours, from light to dark, represent 0.1%, 0.5%, 1.0%, and 2.0%.

Fig. 8. (a) Frequency distribution of \(N_{sw}\) predicted by the CONTROL simulation. (b) Joint PDF (%) of \(N_{sw}\) and height predicted by the CONTROL simulation. Contours, from light to dark, represent 0.1%, 0.5%, 1.0%, and 2.0%.

The L83 value of \(N_{sw}\) used in the N0S experiment was obtained from a midlatitude frontal cloud system. Therefore, it is not surprising that this \(N_{sw}\) is different from that predicted in Arctic mixed-phase clouds. Snapshots of the predicted \(N_{sw}\) (not shown) exhibit horizontally inhomogeneous distributions that vary with time. The significantly larger value used in the N0S experiment resulted in stronger depositional growth of snow (i.e., an enhanced Bergeron–Findeisen mechanism) and more significant accretion of cloud droplets by snow, both contributing to smaller (larger) LWP (SWP). The smaller IWP with larger \(N_{sw}\) is due to more water vapor deposited to snow and therefore less to cloud ice.

Compared to the OneM experiment, the N0S experiment produced about the same SWP (\(~26\) g m\(^{-2}\)) between 3 and 10 h, but less SWP after 10 h and higher LWP during most of the simulation period. The agreement in SWP between the OneM and N0S experiments for the 3–10-h period resulted from similar snow deposition rates and accretion rates of droplets by snow. The larger SWP in the OneM experiment after 10 h (up to \(37\) g m\(^{-2}\)) is related to its larger cloud ice deposition rates caused by the application of the water–ice saturation adjustment scheme of Lord et al. (1984; Fig. 5b), and subsequently more cloud ice converted to snow. The smaller LWP of the OneM experiment versus the N0S experiment (54.2 versus 142.6 g m\(^{-2}\)) is probably also a result from the water–ice saturation adjustment scheme utilized in the OneM experiment, as will be further explained below.

2) Effects of predicting snow number concentration

Compared to the CONTROL, the N0S experiment produced less LWP (142.6 versus 176.5 g m\(^{-2}\)) and IWP (2.8 versus 4.4 g m\(^{-2}\)) but more SWP (25.8 versus 5.2 g m\(^{-2}\)). The joint PDF (%) of \(N_{sw}\) and height from the CONTROL (Fig. 8b) shows that the \(N_{sw}\) varies with height and is mostly one order of magnitude smaller than the specified constant value of \(~3\times10^6\) m\(^{-4}\) in the N0S experiment (Fig. 8a). The L83 value of \(N_{sw}\) used in the N0S experiment was obtained from a midlatitude frontal cloud system. Therefore, it is not surprising that this \(N_{sw}\) is different from that predicted in Arctic mixed-phase clouds. Snapshots of the predicted \(N_{sw}\) (not shown) exhibit horizontally inhomogeneous distributions that vary with time. The significantly larger value used in the N0S experiment resulted in stronger depositional growth of snow (i.e., an enhanced Bergeron–Findeisen mechanism) and more significant accretion of cloud droplets by snow, both contributing to smaller (larger) LWP (SWP). The smaller IWP with larger \(N_{sw}\) is due to more water vapor deposited to snow and therefore less to cloud ice.

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3) Effects of water–ice saturation adjustment

There are some similar results between the SAT and OneM simulations, that is, an earlier decay of LWP and an increase of IWP and SWP after 9 h, compared to the other simulations shown in Figs. 7a–c. The relatively steady LWP between 3 and 8 h has a magnitude of \(~100\) g m\(^{-2}\) in the SAT experiment compared to \(~177\) g m\(^{-2}\) for the CONTROL and \(~65\) g m\(^{-2}\) for the OneM simulation. Other aspects of hydrometeor paths simulated by SAT, such as the significant variability in LWP and SWP, are more similar to those from the OneM than the CONTROL (Table 2). These results suggest that the application of the Lord et al. saturation adjustment scheme to M05 to determine the condensation of cloud water and deposition of cloud ice could significantly change the simulated MPS clouds and could result in an underestimation of LWP. Why did the SAT experiment produce less LWP than the CONTROL? It is partly because condensation rates near the cloud top are smaller in the SAT simulation than in CONTROL (Fig. 9). It is also due to more accretion by snow resulting...
from a greater amount of snow in the SAT experiment. Why did the SAT experiment produce less IWP and more SWP than the CONTROL? Deposition rates in the SAT experiment (Fig. 9) are greater than those in CONTROL, especially near the cloud base and cloud top, suggesting that more cloud ice is produced from the deposition of water vapor and subsequently more efficient conversion from cloud ice to snow.

4) Effects of IFN number concentration

The IN20 experiment produces greater ice crystal number concentration (∼3 L⁻¹) than in the CONTROL and much closer to the aircraft observations (1–3 L⁻¹, Fig. 6c). This is attained by increasing the IFN number concentration to 20 times of the measured value in the IN20 experiment. However, the increase in IFN number concentration (and, hence, increase in the ice crystal number concentration) transfers the solid, largely liquid, stratus deck into a broken cloud system, consistent with previous modeling studies (Harrington et al. 1999; Jiang et al. 2000). As shown in Fig. 7e, the initially thick liquid cloud layer (LWP of 150 g m⁻²) decays monotonically with time in the IN20 experiment and is completely dissipated after 7 h. This is not realistic since a persistent cloud layer was observed. The decay of the simulated liquid cloud layer results from the significantly enhanced consumption of cloud water through the Bergeron processes when the ice crystal number concentration is increased in the simulation.

b. Precipitation and radiative flux at the surface

Arctic clouds are linked to the hydrological cycle and oceanic processes through precipitation that affects freshwater input into the Arctic Ocean. Surface precipitation rate is therefore an important parameter to be reproduced by models. Frequent, light snow events were reported in the ARM ground measurements during the 9–14 October period. However, as mentioned in section 3, the ARM-observed surface precipitation rate could be overestimated because of blowing snow (Xie et al. 2006). The temporally averaged surface precipitation rate from the ARM observation at Barrow was 1.7 mm day⁻¹ for the simulation period. Accumulated (horizontally averaged) surface precipitation rates in the CRM simulations are shown in Table 3. The column denoted as “snow” represents the liquid water equivalent of the snow rate. The 12-h averaged surface precipitation ranges from ∼0.3 to 1.2 mm day⁻¹ in the simulations, less than the ARM observations. The smallest 12-h averaged surface precipitation occurs in the μ0 and N0S experiments. However, the surface precipitation is mainly rain in the μ0 experiment and snow in the N0S experiment. The SAT and OneM experiments produce more surface precipitation in the form of snow (∼0.8 mm day⁻¹) than the other simulations (<0.3 mm day⁻¹), except for IN20 (1.1 mm day⁻¹), consistent with their more significant ice depositional rates. Note that rain does not occur in the OneM experiment because the threshold for activating the autoconversion parameterization from cloud water to rain is 0.5 g kg⁻¹, which is never reached. The IN20 experiment produces the largest surface precipitation among the simulations, owing to the strong depositional growth of cloud ice and snow.

Differences in the representation of microphysical processes affect surface radiation budgets through their influences on the simulated cloud microphysical and optical properties. This is illustrated by the temporal variations of the half-hourly and horizontally averaged downwelling infrared (IR) and shortwave (SW) radiative fluxes (Fig. 10). A striking feature appearing in Fig. 10 is a dramatic reduction of the downwelling IR flux (∼10 W m⁻² h⁻¹) when the liquid water begins to decrease, that is, after 6 h in the IN20 simulation, 10 h in the OneM simulation, and 11 h in the SAT experiment.

Table 3. Surface precipitation rates (mm day⁻¹) averaged over the entire 12-h and 3- to 12-h simulation periods.

<table>
<thead>
<tr>
<th>Simulation</th>
<th>0–12 h</th>
<th>3–12 h</th>
<th>0–12 h</th>
<th>3–12 h</th>
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</thead>
<tbody>
<tr>
<td>CONTROL</td>
<td>0.23</td>
<td>0.18</td>
<td>0.17</td>
<td>0.20</td>
<td>0.39</td>
<td>0.38</td>
</tr>
<tr>
<td>μ0</td>
<td>0.28</td>
<td>0.24</td>
<td>0.02</td>
<td>0.02</td>
<td>0.30</td>
<td>0.27</td>
</tr>
<tr>
<td>N0S</td>
<td>0.03</td>
<td>0.00</td>
<td>0.27</td>
<td>0.30</td>
<td>0.30</td>
<td>0.30</td>
</tr>
<tr>
<td>SAT</td>
<td>0.06</td>
<td>0.00</td>
<td>0.83</td>
<td>0.76</td>
<td>0.89</td>
<td>0.76</td>
</tr>
<tr>
<td>OneM</td>
<td>0.00</td>
<td>0.00</td>
<td>0.76</td>
<td>0.71</td>
<td>0.76</td>
<td>0.71</td>
</tr>
<tr>
<td>IN20</td>
<td>0.05</td>
<td>0.00</td>
<td>1.13</td>
<td>1.28</td>
<td>1.19</td>
<td>1.28</td>
</tr>
</tbody>
</table>
The downwelling IR radiative fluxes in the CONTROL, μ0, and N0S simulations differ from one another by only ~2 W m\(^{-2}\) because these simulations all produced clouds that emitted as near blackbodies. The downward SW flux from the CONTROL has a maximum of ~41 W m\(^{-2}\) at 5–6 h, while that from the OneM and SAT simulations differs from the CONTROL by up to 25 and 13 W m\(^{-2}\), respectively, due to substantial differences in the amounts of liquid droplets from the CONTROL.

### c. Sensitivity to surface heat flux

Figure 11 shows the temporal variations of LWP, IWP, SWP, and RWP when surface turbulent heat fluxes are changed by either ±10% or ±25%, respectively, with either the M05 or the L83 microphysics scheme. Comparisons are also made against those of the CONTROL and OneM simulations. The 4–12-h averages and standard deviations of the vertically integrated water amounts are given in Table 2. Compared to the experiments with the M05 scheme, the effects of changing the surface fluxes are more complicated with the L83 scheme with larger variability, which may be related to the oscillation in the OneM simulated TKE field (Fig. 5d). Most importantly, it is obvious that the differences caused by the two distinct microphysics schemes are more significant than those due to changes in the surface heat fluxes by either 10% or 25%. For example, the time-averaged LWPs from the tests utilizing the M05 scheme range between 153.3 ± 3.3 and 180.1 ± 3.2 g m\(^{-2}\), while those with the L83 scheme range between 37.9 ± 9.6 and 58.6 ± 17.3 g m\(^{-2}\) (Table 2).

### 7. Summary and conclusions

There have been few CRM simulations of boundary layer MPS clouds, even though these clouds occur frequently in the Arctic and may potentially impact global climate and regional weather. In this study, a CRM has been used to simulate single-layer MPS clouds observed at the NSA sites during the ARM MPACE. This CRM was implemented with both a commonly used one-moment microphysics scheme (L83) and an advanced two-moment microphysics scheme (M05) as well as a state-of-the-art radiative transfer scheme (Fu and Liou 1993). A set of simulations with different treatments of microphysical processes and different specifications of microphysical parameters was performed to examine the sensitivity of the CRM-simulated MPS clouds to cloud microphysics parameterizations. Modeled cloud fields have been compared to the vertical profiles of the bulk microphysical properties derived from aircraft measurements (McFarquhar et al. 2007) and retrievals of LWP obtained from ground-based observations (Turner et al. 2007), as well as surface precipitation measurements.

The aircraft observations suggest that the LWCs and
droplet effective radii increased with height in the cloud layer while the droplet number concentrations, and the masses, number concentrations, and effective radii of ice crystals were relatively constant with height. Using the newly implemented two-moment scheme and the observed aerosol size distribution and IFN number concentration, the CONTROL simulation was able to reproduce the magnitudes and vertical structures of cloud liquid water, cloud ice water, droplet number concentration, and droplet effective radius as revealed by the aircraft observations, although it underestimated the number concentration of the ice crystals by an order of magnitude and overestimated the effective radii of the ice crystals by a factor of 2–3. With the one-moment (OneM) microphysics scheme, the CRM produced values of LWP and ISWP that were about 30% and 4 times, respectively, those produced by the CONTROL. In addition, the vertical profile of IWC exhibited a bimodal distribution in contrast to the constant distribution of IWC produced with the two-moment approach.
The deficiencies in the OneM-simulated cloud fields are largely associated with the ice–water saturation adjustment of Lord et al. (1984), which overestimated ice depositional rate near the top and base of the MPS cloud layer and caused some strange nonphysical feedback between the deposition and the turbulence. The deficiencies are also closely related to the constant large value of the snow spectra intercept parameter ($n_{0s}$) used in L83. Moreover, simultaneously changing the surface latent and sensible heat fluxes by either 10% or 25%, respectively, which is an estimate of the possible uncertainties associated with these fluxes, caused smaller differences in the simulated cloud fields than those caused by application of the two- and one-moment microphysics schemes.

When the observed IFN number concentration was used, this CRM and many other models that participated in a model intercomparison project (Klein et al. 2006; Fridlind et al. 2007) could not reproduce the observed ice concentrations, which greatly exceeded those of ice nuclei (a few per liter versus 0.16 L$^{-1}$). On the other hand, the MPS clouds glaciated in the model when the ice concentration was initially close to the observed value, which was obtained by increasing the observed IFN number concentration by a factor of 20 (from 0.16 to 3.2 L$^{-1}$). The rapid glaciation of cloud liquid water through the enhanced Bergeron–Findeisen process at the higher ice number concentration in the CRM is consistent with previous modeling studies (Harrington et al. 1999; Jiang et al. 2000).

Results from a sensitivity test that used the Lord et al. water–ice saturation adjustment in the CONTROL simulation were more similar to those from the OneM simulation than the CONTROL. This indicates that the CRM-simulated MPS clouds are very sensitive to the representations of cloud water condensation and cloud ice deposition. One of the major assumptions on which the saturation adjustment of Lord et al. (1984) is based is that the saturation vapor mixing ratio ($q^*$) is a mass-weighted average of the respective saturation values over liquid and ice when both cloud water and cloud ice are present. Although this assumption was supported by the aircraft data collected during the SHEBA/FIRE-ACE campaign (Fu and Hollars 2004), its utilization in models is problematic because $q^*$ depends on the model-simulated $q_l$ and $q_c$, and the accuracy of $q_l$ and $q_c$ prediction is influenced by other aspects of the model. Use of this assumption resulted in a significant underestimation of LWP during the fall season in an SCM simulation using SHEBA data (Yuan et al. 2006), qualitatively consistent with our findings.

It is found that the two-moment scheme predicted smaller values of $N_{0r}$ (mostly $<0.5 \times 10^6$ m$^{-3}$) than the constant value ($3 \times 10^6$ m$^{-3}$) used in the one-moment scheme of L83. Using the larger constant value of $N_{0r}$ in the CONTROL resulted in 20% less LWP and 5 times more SWP than the CONTROL. Furthermore, representing the cloud ice spectra with an exponential size distribution rather than the gamma distribution resulted in smaller IWP and SWP (2.9 and 3.7 g m$^{-2}$) and larger LWP (2.1 g m$^{-2}$), owing to a slower ice deposition process.

Note that the modeled results may be sensitive to some other parameters (such as the assumed bulk densities and fall speeds of cloud ice and snow) involved in the two-moment microphysics scheme. The mechanisms for the formation of ice concentrations that greatly exceed those of ice nuclei in the MPS Arctic clouds and their representation in the models should be studied further. It should be noted that this is a single case study of a large-scale stratocumulus deck caused by off-ice flow in autumn with cloud-layer temperature from $-10^\circ$ to $15^\circ$C. Additional observations and simulations are needed to further narrow down the uncertainties associated with these microphysical parameters.

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