Multi-layer arctic mixed-phase clouds simulated by a cloud-resolving model: Comparison with ARM observations and sensitivity experiments

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[1] A cloud-resolving model (CRM) is used to simulate the multiple-layer mixed-phase stratiform (MPS) clouds that occurred during a three-and-a-half day subperiod of the Department of Energy-Atmospheric Radiation Measurement Program’s Mixed-Phase Arctic Cloud Experiment (M-PACE) and to examine physical processes responsible for multilayer production and evolution. The CRM with a two-moment cloud microphysics is initialized with concurrent meteorological, aerosol, and ice nucleus measurements and is driven by time-varying large-scale advective tendencies of temperature and moisture and surface sensible and latent heat fluxes. The CRM reproduces the dominant occurrences of the single- and double-layer MPS clouds as revealed by the M-PACE observations although the simulated first cloud layer is lower and the second cloud layer is thicker compared to observations. The aircraft measurements suggest that the CRM qualitatively captures the major characteristics in the vertical distribution and interperiod variation of liquid water content (LWC), droplet number concentration, total ice water content (IWC), and ice crystal number concentration (n_i). However, the magnitude of LWC is overestimated and those of IWC and n_i are underestimated. In particular, the simulated n_i is one order of magnitude smaller than the observed. Sensitivity experiments suggest that both the surface fluxes and large-scale advection control the formation of the lower cloud layer while the large-scale advection initiates the formation of the upper cloud layer but the maintenance of multilayer structures relies on the longwave (LW) radiative effect. The LW cooling near cloud top produces a more saturated environment and a stronger dynamical circulation while cloud base radiative warming of the upper layer creates the stability gap between the two cloud layers. Both cloud layers are sensitive to ice-forming nuclei number concentration since ice-phase microphysics provides a strong sink of cloud liquid water mass.


1. Introduction

[2] Arctic clouds have been identified as playing a central role in the Arctic climate system that has been changed significantly in the recent decades [ACIA, 2005] and can potentially impact global climate [Curry et al., 1996; Vavrus, 2004]. A few field campaigns have been conducted to improve the understanding of cloud-radiative interactions in the Arctic: the Beaufort Arctic Sea Experiment (BASE; [Curry et al., 1997]), the First International Satellite Cloud Climatology Project (ISCCP) Regional Experiment (FIRE) - Arctic Cloud Experiment (ACE; [Curry et al., 2000]), the Surface Heat Budget of the Arctic (SHEBA; [Uttal et al., 2002]), and the US Department of Energy Atmospheric Radiation Measurement (ARM; [Stokes and Schwartz, 1994; Ackerman and Stokes, 2003]) program’s Mixed-Phase Arctic Cloud Experiment (M-PACE [Harrington and Verlinde, 2004; Verlinde et al., 2007]). These field campaigns identified that mixed-phase stratiform (MPS) clouds were prevalent in Arctic transition seasons [Intrieri et al., 2002; Verlinde et al., 2007], especially during the fall over Barrow at the ARM North Slope of Alaska (NSA) site [Wang et al., 2005; Shupe et al., 2005]. This type of mixed-phase cloud is
a water-dominated cloud layer with precipitating ice, yet they persist for long periods of time [e.g., Hobbs and Rango, 1998; McFarquhar et al., 2007].

[3] Previous observational analysis and modeling studies revealed that large-scale advection, surface flux, microphysics, and radiation, as well as their interactions with clouds, could affect the formation and evolution of mixed-phase Arctic clouds. Observations from 12 research flights during BASE suggested local interactions between the clouds and the underlying surface [Curry et al., 1997]. The analysis of Curry et al. also suggested that large-scale advection and leads (areas of open water between ice floes) appear to play a role in forming and maintaining the cloud systems. Utilizing aircraft measurements from the BASE experiment and the National Center of Environmental Prediction (NCEP) reanalysis, Pinto [1998] suggested the importance of large-scale moisture and temperature advection, and cloud top radiative cooling for the evolution of these clouds. Pinto also speculated on the importance of ice forming nuclei (IFN) to cloud stability. In the work of Harrington et al. [1999], the soundings from a summer case were consistently cooled in cloud-resolving model (CRM) simulations to produce physically plausible mixed-phase situations, because of lack of observational soundings for mixed-phase Arctic low clouds at that time. The temperature, ice concentration, and the habit of the ice crystals were found to affect the stability of the simulated mixed-phase cloud layer. In particular, cloud layer stability was shown to be strongly dependent upon the concentration of IFN. It was also shown that ice production and sedimentation could assist the formation of a second, lower cloud layer, resulting in a multilayer cloud. Harrington and Olsson [2001] illustrated that IFN concentration could significantly impact evolution of the simulated mixed-phase clouds that occurred in an environment with a strong surface heat flux. A few recent modeling studies have examined ice formation mechanisms [e.g., Jiang et al., 2000; Morrison and Pinto, 2006; Prenni et al., 2007; Fridlind et al., 2007; Luo et al., 2008; hereafter Luo08]. These studies indicated that models were unable to produce the observed magnitudes of ice water content and ice crystal number concentration with well-known heterogeneous nucleation mechanisms and the H-M mechanism [Hallett and Mossop, 1974], especially with temperatures warmer than about −15°C [e.g., Hobbs, 1969; Beard, 1992].

[4] The M-PACE field campaign took place over the NSA during the period of 27 September to 22 October 2004 [Harrington and Verlinde, 2004; Verlinde et al., 2007]. During the field campaign, Arctic clouds were measured in detail using a wide range of instruments such as the ARM millimeter wavelength cloud radar (MMCR), micropulse lidar (MPL), laser ceilometers, and two instrumented aircraft [Verlinde et al., 2007]. The CRM/SCM (Single-Column Model) forcing data were derived from a sounding network in the Arctic region for an Intensive Operational Period in October 2004 [Xie et al., 2006]. The constrained variational analysis approach developed by Zhang and Lin [1997] and Zhang et al. [2001] was applied to this analysis. The M-PACE observations [e.g., McFarquhar et al., 2007] and the large-scale forcing data [e.g., Xie et al., 2006; Klein et al., 2006] have been used to both initialize and evaluate the results of large-eddy simulation (LES), CRM and SCM simulations. An intercomparison project between LES, CRM, and SCM models and observations have focused on both the single-layer MPS clouds (S. Klein et al., Intercomparison of model simulations of mixed-phase clouds observed during the ARM Mixed-Phase Arctic Cloud Experiment. Part I: Single-layered cloud, submitted to Quarterly Journal of the Royal Meteorological Society, 2008) and the more complicated multiple-layer MPS clouds (H. Morrison et al., Intercomparison of model simulations of mixed-phase clouds observed during the ARM Mixed-Phase Arctic Cloud Experiment. Part II: Multi-layered cloud, submitted to Quarterly Journal of the Royal Meteorological Society, 2008) using the M-PACE data. Individual studies [e.g., Fridlind et al., 2007; Luo08; Morrison et al., 2008] have used this data set to perform sensitivity simulations to understand the mechanisms of single-layer cloud formation and longevity.

[5] Despite the rapid progress in the understanding of single-layer Arctic clouds through modeling studies [e.g., Jiang et al., 2000; Morrison and Pinto, 2006; Fridlind et al., 2007; Luo08], multilayer Arctic clouds have been seldom modeled except for a few idealized [e.g., Herman and Goody, 1976; Harrington et al., 1999] or real-case studies [e.g., Smith and Kao, 1996]. Herman and Goody [1976] found that a multilayer (liquid) stratus cloud in the summertime Arctic could be viewed as a single stratus cloud layer that had suffered incomplete dissipation by solar radiation. They further demonstrated that the near-steady conditions in the radiation field in the summertime Arctic were necessary for the development of clear interstices within a stratus layer. Harrington et al. [1999] showed that a lower MPS cloud layer might be formed by cooling and moistening through ice precipitation from an upper MPS cloud layer. The upper and lower cloud layers were found to maintain due to radiative cooling. Smith and Kao [1996] examined the evolution of Arctic (liquid) status clouds observed during the Arctic Stratus Experiment conducted in June 1980 in the Beaufort Sea region. They found that the upper cloud layer was maintained by LW radiative flux divergence; the lower cloud layer was formed by the advection of warm moist air over a cool sea surface and dissipated due to the transport of moisture in the surface layer downward toward the sea surface.

[6] In the present modeling study, the same CRM as used in Luo08 is used to simulate a three-and-a-half-day subperiod of M-PACE, during which multilayer MPS clouds were observed at the NSA sites. In addition to the contrast between single-layer MPS clouds and multilayer MPS clouds, there are other differences in configurations of the simulations between Luo08 and this study. Most importantly, the large-scale forcing data were constant during the 12 h simulation period by Luo08 but vary with time during the three-and-a-half-day simulation period here. Second, an ocean surface was assumed by Luo08 as the clouds were caused by off-ice flow over the open ocean that was adjacent to the northern coast of Alaska. A land surface is considered here. Accordingly, the surface latent and sensible heat fluxes used by Luo08 were relatively larger (136.5 W m$^{-2}$ and 107.7 W m$^{-2}$, respectively) than those used in this study (18 ± 5 W m$^{-2}$ and 3 ± 5 W m$^{-2}$). The single-layer MPS clouds by Luo08 were maintained by the relatively
large surface turbulent fluxes. The formation and maintenance mechanisms for the observed multiple-layer MPS are more complicated, which is the focus of the present study.

The objectives of this study are twofold. The first objective is to examine how well the CRM simulates the occurrences, evolution and vertical structures of the multilayer MPS clouds by comparing the Baseline simulation with the M-PACE observations. The second objective is to explore the possible mechanisms for the formation, maintenance, and decay of the multilayer MPS clouds. To achieve the second objective, a set of sensitivity experiments are performed to examine the impacts of the large-scale advection, radiative cooling, surface heat flux, ice-phase microphysical processes, IFN number concentration, and latent heating caused by phase change of hydrometeor on the evolution and structures of these clouds.

Section 2 gives a description of the field measurements including the large-scale environment, cloud properties and aerosol properties. The numerical simulations are described in section 3. Analyses of the Baseline experiment and comparison with the observations are presented in section 4. Section 5 represents the results from the sensitivity experiments. Section 6 contains the summary and conclusions.

2. Field Measurements

2.1. Large-Scale Environment

The NSA was under three different synoptic regimes with two transition periods during M-PACE [Verlinde et al., 2007]. This study focuses on a three-and-a-half-day subperiod (14Z 5 October to 02Z 9 October) of the second regime (between 4 and 13 October). According to Verlinde et al. [2007], this synoptic regime was characterized by high pressure building over the pack ice to the northeast of the Alaska coast. As the high pressure system dominated the NSA until 15 October, a small midlevel low pressure system drifted along the northern Alaska coast from 5 to 7 October, and dissipated between Deadhorse and Barrow on 7 October. This midlevel low brought a considerable amount of middle- and upper-level moisture to the NSA. The low-level northeasterly flow out of the high pressure and the small midlevel disturbance related to the low pressure system combined to produce a complicated multilayer cloud structure over the NSA.

2.2. Cloud Properties

Clouds were observed by a wide range of instruments [Harrington and Verlinde, 2004], which were deployed at the ARM NSA surface sites (Barrow, Oliktok Point and Atqasuk; Figure 1) or aboard the two aircraft participated in the M-PACE. The University of North Dakota (UND) Citation served as an in situ platform. Cloud properties are derived from these surface and air-based measurements. Liquid water path (LWP) and precipitable water vapor were derived from the 2-channel (23.8 and 31.4 GHz) microwave radiometers (MWRs) deployed at the ARM NSA surface sites [Turner et al., 2007]. The time interval of the LWP is ~30 s. Other cloud properties that are used in the present study are described here.

2.2.1. Occurrences and Locations of Mixed-Phase Cloud Layers

The boundaries of the multiple mixed-phase cloud layers are determined by combing measurements from the MPL (Micropulse Lidar) and MMCR (Millimeter Wavelength Cloud Radar) deployed at Barrow (Figure 1). These measurements were available at a time interval of ~35 s. The vertical resolution of the MMCR is ~45 m and that of the MPL is ~30 m.

The cloud base height (Hcb) of the lowest water-dominated mixed-phase cloud layer is derived from the MPL measurements [Wang and Sassen, 2001]. If the optical thickness of the lowest mixed-phase cloud layer does not exceed approximately three, i.e., the MPL measurements are still available for the second layer, the lowest layer top (Hct) and the second layer base are also determined from MPL measurements. However, it is often that MPL signals are completely attenuated by the lowest mixed-phase cloud layer. Under this condition, MMCR measurements of reflectivity (Ze) and spectral width (σ) are used to determine the lowest layer top and the boundaries of the higher layers. A mixed-phase cloud is generally associated with larger values of σ than a water- or ice-only cloud, because the coexistence of small water droplets and large ice crystals increases σ. Therefore, the boundaries of a mixed-phase cloud are defined as the heights where σ becomes larger than a threshold σc of 0.4 m s⁻¹. However, σ could be larger than σc at heights right below cloud bases (see Figure 2e for an example). Hence, using σc alone could underestimate Hcb and overestimate cloud physical thickness (Δh). To solve this problem, Δh of the second or higher layer with Hct above 1.5 km is limited to be smaller than Δh (300 m), which is selected based on coincident MMCR and MPL measurements, by adjusting Hcb. Figure 2 shows that the second layer boundaries based on σc and Δh agree well with those based on MPL measurements. The uncertainties

![Figure 1](image-url)
are ~30 m for the lowest boundary (one MPL bin) and better than ~90 m for others.

2.2.2. Bulk Cloud Microphysical Properties

The bulk microphysical properties of the multilayer MPS clouds were derived from the UND Citation measurements on 5, 6, and 8 October (see details by Zhang et al. [2006] and Prenni et al. [2007]). The properties used in the present study include liquid water content (LWC), total ice water content (IWC), total water droplet number concentration \(n_c\), and total ice crystal number concentration \(n_{is}\).

The bulk properties are available at a 10 s interval, but represent a 30 s running average of the measured ice properties. McFarquhar and Cober [2004] and McFarquhar et al. [2007] gave a detailed description of the procedure to derive the bulk microphysical properties of the MPS clouds and the uncertainties associated with the derived products. A concise description of the aircraft observations is given below.

The UND Citation flew three missions dedicated to characterizing microphysics of the multilayer MPS clouds on 5, 6, and 8 October by executing spiral ascents and descents over Barrow and Oliktok Point and by flying ramped ascents and descents between. A typical flight pattern that the UND Citation took was presented in Verlinde et al., 2007, their Figure 5. The mission on 5 October started from about 1930 UTC (1130 local time) and lasted about two hours and fifteen minutes. The second mission was performed between 1830 UTC (1030 local time) and 2130 UTC (1330 local time) 6 October. The flight taken on 8 October lasted about two and half hours starting at about 2000 UTC (1200 local time). There are 628, 829, and 289 in-cloud observations obtained during the three missions, respectively, covering a total in-cloud period of about five hours. Here, in-cloud means the total condensed (liquid + ice) water content derived from probes on the Citation was greater than 0.001 g cm\(^{-3}\). The numbers of the samples of LWC and IWC within each of the 400 m height bins are represented in Figure 3. The sample numbers in the height bins vary from zero to 210 with relatively more samples taken between 400 m and 2 km. There are no samples at heights below 400 m for all three missions and few samples above 2 km for the 5 October and 8 October missions.

2.3. Aerosol Properties

In CRM simulations to be described in section 3, aerosol size distribution and chemical composition are specified for the calculation of droplet activation [Abdul-Razzak et al., 1998; Abdul-Razzak and Ghan, 2000] and ice nuclei (IN) concentration is also specified for the calculation of heterogeneous ice nucleation. In the absence of useful condensation nucleus data for aerosol size distribution during the simulation period (14Z 5 October to 02Z 9 October), and because the IN concentrations from the Continuous Flow Diffusion Chamber (CFDC; [Rogers et al., 2001]) aboard the Citation during this period show mean values and scatter similar to those recorded on the 9 and 10 October flights, we specify the aerosol properties and IN concentration based on the measurements obtained on 9 and 10 October, i.e., the same as by Luo08, Klein et al. (submitted manuscript, 2008), and Morrison et al. (submitted manuscript, 2008). It is further assumed that concentrations of aerosols and IN are horizontally and vertically homogeneous in the CRM domain, except for the contact IN explained below.

A bimodal lognormal aerosol size distribution was fitted to the average size-segregated Hand-Held Particle Counter (HHPC-6) measurement on 10 October, with the total aerosol concentration constrained by the average NOAA Earth System Research Laboratory condensation nuclei measurements [Morrison et al., 2008]. The geometric mean radii are 0.052 and 1.3 \(\mu m\), standard deviations are

![Figure 2. An example of vertical profiles of the (a) range corrected MPL return power (relative unit), (b) MMCR reflectivity Ze, and (c) MMCR spectral width for multiple layer mixed-phase clouds observed on 7 October 2004. Note the horizontal solid and dashed lines indicating mixed-phase layer base and top heights, respectively; and the vertical dotted line in Figure 2c indicating the threshold of spectral width used for detecting the second or higher mixed-phase layer.](image-url)
2.04 and 2.5, and the total number concentrations are 72.2 and 1.8 cm$^{-3}$ for the small and large modes of the aerosol size distribution, respectively. The measurements of active IFN concentration represent the sum of IFN with a diameter less than 2 μm acting in deposition, condensation-freezing, and immersion-freezing modes. They indicate locally high concentrations of IFN up to 24 L$^{-1}$, and a mean of about 0.16 L$^{-1}$ assuming that concentrations below the detection threshold are zero. The observed mean IFN number concentration is used to represent the aforementioned nucleation modes. Higher and lower IFN number concentrations are used in two sensitivity experiments, respectively (see section 3). No direct measurements are available for the number of IFN acting in contact-freezing mode. Thus the contact IFN number is a function of temperature following Meyers et al. [1992].

3. Numerical Simulations

[17] The CRM used in this study is the University of California at Los Angeles/Chinese Academy of Meteorological Sciences (UCLA/CAMS) CRM, which was originally developed by Steve Krueger and Akio Arakawa at UCLA [Krueger, 1988]. A modified version of this CRM [Xu and Krueger, 1991] was brought to the Colorado State University [Xu and Randall, 1995] and later to NASA Langley Research Center [Xu et al., 2005] where a few modifications were made to the CRM [Cheng et al., 2004; Luo et al., 2007; Luo08]. The CRM is based on the anelastic dynamic framework in two dimensions (x and z) with a third-order turbulence closure [Krueger, 1988]. The two-moment microphysics scheme of Morrison et al. [2005] and the radiative transfer scheme of Fu and Liou [1993] are coupled to the dynamic framework. Luo08 described the newly added prognostic variables of number concentrations of four hydrometeor types (cloud water, cloud ice, rain, and snow). Further details of this model can be found in the aforementioned references.

[18] Eight numerical experiments are performed, including the Baseline simulation and seven sensitivity studies (Table 1). The Baseline simulation is prescribed with time-varying large-scale advective tendencies of heat and moisture (Figures 4a and 4b) and surface fluxes of latent and sensible heat (Figure 4c). All simulations start from the same initial atmospheric state at 14 Z 5 October and are run for 84 h. They are performed with the same grid spacing of 2 km in the horizontal. The vertical grid spacing stretches from 100 m at the surface to 500 m at ~5 km and is 500 m above 5 km. The domain width is 256 km in the horizontal and 20 km in the vertical. A time step of 5 s is used. Vertical velocity is specified as zero at the upper and lower boundaries. Cyclic boundary conditions are used at the lateral boundaries. At the lower boundary, the vertical turbulent fluxes of momentum are diagnosed using flux-profile relationships based on Monin-Obukhov surface-layer similarity theory [Businger et al., 1971]. For radiation purpose, the spectral surface albedos for the six bands of Fu and Liou [1993] radiative transfer scheme are determined by combining the 3-hourly broadband albedo from the ARM analysis [Xie et al., 2006] with a curve of spectral albedo over fresh snow. The curve of snow spectral albedo is based on the data downloaded from the Clouds and the Earth’s Radiant Energy System/Surface and Atmospheric Radiation Budget (CERES/SARB) website (ftp://snowdog.larc.nasa.gov/pub/surf/data_tables.asc). Figure 4d shows the spectral albedos corresponding to a broadband albedo of 0.86. The skin temperature from the ARM analysis is used in all simulations for the calculation of surface upward longwave (LW) radiation. Radiative effects of the aerosols are not considered.

[19] The sensitivity simulations (Table 1) consist of noSfcFlx, noLSadv, noLWrad, noIce, IN50th, IN50, and noMicLat simulations, which are identical to the Baseline simulation except that one aspect of the experimental designs is artificially altered. These simulations are designed, as in previous modeling studies of single-layer clouds [e.g., Harrington and Olsson, 2001] and multilayer clouds [e.g., Harrington et al., 1999], to examine the roles of surface turbulent flux, large-scale advection, LW radiation, ice-phase microphysical processes, IFN, and latent heating/cooling in the formation and evolution of multilayer.

Figure 3. Profiles of the sample numbers for liquid water content (solid lines) and ice water content (dashed lines), respectively, in each height bin of 400 m during the three missions that the UND Citation took on (a) 5 October (a), (b) 6 October and (c) 8 October 2004.
Arctic clouds. The noSfcFlx simulation assumes that the surface turbulent fluxes of sensible and latent heat are zero. The noLSadv simulation neglects the large-scale advective tendencies of temperature and water vapor mixing ratio provided by the ARM analysis (Figures 4a and 4b) [Xie et al., 2006; Klein et al., 2006]. The noLWrad simulation sets the LW radiative cooling (heating) rates as zero. (We also performed another simulation in which the effects of both longwave and shortwave radiation are ignored. The results from this simulation are essentially the same as those from the noLWrad simulation and, therefore, are not included in this paper. This suggests that solar radiation is unimportant for the formation and evolution of the simulated MPS clouds, while it was found to cause the layering structure of the summertime Arctic liquid clouds [Herman and Goody, 1976].) The noIce simulation turns off all ice-phase microphysical processes. Only liquid-phase microphysical processes operate in this simulation. The number concentration of IFN acting in deposition, condensation-freezing, and immersion-freezing modes is increased and decreased, respectively, by a factor of 50 in the IN50 and IN50th simulations. The increased and decreased IFN concentrations represent the upper and low end of the IFN measurements. The noMicLat simulation neglects the latent heating or cooling due to all liquid- and ice-phase microphysical processes.

4. Baseline Results

4.1. Temperature, Moisture, Surface Precipitation

[26] The atmospheric temperature (T) and water vapor mixing ratio (q_v) decrease with height from nearly 0°C and

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Description</th>
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<tbody>
<tr>
<td>Baseline</td>
<td>standard baseline simulation</td>
</tr>
<tr>
<td>noSfcFlx</td>
<td>neglecting surface turbulent fluxes of latent and sensible heat</td>
</tr>
<tr>
<td>noLSforcing</td>
<td>neglecting large-scale advective forcing</td>
</tr>
<tr>
<td>noLWrad</td>
<td>neglecting longwave radiative cooling/heating</td>
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<tr>
<td>noIce</td>
<td>neglecting ice-phase microphysical processes</td>
</tr>
<tr>
<td>IN50</td>
<td>increasing ice-forming nuclei concentration from 0.16 L⁻¹ to 8.0 L⁻¹</td>
</tr>
<tr>
<td>IN50th</td>
<td>decreasing ice-forming nuclei concentration from 0.16 L⁻¹ to 0.003 L⁻¹</td>
</tr>
<tr>
<td>noMicLat</td>
<td>neglecting cooling/heating caused by phase changes of hydrometeors</td>
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*See text for further explanations.
at the surface to $-24^\circ$C and 0.5 g kg$^{-1}$ at $\sim 500$ hPa ($\sim 4.7$ km) in the Baseline simulation (Figures 5a and 5b). Typical differences in temperature between the Baseline simulation and the ARM analysis [Xie et al., 2006; Klein et al., 2006] are between $-2^\circ$C and $+2^\circ$C and those in $q_v$ are between $-0.25$ g kg$^{-1}$ and 0.25 g kg$^{-1}$. There are larger differences within the 850–700 hPa layer, i.e., cold (dry) biases up to $-4$ K ($-0.5$ g kg$^{-1}$) before 48 h and opposite biases of the same magnitudes after 48 h (Figures 5c and 5d). A primary reason for the large T biases is the unreasonable partitioning of ice water content (IWC) and liquid water content (LWC). IWC is underestimated and LWC is probably overestimated between 12–24 h in the simulation, resulting in extra radiative cooling (Figure 6b) and negative T biases near the cloud tops, as evidenced by the elevation of negative T biases with time during the first 48 h. This is due to the fact that optical properties of ice crystals and water droplets differ greatly for the same amount of ice/water. The large dry biases are unlikely caused by microphysical drying, as the surface precipitation is underestimated (Figure 5e). One possible cause is that the moistening associated with the midlevel low is underestimated in the large-scale forcing data. The positive biases in T and $q_v$ after 48 h may be related to the underestimation in clouds and precipitation during 44–60 h (Figure 5e).
4.2. Cloud Properties

4.2.1. Occurrences of Multilayer MPS Clouds

One of the unique features of the Arctic MPS clouds under study is that there are multiple mixed-phase cloud layers coexisting. Statistics of their occurrences are computed using the MMCR-MPL observations at Barrow. To compare with the observations, the number of mixed-phase cloud layers at each individual CRM grid column, as well as the base and top heights of the cloud layers, is determined by analyzing the profiles of cloud water mixing ratio (q_c) and cloud ice plus snow mixing ratio (q_ice) at a 5-h temporal interval from the Baseline simulation. A grid cell is considered as mixed-phase if q_c is larger than 0.01 g kg\(^{-1}\) and q_ice is larger than 0.0001 g kg\(^{-1}\); otherwise, it is clear. Using a threshold value of 0.0001 g kg\(^{-1}\) for both q_c and q_ice causes an increase in the occurrence frequency of 1% and 2%, respectively, for three-layer and double-layer mixed-phase clouds and a decrease of 1% for single-layer mixed-phase clouds. However, the major analysis results remain unchanged.

4.2.2. Mixed-Phase Cloud Layer Boundaries

An adequate simulation of cloud base and top heights is important since they are highly correlated with the downward LW radiative flux at the surface and the outgoing LW radiation at the top-of-the-atmosphere, respectively. The top and base heights of the single- and double-layer cloud layers are 41% and 31%. The Baseline reproduces the dominance of single- and double-layer clouds, as well as the increase of the single-layer cloud fraction and decrease of the double-layer cloud fraction, respectively, from 6 October to 7 October. For 8 October, 90% of the observed clouds is single-layer and 10% is double-layer. The Baseline simulation produces a larger fraction for the single-layer clouds (66%) than for the double-layer clouds (34%), qualitatively consistent with the observations.

4.2.3. Multilayer MPS Clouds

The occurrences and relative fractions of single-, double-, and three-layer mixed-phase clouds from the observations and the Baseline simulation are shown in Table 2. During 6 and 7 October, the observations reveal the occurrences of mostly single- or double-layer clouds. The fractions of the observed single-layer clouds are 49% on 6 October and 66% on 7 October and those of the double-layer clouds are 41% and 31%. The Baseline reproduces the occurrence of mostly single-layer clouds and decrease of the double-layer cloud fraction, respectively, from 6 October to 7 October.
Distribution of the observed cloud base height shows a mode at 625 m with about 70% between 250 m and 1 km (Figure 7d). Distribution of the observed cloud top height has a mode at 1.125 km and about 70% between 750 m and 1.5 km (Figure 7e). Compared to the observations, the simulated cloud bases and tops are lower. The cloud base-height distribution has a mode at the lowest bin (0–250 m) and about 70% below 500 m (Figure 7a). The cloud top-height distribution shows a mode of 875 m and ~60% below 1 km (Figure 7b). Too many occurrences

<table>
<thead>
<tr>
<th></th>
<th>1 Layer</th>
<th>2 Layers</th>
<th>3 Layers</th>
<th>1 Layer, %</th>
<th>2 Layers, %</th>
<th>3 Layers, %</th>
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<tr>
<td>MMCR-MPL 10/06</td>
<td>1186</td>
<td>997</td>
<td>206</td>
<td>49</td>
<td>41</td>
<td>9</td>
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<td>CRM 12–36 h</td>
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<td>23825</td>
<td>2584</td>
<td>29</td>
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<td>CRM 36–60 h</td>
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<td>225</td>
<td>8</td>
<td>90</td>
<td>10</td>
<td>0</td>
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<tr>
<td>CRM 60–84 h</td>
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<tr>
<td>MMCR-MPL 10/06–10/08</td>
<td>4728</td>
<td>1943</td>
<td>284</td>
<td>68</td>
<td>28</td>
<td>4</td>
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<td>CRM 12–84 h</td>
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Figure 7. Histograms of base height (a and d), top height (b and e), and physical thickness (c and f) of the first mixed-phase cloud layer above the surface from the Baseline simulation (left column) and the MMCR-MPL observations at Barrow (right column).
of the clouds near the surface are probably related to the moist bias below 900 hPa (~800 m) in the simulation (Figure 5d). Both the observations and the simulation suggest that most of the cloud layers are physically thin (Figures 7f and 7c) with about 93% and 80%, respectively, of the clouds being thinner than 750 m.

The observed cloud bases and tops of the second cloud layers are distributed quite evenly between 1 km and 4 km (Figures 8d and 8e), suggesting that there are small-scale variations in the cloud boundaries. These cloud layers are physically thin with thicknesses less than 500 m (Figure 8f). The histograms from the simulation appear significantly different from the observed ones. The simulated cloud base-height has a bimodal distribution. The mode at ~3.2 km is mainly caused by the clouds near the end of the simulation period (Figure 6a). The other mode at ~1.5–2.0 km is associated with the clouds during 12–36 h simulation period. The simulated tops are located at a few bins (Figure 8b), which can also be seen from Figure 6a, suggesting that the simulated cloud top heights are quite homogeneous in the horizontal. Moreover, the simulated clouds are physically thicker than the observed (Figures 8c and 8f).

Several factors may be responsible for the discrepancies in the vertical locations of the MPS cloud layers between the Baseline and MMCR-MPL observations. The large-scale forcing data used to drive the CRM may contain errors [Xie et al., 2006], possibly caused by the low data density during M-PACE and/or associated with the background field used to generate the forcing data. The forcing data are horizontally homogeneous within the CRM domain, which may contribute to the small horizontal variation in the simulated upper-layer tops. The vertical resolutions of the forcing data (25 hPa) and the CRM grid (100–500 m) are much coarser than those of the MMCR (30 m) and MPL (45 m). The CRM grid-spacing in the horizontal (2 km) could also play a role. Uncertainties associated with the model’s physics, such as turbulence and microphysics, cannot be ruled out as possible causes of the discrepancies.

Figure 8. Same as Figure 6 except for the second mixed-phase cloud layer above the surface.
4.2.3. Liquid Water Path (LWP)

The vertically integrated liquid water amount, i.e., liquid water path (LWP), is compared between the Baseline and the MWR-based retrievals [Turner et al., 2007] for the ARM surface sites at the NSA (Barrow, Atqasuk, and Oliktok Point). When temporally averaged over 78 h starting from 20 Z 6 October, i.e., the first 6 h of the simulation period is excluded in the averaging, the Baseline domain-averaged LWP is about the same as the MWR-based LWP averaged at the three sites (79 g m$^{-2}$ versus 81 g m$^{-2}$). However, the temporal variations of the simulated and retrieved 3-hourly averaged LWPs are different (thick lines in Figure 9). The simulated LWP decreases with time from 12 h to 48 h and increases at \(\sim 60\) h. The retrieved LWP, when averaged among the three sites, is relatively more constant with time. It is likely that the retrievals averaged among the three sites may not represent the evolution of the domain-averaged LWP. The retrieved LWPs at the three sites not only differ in the 78-h-averaged values: 124 g m$^{-2}$ at Barrow, 61 g m$^{-2}$ at Oliktok Point, and 57 g m$^{-2}$ at Atqasuk, but also evolve with distinct patterns (thin lines in Figure 9).

4.2.4. Bulk Microphysical Properties

The bulk microphysical properties of the MPS clouds including LWC, \(n_c\), total ice water content (i.e., ISWC), and total ice crystal number concentration (\(n_i\)), which are derived from the Citation measurements obtained during the missions on 5, 6 and 8 October [Zhang et al., 2006], are compared to those from the Baseline simulation during the subperiods of 12–24 h, 24–36 h, and 72–84 h, respectively. The three subperiods are denoted as subperiods A, B, and C hereafter. The Student's t-test is performed for LWC, \(n_c\), ISWC, and \(n_i\) respectively. Because of the small numbers of sampled LWC and \(n_c\) between 2 and 4 km obtained during the missions on 5 and 8 October (Figure 3), the simulated LWC and \(n_c\) located between 400 m and 2 km during the subperiods A and C and those located between 400 m and 4 km during the subperiod B are used in the Student's t-test, whereas the simulated ISWC and \(n_i\) located between 400 m and 4 km during the subperiods A, B, and C are used. Results from the Student's t-test (Table 3) suggest that the simulated and observed cloud properties have significantly different means, except for the LWC during the subperiod B, with the simulated LWC being larger and ISWC and \(n_i\) being smaller than the observations. The Student's T-statistics suggest that the simulated means of LWC and \(n_c\) are closer to the observed means than those of ISWC and \(n_i\), except for subperiod C.

Although the simulated and observed means are significantly different, the Baseline simulation qualitatively reproduced the major characteristics in the vertical distributions of LWC, \(n_c\), ISWC and \(n_i\) and their interperiod differences suggested by the Citation measurements (Figures 10–13, discussed below). In the following, the discussion will be focused upon the interperiod differences and the contrast with the single-layer clouds presented in Lu08.

The observations indicate that there are large inter-period variations in vertical distribution of the LWC. For example, the means and variations of LWC at heights of \(\sim 1\) km are larger on 8 October than those on 5 and 6 October (Figures 10d–10f). This change is qualitatively reproduced by the Baseline (Figures 10a–10c). The LWPs obtained during the 5 October mission have average values of about 0.05 g m$^{-3}$ for the 400 m to 1.6 km layer, with standard deviations of about the same magnitudes as the averages (Figure 10d). At the same layer, the Baseline LWPs averaged over the subperiod A are 0.06–0.08 g m$^{-3}$ (Figure 10a). For the subperiod B, the observations suggest that the LWPs have a relatively constant vertical distribution for the 500 m to 3.5 km layer (Figure 10e), which is nearly reproduced except that there is a minimum at 1.5 km in the simulation with averages of about 0.05–0.1 g m$^{-3}$ (Figure 10b). During subperiod C, the observed LWPs increase with height from 0.06 g m$^{-3}$ at 600 m to 0.15 g m$^{-3}$ at \(\sim 1\) km, with variations which are comparable to or larger than the means. For subperiod C, the simulated LWPs increase with height from about 0.06 g m$^{-3}$ at 500 m to 0.20 g m$^{-3}$ at \(\sim 1\) km, generally consistent with the observations.

Comparing with single-layer clouds that occurred during another subperiod of M-PACE [McFarquhar et al., 2007; Lu08], the monotonic increase of LWC with height is not produced, due to averages over a large number of clouds with varying cloud base heights in the present case (Figures 7a, 7d, 8a, and 8d). The single-layer feature is attributed to adiabatic growth of liquid water droplets as they ascend in an updraft. Tsay and Jayaweera [1984] observed similar characteristics for single-layer clouds and the upper layer of multilayer clouds.

The observations reveal that the droplet number concentrations have means of about 10–30 cm$^{-3}$ and standard deviations of about the same magnitude as the means. The simulated \(n_c\) is less than 60 cm$^{-3}$ in all three subperiods. The vertical distributions of the simulated \(n_c\) are similar for subperiods A and B, decreasing with height within the lower cloud layer and being relatively constant within the upper cloud layer. There is no observation below...
400 m to evaluate the simulated results, however. During the subperiod C, the simulated $n_c$ in the lower cloud layer is about two times that from the observations ($30–40 \text{ cm}^{-3}$ versus $15 \text{ cm}^{-3}$) although the LWCs are comparable (Figures 10c and 10f). In the upper cloud layer, the simulated $n_c$ has a value of $20–40 \text{ cm}^{-3}$, comparable to the observations.

[35] The decrease of $n_c$ with height in the first cloud layer above the surface (Figures 11a and 11b) differs from the constant vertical distribution of $n_c$ in the single-layer MPS clouds [McFarquhar et al., 2007; Luo08]. This distribution is expected from the cloud condensation nuclei activation parameterization used in the CRM [Abdul-Razzak et al., 1998; Abdul-Razzak and Ghan, 2000]. That is, the production of $n_c$ is dominated by the subgrid-scale vertical velocity, which decreases with height below 1 km (not shown). Seeding of the lower layer with ice crystals falling from the upper layer probably also contributes to the decrease of $n_c$ with height in the lower layer.

[36] The observations from 5 and 6 October suggest that the total IWCs have larger mean values and standard

<table>
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<tr>
<th>Subperiod</th>
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<th>$n_c$</th>
<th>ISWC</th>
<th>$n_{is}$</th>
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<td>0.18</td>
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<tr>
<td>C</td>
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<td>0.00</td>
<td>-16.25</td>
<td>0.00</td>
</tr>
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</table>

**Figure 10.** Vertical profiles of liquid water content from the Baseline simulation during (a) 12–24 h, (b) 24–36 h, and (c) 72–84 h and from the Citation measurements taken on (d) 5 October, (e) 6 October, and (f) 8 October. The solid lines represent the means and the shades represent plus and minus one standard deviation from the means.
deviations in the 400 m to 1.5 km layer than those at higher levels (Figures 12d and 12e). This vertical variation in IWC is reproduced by the Baseline simulation (Figures 12a and 12b). A major discrepancy in the results shown in Figure 12 is that the simulated ISWC is a few times smaller compared to the observed total IWC at the same height range above the surface. It is speculated that the near surface ISWC would also be underestimated if observations were available, based upon the underestimated surface precipitation rate (Figure 5e) because snow is the dominant form of precipitation in the M-PACE environment [Luo08].

Both the observations (Figures 13d–13f) and the Baseline results (Figures 13a–13c) suggest more ice crystals in the lower MPS cloud layer than in the upper cloud layer, partially due to the effect of seeding the lower layer with ice falling from the upper layer. In the CRM microphysics scheme, the H-M mechanism [Hallett and Mossop, 1974], which operates at temperatures between $-8^\circ C$ and $-3^\circ C$, is included for ice enhancement. The ice crystal number concentration is increased by the H-M mechanism at a horizontally averaged rate of several $L^{-1} h^{-1}$, which is too small to reproduce the observed number concentration of ice crystals (Figure 13). This suggests that some ice production mechanisms might be missing in the cloud microphysics scheme. The underestimate of $n_i$ was recently seen in simulations of the M-PACE single-layer MPS clouds [e.g., Fridlind et al., 2007; Luo08; Morrison et al., 2008], where ice enhancement through the H-M mechanism was not significant because the temperature ranged from $-15^\circ C$ (cloud top) to $-10^\circ C$ (cloud base), colder than the temperatures at which the H-M mechanism operates. The underestimate of $n_i$ in previous studies using cloud models was simply due to discrepancy between ice nucleus and ice particle concentrations [e.g., Hobbs, 1969; Koenig and Murray, 1976; Beard, 1992]. Several hypotheses have been proposed addressing this issue (see Beard [1992] for a
review), but they are uncertain and this remains an area of active research.

4.3. Summary

The Baseline simulation reproduced the dominance of single- and double-layer MPS clouds revealed by the MMCR-MPL observations and qualitatively captured the major characteristics in the vertical distributions of LWC, \(n_c\), ISWC and \(n_{is}\) and their interperiod differences suggested by the aircraft observations. However, the simulated first MPS cloud layer is too low and \(n_c\) within the lower layer decreases with height, in contrast to the relatively constant \(n_c\) revealed by the observations. This could be due to uncertainties associated with the parameterizations (e.g., turbulence, droplet activation), surface fluxes, and/or radiation. The second cloud layer is physically too thick with too large LWC, causing too strong LW cooling and negative biases in temperature. Both cloud layers contain too few ice crystal numbers and too small ice crystal masses, indicating missing of ice enhancement mechanisms in the microphysics scheme and resulting in the underestimate of surface precipitation rates.

5. Results From Sensitivity Experiments

The time-height cross sections of the horizontal-averaged LWC and ISWC from the sensitivity experiments (Figures 14b–14h) are compared to those from the Baseline simulation (Figure 14a) in order to examine the possible effects of surface latent and sensible heat fluxes, large-scale advection, LW radiation, ice-phase microphysics, IFN concentration, and heating/cooling by phase changes on the simulated cloud vertical structure and temporal evolution. To explore possible effects of the processes on dynamical circulations, eddy kinetic energy (EKE) is analyzed for the CRM simulations. Total EKE is the sum of the resolved EKE (RKE) and turbulent kinetic energy (TKE). The RKE at each grid point is defined as \(EKE = u'^2 + v'^2 + w'^2\), where \(u', v',\) and \(w'\) are the deviations of the velocities in the x-, y-, and z-directions from their horizontal averages. At the
horizontal resolution of 2 km, the RKE is the energy within mesoscale circulations. Vertical profiles of the horizontally and 12–84 h averaged total EKE, RKE, and TKE are compared among the Baseline, noSfcFlx, noLSadv, and noLWadv simulations (Figure 15). In the simulations, TKE decreases from \(1.0 \, m^2 \, s^{-2}\) at the surface to \(0.1 \, m^2 \, s^{-2}\) at 500 m and is very small above 500 m, where EKE is mainly contributed by RKE (Figures 15a–15d).

In the noSfcFlx experiment, the first event of the lower MPS cloud layer is significantly weakened and the cloud completely disappears after 36 h (Figure 14b). The second event of the lower cloud layer between 52 h and 84 h simulated in the Baseline does not appear in the noSfcFlx experiment. The upper-level cloud layer is nearly unchanged except for a slightly longer duration for the first event. The EKE is weakened in the 200 m to 1.5 km layer but stronger in the 2.5–3.5 km layer compared to the Baseline (Figure 15e), suggesting significant influences of clouds on dynamical circulation probably due to LW cooling near the cloud tops. This suggests that the lower MPS cloud layer in the Baseline simulation is closely related to the surface fluxes even though the magnitudes of these fluxes are small (18 ± 5 W m\(^{-2}\) and 3 ± 5 W m\(^{-2}\)). The switch-off of these fluxes mainly impacts the atmosphere in the lowest 1 km (Figure 16). The differences in potential temperature (\(\Theta\)) and q, between the Baseline and noSfcFlx experiments move to higher altitudes with time. The atmosphere is about 1 g kg\(^{-1}\) drier and 1 K cooler in the noSfcFlx experiment at the surface during the last 36 h. The negative values of \(\Theta\) difference at higher levels up to 2.5 km are related to the disappearance of the lower cloud layer, i.e., removing the radiative cooling effect at the cloud top (Figure 6b), while the positive values of \(\Theta\) difference above 2.5 km after 48 h are related to the longer duration of the upper-layer clouds, which end at 54 h in the noSfcFlx but at 45 h in the Baseline. These clouds increase the LW radiative cooling effects.

In the noLSadv experiment, the tops of the single-layer MPS clouds rise steadily from below 1 km at 6–12 h to ~3 km near the end of the simulation. The LWCs are up to one order of magnitude larger than those in the Baseline (Figure 14c). No second cloud layer is produced. The

Figure 13. Same as Figure 10 except for ice crystal number concentration.
The absence of the upper MPS cloud layer in the noLSadv experiment suggests that the cooling and moistening effects due to large-scale advection at the beginning of the simulation period (Figures 4a and 4b) must be a main factor for the initiation of the upper MPS cloud layer. The middle-level low that propagated over the region, as mentioned in section 2, brought the cloudiness with it. In the Baseline simulation, the initiation of the upper-layer clouds is delayed due to the fact that no initial condensate is prescribed. The steady rise of the boundary layer in the noLSadv experiment is driven by the surface fluxes, which initiate the lower cloud layer. Radiative cooling then further deepens the boundary layer through entrainment in order to achieve radiative-convective equilibrium. The deepened boundary layer clouds have smaller LWCs than those at the early stage of the simulation. The magnitude of EKE is a few times that from the Baseline simulation (Figure 15e), probably caused by strong LW cooling associated with the large LWCs.

From the noSfcFlx and noLSadv experiments, it is apparent that large-scale advective forcing and surface fluxes can be crucial factors for producing the upper-level and lower-level cloud layers, respectively. What are the additional factors that could destroy or prolong the structures of multilayer clouds? Five additional experiments are presented in order to answer this question.

In the noLWrad experiment, two events of single-layer MPS clouds appear between 6 and 48 h and between 62 and 84 h, respectively (Figure 14d). The timing of these two events is matched closely with that of the Baseline simulation, but the magnitudes of LWC are slightly different. This suggests that the lower cloud layer is closely associated with large-scale advective forcing and surface fluxes. The most striking feature of this experiment is that the upper cloud layer simulated in Baseline disappears when the LW cooling effect is switched off, similar to that in the noLSadv experiment in the sense that the upper cloud layer does not form. The total EKE is very small in the layer above 1.5 km (Figure 15d). As shown in Figure 6b, there is radiative heating in the cloud layer (a few degrees per day) that is accompanied by large radiative cooling near the cloud top (20 K d\(^{-1}\)) in the Baseline simulation. Without the radiative effect, the stability of the atmosphere is not favorable for the maintenance of the upper cloud layer because the large-scale advective cooling (6 K d\(^{-1}\)) is much smaller than the radiative cooling (Figures 4b and 6b).
the other hand, the radiative warming of the upper layer effectively changes the stability below the upper cloud layer, which creates the gap in the stability between the two cloud layers and enables for the maintenance of multilayer clouds. Combined with the results of the noLSadv experiment (Figure 14c), these noLWrad results suggest that (1) the upper MPS cloud layer in the Baseline is probably initialized by the large-scale advective forcing and maintained through the LW radiative cooling near the cloud top, (2) the LW radiative cooling contributes to the production of clouds probably through an enhancement of dynamical circulation, and (3) the cloud base warming of the upper cloud layer creates the stability gap that enables the separation of the two cloud layers.

The vertical distributions of clouds produced in the noIce experiment (Figure 14e) are significantly different from those in the Baseline simulation (Figure 14a), although the times of occurrence are more or less similar to the Baseline simulation. In particular, both cloud layers extend upwards and the LWP (224 g m\(^{-2}\)) is larger by a factor of 3, compared to the Baseline (79 g m\(^{-2}\)). The larger noIce LWP results from (1) the lack of cloud water consumption through the Bergeron-Findeisen mechanism and (2) the interactions between the simulated clouds and radiation, as more liquid droplets could result in a stronger radiative cooling which favors more condensation and thus a positive feedback could be formed.

The IN50th experiment produces cloud distributions (Figure 14f) that are somewhat similar to noIce, as the smaller IFN concentration results in less ice crystals and hence weakened consumption of cloud water through the Bergeron-Findeisen process. In the IN50 experiment, essentially no liquid droplets exist (Figure 14g) while the magnitude of the vertically integrated ice and snow mass increases by a factor of 6 from Baseline 41 g m\(^{-2}\) to 260 g m\(^{-2}\), suggesting that the CRM cannot produce the observed

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**Figure 15.** Vertical profiles of eddy kinetic energy (EKE) in the CRM simulations. Figures 15a to 15d represent total EKE (solid lines), TKE (short dashed lines) and resolved EKE (RKE; long dashed lines) in the Baseline, noSfcFlx, noLSadv, and noLWrad simulations, respectively. Figure 15e shows the total EKE from the Baseline (solid line), noSfcFlx (dotted line), noLSadv (short dashed line) and noLWrad (long dashed line) simulations. All results are horizontal means averaged over 12–84 h.
mixed-phase clouds when the upper end of the IFN measurements is used in the simulation. The rapid glaciation of cloud liquid water through the enhanced Bergeron-Findeisen process at the higher IFN concentration in the IN50 experiment is consistent with previous modeling studies [e.g., Harrington et al., 1999; Jiang et al., 2000; Lu08].

The cloud distributions produced in the noMicLat experiment (Figure 14h) are generally similar to those in the Baseline simulation (Figure 14a) except for a larger magnitude of LWC in the interior of the MPS cloud layers. As shown in Figure 6c, phase changes among the hydrometeors cause warming of several K d−1 below the cloud tops, which partially cancels out the strong LW radiative cooling effect there. Neglect of this warming effect could result in a stronger net cooling effect near the cloud top, which favors more condensation than in the Baseline simulation.

In summary, these sensitivity experiments show that both the surface fluxes and large-scale advective forcing control the formation of the lower cloud layer while the large-scale advective forcing initiates the formation of the upper cloud layer but maintenance of multilayer structures relies on the LW radiative effect, which favors condensation in the upper cloud layer through cloud top cooling and creates the stability gap between the two cloud layers through cloud base warming of the upper layer. Moreover, ice crystals consume cloud liquid droplets through the Bergeron-Findeisen mechanism and remove condensate from the cloud layer through precipitation. Therefore, without large-scale advection, which resupplies moisture, and cloud top LW cooling, which produces a more saturated environment, the upper cloud layer will dissipate. These results are consistent with Jiang et al. [2000] and Smith and Kao [1996], who discussed the importance of large-scale advection to the maintenance of single-layer MPS clouds [Jiang et al., 2000] and near surface liquid cloud layer [Smith and Kao, 1996], respectively. The results are also consistent with Harrington et al. [1999] and Smith and Kao [1996], who showed that MPS clouds [Harrington et al., 1999] and liquid clouds [Smith and Kao, 1996] decoupled from the surface are maintained by continued cloud top LW cooling. However, Harrington et al. [1999] did not include large-scale advection so the balance seems to have been a tenuous one. The present study clearly shows that it is the combination of large-scale forcing and LW cooling that allows the upper-layer clouds to be maintained. Furthermore, both the upper and lower mixed-phase cloud layers are very sensitive to IFN concentration because of the Bergeron-Findeisen mechanism. Finally, the microphysical changes, which feed into the LW radiative cooling, could significantly influence the mesoscale circulation that helps maintain the cloud layer. Morrison et al. [2008] showed that mesoscale circulations seemed to be driven by diabatic processes associated with clouds (e.g., cloud top radiative cooling), but they were also strongly modulated by the coastal topography, which is neglected in the CRM. Therefore, the present study is consistent with Morrison et al. [2008], except for the topographic effects.

6. Summary and Conclusions

Multiple-layer mixed-phase stratiform (MPS) clouds that occurred during a three-and-a-half-day subperiod of the DOE-ARM Program M-PACE [Harrington and Verlinde, 2004; Verlinde et al., 2007] have been simulated using a CRM. Several sensitivity simulations have been performed to provide insight into the physical processes responsible for multilayer production and evolution. This CRM includes a two-moment microphysics scheme [Morrison et al., 2005], a 6-four-stream radiative transfer parameterization [Fu and Liou, 1993], and a third-order turbulence closure [Krueger, 1988]. Concurrent meteorological, aerosol, and ice nucleus measurements are used to initialize the CRM. Time-varying large-scale advective tendencies of temperature and moisture and surface sensible and latent heat fluxes [Xie et al., 2006; Klein et al., 2006] are prescribed to the CRM simulations. The Baseline simulation results have been analyzed and compared to the M-PACE observations, including the analysis of atmospheric temperature and moisture biases, surface precipitation rate, and a variety of cloud properties, which can be summarized as follows.

The ARM analysis [Xie et al., 2006] suggests the occurrences of several precipitation events during the simulation period. The CRM captures the timing of the three of the five events except for the first event due to the model spin up and the fourth event due to the underestimate of cloud production. The magnitudes of the simulated precipitation rates are smaller or comparable to the ARM observations, due to underestimated ice water content. The magnitude of the simulated liquid water path (LWP), when averaged over the domain and the three-and-a-half-day simulation period, agrees with the MWR-retrievals averaged over three surface sites [Turner et al., 2007]. The MMCR-MPL measurements reveal mostly single- or double-layer MPS clouds. The Baseline simulation reasonably reproduces the relative frequencies of occurrence of the single- and double-layer MPS clouds. However, there are several discrepancies in the vertical locations of the MPS clouds between the Baseline simulation and the measurements. In particular, the simulated bases and tops of the lower MPS cloud layer are too low and the physical
thicknesses of the upper MPS cloud layer appear to be too large.

[50] The bulk microphysical properties derived from the Citation aircraft measurements taken on 5, 6, and 8 October [Zhang et al., 2006] have been compared to the Baseline results. The observations reveal that the liquid water contents (LWCs) during the 5 October and 6 October missions have relatively constant vertical distributions with means of about 0.05–0.1 g m⁻³ whereas those of 8 October have maxima at heights of ~1 km (~0.15 g m⁻³) and ~2.5–3.0 km (~0.01 g m⁻³). The droplet number concentrations (n_d) have mean values of 10–40 cm⁻³. The total ice water content (IWC) and ice crystal number concentration (n_ic) are several times larger in the lower MPS cloud layer (~0.05 g m⁻³ and a few tens L⁻¹) than in the upper MPS cloud layer probably due to the effect of seeding the lower layer with ice falling from the upper level. Comparison of the simulation with these measurements indicates that the Baseline simulation qualitatively reproduces the major characteristics in the vertical distributions and interperiod variations of LWC, n_d, ISWC, and n_ic. However, it overestimates LWC and underestimates IWC and n_ic. In particular, the simulated n_ice is one order of magnitude smaller than the observed. This is consistent with the simulation of single-layer MPS clouds performed by Luo et al. [2008]. The underestimate of IWC probably causes the underestimate of surface precipitation rate. The overestimate of LWC causes too strong LW cooling and negative temperature biases.

[51] Possible causes for the discrepancies in temperature, moisture, and cloud properties between the Baseline simulation and the M-PACE observations are likely due to the response of imperfect model physics to the imposed large-scale forcings although parts of the discrepancies can be due to errors in the imposed large-scale forcings. For example, the underestimate of n_ice and IWC by simulations with LES, CRM, SCM has been noticed by other modeling studies [e.g., Koenig and Murray, 1976; Fridlind et al., 2007; Luo08; Morrison et al., 2008]. The present study supports the hypothesis that some ice forming mechanisms may be missing in all models. On the other hand, the discrepancies could also be related to the small number of samples in the M-PACE observations (aircraft, in particular), uncertainties associated with the algorithms used to derive the cloud properties, and the inability of the field campaigns to measure all the spatial and temporal variability associated with these multilayer MPS clouds.

[52] Analyses of the sensitivity experiments have shown the following physical mechanisms for the formation and maintenance of multilayer MPS clouds. First, both the surface fluxes and large-scale advection control the formation of the lower cloud layer although the surface fluxes of latent and sensible heat used in the present study are small (18 ± 5 W m⁻² and 3 ± 5 W m⁻², respectively), compared to those used in Luo08 (136.5 W m⁻² and 107.7 W m⁻², respectively) and Harrington and Olsson [2001] (about 150 W m⁻² and 300 W m⁻², respectively). Second, the large-scale advective forcing initiates the formation of the upper cloud layer but the maintenance of multilayer structure relies on the LW radiative effect through the interactions between the cloud top LW cooling and cloud condensate, which results in stronger mesoscale circulation. The cloud base radiative warming of the upper layer also creates the stability gap between the two cloud layers and helps maintain the double-layer structure. Third, the ice-phase microphysics provides a strong sink of cloud liquid water mass. Therefore, both the upper and the lower MPS cloud layers are sensitive to the IFN number concentration: no MPS clouds are produced in the CRM when the upper end of IFN measurements is used. Harrington et al. [1999] also showed that the upper MPS cloud layer decoupled from the surface is maintained by continued cloud top LW cooling. However, the roles of large-scale advective forcing as the initiation mechanism of the upper-layer clouds and the control of the timing of the lower-layer clouds are crucial in maintenance and evolution of the multilayer MPS clouds examined in this study. Moreover, the latent heating counterbalances the LW cooling near the top of the upper cloud layer to reduce the production of condensate. Finally, the strength of the mesoscale circulation in the CRM is significantly impacted by cloud top LW radiative cooling and appears to have contributed greatly to the MPS cloud layers.

[53] The major contribution of this study can be summarized as follows. A detailed statistical comparison between the observed and CRM-simulated cloud properties has been provided for the multilayer MPS clouds, which is largely similar to Luo et al. [2003] for thin cirrus clouds. In the present study, not only the macroscopic properties of the lower- and upper-cloud layers but also the vertical distributions and interperiod variations of cloud microphysical properties are compared. This comparison provides an ideal framework for future modeling studies of multilayer clouds of any type. Analyses of sensitivity experiments have advanced the basic understanding of the formation and evolution of multilayer Arctic MPS clouds, which should be helpful to interpret the results of model intercomparison of this M-PACE subperiod (Morrison et al., submitted manuscript, 2008).

[54] The spatial resolutions used in the simulations are probably too coarse given the fact that many observed clouds are only a few hundred meters thick. The simulated results could be sensitive to the spatial resolutions. The possible effects of varying resolutions on the CRM simulations will be investigated in the future. Furthermore, the conclusions drawn from this study are limited by the short period of simulation time and small number of observational samples. Further studies of this and similar cases, especially with an improved cloud microphysics scheme that can better reproduce the observed ice-phase microphysical properties, will be needed to confirm the conclusions drawn from this preliminary study.

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