Cloud resolving simulations of Arctic stratus
Part II: Transition-season clouds

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Abstract

Two-dimensional simulations of transition (fall and spring) season Arctic stratus clouds (ASC) were conducted using a sophisticated cloud resolving model with bin microphysics coupled to a two-stream radiative transfer model. The impacts of temperature variation and various ice microphysical processes on the evolution of the simulated mixed-phase ASC layer are studied. Cloud layers either collapse through rapid glaciation and ice precipitation from the cloud layer or maintain a quasi-steady state. Sensitivity studies show that the stability of the mixed-phase cloud layer is dependent upon the temperature, ice concentration, and the habit of the ice crystals. In particular, cloud layer stability is shown to be most strongly dependent upon the concentration of ice forming nuclei (IFN). In addition, it is shown that ice production and sedimentation can assist the formation of a second, lower cloud layer suggesting a new mechanism of multiple-layer formation in the Arctic. © 1999 Elsevier Science B.V. All rights reserved.

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1. Introduction

During the fall and spring seasons within the Arctic Basin north of Alaska, transitions in low cloud types and amounts occur. The summer season produces low-lying stratus cloud over the Arctic ocean regions which are widespread and persistent in nature with...
cloud fractions between 60 and 90% (Herman and Goody, 1976). Low level cloudiness during the summer months is dominated by stratus clouds which are predominantly liquid and appear to have regular cellular structures (Curry et al., 1988). In addition to this, these stratus seem to be self-maintained entities, requiring little in the way of synoptic scale forcing to maintain the system once it has formed (Herman and Goody, 1976; Curry and Herman, 1985).

Traditionally, cloud climatologies have shown a transition in low cloud amount during a one month time period in fall and spring, with low-level cloud fractions dropping as low as 30% during winter and being dominated by ice crystals. However, this wintertime low level decrease in cloud cover has recently been called into question (Curry and Ebert, 1992) as clear sky ice crystal precipitation is frequently ignored or erroneously classified in cloud climatology data bases. Part of the reason for such discrepancies is the lack of and difficulty with remote sensing measurements (Curry et al., 1996). If low-level cloud fractions in the winter are as large as has been suggested (Curry et al., 1996) then some mechanism must be maintaining the cloud systems; leads and large scale flow (Pinto and Curry, 1995; Curry et al., 1997) appear to play a role in the process. Because of the existence of the ice phase, there is no reason to assume that the mixed-phase clouds are maintained by the same mechanism as the summertime Arctic stratus clouds (ASC). The initiation of the ice phase can have a large impact on cloud evolution as ice crystals attain large sizes through vapor growth alone (unlike water drops) thus increasing the removal of water mass from the cloud system through precipitation (as compared to pure water clouds). For this reason, mixed-phase clouds are considered colloidally unstable. Whether or not fall and spring low-level clouds are self-maintaining 1 entities as summertime ASC may, therefore, be called into question.

Although observations of mixed-phase ASC are sparse, experiments such as the Beaufort Arctic Sea Experiment (BASE) in 1994 have produced a source of data. Curry et al. (1997) results, for example, illustrate the persistence of mixed-phase clouds over the Arctic in the fall. Pinto (1998) analysis of two fall ASC cases illustrate the importance of large scale moisture and temperature advection for the maintenance of these clouds. In addition, Pinto speculates on the importance of ice forming nuclei (IFN) to cloud stability and shows that some observed cases can have thermodynamic structures similar to summertime clouds. These results are consistent with the preliminary modeling results of Harrington et al. (1997). It is apparent, nevertheless, that low clouds undergo a transition in microphysical structure (liquid to mixed-phase) that may have a strong impact on their overall stability.

The nature of seasonal cloud-type transitions is, for the most part, neglected in current research on Arctic low level cloudiness. There is, however, a growing interest in these transitions which are motivated by climate studies showing the sensitivity of the region to small changes in model parameters (e.g., Royer et al., 1990; Walsh and Crane, 1992), including responses of the Arctic sea ice cover (Curry et al., 1993; Ebert and

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1 We define self-maintaining cloud systems as persistent cloud systems that can be maintained almost exclusively by cloud-scale processes. Liquid-phase ASC may be simulated in a limited-domain framework (e.g., CRM or LES) as resolved cloud scale processes can maintain the system (Olsson et al., 1998).
Curry, 1993 and sensitivity to aerosol concentrations (Curry, 1995). Simulations in which sea ice cover is reduced showed that low cloud fractions can either increase or decrease depending upon how model convection is parameterized (Royer et al., 1990). Transitions during the fall and spring can have a significant impact on the heat budget of the Arctic through changes in cloud and radiative properties which impact the equilibrium sea-ice thickness (Curry et al., 1993; Pinto and Curry, 1995). This can feed back positively into cloud fractional coverage (Pinto and Curry, 1995) and may have an impact on climate simulations. It is certain that changes in sea ice/ocean heat budgets and large scale flow patterns must play some role in the processes that control the fall and spring seasons ASC structure. It seems plausible, however, that changes in the microphysical nature of the system should also play a role.

In an initial attempt to explore the microphysical impacts on the evolving structure of mixed-phase ASC that may exist during the fall and spring, a detailed mixed-phase microphysical model (Reisin et al., 1996) and a sophisticated radiation scheme (Harrington, 1997) are coupled to a cloud-resolving model (CRM) version of the Regional Atmospheric Modeling System (RAMS) developed at Colorado State University (Pielke et al., 1992). Specifically, the microphysical conditions necessary to produce self-maintaining mixed-phase systems and collapsing boundary layers during a simulated transition from predominantly liquid to mixed-phase ASC are explored. This is accomplished by consistently cooling a summertime ASC sounding to produce physically plausible mixed-phase situations. We note at the outset that these sets of studies are hypothetical in nature; we are not attempting to reproduce actual cases but, instead, are using these simulations to build ideas concerning how mixed-phase ASC (or stratus in general) might behave under the influences of ice forming nuclei. Mixed-phase clouds are produced that either persist for the entire 8 h simulation period or completely collapse causing buoyancy-driven circulations within the boundary layer to cease. Examinations of the dependence of boundary layer stability on the ice phase is examined by conducting sensitivity experiments in which IN concentrations are altered, model processes are removed, and the habit of the ice crystals is changed.

2. The numerical model

2.1. Description of the dynamical model

The dynamical modeling framework used for this study is the two-dimensional analog (Cloud Resolving Model or CRM) of the Large Eddy Simulation technique (or LES) which is a fine resolution version of RAMS. This model has been used successfully to simulate the impact of drizzle production by stratocumulus on the marine boundary layer (Stevens et al., 1996; Feingold et al., 1996). The version of the model used here integrates equations for the wind components ($u$ and $w$ in the 2-D framework).

Because of the lack of data at the time (no soundings available) for mixed-phase Arctic low clouds, this was the only reasonable alternative. The physical plausibility of the method will be examined in relation to recently published observations (Pinto, 1998).
the perturbation Exner function ($\pi$), the total water mixing ratio ($r_w$), and the ice–liquid water potential temperature ($\theta_{ul}$) on a vertically stretched Arakawa C-grid. Further details of the model may be found in Stevens et al. (1996) and Olsson et al. (1998).

2.2. The explicit microphysical model

In order to examine the details of the ice–liquid microphysical interactions that occur within transition season ASC, we couple a detailed, bin resolving liquid–ice microphysical model (Reisin et al., 1996) to the dynamical framework of the CRM. Earlier studies which couple detailed, bin microphysical models to two and three-dimensional LES models have considered the effects of the liquid phase only (e.g., Feingold et al., 1994; Kogan et al., 1995). To our knowledge, this is a first attempt at modeling mixed-phase stratus clouds with a detailed, binned representation of ice microphysics. Bin microphysical models have enormous advantages over conventional bulk (or explicit) microphysical schemes (e.g., Ferrier, 1994; Walko et al., 1995) because the explicit resolution of the ice-drop spectra with a fixed-grid precludes the necessity of certain parameterized effects (e.g., transfer rates of crystals, distribution shape assumptions, terminal fall speed methods among others). Thus, the drop-ice distribution functions are allowed to evolve more freely than in recently developed single-moment or double-moment bulk microphysical schemes (Meyers et al., 1997). This increased accuracy comes, however, at an enormous computational expense.

The ice–liquid bin microphysical model (BM) solves equations for condensation (deposition) $^3$ (Stevens et al., 1996), stochastic collection and sedimentation (Reisin et al., 1996) on an Eulerian grid in radius space. The grid covers a diameter range of 3.125 to 1000 $\mu$m using 25 bins. Bin edges are defined by a mass-doubling formula,

$$m_k = 2^{k-1}m_1,$$

where $m_1$ is the mass of a drop or ice crystal of diameter 3.125 $\mu$m and $k$ is the bin edge which ranges from 1 to 26.

The BM model predicts on the zeroth (number) and first (mass) moments for the mass distributions of four hydrometeor classes; this constitutes a total of 200 scalars which must be predicted by the model. The hydrometeor classes include a liquid phase, pristine ice crystals, aggregates, and graupel. The liquid phase is initiated by activation of a log-normal population of CCN as a function of supersaturation (Feingold et al., 1994). The CCN concentration utilized in this study, 100 cm$^{-3}$, is within the range of measured aerosol concentrations in the arctic (e.g., Hegg et al., 1995). The pristine ice class has ice nucleation (Meyers et al., 1992; Harrington et al., 1995) as a mass and number source; the crystals may grow by vapor diffusion and retain small amounts of rime. Ice concentrations predicted by the Meyers et al. (1992) formula are well within the bounds of measured values over the arctic ocean (e.g., Jayaweera and Ohtake, 1973; 1993).

$^3$ Here, we adopt the standard terminology in the vernacular proposed by McDonald (1958). The terms condensation and evaporation are, as is standard, reserved for vapor–liquid growth mechanism while deposition and sublimation are used for the analogous vapor–ice growth mechanism.
Ohtake et al., 1982; Curry et al., 1990), although they do appear to be closer to the large-end of the concentration range (Bigg, 1996). In addition, comparisons (shown later) of our simulated ice concentrations to those of Pinto (1998) case show that our ice concentrations are, at times, somewhat high. In situ sources of IFN over the ice pack appear to be rather limited and associated with leads, open water, and large scale advection (Bigg, 1996; Curry et al., 1996; Pinto, 1998). Because of this, and because our case is assumed to be over the ice pack, we allow IFN to be removed through the nucleation–precipitation process. The aggregate class is affected by pristine ice–aggregate and aggregate–aggregate collection events and may also retain small amounts of rime. The significant riming of pristine ice and aggregates produces graupel (see Reisin et al., 1996 for a full discussion).

2.3. The two-stream radiation model

The two-stream model described in Harrington (1997) is coupled to the bin microphysical framework for these studies. The two-stream model solves the equation of transfer for three gaseous constituents, \(H_2O\), \(O_3\) and \(CO_2\) and the optical effects of the hydrometeor size spectra. The gaseous absorption problem is solved following the FESFT method proposed by Ritter and Geleyn (1992), however the number of \(k\)-coefficients is reduced through trial and error fitting which reduces computational costs further (Harrington, 1997). The two-stream fluxes are updated every 20 s throughout the course of the simulation.

Optical properties are computed consistently with the various microphysical schemes employed in the CRM. Lorenz–Mie theory is used to compute the optical properties for water drops, while the theories of Mitchell and Arnott (1994) and Mitchell et al. (1996) are used for non-spherical ice crystals. The bin microphysical model uses a method of computing the optical properties whereby bin averages of the appropriate quantities are computed beforehand, and then summed with appropriate weights during the simulation (Harrington et al., 1998). This gets around the problem of having to specify an a priori shape of the ice and water distribution functions. In addition, the method allows for the variation in hydrometeor distribution shape to feedback into the radiative fluxes in a realistic manner. The method has been tested against gamma distribution functions for which analytical solutions are known. Comparisons show that the method produces excellent accuracy with errors never greater than about 2% (Harrington, 1997).

3. Initiation of the transition season clouds

Since a lack of data exists for mixed-phase clouds that occur over the Arctic ocean and ice pack during fall and spring (transition seasons), we produce what we consider a physically plausible situation by cooling a sounding obtained from a summer season ASC case until ice forms. The case utilized here is the June 28, 1980 ASC case which has been successfully simulated with one-dimensional models (McInnes and Curry, 1995; Smith and Kao, 1996) and, recently, with a two-dimensional CRM coupled to a liquid phase bin microphysical model (see Olsson et al., 1998; Harrington et al., 1998).
This case was characterized, microphysically, by multiple liquid layers (common during the Arctic summer) with intermittent drizzle produced by the upper cloud layer (Tsay and Jayaweera, 1984). The upper cloud deck was about 300 m thick and existed within a well-mixed layer in which cloud top radiative cooling was the strongest driving mechanism for turbulence (Curry et al., 1988). The lower cloud existed within a strongly sheared and very stable (lapse rate of 30°C km⁻¹) layer; heat and moisture fluxes from this layer had little to no effect on the upper cloud deck (Curry et al., 1988). Since the lower layer is of little importance to the studies of the mixed-phase upper cloud deck, this lower fog layer is removed from the initialization data. This was accomplished by keeping the RH approximately constant from about 400 m down to the surface. Some initial testing showed that the removal of this layer had little impact on our simulations because the fog layer existed within a very stable surface layer. Little to no mixing occurs within this region and, a study in which parcels were released within this lower layer showed zero flux of parcel tracers out of the stable surface inversion. A complete description of the observations associated with this case may be found in Curry (1986).

Since cloud top temperatures for the June 28, 1980 case hovered just a few degrees above zero, we produce a natural transition in cloud type by cooling the representative sounding in 5°C increments while keeping the relative humidity constant. The two cases are denoted by the amount of cooling of the sounding, which is 5°C and 10°C. The mean θ and θe profiles for each case are shown in Fig. 1a and are characterized by the same features as the summertime ASC case. The winds (Fig. 1b) show that little shear occurs within the upper cloud layer which is consistent with the strong mixing. Recently, the Beaufort Arctic Sea Experiment (BASE) which occurred off of the northern coast of Alaska and Canada in the autumn of 1994 produced some of the first observational data on the microphysics of these mixed-phase cloud cases. The soundings used here, along with the morphology of the cloud layer, compare well with the soundings shown by Pinto (1998) for one of the cases observed during BASE. In that case, a 300 m thick

![Fig. 1. Initialization: (a) θ is given by the solid line while θe is given by the dashed line, (b) horizontal winds (u and v components) used in the model initialization are also shown.](image-url)
mixed-phase cloud layer capped the Arctic boundary layer over a solid sheet of sea ice. The cloud layer existed within a mixed layer and showed a thermodynamic ($\theta$) profile which is strikingly similar to the ones used for our cases. In addition, the similarity in the analysis (see below) between our simulated cases and those of Pinto (1998) suggests that our soundings are physically appropriate.

3.1. Results for initiated clouds

The CRM model is run for both the 5°C and 10°C cooled soundings using the bulk microphysics of Walko et al. (1995) over a four hour period so that a mixed-phase cloud layer is produced with minimal computational cost. The model is initialized with a random perturbation in $\theta$ ($-0.1$ K $\leq \delta \theta \leq 0.1$ K) to add horizontal inhomogeneity; the perturbations are forced to satisfy, $\int \delta \theta(x, z = \text{const})dx = 0$, so that the average thermodynamic structure of the boundary layer is not altered. Simulations with the bin microphysical model are conducted for a further four hour period with the spin-up simulations being used as a common point of departure.

Profiles of the relevant liquid water variables are shown in Fig. 2 after four hours of the initial spin-up simulation. All profiles are averaged over the horizontal domain and temporally in 15 min blocks. After four hours, the 5°C cooled simulation produces a cloud with greater liquid water content (LWC) spread over a larger depth than the 10°C cooled case. More ice water content (IWC) covering a greater vertical extent and, with a maximum at cloud base, is produced in the 10°C case. This maximum in IWC is due to the larger sizes and, hence, larger fall-speeds of the ice hydrometeors in 10°C ($r_{\text{c}}$). These results are similar to Pinto’s (1998) observations which showed smaller ice persisting within the liquid layer and larger ice crystals near cloud base. More ice hydrometeors are nucleated in the 10°C cooled case ($N_i$) because of larger ice supersaturations (10% vs. 6.5%); a prominence in $N_i$ at about 300 m in the 10°C case is due to ice precipitation. The larger IWCs produced in the 10°C cooled simulation are due to the large ice supersaturations and the Bergeron–Findeisen process (Pruppacher and Klett, 1997). Differences between the equilibrium vapor pressure $^5$ over water ($e_w$) and ice ($e_i$) maximize near the cloud top temperatures of the 10°C cooled case (about $-13°C$), thus ice will grow more rapidly at the expense of liquid drops.

Total integrated condensed water paths for all of our simulations fall within the range of 20 to 80 g m$^{-2}$ (most frequently within 40 to 60 g m$^{-2}$). Pinto (1998) data shows total water paths between about 15 and 60 g m$^{-2}$ for a case which is 7 to 10°C cooler than our cases. Thus, it seems that our hypothesized cases are not excessively moist.

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$^4$ Ice and water effective radii ($r_e$ and $r_{\text{c}}$) are defined as the ratio of the third to the second moment of the size distribution functions.

$^5$ We adopt the terminology of Bohren (1987) in which the standard phrase, ‘saturation vapor pressure’, is replaced by, ‘equilibrium vapor pressure.’ Indeed, as discussed by Bohren (1987), the word equilibrium better describes the physics behind the vapor pressure derivation.
In the following sections we explore the evolution of the cloud systems utilizing the bin microphysical model. The control simulations will be discussed first, followed by analysis of sensitivity studies.

4. Control simulations of two mixed-phase ASC cases

The main difference between the simulated mixed-phase ASC cases is that the 5°C cooled cloud can maintain itself against collapse while the 10°C case cannot. This is an important effect as it is not known what causes the maintenance of collooidally unstable clouds in the Arctic, nor is the reason for the seasonal decrease in stratus cloudiness known. In Sections 4.1 and 4.2, the differences (microphysically and thermodynamically) between these two cases are explored.
4.1. The 5°C cooled control simulation

Since the mixed-phase system is colloidally unstable, one expects that the coexistence of liquid water and ice will cause the depletion of the liquid over a certain period of time. Thus, in order for mixed-phase ASC to attain some sort of self-maintaining state (which guards against cloud layer collapse) a balance between ice water production and precipitation must be realized.

As the time-series plot of LWC and IWC in Fig. 3 shows, the mixed-phase cloud layer undergoes a rapid glaciation period after 4 h of simulation time (at 1900 Z). Ice water content is rapidly produced, at the expense of the liquid, and removed through sedimentation during the 4–5 h period of the simulation. Throughout the remainder of the simulation, the LWC slowly increases while the IWC appears to maintain a quasi-constant vertical profile. The rapid glaciation period is due to the fact that, once enough ice is produced, the Bergeron–Findeisen process can act very rapidly. Fig. 4 shows the rapid reduction of ice concentrations ($N_i$), which originally peaked around 2 l−1, over the 4–5 h period. These concentrations, produced by the Meyers et al. (1992) formula, are somewhat higher than some measurements (0.5 l−1 ice concentrations in Pinto (1998) data). In all of these cases, almost all of the ice is produced through vapor depositional growth; ice aggregation and graupel production (see Harrington, 1997) make-up only a very small fraction of the total IWC (less than 0.03%), thus IWC and $N_i$ will be assumed to be synonymous with pristine ice for the remainder of this paper.

![Fig. 3. Time series: LWC (shaded) and IWC (contoured) for SCTRL in g m−1.](image-url)
Because of the large precipitation rates that occur during the 4–5 h period, significant sublimation of ice occurs within the dry regions below cloud base. As Fig. 5a shows, water vapor content \( r_v \) is reduced within the region of the cloud layer while the regions below cloud are substantially moistened through ice sublimation. This causes significant cooling of the lower layers and leads to substantial stabilization of the lower BL within only a one hour time-period \( (\theta, \text{Fig. 5b}) \). In addition, the upper part of the cloud deck continues to cool in spite of the fact that significant latent heat is released.
Fig. 6. Profiles of: (a) radiative heating rate ($\partial \theta / \partial t_{\text{radiation}}$) and (b) heating due to microphysics ($\partial \theta / \partial t_{\text{mic}}$) for 5CTRL. The solid line denotes fields at hour 4, long-dashed line denotes 4.5 h, and short-dashed line denotes 5 h.

through ice growth. An examination of Fig. 6 shows why, the microphysical heating rate ($\partial \theta / \partial t_{\text{mic}}$) is not strong enough to dominate over the radiative cooling ($\partial \theta / \partial t_{\text{radiation}}$) that is occurring, even though radiative cooling is weakening in time. This leads to a total heating rate which is negative within the cloud layer (as expressed by the $\theta$ plots), thus allowing for further condensation and deposition.

Since cloud-scale circulations are driven primarily by cloud-top radiative cooling in this case (Curry et al., 1988; Olsson et al., 1998), vertical motions weaken with the

Fig. 7. Profiles of: (a) buoyancy production of $\langle w' \hat{w}' \rangle$ and (b) vertical velocity variance ($\langle w' w' \rangle$) for 5CTRL. The solid line denotes fields at hour 4, long-dashed line denotes 4.5 h, and short-dashed line denotes 5 h.
Fig. 8. Microphysical profiles of: (a) ice concentration \(N_i\) and (b) ice effective radii \(r_{ei}\). Solid line denotes 6 h, long-dashed line denotes 7 h and the short-dashed line denotes 8.

Reduced radiative cooling, however the motions are still strong enough to maintain the cloud layer. This is shown in a plot of the vertical velocity variance \((w'w'')\); Fig. 7) and the buoyancy production of \(w'w'\). Buoyancy production is initially reduced because of the reduction in LWC; this closely follows the evolution of the LWC, however circulations remain shallow and weaker because of the reduced radiative cooling associated with the reduced cloud depth (Fig. 3). Interestingly enough, observations of shallow circulations with similar cloud morphology exist for autumnal ASC (Pinto, 1998).

Through the remainder of the simulation (5–8 h), LWCs slowly increase while IWC and \(N_i\) remain low with quasi constant-in-height profiles (Figs. 3 and 8). Ice crystal distributions are dominated by few large crystals \(r_{ei}\) which grow in the vapor-rich, low \(N_i\) cloudy environment and subsequently precipitate from the cloud layer. The shape of these profiles shows that this process of creation of crystals within the cloud and removal by precipitation is fairly constant. The balance is maintained by radiative cooling \((\partial \theta/\partial t_{\text{Radiation}})\) and microphysical heating \((\partial \theta/\partial t_{\text{Mic}})\) at the top of the cloud, which have both equilibrated (Fig. 9). Radiative cooling dominates the heat budget causing an overall cooling of the cloud \((\theta)\). It is noteworthy that, once this quasi-equilibrium state is achieved, \(N_i\) values are closer to observed values (Pinto, 1998).

Since the liquid water cloud is not depleted, the continued cloud top cooling assists in the production of vertical circulations which drive the condensation/evaporation processes within the cloud layer. In addition, this mixing redistributes heat throughout the cloud layer, thus mixing down radiatively cooled air \((\theta, \text{Fig. 9c})\). As LWCs increase in time, so do the turbulent fluxes of liquid and ice water \((<w'r_i'>, <w'r_i''); \text{Fig. 10b and d})\). By 8 h, significant drizzle \(F_{\text{drizzle}}\) in the form of liquid and ice dominates the

\(^6\)This fluxes is defined as, \(F_{\text{drizzle}} = \sum_{k=1}^{25} m_k v_{tk}\), where \(k\) is the bin number, \(m_k\) is the mass of drops within bin \(k\), and \(v_{tk}\) is the terminal fall speed of drops in bin \(k\).
Fig. 9. Profiles of: (a) radiative heating rates \( (\partial \theta / \partial \text{Radiation}) \), (b) microphysical heating rate \( (\partial \theta / \partial t_{\text{mic}}) \), and (c) \( \theta \) for SCTRL. The solid line denotes fields at 6 h, long-dashed line denotes 7 h, and short-dashed line denotes 8 h.

The precipitation from the cloud layer redistributes moisture and cools the lower portion of the boundary layer \( (\partial \theta / \partial t_{\text{mic}}) \), Fig. 9b which leads to increasing stability. In addition, the moistening of the layers below cloud permits the extension of the cloud to lower levels, as is shown in Fig. 3. This is consistent with recent observations of mixed-phase ASC over the Beaufort sea in autumn (Curry et al., 1997; Pinto, 1998) which show that mixed-phase clouds evolved to deeper, primarily crystalline clouds in

\[ F_{\text{liq}} \] and \[ F_{\text{ice}} \] below cloud base. Reduced drop and ice crystal concentrations in conjunction with increasing vertical motions and continued cooling of the layer helps to produce the large hydrometeors that contribute to this flux.

The total flux is defined as \[ F_{\text{liq}} = F_{\text{liq},l} + <w'r_i'> \] for liquid and analogously for ice.
time; it was also observed that the mixed-phase ASC were associated with stable lapse rates. Our results suggest that these features may be due, in part, to ice production and precipitation from cloudy, mixed-phase layers over the ice pack in autumn.

4.2. The 10°C cooled control simulation

Cooling the representative sounding a further 5°C (cloud top temperatures around −13°C), produces a distinctly different evolutionary scenario than the 5°C cooled case. After 4 h of model integration, the cloud once again undergoes a rapid glaciation period in which significant ice, once produced, quickly grows at the expense of the drops (Fig. 11). In this case, however, the growth of the ice crystals is so rapid that the available water mass is lost to the surface through precipitation (analogous to Ackermann et al., 1993 for warm stratocumulus) within an hour (4 to 5 h). The combination of larger numbers of ice crystals (up to 4 l⁻¹; Fig. 12) and higher ice supersaturations (10% vs.
6.5% in 5°C produces the larger ice crystal growth rates (compare ∂θ/∂t|_{fin} for 5°C, Fig. 6b, and 10°C, Fig. 13b). Again, these concentrations are on the high-side for Arctic conditions (e.g., Bigg, 1996).

Fig. 11. Time Series: LWC (shaded) and IWC (contoured) for 10CTRL in g m⁻³.

Fig. 12. Profiles of ice concentration \( \langle N_i \rangle \) for 10CTRL during the 4–5 h simulation period. Solid line denotes 4 h, long-dashed line denotes 4.25 h, and short-dashed line denotes 4.5 h.
Fig. 13. Profiles of: (a) radiative heating rate \( \frac{\partial \theta}{\partial t_{\text{Radiation}}} \), (b) microphysical heating rate \( \frac{\partial \theta}{\partial t_{\text{mic}}} \), and (c) \( \theta \) for 10CTRL. Solid line denotes 4 h, long-dashed line denotes 4.25 h, and short-dashed line denotes 4.5 h.

The efficient precipitation processes in the 10°C case, coupled with the large ice crystal growth rates, produces an overall heating of the cloud layer in time (\( \theta \), Fig. 13c). Radiative cooling at cloud top is too weak to offset the latent heat release through ice growth in this case. As a result, the upper portion of the cloud is stabilized (\( \theta \), Fig. 13c) rapidly through the growth of the ice crystals while the layers beneath cloud base are stabilized by the sublimation of precipitating ice which cools and moistens this region. This process effectively shuts off the generation of LWC due to cloud top radiative cooling. In conjunction with this effect, the radiatively-driven circulations cease by the 5th hour of the simulation.

Thus, the cooler sounding produced a liquid cloud that cannot maintain itself against the efficient ice precipitation mechanism which effectively stabilizes the layer and shuts off the production of cloud-scale circulations. This demonstrates the impact that cooling.
a representative atmosphere might have on the production of colloidally stable mixed-phase ASC. Pinto (1998) has postulated, based on observations taken during BASE, that cloud collapse due to complete glaciation may occur during autumn. Thus, colloidal instability may be one of the reasons for the reduction in stratus cloud amount that appear to occur over the Arctic during fall (Herman and Goody, 1976).

In the next sections, sensitivity simulations to the 5°C and 10°C cases are explored in order to ascertain the important factors that determine the colloidal stability of a mixed-phase ASC layer.

5. Sensitivity studies of mixed-phase ASC

Sensitivities were conducted to explore how microphysical parameters utilized by the ice microphysical model affect the colloidal stability of the cloud layer. Because of the intense computational needs of the ice bin microphysical model, the number of sensitivity studies was limited.

Table 1 lists the set of sensitivity simulations conducted with the bin microphysical model; the simulations of the previous sections are included as the control group (5°C is 5CTRL and 10°C is 10CTRL). Since it is likely that IFN concentrations \(N_{IFN}\), collection, sedimentation and ice crystal habit all contribute in some form to the colloidal stability of the simulated mixed-phase clouds, sensitivity experiments are designed around these processes. The sensitivity to collection is explored by disabling this process in both the 5°C and 10°C simulations (5NC and 10NC); thus neither aggregates nor graupel are produced by the model. The effects of precipitation are explored by disabling sedimentation in the 5°C case (5NC only the 10°C case only collapses faster). Ice forming nuclei concentrations are doubled in the 5°C simulation (5C2IN) in order to decrease the colloidal stability of the cloud layer while in the 10°C case, ice concentrations are reduced by half (10INH) and one-tenth (10INT). The impact

<table>
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<td>Tenth</td>
<td>Yes</td>
<td>Yes</td>
<td>Oblate</td>
</tr>
<tr>
<td>10HAB</td>
<td>Standard</td>
<td>Yes</td>
<td>Yes</td>
<td>Sphere</td>
</tr>
</tbody>
</table>

Listed is the acronym for the simulation and the physical processes utilized.

*Standard means the standard ice nucleation formulas are utilized; sensitivities either increase or decrease these amounts as specified.

*Oblate spheroids are the habit used for most simulations; the second habit option is that of spheres.
of reduced in-cloud residence times of the ice crystals is explored in the collapsing case (10°C) by using a faster precipitating ice habit, an ice sphere (10HAB).

This set of tests is by no means all-inclusive, however the goal is to suggest the importance of various microphysical processes in producing colloidally stable vs. unstable mixed-phase cloud layers. As we shall discover, IFN concentrations, and their removal, play an important role in the maintenance of these cloud systems.

5.1. Sensitivity to collision–coalescence

The collision–coalescence processes appears to have a small impact on the stability of the mixed-phase cloud layer in both the 5°C and 10°C cooled cases. A cursory comparison of the LWC/IWC time-series for 5NC (Fig. 14, only 2 h shown) with that of the control (5CTRL, Fig. 3), shows a qualitatively similar evolution. One might expect that, since large ice crystals are produced through aggregation, the larger ice precipitation flux out of the cloud may increase cloud stability. However, few aggregates are produced in these simulations of mixed-phase ASC (see Section 4.1) and ice crystals grow very efficiently to large sizes by deposition at these temperatures. Thus, ice aggregation is not extremely important for the production of precipitation-sized ice.

Without the effects of collection, note that the vertical extent of the liquid cloud is reduced as compared to 5CTRL by the end of the simulation. In addition, LWCs are higher in the 5NC simulation, suggesting that collision–coalescence actually reduces cloud stability. These effects are due primarily to the disabling of water drop self-collection. The suppression of the drizzle process causes the LWC profiles to be confined to a

![Fig. 14. Time series: LWC (shaded) and IWC (contoured) for 5NC in g m⁻³.](image)
shallower layer. Thus, more LWC and larger drop concentrations exist at cloud top and this increases the cloud top longwave cooling by 0.9 K h$^{-1}$. The stronger cooling rates produce greater downward buoyancy at cloud top, thus increasing the strength of the eddy circulations which allow for further condensation and, hence, larger LWCs. Little difference with the control is produced by 10NC and, therefore, it is not shown.

The results of this section suggest that, in cases where aggregation is not the dominant mechanism of producing large ice crystals, cloud stability may be enhanced in the presence of weak to no collection for both ice and water.

5.2. Sensitivity to sedimentation

The sedimentation of ice crystals from the liquid layer is the only mechanism whereby cloud collapse can be avoided. Thus, we might ask how quickly ice crystals can deplete the available liquid water once significant ice has been produced. This should give us an idea of the time-scale required for the removal of ice from the layer before cloud collapse occurs. By deactivating the ice sedimentation process, we can gain an upper estimate on this time scale.

As expected, treating the ice and water drops as passive tracers causes complete glaciation of the cloud (5NS) between 4.0 and 4.5 h of the simulation (the collapsing cloud is very similar to Fig. 11 and is, therefore, not shown). Thus, removal of ice crystals during the rapid glaciation phase is vital to the stability of the cloud layer. In addition, without ice sedimentation, the lower portions of the boundary layer are not thermally modified through sublimational cooling.

Other factors such as IFN concentrations will impact stability through modifying in-cloud residence times of the ice, a process we examine next.

5.3. Sensitivity to ice concentration: 5C2IN

In addition to the temperature differences of the clouds, 10CTRL has almost twice the number of ice crystals (4 l$^{-1}$ vs. about 2 l$^{-1}$) which impacts the total mass growth rate and the ice precipitation rates drastically. Increasing the number of IFN in the 5CTRL simulation illustrates the importance of IFN in the stabilization of the mixed-phase layer, thus $N_{IFN}$ is doubled in the 5°C simulation. This brings the maximum ice crystal concentration close to that of the 10°C control simulation (10CTRL) and quantifies the importance of IFN in mixed-phase cloud stability. The results not only elucidate the importance of IFN concentrations, but also shows that multiple liquid cloud layers may be formed through ice microphysical, radiative and dynamical interactions.

The larger ice concentrations increase in-cloud residence times by reducing ice crystal sizes which causes faster reductions in LWCs than 5CTRL (Fig. 15). By the fifth hour of the simulation all that remains is a thin, tenuous cloud layer of about 150 m depth. In addition, the doubling of $N_{IFN}$ from about 2 l$^{-1}$ (Fig. 4) to 4 l$^{-1}$, which is well within the range of observed Arctic $N_{IFN}$ (Curry et al., 1996), produces a lower liquid cloud layer. Since this layer is not a transient feature, its formation and maintenance warrants discussion (particularly because of the frequent layering observed in Arctic cloud systems).
5.3.1. Comparison with 5CTRL and 10CTRL

Even though cloud ice concentrations are similar to 10CTRL, ice production rates are smaller in 5C2IN because the equilibrium vapor pressure differences are greater in 10CTRL, which allows ice to grow faster at the expense of liquid drops.

As ice contents are increased through enhanced deposition rates, in-cloud water vapor amounts are reduced while $\theta$-values are increased over 5CTRL (Fig. 16a). Only near cloud top does radiative cooling reduce $\theta$ over the 4 to 5 h period (Fig. 16c and d), unlike 5CTRL where cooling over the depth of the cloudy layer occurred (Fig. 5b). The sub-cloud layers are cooled radiatively as the cloud becomes optically thin. In addition, strong ice precipitation causes rapid cooling and moistening of these layers (d$\theta$/dt, and $\theta$; Fig. 16c and a). This stabilization process is stronger than in 5CTRL, again showing the strong impact ice precipitation can have on the Arctic BL.

Circulation strengths ($\langle w'w' \rangle$) weaken as buoyant production of TKE is reduced concomitantly with radiative cooling (Fig. 17). By 5 h, the circulations are confined to a shallow layer near 1000 m; such shallow, coherent structures have recently been detected in thin, tenuous ASC (Andreas Reuter, personal communication). Lower level circulations are initiated by 4.5 h as cooling, moistening and ice mass loading increase buoyancy production of TKE. Once the ice has been removed, these circulations are maintained by radiative cooling at the top of the ice-produced humidity inversion (5 h line in Fig. 16d).

5.3.2. Multi-layer formation and maintenance

The cooling and moistening of the atmosphere between 200–400 m, along with the production of circulations through radiative cooling, causes the formation of a second
Fig. 16. Profiles of: (a) θ, (b) water vapor mixing-ratio ($r_v$), (c) total θ tendency ($d\theta/dt$), and (d) radiative heating rate ($\Delta\theta/\Delta t_{\text{Radiation}}$) for SC21N. Solid line denotes 4 h, long-dashed line denotes 4.5 h, and short-dashed line denotes 5 h.

liquid cloud between 5 and 5.5 h of model integration (Fig. 15). Once the majority of the ice mass precipitates out of the lower atmosphere, particularly the small ice crystals which have long in-cloud residence times, water saturations build up. Droplets are activated as vertical motions (Fig. 17) produce water supersaturations by about 5 h (as is shown by $n(D)$ in Fig. 18a). One can see from the increasing total drop concentration ($N_t$, Fig. 18b) with time, and the small numbers, that this is a tenuous cloud layer formed in a region of weak supersaturation. The low ice concentrations ($N_i$, Fig. 18c) reduce the depletion of liquid drops by the Bergeron–Findeisen process, thus allowing the water drops to persist.

Once the upper layer has thinned, and the lower layer has formed, significant θ-reductions occur within each cloud layer (Fig. 19). This cooling is caused by radiative flux divergence which strongly influences the total cooling. The high correlation
between the radiative cooling rates and $\theta$ reductions indicates that condensational heating must be small; this result is corroborated by the small increases in $N_i$ with time (Fig. 18). Since condensation is slow, both layers remain quite tenuous which keeps radiative cooling rates small. Circulations in the lower cloud are weak (Fig. 20, inset), which has a small impact on the supersaturation. In time, the lower liquid layer forms a cloud-top inversion through the strong differential in radiative cooling and takes on the character of a cloudy mixed-layer.

The slow changes in LWC and IWC for both the upper and lower decks suggests a balance between in-cloud production and precipitation removal. Indeed, the relatively constant $N_i$ profiles with time and the constant-with-height $N_i$ profiles corroborate this (Fig. 18). The fact that the drop distribution, $n(D)$, varies slowly in the lower cloud deck suggests that the microstructure here is also fairly constant. Ice crystals precipitating from the upper layers are dominated by large ice crystals (not shown) which is to the benefit of the lower liquid layer as these crystals quickly pass through with little impact. Such behavior (in-cloud production balanced by precipitation removal) appears to be characteristic of autumnal mixed-phase cloud layers, as suggested by the data of Pinto (1998).

The maintenance of the upper and lower deck, in the presence of the continued loss of water mass through precipitation, appears to be mostly due to radiative cooling which allows for further condensation and the maintenance of buoyancy production. As the upper deck attains larger water mass, circulations within the deck strengthen and deepen ($<w'w'>$, Fig. 20) in response to the stronger cooling rates and larger water mass loadings. The LWC of the lower deck does not change rapidly (Fig. 15), and is maintained by continued radiative cooling (Fig. 19).
Fig. 18. Profiles of: (a) drop distribution ($n(D)$), (b) drop concentration ($N$), and (c) ice concentration ($N_i$), for 5C2IN. Solid line denotes 5.5 h, long-dashed line denotes 6 h, and short-dashed line denotes 7 h. Times are similar for $n(D)$ except that the solid line is for 5 h.

It should be noted that the formation of this lower layer occurs above the surface inversion and, hence, is not associated with the region from which the original fog layer was removed. Indeed, it would be difficult for the above mechanism to form a cloud layer within the surface inversion. This inversion is strong enough to suppress turbulent motions and, thus, there is no dynamic support for the formation of a lower cloud layer.

This potential form of layering falls outside the classification regime defined by Curry et al. (1988) as the lower deck is formed by cooling and moistening through ice precipitation. The production of this lower layer is dependent upon the rapid glaciation of the upper cloud deck and reduction of its optical depth in time. Whether such glaciation periods occur within true mixed-phase ASC is not known, although this has been suggested by some observations (Curry et al., 1997). In addition to this, the results shown above illustrate that transitions from mixed-layers to stable-layers may occur.
through ice precipitation and sublimation, a result which is consistent with observations (e.g., Curry et al., 1997; Pinto, 1998).

5.4. Sensitivity to reduced concentrations: 10INH and 10INT

Since increasing ice concentrations decreases stability in the 5°C simulations, then reductions in ice concentrations should increase stability in the 10°C cases, which is examined in this section.
Reducing concentrations by half (10INH), so that ice concentration maxima match those of the 5CTRL simulation, still produces a collapsing cloudy layer (although this is more gradual). Differences in comparison to 10CTRL are not large and, therefore, are now shown. Since the Bergeron–Findeisen process is more efficient at these cooler temperatures, reducing concentrations to 5CTRL levels does not produce a colloidally stable layer. Thus, concentrations of IFN must be reduced further in colder cloud cases in order for self-maintaining cloud layers to be formed. Because of this, we reduce concentrations in the control by a factor of 10 (10INT) and, again, examine the cloud layer stability.

The reduction of ice concentrations by 1/10 the initialization values produces a mixed-phase system that is colloidally stable (Fig. 21). In this case, IWCs are produced more slowly than in 10CTRL (Fig. 11) and precipitate much more quickly from the liquid layer since fewer crystals compete for the available vapor. In fact, IWCs are less than half that produced in 10CTRL (Fig. 11) and are even slightly less than those produced in 5CTRL (Fig. 4). Behavior of the IWC layer over the 4 to 5 h time period is similar to that of 5CTRL (compare Figs. 21 and 3), however reductions in LWC are much lower owing to the small ice concentrations.

The rapid removal of large ice crystals, produced because of the low ice concentrations, from the upper layers of the cloud affects the water vapor ($r_v$) and $\theta$ profiles (Fig. 22a and b) differently than in 10CTRL (Fig. 13c). Because IWC production rates are smaller than 10CTRL, but similar to 5CTRL, net depositional heating does not offset cloud-top radiative cooling, allowing the cloud layer to cool in time, unlike 10CTRL. The weaker ice production rates, in conjunction with rapid sedimentation, causes less

![Fig. 21. Time series: LWC (shaded) and IWC (contoured) for 10INT in g m$^{-3}$.](image-url)
moistening and cooling of the sub-cloud layer ($\theta$ and $r_v$) than in the 5CTRL or 5C21N cases. The weaker stabilization of the lower BL, due to smaller ice production and greater precipitation rates, produces a boundary layer structure which is more reminiscent of a pure liquid phase mixed-layer. These results suggest a small range of IFN concentration ($\sim 0.4$ to $4 \text{ l}^{-1}$) over which IWC production can actually affect the cloudy layers (i.e., cause them to deviate significantly from the behavior of a pure liquid cloud) without causing cloud collapse due to colloidal instability.

In all of these cases, it is apparent that ice sedimentation rates are important for cloud layer stability. Ice habit plays a large role in determining ice terminal fall speeds and this is explored next.

5.5. Sensitivity to ice habit: 10HAB

Ice crystals have a broad spectrum of terminal fall speeds with dendrite crystals (which fall slowly) and ice spheres (which fall quickly) making up the two extremes (Pruppacher and Klett, 1997). Since oblate spheroids (used in 10CTRL) have fall speeds more closely associated with plate crystals (and, thus, have slower fall speeds), we choose to use ice spheres so that the two simulations, 10CTRL and 10HAB, will approximately cover the two extremes. Since ice sedimentation velocities are large in 10HAB, in-cloud residence times of ice crystals are reduced dramatically as ice crystals quickly sediment out of the liquid layer (Fig. 23). The initial thinning of the liquid layer is greater than in 5CTRL (Fig. 3), however enough LWC persists to drive cloud circulations through cloud top radiative cooling.

Ice crystals precipitate rapidly enough away from the cloud top region so that depositional heating does not dominate the total heat budget, thus cloud top cools in time through radiative effects ($\theta$; Fig. 24a). The weak deposition rates produce only slight reductions in water vapor contents ($r_v$; Fig. 24b) at cloud top. This contrasts the
10CTRL simulation in which rapid reductions in \( r_v \) and an increase in \( \theta \) occur because of the large deposition rates. Ice convergence (Fig. 23) at mid-cloud levels (700 to 900 m) increases deposition rates causing rapid LWC and \( r_v \) reductions and, thus, heating of the mid-cloud layer. The precipitation of large amounts of ice through the sub-cloud

Fig. 23. Time series: LWC (shaded) and IWC (contoured) for 10HAB in g m\(^{-3}\).

Fig. 24. Profiles of: (a) \( \theta \) and (b) water vapor mixing-ratio \( (r_v) \) for 10HAB. Solid line denotes 4 h, long-dashed line denotes 4.5 h, and short-dashed line denotes 5 h.
regions causes rapid cooling and moistening ($\theta$ and $r$,) which stabilizes the layer. This effect is not as strongly pronounced as it is in the $5^\circ$C cases since the ice spheres fall much more quickly. Since LWCs at cloud top are not completely depleted, radiative cooling continues to drive circulations allowing the persistence of the upper cloud deck.

Thus, if ice crystals with large sedimentation velocities are produced by the cloud layer, a cooler layer can accommodate larger ice concentrations (in this case, up to 10 times as much) and still retain its stability. The key to stability of the cloud layer is related to whether or not ice crystals within the vicinity of cloud top have deposition rates that are large enough to deplete the LWC there. If this is not the case, and as long as the LWC is sufficient to continue to drive some weak eddies (as in the 5C2IN simulation) through radiative cooling, the cloud layer may persist without collapse of the cloudy layer.

6. Concluding remarks

A detailed ice–liquid phase bin microphysical model (Reisin et al., 1996) was coupled in a consistent fashion to a radiative transfer scheme (Harrington, 1997) and a CRM version of the RAMS model (see Stevens et al., 1996). This modeling framework was used to carry out experiments on mixed-phase ASC which were designed around the June 28, 1980 ASC case (Curry et al., 1996). The soundings from that case were cooled in $5^\circ$C increments to produce hypothetical mixed-phase clouds. The validity of these hypothesized mixed-phase clouds was considered in light of current case study analysis from the BASE experiment (Pinto, 1998).

All simulated cloud systems undergo periods of rapid glaciation as small ice crystals with significant mass build up within the liquid cloud layer. The colloidal stability of the simulated cloud layer is then dependent upon the rapidity of ice crystal growth vs. the sedimentation velocity of the ice crystals. In cases with warmer cloud top temperatures, fewer IFN are activated which, in combination with a weaker Bergeron–Findeisen process, causes less depletion of the available LWC. Colder cloud temperatures require smaller numbers of IFN (or faster precipitating ice habits) in order to attain colloidal stability. Removal of IFN from the cloud layer appears to be of prime importance to the stability of the cloudy layer. Since there are few in situ IFN sources over the arctic ice pack (Bigg, 1996; Pinto, 1998), precipitation rapidly reduces the available surface area for ice vapor deposition. Thus, in time, fewer ice crystals are produced (as IFN numbers are reduced) and this invariably leads to stable mixed-phase cloudy layers. Pinto (1998) paper comes to similar conclusions about the importance of IFN sources and removal using data from the BASE experiment conducted in 1994. Additionally, a recent paper by Hobbs and Rangno (1998) has shown that such quasi-stable liquid topped clouds that precipitate ice are frequently observed over the Arctic Ocean.

The discussions presented above suggest a three-fold importance of ice concentrations. First, ice concentrations affect net depositional growth and, thus heating (cooling) and drying (moistening). Second, ice concentrations affect the production of large ice crystals and, therefore the vertical structure of the IWC. In-cloud residence times of IWC (which is also dependent upon ice habit) feeds into the importance of depositional
growth. Third, ice concentrations directly and indirectly affect cloud top radiative cooling rates by converting large numbers of droplets to lesser numbers of ice crystals (thus, reducing the integrated surface area) which have smaller projected areas than liquid drops.

These simulations also show the importance of ice precipitation in the moistening and cooling of the lower portions of the boundary layer. In time, colloidally stable clouds tended to deepen and thermodynamically stabilize their environment. This result is consistent with the autumnal boundary layer evolution observed by Curry et al. (1997) during BASE. Our simulations show that ice sedimentation and sublimation can rapidly (within a matter of hours) cause stabilization and moistening of the lower boundary layer. This suggests that the frequent stable regions found beneath autumnal mixed-phase ASC may be part of the microphysical evolution of the ASC system.

Our studies point to the possibility of layer formation through ice precipitation mechanisms. If ice sedimentation is rapid enough, and the cloud capping the boundary layer becomes optically thin through ice depositional processes, a lower cloud layer may form through the radiative cooling of a moisture inversion. In this case, the moisture excess and cooling was produced by ice sedimentation from the upper cloud layer. The radiative cooling of the lower layer caused droplet activation and multiple-layer formation. It appears possible that any mechanism (advection, open leads etc.) which could modify the lower boundary layer in the presence of a thin upper cloud layer might cause multiple layer formation.

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