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The role of ice nuclei recycling in the maintenance of cloud ice in Arctic mixed-phase stratocumulus

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Abstract

This study investigates the maintenance of cloud ice production in Arctic mixed phase stratocumulus in large-eddy simulations that include a prognostic ice nuclei (IN) formulation and a diurnal cycle. Balances derived from a mixed-layer model and phase analyses are used to provide insight into buffering mechanisms that maintain ice in these cloud systems. We find that for the case under investigation, IN recycling through subcloud sublimation considerably prolongs ice production over a multi-day integration. This effective source of IN to the cloud dominates over mixing sources from above or below the cloud-driven mixed layer. Competing feedbacks between dynamical mixing and recycling are found to slow the rate of ice lost from the mixed layer when a diurnal cycle is simulated. The results of this study have important implications for maintaining phase partitioning of cloud ice and liquid that determine the radiative forcing of Arctic mixed-phase clouds.

1 Introduction

Reliable climate projections require realistic simulations of Arctic cloud feedbacks. Of particular importance is accurately simulating Arctic mixed-phase stratocumuli (AMPS), which are ubiquitous and play an important role in regional climate due to their impact on the surface energy budget and atmospheric boundary layer structure through cloud-driven turbulence, radiative forcing, and precipitation (Curry et al., 1992; Walsh and Chapman, 1998; Intrieri et al., 2002; Shupe and Intrieri, 2004; Sedlar et al., 2011; Persson, 2012). For example, Bennartz et al. (2012) showed that the extreme melt events observed at Summit, Greenland in July 2012 would not have occurred without the surface radiative forcing produced by AMPS.

AMPS are characterized by a liquid cloud layer that precipitates ice crystals even at temperatures well below freezing (Hobbs and Rangno, 1998; Intrieri et al., 2002; McFarquhar et al., 2007). Radiative cooling near cloud top generates turbulence that

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maintains the liquid layer and forms an approximately well-mixed layer that extends as far as 500 m below cloud base. These cloud-driven mixed layers are frequently decoupled from the surface layer, limiting the impact of fluxes of heat, moisture, and aerosols on the cloud layer from below (Solomon et al., 2011; Shupe et al., 2013). However, unlike subtropical cloud-topped boundary layers, decoupled AMPS can persist for extended periods of time due to weak precipitation fluxes out of the mixed layer and relatively moist air entrained into the cloud layer at cloud top (Tjernström et al., 2004; Solomon et al., 2011, 2014; Sedlar et al., 2012).

AMPS are challenging to model due to uncertainties in ice microphysical processes that determine phase partitioning between ice and radiatively-important cloud liquid water (Sandvik et al., 2007; Tjernström et al., 2008; Klein et al., 2009, Karlsson and Svensson, 2011; Barton et al., 2012; Birch et al., 2012; de Boer et al., 2012), which drives turbulence that maintains the system. Phase partitioning depends upon the number, shape, and size of ice crystals, since these determine the efficiency of water vapor uptake by ice and hence the availability of water vapor for droplet formation (Chen and Lamb, 1994; Sheridan et al., 2009; Ervens et al., 2011; Hoose and Möhler, 2012).

Since temperatures in AMPS are too warm for homogenous ice nucleation, ice must form through heterogeneous nucleation. Aerosols with properties to serve as seeds for heterogeneous ice crystal formation are referred to as ice nuclei (IN). A number of different aerosols such as mineral dust, soot, sea salts, and bacteria have been observed to act as IN, all of which nucleate at different temperatures and supersaturation ranges (Levin and Yankofsky, 1983; DeMott, 1990; Möhler et al., 2006; Welti et al., 2009; Lüönd et al., 2010; Kanji et al., 2011; Broadley et al., 2012; Kulkarni et al., 2012; Pinti et al., 2012; Wise et al., 2012). In addition, observations indicate that nucleation properties are modified by aging and coating of aerosols (Möhler et al., 2005; Cziczo et al., 2009). Heterogeneous ice nucleation can occur by a number of modes; either in the presence of super-cooled droplets, when an aerosol comes into contact with a droplet (contact freezing), or is immersed in a droplet (immersion freezing), or by vapor deposition on IN (deposition freezing) (Pruppacher and Klett, 1997).

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IN can be entrained into the cloud-driven mixed layer through turbulent mixing from above and/or below. Recent studies indicate that entrainment alone cannot account for observed ice crystal number concentration (N_{NI}) (Fridlind et al., 2012), motivating the use of diagnostic formulations for ice formation to produce model simulations of AMPS with realistic phase partitioning (Ovchinnikov et al., 2011). While this modeling strategy constrains N_{NI} to be close to the measured values it eliminates the dynamical-microphysical feedbacks that regulate ice/liquid phase partitioning (Avramov et al., 2011).

Here we investigate a relatively unexplored source of ice production-recycling of ice nuclei in regions of ice subsaturation. AMPS frequently have ice-subsaturated air near the cloud-driven mixed-layer base where falling ice crystals can sublimate, leaving behind IN. Recycling was found to be significant in large eddy simulations of a single-layer stratocumulus observed during the D.O.E. Atmospheric Radiation Measurement Program's Mixed-Phase Arctic Cloud Experiment (M-PACE; Verlinde et al., 2007; Fan et al., 2009). AMPS observed during M-PACE formed due to a cold-air outbreak, where large fluxes of heat and moisture over the open ocean forced turbulent roll clouds that were coupled to the surface layer. This coupling with the surface layer prevented the identification of the role of dynamics internal to the cloud-driven mixed layer in maintaining phase-partitioning.

In this study we focus on the internal microphysics and dynamics of the cloud-driven mixed layer by investigating processes in an AMPS decoupled from surface sources of moisture, heat, and ice nuclei. We posit that recycling plays a significant role more generally since, for example, assuming an adiabatic vertical profile, a 650 m-deep mixed layer with a cloud-top temperature of $-16\,^{\circ}\mathrm{C}$ requires a water vapor mixing ratio of at least $1.7\,\mathrm{g\,kg^{-1}}$ at mixed-layer base to be saturated with respect to ice, i.e., in order for recycling to be a *negligible* source of ice nuclei in the mixed layer. This value is typically only seen in the Arctic between May–September (Serreze et al., 2012), while AMPS frequently occur outside of these months (Shupe et al., 2011).

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We examine the role of IN recycling in maintaining ice production using large eddy simulations of a springtime decoupled AMPS. Three simulations are analyzed; a "Control" with shortwave radiation turned off to compare with previous simulations of this case that use different IN formulations, "NoRecycle" with IN recycling turned off to identify the impact of recycling on the cloud life-time and phase partitioning, and "SW" with recycling and shortwave radiation turned on to identify the impact of realistic diurnal heating and cooling tendencies on the recycling process. This study builds on previous studies of this case, all of which exclude shortwave radiation (Avramov et al., 2011; Solomon et al., 2011, 2014), by including a prognostic equation for IN and a diurnal cycle. Within this modeling framework we investigate the relative roles of recycling and entrainment of IN in maintaining cloud ice production.

2 Case description

The case derives from observations of a persistent single-layer Arctic mixed-phase stratocumulus cloud observed near Barrow, AK on 8 April 2008 during the Indirect and Semi-Direct Aerosol Campaign (McFarquhar et al., 2011) (see Fig. 1). The adjacent Beaufort Sea was generally ice covered during this time, with significant areas of open water observed east of Barrow. A 4 K temperature inversion with inversion base at 1.05 km was observed via a radiosonde at 17:34 UTC; static stability was near neutral within the mixed layer overlaying a stable near-surface layer with static stability greater than 2 K km^{-1} below 500 m. The water vapor mixing ratio, q_v, decreased from 1.7 g kg^{-1} at the surface to 1.2 g kg^{-1} at cloud top, above which a secondary maximum of 1.6 g kg^{-1} was observed. Winds were east-southeasterly throughout the lowest 2 km.

Measurements from ground-based, vertically pointing, 35 GHz cloud radar, micropulse lidar, and dual-channel microwave radiometer at Barrow indicated a cloud layer extending into the inversion by 100 m, cloud base at 0.9 km, and cloud top at 1.15 km. Cloud ice water path (IWP), derived from cloud radar reflectivity measurements, varied from 20–120 g m^{-2} within 10 min of the sounding, with an uncertainty

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of up to a factor of 2 (Shupe et al., 2006). Concurrently liquid water path (LWP), derived from dual-channel microwave radiometer measurements, was 39–62 $\mathrm{g\,m^{-2}}$, with an uncertainty of 20–30 $\mathrm{g\,m^{-2}}$ (Turner et al., 2007).

Research flights were conducted by the National Research Council of Canada Convair-580 at 22:27–23:00 UTC on 8 April 2008 over the ocean northwest of Barrow (McFarquhar et al., 2011). Droplet concentrations measured by a Particle Measuring Systems Forward Scattering Spectrometer Probe varied between 100 and 200 $\mathrm{cm^{-3}}$. Ice crystal number concentrations measured by Stratton Park Engineering Company 2D-S and Particle Measuring Systems 2D-P optical array probes for sizes larger than 100 µm averaged 0.4 $\mathrm{L^{-1}}$. IN concentrations measured with the Texas A&M Continuous Flow Diffusion Chamber varied from 0.1 L to above 20 $\mathrm{L^{-1}}$. Ice crystal habit estimated using the automated habit classification procedure of Korolev and Sussman (2000) indicated primarily dendritic crystal habits.

3 Model description

We use the large-eddy simulation mode of the Advanced Research WRF model (WRFLES) Version 3.3.1 (Yamaguchi and Feingold, 2012) with the National Center for Atmospheric Research Community Atmospheric Model longwave radiation package (Collins et al., 2004), RRTMG shortwave package (Iacono et al., 2008), the Morrison two-moment microphysical scheme (Morrison et al., 2009), a Monin–Obukhov surface layer (Paulson, 1970), and a 1.5-order turbulent kinetic energy prediction scheme (Skamarock et al., 2008).

All model runs are initialized with winds, temperature, and water vapor from the 17:00 UTC, 8 April 2008 sounding at Barrow, AK (see Fig. 1). Divergence is assumed to be $2.5 \times 10^{-6}\,\mathrm{s^{-1}}$ below the temperature inversion and zero above, giving a linear increase in large-scale subsidence from zero at the surface to 2.7 $\mathrm{mm\,s^{-1}}$ at the base of the initial inversion (z = 1.1 km).

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All simulations are run on a domain of 3.2 × 3.2 × 1.8 km with a horizontal grid spacing of 50 m and vertical spacing of 10 m. The domain has 65(x)× 65(y) × 180(z) gridpoints and is periodic in both the x and y directions. The top of the domain is at 1.8 km, which is 0.7 km above cloud top in this case. The model time step is 0.75 s. The structure of the cloud layer is insensitive to changes in resolution and domain size. For example, tests run for Solomon et al. (2014) demonstrated that increasing the vertical and horizontal resolution by a factor of two resulted in an increase in LWP and IWP by 5 and 1 %, respectively, while increasing the domain size by a factor of two in both the x and y directions results in an increase in LWP and IWP of less than 1 %.

Cloud droplets are activated using resolved and subgrid vertical motion (Morrison and Pinto, 2005) and a log-normal aerosol size distribution (assumed to be ammonium bisulfate and 30 % insoluble by volume) to derive cloud condensation nuclei spectra following Abdul-Razzak and Ghan (2000). The aerosol accumulation mode is specified with concentrations of 165 cm^{-3}, modal diameter of 0.2 µm, and geometric SD of 1.4 µm, based on in situ ISDAC measurements. In this formulation, IN and cloud condensation nuclei are treated as separate species.

Temperature and moisture profiles are nudged to the initial profiles in the top 400 m of the domain with a time scale of 1 h. The model is initialized with winds, temperature, and water vapor similar to the Control integration from Solomon et al. (2014). Horizontal winds are nudged to the initial profiles at and above the initial inversion base with a timescale of 2 h. Initial temperature and subgrid turbulent kinetic energy (TKE) are perturbed below the top of the mixed layer with pseudo-random fluctuations with amplitude of ±0.1 K and 0.1 $m^2 s^{-2}$, respectively. The liquid layer is allowed to form in the absence of ice during the first hour of the integration to prevent potential glaciation during spinup.

The cloud-driven mixed layer is defined as the region where the liquid-ice water static energy is approximately constant with height. We define the boundaries of the mixed-layer top and base to occur where the slope of liquid-ice static energy exceeds 7×10^{-3}

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and 1×10^{-3} K m^{-1}, respectively. Cloud top and base are defined as the heights where cloud water mixing ratio, q_c, is equal to 1×10^{-4} g kg^{-1}.

Nested Weather Research and Forecasting (WRF) model simulations of this case performed with an inner grid at LES resolution (Solomon et al., 2011) demonstrate that moisture is provided to the cloud system by a total water inversion at cloud top and that the mixed layer does not extend to the surface, i.e., the mixed layer is largely decoupled from surface sources of moisture. In addition, the nested simulations indicate that cloud liquid water, q_c, is maintained within the temperature inversion by downgradient turbulent fluxes of water vapor mixing ratio q_v from above and direct condensation driven by radiative cooling. These processes cause at least 20 % of q_c to extend into the temperature inversion.

The Control integration is run with shortwave radiation turned off in order to compare with previous LES studies of this case (Avramov et al., 2011; Solomon et al., 2014). The results of Control are compared to two additional simulations; one with IN recycling turned off (hereafter "NoRecycle") and one with recycling and shortwave radiation both turned on (hereafter "SW"). SW is used to investigate how the diurnal cycle impacts IN recycling and ice formation. All runs use the same setup except SW has subsidence reduced by 30 % to keep the mixed-layer top from lowering appreciably because of smaller LWPs. This allows for direct comparisons of mixed layer structure and fluxes at the mixed layer boundaries. The first six hours of integration are not used in the analysis to allow for the spinup of cloud ice. Hours 6–40 are used for analysis of the Control and NoRecycle simulations and hours 16–76 are used for analysis of the SW simulation to allow for multiple diurnal cycles.

WRFLES has been modified to include a prognostic equation for IN number concentration (N_{IN}),

$$\frac{\partial N_{IN}}{\partial t} + \mathrm{ADV} + \mathrm{DIFF} = \frac{\delta N_{IN}}{\delta t}\Big|_{\mathrm{activation}} + \frac{\delta N_{IN}}{\delta t}\Big|_{\mathrm{sublimation}} \quad (1)$$

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where ADV represents advection and DIFF represents turbulent diffusion. Activation is also referred to as nucleation of ice and sublimation is also referred to as recycling of IN.

Here we adopt an empirical approach by initializing N_{IN} with an observationally based relationship expressing the number of available IN as a function of temperature in regions of water-saturation (DeMott et al., 2010),

$$N_{IN} = F \cdot 0.117 \exp(-0.125 \cdot (T - 273.2)) \quad (2)$$

where F is an empirically derived scale factor and T is temperature in Kelvin. Sixteen prognostic equations are integrated for N_{IN} in equally spaced temperature intervals with nucleation thresholds between −20.2 and −15.5° C (see Fig. 2). Therefore, additional IN become available for activation as the cloud layer cools. Initial N_{IN} concentrations are a function of the nucleation threshold temperatures and are independent of the in-situ temperature. The in-situ temperature in regions of water saturation determines how many IN are activated. To take deviations from the empirical derivation into account, IN are activated with 50 % efficiency (by multiplying the activation tendency in Eq. (1) by 0.5), however results are insensitive to this parameter (not shown). Due to the pristine dendritic nature of the observed crystals, ice shattering and aggregation are neglected in the simulations and sublimation returns one N_{IN} per crystal.

N_{IN} (in units of L^{-1}) integrated over the domain in each temperature bin k at time t is equal to

$$\overline{N}_{IN}(k,t) = \iiint N_{IN}(x,y,z,k,t) \; dx\,dy\,dz. \quad (3)$$

Upon sublimation, the modification of activation thresholds that can occur for previously nucleated IN, i.e. preactivation (Roberts and Hallett, 1967), is not considered and N_{IN} are returned to each bin k with weighting

$$W_k = \left[\overline{N}_{IN}(k,0) - \overline{N}_{IN}(k,t)\right] / \overline{N}_{IN}(k,0) \quad (4)$$

where W_k is normalized such that $\sum W_k = 1$. The W_k are recalculated each time step. In this way, IN are recycled preferentially to each of the 16 temperature bins from which they originated (Feingold et al., 1996).

The factor F in Eq. (2) is set to 4 for all simulations yielding an initial N_{IN} at every gridpoint equal to 5.8 L^{-1}, compared to 10 L^{-1} used in LES studies of the same case presented in Avramov et al. (2011). Supplement integrations were done to test for robustness of the results presented in Sect. 4 by varying initial IN concentrations, i.e., the factor F, (shown in Fig. 3) and by varying snow density and fall speeds (shown in Fig. 4). Figure 3 shows that the simulation maintains ice production when the initial N_{IN} is increased or decreased by ~ 3 L^{-1} relative to control. Figure 4 shows that the simulations maintain quasi-steady ice and liquid water paths after an initial spinup but the amount of ice produced is sensitive to the snow fall speed.

Crystal size distributions for averaged values of ice water mixing ratio and number concentration from the Control integration are shown in Fig. 5. These crystal size distributions are consistent with the Avramov et al. (2011) simulations of this case where crystal habits are assumed to be high density pristine dendrites. The distribution shown in Fig. 5 underestimates the number of large (greater than 5 mm) crystals as estimated by the 2D-S and 2D-P probes (see Avramov et al. (2011) for a detailed discussion of the measurements).

4 Model results

4.1 Control integration

In the quasi-steady control integration, the mixed-layer depth is approximately 850 m and comprises a 375 m deep liquid cloud layer (henceforth "the cloud layer"), extending above the mixed-layer top by 25 m, and a 500 m subcloud layer below (Fig. 6). IN are produced by sublimation of ice crystals below the cloud layer, advected to the cloud layer by turbulence, and activated as ice crystals (Fig. 6). Ice that forms in the cloud layer is transported vertically by turbulence, precipitates to cloud base and below, and sublimates below the cloud layer. At the mixed-layer base, an increase in N_{NI} due to precipitation approximately balances a decrease in N_{NI} due to sublimation. These processes constitute a feedback through which ice production and IN recycling are closely related. This feedback between ice production and IN in the mixed layer is linked to dynamic-thermodynamic tendencies, which sustain a subsaturated subcloud layer because the decrease in relative humidity due to an upward turbulent vapor flux exceeds the increase due to sublimation.

LWP and IWP remain steady until hour 16 of the simulation, and decrease slowly thereafter (solid lines in Fig. 7a). LWP and IWP magnitudes are within the observational estimates for this case. In addition, the cloud system is sustained over a multi-day period similar to measurements taken during ISDAC. Continuous cloud-top cooling causes the minimum horizontally-averaged temperature (near cloud top) to decrease from −17.5 to −20 °C from hour 10 to hour 40 (Fig. 7b).

Over the 40 h integration, the cloud layer remains decoupled from the surface (Fig. 7c). However, this does not prevent the number concentration of ice crystals (N_{NI}) in the cloud layer from remaining relatively steady, decreasing from vertically integrated values of 372 to 365 m L^{-1} (Fig. 7d, or in terms of vertically averaged cloud layer values, 1.2 to 1.1 L^{-1}). By contrast, while N_{NI} is maintained in the cloud layer, N_{IN} in the sub-cloud layer decreases significantly from 2 to 0.2 L^{-1} over the same period. Therefore, even though more N_{NI} are lost from the cloud than are activated (Fig. 8a), the relatively constant flux of IN into the cloud layer (Fig. 8b) allows N_{NI} in the cloud to decrease at a slower rate than N_{IN} in the subcloud layer. The continuous loss of N_{IN} in the subcloud layer is due to the IN flux into the cloud layer exceeding the N_{IN} gained through sublimation and turbulent advection at mixed-layer base (Fig. 8b). This loss is not mitigated by entrainment at mixed-layer top, which is found to be negligible (Fig. 8c), consistent with Fridlind et al. (2011).

The feedback loops discussed above are illustrated by the conceptual diagram in Fig. 9, where any change to one link in the cycle leads to an increase or decrease in ice production. For example, a decrease in the turbulent advection of N_{IN} into the cloud layer, slows the activation of IN, reduces the precipitation flux into the subcloud layer, reducing sublimation and availability of IN below cloud base. Both dynamics and thermodynamics play a role in the buffering aspect of these feedback loops since, for example, the slowing of IN activation in the example above would lead to increased cloud liquid production, cloud-top radiative cooling, and enhanced turbulent mixing, which would lead to increased transport of IN into the cloud layer and therefore increased activation of IN.

4.2 Impact of turning off recycling

When IN recycling is turned off, all IN that activate are lost from the system. This results in a more rapid loss of IN, a decrease in IWP, and a rapid increase in LWP (Fig. 7a and d, dashed lines), in contrast to the measurements that show a steady liquid layer and consistent ice production. These results demonstrate the importance of IN recycling in regulating phase partitioning. The rapid increase in LWP increases cloud-generated turbulence via enhanced radiative cooling and increases the turbulent mixing of IN from the subcloud layer into the cloud layer, contributing to a more rapid depletion of IN relative to the control integration. This process eventually becomes limited due to depletion of IN in the reservoir below (Fig. 8b). Due to the additional activation of IN as the cloud layer cools, ice production is maintained in the absence of recycling and the activation of IN in the cloud layer exceeds the upward IN flux at cloud base (Fig. 8a and b). However, the diminishing N_{IN} in the subcloud layer limits IN activation and N_{NI} rapidly decreases in the cloud layer (Fig. 7d).

4.3 Impact of diurnal cycle

A diurnal cycle is added to the control simulation in order to investigate how the feedback loops identified in the control and NoRecycle runs are modified with realistic transient heating and cooling tendencies due to variations in incoming shortwave radiation.

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A question that is addressed in this diurnal simulation is, to what extent is the continuous production of ice in the control simulation due to the lack of incoming shortwave radiation, which may overestimate the cooling tendencies in the cloud layer, resulting in an overestimate of IN activation? In addition, we investigate whether allowing for a realistic diurnal cycle provides for additional buffering feedbacks.

Adding a diurnal cycle to the control simulation produces a diurnal peak in downwelling surface shortwave radiation of 510 W m^{-2} and 6 h of total darkness per day (Fig. 10b). As shortwave radiation increases, the net radiative cooling near cloud top diminishes, which decreases cloud-generated turbulence, decreasing LWP and cloud-layer thickness. In addition, it is seen that the peak daily LWP coincides with zero shortwave radiation when in-cloud turbulence and cloud thickness are largest (Fig. 10a). These values are on the low end but within the measurements for this ISDAC case.

Figure 10a and b show that LWP and IWP variability is predominantly driven by the diurnal cycle. However, IWP variability is seen to lag LWP by 3–4 h because as shortwave radiation decreases the cloud layer cools, which increases activation of IN, increasing N_{NI}, allowing more ice crystals to grow, which increases IWP (Fig. 10a and b). Similar to the control simulation subcloud N_{IN} decreases at a faster rate than cloud layer N_{NI}, but allowing for the warming and cooling tendencies in the diurnal cycle results in cloud layer N_{NI} that decreases 40 % more slowly than in the control simulation (Fig. 10c).

Precipitation and turbulent mixing of N_{NI} (hereafter turbulent mixing is referred to as "T_{NI}") at cloud base are out of phase by 10 h (Fig. 10d), with turbulence leading precipitation. When shortwave radiation is weak or absent, the increase in N_{NI} eventually becomes limited by a decreasing turbulent mixing of IN ("T_{IN}") into the cloud layer from below, as recycling slows due to a decrease in N_{NI} flux from the cloud layer (Fig. 10d and f). When shortwave radiation is strong, reduction in IWP is limited by weaker precipitation losses, and attendant weaker sublimation and IN flux into the cloud layer (Fig. 10d and f). Entrainment of N_{IN} at the mixed-layer top is insignificant throughout the integration (Fig. 10e).

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5 Analysis from a mixed-layer perspective

The results discussed in Sect. 4 can be understood from balances in a well-mixed layer with sources/sinks at the upper and lower boundaries. Total particle concentration ($N_{IN} + N_{NI}$) is only changed by fluxes at the mixed-layer boundaries when recycling is allowed. These fluxes are entrainment of N_{IN} at mixed-layer top and turbulent mixing of both N_{NI} and N_{IN} (T_{NI} and T_{IN}) and precipitation of N_{NI}(P) at mixed-layer base. Since there are no sources and sinks of $N_{IN} + N_{NI}$ within the mixed layer, the horizontally-averaged $N_{IN} + N_{NI}$ flux ($f(z)$) must vary linearly from mixed-layer base to mixed-layer top (Lilly, 1968; Bretherton and Wyant, 1997). If it is assumed that f at the mixed-layer base is downward (assumed negative in this formulation) and f at the mixed-layer top is negligible (robust assumptions for a scenario where ice is precipitating from the mixed layer and entrainment is weak), then

$$f(z) = R \cdot \frac{H-z}{H-B}, \quad B \le z \le H \tag{5}$$

where H is the mixed-layer height, B is the mixed-layer base and R is the total $N_{IN} + N_{NI}$ flux at the mixed-layer base,

$$R = f|_{\text{Mixed-Layer Base}} = [P + T_{NI} + T_{IN}]_{\text{Mixed-Layer Base}}, \tag{6}$$

and

$$[T_{NI} + T_{IN}]_{\text{Cloud Base}} \approx [f - P]_{\text{Cloud Base}}. \tag{7}$$

Since $f < 0$, the turbulent flux of N_{IN} into the cloud layer plus the turbulent flux of N_{NI} into the subcloud layer is always less than precipitation of N_{NI} at cloud base. In addition, in a slowly evolving state where $T_{IN}|_{\text{Mixed-Layer Base}} > 0$, total IN flux due to sublimation in

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the mixed layer, S, can be written as

$$S \approx [P + T_{\mathrm{NI}}]_{\text{Mixed-Layer Base}} - [P + T_{\mathrm{NI}}]_{\text{Cloud Base}} \tag{8a}$$

$$\approx [f - T_{\mathrm{IN}}]_{\text{Mixed-Layer Base}} - [f - T_{\mathrm{IN}}]_{\text{Cloud Base}} \tag{8b}$$

$$< T_{\mathrm{IN}}|_{\text{Cloud Base}} - T_{\mathrm{IN}}|_{\text{Mixed-Layer Base}} \tag{8c}$$

$$< T_{\mathrm{IN}}|_{\text{Cloud Base}}. \tag{8d}$$

Thus in a well-mixed layer with an upward $T_{\mathrm{IN}}|_{\text{Mixed-Layer Base}}$, sublimation is always less than the flux of N_{IN} into the cloud layer.

Based on results from control, precipitation of N_{NI} at cloud base is sufficient to balance the upward turbulent flux of N_{IN} (i.e., $|T_{\mathrm{IN}}| \gg |T_{\mathrm{NI}}|$ at cloud base). Therefore, in a well-mixed layer with precipitation of N_{NI} at the mixed-layer base that is larger in magnitude than an upward turbulent N_{IN} flux at the mixed-layer base, and assuming negligible entrainment at the mixed-layer top

$$|P|_{\text{Cloud Base}} > T_{\mathrm{IN}}|_{\text{Cloud Base}} > S. \tag{9}$$

However, if all N_{NI} sublimate in the mixed layer and the upward turbulent flux of N_{IN} dominates at the mixed-layer base then $f > 0$ and

$$T_{\mathrm{IN}}|_{\text{Cloud Base}} > |P|_{\text{Cloud Base}} = S, \tag{10}$$

the mixed layer gains $N_{\mathrm{IN}} + N_{\mathrm{NI}}$ over time, resulting in a continuously increasing ice production in the cloud layer. In SW, $T_{\mathrm{IN}}|_{\text{Cloud Base}}$ is also greater than $|P|_{\text{Cloud Base}}$ after a period of weakened turbulence and weaker precipitation at the mixed-layer base, due to increased activation of N_{IN} due to decreasing shortwave radiation.

If IN entrainment at the mixed-layer top is not negligible then $f(z)$ must be modified to include fluxes at the mixed-layer top and $|f|_{\text{Cloud Base}}$ will increase. If $|f|_{\text{Cloud Base}}$ increases such that $f_{\text{Cloud Base}} < P_{\text{Mixed-Layer Base}}$, then sublimation will exceed $T_{\mathrm{IN}}|_{\text{Cloud Base}}$.

6 Analysis of buffered feedbacks in SW

Phase diagrams highlight the processes involved in ice production when a diurnal cycle is allowed (following the arrows from green to blue to black to red in Figs. 11a and b).

When incoming shortwave radiation is a maximum, recycling (sublimation) is seen to be at a minimum. This is counterintuitive since subcloud relative humidity is low at this time, which would be expected to produce increased sublimation. However, due to weak turbulent mixing between the cloud and subcloud layers the net N_{NI} flux into the subcloud layer is weak, resulting in weak sublimation and recycling. This situation is reversed as shortwave radiation decreases, since increased cloud-top cooling increases cloud-driven turbulent mixing, which allows recycling to increase in the regions of reduced subcloud relative humidity. As is seen in the conceptual diagram (Fig. 9), this then leads to an increased N_{NI} flux into the subcloud layer (green arrows, Fig. 11). However, N_{NI} in the cloud layer does not begin to increase until activation in the cloud layer exceeds the flux of N_{NI} into the subcloud layer (green arrows). This cycle is further amplified as shortwave radiation decreases, namely, decreased shortwave radiation increases cloud-driven turbulence, increasing the flux of IN into the cloud layer, increasing the activation of IN, which increases N_{NI} in the cloud layer and the N_{NI} flux from the cloud layer into the subcloud layer (blue arrows).

When incoming shortwave radiation is a minimum, more N_{IN} are activated because the cloud layer cools. However, again we see that N_{NI} tendencies due to thermodynamics are buffered by the slowing of turbulence-driven feedbacks due to a thickening of the cloud layer. Thus, a net increase in N_{NI} in the cloud layer, commensurate with an increased IWP and precipitation (black arrows), is buffered by a decrease in the downward turbulent mixing of N_{NI}, which reduces recycling, slowing the feedback loop (see Fig. 9). During the morning hours, as the cloud layer warms and thins and ice activation becomes less efficient, turbulence continues to decline, slowing the recycling feedback process to the point where limited IN fluxes to the cloud layer inhibit ice production and N_{NI} declines (red arrows).

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7 Summary

We have demonstrated that sustained recycling of IN through a drying subcloud layer and additional activation of N_{IN} due to a cooling cloud layer are sufficient to maintain ice production, and regulate liquid production over multiple days in a decoupled AMPS.

This study provides an idealized framework to understand feedbacks between dynamics and microphysics that maintain phase-partitioning in AMPS. In addition, we have shown that modulation of the cooling of the cloud layer and the humidity of the subcloud layer by the diurnal cycle buffers the mixed-layer system from a loss of particles. The results of this study provide insight into the mechanisms and feedbacks that may maintain cloud ice in AMPS even when entrainment of IN at the mixed-layer boundaries is weak. While the balance of these processes changes depending upon the specific conditions of the cloud layer, for example whether the cloud layer is coupled to the surface layer, the mechanisms detailed in this paper will manifest to some degree and therefore the current study provides a framework for understanding the role of recycling in maintaining phase-partitioning in AMPS.

Author contributions. A. Solomon, G. Feingold, and M. D. Shupe conceived and designed the experiments; A. Solomon performed the simulations; A. Solomon, G. Feingold, and M. D. Shupe analyzed the model results and co-wrote the paper.

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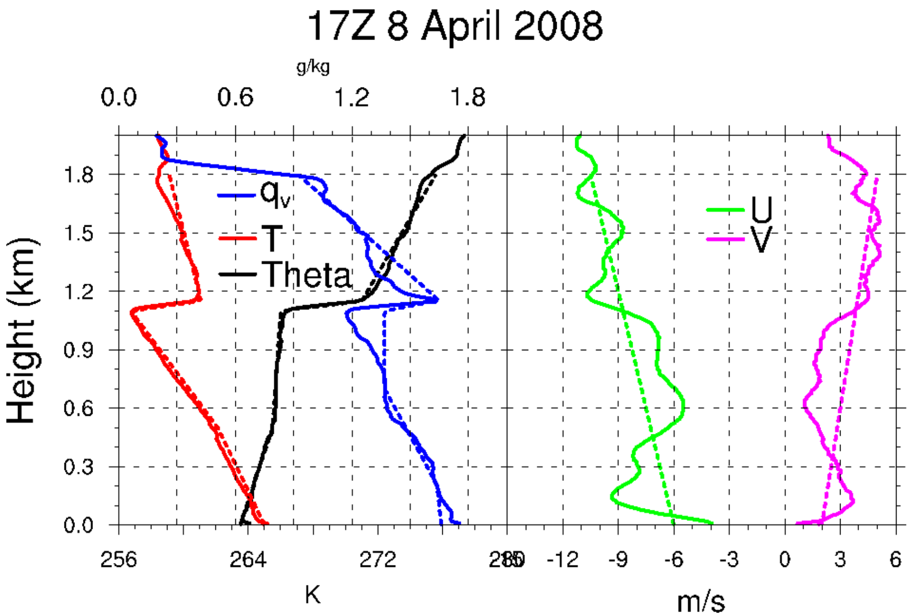

Figure 1. Sounding measured at 17:34 UTC, 8 April 2008 at Barrow, Alaska (71.338° N, 156.68° W). (Left) Water vapor mixing ratio (q_v), temperature T, and potential temperature (Theta), in units of $\mathrm{g\,kg^{-1}}$, degrees Kelvin, and degrees Kelvin respectively. (Right) Zonal wind U and meridional wind V, in units of $\mathrm{m\,s^{-1}}$. Gray shading marks the extent of the cloud layer. The dashed lines show the initial profiles used in the WRFLES experiments. The dashed line overlaying water vapor mixing ratio is the initial profile for the total water mixing ratio.

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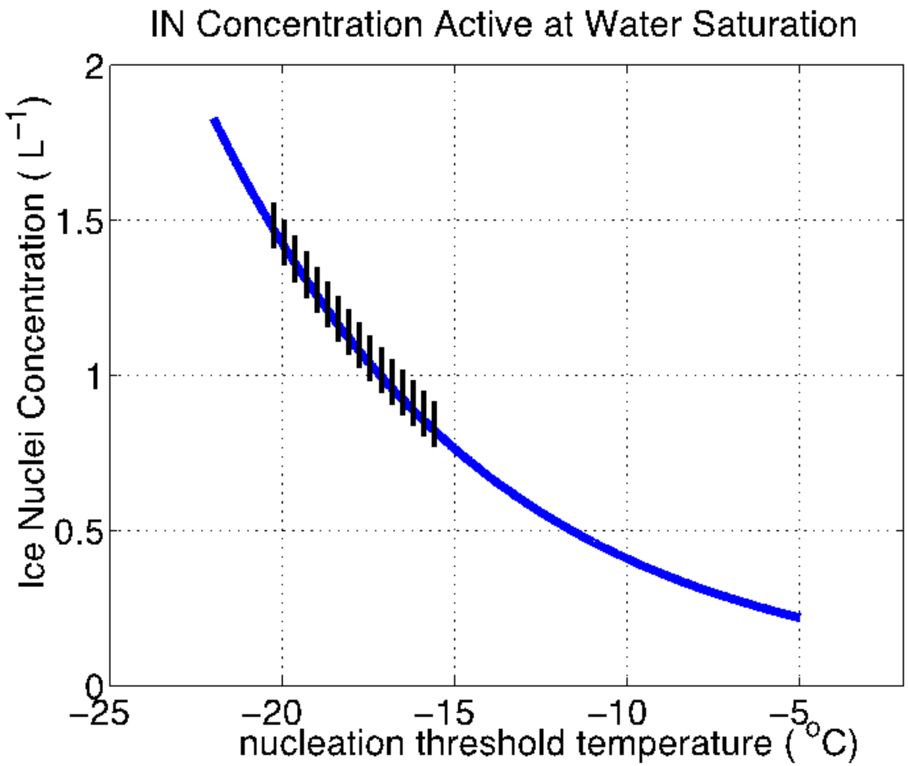

Figure 2. IN number concentration active at water saturation vs. temperature based on the empirical relationship derived in DeMott et al. (2010) (blue line). Black vertical lines indicate threshold temperatures for nucleation in the 16 IN bins. IN increments between lines indicate the additional IN available for nucleation at colder temperatures.

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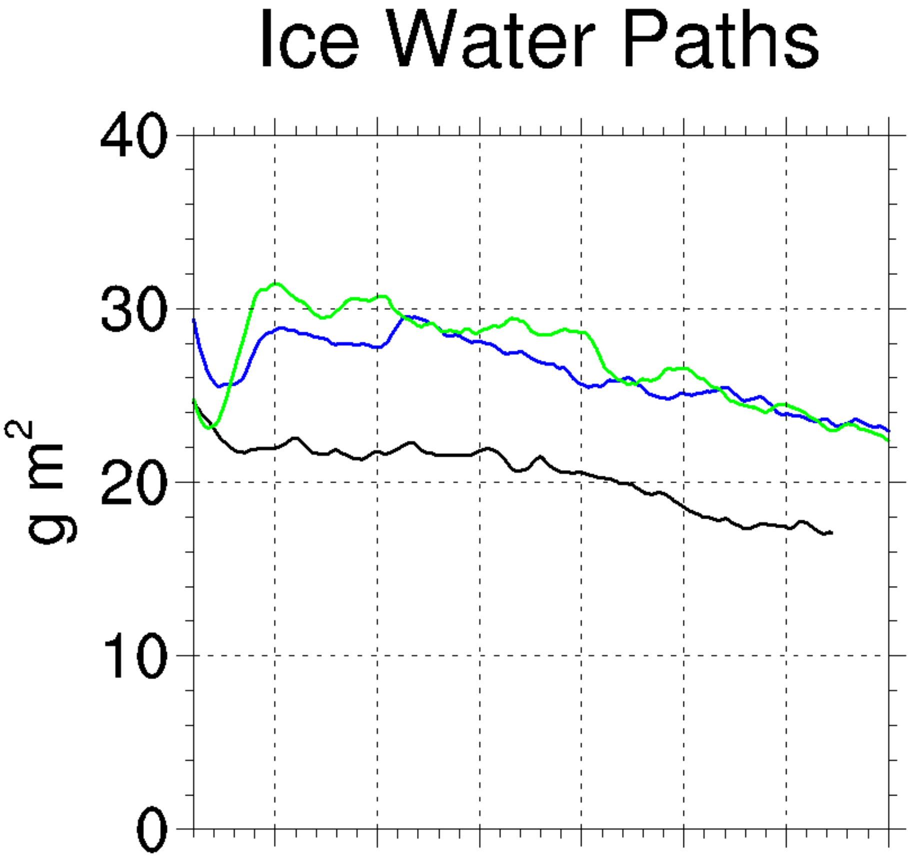

Figure 3. Sensitivity of ice water path to the parameter F in Eq. (2). Note the similar ice water paths for $F = 4$ and $F = 6$ (total N_{IN} initial values of 5.8 and 8.7 L^{-1}, respectively).

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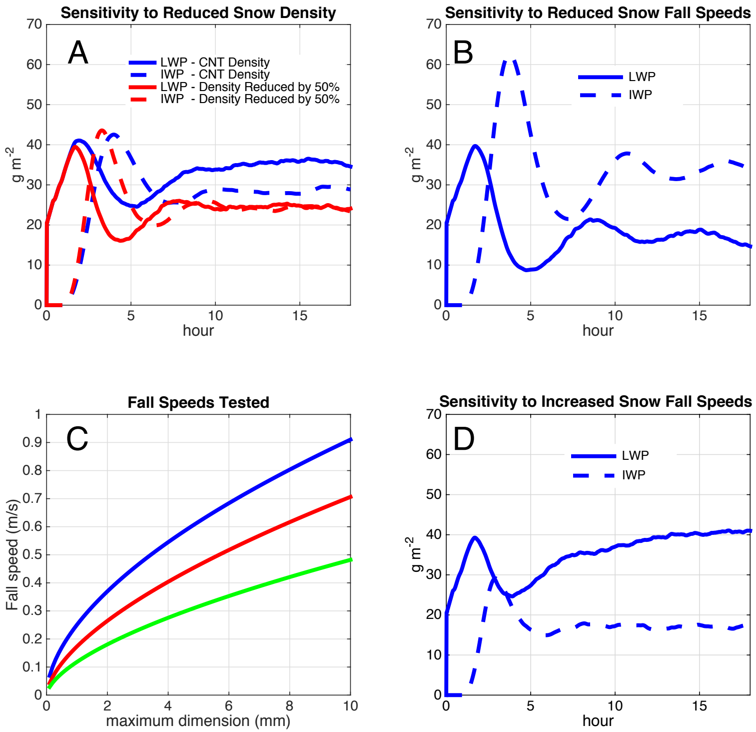

Figure 4. Sensitivity of LWP and IWP to snow density and fall speeds. LWP shown with solid lines in (**a**, **b** and **d**) and IWP shown with dashed lines in (**a**, **b** and **d**), in units of $\mathrm{g\,m^{-2}}$. Fall speeds used in sensitivity studies shown in **(b)**, in units of $\mathrm{m\,s^{-1}}$. **(a)** Sensitivity to reducing snow density from $100\,\mathrm{kg\,m^{-3}}$ (the control value, blue) to $50\,\mathrm{kg\,m^{-3}}$ (red) using control fall speeds (red line in **c**). **(b)** Sensitivity to reducing snow fall speeds (green line in **c**) using control snow density (compare to blue lines in **a**). **(d)** Sensitivity to increasing snow fall speeds (blue line in **c**) using control snow density (compare to blue lines in **a**).

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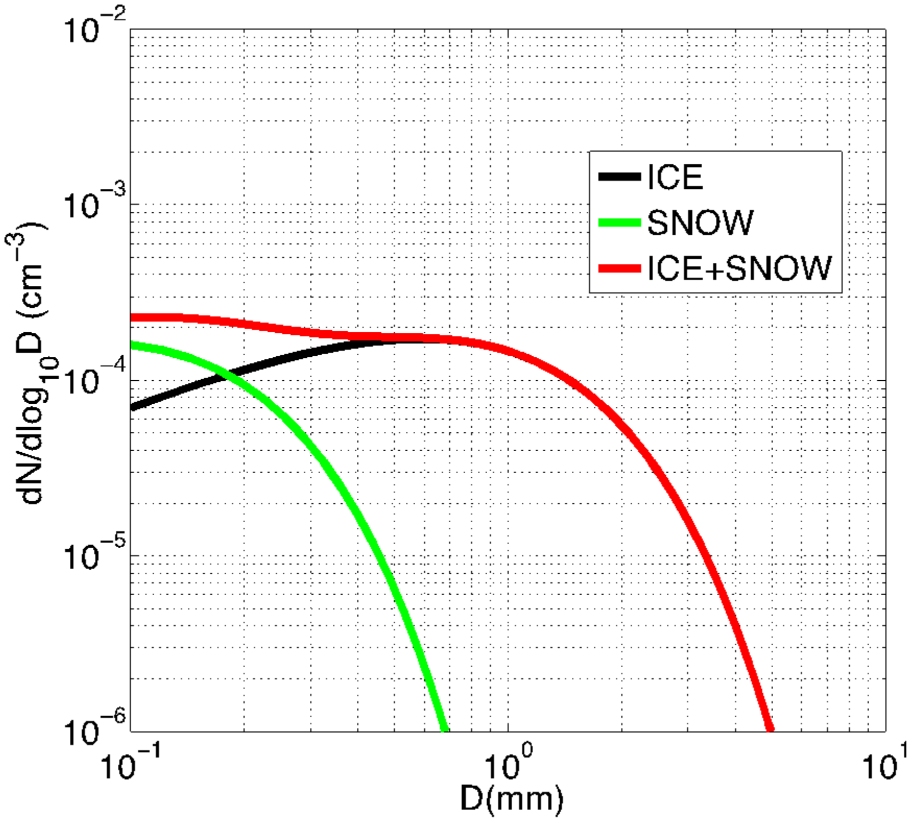

Figure 5. Simulated ice particle number size distributions using in-cloud mass and number concentrations. Ice water mixing ratio = $3 \times 10^{-4}\,g\,kg^{-1}$, ice number concentration = $0.4\,L^{-1}$, snow water mixing ratio = $2.4 \times 10^{-2}\,g\,kg^{-1}$, snow number concentration = $0.45\,L^{-1}$.

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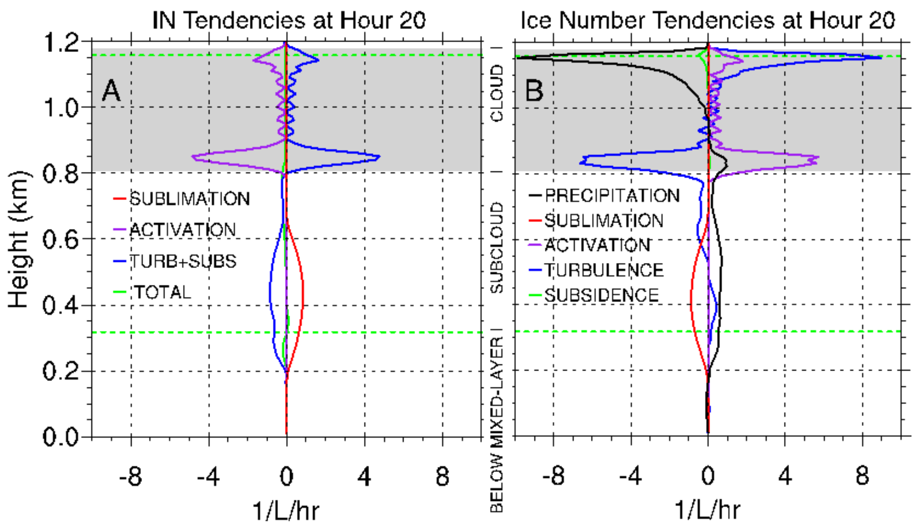

Figure 6. (a) N_{IN} and **(b)** N_{NI} averaged over 0.5 h at hour 20, in units of $L^{-1}\ h^{-1}$. Grey shading indicates the extent of the cloud layer. Green dash lines indicate the top and bottom of the mixed layer.

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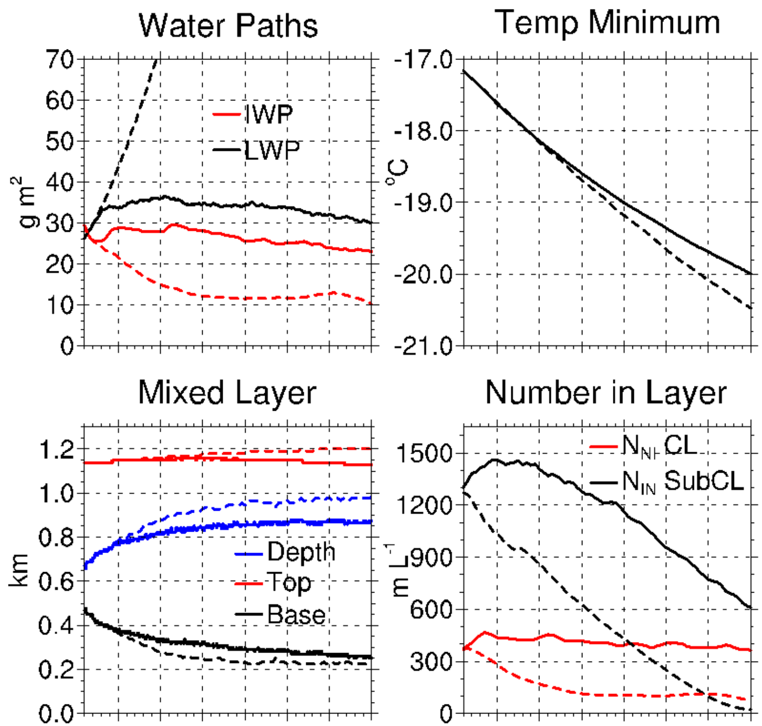

Figure 7. Control and NoRecycle time series for hours 6–40 (smoothed with 90 min running average). NoRecycle shown with red and black dashed lines. **(a)** LWP (black) and IWP (red), in units of $\mathrm{g\,m^{-2}}$. **(b)** Minimum horizontally-averaged temperature, in units of °C. **(c)** Mixed-layer depth (blue), top height (red), and base height (black), in units of km. **(d)** N_{NI} integrated over cloud layer (referred to as CL, red) and N_{IN} integrated over subcloud layer (referred to as SubCL, black), in units of $\mathrm{m\,L^{-1}}$.

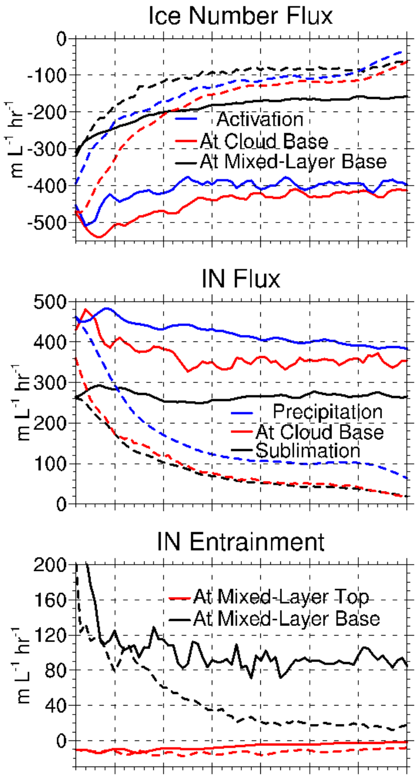

Figure 8. Horizontally-averaged fluxes from control and NoRecycle integrations for hours 6–40 (smoothed with 90 min running average). NoRecycle shown with dashed lines. **(a)** N_{NI} flux at cloud base due to turbulence + subsidence + precipitation (red), mixed-layer base due to turbulence + subsidence + precipitation (black), and due to activation (multiplied by −1, blue), in units of m L^{-1} h^{-1}. **(b)** N_{IN} flux at cloud base due to turbulence (red), N_{IN} flux due to sublimation (black), and precipitation of N_{NI} at cloud base (multiplied by −1, blue), in units of m L^{-1} h^{-1}. **(c)** N_{IN} entrainment at mixed-layer top (red) and base (black), in units of m L^{-1} h^{-1}.

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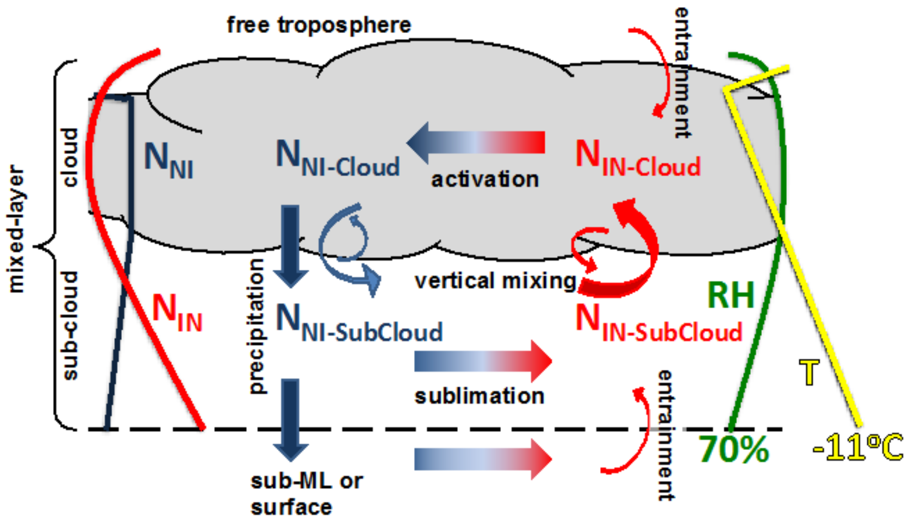

Figure 9. Schematic of feedback loops that maintain ice production and the phase-partitioning between cloud liquid and ice in AMPS when recycling is allowed. Red colors denote N_{IN}. Blue colors denote N_{NI}. Vertical profiles of N_{NI}, N_{IN}, relative humidity, and temperature shown with thin blue, red, green, and yellow lines, respectively.

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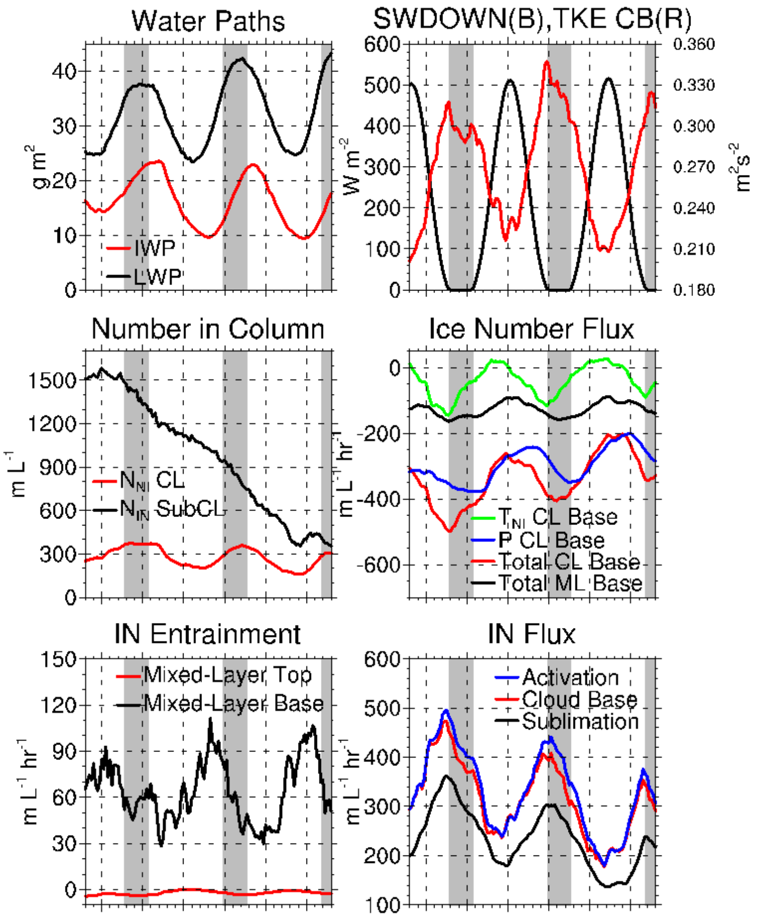

Figure 10. SW time series (see figure captions).

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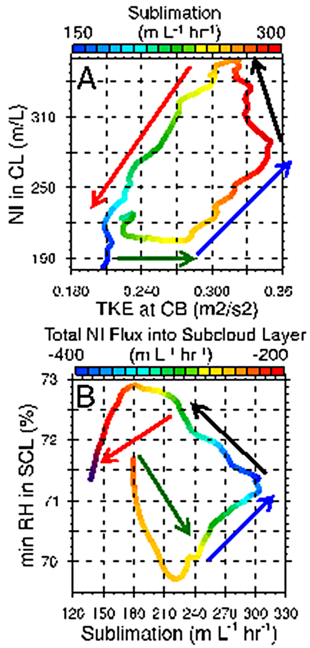

Figure 11. (a) Phase diagram of TKE at cloud base vs. NI in the cloud layer starting at peak shortwave hour 40, in units of m L^{-1} and m L^{-1} h^{-1}, respectively. Colors show sublimation in units of m L^{-1} h^{-1}. **(b)** 24 h phase diagrams of sublimation vs. minimum relative humidity in the subcloud layer starting at peak shortwave hour 40, in units of m L^{-1} h^{-1} and %, respectively. Colors show total N_{NI} flux at cloud base, m L^{-1} h^{-1}. Hours 42–47, 47–50, 50–56, and 57–62 indicated with green, blue, black, red arrows, respectively. Minimum shortwave indicated with the moon symbol. Maximum shortwave indicated with the sun symbol.

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