# CAN ICE-NUCLEATING AEROSOLS AFFECT ARCTIC SEASONAL CLIMATE?

BY ANTHONY J. PRENNI, JERRY Y. HARRINGTON, MICHAEL TJERNSTRÖM, PAUL J. DEMOTT, ALEXANDER AVRAMOV, CHARLES N. LONG, SONIA M. KREIDENWEIS, PETER Q. OLSSON, AND JOHANNES VERLINDE

The inability of regional models and global climate models to reproduce Arctic clouds and the Arctic radiation budget may be due to inadequate parameterizations of ice nuclei.

The Arctic near-surface warming since the late 1960s is approximately twice that of the global average (MacBean 2004; Serreze and Francis 2006). This trend is projected to proceed through this century (Kattsov and Källén 2004). Despite much attention, there is no consensus regarding the underlying reasons for this enhanced climate sensitivity. While many suggestions have been proposed, and some are certainly relevant, we are still unable to reconcile the hypotheses with observations and reanalyses. This reflects a lack of understanding of the Arctic climate system, which is at least partly caused by a paucity of observational data. Despite constant improvement, global climate models reflect this deficiency, with their greater difficulty in reproducing the current climate in the Arctic than elsewhere (Walsh et al. 2002). Further, the scatter between projections from different climate models is much larger in the Arctic than for other regions. For example, the projected summer ice cover toward the end of this century ranges from almost the same as today to a perennially ice-free Arctic (Walsh 2004). The intermodel scatter in the projected near-surface temperature is also an order of magnitude larger in the Arctic than farther south (Holland and Bitz 2003). Much of this uncertainty is related to the inability of the models to describe accurately many of the physical processes and feedbacks involved, some of which appear to be exclusive to the Arctic.

Clouds play an important role in the Arctic climate, but are not well represented by current models (Curry et al. 1996). Clouds determine the net longwave radiation at the surface and also regulate incoming solar radiation in summer. Low-level boundary layer (BL) clouds tend to dominate in the Arctic, with very high

**AFFILIATIONS:** PRENNI, DEMOTT, AND KREIDENWEIS—Department of Atmospheric Science, Colorado State University, Fort Collins, Colorado; HARRINGTON, ARRAMOV, AND VERLINDE—Department of Meteorology, The Pennsylvania State University, State College, Pennsylvania; TJERNSTRÖM—Department of Meteorology, Stockholm University, Stockholm, Sweden; LONG—Pacific Northwest National Laboratory, Richland, Washington; OLSSON—Alaska Experimental Forecast Facility, University of Alaska, Anchorage, Alaska
**CORRESPONDING AUTHOR**: Anthony J. Prenni, Campus Delivery 1371, Department of Atmospheric Science, Colorado State University, Fort Collins, CO 80523-1371
E-mail: prenni@lamar.colostate.edu

*The abstract for this article can be found in this issue, following the table of contents.*

DOI:10.1175/BAMS-88-4-541

In final form 8 November 2006
©2007 American Meteorological Society

temporal frequencies in all seasons (Curry et al. 1996; Intrieri et al. 2002) and somewhat uniform spatial distributions (Vavrus 2004). Arctic boundary layer clouds are often characterized by the presence of persistent temperature inversions, humidity inversions, and strong, stably stratified layers (Pinto 1998). A particularly important feature is that these clouds are often mixed phase even at quite low temperatures, consisting of liquid water tops that precipitate ice (Pinto 1998). To date, successfully modeling these clouds and their phase state has proven challenging (Beesley and Moritz 1999; Harrington and Olsson 2001a; Harrington et al. 1999; Morrison and Pinto 2005; Morrison et al. 2003; Vavrus 2004). Accurate modeling of cloud water phase is particularly important for the radiation balance at the surface. Liquid water clouds are optically thicker than ice clouds, and therefore increase the longwave radiation at the surface. Cloud liquid water and ice are also tied to cloud-scale dynamics, sea ice coverage and thickness, and climate (Curry and Ebert 1990; Curry et al. 1996; Harrington and Olsson 2001b; Jiang et al. 2000; Vavrus 2004), which adds to the modeling challenges. Such difficulties are increased further given that the temperatures at which ice crystals have been observed to form (Curry et al. 1996), and their concentrations (Hobbs and Rangno 1998), cover a large range in lower-tropospheric Arctic clouds, with liquid water detected at temperatures as low as –34°C (Intrieri et al. 2002). There is also a potential feedback, where enhanced winter warming, primarily due to an increase of greenhouse gases, increases the fraction of liquid water in these clouds, thereby increasing the longwave radiation at the surface and enhancing warming.

Some of the model difficulties with Arctic clouds may result from inaccuracies in ice nucleation parameterizations. Primary nucleation by specific aerosol particles known as ice nuclei (IN) is responsible for initial ice formation in mixed-phase stratus clouds. After initiation, secondary ice production, from cloud ice and liquid particle interactions, has been observed in Arctic clouds at temperatures between –2.5° and -8°C (Hobbs and Rangno 1998; Rangno and Hobbs 2001). Detailed cloud-resolving model (CRM) studies have suggested that mixed-phase Arctic clouds are very sensitive to modest changes in IN concentrations; this appears to be the case whether the clouds either exist over the pack ice (Harrington et al. 1999; Jiang et al. 2000) or are strongly forced by surface fluxes (Harrington and Olsson 2001a). In all of these cases, simulations in which IN concentrations are increased to 2–3 times the base-case values can transform a largely liquid stratus deck of wide areal coverage into a broken, optically thin cloud system. This occurs largely because of the lower vapor pressure of ice compared to that of liquid water, which causes ice crystals to grow and precipitate at the expense of liquid drops (the so-called Bergeron–Findeisen process). Freezing of cloud water onto ice particles amplifies this process. An active Bergeron–Findeisen process can produce rapid ice precipitation that dries, and sometimes dissipates, the cloud layer. In contrast, the absence of IN can lengthen the cloud lifetime and invigorate cloud circulations.

Arctic IN concentrations generally are reported to be much lower than those found at lower latitudes (e.g., Bigg 1996). Fountain and Ohtake (1985) measured mean IN concentrations of 0.17 $L^{-1}$ for filter samples collected from August through April and processed in a diffusion chamber at –20°C at a surface site in Barrow, Alaska. In contrast, Cooper (1986) cites typical average values of ice crystal concentrations from primary processes in midlatitudes to be 5 $L^{-1}$ at –20°C, decreasing to 0.1 $L^{-1}$ at –10°C. Most models use ice nucleation parameterizations based on a selection of IN data collected at the surface in the midlatitudes. These parameterizations equate to at least the upper end of the cloud ice crystal concentration values listed by Cooper (1986), nucleating, on average, ice concentrations greater than or at the upper limit of the observed Arctic values. Detailed CRM studies (Harrington et al. 1999) have suggested that using such parameterizations for the Arctic leads to the rapid depletion of liquid in modeled mixed-phase clouds, which in turn alters the cloud's lifetime and radiative properties.

An additional consideration for the Arctic is the seasonal cycle of aerosol concentration, resulting from transport from the midlatitudes from about December to April (Barrie 1986). This "Arctic haze" has been tied to variations in IN concentrations (Borys 1989), suggesting that IN concentrations also may exhibit seasonable variability. Rogers et al. (2001a) measured springtime Arctic IN concentrations; average IN concentrations were ~10 $L^{-1}$ and showed spatial variability that covered five orders of magnitude (Rogers et al. 2001a).

Here, we present IN measurements collected in the vicinity of Arctic clouds during the fall using the same instrument as that in the Rogers et al. (2001a) study. These data are used as input into a detailed cloud-resolving model for a specific mixed-phase cloud case encountered during the study period to determine the potential role of IN in affecting Arctic cloudiness and climate. Based on this case study, we conclude by exploring possible implications for climate modeling in the Arctic.

**CURRENT CLIMATE MODELS.** The ability of current models to simulate mixed-phase clouds in the Arctic and their effects on climate has recently been explored as part of the Arctic Climate Model Intercomparison Project (ARCMIP; Curry and Lynch 2002). In ARCMIP, six regional models were compared to observations from the Surface Heat Budget of the Arctic Ocean (SHEBA) experiment (Uttal et al. 2002). Tjernström et al. (2005) and Rinke et al. (2006) provide extensive evaluations of these following six models: Arctic Regional Climate System Model (ARCSyM; Lynch et al. 1995), Coupled Ocean–Atmosphere Mesoscale Prediction System (COAMPS; Hodur 1997), High-Resolution Limited Area Model [HIRLAM; with physics from ECHAM4, a GCM based on European Centre for Medium-Range Weather Forecasts (ECMWF) forecast models modified and extended in Hamburg; Christensen et al. 1996)], polar version of the fifth-generation Pennsylvania State University (PSU)–National Center for Atmospheric Research (NCAR) Mesoscale Model (Polar MM5; Cassano et al. 2001), Regional Model from the Max Planck Institute (REMO; Jacob 2001), and Rossby Centre Atmospheric Model (RCA; Jones et al. 2004). Some of these models were developed as regional weather forecast models (e.g., COAMPS), and some as regional climate models (e.g., ARCSyM), while some are regional models but with the same model physics as that of a global climate model (e.g., HIRHAM and RCA). All models were run for over a full year. To facilitate intermodel comparisons, all model simulations were set up on the same model domain with the same horizontal resolution, using identical forcing at the lateral boundaries. Sea and ice surface temperature were also prescribed to be the same for all models, taken from Advanced Very High Resolution Radiometer satellite measurements, while ice fraction was prescribed from Special Sensor Microwave Imager (SSM/I) satellite observations. Surface temperatures over land were derived independently for each model from internal surface energy balance considerations.

Figure 1 shows some results from this effort, comparing the vertically integrated cloud liquid water (liquid water path; LWP) from the ARCMIP models, with measured values taken during the SHEBA experiment. Measured LWP is taken from microwave radiometer measurements, with an uncertainty of 25 g m$^{-2}$ (Westwater et al. 2001). During the summer months, all of the models predicted liquid water paths that were similar to those observed, although differences in detail occur. During winter, however, there was very little or no liquid water in the model simulations, while liquid water was observed to persist in the clouds. Of these six models, three have moist-physics schemes that explicitly predict solid and liquid forms of cloud and precipitation separately, while the remaining three partition between solid and liquid water empirically, based on temperature. Paradoxically, the more advanced schemes, in principle being capable of describing more complex aerosol–cloud interactions, fail to produce any liquid water at all in winter, while the simple models perform (marginally) better. To fully utilize the capabilities of the more advanced schemes, more data are needed for development and evaluation.

The effect this particular error causes on, for example, downwelling longwave radiation is more difficult to evaluate. An observed error in radiation may depend on issues unrelated to the cloud water phase. Nevertheless, cloud particle phase is closely tied to the cloud's physical and radiative properties, and failure

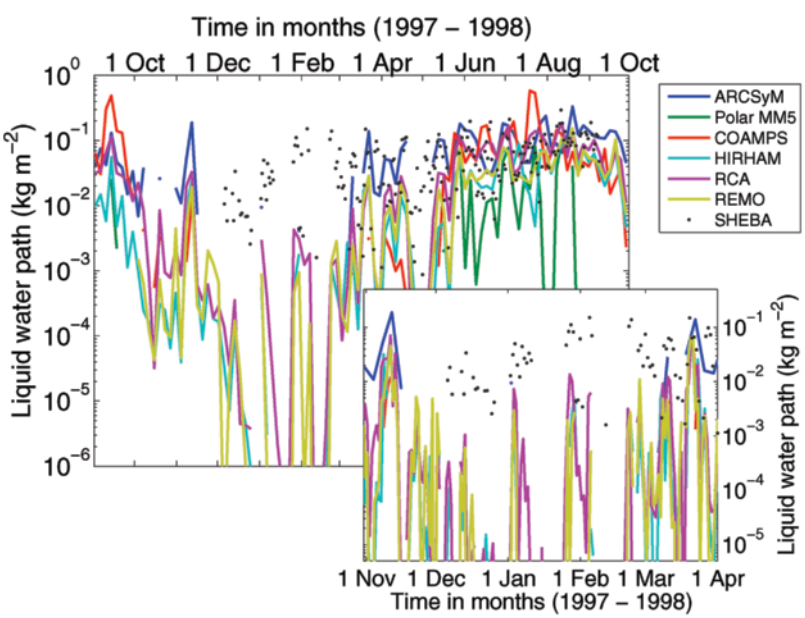


**Fig. 1. Time series of liquid water path (kg m$^{-2}$) for the SHEBA year. Solid lines are from the six ARCMIP model integrations (see the legend), while black dots are measured by a microwave radiometer at the SHEBA site. (top) Weekly averaged model values for the whole year, and (bottom) diurnal averages for winter are shown; SHEBA data are diurnal averages.**

to accurately simulate phase may lead to considerable errors. Figure 2 shows the relative probability for error in downwelling longwave radiation for clear and cloudy winter conditions. These were determined by an analysis of the observed net longwave radiation and the modeled total water path; requiring the former to be small and the latter to be simultaneously large isolates cases when there were dense clouds in both the model and reality, and vice versa for the coinciding clear occasions. There is considerable scatter between the models, more so for the cloudy than for the clear cases. While the errors for the clear cases are scattered around zero, the errors for the cloudy cases are more widespread, ranging from approximately –75 to +20 W $m^{-2}$, with a model ensemble median error around –25 W $m^{-2}$. This is considerable given that the mean observed total flux in winter is –36 W $m^{-2}$, with a standard deviation of 24 W $m^{-2}$; the simulated error is thus of the same order of magnitude as the total mean flux and is similar in magnitude to its variability. This error is likely due to the lack of liquid water in the modeled clouds.

**RECENT MEASUREMENTS.** The Mixed-Phase Arctic Cloud Experiment (M-PACE) was conducted from late September through October 2004 in the vicinity of the Department of Energy's North Slope of Alaska (NSA) field site. The overall objective of the project was to collect a focused set of observations needed to advance understanding of the dynamical and microphysical processes that lead to long-lived mixed-phase Arctic clouds. To this end, measurements of cloud and aerosol properties were made by aircraft and a suite of remote sensing devices (Verlinde et al. 2007). The IN measurements were made using a continuous-flow ice thermal diffusion chamber (CFDC) on the University of North Dakota's Citation II aircraft. This instrument permits the processing of aerosol particles sampled through an aircraft inlet in real time in and around cloud levels to determine IN concentrations (Rogers 1988; Rogers et al. 2001b). An inlet impactor upstream of the CFDC ensures that aerosol particles larger than ~1.5 $\mu$m (aerodynamic diameter) are removed prior to entering the instrument (Rogers et al. 2001b), so that large aerosol particles are not erroneously identified as ice; IN, which are larger than this cut point, are not sampled by the instrument. The CFDC is sensitive to all nucleation modes, except contact freezing. Using this technique, IN concentrations can be determined as a function of the processing temperature and supersaturation (relative humidity with respect to water or ice, in percent, minus 100) in the instrument.

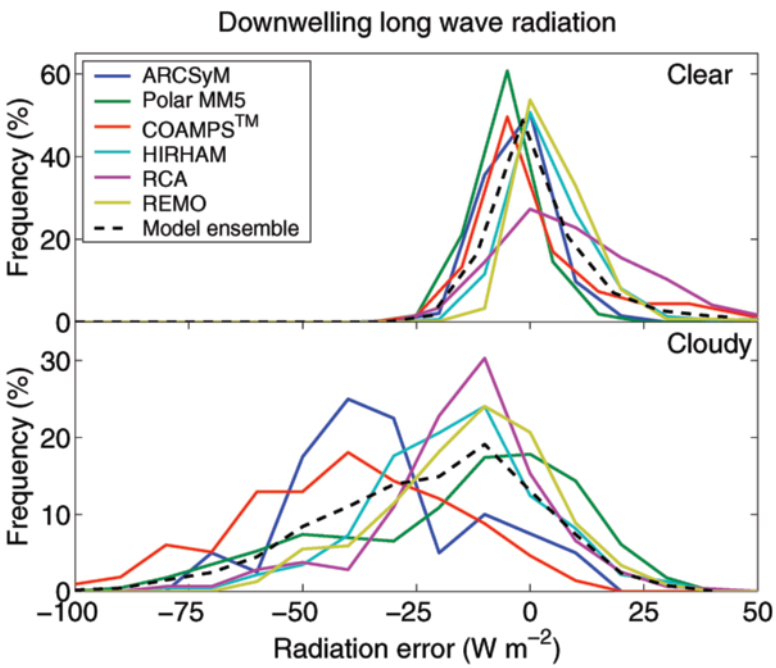


**Fig. 2. Relative frequency of occurrence of the model error in downwelling longwave radiation, defined as model minus observed flux. Solid colored lines represent models (see the legend) while the black dashed line is the across-model average. (top) Clear and (bottom) cloudy cases are shown, both defined as being clear or cloudy in both observations and model.**

In Fig. 3 the IN measurements from M-PACE are presented in composite form. Data are given as concentrations, binned into unit ranges of processing temperature, water supersaturation, and ice supersaturation. These average concentrations include a substantial contribution (~87%) from measurements for which no IN were detected. As such, the figure gives a good representation of average IN concentrations during the project, but not the range of concentrations encountered during individual flights. Measured IN concentrations are quite low, more than an order of magnitude lower than those reported by Rogers et al. (2001a) for measurements made in the spring in the Arctic. Shown in the third panel of the figure is the parameterization of Meyers et al. (1992), which is used in many models, often without regard to the location, season, or altitude being modeled. In the Meyers formulation,

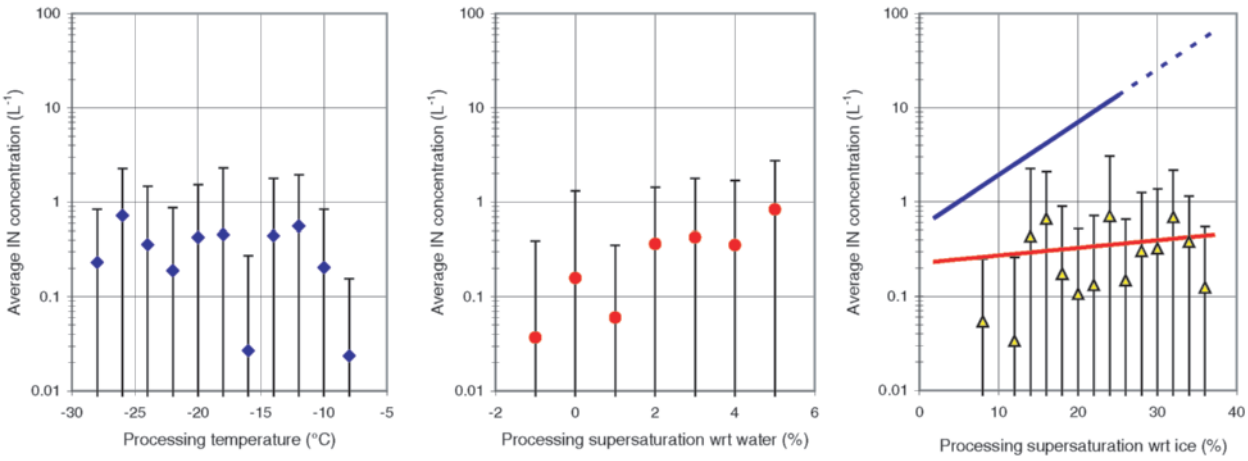


**Fig. 3. M-PACE IN data processed for project-averaged concentrations (60-s-integrated values) in finite bin intervals of CFDC processing temperature, water supersaturation, and ice supersaturation. Error bars indicate one std dev. The ice nucleation parameterization of Meyers et al. (1992), used as a standard in many cloud-resolving models, is shown for comparison in the third panel as a solid (dashed) blue line for the supersaturation range for which it can be strictly applied (is extended to higher supersaturations). A best fit for the current binned and weighted data is shown as a solid red line, using the same functional form as that in Meyers et al. (1992).**

$N_{IN} = \exp(a + bS_i)$, where $N_{IN}$ is the number of IN ($L^{-1}$), $a$ and $b$ are empirically determined from midlatitude data ($a$ = –0.639, and $b$ = 0.1296), and $S_i$ is the ice supersaturation (%). The Meyers et al. (1992) formulation is derived from measurements taken from –7° to –20°C. The majority (> 70%) of the M-PACE IN measurements were collected within this temperature range, although the data in Fig. 3 indicate that processing temperature does not greatly affect measured IN concentrations. It is clear that the Meyers et al. (1992) parameterization is not representative of average IN behavior encountered during M-PACE flights, and the use of this parameterization will impair the ability to predict cloudiness and related radiative forcing in this region (see below). Using the same function form as that of Meyers et al., we determine best-fit parameters of $a$ = –1.488 and $b$ = 0.0187 ($X^2$ = 107.5) for these binned and weighted data. Given the poor fit of the Meyers et al. formulation to these data, other functional forms were attempted as input for the simulations described below, but the results were essentially unchanged. Next, we consider the potential effects of these smaller IN concentrations on modeled Arctic cloudiness.

**SIMULATIONS USING M-PACE ICE NUCLEI MEASUREMENTS.** To illustrate the strong dependence of liquid water mass and the surface radiative budget on IN concentrations in mixed-phase Arctic clouds, we use the Regional Atmospheric Modeling System (RAMS; Cotton et al. 2003). RAMS, similar to many of the models used in ARCMIP, includes detailed computations of mixed-phase cloud processes. RAMS is a nested-grid model that allows for the simulation of large-scale features (up to global) with nested grids that allow for fine resolution down to the scale of individual clouds. The configuration used here has three nested grids (Fig. 4), with communication between the grids. All of the model results shown are from the finest grid. The model is initialized on 1200 UTC 9 October 2004 using the analyses from the National Center for Environmental Prediction's (NCEP's) Meso-ETA Model grid over Alaska. Open boundary conditions are used and the model is nudged toward the Eta Model analyses. Surface conditions are initialized using the Eta Model surface data over the terrestrial regions. The sea ice scheme is initialized with the daily Defense Meteorological Satellite Program (DMSP) SSM/I ice dataset and the ocean temperatures are initialized using the NCEP optimal interpolation (OI) SST weekly data.

The latest version of RAMS has many components that make it appropriate for Arctic simulations. It has detailed sea ice and surface models suitable for the Arctic (Cotton et al. 2003). The sea ice model is critical, because it regulates heat and moisture fluxes from the surface, which strongly influence cloud processes. Cloud processes are represented by the following seven hydrometeor types: cloud drops, rain, pristine ice crystals, snow crystals, crystal aggregates, graupel, and hail. The growth, evaporation, and sedimentation of all hydrometeors are explicitly computed in the model (Meyers et al. 1997; Walko et al. 1995). Cloud drops and pristine ice are nucleated on a

**Fig. 4. Grid spacing used by RAMS for the M-PACE study. The outer grid has a horizontal grid spacing of 64 km and covers much of Alaska and the southern Arctic Ocean from eastern Siberia to western Canada. The second grid is situated over most of northern Alaska and the adjacent Arctic Ocean, with a grid spacing of 16 km. The third grid, with a fine grid spacing of 4 km, was placed over the M-PACE domain. Model results given in the paper are from the smallest grid.**

prespecified distribution of cloud condensation nuclei (Saleeby and Cotton 2004) and IN, respectively. No in situ sources of IN are parameterized in the model. The IN can nucleate crystals in four different ways: deposition, condensation–freezing, immersion, and contact. Each mode except immersion nucleation is parameterized in our model (Cotton et al. 1986, 2003; Saleeby and Cotton 2004; Walko et al. 1995). Deposition and condensation–freezing nucleation follows Meyers et al. (1992). Contact nucleation rates are computed by using the number of nuclei from the temperature-dependent parameterization of Meyers et al. (1992), along with thermophoretic and Brownian diffusion formulations given in Cotton et al. (1986). As noted, the CFDC does not measure contact nucleation. However, it is reasonable to assume that all IN concentrations vary concomitantly. Because the deposition/condensation–freezing IN concentrations were ~26 times smaller than the midlatitude values, we reduce our contact IN concentrations by a factor of 26. Sensitivity studies conducted of contact IN concentrations up to a factor of 40 times the assumed ambient values showed no strong sensitivity to contact IN. This is in contrast to a recent modeling study (Morrison et al. 2005), though this case study from SHEBA was colder than our case, perhaps making contact nucleation more important. The RAMS model is initialized with a profile of IN that decreases with height using a polynomial dependence that mimics observed aerosol profiles. During a simulation, the model tracks the number of IN for each heterogeneous nucleation model. Hence, the IN concentration is advected by the wind, diffused by turbulence, and depleted when ice crystals nucleate and precipitate out of the atmosphere. The depletion mechanism keeps track of the nucleated aerosol (which is also advected), so that these IN are not nucleated in the future. For instance, in the case of deposition/condensation–freezing we store and track the nucleated IN at a given ice supersaturation. Thus, the most easily nucleated IN are removed first from the population, allowing for realistic depletion of IN through ice precipitation. Over time, the IN profile within the boundary layer rapidly declines to lower values. The mixing in of IN from above the boundary layer resupplies IN to the cloud layer, which causes snow precipitation bursts. Cloud and atmospheric properties are coupled to a two-stream radiative transfer code (Harrington and Olsson 2001b) that computes fluxes of solar and infrared radiation.

We simulated the time period of 9–11 October from M-PACE, during which time the North Slope and Arctic Ocean were covered by extensive mixed-phase clouds. A *Terra* satellite image of northern Alaska from 10 October is shown in Fig. 5a. High pressure over the pack ice to the northeast of the Alaskan coast dominated the NSA during this period. The high pressure system produced a low-level northeasterly flow, moving cold (~–20°C) air off of the pack ice and over the relatively warm ocean. This produced vigorous convection and persistent low-level clouds under a sharp inversion, which ultimately were advected over the NSA. No mid- or upper-level clouds were present during this period. Lidar and radar indicate a liquid cloud deck above 800-m elevation (which was verified by aircraft), with ice precipitation shafts falling from the liquid cloud deck (Verlinde et al.

2007). Representative cloud-top temperatures were approximately –17°C (Verlinde et al. 2007).

Our base simulation, shown in Fig. 5b, uses the Meyers et al. (1992) heterogeneous ice parameterization (Standard IN Dep), which includes depletion of IN through precipitation. As the figure shows, when this parameterization is used, the clouds rapidly glaciate. Unlike the observed clouds, very little liquid remains and most of the region is covered by thin ice clouds, consistent with the ARCMIP results discussed above. The lack of liquid water is due to the large number of IN predicted by the Meyers et al. (1992) parameterization. The simulation was repeated using the new IN parameterization from M-PACE (M-PACE IN Dep), which also includes depletion of IN through precipitation. As Fig. 5c shows, extensive decks of liquid clouds with smaller amounts of ice are now predicted, in better accord with observations. This difference between the two simulations occurs throughout the 2-day period, and is demonstrated in Fig. 6a, which shows the measured and modeled LWP over one of the observation sites (Oliktok Point). The standard IN case (Standard IN Dep) produces some liquid, but this is rapidly removed through glaciation, whereas the M-PACE IN case (M-PACE IN Dep) maintains a thick liquid cloud that continually precipitates ice, although the modeled LWP is smaller than the observations. The average modeled ice concentrations compare favorably to the observations (from ~0.1 to 1 $L^{-1}$), but never reached some of the large observed values (10s to 1000 $L^{-1}$). The model could not capture some of the high ice water contents (IWCs) because of precipitation bursts, but the average IWC (~0.05 g $m^{-3}$) compares roughly with that observed by the aircraft. Nevertheless, the simulations that use M-PACE IN concentrations are able to maintain a liquid cloud deck, unlike the standard model. These results are consistent with the earlier modeling results of Harrington et al. (1999) and Harrington and Olsson (2001a).

It is important to note that IN are depleted due to ice crystal nucleation and precipitation in the simulations described above. Most models do not deplete IN concentrations, and instead use formulations like Meyers et al. (1992) in a static fashion; that is, $N_{IN}$ is constant for each $S_i$ throughout a simulation. This is unrealistic because, as Harrington and Olsson (2001a) showed, ice nucleation and subsequent precipitation removal of IN from the cloud layer increases the mixed-phase lifetime. As Fig. 6a shows, simulations in which IN concentrations were increased by a factor of 2 (2 × M-PACE IN Dep) and 10 (10 × M-PACE IN Dep) are still able to maintain the observed mixed-phase structure, albeit with smaller liquid water amounts. Without depletion, even the MPACE-derived IN concentrations (M-PACE IN, No Dep) lead to rapid glaciation and the loss of all liquid water, which is consistent with Harrington and Olsson (2001).

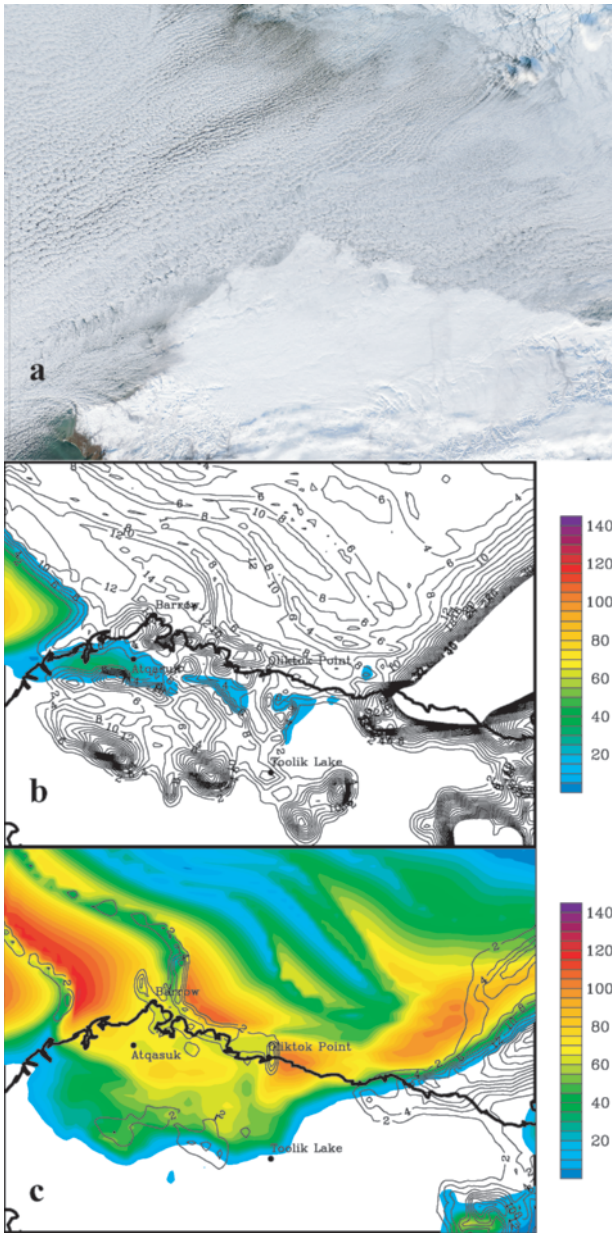


**Fig. 5. (a) *Terra* image for 10 Oct 2004 over study area, showing extensive cloud deck. Moderate Resolution Imaging Spectroradiometer (MODIS) satellite data courtesy of National Aeronautics and Space Administration (NASA) Goddard Space Flight Center. Liquid water path (shaded, g $m^{-2}$) and ice water path (contoured, g $m^{-2}$) at 1800 UTC for simulations with (b) Standard IN concentrations and (c) M-PACE IN concentrations. Both simulations include depletion of IN concentrations through ice precipitation.**

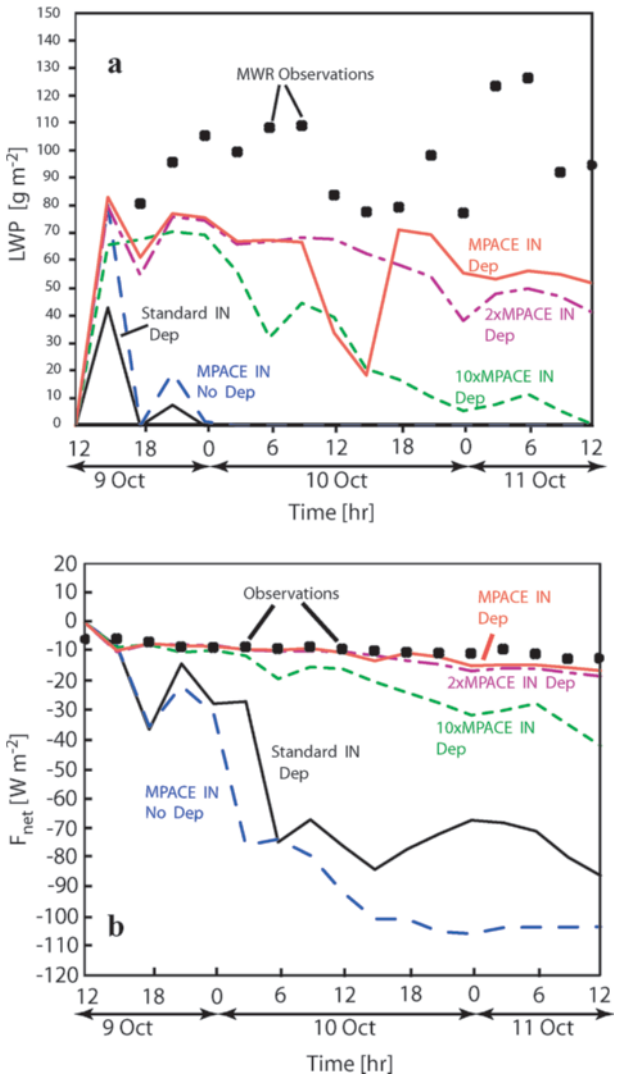


**Fig. 6. Time series for the 2-day simulation plotted over Oliktok Point for Standard IN and M-PACE IN concentrations: (a) liquid water path (g m$^{-2}$) and (b) net infrared surface flux (W m$^{-2}$, $F^- - F^+$; negative values indicate surface cooling by infrared emission). The solid black points represent observed values: (a) microwave radiometer measurements over Oliktok Point, and (b) surface infrared flux measurements at Oliktok Point. Multipliers in front of M-PACE IN indicate the factor by which the M-PACE IN concentrations were increased in the model. "Dep" indicates that IN were depleted by ice precipitation whereas "No Dep" indicates that IN were not depleted.**

These results suggest that to capture the radiatively important liquid phase of Arctic clouds, models must correctly predict both the number of heterogeneously nucleated ice and the cloud processing and removal of IN through precipitation.

Correctly simulating mixed-phase clouds is important for many reasons, but arguably the most important is the influence of these clouds on the Arctic radiative budget. Figure 6b shows the net infrared radiative flux at the surface for each of the simulations in Fig. 6a. The simulation with the new IN parameterization and IN depletion (M-PACE IN Dep) compares very well with surface observations of downwelling and upwelling broadband longwave radiation made using Eppley Precision Infrared Radiometers. As IN concentrations increase in the model, liquid water is converted rapidly to ice, leading to optically thin clouds. This causes the net infrared radiative loss to change dramatically. Thus, small changes in the IN population, and whether or not IN are depleted in the model, can lead to substantial changes in the surface radiative budget (up to 100 W m$^{-2}$) for boundary layer clouds that develop in response to off-ice flow in autumn. Although these simulations are based on a limited IN dataset collected in autumn and are for a specific cloud event, these results support the notion that the phase of water in the Arctic BL clouds plays a critical role in Arctic regional climate. Further, in comparison with those of the ARCMIP simulations, these limited results suggest that the lack of liquid water predicted by models in Arctic clouds, and hence the errors in the surface radiative budget, is possibly linked to inadequate parameterization of ice processes, and ice nuclei, in the Arctic.

**CONCLUSIONS.** Recent studies suggest that Arctic climate is more sensitive to changes in climate forcing than other regions on Earth, while global climate models are less reliable in this region (ACIA 2004). Clouds play an important role for the Arctic surface energy balance and are difficult to model. Based on IN measurements collected during M-PACE and simulations based on a case encountered during M-PACE, we show one possible reason for this: a difference in the aerosol properties of the Arctic compared to lower latitudes. The global climate models that form the basis for assessments such as the Arctic Climate Impact Assessment (ACIA) use the same cloud and aerosol descriptions in the Arctic as anywhere else on Earth, and are calibrated to provide a reasonable global climate. For the global climate, ice clouds such as those found in tropical cirrus anvils are probably more important; however, applying formulations optimized for midlatitude and tropical conditions to the Arctic, where conditions are clearly different, results in a poor representation of this apparently very sensitive region. It also means that global models that underpredict liquid water clouds may feature a larger shift from ice to liquid clouds as the model climate

warms. This may constitute an enhanced, unrealistic feedback on climate change. Moreover, global models that underpredict liquid could miss an important aerosol feedback; namely, an increase in Arctic IN concentrations may cause a reduction in liquid cloud amounts. The challenge is to improve model process representations in special regions like the Arctic, without disrupting apparently well-working formulations for other regions. An important conclusion from these results is a necessity to include realistic treatment of aerosols and aerosol–cloud interactions in future climate simulations.

**ACKNOWLEDGMENTS.** AJP and PJD were supported by the Battelle Memorial Institute, U.S. Department of Energy, Grant 12441. For the M-PACE field study, JYH, AA, and JV were supported by the Office of Science (BER), U.S. Department of Energy, Grant DE-FG02-05ER64058. MT is funded by the Swedish Research Council through Grants 2002-5610 and 2004-4271. MT wishes to acknowledge collaborators in the ARCMIP program for allowing the use of their model data: John Cassano, Klaus Dethloff, Colin Jones, Susanne Pfeifer, Annette Rinke, Michael Shaw, Tido Semmler, Gunilla Svensson, Klaus Wyser, and Mark Žagar. Additional thanks to those who did the work of setting up the model experiment: Judy Cury, Amanda Lynch, Elizabeth Cassano, and Jeff Key, and also to the SHEBA team. CNL acknowledges the support of the Climate Change Research Division of the U.S. Department of Energy as part of the ARM Program. The authors are grateful for the useful comments of the two anonymous reviewers, and to Nat Johnson, for comments that greatly improved the manuscript. Recognition is also extended to those responsible for the operation and maintenance of the instruments that produced the data used in this study. Finally, the authors wish to thank Michael Poellot, Tony Grainger, and the UND flight crew and staff. MODIS satellite data courtesy of NASA Goddard Space Flight Center.

## REFERENCES

ACIA, 2004: *Impacts of a Warming Arctic: Arctic Climate Impact Assessment*. Cambridge University Press, 1020 pp.

Barrie, L. A., 1986: Arctic air-pollution—An overview of current knowledge. *Atmos. Environ.,* **20,** 643–663.

Beesley, J. A., and R. E. Moritz, 1999: Toward an explanation of the annual cycle of cloudiness over the Arctic Ocean. *J. Climate,* **12,** 395–415.

Bigg, E. K., 1996: Ice forming nuclei in the high Arctic. *Tellus,* **48,** 223–233.

Borys, R. D., 1989: Studies of ice nucleation by Arctic aerosol on AGASP-II. *J. Atmos. Chem.,* **9,** 169–185.

Cassano, J. J., J. E. Box, D. H. Bromwich, L. Li, and K. Steffen, 2001: Evaluation of Polar MM5 simulations of Greenland's atmospheric circulation. *J. Geophys. Res.,* **106,** 33 867–33 890.

Christensen, J. H., O. B. Christensen, P. Lopez, E. Van Meijgaard, and A. Botzet, 1996: The HIRHAM4 regional atmospheric model. DMI Scientific Rep. 96-4, 51 pp.

Cooper, W. A., 1986: Ice initiation in natural clouds. *Precipitation Enhancement—A Scientific Challenge, Meteor. Monogr.,* Vol. 21, Amer. Meteor. Soc., 29–32.

Cotton, W. R., G. J. Tripoli, R. M. Rauber, and E. A. Mulvihill, 1986: Numerical simulation of the effects of varying ice crystal nucleation rates and aggregation processes on orographic snowfall. *J. Climate Appl. Meteor.,* **25,** 1658–1680.

——, and Coauthors, 2003: RAMS 2001: Current status and future directions. *Meteor. Atmos. Phys.,* **82,** 5–29.

Curry, J. A., and E. E. Ebert, 1990: Sensitivity of the thickness of Arctic sea ice to the optical properties of clouds. *Ann. Glaciol.,* **14,** 43–46.

——, and A. H. Lynch, 2002: Comparing Arctic regional climate models. *Eos, Trans. Amer. Geophys. Union,* **83,** 87.

——, W. B. Rossow, D. Randall, and J. L. Schramm, 1996: Overview of Arctic cloud and radiation characteristics. *J. Climate,* **9,** 1731–1764.

Fountain, A. G., and T. Ohtake, 1985: Concentrations and source areas of ice nuclei in the Alaskan atmosphere. *J. Climate Appl. Meteor.,* **24,** 377–382.

Harrington, J. Y., and P. Q. Olsson, 2001a: On the potential influence of ice nuclei on surface-forced marine stratocumulus cloud dynamics. *J. Geophys. Res.,* **106,** 27 473–27 484.

——, and ——, 2001b: A method for the parameterization of cloud optical properties in bulk and bin microphysical models: Implications for arctic cloudy boundary layers. *Atmos. Res.,* **57,** 51–80.

——, T. Reisin, W. R. Cotton, and S. M. Kreidenweis, 1999: Cloud resolving simulations of Arctic stratus—Part II: Transition-season clouds. *Atmos. Res.,* **51,** 45–75.

Hobbs, P. V., and A. L. Rangno, 1998: Microstructures of low and middle-level clouds over the Beaufort Sea. *Quart. J. Roy. Meteor. Soc.,* **124,** 2035–2071.

Hodur, R. M., 1997: The Naval Research Laboratory's coupled ocean–atmosphere mesoscale prediction system (COAMPS). *Mon. Wea. Rev.,* **125,** 1414–1430.

Holland, M. M., and C. M. Bitz, 2003: Polar amplification of climate change in coupled models. *Climate Dyn.,* **21,** 221–232.

Intrieri, J. M., M. D. Shupe, T. Uttal, and B. J. McCarty, 2002: An annual cycle of Arctic cloud characteristics

observed by radar and lidar at SHEBA. *J. Geophys. Res.,* **107,** 8030, doi:10.1029/2000JC000423.

Jacob, D., 2001: A note to the simulation of annual and interannual variability of the water budget over the Baltic Sea drainage basin. *Meteor. Atmos. Phys.,* **77,** 61–73.

Jiang, H. L., W. R. Cotton, J. O. Pinto, J. A. Curry, and M. J. Weissbluth, 2000: Cloud resolving simulations of mixed-phase Arctic stratus observed during BASE: Sensitivity to concentration of ice crystals and large-scale heat and moisture advection. *J. Atmos. Sci.,* **57,** 2105–2117.

Jones, C. G., K. Wyser, A. Ullerstig, and U. Willén, 2004: The Rossby Center regional atmospheric climate model. Part II: Application of the Arctic. *Ambio,* **33,** 211–220.

Kattsov, V. M., and E. Källén, 2004: Future climate change: Modeling and scenarios for the Arctic. *Impacts of a Warming Arctic: Arctic Climate Impacts Assessment,* S. J. Hassol, Ed., Cambridge University Press, 99–150.

Lynch, A. H., W. L. Chapman, J. E. Walsh, and G. Weller, 1995: Development of a regional climate model of the western Arctic. *J. Climate,* **8,** 1555–1570.

MacBean, G., 2004: Arctic climate—Past and present. *Impacts of a Warming Arctic: Arctic Climate Impacts Assessment,* S. J. Hassol, Ed., Cambridge University Press, 21–60.

Meyers, M. P., P. J. Demott, and W. R. Cotton, 1992: New primary ice-nucleation parameterizations in an explicit cloud model. *J. Appl. Meteor.,* **31,** 708–721.

——, R. L. Walko, J. Y. Harrington, and W. R. Cotton, 1997: New RAMS cloud microphysics parameterization. 2. The two-moment scheme. *Atmos. Res.,* **45,** 3–39.

Morrison, H., and J. O. Pinto, 2005: Mesoscale modeling of springtime Arctic mixed-phase stratiform clouds using a new two-moment bulk microphysics scheme. *J. Atmos. Sci.,* **62,** 3683–3704.

——, M. D. Shupe, and J. A. Curry, 2003: Modeling clouds observed at SHEBA using a bulk microphysics parameterization implemented into a single-column model. *J. Geophys. Res.,* **108,** 4255, doi:10.1029/2002JD002229.

——, ——, J. O. Pinto, and J. A. Curry, 2005: Possible roles of ice nucleation mode and ice nuclei depletion in the extended lifetime of Arctic mixed-phase clouds. *Geophys. Res. Lett.,* **32,** L18801, doi:10.1029/2005GL023614.

Pinto, J. O., 1998: Autumnal mixed-phase cloudy boundary layers in the Arctic. *J. Atmos. Sci.,* **55,** 2016–2038.

Rangno, A. L., and P. V. Hobbs, 2001: Ice particles in stratiform clouds in the Arctic and possible mechanisms for the production of high ice concentrations. *J. Geophys. Res.,* **106,** 15 065–15 075.

Rinke, A., and Coauthors, 2006: Evaluation of an ensemble of Arctic regional climate models: Spatiotemporal fields during the SHEBA year. *Climate Dyn.,* **26,** 459–472.

Rogers, D. C., 1988: Development of a continuous flow thermal gradient diffusion chamber for ice nucleation studies. *Atmos. Res.,* **22,** 149–181.

——, P. J. DeMott, and S. M. Kreidenweis, 2001a: Airborne measurements of tropospheric ice-nucleating aerosol particles in the Arctic spring. *J. Geophys. Res.,* **106,** 15 053–15 063.

——, ——, ——, and Y. Chen, 2001b: A continuous-flow diffusion chamber for airborne measurements of ice nuclei. *J. Atmos. Oceanic Technol.,* **18,** 725–741.

Saleeby, S. M., and W. R. Cotton, 2004: A large-droplet mode and prognostic number concentration of cloud droplets in the Colorado State University Regional Atmospheric Modeling System (RAMS). Part I: Module descriptions and supercell test simulations. *J. Appl. Meteor.,* **43,** 182–195.

Serreze, M., and J. A. Francis, 2006: The Arctic amplification debate. *Climate Dyn.,* **76,** 241–264.

Tjernström, M., and Coauthors, 2005: Modelling the Arctic boundary layer: An evaluation of six ARCMIP regional-scale models with data from the SHEBA project. *Bound.-Layer Meteor.,* **117,** 337–381.

Uttal, T., and Coauthors, 2002: Surface heat budget of the Arctic Ocean. *Bull. Amer. Meteor. Soc.,* **83,** 255–276.

Vavrus, S., 2004: The impact of cloud feedbacks on Arctic climate under greenhouse forcing. *J. Climate,* **17,** 603–615.

Verlinde, J., and Coauthors, 2007: The mixed-phase Arctic cloud experiment (M-PACE). *Bull. Amer. Meteor. Soc.,* in press.

Walko, R. L., W. R. Cotton, M. P. Meyers, and J. Y. Harrington, 1995: New RAMS cloud microphysics parameterization. 1. The single-moment scheme. *Atmos. Res.,* **38,** 29–62.

Walsh, J. E., 2004: Cryosphere and hydrology. *Impacts of a Warming Arctic: Arctic Climate Impacts Assessment,* S. J. Hassol, Eds., Cambridge University Press, 183–242.

——, V. M. Kattsov, W. L. Chapman, V. Govorkova, and T. Pavlova, 2002: Comparison of Arctic climate simulations by uncoupled and coupled global models. *J. Climate,* **15,** 1429–1446.

Westwater, E. R., Y. Han, M. D. Shupe, and S. Y. Matrosov, 2001: Analysis of integrated cloud liquid and precipitable water vapor retrievals from microwave radiometers during the Surface Heat Budget of the Arctic Ocean project. *J. Geophys. Res.,* **106,** 32 019–32 030.