Suppression of Arctic air formation with climate warming: Investigation with a 2-dimensional cloud-resolving model

Timothy W. Cronin,*

Department of Earth, Atmospheric, and Planetary Sciences, MIT, Cambridge, Massachusetts

Harrison Li

Harvard University, Cambridge, Massachusetts

Eli Tziperman

Department of Earth and Planetary Sciences and School of Engineering and Applied Sciences, Harvard University, Cambridge, Massachusetts

*Corresponding author address: Department of Earth, Atmospheric, and Planetary Sciences, MIT, Cambridge, Massachusetts, USA

E-mail: twcronin@mit.edu
Arctic climate change in winter is tightly linked to changes in the strength of surface temperature inversions, which occur frequently in the present climate as Arctic air masses form during polar night. Recent work proposed that increasing low cloud optical thickness in a warmer climate could suppress formation of Arctic air masses, amplifying winter warming over continents and sea ice relative to open ocean – but this mechanism was based on single-column simulations that did not allow assessment of how fractional cloud cover might change in warmer climates. This paper examines the potential for suppression of Arctic air formation with climate warming, using a 2-dimensional cloud-resolving model with several different microphysics schemes.

The cloud-resolving model supports the single-column model results: low cloud optical thickness and duration increase strongly with initial air temperature, slowing the surface cooling rate as the climate is warmed. The cloud-resolving model cools less at the surface than the single-column model and is more sensitive to warmer initial states, because it produces cloudier atmospheres with stronger lower-tropospheric mixing, and distributes cloud-top cooling over a deeper atmospheric layer with larger heat capacity. Resolving cloud turbulence has largest impact on the microphysics schemes that best represent mixed-phase clouds, increasing their sensitivity to climate warming. These findings support the hypothesis that increasing insulation of the high-latitude land surface by low clouds in a warmer world could act as a strong positive feedback in future climate change, and suggest studying Arctic air formation in a 3-dimensional climate model as a future research direction.
1. Introduction

In recent decades, the Arctic has warmed faster than the rest of the globe, particularly during winter (Chylek et al. 2009; Hartmann et al. 2013), and Arctic-amplified warming is expected to continue through the 21st century according to climate model simulations (Holland and Bitz 2003; Pithan and Mauritsen 2014). Arctic amplification of temperature change occurs in models primarily due to positive feedbacks that act at high latitudes, especially those related to decreased surface albedo from melting of ice and snow, and to changes in the tropospheric lapse rate (Pithan and Mauritsen 2014), though it can also occur without locally amplifying feedbacks due to increased poleward atmospheric heat transport (e.g., Alexeev and Jackson 2013). Although changes in tropospheric lapse rate contribute strongly to Arctic amplification, assessment of lapse rate changes in models has often been diagnostic; a more process-based understanding of the controls on high-latitude lapse rates and surface inversion strength is needed.

Recent work by Cronin and Tziperman (2015) found amplified surface warming and weaker surface inversions over high-latitude winter continents due to increasing insulation of the surface by optically thick liquid clouds in a single-column model. The goal of this study is to test the viability and strength of this mechanism – increased surface insulation by low clouds in a warmer world – in a model that resolves clouds, and to compare cloud-resolving and single-column model results.

Arctic air formation is an initial value problem in which an air mass with a prescribed initial temperature and moisture profile is allowed to cool by radiation to space over a low heat capacity surface in the absence of sunlight. This idealized problem represents the advection of maritime air over high latitude land or sea ice during polar night. The concept of Arctic air formation was first introduced by Wexler (1936), who explored the transformation of polar maritime air into Arctic air.
by longwave radiative cooling in the absence of sunlight or clouds. Based on observed soundings
during events of high-latitude air mass stagnation over land, Wexler (1936) suggested that the
temperature profile of an Arctic air mass could be represented by a relatively cold isothermal layer
overlying an even colder surface inversion layer. Both the isothermal layer and the inversion layer
cool as time progresses and the isothermal layer deepens; in essence, radiative cooling of the
surface to space steadily consumes the heat content of the lower troposphere from beneath.

Work since the study of Wexler (1936) has slowly built from his original idea. Curry (1983)
used a one-dimensional model with more detailed radiative transfer to examine the role of clouds,
subsidence, and turbulence on the formation of Arctic air – and found that the process is especially
sensitive to the amount of condensate in the atmosphere and its partitioning between liquid and ice.
Overland and Guest (1991) conducted longer simulations, including a modification of the problem
to explore equilibrium Arctic atmospheric temperature structure in winter. They also examined
the rapid adjustment of surface temperature and inversion strength to transitions between clear
and cloudy skies. Emanuel (2008) performed single-column model calculations of Arctic air mass
formation, and showed that, owing to its large static stability, Arctic air has greater saturation
potential vorticity relative to the rest of the troposphere, which may allow it to play an important
role in development of mid-latitude weather systems.

Arctic air formation has also been linked to the broader investigation of Arctic mixed-phase
clouds and boundary layer dynamics, air mass transformation, and the role of the surface energy
balance throughout the year in sea ice loss. The longwave radiative effect of clouds is a large term
in the Arctic surface energy balance in all seasons, and consequently regulates ice loss over both
land and sea, with more ice loss typical under cloudy conditions despite less absorption of short-
wave radiation (Kapsch et al. 2013; VanTricht et al. 2016; Mortin et al. 2016). From a Lagrangian
standpoint, clouds form as warm and moist air masses are advected and cooled over the Arctic
ocean in both summer – leading to ice melt and fog (Tjernstrom et al. 2015) – as well as winter – leading to bottom-heavy warming and weakening of surface inversions (Woods and Caballero 2016). A key finding from recent field studies in the Arctic is bimodality of the Arctic winter boundary layer: the coupled system of boundary layer and surface prefer to reside in either a “radiatively clear” state or an “opaquely cloudy” state (Stramler et al. 2011; Morrison et al. 2012). The radiatively clear state is characterized by clear skies or optically thin ice cloud, a colder surface and stronger surface inversion, and net surface radiative cooling of $\sim 40$ W m$^{-2}$, whereas the opaquely cloudy state is characterized by presence of cloud liquid, a warmer surface under a weaker elevated inversion, and near-zero surface net radiation. Pithan et al. (2014) found that similar distinct clear and cloudy states emerge in a single-column simulation of Arctic air formation, and that a bimodal distribution of surface longwave radiation emerges when the transient cooling process is sampled in time. They also found that many climate models fail to capture this bimodality, due to insufficient maintenance of supercooled cloud liquid and thus a poor representation of Arctic mixed-phase clouds. Pithan et al. (2016) followed on this work, and used an intercomparison of Arctic air formation in single-column versions of several weather and climate models to understand the strengths and biases in model representation of both the clear and cloudy boundary layer states.

Cronin and Tziperman (2015) focused on the climate sensitivity of the process of Arctic air formation, by modifying the temperature of the initial sounding (representing maritime air) while holding relative humidity fixed. They found that a warmer initial atmosphere leads to longer-lasting low clouds and a reduced surface cooling rate, suppressing Arctic air formation and potentially amplifying winter continental warming – which would help explain continental warmth in past climates (e.g., Greenwood and Wing 1995). Increasing optical thickness of low clouds with warming results from both increasing condensate amount and from the change in cloud phase from
ice to liquid. From the perspective of boundary layer states, a warmer initial atmosphere increases the fraction of the cooling period spent in the opaquely cloudy state.

One potential concern that could be raised with these findings, however, is that the single-column model used in Cronin and Tziperman (2015) did not allow for fractional cloud cover. A decrease in cloud fraction with warming would damp the effect of more cloud condensate on reducing the surface cooling rate. Furthermore, all single-column models rely on parameterizations to represent processes of convection, cloud-top mixing, and entrainment, which are important in Arctic mixed-phase clouds, and may change with warming (e.g., Morrison et al. 2012). It is not trivial that a 2-dimensional model, which explicitly resolves these processes down to a scale of a few hundred meters, would have the same sensitivity to warming as a single-column model.

To test the suppression of Arctic air formation with climate warming found by Cronin and Tziperman (2015), this paper studies the process of Arctic air formation and its sensitivity to temperature in a 2-dimensional cloud-resolving model. We find that clouds tend to have large fractional area coverage, and that increased mixing and deeper cloud layers actually lead to more cloud cover and less surface cooling in the 2-dimensional model than in the single-column model. Thus, sensitivity of the 2-dimensional model is broadly consistent with Cronin and Tziperman (2015): warming of the initial atmosphere leads to more low clouds and a reduced surface cooling rate. The 2-dimensional model, however, is even more sensitive than the single-column model to warming of the initial atmosphere, especially for the microphysics schemes with the best representation of mixed-phase clouds. We describe the model setup in section 2, present results in section 3, and discuss our findings in section 4.
2. Model description

We use a 2-dimensional idealized configuration of the Weather Research and Forecasting (WRF, version 3.4.1) model to simulate the process of Arctic air formation over a low heat capacity land surface for a 14-day period during polar night. Simulations span a range of different initial temperature profiles and microphysics schemes. Many aspects of the model setup follow Cronin and Tziperman (2015), and these are summarized below. The surface is a uniform moist slab with the roughness of an ocean surface and heat capacity $C_S = 2.1 \times 10^5$ J m$^{-2}$ K$^{-1}$. This heat capacity is equivalent to a water layer of depth 5 cm, and corresponds to the heat capacity of the surface layer of a deep snowpack that communicates with the atmosphere on a 1-day time scale. Roughness lengths in the main set of simulations are $\sim 1 - 4 \times 10^{-5}$ m; sensitivity tests with larger roughness lengths of $10^{-4}$ and $10^{-3}$ m differ little in a qualitative sense but have slightly colder 2-meter air temperatures because the cold surface is more strongly coupled to the atmosphere. Initial temperature profiles are defined by a single parameter, the initial 2-meter air temperature $T_2(0)$ (note that we use the terms “2-meter air temperature” and “surface air temperature” interchangeably in this paper). Above the surface, initial soundings have a tropospheric lapse rate that is either moist-adiabatic or $-8$ K km$^{-1}$, whichever is more stable, and an isothermal stratosphere at $-60^\circ$C. The initial relative humidity profile decreases from 80 percent at the surface to 20 percent at 600 hPa; above 600 hPa the relative humidity is constant at 20 percent up to the tropopause, and set to a mixing ratio of 0.003 g kg$^{-1}$ in the stratosphere. These temperature and humidity profiles are chosen to mimic those in Pithan et al. (2014) and Pithan et al. (2016), but with moist-adiabatic stratification allowing for extension to warmer initial profiles without leading to moist convective instability. In this paper we use the phrase “climate warming” synonymously with “warming of the initial atmospheric state”.
The model domain is 15 km long, \( \sim 15 \) km high, and periodic in the horizontal, with a horizontal grid spacing of 100 m and a uniform initial vertical grid spacing of 50 m. Cronin and Tziperman (2015) used a stretched grid, with spacing increasing from \( \sim 20 \) m at the surface to \( \sim 200 \) m at 2 km altitude and becoming even coarser higher up, so this study resolves all clouds that are not fog with more vertical detail. The model time step for dynamics is 1.5 seconds, the Coriolis parameter is set to zero, and no cumulus parameterization scheme is used. The spacing of levels in height decreases with time in each simulation because the vertical coordinate in the model is based on hydrostatic pressure, so cooling causes compression (by roughly 20\% for 55\(^\circ\)C of cooling at a starting temperature of 0\(^\circ\)C). Although the simulations in Cronin and Tziperman (2015) extend to 35 km height, we reduce the vertical extent of the model here because the stratosphere is not essential for this problem. We also run the model with varied grid spacing in both the horizontal and vertical to test the robustness of the main results (see section 4d). To stimulate turbulent mixing near the surface (both parameterized mixing and resolved eddies) and prevent the development of stationary surface temperature heterogeneities, a uniform initial 5 m s\(^{-1}\) zonal wind is imposed everywhere, and the horizontal-mean flow is relaxed to this value over a 1-day period. To allow development of horizontal asymmetries, we initialize the lowest four levels in the model with random temperature perturbations that are spatially white and chosen from a uniform distribution between -0.1 K and +0.1 K.

Longwave radiative transfer in the model is parameterized using the RRTMG scheme (Iacono et al. 2008), which is called every 2 minutes. Sensitivity tests with 6 seconds between radiation calls lead to minimal difference in the results, but are substantially more computationally costly. Shortwave radiation is zero at all times, because the latitude is set to 90\(^\circ\)N and the model runs occur during the first two weeks of January. We use the Yonsei University (YSU) boundary layer scheme, which diffuses heat and moisture non-locally within a layer of depth diagnosed using a
bulk Richardson number (Hong et al. 2006). Because the near-surface stability in our model setup is often very high, and winds are relatively weak, diagnosed boundary layer depths are small – rarely up to a few hundred meters and often less than 50 m in depth. Resolved turbulent flow dominates transport above this height. Surface turbulent fluxes are parameterized using Monin-Obukhov similarity with separate look-up functions for different stability classes, and the latent heat flux is not allowed to be negative (i.e., there is no dew or frost). Test simulations allowing for dew and frost show small effects on the surface energy balance and do not substantially alter our findings.

One advantage of using WRF is the ability to easily test different parameterizations of cloud microphysics. The six schemes are denoted Lin-Purdue, WSM-6 (WRF single-moment 6-class), Goddard, Thompson, Morrison, and Stony Brook, and some basic information on each is given in the Appendix, with a focus on generation processes for cloud ice. Microphysics schemes are called at every model time step. As in Cronin and Tziperman (2015), some results are shown only for the Lin-Purdue microphysics scheme, but many plots show mean and spread of the 6 microphysics schemes used. The Lin-Purdue scheme is a relatively simple bulk (single-moment) scheme that models six hydrometeors – water vapor, cloud liquid, cloud ice, snow, rain, and graupel – assuming exponential size distributions of rain, snow, and graupel particles (Lin et al. 1983). We also perform sensitivity tests labeled “no-CRF” where cloud-radiation interactions are disabled; the Lin-Purdue scheme is still used but cloud water concentrations are set to zero in the radiative transfer scheme (latent heat associated with phase changes is still considered). Comparing no-CRF results to the results with radiatively active clouds allows us to quantify the impact of cloud longwave forcing on Arctic air formation.

To explore the effects of surface heterogeneity, we also perform a set of sensitivity tests with the Lin-Purdue scheme, where the surface heat capacity is heterogeneous but has the same domain-
average value. We set $C_S = 9.66 \times 10^5$ J m$^{-2}$ K$^{-1}$ (23 cm of water equivalent) in a contiguous 10% of the domain, and $C_S = 1.26 \times 10^5$ J m$^{-2}$ K$^{-1}$ (3 cm of water equivalent) elsewhere. This heterogeneity in surface heat capacity is intended as a crude representation of leads in sea ice, or areas of shallow water over land (i.e., lakes, wetlands, streams). Although the true heat capacity of such heterogeneities in land cover would likely be larger, we do not want to complicate the interpretation of these sensitivity tests by also modifying the domain-average surface heat capacity. Limitations in our treatment of the surface are further discussed in Section 4d.

One important caveat of this modeling setup is that, although some of the microphysics schemes predict particle sizes in different categories, this information is not passed to the RRTMG radiation scheme. Rather, in the version of WRF that we use, only the total cloud liquid and ice content in each column and vertical level is passed to RRTMG, and the radiation scheme makes assumptions about the cloud particle sizes. With more complete radiation-microphysics coupling, differences in microphysics schemes would likely affect the results more than indicated here.

3. Results

a. Time evolution of temperature and clouds

An increase in cloudiness with warmer initial states is evident from snapshots of modeled Arctic air formation after four days of cooling from initial 2-m air temperatures ($T_2(0)$) of 0, 10, and 20°C and the Lin-Purdue microphysics scheme (Figure 1). Video S1 also shows the evolution of the cloud fields and temperature profiles over the whole two-week period; below we summarize this evolution over the first four days to explain the origins of different cloud structures in Figure 1. Analogous supplemental Videos S2-S6 are included to show simulations with the other microphysics schemes, but are not discussed in detail.
When the initial surface temperature is $T_2(0) = 0^\circ$C (Figure 1a), a physically thin but optically thick mixed-phase fog forms after about 6 hours of initial clear-sky cooling, but dissipates in roughly a day due to scavenging of cloud water by snow, leaving behind tenuous ice clouds and snow. Although the surface temperature rebounds slightly when fog forms, the first four days are dominated by clear-sky surface cooling by more than $25^\circ$C and development of a strong surface inversion. With a warmer initial surface temperature of $T_2(0) = 10^\circ$C (Figure 1b), a surface fog layer again forms after about 6 hours, but it is all liquid and deepens rapidly. An elevated stratus layer detaches from the surface fog around hour 36, grows upward into the clear background air, develops some ice after about hour 48 as it continues to cool, and then dissipates rapidly by precipitation just before the end of day 3. A new cloud layer beneath, limited in its growth by effects of the overlying cloud layer (weakened radiative cooling and precipitation), then rapidly thickens and deepens, and repeats the cycle of growth and decay over day 4. At the end of day 4, the broken remnants of this second mixed-phase stratus cloud layer lie at about 1.5 km altitude, and a new liquid fog layer is beginning to form at the surface. As a consequence of strong longwave cloud radiative forcing during this sequence of two cloud layer life cycles, the surface has cooled only by about $12^\circ$C by the end of the fourth day, and there is only a very weak surface inversion. For an even warmer initial surface temperature of $T_2(0) = 20^\circ$C (Figure 1c), the initial evolution and growth of the surface fog layer is very similar to the $T_2(0) = 10^\circ$C case. The stratus layer at $\sim 2.5$ km, however, has persisted continuously since it detached from the initial fog layer around hour 32, and is only just beginning to develop ice and dissipate in the snapshot shown in Figure 1c. There is also a second optically thick liquid cloud layer below it. This abundant and persistent cloud cover displaces radiative cooling upwards from the surface to the top of the upper cloud layer, resulting in only $\sim 10^\circ$ C of surface cooling by the end of the fourth day, with the low-level lapse rate remaining nearly moist adiabatic as in the initial state. At the end of day four, the
region between the two cloud layers for $T_2(0) = 20^\circ$C is convective (see Video S1), with horizontal fluctuations of vertical velocity $\sim 0.2$ m s$^{-1}$.

The behavior and structure of mixed-phase clouds in the warm and very warm simulations ($T_2(0) = 10^\circ$C and $T_2(0) = 20^\circ$C) in Figure 1 bear some resemblance to real Arctic mixed-phase clouds. Supercooled liquid overlies ice, ice and snow fall gradually, updrafts and entrainment through a moist inversion supply new cloud water, and multiple cloud layers form in areas of both stable and moist-adiabatic stratification (e.g., Morrison et al. 2012; Sedlar et al. 2011; Verlinde et al. 2013). The most serious deficiencies of the simulations with the Lin-Purdue scheme are that new clouds are too commonly mixtures of ice and liquid (see Appendix A), and that the scheme does not maintain enough supercooled liquid at low temperatures. Inspection of the WRF implementation of the Lin-Purdue scheme reveals that it cannot form any new supercooled liquid by condensation below $-25^\circ$C, and thus cannot simulate persistent mixed-phase clouds when the cloud layer cools below $-10^\circ$C – instead switching rapidly into the radiatively clear boundary layer state. Other schemes generally share this deficiency of maintaining too little supercooled liquid – or equivalently, a too-high glaciation temperature (see Appendix A) – similar to findings by Pithan et al. (2014) for global models. In broad terms, the WSM-6 (Video S2) and Goddard (Video S3) schemes do a poorer job than the Lin-Purdue scheme of representing mixed-phase clouds; they both lack the basic supercooled liquid-over-ice morphology. The WSM-6 scheme simulates much more cloud ice than any other scheme, and has a glaciation temperature only slightly below $0^\circ$ C.

The Thompson (Video S4) and Morrison (Video S5) schemes represent mixed-phase clouds better than the Lin-Purdue scheme – both are able to maintain a mixed-phase cloud layer that grows upward and persists for 2-3 days in the $T_2(0) = 0^\circ$C case. One of the primary papers documenting the Morrison scheme, in fact, applied it in WRF to a case study of Arctic mixed-phase stratus (Morrison and Pinto 2005). The Stony Brook (Video S6) scheme has nearly identical cloud ice and
cloud liquid physics to the Lin scheme, and thus has similar strengths and weaknesses in depicting
Arctic mixed-phase clouds.

Snapshots of domain-average soundings and cloud properties every 2 days during the cooling process also show the strong increase in cloudiness with warming of the initial state for the Lin-Purdue microphysics scheme (Figure 2). Whenever low-level liquid clouds are present, they prevent the development of the surface inversion characteristic of Arctic air, instead giving rise to a strong cloud-top inversion with temperatures increasing from the cloud top down to the surface (Figure 2). Consequently, the simulation with $T_2(0) = 20^\circ$C has a total surface air temperature decrease of 39.7°C over two weeks, compared to 52.2°C for $T_2(0) = 10^\circ$C and 57.3°C for $T_2(0) = 0^\circ$C. Comparing the evolution of surface longwave fluxes (Figure 3a) with the soundings in Figure 2 shows the correspondence between suppressed surface cooling rates and the presence of low-level liquid clouds. Figure 3c explicitly shows that warmer initial states have a more positive and more persistent surface cloud longwave forcing; for the warmest case, surface cloud longwave forcing increases until after day 10. Turbulent surface heat fluxes also regulate the total cooling rate under clear skies; once the cloud layer dissipates in each simulation, about half of the surface radiative cooling is offset by sensible heating of the surface (by warmer overlying air in the surface inversion; negative fluxes indicate warming of the surface in Figure 3b).

Synchronous spikes in surface longwave cooling (upward) and surface cloud radiative effect and condensate path (downward) appear prominently for the warmest simulation in Figure 3. These spikes occur due to quasi-periodic formation and upward propagation of a near-surface fog and stratus cloud layer, which repeatedly transforms into cumulus convection and dissipates, allowing the surface to cool and again form a fog layer (see Video S1; frames near and leading up to a spike at day 4, hour 21 provide a good example). The frequency of these spikes is $\sim$1-2 d$^{-1}$, but the timing, regularity, and amplitude varies substantially across microphysics schemes (see Videos
S2-S6). Although these spikes are an intriguing feature of the warmer simulations, they do not seem to play a key role in any of the main conclusions, and consequently we do not discuss them further.

Simulations with heterogeneous surface heat capacity share many of these features, with similar time series of surface energy balance, cloud radiative forcing, vertically integrated cloud condensate, and even similar timing of spikes in the warmest simulation (dash-dotted lines in Figure 3). An additional result from these simulations is that thick clouds not only suppress surface cooling, they also reduce the spatial variability in surface air temperature (Figure 3e). In clear-sky conditions, the surface cools less quickly where it has high heat capacity, and more quickly elsewhere. This differential surface cooling rate leads to 2-meter air temperatures that vary spatially by $\sim 4^\circ$C in the clear state, but only $\sim 0.5^\circ$C in the cloudy state. Increased duration of optically thick clouds in a warmer initial atmosphere thus allows not only for maintenance of warmer conditions overall, but also for better redistribution of heat within the domain, from locally warm areas to locally cold ones. A speculative consequence of this finding would be that inland water bodies – even if they are relatively shallow – could play more of a role in moderating continental temperatures in warmer climates because their heat could be redistributed nonlocally by advection in a deeper cloud-topped boundary layer, rather than being lost locally by radiation to space.

Surface air temperature evolution across the six microphysics schemes shows that warming the initial atmosphere leads to a strong reduction in the surface cooling rate over the two-week cooling period, especially over the first week (Figure 4a). Time series of the surface air cooling ($T_2(0) - T_2(t)$) averaged across microphysics schemes show a brief initial clear-sky cooling period (lasting well under a day), followed by progressively longer plateaus in surface air temperature for warmer initial states (Figure 4b). The no-CRF simulations show that surface air cooling would be much
less sensitive to the initial atmospheric temperature under radiatively clear skies (dash-dotted lines in Figure 4b).

Time series of cloud fraction (Figure 5) aid in understanding the large spread in 2-meter temperature across microphysics schemes shown in Figure 4a. We define the cloud fraction as the fraction of grid cells with cloud condensate path – both liquid and ice – greater than 20 g m$^{-2}$. Most microphysics schemes show a rapid transition from the opaquely cloudy to the radiatively clear state as time passes (Figure 5, also visible in Figure 3c,d). The greatest spread in surface air temperature across microphysics schemes occurs when some microphysics schemes still have optically thick cloud layers but others do not, and this time window occurs progressively later for warmer initial atmospheres – around days 1-4 for $T_2(0) = 0^\circ$C, days 4-10 for $T_2(0) = 10^\circ$C, and day 11 onward for $T_2(0) = 20^\circ$C (Figure 4a, Figure 5). For $T_2(0) = 10^\circ$C, the WRF 1-Moment, Goddard, and Morrison schemes all switch back into a sustained overcast or mostly-cloudy state from the clear state – for days 4-11, 8-14, and 10-12, respectively. These three transitions are each associated with a physically deep layer of tenuous ice cloud that exceeds our 20 g m$^{-2}$ criterion for cloud condensate path, but has a weak longwave radiative effect and cannot prevent development of a surface inversion. The depth of these optically thin ice-cloud layers is probably exaggerated by the lack of large-scale dynamics (e.g., subsidence), but optically thin ice clouds are a common occurrence in the Arctic winter boundary layer (e.g., Curry et al. 1996).

\section*{b. Time-mean cooling and cloud properties}

To synthesize many cooling time series such as those shown in Figure 4, we define a metric of time-mean surface air cooling, $\Delta T_2$ by

\[ \Delta T_2 = T_2(0) - \overline{T_2(t)}, \] (1)
where the overline on $T_2(t)$ indicates a time-mean over the two week simulation. Time-mean surface air cooling decreases nonlinearly with initial temperature, with relatively little spread across microphysics schemes (Figure 6a), and is much greater for the no-CRF simulations (dash-dotted line in Figure 6a). A least-squares fit of $\Delta T_2$ averaged over all microphysics schemes gives:

$$\Delta T_2 = 36.9 - 0.615 \times T_2(0) - 0.027 \times T_2(0)^2$$

and for the no-CRF simulations gives:

$$\Delta T_{2}^{\text{no-CRF}} = 40.4 - 0.239 \times T_2(0) - 0.0095 \times T_2(0)^2,$$

with $T_2(0)$ in °C. The slope $\gamma = -\frac{\partial \Delta T_2}{\partial T_2(0)}$ is a dimensionless feedback measure that indicates the extent to which the Arctic air formation process amplifies initial surface air temperature differences. In the mean across microphysics schemes, each degree of warming of the initial state leads to a decrease of time-mean surface air cooling by $\gamma = 0.615 + 0.054 \times T_2(0)$ degrees. For example, for the warmest case ($T_2(0) = 20$°C), each degree of initial warming implies a $\gamma \approx 1.7$ degree decrease in average cooling. In comparison, the no-CRF feedback is about 65% smaller: $\gamma_{\text{no-CRF}} = 0.239 + 0.019 \times T_2(0)$. Thus, cloud radiative effects alone reduce surface cooling by about a degree for each degree warming of $T_2(0)$ in the warmest cases that we simulate.

The time for the 2-meter temperature to reach freezing, $\tau_0$, remains well under a day for all microphysics schemes when $T_2(0)$ is less than 8°C. For a two-degree warming above 10° C (8°C for the Goddard microphysics scheme), however, the time to freezing very rapidly jumps to 4-6 days, then increases further to 10-12 days for $T_2(0) = 20$°C. The rapid jump and subsequent continued increase occur because of the shape of the cooling time series, and because the time to freezing is the time to cross a fixed threshold of 0°C (Figure 4a). When $T_2(0)$ is less than 10°C, the surface drops below freezing during the period of rapid clear-sky cooling; for the warmer cases, clouds form before the surface has the chance to cool to freezing, and the threshold is crossed.
during a period of slow cooling. The abrupt jump in $\tau_0$ near $T_2(0) = 10^\circ C$ is thus an indicator of the flatness of the cooling time series in Figure 4a.

Time-mean cloud condensate path – dominated by cloud liquid path, $\bar{q}_{cl}$ – increases with warming across all microphysics schemes (Figure 7a). Averaged across microphysics schemes, time-mean cloud liquid path increases from $\sim 10 \text{ g m}^{-2}$ at $T_2(0) = 0^\circ C$ to $\sim 50 \text{ g m}^{-2}$ at $T_2(0) = 10^\circ C$ to $\sim 125 \text{ g m}^{-2}$ at $T_2(0) = 20^\circ C$. The spread across microphysics schemes is large, greater than a factor of three at all initial temperatures – but all schemes nonetheless show large relative increases in cloud liquid path with warming. Although part of the increase occurs due to thickening of clouds in the opaquely cloudy state, most of this increase owes to an increasing fraction of the two-week period spent in the opaquely cloudy state (Figure 5). The increasing fraction of time spent in the cloudy state with warming can also be seen from Figure 7c, which shows the temperature-dependence of time-mean cloud fraction – defined as the fraction of grid cells for which total condensate path exceeds $20 \text{ g m}^{-2}$ and averaged across all time steps. The time-mean cloud fraction averaged across microphysics schemes increases from $\sim 10\%$ at $T_2(0) = 0^\circ C$ to $\sim 90\%$ at $T_2(0) = 20^\circ C$. Time-mean cloud ice path changes less consistently with warming than does cloud liquid path or cloud fraction (Figure 7b). Outliers include the five-fold increase in cloud ice path from $8 \text{ g m}^{-2}$ at $T_2(0) = 0^\circ C$ to $40 \text{ g m}^{-2}$ at $T_2(0) = 20^\circ C$ in the WSM-6 scheme, and values of $< 1 \text{ g m}^{-2}$ of cloud ice path with non-monotonic dependence on $T_2(0)$ in the Thompson scheme. The phase of cloud water that contributes to the cloud fraction (condensate path exceeding $20 \text{ g m}^{-2}$) is almost entirely liquid for the Lin-Purdue, Thompson, Morrison, and Stony Brook schemes, but ice as well for the Goddard scheme and especially the WSM-6 scheme. Comparing Figures 6 and 7, we see that the spread in two-week average cooling across microphysics schemes is linked to the spread in cloud fraction and time-mean cloud liquid path – with the Thompson and
Morrison schemes cooling least, the Goddard scheme close to the multi-microphysics mean, and
the Lin-Purdue, WSM-6, and Stony Brook schemes cooling the most.

Mean statistics averaged over time and across microphysics schemes are given in Table 1, for
simulations with $T_2(0)$ in $\{-10, -5, 0, 5, 10, 15, 20\}^\circ C$, and will be discussed in more detail below.

Note that the time-mean surface skin temperature $T_S$ is $\sim 3^\circ C$ colder than the time-mean surface
air temperature $T_2$ in the coldest simulations, because the strong surface inversion includes a near-
discontinuity between the surface and near-surface air (this surface temperature jump is weakened
by about a third in sensitivity tests where the roughness length is increased from $\sim 2 \times 10^{-5} m$ to
0.001 m). Evolution of the surface temperature jump $T_S - T_2$ closely follows the surface turbulent
heat flux, with slightly positive values in the opaquely cloudy state and negative values in the ra-
diatively clear state. Because the cloudy state dominates for higher values of $T_2(0)$, the difference
between surface skin temperature and surface air temperature is much reduced for the warmest
simulations. Thus, a cooling metric defined in terms of time-mean surface skin temperature would
be even more sensitive to the initial state than $\Delta T_2$.

4. Discussion

a. Comparison of cloud-resolving and single-column model results

The results above are both qualitatively and quantitatively consistent with the findings of Cronin
and Tziperman (2015), who used a single-column setup of WRF. A key conclusion of this paper is
to support the single-column model result that Arctic air formation is suppressed for a warm initial
maritime profile.

Nevertheless, it is useful to analyze in depth the differences between single-column and cloud-
resolving results, to determine where differences lie. Our Figure 6 shows in gray the multi-
microphysics mean curves from Figure 2 of Cronin and Tziperman (2015). For the cloud-resolving simulations presented here, time-mean surface cooling over two weeks is slightly reduced (by $\sim 1-3^\circ$C) and the feedback $\gamma$ is $3 - 8\%$ larger relative to the single-column results. The cooling curves are less flat for warm $T_2(0)$ in the simulations of Cronin and Tziperman (2015) as compared to the results found here, so the sharp jump in $\tau_0$ around $T_2(0)=10^\circ$C was not found in Cronin and Tziperman (2015). These differences could result from numerical issues such as changes in domain height and vertical level spacing, or from physical differences between the cloud-resolving and single-column models related to changes in clouds and turbulence.

To compare single-column and cloud-resolving results more directly, we have re-run single-column simulations with the same vertical level spacing (301 vertical levels, model top at 15 km) used here in all of our 2-dimensional simulations. Figure 8 shows differences in time-mean surface air temperature between cloud-resolving and single-column simulations, $T_{2D} - T_{2SCM}$, as well as differences in time-mean cloud condensate amount and time-mean cloud fraction. Differences are small at low $T_2(0)$ for all but the Morrison scheme, but generally grow for warmer initial atmospheric states (some schemes have greatest disagreement for $T_2(0) = 10 - 15^\circ$C). Cloud-resolving simulations are generally warmer near the surface than single-column simulations (Figure 8a), by $\sim 2^\circ$C in the mean across microphysics schemes for initial 2-meter temperatures above $10^\circ$C, and up to $5^\circ$C for the Morrison scheme. Surface air temperature differences between the cloud-resolving and single-column models also have strong time-dependence, with most warming at times when the cloud-resolving model still has optically thick clouds but the single-column model clouds have dissipated (not shown).

Warmer surface conditions in the cloud-resolving simulations are associated with a larger time-mean cloud liquid path and cloud fraction (Figure 8b,d). The difference in cloud fraction is $\sim 0.1$ in the mean across microphysics schemes for $T_2(0) > 10^\circ$C – with the cloud resolving model
maintaining cloud layers for considerably longer than the single-column model, especially in the
Thompson and Morrison schemes. The Goddard scheme is an outlier, in that it has less cloud liquid
and ice, and a smaller cloud fraction in the 2-dimensional simulations than in the single-column
simulations, yet it still maintains a modestly warmer surface at higher initial temperatures. The
pattern of overall differences in both temperature and cloud condensate between cloud-resolving
and single-column simulations indicates more active cloud-top turbulence and entrainment in the
cloud-resolving simulations, leading to cloud layers that grow more rapidly and ultimately become
deeper and condense more water than in the single-column simulations (this difference occurs for
all microphysics schemes at $T_2(0) = 20^\circ C$). Because the radiative cooling to space from lower-
tropospheric cloud tops is spread over a deeper layer, cooling is weaker at the surface but extends
to greater height, with additional warmth near the surface offset by colder conditions near the
highest cloud-top inversion. Thus, in spite of our initial expectation that a cloud-resolving model
might give more surface cooling because it allows for cloud layers with less than 100% cover, it
appears that the reverse occurs due to increasing cloud condensate, cloud fraction, and cloud layer
thickness.

The differences between cloud-resolving and single-column model simulations are relatively
small when compared with the overall changes to the process of Arctic air formation with climate
warming. The largest differences, however, occur for the Thompson and Morrison schemes, which
seem to give the most credible mixed-phase clouds. Using a 2-dimensional model rather than a
single-column model leads to a $\sim 50\%$ increase in $\gamma$ for both of these schemes for $T_2(0)$ of 0
to $10^\circ C$. Comparing 2-dimensional and single-column model results thus suggest that Arctic air
formation may be suppressed more rapidly with warming near the present climate than found in
Cronin and Tziperman (2015).
b. Vertical structure of cooling

As the initial atmospheric state is warmed, the surface cooling rate decreases, but the top-of-atmosphere outgoing longwave radiation increases. Reconciling these opposing tendencies requires that the cooling rate of the atmosphere must increase with $T_2(0)$ at some height above the surface. Plotting the profile of average cooling rate, we indeed see that cooling decreases with $T_2(0)$ at and just above the surface, but increases with $T_2(0)$ above the lowermost troposphere (Figure 9). The strongest reduction in surface cooling as $T_2(0)$ is increased occurs over the first week of the simulation (Figure 9a), with a non-monotonic response of surface cooling to $T_2(0)$ during the second week (Figure 9b). Colder initial atmospheres have a surface cooling rate of $\sim 4^\circ$C d$^{-1}$ over the two-week period and $\sim 5^\circ$C d$^{-1}$ during the first week, but the atmosphere above 850 hPa cools at $2^\circ$C d$^{-1}$ or less. The warmest cases ($T_2(0) = 20^\circ$C) have a cooling rate of about $2^\circ$C d$^{-1}$ that is relatively constant in time and height over a deep layer (up to approximately 300 hPa).

There is significant spread across microphysics schemes in the timing of cooling – as shown by the shaded areas in the plots of cooling broken down by week (Figure 9a,b) – but the total amount of cooling over the two-week period differs much less (Figure 9c). The mass-weighted average cooling rate of the atmosphere increases with $T_2(0)$, from $1.4^\circ$C d$^{-1}$ at $T_2(0) = 0^\circ$C to $1.8^\circ$C d$^{-1}$ at $T_2(0) = 20^\circ$C (Table 1). This cooling rate is an average temperature change from the sum of all physical processes simulated (radiation, condensation, convection), and although it is dominated by longwave radiative cooling, it should not be interpreted as a radiative cooling rate alone. Surface cooling rates in the no-CRF simulations show much less sensitivity to the initial atmospheric temperature, with $\sim 4 - 5^\circ$C d$^{-1}$ surface cooling over the first week and $\sim 2 - 3^\circ$C d$^{-1}$ surface cooling over the second week, and surface-amplified cooling over the whole two-week period (Figure 10). An upward shift and increase in magnitude of mid- and upper-tropospheric cooling
with warming of the initial atmosphere is seen in simulations with and without cloud-radiation
interactions (compare Figures 9 and 10 above 600 hPa).

The energy loss of the atmosphere and surface by radiation to space can be divided into contrib-
utions from changes in internal heat, latent, and gravitational potential energy of the atmosphere,
and changes in heat content of the surface (Table 1; angle brackets indicate mass-weighted vertical
integrals). Condensation \( L_v \Delta \langle q_v \rangle / \Delta t \) plays an increasingly important role in the column energy
budget at the warmest temperatures, amounting to \( \sim 40 \) W m\(^{-2}\) of heating in the warmest cases.
For extremely warm initial temperatures, the role of condensation alone in increasing the effective
heat capacity of the atmosphere could lead to a reduction in average tropospheric cooling rates with
increasing \( T_2(0) \) (Cronin and Emanuel 2013), but our simulations do not reach that limit, in part
because our initial soundings are rather dry. Note that changes in internal heat energy \( (c_v \Delta \langle T \rangle / \Delta t) \)
and gravitational potential energy \( \Delta \langle gz \rangle / \Delta t \) of the column do not sum to \( c_p \Delta \langle T \rangle / \Delta t \), because the
upper bound of integration is only \( \sim 15 \) km and the equality only holds for a full column integral
(e.g., Peixoto and Oort 1992, chapter 13). Loss of heat by the surface, \( C_S \Delta T_S / \Delta t \), supplies \( \lesssim 10 \) W
m\(^{-2}\) to the atmosphere; the magnitude of this term is limited because the surface heat capacity is
small by design \( (C_S = 2.1 \times 10^5 \) J m\(^{-2}\) K\(^{-1}\)).

c. Top-of-atmosphere radiation and feedbacks

To attempt to compare our results with the feedbacks diagnosed from global climate models
(e.g., Pithan and Mauritsen 2014), we can analyze how top-of-atmosphere outgoing longwave
radiation (OLR) depends on surface temperature. The main idea is similar to a climate model
feedback analysis: that the derivative of OLR with respect to \( T_S \) is a measure of the total radiative
feedback – or how strongly top-of-atmosphere energy loss tends to damp surface warming. Be-
cause our model for Arctic air formation is an idealized depiction of a complex transient process,
this analysis is not equivalent to a climate model feedback analysis, and we do not decompose
the sensitivity of OLR to surface temperature $T_S$ into specific feedbacks – e.g., Planck, lapse-rate,
water vapor, cloud (such a decomposition would also require additional radiative transfer calcula-
tions outside of WRF). We infer from this analysis that there are strong positive cloud and clear-sky
feedbacks associated with suppression of Arctic air formation, that make $d\text{OLR}/dT_S$ much smaller
in magnitude – less negative – for the time-mean atmospheric states (during the process of Arctic
air formation) as compared to the initial atmospheric states.

Figure 11a shows the surface temperature-dependence of three choices of OLR: initial OLR(0)
plotted against initial surface temperature, time-mean clear-sky $\overline{\text{OLR}}_{\text{clear}}$ plotted against time-
mean surface temperature, and time-mean all-sky $\overline{\text{OLR}}$ plotted against time-mean surface temper-
ature (values also given in Table 1). A blackbody with temperature 25 degrees less than the surface
is also plotted for reference, as an example with only the Planck feedback present. Denoting the
negative slope of OLR with respect to surface temperature as a feedback, we then also have three
different feedback metrics for our simulations, corresponding to each choice of OLR:

$$\lambda(0) = -\frac{d\text{OLR}(0)}{dT_S(0)}$$
$$\overline{\lambda}_{\text{clear}} = -\frac{d\overline{\text{OLR}}_{\text{clear}}}{dT_S}$$
$$\overline{\lambda} = -\frac{d\overline{\text{OLR}}}{dT_S}$$

(4)

These feedback parameters are plotted with negative sign, and represent a stabilizing total long-
wave feedback in all cases – a warmer atmosphere and surface lead to more OLR (Figure 11b).
The contrast in the feedback parameter, $\lambda$, and its temperature dependence, between the time-
mean and initial OLR values provides a measure of how the suppression of cold air formation
alters the local top-of-atmosphere longwave feedback. We wish to emphasize a few features of

Figure 11b:
The longwave feedback in the initial state, $\lambda(0)$, increases in magnitude from $-2.2 \text{ W m}^{-2} \text{K}^{-1}$ for the coldest initial temperatures to $-2.8 \text{ W m}^{-2} \text{K}^{-1}$ for the warmest – presumably due to negative Planck and lapse-rate radiative feedbacks which increase in strength with warming of the initial state.

The longwave clear-sky feedback in the time-mean state, $\lambda_{\text{clear}}$, has a near-constant value $\sim -1.7 \text{ W m}^{-2} \text{K}^{-1}$. The smaller magnitude of $\lambda_{\text{clear}}$ relative to $\lambda(0)$ – by $\sim 0.5 \text{ W m}^{-2} \text{K}^{-1}$ in the range of overlapping temperature – indicates that there are positive clear-sky OLR feedbacks associated with the suppression of Arctic air formation. We interpret this offset between $\lambda_{\text{clear}}$ and $\lambda(0)$ as primarily a positive lapse rate feedback, though we cannot rule out a water vapor feedback component.

The longwave all-sky feedback in the time-mean state, $\lambda$, decreases in magnitude from $-1.9 \text{ W m}^{-2} \text{K}^{-1}$ for the coldest initial temperatures to $-0.9 \text{ W m}^{-2} \text{K}^{-1}$ for the warmest. The smaller magnitude of the feedback $\lambda$ with warming is due to a positive top-of-atmosphere cloud longwave feedback that rapidly increases to $\sim 1 \text{ W m}^{-2} \text{K}^{-1}$ for the warmest initial conditions (compare $\lambda$ and $\lambda_{\text{clear}}$ in figure 11b).

The strong top-of-atmosphere cloud feedback under warm conditions shown in Figure 11b can also be inferred from time-mean top-of-atmosphere cloud radiative forcing, which is generally small compared to the surface cloud radiative forcing (and can even be negative), but which increases rapidly with warming for warm initial atmospheres (Table 1, last two rows). From a conventional top-of-atmosphere standpoint, clouds may thus act as a stronger positive feedback than was realized by Cronin and Tziperman (2015). Our results thus support the potentially important role of positive top-of-atmosphere longwave cloud feedback, particularly in much warmer climates (e.g., Abbot et al. 2009).
The surface in our model is treated very simply – essentially as a heat-storing slab that is dragged along with the atmosphere as it moves. By using identical values of $T_S(0)$ and $T_2(0)$, and giving the surface a relatively low heat capacity, we have tried to minimize biases in sensitivity $\gamma = -\frac{\partial \Delta T_2}{\partial T_2(0)}$ that might be related to the surface, although one clear limitation of this work is that properties of the surface do not change at all with temperature, even though major changes in real surface properties occur across the freezing point. Cronin and Tziperman (2015) found that a surface with higher heat capacity leads to smaller values of $\gamma$, whereas a surface with a lower heat capacity leads to a larger $\gamma$. We speculate that greater surface heat capacity will generally make surface temperatures less sensitive to changes in initial atmospheric state. This would have implications for previous work by Curry (1983), Pithan et al. (2014), and Pithan et al. (2016), who used a sea ice-like surface, with a prescribed thickness and temperature profile. Such a structural choice makes sense for exploring Arctic air formation in the present climate, but likely makes $\gamma$ smaller than in our work, because ice with a prescribed temperature that is independent of the initial 2-meter air temperature will conduct heat out of a warm initial atmosphere or into a cold initial atmosphere.

Our default model resolution $(\Delta x, \Delta z) = (100$ m, 50 m) has allowed us to run a large number of simulations across different microphysics schemes and temperatures, but may be insufficient to fully resolve the liquid layer at the top of Arctic mixed-phase clouds (e.g., Ovchinnikov et al. 2014), and the sharpness of inversions that form near stratus cloud tops (e.g., Blossey et al. 2013). Sensitivity tests with varied resolution using the Lin-Purdue scheme at $T_2(0)$ of 0 and 20°C show that altering the resolution by a factor of 2 typically leads to a change of a few W m$^{-2}$ in surface cloud forcing, $\sim 1$°C in time-mean surface air temperature, and that decreasing the resolution for
a cold initial state can completely suppress the low-level turbulence (Table 2). Increasing vertical resolution leads to surface warming for $T_2(0) = 20^\circ$C, but to surface cooling for $T_2(0) = 0^\circ$C. This reversal occurs because the cloudy state dominates in the warm simulation, but the clear state dominates in the cold simulation, and its sharp surface inversions are strengthened at high vertical resolution. The Thompson and Morrison schemes, which simulate better mixed-phase clouds at low temperature, might be more sensitive to resolution for cold initial states. Use of 2-dimensional rather than 3-dimensional geometry is also a limitation of this study; we originally imagined that the upscale cascade of turbulent kinetic energy in 2 dimensions might allow for larger-scale circulations to develop and disrupt layered clouds, but it is not clear to what extent this occurred in any of our simulations.

Another limitation of this work is the lack of large-scale subsidence or wind shear. Moderate subsidence in single-column simulations of Cronin and Tziperman (2015) was found to limit the upward growth of the cloud layer, but did little to modify the suppression of Arctic air formation in a warmer initial atmosphere; we anticipate similar sensitivity to subsidence in the cloud-resolving model used here. A stochastically varying vertical velocity profile that includes stronger ascent and stronger subsidence, as in Brient and Bony (2013), might provide a stronger and more realistic test of our mechanism than steady subsidence. Regarding wind shear, we expect more rapid horizontal advection of air aloft than near the surface, so realistic vertical shear would likely lead to less cooling in the mid- and upper-troposphere as an air mass traverses a continent than we have simulated here. The 2-week time period during which we follow the cold air formation process is also somewhat arbitrary, and should be informed by investigation of the observed time scales of continental traversal for air parcels in the lower troposphere, and projections of how these time scales may change with warming. Recent work by Woods and Caballero (2016) has suggested a link between observed Arctic warming and increasing frequency of injection events of very warm
and moist air into the polar region – a result consistent with the ideas presented in this paper. Questions about the role of large-scale dynamics – both vertical advection and the statistics of horizontal flow – point to investigation of cold air formation in a global climate model as a productive future research direction.

An additional limitation was noted above: in the version of WRF we use, the coupling between radiation and microphysics is incomplete. Only the total liquid and ice content in each column and vertical level are passed from the microphysics scheme to the radiation scheme, and the radiation scheme makes its own assumptions about cloud particle sizes. The spread across microphysics schemes shown here thus includes temperature-dependent differences in liquid condensate fraction and microphysical removal rates, but does not include the coupling between varying particle sizes and radiation. Investigation of cloud ice effective radii in the Thompson and Morrison scheme simulations (not shown) reveals large differences between the two schemes, but only weak sensitivity to $T_2(0)$ in each. Lack of temperature-dependence within each scheme is reassuring for our overall mechanism for suppression of Arctic Air formation, but the large difference between schemes indicates that we have underestimated the spread in cooling rates that would occur for complete coupling of radiation and microphysics. Finally, we do not include in our simulations any link between aerosols and clouds or precipitation interaction, although interactions between aerosol, cloud, and precipitation are likely important for determining cloud particle size and lifetime in the real Arctic (Mauritsen et al. 2011; Solomon et al. 2015).

5. Conclusions

We have used a 2-dimensional cloud-resolving configuration of the WRF model to investigate the idea that Arctic air formation is suppressed in warmer climates. Our results indicate that warming of the initial state leads to substantial inhibition of Arctic air formation via the development
of optically thick low-level liquid clouds that hamper surface radiative cooling, and are consistent with single-column model simulations in Cronin and Tziperman (2015). Two-week averaged cooling decreases by roughly 0.6 degrees for each degree warming of the initial state at initial air temperatures near 0°C, and this amplification factor rises for very warm initial air temperatures to as much as 1.7 degrees of reduced cooling for each degree warming of the initial state. These results are robust across several different cloud microphysics schemes, but suppression of cold air formation is stronger in the schemes that produce the most credible mixed-phase clouds, and also more sensitive in these schemes to the use of a 2-dimensional model rather than a single-column model. Low-level clouds in the 2-dimensional simulations grow more rapidly, extend to greater altitude, and persist longer as compared to analogous single-column simulations, which increases the strength of the cloud feedback. The greater altitude reached and faster growth of the lower-tropospheric cloud layer appear related to cloud-top entrainment and turbulence, which is not resolved in the single-column model.

This paper has found that resolving smaller-scale processes, especially convection and turbulence, strengthens the results of Cronin and Tziperman (2015). Many questions for future research on Arctic air formation relate to larger spatial scales, and the interaction of Arctic air formation with the general circulation of the mid- and high-latitude atmosphere. How does the upward shift in cooling during the formation of cold air affect mid- and high-latitude dynamics? What pressure levels in the Arctic atmosphere are most linked to extreme cold outbreaks in mid-latitudes? How robust is the low-cloud insulation mechanism to full 3-dimensional wind variability in a global model? What is the joint probability density function of age of terrestrial air and sea surface temperature last encountered, and how does this joint distribution change with global warming? Addressing these questions may offer new ways of understanding Arctic amplification and its coupling to changes in mid-latitude extreme weather.
Acknowledgments. This work was supported by a Harvard undergraduate PRISE fellowship (HL), NOAA Climate and Global Change Postdoctoral Fellowship and by the Harvard University Center for the Environment (TWC), and by the NSF climate dynamics program under grant AGS-1303604. ET thanks the Weizmann institute for its hospitality during parts of this work. We thank three anonymous reviewers for thorough and helpful comments that helped to improve the manuscript.

APPENDIX

Cloud microphysics schemes

We use six different microphysics schemes in this paper, denoted Lin-Purdue, WSM-6 (WRF single-moment 6-class), Goddard, Thompson, Morrison, and Stony Brook. Each is described in its own subsection below, emphasizing processes that produce and consume cloud ice (critical for understanding mixed-phase clouds). Most of the models are bulk schemes that use a single prognostic variable – mass mixing ratio – to model six classes of water: vapor, cloud liquid, cloud ice, rain, snow, and graupel, with exceptions noted below.

Previous work highlights the importance of liquid-ice partitioning as a function of temperature in determining the magnitude of low cloud optical thickness feedback over the Southern Ocean in global climate models (McCoy et al. 2015). The dependence of liquid cloud condensate fraction on temperature is an informative emergent property of both the microphysics scheme and the model setup (Figure A1), and has also been shown to be related to model biases in mixed-phase clouds (Pithan et al. 2014). As in McCoy et al. (2015), we denote the glaciation temperature for each scheme as \( T_{50-50} \), or the temperature at which half of cloud water is ice and half is liquid. Curves in Figure A1 were calculated by aggregating the liquid cloud condensate fraction for all model gridpoints with at least \( 10^{-4} \) g kg\(^{-1} \) of total cloud condensate among the simulations with
initial surface temperatures in \{-10, -5, 0, 5, 10, 15, 20\}°C into bins of width 1°C. The glaciation temperatures in our simulations are around 10°C higher than the observations of Hu et al. (2010) and the multi-model mean in McCoy et al. (2015), and we also find a more rapid liquid-ice transition with temperature than in observations (Figure A1). The implications of these biases are unclear; McCoy et al. (2015) suggest that higher glaciation temperature corresponds to a stronger cloud optical thickness feedback over the Southern Ocean, but clouds in our model setup often form at a much lower temperature than experienced in the lower troposphere over the Southern Ocean. We speculate that lowering the glaciation temperature of a specific scheme might lead to larger cloud optical thickness changes with warming at lower \(T_2(0)\) – perhaps shifting curves in Figures 6 and 7 to the left.

\textbf{a. Lin-Purdue scheme}

The Lin-Purdue scheme is described in Lin et al. (1983) and cloud liquid and cloud ice microphysics are described in more detail in Hsie et al. (1980). Cloud ice in the model is produced primarily by supersaturation adjustment, with the fraction of supersaturation partitioned into cloud ice varying linearly from 0 at 0°C to 1 at -25°C, and supersaturation calculated as a linear mix of that over liquid and ice with the same temperature-dependent weighting. Offline calculations show that the production of cloud ice from cloud liquid by depositional growth – an arrow shown in Figure 1 of Lin et al. (1983) – is entirely negligible. This simple linear ramp is responsible for the near-linear dependence of liquid condensate fraction on temperature in the Lin-Purdue scheme (linear ramp shown as a dotted black line in Figure A1), with \(T_{50-50} = -13.2°C\) just below the midpoint of the mixed-phase range. Liquid condensate fraction lies above this dotted line in Figure A1 for temperatures between about -10 and 0°C because the model implementation of the Berg-
eron process leads to more rapid removal of cloud ice than it does of cloud liquid – often leading
to supercooled liquid clouds that have snowed out all of their ice.

b. WSM-6 scheme

The WRF single-moment 6-class (WSM-6) scheme is described in Hong and Lim (2006), but
many essential details of the cloud ice scheme are presented in Hong et al. (2004). Cloud ice is
produced rapidly by transfer from the vapor phase, both onto existing particles and forming new
particles, even at temperatures just below freezing. If little existing ice present, then the scheme
produces ice by initiation of new particles, at a rate sufficient to deplete the supersaturation over
ice in ≲1500 s at all temperatures below freezing. If existing cloud ice is present, then vapor
deposition onto existing ice particles can be a faster process, especially at temperatures slightly
below freezing. This aggressive formation of cloud ice from vapor is responsible for the extremely
high glaciation temperature $T_{50-50} = -0.5^\circ C$ in the WSM-6 scheme. WSM-6 also maintains
clouds with the largest ice path, in part because autoconversion of cloud ice to snow has a high
threshold $0.08/\rho$ g kg$^{-1}$, but also because the parameterization of the Bergeron process does not
directly transfer mass from cloud ice to snow.

c. Goddard scheme

The Goddard scheme is described in Tao et al. (1989) and Tao et al. (2003). Cloud ice production
is dominated by initiation of new ice particles if concentrations are very low. Once the existing
cloud ice content becomes appreciable, cloud ice production is dominated by the parameterization
of the Bergeron process, which draws directly from cloud liquid, and has an increase in its rate
coefficient by a factor of $\sim 500$ as temperatures drop from -8$^\circ C$ to -15$^\circ C$. This rapid increase
in conversion rate from cloud liquid to cloud ice with cooling is likely responsible for the value
of the glaciation temperature $T_{50-50} = -12.2^\circ C$, as well as the sharpness of the transition from liquid to ice clouds in the Goddard scheme (Figure A1). The Goddard scheme has an even larger autoconversion threshold than the WSM-6 scheme for forming snow from cloud ice – 0.6 g kg$^{-1}$ – but it also has rapid transfer of mass from cloud ice to snow through the Bergeron process; the competition of these processes leads to it being the scheme with the second largest cloud ice path.

d. Thompson scheme

The Thompson scheme is described in Thompson et al. (2008), and simulates two-moment rain and cloud ice. The Thompson scheme also has the lowest glaciation temperature of $T_{50-50} = -17.5^\circ C$ – which is related both to its maintenance of supercooled water, and its very low cloud ice mixing ratios. Two-moment cloud ice makes it far more difficult to disentangle the dominant sources and sinks of cloud ice by offline coding of process rates. Simulated cloud ice production includes heterogeneous nucleation of new ice particles at ice supersaturation greater than 25% or temperatures less than $-12^\circ C$ and supersaturation over liquid, deposition from vapor onto existing particles, freezing of cloud liquid, freezing of small raindrops, and ice multiplication from rime splintering. The rate of cloud ice production from freezing of cloud liquid increases exponentially with supercooling and with the liquid droplet volume, following Bigg (1953).

e. Morrison scheme

The Morrison scheme is described in Morrison et al. (2009), and in more detail by Morrison and Pinto (2005), and has two-moment rain, snow, graupel, and cloud ice. Despite the high glaciation temperature of $T_{50-50} = -4.8^\circ C$, the scheme is nonetheless able to produce layers of supercooled liquid cloud below $-20^\circ C$, especially at low $T_2(0)$. As with the Thompson scheme, two-moment variables make it difficult to disentangle dominant sources and sinks of cloud ice. Simulated ice
production includes heterogeneous nucleation of new ice particles, deposition from vapor onto
existing particles, freezing of cloud liquid (similar to the Thompson scheme), accretion of cloud
liquid droplets by falling ice, and distinct ice multiplication pathways from cloud liquid-snow,
cloud liquid-graupel, rain-snow, and rain-graupel interactions.

f. Stony Brook scheme

The Stony Brook scheme is described in Lin and Colle (2011), and is a direct descendant of
the Lin-Purdue scheme. It has a near-identical treatment of cloud liquid and ice physics, and the
major difference from the Lin-Purdue scheme is that the prognostic graupel variable is replaced
by a continuous riming intensity variable for falling ice. The liquid condensate fraction curve is
almost identical to that of the Lin-Purdue scheme in Figure A1, with $T_{50-50} = -12.6 \degree C$.

References

Abbot, D. S., M. Huber, G. Bosquet, and C. Walker, 2009: High-co$_2$ cloud radiative forcing
feedback over both land and ocean in a global climate model. Geophysical Research Letters,

Alexeev, V. A., and C. H. Jackson, 2013: Polar amplification: is atmospheric heat transport im-

Bigg, E. K., 1953: The formation of atmospheric ice crystals by the freezing of droplets. Quarterly

Blossey, P. N., and Coauthors, 2013: Marine low cloud sensitivity to an idealized climate change:
The CGILS LES intercomparison. Journal of Advances in Modeling Earth Systems, 5, 234–258,


Table 1. Summary statistics for simulations with different initial temperatures averaged across microphysics schemes. Quantities prefaced with “time-mean” are averaged over the whole two-week period, and angle brackets indicate mass-weighted vertical integrals. 42

Table 2. Comparison of simulations with varied grid spacing. Metrics for comparison are the time-mean 2-m air temperature, $T_2$ ($^\circ$C), the time-mean surface cloud radiative forcing $\text{CRF}_S$ (W m$^{-2}$), and the time-mean standard deviation of the instantaneous zonal wind, averaged over levels between 1000 and 850 hPa, $\sigma_u$ (m s$^{-1}$). Values in parentheses indicate deviations from the respective control simulation. Simulations shown use the Lin-Purdue microphysics scheme, with initial 2-meter temperatures of $T_2(0) = 0^\circ$C (left-hand data block) and $T_2(0) = 20^\circ$C (right-hand data block). 43
TABLE 1. Summary statistics for simulations with different initial temperatures averaged across microphysics schemes. quantities prefaced with “time-mean” are averaged over the whole two-week period, and angle brackets indicate mass-weighted vertical integrals.

<table>
<thead>
<tr>
<th>Quantity</th>
<th>units</th>
<th>-10</th>
<th>-5</th>
<th>0</th>
<th>5</th>
<th>10</th>
<th>15</th>
<th>20</th>
</tr>
</thead>
<tbody>
<tr>
<td>Time-mean 2-m temperature, $T_2(t)$</td>
<td>°C</td>
<td>-50.2</td>
<td>-44.0</td>
<td>-36.9</td>
<td>-28.6</td>
<td>-17.5</td>
<td>-5.8</td>
<td>5.4</td>
</tr>
<tr>
<td>Time-mean surface temperature, $T_S(t)$</td>
<td>°C</td>
<td>-53.6</td>
<td>-47.1</td>
<td>-39.5</td>
<td>-30.7</td>
<td>-19.1</td>
<td>-6.8</td>
<td>5.0</td>
</tr>
<tr>
<td>Time-mean atmospheric cooling rate</td>
<td>°C d$^{-1}$</td>
<td>1.15</td>
<td>1.27</td>
<td>1.39</td>
<td>1.52</td>
<td>1.64</td>
<td>1.75</td>
<td>1.84</td>
</tr>
<tr>
<td>Time-mean condensation rate</td>
<td>mm d$^{-1}$</td>
<td>0.13</td>
<td>0.21</td>
<td>0.32</td>
<td>0.48</td>
<td>0.72</td>
<td>1.03</td>
<td>1.37</td>
</tr>
<tr>
<td>Change in atmospheric heat content, $c_v \Delta \langle T \rangle / \Delta t$</td>
<td>W m$^{-2}$</td>
<td>-88.8</td>
<td>-97.2</td>
<td>-106.3</td>
<td>-115.5</td>
<td>-124.5</td>
<td>-131.5</td>
<td>-136.9</td>
</tr>
<tr>
<td>Change in atmospheric potential energy, $\Delta \langle gz \rangle / \Delta t$</td>
<td>W m$^{-2}$</td>
<td>-18.2</td>
<td>-19.8</td>
<td>-21.3</td>
<td>-22.6</td>
<td>-23.1</td>
<td>-22.3</td>
<td>-19.7</td>
</tr>
<tr>
<td>Change in atmospheric latent energy, $L_v \Delta \langle q_v \rangle / \Delta t$</td>
<td>W m$^{-2}$</td>
<td>-3.8</td>
<td>-5.9</td>
<td>-9.2</td>
<td>-14.0</td>
<td>-20.8</td>
<td>-29.8</td>
<td>-39.7</td>
</tr>
<tr>
<td>Change in surface heat content, $C_S \Delta T_S / \Delta t$</td>
<td>W m$^{-2}$</td>
<td>-10.9</td>
<td>-10.8</td>
<td>-10.5</td>
<td>-10.0</td>
<td>-9.3</td>
<td>-8.2</td>
<td>-5.7</td>
</tr>
<tr>
<td>Time-mean energetic imbalance</td>
<td>W m$^{-2}$</td>
<td>2.65</td>
<td>2.22</td>
<td>1.66</td>
<td>0.99</td>
<td>0.27</td>
<td>-0.64</td>
<td>-0.71</td>
</tr>
<tr>
<td>Time-mean OLR</td>
<td>W m$^{-2}$</td>
<td>124.4</td>
<td>135.9</td>
<td>148.9</td>
<td>163.1</td>
<td>177.9</td>
<td>191.2</td>
<td>201.4</td>
</tr>
<tr>
<td>Initial OLR</td>
<td>W m$^{-2}$</td>
<td>192.6</td>
<td>203.9</td>
<td>215.5</td>
<td>227.7</td>
<td>240.4</td>
<td>253.9</td>
<td>268.1</td>
</tr>
<tr>
<td>Time-mean clear-sky OLR</td>
<td>W m$^{-2}$</td>
<td>120.8</td>
<td>131.9</td>
<td>144.7</td>
<td>159.7</td>
<td>179.2</td>
<td>200.4</td>
<td>221.4</td>
</tr>
<tr>
<td>Time-mean net longwave flux at surface</td>
<td>W m$^{-2}$</td>
<td>20.4</td>
<td>18.9</td>
<td>17.1</td>
<td>15.4</td>
<td>13.2</td>
<td>10.6</td>
<td>6.7</td>
</tr>
<tr>
<td>Time-mean net clear-sky longwave flux at surface</td>
<td>W m$^{-2}$</td>
<td>31.2</td>
<td>34.5</td>
<td>39.6</td>
<td>46.4</td>
<td>57.8</td>
<td>67.9</td>
<td>73.8</td>
</tr>
<tr>
<td>Time-mean surface sensible heat flux</td>
<td>W m$^{-2}$</td>
<td>-9.39</td>
<td>-8.17</td>
<td>-6.67</td>
<td>-5.45</td>
<td>-4.10</td>
<td>-2.62</td>
<td>-1.03</td>
</tr>
<tr>
<td>Time-mean surface latent heat flux</td>
<td>W m$^{-2}$</td>
<td>0.01</td>
<td>0.03</td>
<td>0.08</td>
<td>0.08</td>
<td>0.09</td>
<td>0.11</td>
<td>0.12</td>
</tr>
<tr>
<td>Time-mean top-of-atmosphere CRF</td>
<td>W m$^{-2}$</td>
<td>-3.5</td>
<td>-4.0</td>
<td>-4.1</td>
<td>-3.4</td>
<td>1.3</td>
<td>9.2</td>
<td>20.0</td>
</tr>
<tr>
<td>Time-mean surface CRF</td>
<td>W m$^{-2}$</td>
<td>10.8</td>
<td>15.7</td>
<td>22.5</td>
<td>31.1</td>
<td>44.6</td>
<td>57.4</td>
<td>67.1</td>
</tr>
</tbody>
</table>
TABLE 2. Comparison of simulations with varied grid spacing. Metrics for comparison are the time-mean 2-m air temperature, $T_2$ (°C), the time-mean surface cloud radiative forcing $\text{CRF}_S$ (W m$^{-2}$), and the time-mean standard deviation of the instantaneous zonal wind, averaged over levels between 1000 and 850 hPa, $\sigma_u$ (m s$^{-1}$). Values in parentheses indicate deviations from the respective control simulation. Simulations shown use the Lin-Purdue microphysics scheme, with initial 2-meter temperatures of $T_2(0) = 0$°C (left-hand data block) and $T_2(0) = 20$°C (right-hand data block).

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Grid Spacing ($\Delta x, \Delta z$)</th>
<th>$T_2(0) = 0$°C</th>
<th>$T_2(0) = 20$°C</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>$T_2$, °C</td>
<td>$\text{CRF}_S$, W m$^{-2}$</td>
</tr>
<tr>
<td>Control</td>
<td>(100 m, 50 m)</td>
<td>-38.83</td>
<td>15.34</td>
</tr>
<tr>
<td>5× horizontal</td>
<td>(500 m, 50 m)</td>
<td>-38.5 (0.28)</td>
<td>11.5 (-3.87)</td>
</tr>
<tr>
<td>2.5× horizontal</td>
<td>(250 m, 50 m)</td>
<td>-38.5 (0.33)</td>
<td>11.5 (-3.85)</td>
</tr>
<tr>
<td>1/2× horizontal</td>
<td>(50 m, 50 m)</td>
<td>-37.3 (1.54)</td>
<td>18.1 (2.81)</td>
</tr>
<tr>
<td>2× vertical</td>
<td>(100 m, 100 m)</td>
<td>-37.1 (1.68)</td>
<td>16.2 (0.90)</td>
</tr>
<tr>
<td>1/2× vertical</td>
<td>(100 m, 25 m)</td>
<td>-39.5 (-0.67)</td>
<td>16.1 (0.78)</td>
</tr>
</tbody>
</table>
LIST OF FIGURES

Fig. 1. Snapshots at day 4 of clouds as a function of height and x-coordinate, for simulations with initial 2-meter temperatures of a) 0°C, b) 10°C, and c) 20°C. Cloud liquid and ice content are plotted with a two-dimensional logarithmic scale inset on a): light to dark blues indicate cloud ice from 0.001 to 0.1 g kg⁻¹, light to dark greens indicate cloud liquid from 0.01 to 1 g kg⁻¹, and shades of cyan indicate mixed-phase clouds containing both liquid and ice. d) Domain-average temperature profiles show development of a strong surface-based inversion in the cold simulation, and deeper but less cold convective surface layers beneath cloud-top inversions for the warmer simulations. Video S1 shows time evolution of this Figure over the full 14 days of the three simulations. 

Fig. 2. Lower tropospheric temperature profiles as a function of pressure at 2-day intervals (black lines), for simulations using the Lin-Purdue microphysics scheme. Subplots are for initial 2-meter temperatures of a) 0°C, b) 10°C, and c) 20°C. Colored circles display information about the clouds at every other vertical level: circle radius is proportional to the logarithm of total cloud condensate, circle color indicates ice fraction in a linear scale from green (all liquid) to blue (all ice), and the fractional fill of a circle at a given level corresponds to the cloud fraction at that level. Grid cells are considered cloud if condensed water exceeds 0.001 g kg⁻¹, so small filled blue dots in a) and b) indicate very optically thin ice clouds that cover the full domain.

Fig. 3. Time series of surface energy balance in simulations with $T_2(0)$=0, 10, and 20°C, using the Lin-Purdue microphysics scheme. Solid lines show results from normal Lin-Purdue simulations with uniform surface heat capacity, and dash-dotted lines show results from simulations with heterogeneous heat capacity. a) Shows net longwave cooling of the surface, b) shows the total turbulent heat flux out of the surface (sensible plus latent), c) shows the longwave cloud forcing at the surface, d) shows the cloud condensate path for the entire depth of the troposphere (note log scale in d), and e) shows spatial variability in 2-meter air temperature, quantified as the difference between the maximum and minimum 2-meter air temperatures at each time. Surface cloud longwave forcing in c) is defined as the difference between the surface net longwave fluxes in hypothetical clear-sky and all-sky cases. In d), filled markers show the points at which cloud water first crosses thresholds of >10% ice (circles), >50% ice (triangles), and >90% ice (squares).

Fig. 4. Time series of a) 2-meter air temperature and b) cumulative 2-meter cooling ($T_2(0) - T_2(t)$), for initial 2-meter temperatures of -10°C to 20°C at 5°C increments, as labeled. The solid colored lines in both a) and b) represent an average over all cloud microphysics schemes; shaded regions in a) indicate the variation among different microphysics schemes, excluding the no-CRF case ($\pm$ one standard deviation). Dash-dotted lines in b) show cooling time series for the no-CRF simulation at each initial temperature.

Fig. 5. Timeseries of cloud fraction in each microphysics scheme over 2-week simulations for $T_2(0)$=0, 10, and 20°C. Cloud fraction is defined as the fraction of grid cells with a vertically integrated condensed water path of 20 g m⁻² or greater, including both liquid and ice.

Fig. 6. a) Average 2-meter cooling over 2 weeks, defined as the difference between the initial 2-meter temperature and the 2-meter temperature averaged over the 2 week period, and b) the number of days for the 2-meter temperature to drop to freezing. Both quantities are plotted for a range of initial 2-m air temperatures from -10°C to 20°C in 2°C increments, and several cloud microphysics schemes. The mean across microphysics schemes is shown in black, and the same quantity from Cronin and Tziperman (2015) is shown in gray for comparison.
Fig. 7. a) Time-mean cloud liquid path, $\overline{q_{cl}}$ (g m$^{-2}$) and b) time-mean cloud ice path, $\overline{q_{ci}}$ (g m$^{-2}$), each shown across the set of microphysics schemes and values of initial 2-meter temperature $T_2(0)$. c) The time-mean cloud fraction, defined as the fraction of grid cells for which total condensate path -- both liquid and ice -- exceeds 20 g m$^{-2}$, averaged across all time steps. Each panel also shows the mean across microphysics schemes as a thick black line, and a) shows a quadratic fit to the multi-microphysics mean of the cloud liquid path.

Fig. 8. Difference in surface air temperature and cloud statistics between 2-dimensional simulations and analogous single-column model simulations as a function of the initial temperature, quantified by a) time-mean 2-meter temperature difference $T_{2D} - T_{2SCM}$ (°C), b) time-mean cloud liquid path difference $\overline{q_{cl2D}} - \overline{q_{clSCM}}$ (g m$^{-2}$), c) time-mean cloud ice path difference $\overline{q_{ci2D}} - \overline{q_{ciSCM}}$ (g m$^{-2}$), and d) time-mean cloud fraction difference (2D minus SCM). Mean across microphysics schemes is shown as a thick black line in each subplot.

Fig. 9. Vertical profiles of total cooling rate averaged across microphysics schemes, as a function of initial 2-meter air temperature for the first week (a), the second week (b), and both weeks (c). Solid lines represent averages across microphysics schemes, and the shaded area indicates the spread across microphysics schemes (± 1 standard deviation).

Fig. 10. Vertical profiles of total cooling rate for clear-sky radiation simulations (no-CRF), as a function of initial 2-meter air temperature for the first week (a), the second week (b), and both weeks (c).

Fig. 11. a) Outgoing longwave radiation (OLR) as a function of surface temperature for initial conditions (OLR(0)), time-mean values (OLR), and time-mean clear-sky values (OLR$_{clear}$). b) Feedbacks based on slopes of OLR [from a)] with respect to $T_S$. Progressively less negative values of $\lambda_{clear}$ and $\lambda$ relative to $\lambda(0)$ in b) indicate positive cloud and clear-sky feedbacks associated with suppression of Arctic air formation from warmer initial states. A blackbody with temperature 25 degrees lower than the surface, $(T_S - 25)$, is shown in dotted lines to provide reference values for a) OLR and b) the Planck feedback by itself.

Fig. A1. Dependence of liquid condensate fraction on temperature (as in McCoy et al. 2015), taken from simulations with $T_2(0)$ in $\{-10, -5, 0, 5, 10, 15, 20\}$°C and averaged over all time intervals and vertical levels where total cloud condensate is greater than $10^{-4}$ g kg$^{-1}$. The fit to global satellite observations from Hu et al. (2010) equations (1) and (2) is also shown as a thick gray line.
Fig. 1. Snapshots at day 4 of clouds as a function of height and x-coordinate, for simulations with initial 2-meter temperatures of a) 0°C, b) 10°C, and c) 20°C. Cloud liquid and ice content are plotted with a two-dimensional logarithmic scale inset on a): light to dark blues indicate cloud ice from 0.001 to 0.1 g kg\(^{-1}\), light to dark greens indicate cloud liquid from 0.01 to 1 g kg\(^{-1}\), and shades of cyan indicate mixed-phase clouds containing both liquid and ice. d) Domain-average temperature profiles show development of a strong surface-based inversion in the cold simulation, and deeper but less cold convective surface layers beneath cloud-top inversions for the warmer simulations. Video S1 shows time evolution of this Figure over the full 14 days of the three simulations.
Fig. 2. Lower tropospheric temperature profiles as a function of pressure at 2-day intervals (black lines), for simulations using the Lin-Purdue microphysics scheme. Subplots are for initial 2-meter temperatures of a) 0°C, b) 10°C, and c) 20°C. Colored circles display information about the clouds at every other vertical level: circle radius is proportional to the logarithm of total cloud condensate, circle color indicates ice fraction in a linear scale from green (all liquid) to blue (all ice), and the fractional fill of a circle at a given level corresponds to the cloud fraction at that level. Grid cells are considered cloud if condensed water exceeds 0.001 g kg⁻¹, so small filled blue dots in a) and b) indicate very optically thin ice clouds that cover the full domain.
Fig. 3. Time series of surface energy balance in simulations with \( T_2(0) = 0, 10, \) and 20°C, using the Lin-Purdue microphysics scheme. Solid lines show results from normal Lin-Purdue simulations with uniform surface heat capacity, and dash-dotted lines show results from simulations with heterogeneous heat capacity. a) Shows net longwave cooling of the surface, b) shows the total turbulent heat flux out of the surface (sensible plus latent), c) shows the longwave cloud forcing at the surface, d) shows the cloud condensate path for the entire depth of the troposphere (note log scale in d), and e) shows spatial variability in 2-meter air temperature, quantified as the difference between the maximum and minimum 2-meter air temperatures at each time. Surface cloud longwave forcing in c) is defined as the difference between the surface net longwave fluxes in hypothetical clear-sky and all-sky cases. In d), filled markers show the points at which cloud water first crosses thresholds of >10% ice (circles), >50% ice (triangles), and >90% ice (squares).
Fig. 4. Time series of a) 2-meter air temperature and b) cumulative 2-meter cooling ($T_2(0) - T_2(t)$), for initial 2-meter temperatures of -10°C to 20°C at 5°C increments, as labeled. The solid colored lines in both a) and b) represent an average over all cloud microphysics schemes; shaded regions in a) indicate the variation among different microphysics schemes, excluding the no-CRF case (± one standard deviation). Dash-dotted lines in b) show cooling time series for the no-CRF simulation at each initial temperature.
Fig. 5. Timeseries of cloud fraction in each microphysics scheme over 2-week simulations for $T_2(0) = 0$, 10, and 20°C. Cloud fraction is defined as the fraction of grid cells with a vertically integrated condensed water path of 20 g m$^{-2}$ or greater, including both liquid and ice.
Fig. 6. a) Average 2-meter cooling over 2 weeks, defined as the difference between the initial 2-meter temperature and the 2-meter temperature averaged over the 2 week period, and b) the number of days for the 2-meter temperature to drop to freezing. Both quantities are plotted for a range of initial 2-m air temperatures from -10°C to 20°C in 2° increments, and several cloud microphysics schemes. The mean across microphysics schemes is shown in black, and the same quantity from Cronin and Tziperman (2015) is shown in gray for comparison.
Fig. 7. a) Time-mean cloud liquid path, $\overline{q_{cl}}$ (g m$^{-2}$) and b) time-mean cloud ice path, $\overline{q_{ci}}$ (g m$^{-2}$), each shown across the set of microphysics schemes and values of initial 2-meter temperature $T_2(0)$. c) The time-mean cloud fraction, defined as the fraction of grid cells for which total condensate path – both liquid and ice – exceeds 20 g m$^{-2}$, averaged across all time steps. Each panel also shows the mean across microphysics schemes as a thick black line, and a) shows a quadratic fit to the multi-microphysics mean of the cloud liquid path.
FIG. 8. Difference in surface air temperature and cloud statistics between 2-dimensional simulations and analogous single-column model simulations as a function of the initial temperature, quantified by a) time-mean 2-meter temperature difference $T_{2D} - T_{2SCM}$ ($^\circ$C), b) time-mean cloud liquid path difference $\hat{q}_{cl}^{2D} - \hat{q}_{cl}^{SCM}$ (g m$^{-2}$), c) time-mean cloud ice path difference $\hat{q}_{ci}^{2D} - \hat{q}_{ci}^{SCM}$ (g m$^{-2}$), and d) time-mean cloud fraction difference (2D minus SCM). Mean across microphysics schemes is shown as a thick black line in each subplot.
FIG. 9. Vertical profiles of total cooling rate averaged across microphysics schemes, as a function of initial 2-meter air temperature for the first week (a), the second week (b), and both weeks (c). Solid lines represent averages across microphysics schemes, and the shaded area indicates the spread across microphysics schemes (± 1 standard deviation).
Fig. 10. Vertical profiles of total cooling rate for clear-sky radiation simulations (no-CRF), as a function of initial 2-meter air temperature for the first week (a), the second week (b), and both weeks (c).
FIG. 11. a) Outgoing longwave radiation (OLR) as a function of surface temperature for initial conditions (OLR(0)), time-mean values (OLR), and time-mean clear-sky values (OLR\text{clear}). b) Feedbacks based on slopes of OLR [from a)] with respect to \( T_S \). Progressively less negative values of \( \lambda_{\text{clear}} \) and \( \lambda \) relative to \( \lambda(0) \) in b) indicate positive cloud and clear-sky feedbacks associated with suppression of Arctic air formation from warmer initial states. A blackbody with temperature 25 degrees lower than the surface, \( (T_S - 25) \), is shown in dotted lines to provide reference values for a) OLR and b) the Planck feedback by itself.
Fig. A1. Dependence of liquid condensate fraction on temperature (as in McCoy et al. 2015), taken from simulations with $T_2(0)$ in $\{-10, -5, 0, 5, 10, 15, 20\}^\circ$C and averaged over all time intervals and vertical levels where total cloud condensate is greater than $10^{-4}$ g kg$^{-1}$. The fit to global satellite observations from Hu et al. (2010) equations (1) and (2) is also shown as a thick gray line.