Low clouds suppress Arctic air formation and amplify high-latitude continental winter warming

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Low clouds suppress Arctic air formation and amplify high-latitude continental winter warming

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High latitude continents have warmed much more rapidly in recent decades than the rest of the globe, especially in winter, and the maintenance of warm, frost-free conditions in continental interiors in winter has been a long-standing problem of past equable climates. We use an idealized single-column atmospheric model across a range of conditions to study the polar-night process of air mass transformation from high-latitude maritime air, with a prescribed initial temperature profile, to much colder high-latitude continental air. We find that a low-cloud feedback – consisting of a robust increase in the duration of optically thick liquid clouds with warming of the initial state – slows radiative cooling of the surface and amplifies continental warming. This low-cloud feedback increases the continental surface air temperature by roughly two degrees for each degree increase in the initial maritime surface air temperature, effectively suppressing Arctic air formation. The time it takes for the surface air temperature to drop below freezing increases nonlinearly to over 10 days for initial maritime surface air temperatures of 15-20°C. These results, supplemented by an analysis of CMIP5 climate model runs which shows large increases in cloud water path and surface cloud longwave forcing in warmer climates, suggest that the “lapse rate feedback” in simulations of anthropogenic climate change may be related to the influence of low clouds on the stratification of the lower troposphere, and also indicate that optically thick stratus cloud decks could help to maintain frost-free winter continental interiors in equable climates.

Significance

Future greenhouse simulations, and evidence of frost-intolerant species in high latitude continental interiors during past equable climates, show significantly amplified warming at high latitudes over land in winter, whose physical mechanisms are still not understood. We show that the process of Arctic air formation, in which a high-latitude maritime air mass is advected over a continent, cooled at the surface and transformed into a much colder continental polar air mass, may change dramatically and even be suppressed in warmer climates due to an increase in the duration of optically-thick low clouds. This leads to two degree warming over the continent in response to a each degree warming over the nearby ocean, possibly explaining both past and future continental warming.

Reserved for Publication Footnotes
a direct prediction of the lapse rate and how it changes with climate [16].

The goal of this paper is to suggest that both the high-latitude lapse rate feedback, and winter continental warmth in equable climates, may be tied to the process of very cold continental air mass formation during winter. It has been shown that lower-tropospheric mixed-phase clouds play a critical role in the formation of Arctic air [22]. We show that Arctic air formation may be suppressed in warmer climates, leading to significant continental warming during winter. Specifically, we perform simulations with a single-column model of an air mass that begins over the ocean and is advected over high-latitude land. We find that increased lifetime of low-level liquid clouds with warming of the initial atmospheric state leads to a slower surface cooling rate and to a less stably stratified lower troposphere, consistent with the lapse rate feedback in climate models. We also analyze CMIP5 climate model results and show that aspects of Arctic amplification in such models are consistent with the proposed mechanism.

Results

Following studies that have used single-column models to gain insight into how high-latitude maritime air is cooled at the surface and transformed into polar continental air [23, 22, 24, 25, 26], we take a Lagrangian perspective of the cooling that occurs as a column of air moves from high-latitude open ocean, or from lower latitudes, into a region of the Arctic during polar night. We prescribe the initial vertical temperature and humidity profiles of an atmospheric air column, and allow it to evolve for two weeks in the absence of solar heating and over a very low heat capacity surface representing land, snow, or sea ice. To robustly explore the role of clouds, we use several different cloud microphysical parameterizations. Further details of the model setup are provided in the Materials and Methods section.

Figure 1A shows the cooling of the atmosphere for a reference simulation with the initial 2-meter atmospheric temperature \( T(0) = 0 \)°C, corresponding to present-day high-latitude ocean surface conditions. Snapshots of the vertical profile of temperature and clouds are shown every 2 days over a 14-day period. Initial cooling and condensation near the surface lead to an optically thick mixed-phase cloud deck near 950 hPa at day 2 (green line), which gives way to a much deeper but optically thin ice cloud layer by day 6 (thinner blue lines). This layer persists for the rest of the 2-week period, slightly slowing surface cooling, but not preventing development of a strong surface-based inversion. Purple lines in Figure 1C show the evolution of both the 2-m and 850-hPa air temperature; we see that for most of the two-week period, after the mixed-phase cloud layer dissipates, the atmosphere is warmer at 850 hPa than at the surface, by 10-15°C. This simulation illustrates the key features of Arctic air formation in the present-day climate [22, 26]. As explained by [22], the clouds in this situation do little to reduce the cooling of the surface itself, and also facilitate the direct cooling of the lower troposphere to space. The process of surface cooling is qualitatively different for a warmer initial state, with \( T(0) = 20 \)°C (motivated by past equable climates, e.g., [2]). Initial cooling leads to a thick liquid fog and stratus cloud layer that forms by day 2, moves upwards to 800 hPa by day 6, and persists until the last two days of the simulation (Figure 1B). The surface air temperature does not fall below freezing until day 9 (orange line in Figure 1A), which is 2 days later than for the 850-hPa temperature, only by a few degrees at the end of the simulation. Although the warm initial state has higher stability, its stability after two weeks of cooling is much lower than that of the reference simulation initial state, and a surface-based inversion develops only towards the end of the simulation. The final surface air temperature in the warm simulation is ~40°C warmer than in the reference simulation despite an initial surface warming of only 20°C relative to the reference simulation (Figure 1C).

Weakened radiative cooling at the surface due to the persistent cloud layer thus amplifies the initial surface air warming by a factor of roughly two. These dramatic results amount to a suppression of Arctic air formation in a much warmer climate.

The importance of clouds can be seen by comparing net longwave surface cooling rates with and without clouds (Figure 1D). The difference between the solid and dashed purple lines, corresponding to the simulation with \( T(0) = 0 \)°C, indicates that clouds only weakly influence the surface cooling, except in the brief period between days 1 and 2 when a thick mixed-phase stratus layer forms. The difference between the solid and dashed orange lines, corresponding to the \( T(2) = -20 \)°C simulation, is larger and more persistent, indicating that clouds reduce the rate of surface cooling by about 5°C during the entire duration of the simulation. The influence of initial temperature can also be seen in plots of the vertically-integrated cloud liquid and ice amounts; the warm initial state develops and retains more liquid water in clouds (Figure 1E).

The reduced rate of cooling in response to higher initial temperature \( T(0) \) is robust with respect to the microphysics scheme used, as seen in the difference between the initial temperature and the time-mean 2-m air temperature over the duration of the simulation, \( \Delta T = T(2) - T(Figure 2A) \). The average surface cooling across microphysics schemes for \( T(0) = 0 \)°C is \( \Delta T = 38^\circ \)C, and is reduced by \( 21 \)°C to \( \Delta T = 17^\circ \)C for \( T(0) = 20 \)°C. The suppression of Arctic air formation thus amplifies warming of the initial atmospheric state by over a factor of two.

The time taken for the 2-m air temperature to drop below freezing is less than 0.5 days if \( T(2) < 10 \)°C, but rises steeply thereafter, to \( \approx 10 \) days for \( T(2) = 20 \)°C (Figure 2B). This is a consequence of the differential surface cooling rates under clear and cloudy skies; the surface initially cools rapidly under clear skies, but cools much more slowly once clouds form, with a temperature plateau for many days (e.g., solid orange line in Figure 1C). For \( T(2) < 10 \)°C the surface drops below freezing before clouds form; for \( T(2) > 10 \)°C, the surface drops below freezing after clouds form.

Sensitivity tests allow us to decompose the reduced rate of cooling into contributions from the radiative effects of clouds, as well as the latent heat release that accompanies cooling and condensation, and the dependence of clear-sky longwave radiation on the initial atmospheric state. The dash-dotted line marked “no microphysics” in Figure 2A indicates the cooling that takes place in simulations where there is no phase change of water allowed, and thus no cloud formation or latent heat release. The modestly reduced cooling of the “no microphysics” case at higher \( T(2) \) owes to the decrease in clear-sky surface radiative cooling with higher atmospheric temperature (also visible in Figure 1D when comparing formal net longwave cooling rates). The dash-dotted line marked “no CRF” in Figure 2A shows the cooling that takes place when phase change of water is allowed, but clouds have no effect on radiative transfer calculations. The difference between the “no microphysics” and “no CRF” simulations thus indicates that the influence of latent heat release on the reduction of surface cooling is only \( \approx 3 \)°C at \( T(2) = 20 \)°C. The large difference between the “no CRF” dash-dotted line and the set of solid lines, including the black multi-microphysics mean line, shows that cloud-radiation interactions dominate the reduced cooling.
ing with warmer $T_2(0)$. Furthermore, simulations with the Kessler microphysics scheme, which represents clouds as only liquid regardless of temperature (dash-dotted line labeled “no ice”), show reduced cooling for all $T_2(0)$, and also a weaker sensitivity of $\Delta T_2$ to $T_2(0)$. These sensitivity tests demonstrate that most of the simulated reduction in cooling arises from the radiative effects of clouds, and relates to a change in the phase of cloud particles from ice to liquid.

Consider next the role of CO$_2$, first by allowing its concentration to vary, with $T_2(0)$ held constant. Taking the mean across microphysics schemes, we find that each doubling of CO$_2$ leads to a modest 0.56-0.8$^\circ$C increase in the two-week average temperature, $T_2$, with largest effect for $T_2(0) \approx 10^\circ$C (Figure SI-1). Next, specifying a doubling of CO$_2$ along with each 4$^\circ$C increase in $T_2(0)$, we find that the fraction of warming due to clear-sky processes increases (compare “no microphysics” curves in Figures 2A and SI-2A), yet the warming is similar to that obtained by only changing the initial temperature $T_2(0)$. These results indicate that the direct influence of changes in CO$_2$ is small compared to that of changes in clouds.

The warming of the surface by low clouds is weakened for a drier initial atmospheric condition (Figure SI-3), and strengthened for a moister initial atmospheric condition (Figure SI-4); the influence of the initial relative humidity profile is modest, however, because the large decrease in near-surface temperature always leads to supersaturation and cloud formation at some point in the cooling process. Our main results also seem robust to the inclusion of subsidence, which limits the upward growth of the cloud deck for warmer initial conditions (Figure SI-5), and weakens but does not eliminate the role of clouds in suppressing Arctic air formation as the initial state is warmed (Figure SI-6). Simulations with smaller and larger surface heat capacities (Figures SI-7 and SI-8, respectively) show that the sensitivity of $\Delta T_2$ to $T_2(0)$ is larger for a more insulating (lower heat capacity) surface, because inversions at low temperatures corresponding to the present climate are even stronger in such situations.

Output from CMIP5 [27] climate model simulations is consistent with our findings that climate warming leads to more low clouds over land at high latitudes during winter, and that these changes in clouds contribute to amplified warming. We compare boreal winter (DJF) multi-model mean changes between a historical period from 1980-1999 and a future projection for the strongly forced RCP8.5 scenario [27] from 2080-2099 (models used are listed in the materials and methods section). Figure 3 shows that maximum changes in surface air temperature, longwave cloud radiative effect on the surface energy balance, and cloud liquid condensate path occur over the Arctic ocean and high latitude land. The boreal winter surface cloud longwave radiative effect decreases in most regions, but increases over high-latitude land and the Arctic ocean, by an amount comparable to or greater than the forcing from increased greenhouse gas concentrations (Table 1). These large increases in liquid water content and surface longwave cloud radiative effect over the winter Arctic are consistent with suppression of Arctic air formation by low clouds in a warmer world. The findings in Figure 3 and Table 1 extend related previous work [28, 29, 30, 31, 11, 32] by focusing on surface longwave cloud changes in winter, across a set of models, rather than looking at annual-mean or top-of-atmosphere radiative changes, or surface changes in only a single model.

**Discussion**

With regard to explaining warm continental interiors in equable climates, the time-to-freezing metric from Figure 2B can be compared to the time it takes for an air mass to traverse a continent. An air mass moving at 5 m/s moves 4,320 km in 10 days, which is comparable to the east-west width of North America. Figure 2B implies that maintaining frost-free conditions a few thousand kilometers downwind of a warm ocean, even in polar night, can be accomplished with the aid of low clouds. Thus, the long-standing problem of explaining how crocodiles and palm trees could survive winters in the Eocene in Wyoming [1], 2000 km east of the moderating effect of the Pacific ocean, may be resolved by our finding of suppressed Arctic air formation in warmer climates, given a sea-surface temperature $\approx 20^\circ$C at 45$^\circ$N that is consistent with proxy evidence [2]. Furthermore, we have assumed polar-night, and thus zero insolation, but the midlatitudes see insolation $\sim 120$ W m$^{-2}$ at 45$^\circ$N even at the winter solstice, which would delay surface freezing for a longer period of time. Cold air outbreaks in Wyoming may owe to advection of cold air from higher latitudes, rather than cooling of air that flows directly east from the Pacific, but we have also shown that such higher-latitude air will likely be considerably warmer at the surface due to the role of clouds in suppressing Arctic air formation with warming. The findings in Figure 3 and Table 1 extend previous work [27, 28, 29, 30] by focusing on cloud radiative effect changes largely due to changes in high clouds, while surface cloud radiative effect changes are driven largely by changes in low clouds. Our single-column model setup largely precludes the formation of upper-level clouds because the initial condition is taken to be dry in the upper troposphere. The clouds in our single-column model simulations thus consistently have little influence on the top-of-atmosphere energy budget, yet still lead to a dramatically reduced surface cooling. Our examination of CMIP5 model
results for high-latitude land in winter shows that changes in the cloud radiative effect are smaller by about 40% at the top of the atmosphere as compared to the surface (Table 1), indicating an important role for low clouds, consistent with the column model results.

A seasonal perspective is also important when interpreting Arctic cloud changes. Increases in Arctic cloud fraction and liquid water path may lead to a negative annual-mean climate feedback at the top of the atmosphere. This is a consequence of both negative shortwave feedbacks in summer offsetting year-round positive longwave feedbacks, and clouds in the Arctic not increasing in height as much as clouds in other places, which reduces the top-of-atmosphere longwave cloud feedback [31]. Such a seasonal cycle of surface temperature at high latitudes is in line with proxy evidence of equable climates.

An even stronger warming over land than found in our column model may be expected when accounting for additional factors not considered here. These factors include the shorter time spent by air column over low heat capacity surfaces due to reduced winter sea ice cover in warmer climate, increases in surface heat capacity over land due to reduced snow cover, and CO2 increase.

Conclusions
We have analyzed the sensitivity of a cold air formation using a single column model and three dimensional climate model output. Using the column model, we prescribed the initial conditions and the atmosphere corresponding to an air column starting over a high-latitude open ocean, and allowed it to evolve for two weeks in the absence of solar heating over a low heat capacity surface corresponding to high latitude land, snow or sea ice. We find that with a warmer initial conditions corresponding to a warmer ocean surface, cooling is strongly suppressed due to the increasing lifetime and optical thickness of low clouds. These low clouds distribute the cooling over a deeper layer of the atmosphere, and thus amplify the warming at the surface relative to that aloft, which also suggests low clouds as an important potential cause of lapse-rate changes seen in simulations of future CO2-induced warming.

Large changes in cloud radiative effect on the surface energy budget occur in coupled climate models due to cloud phase and microphysics changes [34], temperature effects on adiabatic water content [35, 36, 37], and changes in large-scale moisture convergence and evaporation [29]. Furthermore, observations of internal variability and trends indicate that a warmer Arctic will also likely be cloudier in winter [38, 39]. Taken along with our proposed effects of low clouds on polar air formation and other low cloud effects in warmer climates [32], these lines of evidence point to a robust feedback between low clouds and high-latitude winter warming. Despite having only a weak influence on top-of-atmosphere radiative balance, low clouds at high latitudes may play a key role in equable climates, as well as in current and future climate change.

Materials and Methods

WRF model setup. We use a single-column configuration of the advanced research Weather Research and Forecasting model (WRF v3.4.1, [40]). Key aspects of the model setup include the initial profiles of temperature and humidity, and the physics parameterizations used. The initial temperature profile has a prescribed sub-metre temperature $T_0(0)$, and a lapse rate of either the mean pseudo-adiabatic value or $-8 K/km$, whichever is more stable; the tropopause at -60°C demarcates the base of an isothermal stratosphere. The initial humidity profile decreases from 80% at the surface (1000 hPa) to 20% at 600 hPa, and is constant at 20% up to the tropopause; the water vapor mixing ratio is set to 5 ppm above the tropopause. There is no large-scale subsidence, and the atmosphere is relaxed with a 1-day time scale to an eastward wind of 5 m/s that is uniform with height. The vertical coordinate of the WRF model is $h = (p_{in} - p)/\left(p_{in} - p_{out}\right)$, where $p_{in}$ is the grid-cell pressure, $p_{out}$ is the hydrostatic surface pressure, and $p_{surf}$ is the top-model hydrostatic pressure. We determine the position of model level 1 (out of $N = 51$ total) using a grid with finer spacing in the lower troposphere: $h_1 = 1 - \left(1 - (N-1)/51\right)^{1.6}$ We use the RRTMG scheme [41] for longwave radiative transfer, with CO2 set to 300 ppm unless otherwise noted; no shortwave scheme is required as simulations occur during polar night. We use the Yonsei University PBL scheme [42], and the surface is defined as a slab with a heat capacity equivalent to a water layer of 5 cm depth, independent of temperature. This choice of slab thickness is justified by considering the diffusive penetration depth into a homogeneous snow surface with volumetric heat capacity $C = 6 \times 10^7 J/m^3 K^{-1}$, thermal diffusivity $D = 5 \times 10^{-7} m^2 s^{-1}$, and a time scale $T = 10^5 s$ ($\approx 1 d$). The water-equivalent slab depth, $z_{eff}$, is given by the ratio of heat capacity of the medium to that of water, times the diffusive penetration depth $\sim z_{eff} = (C/C_w)z$, where $z = \sqrt{D\tau}$. These parameters give $z = 0.32 m$, and thus $z_{eff}$ is 5 cm. Sensitivity tests with much smaller and larger values of the surface heat capacity ($z_{eff} = 2 cm, 15 cm$) are shown in Figures SI-4 and SI-5; our general findings hold more strongly for a more insulating (lower heat capacity) surface, because inversions at low temperatures corresponding to the present climate are even stronger in such situations. Note also that a larger effective slab depth $z_{eff} \sim 15 cm$ would be more appropriate for a land surface without snow cover.

We use the following micropysics schemes: "Kessler" [43], "Lin" [44], "WSM-6" (WRF single-moment 6-class) [45], "Goddard" [46], "Thompson" [47], "Morrison" [48], "Stony Brook" [49], "W2M-6" (WRF 2-moment 6-class) [50].

The following 11 CMIP5 models were used in this study to calculate mean changes between historical (1980-1999) and RCP8.5 (2080-2099) periods: (1) BCC-CSM1-1; Beijing Climate Center, China Meteorological Administration; (2) CanESM2; Canadian Centre for Climate Modelling and Analysis; (3) CCSM4; National Center for Atmospheric Research; (4) CNRM-CM5; Centre National de Recherches M´et´eorelogiques; (5) GFDL-CM3; NOAA Geophysical Fluid Dynamics Laboratory; (6) HadGEM2-ES; Met Office Hadley Centre; (7) INMCM4; Institute for Numerical Mathematics; (8) IPSL-CM5A-LR; Institut Pierre-Simon Laplace; (9) MIROC5; Japan Agency for Marine-Earth Science and Technology; (10) MPI-ESM-MR; Max Planck Institute for Meteorology; (11) MRI-CGCM3; Meteorological Research Institute.

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Table 1. CMIP5 zonal-mean, model-mean changes in surface temperature and cloud radiative forcing, for DJF, showing the difference [2080-2099] minus [1980-1999]. Parentheses indicate standard deviation across models.

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Fig. 1. Single-column simulation results of polar air formation for cold and warm initial atmospheric columns. A reference simulation with initial atmosphere with 2 meter temperature of \( T_2(0) = 0^\circ C \) is shown in panel (A); and by purple lines in (C-E). A simulation starting with a warmer initial atmosphere, with \( T_2(0) = 20^\circ C \), is shown in panel (B); and by orange lines in (C-E). In (A) and (B), the black lines show the temperature profiles every two days as a function of pressure; the solid colors overlaying the temperature profile indicate where clouds are found, total cloud water content is indicated by line thickness, and cloud phase is indicated by color, with green being all liquid and blue all ice. Panel (C) shows the evolution of air temperature at 2-m height (solid) and 850 hPa (dashed), panel (D) shows the net longwave cooling of the surface for actual (solid) and hypothetical clear-sky (dashed) conditions, and panel (E) shows the evolution of vertically-integrated cloud liquid (\( q_{c,l} \), solid) and cloud ice (\( q_{c,i} \), dashed). The “Lin” microphysics scheme is used for these two simulations; other microphysics schemes show qualitatively similar features.
Fig. 2. Simulation results for (A) average surface cooling over 2-week period, $\Delta T_2$ ($^\circ$C), and (B) number of days taken for the 2-meter air temperature to drop below freezing, $\tau_0$, both as a function of $T_2(0)$. Black line ("multi-$\mu$physics mean") indicates an average across the solid-line microphysics parameterizations, which contain both liquid- and ice-phase processes. Dash-dotted lines show unrealistic microphysics assumptions used to diagnose the response mechanism: "no microphysics" indicates no phase change of water allowed, and thus no clouds at all; "no CRF" indicates that clouds are allowed to form but do not affect radiative transfer; "Kessler (no ice)" indicates a microphysics scheme that has only liquid condensate, regardless of temperature. A quadratic fit to the solid black line in (A) is shown in black dashes, with the fit shown at the bottom of (A).
Fig. 3. Maps of mean changes in December-January-February variables across 11 coupled climate models, as differences between historical simulations (1980-1999), and future simulations with the RCP8.5 radiative forcing scenario (2080-2099). Panel (A) shows changes in surface air temperature (°C), panel (B) shows changes in surface longwave radiative forcing by clouds (W/m²), and panel (C) shows percentage changes in the vertical integral of cloud liquid, which more than doubles over some high-latitude regions. Note the amplified response over high latitude continents and the Arctic.
Fig. SI-1. Direct influence of $p\text{CO}_2$ on 2-week average surface air temperature, $T_2$, across a range of temperatures. Note the logarithmic horizontal axis; minimum $p\text{CO}_2$ is 75 ppm, and maximum is 9600 ppm. Solid black lines indicate mean across microphysics schemes, numbers on each line indicate the initial 2-m air temperature $T_2(0)$, and dashed black lines indicate fit $T_2 = m\log_2(p\text{CO}_2) + b$. Slopes $m$ vary from 0.61 to 0.84 °C per doubling of $p\text{CO}_2$. 
In all-2-m air temperature, $T_2(0)$ (°C)

![Graph](image)

Fig. SI-2. As in Figure 2, but with $p_{CO_2}$ doubled for each 4°C increase in $T_2(0)$; $p_{CO_2}$ = (75, 150, 300, 600, 1200, 2400, 4800, 9600) ppm for $T_2(0) = (-8, -4, 0, 4, 8, 12, 16, 20)$ °C, respectively. Results are similar to those without CO$_2$ changes (Figure 2), and principal differences from the results in Figure 2 include more cooling at low $T_2(0)$, less cooling at high $T_2(0)$, and an increase in the magnitude of the slopes of the “no microphysics” and “no CRF” dash-dotted lines, which both use only clear-sky radiative transfer.
Fig. SI-3. As in Figure 2, but with a drier initial sounding, having 70% relative humidity at the surface, decreasing to 10% relative humidity at 600 hPa and above to the tropopause. Principal differences from results in Figure 2 include more cooling, especially at high $T_2(0)$, and consequently smaller sensitivity of $\Delta T_2$ to $T_2(0)$.
Fig. SI-4. As in Figure 2, but with a moister initial sounding, having 90% relative humidity at the surface, decreasing to 30% relative humidity at 600 hPa and above to the tropopause. Principal differences from results in Figure 2 include less cooling, especially at high $T_2(0)$, and consequently larger sensitivity of $\Delta T_2$ to $T_2(0)$. 

(A) Two-week average cooling $\Delta T_2$ (°C)

(B) Time $\tau_0$ (days) taken for $T_2$ to drop below freezing
Fig. SI-5. As in Figure 1, but with an imposed subsidence profile, with subsidence peaking at 2 mm/s at $z=4$ km, and decreasing to zero for $z=0$ km and $z=12$ km; the exact expression is given in the Materials and Methods section. Principal differences from results in Figure 1 include more cooling in the lowest 100-150 hPa of the troposphere, a sharper cloud-top inversion for the warmer initial condition, and less cooling aloft.
Fig. SI-6. As in Figure 2, but with an imposed subsidence profile, as described in Figure SI-5 and in the Materials and Methods section. Principal differences from results in Figure 2 include slightly more cooling, reduced sensitivity of $\Delta T_2$ to $T_2(0)$, and shorter times taken for the surface to reach freezing.
Fig. SI-7. As in Figure 2, but with a smaller thickness of the ground layer, $z_{\text{eff}} = 2$ cm instead of $z_{\text{eff}} = 5$ cm. Principal differences from results in Figure 2 include more cooling, especially at low $T_2(0)$, and consequently larger sensitivity of $\Delta T_2$ to $T_2(0)$. 
Fig. SI-8. As in Figure 2, but with a larger thickness of the ground layer, $z_{\text{eff}} = 15$ cm instead of $z_{\text{eff}} = 5$ cm. Principal differences from results in Figure 2 include less cooling, especially at low $T_2(0)$, and consequently smaller sensitivity of $\Delta T_2$ to $T_2(0)$. 
Fig. SI-9. Cloud water residence time plotted against initial 2-m air temperature, $T_2(0)$, for each of the microphysics schemes in Figure 2. Cloud water residence time is defined as the time-mean vertically-integrated cloud condensate (units: kg m$^{-2}$), divided by the time-mean surface precipitation rate (units: kg m$^{-2}$ s$^{-1}$). Note the logarithmic vertical axis; straight lines thus correspond to exponentially increasing or decreasing time scales with temperature; a factor of 10 change across the 30-degree temperature range shown here corresponds to $\sim 7.5\% \, \text{C}^{-1}$.

Most microphysics schemes show an increase in cloud water residence time with temperature, although the increase is not always monotonic (e.g., Stony Brook), and one scheme does not show an average increase in cloud water residence time with temperature (Thompson).