

Resolved Snowball Earth Clouds

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(Manuscript received 2 December 2013, in final form 23 February 2014)

ABSTRACT

Recent general circulation model (GCM) simulations have challenged the idea that a snowball Earth would be nearly entirely cloudless. This is important because clouds would provide a strong warming to a high-albedo snowball Earth. GCM results suggest that clouds could lower the threshold CO_2 needed to deglaciate a snowball by a factor of 10–100, enough to allow consistency with geochemical data. Here a cloud-resolving model is used to investigate cloud and convection behavior in a snowball Earth climate. The model produces convection that extends vertically to a similar temperature as modern tropical convection. This convection produces clouds that resemble stratocumulus clouds under an inversion on modern Earth, which slowly dissipate by sedimentation of cloud ice. There is enough cloud ice for the clouds to be optically thick in the longwave, and the resulting cloud radiative forcing is similar to that produced in GCMs run in snowball conditions. This result is robust to large changes in the cloud microphysics scheme because the cloud longwave forcing, which dominates the total forcing, is relatively insensitive to cloud amount and particle size. The cloud-resolving model results are therefore consistent with the idea that clouds would provide a large warming to a snowball Earth, helping to allow snowball deglaciation.

1. Introduction

Between 600 and 800 million years ago there were at least two periods during which paleomagnetic evidence indicates that glaciers flowed from continents into the ocean at or near the equator¹ (Kirschvink 1992; Hoffman et al. 1998). These events are associated with large carbon isotopic excursions and the glacial sequences are capped with thick layers of carbonates (Hoffman and Schrag 2002), both of which indicate that these glaciations were associated with large perturbations of the carbon cycle and were qualitatively different from the recent Pleistocene glaciations. Moreover, geochemical measurements indicate that the CO_2 during these events reached 0.01–0.1 volume mixing ratio (vmr; Kasemann et al. 2005; Bao et al. 2008, 2009), although there is some debate about the details (Sansjofre et al. 2011; Cao and Bao 2013).

Additionally, in at least some areas banded iron formations, which indicate anoxic ocean conditions, were deposited during these glaciations (Kirschvink 1992; Hoffman and Li 2009). The basic paleomagnetic evidence that glacial deposits (Evans and Raub 2011) and overlying carbonates (Hoffman and Li 2009) were at low paleolatitudes has been confirmed by recent meta-analyses incorporating all available data.

The (hard) snowball Earth hypothesis (Kirschvink 1992; Hoffman et al. 1998) explains these data through the onset of global glaciation, with the entire ocean covered in kilometer-thick ice, the slow accumulation of CO_2 in the atmosphere due to greatly reduced silicate weathering (Walker et al. 1981) leading to warming and deglaciation after millions of years, and the deposition of cap carbonates during the resulting extremely warm period when strong weathering would reduce the atmospheric CO_2 . Alternative hypotheses involving thin tropical ice or open tropical oceans have been proposed (Hyde et al. 2000; McKay 2000; Pollard and Kasting 2005; Abbot et al. 2011), mainly to explain the apparent survival of photosynthetic and animal life in the ocean through the snowball events (e.g., Love et al. 2009; Bosak et al. 2011), although oases in a hard snowball might also be able to permit this survival (e.g., Campbell

¹ There are potentially more such episodes about two billion years ago for which the data are poorer.

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et al. 2011; Tziperman et al. 2012). In this paper I will focus on the hard snowball Earth hypothesis in the spirit of fully investigating its implications to evaluate its consistency with available data, while acknowledging that future data may prove that it is not the correct model for all of the global glaciations.

Pierrehumbert (2004, 2005) performed the first general circulation model (GCM) simulations applied specifically to the hard snowball Earth, using the Fast Ocean Atmosphere Model (FOAM). He found that the snowball was nowhere near deglaciating at $\text{CO}_2 = 0.2$ vmr, which is inconsistent with the available geochemical estimates of the CO_2 (0.01–0.1 vmr; Kasemann et al. 2005; Bao et al. 2008, 2009). This led to the proposal that dust could significantly lower the snowball albedo (Abbot and Pierrehumbert 2010; Abbot and Halevy 2010; Le Hir et al. 2010; see also Schatten and Endal 1982) and a series of investigations in other GCMs (Le Hir et al. 2007; Abbot and Pierrehumbert 2010; Le Hir et al. 2010; Hu et al. 2011; Pierrehumbert et al. 2011). The GCMs yielded sometimes conflicting results. For example, Abbot and Pierrehumbert (2010) found that FOAM could only deglaciate at $\text{CO}_2 = 0.1$ vmr when the albedo was reduced in the tropics to account for a mixture of dust with snow and ice, whereas the Community Atmosphere Model (CAM) deglaciated in some configurations without reducing the albedo at $\text{CO}_2 = 0.1$ vmr.

This led to a comparison of six different GCMs run in consistent snowball Earth conditions (Abbot et al. 2012, 2013). Most of these GCMs calculate clouds using subgrid-scale schemes, but one, the superparameterized Community Atmosphere Model (SP-CAM), has an embedded two-dimensional cloud-resolving model² in each grid box. The main finding of Abbot et al. (2012, 2013) was that all of the other GCMs produced a much higher cloud radiative forcing (CRF) than FOAM, which led to a warming of the tropics by 7–11 K relative to FOAM, which is equivalent to an increase of the CO_2 by a factor of 10–100 in the snowball climate. Although the complex processes involved in deglaciation were not modeled, this is roughly enough warming to allow deglaciation to occur at a CO_2 level consistent with the geochemical data even without the albedo-reducing effects of dust (which could still be relevant).

Here I perform similar tests to those performed using GCMs by Abbot et al. (2012, 2013), but instead I use a cloud-resolving model [the System for Atmospheric Modeling (SAM); Khairoutdinov and Randall 2003]. A

cloud-resolving model explicitly simulates nonhydrostatic cloud-scale motions, although it still requires a cloud microphysical scheme, whereas GCMs rely on subgrid-scale parameterizations to model convection and cloud behavior. Since clouds and convection in a cloud-resolving model are less parameterized than in a GCM they may be more likely to yield an accurate representation of clouds in a climate different from the modern climate, although cloud-resolving models must be run on small domains and therefore cannot calculate the effects of large-scale motions, which can drive convective behavior. I find that SAM produces cloud condensate and a CRF similar to that of most GCMs (Abbot et al. 2012, 2013). This conclusion is robust to large changes in the cloud microphysical scheme. This work confirms the GCM results that clouds would likely provide significant warming to a snowball Earth.

2. SAM

a. Model description

I use the System for Atmospheric Modeling (Khairoutdinov and Randall 2003), version 6.10.4. SAM is a cloud-resolving model that solves the anelastic equations of motion and has liquid water and ice moist static energy, total nonprecipitating water, and total precipitating water as prognostic thermodynamical variables. Cloud condensate is diagnosed as occurring when an air parcel reaches saturation and supersaturation is not allowed. Longwave and shortwave radiative fluxes are computed using the radiation scheme from the National Center for Atmospheric Research (NCAR) Community Atmosphere Model, version 3 (Collins et al. 2004). This radiation scheme is accurate to within a few watts per meter squared at $\text{CO}_2 = 0.1$ vmr (Abbot et al. 2012). I use the single-moment cloud-microphysics scheme described in Khairoutdinov and Randall (2003). My main results are robust to large changes in the microphysical scheme parameters (section 4). I also tested the double-moment scheme described in Morrison et al. (2009), but found that the model was unstable when I used it in the configuration described here.

I ran the model on a doubly periodic square domain with a side length of 128 km that ranged vertically from the surface to a model top height of 17.5 km. I used a horizontal grid size of 1 km, a variable vertical grid size (368 m above 2 km and finer below this) for a total of 59 vertical levels, and a time step of 10 s. I confirmed that the model had converged with respect to these parameters (appendix B). I applied a small linear shear of $-0.01 \text{ m s}^{-1} \text{ mb}^{-1}$ (1 mb = 1 hPa) to avoid convective self-aggregation and domain-size dependence of results

² The cloud-resolving model embedded in SP-CAM is actually a version of SAM, the model used in this paper.

TABLE 1. This table compares the surface temperature (T_s) and top-of-atmosphere shortwave (CRF_{sw}), longwave (CRF_{lw}), and total (CRF) cloud radiative forcing for the reference SAM snowball simulation with $\text{CO}_2 = 10^{-4}$ vmr, a SAM snowball simulation with $\text{CO}_2 = 10^{-2}$ vmr, and the reference modern tropics SAM simulation.

Simulation	T_s (K)	CRF_{sw} (W m^{-2})	CRF_{lw} (W m^{-2})	CRF (W m^{-2})
Snowball, $\text{CO}_2 = 10^{-4}$ vmr	245	−2	16	14
Snowball, $\text{CO}_2 = 10^{-2}$ vmr	263	−5	23	18
Modern	299	−30	29	−1

([appendix A](#)). This is reasonable given the prevalence of shears in real atmospheres. I applied no other large-scale forcings to the model, so the simulations can be thought of as representing the tropics as a whole, rather than a specific tropical region such as the region of ascent of the Hadley circulation.

I ran the model with a passive mixed layer surface with a fixed constant surface albedo that was independent of wavelength. Following [Khairoutdinov and Yang \(2012\)](#), I allowed heterogeneity in surface fluxes across the domain, but homogenized the surface temperature, while allowing it to vary with time. I used a heat capacity of the mixed layer equivalent to 2 m of water for all simulations and ran the model until the surface temperature equilibrated, at which point the surface and top of the atmosphere were in energy balance to less than 1 W m^{-2} . It took 300–1000 days for the surface temperature to equilibrate, depending on the simulation, and I averaged variables presented below over 100 days after the model had equilibrated. To simulate the tropics, I applied a heat export to the mixed layer. This assumes that the effect of atmospheric heat transport can be reasonably approximated by ocean heat transport, and should tend to reduce convection. Although this assumption would be invalid for many purposes, I am able to obtain a reasonable proxy for the modern tropics in SAM despite making it. I used a solar constant of 566 W m^{-2} and a constant solar zenith angle of 45° , which yields an insolation of 400 W m^{-2} , which I calculated to be the modern tropical average using the code of [Huybers and Eisenman \(2006\)](#).

b. Reference modern climate state

I first confirm that SAM provides a reasonable simulation of the modern tropics. The goal is to determine whether SAM can produce a climate that is comparable to that of the modern tropics without extensive parameter tuning, rather than to reproduce features of the modern tropics in detail. To do this I use the model setup described in [section 2a](#), with the surface albedo set to 0.05, the heat export set to 85 W m^{-2} , and the CO_2 concentration set to 355 ppm. This results in a surface temperature of 299 K ([Table 1](#)), which is reasonable for the modern tropics.

The cloud behavior SAM produces in the modern configuration is broadly similar to that of the modern tropics. SAM produces deep convective plumes and tropical anvil clouds ([Fig. 1](#)). SAM's longwave cloud radiative forcing (CRF_{lw} ; [Table 1](#)) is very similar to the modern tropical value of approximately 30 W m^{-2} ([Harrison et al. 1990](#)), which is important because CRF_{lw} dominates snowball Earth cloud effect. SAM's shortwave cloud radiative forcing ($\text{CRF}_{\text{sw}} = -30 \text{ W m}^{-2}$, [Table 1](#)) is somewhat smaller than that of the modern tropics (-50 W m^{-2} ; [Harrison et al. 1990](#)), but this difference is not important when we consider the snowball climate since CRF_{sw} is so small in the snowball simulations [magnitude of $O(1 \text{ W m}^{-2})$]. SAM also reproduces the broad features of the mean cloud condensate profile as obtained by *CloudSat* observations of the tropical ocean [[Fig. 2](#); *CloudSat* data from [Su et al. \(2011\)](#), and averaged from August 2006 to July 2010]. Both the lower-level cloud liquid and upper-level cloud ice maxima occur at roughly the right pressures in SAM and are within a factor of 2–3 of the values from the satellite retrieval.

3. Resolved Snowball Earth clouds

I produced a reference snowball climate by making three changes to the parameters I used to generate the reference modern climate. First, I increased the surface albedo to 0.6,³ the value used in the comparison of GCMs in a snowball state ([Abbot et al. 2012, 2013](#)). There is uncertainty in the appropriate surface albedo for a snowball climate ([Warren et al. 2002](#); [Dadic et al.](#)

³ To reduce the number of variables changed and make comparison between the snowball and modern reference climates in SAM more straightforward, I did not reduce the insolation by 6% in the snowball simulations, as [Abbot et al. \(2012, 2013\)](#) did. This will have a negligible effect on the comparisons of the shortwave cloud radiative forcing in the snowball state between the SAM simulations described here and the GCM simulations of [Abbot et al. \(2012\)](#) because the shortwave cloud forcing is so small. Additionally, a 6% decrease in the insolation causes a forcing about half as large as the range in heat export from the tropics in the GCMs ($\sim 15 \text{ W m}^{-2}$; [Abbot et al. 2013](#)). Since I pick a single heat export for the SAM simulations, it is reasonable to neglect the smaller change in the insolation.

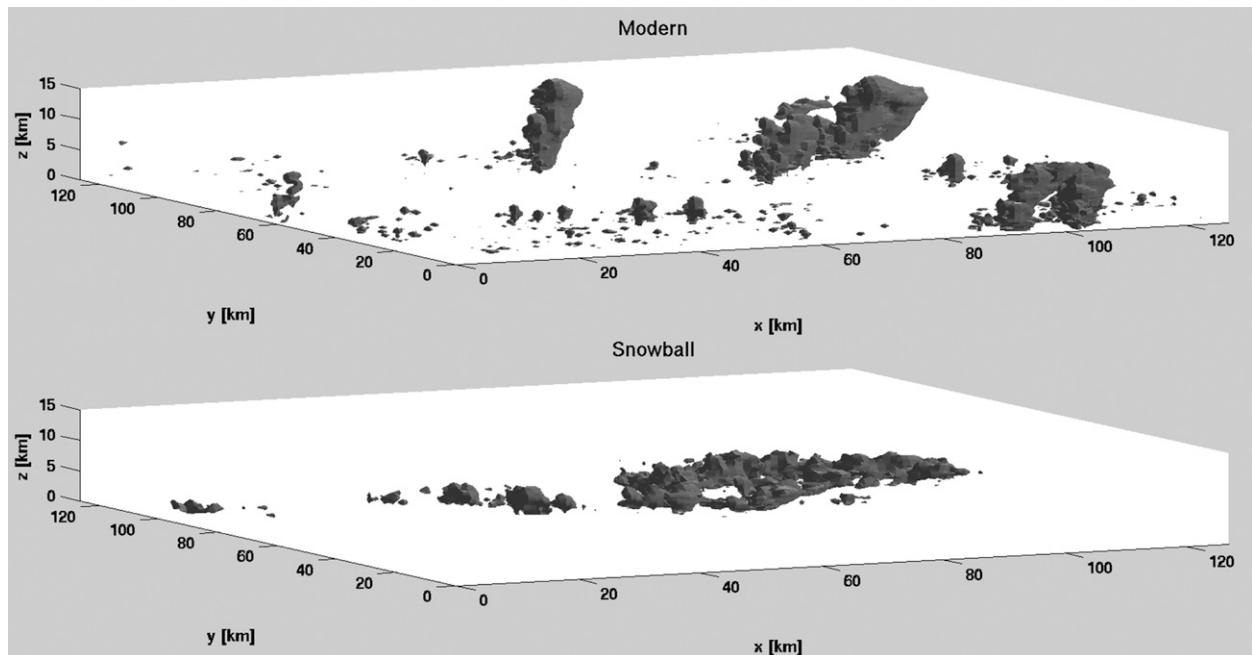


FIG. 1. Snapshot of cloud surfaces in (top) the reference modern climate state in SAM and (bottom) the reference snowball climate state (with $\text{CO}_2 = 10^{-4}$) in SAM. A cloud surface is defined as the contour on which the sum of the total precipitation and cloud condensate equals 0.4 g kg^{-1} for the modern simulation and 0.04 g kg^{-1} for the snowball simulation.

2013), particularly if there are large quantities of dust (Abbot and Pierrehumbert 2010; Le Hir et al. 2010), but this is a reasonable average value assuming some regions of snow, bare ice, and dusty ice. Second, I decreased the CO_2 to 10^{-4} vmr for consistency with the GCM simulations (Abbot et al. 2012, 2013). This low CO_2 value is meant to represent an early period in the snowball life cycle. Finally, I reduced the heat export to 15 W m^{-2} , which is roughly the heat export from the tropics in the snowball GCM simulations (Abbot et al. 2013). These changes resulted in a surface temperature of 245 K (Table 1), which is in the middle of the tropical temperature range of the snowball GCM simulations (Abbot et al. 2012, 2013).

SAM produces low-level clouds in the reference snowball state that roughly resemble the stratocumulus clouds that form on modern Earth when there is relatively low-level convection beneath a stable layer (Fig. 1). These snowball clouds are produced by convection, which we can start to understand by looking at the temperature lapse rate. In the reference snowball simulation (with $\text{CO}_2 = 10^{-4}$ vmr), a snowball simulation with CO_2 increased to 10^{-2} vmr, and the reference modern simulation, the atmospheric temperature closely follows a moist adiabat above the boundary layer (Fig. 3), which demonstrates the importance of moist convection for setting the temperature profile. In the reference snowball simulation the air is so cold and dry that the moist and dry

lapse rates are nearly indistinguishable from each other. This means that the convection is effectively dry, in the sense that latent heat released by condensation makes a negligible contribution to setting the lapse rate in the snowball case. Since small amounts of cloud and precipitation do form during snowball convection, however, it differs from the dry convection that occurs on modern

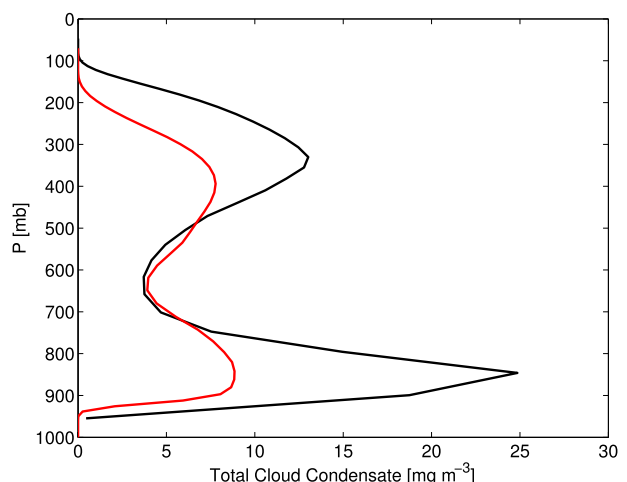


FIG. 2. Vertical profiles of total (liquid plus ice) cloud condensate in the reference modern SAM simulation (red) and for *CloudSat* observations (Su et al. 2011) averaged over the modern tropical ocean (black).

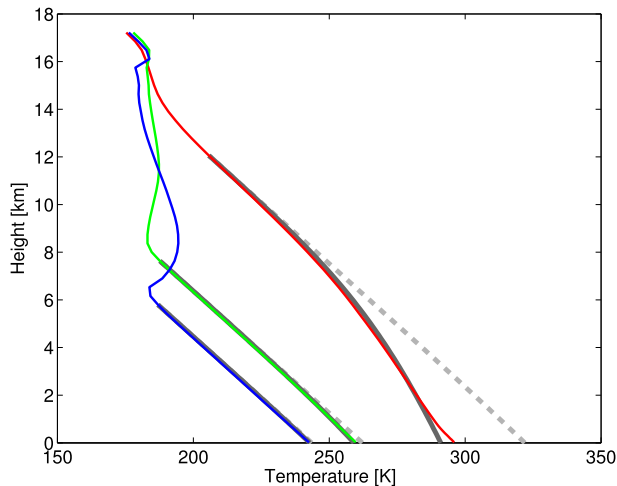


FIG. 3. Vertical profiles of air temperature for the reference SAM snowball simulation with $\text{CO}_2 = 10^{-4}$ vmr (blue), a SAM snowball simulation with $\text{CO}_2 = 10^{-2}$ vmr (green), and the reference SAM modern tropics simulation (red). Dry (light gray dashed) and moist (dark gray solid) adiabats, calculated following Emanuel (1994), are also plotted.

Earth on warm days in dry regions and is associated with clear skies. When the CO_2 is increased to 10^{-2} vmr in the snowball, the air near the surface warms enough that a small deviation between the moist and dry adiabats develops, and the temperature profile is better approximated by the moist adiabat.

The convection that SAM produces in the snowball lifts enough moisture to create an optically thick layer of cloud ice in the midtroposphere (Fig. 4). This cloud ice is particularly radiatively effective because the clear-sky air is so cold and dry that relatively little greenhouse warming, other than that provided by CO_2 , exists in the absence of clouds. Moreover, since the atmosphere is very dry the lapse rate is higher than in the modern tropics, so that the temperature difference between the surface and the cloud top is larger than one might expect from the (relatively shallow) depth of convection. The quantity of cloud ice in the reference SAM simulation is of similar magnitude to what most GCMs produce in snowball configuration (Abbot et al. 2012). For comparison, we can consider the annual and tropical mean cloud ice profiles produced by the CAM GCM (Collins et al. 2004) and a superparameterized version of CAM with an embedded 2D cloud-resolving model in each grid box (SP-CAM; Khairoutdinov and Randall 2001; Khairoutdinov et al. 2008). CAM and SP-CAM produced the most cloud ice in the comparison of GCMs (Abbot et al. 2012), and they produce somewhat more than SAM does in the reference snowball state (Fig. 5).

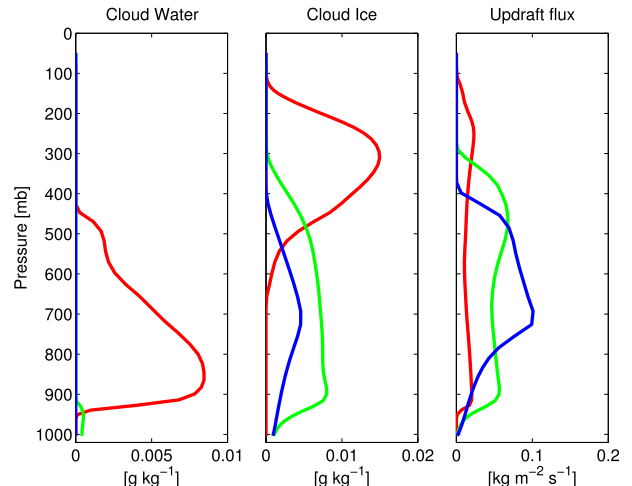


FIG. 4. Vertical profiles of cloud water, cloud ice, and updraft cloud mass flux for the reference SAM snowball simulation with $\text{CO}_2 = 10^{-4}$ vmr (blue), a SAM snowball simulation with $\text{CO}_2 = 10^{-2}$ vmr (green), and the reference SAM modern tropics simulation (red).

The top-of-atmosphere CRF in the reference SAM snowball state is 14 W m^{-2} (Table 1), which is in the middle of the range of $11\text{--}21 \text{ W m}^{-2}$ for the GCMs with prognostic cloud condensate (Abbot et al. 2012, 2013). If we make a specific comparison to the CAM and SP-CAM GCMs we find that they both produce CRF_{lw} and CRF_{sw} that are slightly larger in magnitude, but a similar total CRF (Table 2). I next increased the CO_2 to 10^{-3} , 10^{-2} , and 10^{-1} vmr to test the effect of warming

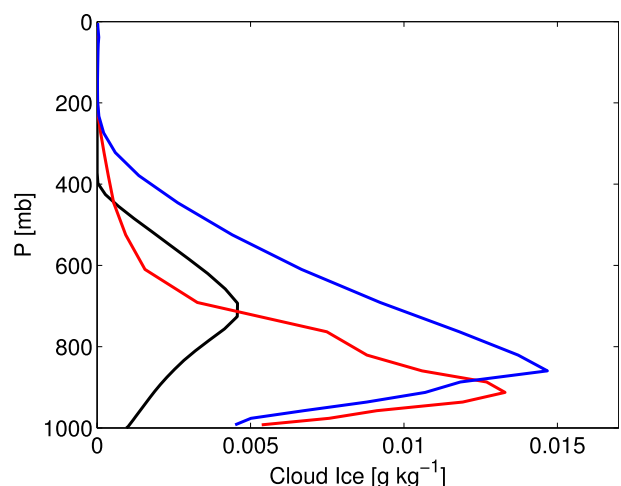


FIG. 5. Vertical profiles of cloud ice in the reference SAM snowball simulation (black), and in the GCMs CAM (red) and SP-CAM (blue) averaged over the tropics ($20^\circ\text{S}\text{--}20^\circ\text{N}$) and in the annual mean. $\text{CO}_2 = 10^{-4}$ vmr for all simulations. The GCM simulations are described in Abbot et al. (2012).

TABLE 2. This table compares the surface temperature (T_s) and top-of-atmosphere shortwave (CRF_{sw}), longwave (CRF_{lw}), and total (CRF) cloud radiative forcing for the reference SAM snowball simulation and for the annual mean in the tropics (20°S – 20°N) in the GCMs CAM and SP-CAM. $\text{CO}_2 = 10^{-4}$ vmr for all simulations. The GCM simulations are described in [Abbot et al. \(2012\)](#).

Simulation	T_s (K)	CRF_{sw} (W m^{-2})	CRF_{lw} (W m^{-2})	CRF (W m^{-2})
SAM	245	−2	16	14
CAM	245	−4	18	15
SP-CAM	247	−4	23	19

caused by increased CO_2 as a snowball ages. As CO_2 increases, both CRF_{lw} and CRF_{sw} increase in magnitude, but CRF_{lw} increases slightly more, leading to a small net increase in CRF with increasing CO_2 until the surface temperature exceeds the melting point ([Fig. 6](#)). This is consistent with the behavior of the GCMs ([Abbot et al. 2012](#)), for which the net change in CRF was small and not of consistent sign because of competing increases in the magnitude of CRF_{lw} and CRF_{sw} . Much more warming occurs as the CO_2 is increased in SAM than in the tropics of the GCMs, such that the temperature in SAM at $\text{CO}_2 = 10^{-2}$ vmr is similar to that in the GCMs at $\text{CO}_2 = 10^{-1}$ vmr. This is because I have held the heat export constant in SAM, whereas heat export from the tropics increases as the tropics warm in the GCMs ([Abbot et al. 2013](#)).

Although snowball convection is shallower than in the modern tropics, the updraft cloud mass flux is actually significantly stronger ([Fig. 4](#)). This might be surprising since less solar radiation is absorbed at the surface in the high-albedo snowball, but it can be explained by the fact that the air is much drier in the snowball. Because of this the vertical transport of moisture is much smaller in the snowball than in the modern tropics ([Fig. 7](#)), so that convection must be stronger in order to move enough heat vertically to balance radiative cooling aloft. The convective mass flux is high enough in the snowball that significant (in relative terms) amounts of cloud ice are advected vertically ([Fig. 7](#)). Cloud microphysics sensitivity experiments show that ice sedimentation is the dominant removal process for cloud ice in the reference snowball simulation ([section 4](#)). Therefore, the qualitative picture of convection in the reference snowball simulation is that it is dry and vigorous, although fairly shallow, and lifts water vapor to form cloud ice by condensation. Convection then lifts this cloud ice farther and eventually it is removed from the atmosphere by slow settling (rather than forming or being absorbed onto precipitation).

The fixed anvil temperature (FAT; [Hartmann and Larson 2002](#)) hypothesis is a useful guide for thinking

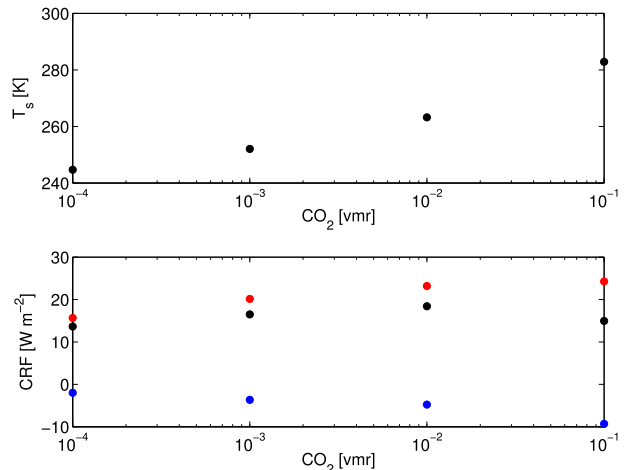


FIG. 6. Surface temperature (T_s) and top-of-atmosphere cloud radiative forcing (CRF) as a function of CO_2 volume mixing ratio in the snowball state in SAM. The longwave (red), shortwave (blue), and net (black) CRF are all plotted.

about similarities of convection in SAM between the vastly different modern and snowball climate states. The FAT hypothesis states that convective outflow and the cloud top should occur at roughly the same air temperature regardless of the surface temperature. This can be understood by considering the heat balance in the upper troposphere, which is mainly between heat released by condensation of water vapor lifted by convection and radiative cooling in nearby clear-sky regions. Convection only reaches as high as clear-sky radiative cooling

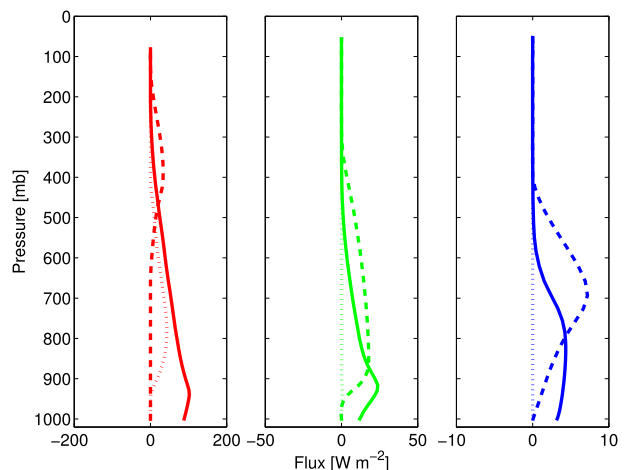


FIG. 7. The vertical flux of water vapor (solid), cloud liquid (dotted), and cloud ice (dashed) for the reference modern SAM simulation with $\text{CO}_2 = 10^{-4}$ vmr (red), a snowball simulation with $\text{CO}_2 = 10^{-2}$ vmr (green), and the reference snowball simulation with $\text{CO}_2 = 10^{-4}$ vmr (blue). Note the different horizontal scales in the three plots.

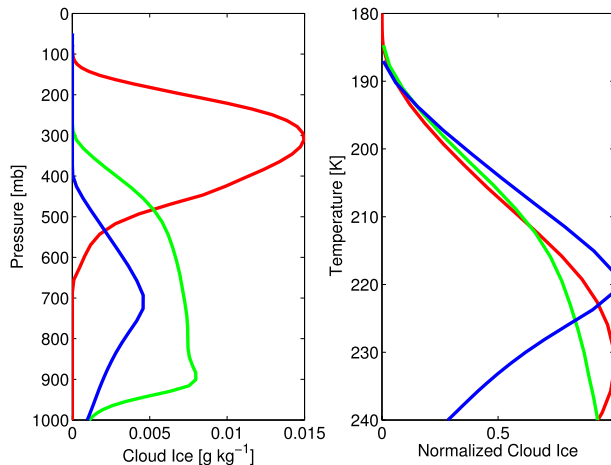


FIG. 8. Vertical profiles of cloud ice as a function of (left) pressure and (right) air temperature for the reference SAM snowball simulation with $\text{CO}_2 = 10^{-4}$ vmr (blue), a SAM snowball simulation with $\text{CO}_2 = 10^{-2}$ vmr (green), and the reference SAM modern tropics simulation (red).

can operate efficiently and balance the convective heat flux. The clear-sky radiative cooling is dominated by water vapor emission, and the water vapor concentration is a strong function of temperature by the Clausius–Clapeyron relationship. This means that when the temperature drops below a certain value, the water vapor concentration gets low enough that clear-sky radiative cooling is inefficient. The FAT hypothesis states that regardless of the surface climate, convection should extend up to this temperature.

When I plot vertical profiles of cloud ice as a function of air temperature, rather than pressure, I find that the cloud-top location is roughly determined by the local air temperature (Fig. 8). The fact that the cloud top exists at a roughly constant temperature is likely due to FAT, as opposed to convection simply reaching a stratosphere roughly defined by the skin temperature, since the skin temperature is approximately 215 K in the reference modern simulation and about 190–195 K in the snowball simulations (and therefore is not constant). Figure 8 indicates that it is reasonable to think of the snowball atmosphere as similar to the upper levels of the modern atmosphere, or as having a much lower tropopause. It also suggests that we should expect some CRF for a snowball climate with a surface temperature above the cloud-top temperature of approximately 215 K. I confirm this idea by increasing the heat export in the snowball reference state. When this decreases the surface temperature below 215 K, convection ceases to be an important factor and the cloud content and cloud radiative forcing drop to near zero (Fig. 9).

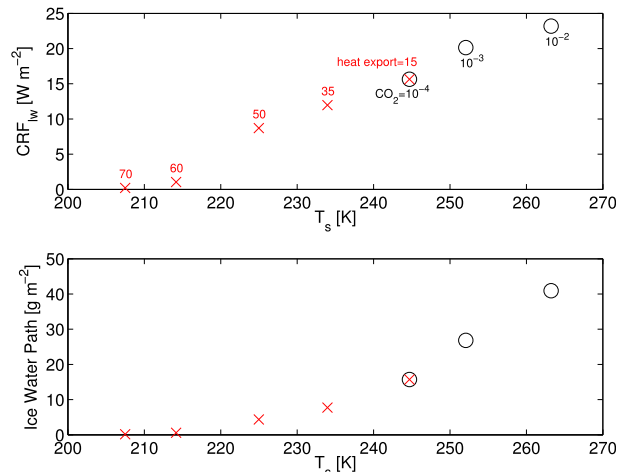


FIG. 9. (top) Top-of-atmosphere longwave cloud radiative forcing (CRF_{lw}) and (bottom) cloud ice water path as a function of surface temperature (T_s) in snowball Earth states in SAM. The reference snowball simulation (heat export of 15 W m^{-2} and $\text{CO}_2 = 10^{-4}$ vmr) is represented by both a black circle and a red cross. For the simulations represented by black circles the heat export is held at 15 W m^{-2} and the CO_2 is increased as labeled below data points in the top panel in black. For the simulations represented by red crosses the CO_2 is held at 10^{-4} vmr and the heat export is increased as labeled above the data points in the top panel in red.

4. Sensitivity to microphysical parameters

Although SAM resolves cloud-scale motion, it cannot resolve the cloud microphysics essential for determining the quantity and size of cloud condensate, which strongly affect cloud radiative properties. SAM relies on parameterizations of cloud microphysical quantities (Khairoutdinov and Randall 2003), which are subject to uncertainty. Because of this I performed a series of sensitivity tests in which I varied parameters in the cloud microphysical scheme.

I first varied a number of parameters related to the formation of precipitation from cloud ice. This is important because precipitation is rapidly removed from the atmosphere. In the SAM microphysical scheme the increase in snow due to accretion of condensate follows a continuous growth equation with a proportionality constant E_{si} (ice to snow) and E_{gi} (ice to graupel). I increased and decreased these parameters by a factor of 2. Additionally, aggregation of cloud ice onto snow is assumed to be proportional to $\beta e^{0.025(T-273.16)}(q_i - q_{\text{io}})$ when $q_i > q_{\text{io}}$, where T is the temperature, q_i is the cloud ice mixing ratio, q_{io} is the threshold cloud ice mixing ratio for aggregation, and β is the ice aggregation rate. I increased and decreased q_{io} by a factor of 2 and β by an order of magnitude. None of these changes had a significant effect on the vertical profile of cloud condensate (not shown), the cloud radiative forcing, or the surface

TABLE 3. This table compares the surface temperature (T_s) and top-of-atmosphere shortwave (CRF_{sw}), longwave (CRF_{lw}), and total (CRF) cloud radiative forcing in simulations designed to test the sensitivity of the model to cloud microphysical assumptions. Microphysical parameters are perturbed from the reference snowball state with $\text{CO}_2 = 10^{-4}$ vmr and from a snowball state with $\text{CO}_2 = 10^{-2}$ vmr. Here E_{si} is the collection efficiency of snow for cloud ice parameter; its standard value is 0.1 and it is decreased (denoted by down arrow) to 0.05 and increased (denoted by up arrow) to 0.2; E_{gi} is the collection efficiency of graupel for cloud ice parameter; its standard value is 0.1 and it is decreased to 0.05 and increased to 0.2; q_{ci} is the threshold cloud ice mixing ratio for aggregation of ice to precipitation; its standard value is $10^{-4} \text{ kg kg}^{-1}$ and it is decreased $10^{-5} \text{ kg kg}^{-1}$ and increased to $10^{-3} \text{ kg kg}^{-1}$; β is the rate constant for aggregation of ice to precipitation; its standard value is 10^{-3} s^{-1} and is decreased to 10^{-4} s^{-1} and increased to 10^{-2} s^{-1} ; T_{10} is the lowest temperature for which cloud water can exist (only cloud ice exists below T_{10}); its standard value is -30°C and it is decreased to -40°C ; and v_i is the ice sedimentation velocity, which is a specified function of cloud ice density in SAM; it is both decreased and increased by a factor of 2 for all cloud ice densities. Finally, r_{ei} is the ice cloud effective radius and r_{el} is the liquid cloud effective radius; r_{ei} is specified to be a roughly exponential function of temperature and r_{el} is set to a constant value of $14 \mu\text{m}$. They are both decreased and increased by a factor of 2.

Test simulation	T_s (K)	CRF_{sw} (W m^{-2})	CRF_{lw} (W m^{-2})	CRF (W m^{-2})
Reference, $\text{CO}_2 = 10^{-4}$ vmr	245	-2	16	14
$E_{\text{si}} \downarrow$	245	-2	16	14
$E_{\text{si}} \uparrow$	245	-2	16	14
$E_{\text{gi}} \downarrow$	245	-2	16	14
$E_{\text{gi}} \uparrow$	245	-2	16	14
$q_{\text{ci}} \downarrow$	245	-1	16	15
$q_{\text{ci}} \uparrow$	245	-2	16	14
$\beta \downarrow$	245	-2	16	14
$\beta \uparrow$	245	-2	16	14
$T_{10} \downarrow$	245	-2	18	15
$v_i \downarrow$	257	-5	37	31
$v_i \uparrow$	242	0	10	9
$r_{\text{ei}}, r_{\text{el}} \downarrow$	244	-5	19	14
$r_{\text{ei}}, r_{\text{el}} \uparrow$	245	1	13	14
Reference, $\text{CO}_2 = 10^{-2}$ vmr	263	-5	23	18
$r_{\text{ei}}, r_{\text{el}} \downarrow$	264	-9	29	20
$r_{\text{ei}}, r_{\text{el}} \uparrow$	263	0	19	18

temperature (Table 3), which indicates that conversion of cloud ice to precipitation is not a significant cloud ice sink in the SAM snowball simulations.

SAM partitions cloud condensate between cloud ice and cloud liquid as a function of temperature. The lowest temperature for which cloud water can exist is T_{10} , with only cloud ice existing below T_{10} . The standard value of T_{10} is -30°C , which means that all cloud condensate is in the form of cloud ice in the reference snowball state. When I decrease T_{10} to -40°C , there is a very small increase in the net CRF and a less than 1-K increase in surface temperature (Table 3) due to some of the low-level cloud condensate existing as liquid rather than ice. Reasonable changes in this parameter therefore do not alter the conclusions of this paper.

Next I considered the effect of changing the direct sedimentation of cloud ice out of the atmosphere. The ice sedimentation scheme in SAM is based on a fit to observations of ice sedimentation velocity as a function of ice water content described in Heymsfield (2003). The maximum scatter of data points is about a factor of 2 in either direction in these data. I vary the ice sedimentation velocity by this much, which represents a large overestimate of the potential variation in the

actual line of best fit to the data. When I decrease the ice sedimentation velocity by a factor of 2, the net CRF increases by 17 W m^{-2} and the surface temperature increases by 12 K (Table 3). This indicates that ice sedimentation is the main removal process of cloud ice in the SAM snowball simulations. When I increase the ice sedimentation velocity by a factor of 2, the surface temperature only decreases by a few kelvins and the net CRF is still $O(10 \text{ W m}^{-2})$ (Table 3). This shows that even if the ice sedimentation scheme used here dramatically underestimates the ice sedimentation velocity, the main conclusions of this paper would still hold.

Finally I varied the ice cloud effective radius r_{ei} and the liquid cloud effective radius r_{el} used by the radiation scheme. Changing the cloud particle effective radius in the radiation scheme does not change the sedimentation velocity, which I investigated separately above. The effective radius is important for the radiative calculation because it determines how much cloud condensate is required for a cloud to be optically thick. I used the CAM radiation scheme (Collins et al. 2004), which assumes r_{el} has a constant value of $14 \mu\text{m}$ and r_{ei} is a roughly exponential function of temperature rising

from $6\text{ }\mu\text{m}$ at 180 K to $250\text{ }\mu\text{m}$ at 274 K. I doubled or halved r_{el} and the value of r_{ei} after computation (as a function of temperature) by the code. I did this for both the reference snowball simulation ($\text{CO}_2 = 10^{-4}\text{ vmr}$) and for a warmer snowball simulation ($\text{CO}_2 = 10^{-2}\text{ vmr}$). Increasing or decreasing r_{el} and r_{ei} by a factor of 2 changed CRF_{lw} by approximately 20% (Table 3), which does not change the conclusion of this paper that clouds in a hard snowball are capable of supplying a radiative forcing of $O(10\text{ W m}^{-2})$. This is consistent with the fact that the cloud longwave effect is generally insensitive to cloud particle size because relatively thin ice clouds act effectively as a blackbody (Fig. 5.9 of Pierrehumbert 2010). One expects changes in cloud particle effective radius to have a much larger effect on CRF_{sw} than CRF_{lw} (Pierrehumbert 2010), and consistent with this I find that CRF_{sw} changes by 100%–150% in these sensitivity experiments (Table 3). Since the magnitude of CRF_{sw} is small because the clouds are relatively thin and the surface albedo is high, however, this only amounts to a change in CRF_{sw} of a few watts per meter squared. In these sensitivity experiments the changes in CRF_{lw} and CRF_{sw} when the cloud particle effective radius is changed happen to cancel so that the net CRF and the surface temperatures are nearly unaffected (Table 3). This is of less significance than the fact that the magnitude of changes in both CRF_{lw} and CRF_{sw} is only a few watts per meter squared, so that the results of this paper are not sensitive to the assumed cloud particle effective radius.

5. Discussion

This work is important not only because of its geological implications, but also because it can be considered a success of climate theory. It is noteworthy that climate theory as embodied by sophisticated numerical models including multiple GCMs, a superparameterized GCM, and a cloud-resolving model can paint a consistent picture (at least to within a factor of 2) of cloud behavior during a snowball Earth event, a climate substantially different from the modern climate. This emphasizes the importance of subjecting climate theory to tests in regimes outside the range normally considered.

Many different types of microphysical schemes are used in cloud-resolving models. I used a relatively simple single-moment scheme. One would obtain somewhat different results using different microphysical schemes; however, it is unlikely that a different microphysical scheme would reduce the CRF by an order of magnitude (the amount needed to change the conclusions of this paper) given that the results were robust to large changes in the parameters of the microphysical scheme (section 4).

The basic fact is that resolved convection extends to a temperature similar to that of modern tropical convection in the cloud-resolving model. As long as the surface temperature is significantly higher than the cloud-top temperature of approximately 215 K, convection will lead to condensation and produce clouds. It takes relatively little cloud ice to form clouds that are optically thick in the infrared; given the high albedo of the snowball, these clouds will cause warming.

Spatially, the strongest convection and most optically thick clouds in the GCMs occurred where there was convergence and ascent caused by the strong Hadley cell in a snowball Earth (Abbot et al. 2012, 2013). I applied no vertical ascent to SAM here, but nevertheless the model still produced a CRF similar to that in the tropics of the GCMs. Similarly, the CRF SAM produces increases with shear if the magnitude of the shear is increased beyond $0.01\text{ m s}^{-1}\text{ mb}^{-1}$ (appendix A), which may make the CRF estimate made here conservative, since a strong Hadley cell would lead to large shears. In a real snowball Earth these effects of large-scale dynamics, which might enhance the CRF beyond the estimates made here, would be strongly influenced by the vertical transport of momentum, which is a parameterized process in GCMs (Voigt et al. 2012; Voigt 2013). In any case, the current work is sufficient to demonstrate the order of magnitude of CRF during a snowball.

A final uncertainty that is worthy of discussion is the amount of cloud condensation nuclei that would be available in a snowball. Likely sources of cloud condensation nuclei would be aeolian and volcanic dust, which could reach significant levels because of minimal sinks (Abbot and Pierrehumbert 2010; Abbot and Halevy 2010; Le Hir et al. 2010). Nevertheless, there might be less cloud condensation nuclei than in the modern climate, which would lead to larger, less radiatively active cloud particles. Even when I doubled the cloud particle radius, however, the snowball longwave cloud radiative forcing SAM produced only decreased by a few watts per meter squared, which is not enough to alter the conclusions of this work.

6. Conclusions

Early thinking and calculations suggested that a snowball Earth would be so cold that there would be few optically thick clouds, and consequently snowball deglaciation would not be possible at a CO_2 consistent with geochemical data. I have shown that clouds provide a radiative forcing of $10\text{--}20\text{ W m}^{-2}$ in a hard snowball Earth climate state in the cloud-resolving model System for Atmospheric Modeling (SAM). The model produces nearly dry convection that extends vertically to a temperature of

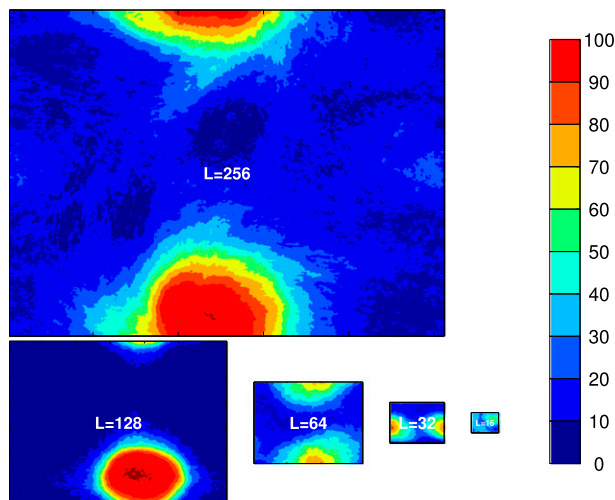


FIG. A1. Snapshots of cloud fraction for SAM snowball simulations with a horizontal resolution of 1 km and no shear applied. The surface temperature is between 243 and 254 K in these simulations, yet convective self-aggregation is evident. The domains are doubly periodic and square with side length $L = 256, 128, 64, 32$, and 16 km.

about 215 K and leads to clouds that resemble stratocumulus clouds on modern Earth. These clouds slowly dissipate by cloud ice settling out of the atmosphere, and are then replenished through further convection. SAM robustly produces a cloud radiative forcing of $O(10 \text{ W m}^{-2})$ in the snowball even when I make a variety of large changes to parameters in the cloud microphysics scheme, which is consistent with the tropical cloud radiative forcing in most general circulation models (GCMs) run in snowball configuration. A cloud radiative forcing of this magnitude is sufficient to allow consistency of the snowball CO_2 deglaciation threshold with geochemical data, and preserve the hard snowball Earth as a viable hypothesis for the global glaciations that occurred between 800 and 600 million years ago.

TABLE B1. This table compares the surface temperature (T_s) and top-of-atmosphere shortwave (CRF_{sw}), longwave (CRF_{lw}), and total (CRF) cloud radiative forcing for the convergence tests I have done on the SAM snowball simulation with a shear of $-0.01 \text{ m s}^{-1} \text{ mb}^{-1}$. The initial test horizontal resolution in both directions (dx) is 1 km, the time step (dt) is 10 s, the vertical resolution (dz) is 368 m above approximately 2 km, and the length of the square domain (L) is 64 km. The convergence tests are self-explanatory, except that in “reduced dz ” I increase the number of vertical levels from 59 to 89, reducing dz from 368 m to 215 m above approximately 2 km. These tests demonstrate that the model has converged with respect to all variables except L in the initial test simulation. Unless otherwise noted, a domain length of 128 km ($2L$) is used in other simulations throughout this paper.

Test simulation	T_s (K)	CRF_{sw} (W m^{-2})	CRF_{lw} (W m^{-2})	CRF (W m^{-2})
Initial test	249	−3	23	19
$dx/2$	248	−3	21	18
$dt/2$	249	−3	23	20
Reduced dz	248	−3	21	17
$2L$	245	−2	16	14
$4L$	245	−2	16	14

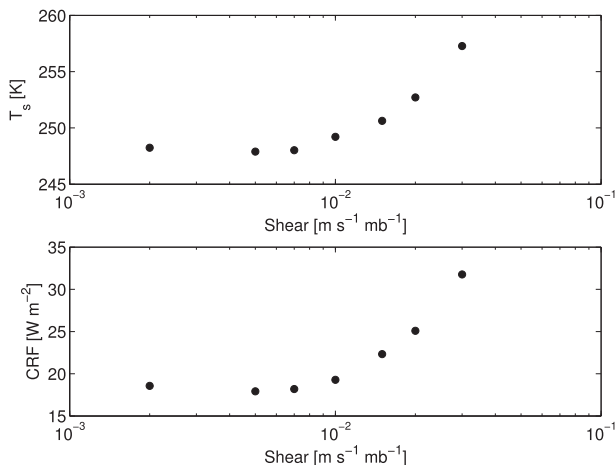


FIG. A2. Surface temperature (T_s) and top-of-atmosphere cloud radiative forcing (CRF) as a function of the magnitude of the applied linear shear in SAM run in the snowball state.

Acknowledgments. I thank Marat Khairoutdinov for developing SAM and making it available, Hui Su for providing the *CloudSat* data, and Mark Branson for sharing the CAM and SP-CAM output. I thank Aiko Voigt and Nathan Arnold for careful readings of an early version of this manuscript and Christopher Bretherton, Rodrigo Caballero, Zhiming Kuang, Caroline Muller, and David Romps for helpful comments and advice as I worked on this project. I thank two anonymous reviewers for constructive reviews. I acknowledge support from an Alfred P. Sloan Research Fellowship.

APPENDIX A

Convective Self-Aggregation and Shear

When I run SAM in snowball configuration without vertical shear, convective self-aggregation ([Bretherton](#)

et al. 2005) spontaneously occurs (Fig. A1). Interestingly, and in contrast to modern climate simulations (Muller and Held 2012), convective self-aggregation occurs even when the domain length is reduced to as little as 16 km. Some previous work has implicated high surface temperature as an important determinant of convective self-aggregation (Khairoutdinov and Emanuel 2010), but this cannot be the main determinant as the surface temperatures in the simulations presented in Fig. A1 are between 243 and 254 K. It is likely that snowball convective self-aggregation is driven either by a lack of cold pools in the snowball, which are more common with vigorous deep convection and destroy convective self-aggregation (Jeevanjee and Roms 2013), or stronger longwave cooling in cold, dry regions, which would promote convective self-aggregation (Muller and Held 2012).

In the simulations that show convective self-aggregation, the surface temperature changes by approximately 10 K and the CRF changes by approximately 15 W m^{-2} as the domain size is changed. Since we are interested in estimating cloud behavior during a snowball and have no way to determine the appropriate domain size a priori, this is unacceptable. Fortunately, introducing a small linear (in pressure) shear to the model destroys the convective self-aggregation (and domain size dependence of results). It is reasonable to apply a small shear because shears are common throughout real atmospheres. I found that the magnitude of the applied shear has little effect on the equilibrated climate if it is $0.01 \text{ m s}^{-1} \text{ mb}^{-1}$ or less (Fig. A2), and therefore used a shear of $-0.01 \text{ m s}^{-1} \text{ mb}^{-1}$ throughout the rest of the paper.

APPENDIX B

Resolution and Domain Size

I performed a number of tests to make sure that the snowball state that I investigate in this paper has converged with respect to changes in domain size and horizontal, vertical, and time resolution (Table B1). Halving the horizontal grid size and time step and decreasing the vertical grid size by 50% change the equilibrated surface temperature by 1 K or less and the equilibrated CRF by less than 2 W m^{-2} . This is an acceptable tolerance for investigation of the snowball climate. When I increased the side length of the square domain from 64 to 128 km the surface temperature decreased by 4.6 K and the CRF decreased by approximately 6 W m^{-2} , but both barely changed when I further increased the side length to 256 km. I therefore used a side length of 128 km throughout this paper.

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