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Mudball: Surface dust and Snowball Earth deglaciation

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[1] Recent modeling results have raised doubts about the ability to deglaciate from a global glaciation at atmospheric carbon dioxide levels that are realistic for a Neoproterozoic Snowball Earth. Here we argue that over the lifetime of a Snowball event, ice dynamics should lead to the development of a layer of continental and volcanic dust at the ice surface in the tropics that would significantly lower the tropical surface albedo and encourage deglaciation. This idea leads to the prediction that clay drapes found on top of Neoproterozoic glaciations should be thicker in tropical than extratropical regions. We test this idea by running the FOAM general circulation model (GCM) with an added tropical dust layer of different sizes and albedos and find that the tropical dust layer causes Snowball deglaciation at pCO_2 = 0.01–0.1 bar in a reasonable regime of these parameters. We find similar, though more nuanced, results from a limited number of test cases using National Center for Atmospheric Research's CAM GCM.

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1. Introduction

[2] During the Neoproterozoic era (∼1000 to ∼542 Ma), there were at least two global-scale glaciations, at ∼710 Ma and ∼635 Ma, during which glaciers reached sea level in the deep tropics [*Evans*, 2000; *Trindade and Macouin*, 2007]. A fierce debate rages [e.g., *Allen and Etienne*, 2008] as to whether Earth's oceans were covered in thick sheets of ice during these periods, the so-called "Snowball Earth hypothesis" [*Kirschvink*, 1992; *Hoffman et al.*, 1998], or tropical oceans were ice free, the so-called "Slushball Earth hypothesis" [*Hyde et al.*, 2000]. The Snowball hypothesis encompasses the onset of global sea ice coverage due to a runaway ice albedo feedback, millions of years in the Snowball state during which the CO_2 concentration would build up to immense levels because of greatly reduced weathering, and the escape from the Snowball as a result of warming due to the high CO_2 concentrations [*Kirschvink*, 1992]. Significant, though debated, geological evidence has been collected in support of the Snowball hypothesis. First, iron and iron-manganese deposits are found in glacial sediments [*Hoffman and Li*, 2009], which indicates anoxic ocean conditions and might imply that the oceans were completely ice covered. Second, many of the glacial formations are overlain with carbonate formations [*Hoffman and Li*, 2009], which could have resulted from the rapid deposition of carbonates washed into the ocean by warm rain in the extremely hot and high-CO_2 postglacial period and are much harder to explain if the ocean was not ice covered.

[3] Snowball deglaciation is a crucial aspect of the Snowball debate because as we are not currently in a Snowball, the Snowball story would be rendered implausible if the mechanism for Snowball escape can be shown to be physically inconsistent. Since climate models are a representation of our understanding of the climate system, we should be able to use them to determine the CO_2 concentration that would be required to deglaciate a Snowball Earth event. If this CO_2 concentration is at odds with the concentration we believe to have prevailed at the end of Snowball events, then there must be a problem with either the Snowball hypothesis or the climate models.

[4] Recent measurements of oxygen isotopic anomalies in sulfate minerals suggest that CO_2 levels reached 0.01–0.08 bar during and just after the ∼635 Ma glacial event [*Bao et al.*, 2008, 2009], although these measurements do not necessarily represent the maximum CO_2 level reached. Geochemical modeling indicates that it would take 2–5 Ma for 0.08 bar of CO_2 to accumulate in the atmosphere, depending on the area of sea ice openings and magnitude of deep-sea weathering [*Le Hir et al.*, 2008]. This is consistent with geochronological evidence that the ∼635 Ma glaciation could have lasted as long as 15 Ma and that the ∼710 Ma glaciation may have lasted even longer, but cannot rule out shorter lifetimes [*Hoffman and Li*, 2009]. Therefore, current evidence appears to limit the order of magnitude of the pCO_2 required for deglaciation to no higher than ∼0.1 bar and possibly as low as ∼0.01 bar.

[5] Most climate models require pCO_2 ∼ 0.3 bar to deglaciate a Snowball with thick tropical ice. For example, the energy balance model of *Caldeira and Kasting* [1992] deglaciates at pCO_2 = 0.29 bar [*Pierrehumbert*, 2005], the LMDz global climate model deglaciates at pCO_2 = 0.3 bar [*Le Hir et al.*, 2007], and the FOAM v1.5 global climate model [*Jacob*, 1997] is far from deglaciation at pCO_2 = 0.2 bar *Pierrehumbert* [2004]. We should note, however, that it may

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Table 1. Summary of Models and Configurations Used in the Simulation Presented in This Paper[a]

Model	Configuration	Notes	Section
FOAM	Snowball	Idealized continent surrounded by sea ice with dust strips, modeled as either continent or low-albedo sea ice, of varying sizes added in some cases	2.1, 4
CAM	Modern	Modern continents with different MLD and OHT assumptions	2.1
CAM	Snowball	Idealized unglaciated continent surrounded by sea ice	5
CAM	Snowball	Idealized glaciated continent surrounded by sea ice	5
CAM	Snowball	Idealized continent surrounded by glacial ice	5

[a]We run the NCAR's Community Atmosphere Model v3.1 (CAM) at T42 horizontal resolution with 26 vertical levels. We run the Fast Ocean Atmosphere Model v1.5 (FOAM) at R15 horizontal resolution with 18 vertical levels.

be coincidental that the energy balance model result agrees reasonably well with GCM results, given the fairly arbitrary assumptions about clouds, snow cover, and the effective horizontal diffusivity that must be made in an energy balance model. Furthermore, even though the parameterizations in GCMs are in some ways closer to the basic physics, they still entail a certain amount of arbitrariness. Nonetheless, these results indicate that the $p\mathrm{CO_2}$ observations suggest prevailed at the end of the last Snowball event may not be high enough to produce deglaciation in climate models.

[6] The only climate model that has produced a Snowball deglaciation at a $p\mathrm{CO_2}$ below 0.1 bar is an energy balance model coupled to a dynamic ice sheet model used by *Pollard and Kasting* [2005]. When the bare sea ice albedo is relatively low this model produces a solution with thin (O(1 m)) tropical sea ice that deglaciates at $p\mathrm{CO_2} \sim 0.01$ bar. Surface albedo was also found to be extremely important by *Lewis et al.* [2006], who showed that small changes in surface albedo parameterization can lead to changes by an order of magnitude in the $\mathrm{CO_2}$ concentration required for Snowball deglaciation in the University of Victoria Earth System Climate Model. In this paper we suggest that the Snowball hydrological cycle would drive sea glaciers in a way that would lead to the development of a dust layer at the ice surface in the tropics that would significantly lower the tropical albedo and allow deglaciation at $p\mathrm{CO_2} = 0.01$–0.1 bar. Our proposal is a specific mechanism for lowering the surface albedo, and as we will discuss below, we propose a specific geological test for this idea.

[7] We begin by discussing the reasons for believing that such a dust layer would form during a Snowball (section 2) and arguing that the expected dust accumulation rates could produce a layer of quite significant depth (section 3). We then show that when such a tropical dust layer is added to FOAM v1.5 in the same configuration used by *Pierrehumbert* [2004], it leads to Snowball deglaciation for $p\mathrm{CO_2}$ = 0.01–0.1 bar (section 4). In section 5 we confirm that our main conclusions from section 4 are valid in the National Center for Atmospheric Research (NCAR) Community Atmosphere Model (CAM) v3.1, although the situation is somewhat more complicated. Finally, we discuss the implications and limitations of these results (section 6) and conclude (section 7). Table 1 presents a summary of the models and configurations we use in this paper.

2. Reasons for the Formation of a Tropical Dust Strip

2.1. Dynamics of the Planetary Net Ablation Zone

[8] A planetary net ablation zone (annual average evaporation (E) greater than precipitation (P)) within 10°–25° of the equator is a consistent feature of Snowball simulations (Figure 1a) [e.g., *Pierrehumbert*, 2005; *Pollard and Kasting*, 2005; *Goodman and Pierrehumbert*, 2003]. This is in stark contrast with the situation in the modern climate, in which $P > E$ in the deep tropics and $E > P$ throughout the subtropics (Figure 2) [e.g., *Peixoto and Oort*, 1992]. There are four main differences in surface boundary conditions on atmospheric dynamics between the Snowball and modern that cause differences in the hydrological cycle: the Snowball surface temperature is much lower; far less energy is absorbed at the surface in a Snowball; the Snowball surface heat capacity is much lower; and there is no surface (ocean) heat transport in a Snowball. In this section we will briefly explore how these differences produce a planetary net ablation zone, which deserves thought because it is crucial for the mechanism discussed in this paper and because it is so different from the situation in the modern climate.

[9] Since global and annual mean precipitation is constrained by energy absorbed at the surface [*Pierrehumbert*, 2002; *Held and Soden*, 2006], it would be greatly reduced in a Snowball because of the high surface albedo (Figure 1a). The magnitude of the $P - E$ pattern is reduced even more, by roughly 2 orders of magnitude, such that P nearly equals E at all latitudes in the zonal average. This is because specific humidity is constrained by the Clausius-Clapyron relation to be very low in a Snowball so that even if atmospheric circulation is not greatly affected by the transition to the Snowball climate, the air cannot transport nearly as much moisture.

[10] The reduction in the surface heat capacity appears to be the most important factor leading to the net ablation zone in the tropics. For example, an equatorial net ablation zone can be produced in NCAR's Community Atmosphere Model (CAM) v3.1 (T42 horizontal resolution, 26 vertical levels) when it is run in modern configuration with a slab ocean simply by reducing the mixed layer depth (Figure 2). In contrast, the removal of ocean heat transport concentrates the maximum in $P - E$ on the equator but does not change the qualitative shape of $P - E$.

[11] The reason reducing the surface heat capacity leads to a net ablation zone on the equator is that it causes a "monsoonal" circulation similar to that found at reduced surface heat capacity by *Bordoni and Schneider* [2008] and *Schneider and Bordoni* [2008]. The region of low-level convergence extends much further into each hemisphere at the solstice and spends relatively little time near the equator so that the annual-mean $P - E$ pattern looks more like the sum of the solstitial $P - E$ patterns than it does the equinoctial $P - E$ pattern (Figure 1). For example, in the CAM 50 m mixed layer depth (MLD), zero ocean heat transport (OHT)

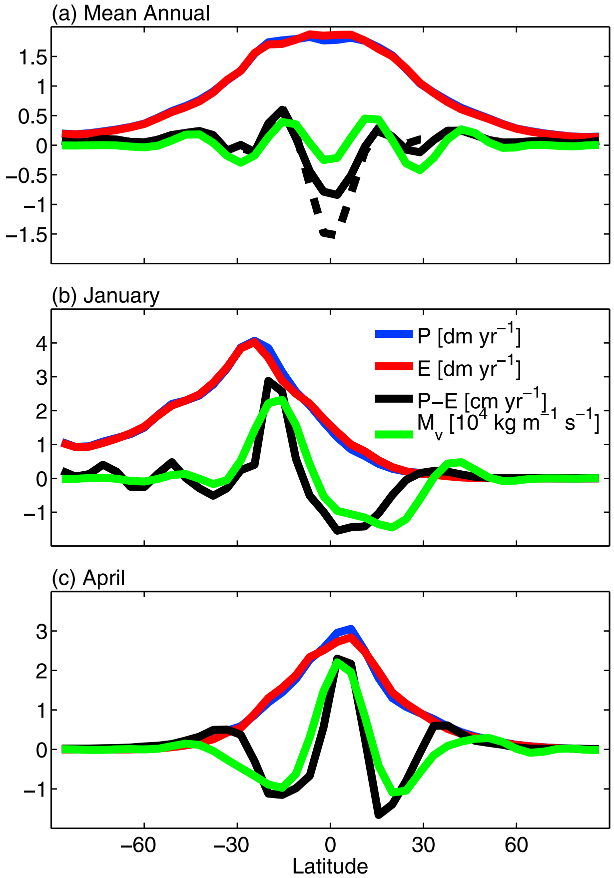

Figure 1. Plots of zonally averaged precipitation minus evaporation in a FOAM Snowball simulation after the fashion of *Pierrehumbert* [2005] with $pCO_2 = 0.1$ bar reveal the distinctive "planetary net ablation zone" in the deep tropics. Averaging period is (a) annual, (b) the month of January, and (c) the month of April. Plotted variables are precipitation in units of decimeters per year (blue), evaporation in units of decimeters per year (red), difference between precipitation and evaporation in units of centimeters per year (black), and zonally integrated vertical mass flux at a pressure of 866 mbar in units of 10^4 kg m^{-1} s^{-1} (green). The black dashed line in Figure 1a is the precipitation minus evaporation averaged over ice only.

simulation of Figure 2, the maximum in low-level vertical mass flux occurs at 1.4°S in January (the closest grid point to the equator), whereas in the CAM 1 m MLD, zero OHT simulation it occurs at 20.9°S. In the Snowball simulation of Figure 1, the maximum in low-level convergence in January occurs at 15.5°S. This value is consistent with that predicted (15.4°S) by the linear model of low-level convergence of *Schneider and Bordoni* [2008], if a damping coefficient of (0.3 d)$^{-1}$ is used based on a rough fit of the meridional momentum equation. Therefore, we can conclude that the main reason for the tropical net ablation zone in Snowball simulations is the reduced surface heat capacity, and we have a reasonable understanding of the dynamical explanation for this.

2.2. Consequences for Tropical Dust Accumulation

[12] If the Snowball is in steady state, then mass balance requires some source of tropical ice to replace that lost by net sublimation at the surface. Even if the Snowball is not completely in equilibrium, it must be nearly so, and there should be some significant source of ice to the tropics, because otherwise the tropical ice would simply sublimate away. For example, if the tropical net sublimation rate is on the order of 0.01 m yr^{-1} (Figure 1a), it would only take 10^5 yr to consume 1 km of ice, a conservatively high estimate of tropical ice thickness [*Goodman and Pierrehumbert*, 2003], which is only on the order of 10% of a conservatively short Snowball lifetime (O(10^6 yr)). A Snowball lifetime of O(10^6 yr), which we will use in this section and section 3, is consistent with *Hoffman and Li* [2009] (see section 1) and makes the estimates presented here conservative. Therefore, at the very minimum, 90% of the ice lost from the tropics due to net sublimation has to be replaced somehow, if the system is to stay in a Snowball.

[13] The tropical ice that is lost because of net sublimation must be replaced in one of the following ways: (1) net freezing at the bottom of the ice sheet, (2) ice sheet flow from higher latitudes, or (3) some combination of the first two. We argue below that both cases (1) and (2), and consequently case (3), must lead to the accumulation of dust at the ice surface in the tropics over the lifetime of the Snowball (see Figure 3 for a schematic diagram).

[14] Net freezing at the bottom of the ice sheet is the only process that can replace sublimated tropical ice in the absence of horizontal ice flow. In this case ice will be advected upward continually, a process, first mentioned by *Hoffman*

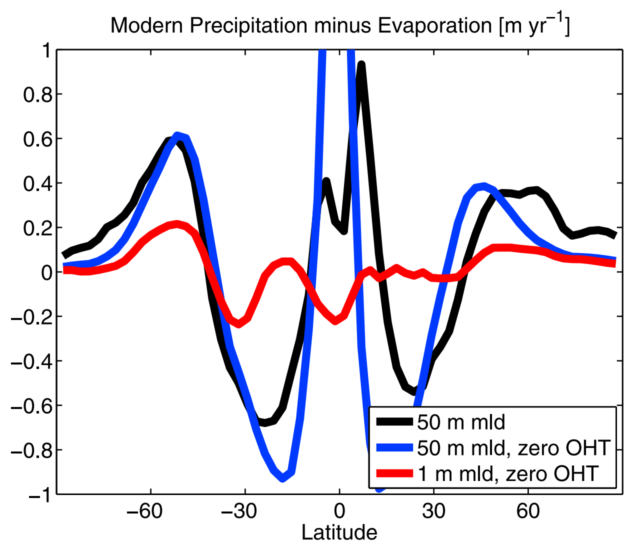

Figure 2. Annually and zonally averaged precipitation minus evaporation in simulations using NCAR's Community Atmosphere Model (CAM) v3.1 run at T42 horizontal resolution with 26 vertical layers, modern boundary conditions, and the following mixed layer depths (MLD) and ocean heat transports (OHT): MLD = 50 m, modern OHT (black); MLD = 50 m, OHT = 0 everywhere (blue); and MLD = 1 m, OHT = 0 everywhere (red). Magnitudes of maxima and minima in precipitation minus evaporation are smaller in the MLD = 1, OHT = 0 case because the sea ice edge advances and the surface temperature is lower than in the other cases.

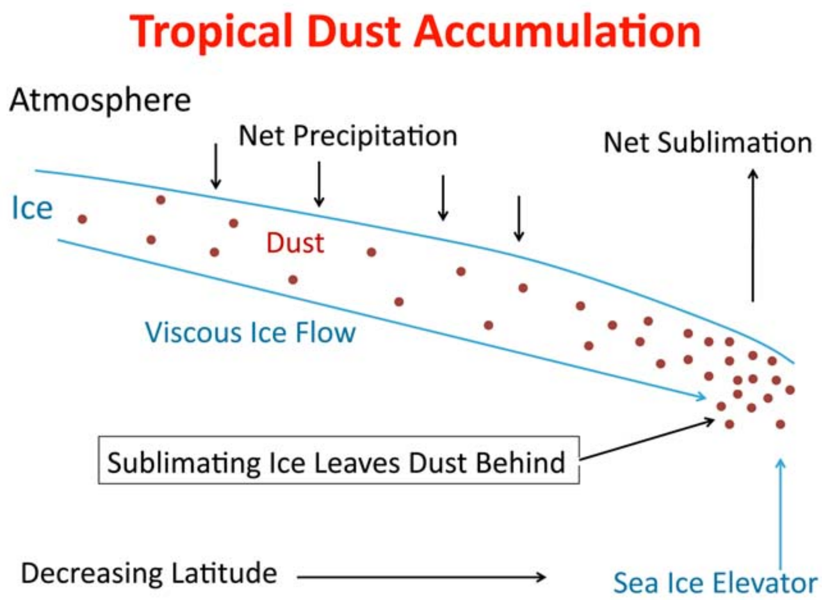

Figure 3. Schematic diagram showing how the "sea ice elevator" [*Hoffman and Schrag*, 1999] and viscous flow of "sea glaciers" [*Goodman and Pierrehumbert*, 2003] would lead to the accumulation of dust at the ice surface in the low-latitude "planetary net ablation zone" (see section 2).

and Schrag [1999], that *Goodman and Pierrehumbert* [2003] termed the "sea ice elevator." Given a net sublimation rate of 0.01 m yr^{-1}, an ice thickness of 10^3 m, and a Snowball lifetime of 10^6 yr, the equatorial ice would be replaced 10 times over the course of the Snowball. Therefore, the timescale of upward advection of ice is relatively short compared to the lifetime of the Snowball, and it is reasonable to assume that all dust deposited in the ice over the lifetime of the Snowball would be trapped near the surface. That said, we will discuss effects that could potentially disrupt this surface dust buildup, such as the formation of cryoconite holes, in section 6.

[15] Given the large thickness to which "sea glaciers" are expected to grow during a Snowball, viscous ice flow from the net snow accumulation region in the extratropics into the net sublimation region in the tropics might occur [*Goodman and Pierrehumbert*, 2003]. This process could offset or replace the net freezing at the bottom of the ice sheet in the equatorial region [see *Goodman*, 2006, Figure 1c]. In fact, so much ice flow might occur that melting rather than freezing might occur at the bottom of the equatorial ice sheet [*Goodman and Pierrehumbert*, 2003]. In any case, as *Goodman* [2006] briefly mentions, this process should lead to slow accumulation of dust in the tropics over the lifetime of the Snowball. Not only would the dust that falls in the tropics be trapped at the surface, as discussed in the previous paragraph, but all the dust that falls on top of the ice that flows into the tropics could also be trapped at the surface of the tropical ice. Since the net sublimation rate in the tropics is governed by meteorological processes, we expect it to be the same regardless of the source of the replacement ice. Therefore, given a net sublimation rate of 0.01 m yr^{-1} and a Snowball lifetime of 10^6 yr, roughly 10^4 m of ice should sublimate and pass through the ice-air interface in the net ablation zone over the lifetime of a Snowball event, leaving behind any deposited dust as it goes, and significantly decreasing the albedo.

3. Thinking About the Dust Accumulation Rate

[16] There are three main potential sources of dust that could lead to significant deposition on the ice surface during a Snowball: (1) continental dust, (2) volcanic dust, and (3) cosmic dust. In this section we will make rough estimates for the magnitude of the dust flux from each of these sources during a Snowball. The purpose of this section is to show that it is reasonable to think that a significant amount of dust could accumulate in the tropics at the ice surface over the lifetime of a Snowball, rather than to make an accurate estimate of the exact amount of dust that would accumulate.

[17] *Kirschvink* [1992] suggested that there could be large unglaciated continental regions during a Snowball because of a significantly reduced hydrological cycle. Climate models have produced conflicting results as to whether perennial snow, which would lead to glaciation, would accumulate on continents during a Snowball event [e.g., *Donnadieu et al.*, 2003; *Pierrehumbert*, 2004, 2005], and they may not be the best tool for this purpose since they have difficulty reproducing regions of perennial snow cover during last glacial maximum [*Krinner et al.*, 2006]. In any case, it is certain that any unglaciated continental regions in a Snowball would produce huge dust fluxes since they would be extremely dry, have no vegetation, and be subjected to large seasonal and diurnal temperature cycles that would cause soil fracturing and cryogenic weathering [*Konishchev*, 1982]. A starting point is to imagine that

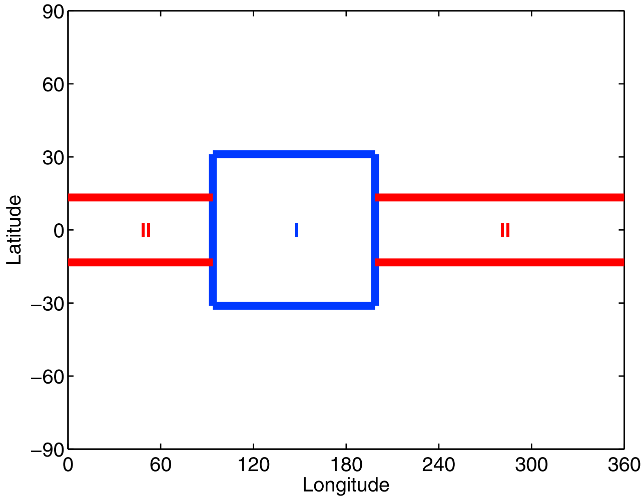

Figure 4. Map of idealized continental configuration used in the FOAM runs. Region I is the idealized supercontinent corresponding to that used *Pierrehumbert* [2004], which has no vegetation and an albedo of 0.4 in the near infrared and 0.24 in the visible. The supercontinent occupies 1/6 of Earth's surface area, which compares to a modern land fraction of roughly 1/4. Region II is the tropical dust strip. We investigate the effect of modifying the size and albedo of this strip in section 4.

the dust flux at low latitudes would be similar to that during the last glacial maximum, or 2–20 times higher than the modern value of roughly 1 g m^{-2} yr^{-1} [*Harrison et al.*, 2001]. If we assume a dust density of 2 g cm^{-3}, this is equivalent to a dust accumulation rate of 10^{-6}–10^{-5} m yr^{-1}.

[18] The long-term global average tephra flux, which includes dust and larger objects, from volcanoes is about 35,000 kg s^{-1} [*Mason et al.*, 2004]. If we assume the tephra has an average density of 2 g cm^{-3}, that a significant fraction of the tephra is small size, that the tephra is spread evenly over the surface area of the planet when averaged over a large number of eruptions, and that the volcanic flux during a Snowball was roughly the same as it is now, we get a volcanic dust flux of 10^{-6} m yr^{-1}. The magnitude of this volcanic dust flux is important because it represents a baseline dust flux that we would expect even if the continents are totally glaciated.

[19] Currently, the cosmic dust flux is about 1.5×10^{8} kg yr^{-1} [*Ceplecha*, 1996], which is equivalent to 1.5×10^{-10} m yr^{-1} if we assume the dust is spread evenly over the planet and has a density of 2 g cm^{-3}. Therefore, unless the cosmic dust flux was significantly larger during Snowball Earth episodes than it is today, it is negligible when compared to the continental and volcanic dust sources.

[20] On the basis of the rough order-of-magnitude estimates made above, it seems reasonable to assume that the globally averaged dust flux during a Snowball would be roughly 10^{-6}–10^{-5} m yr^{-1}. Even neglecting horizontal concentration, this should lead to the accumulation of 1–10 m of dust at the ice surface in the tropics over a 10^{6} yr Snowball by the mechanisms discussed in section 2. It seems highly probably that 1–10 m of dust would have a significant effect on the surface albedo and potentially other important processes such as evaporation.

4. Tropical Dust Snowball Escape Mechanism in FOAM

4.1. Plausibility of the Idea as Indicated by FOAM Tests

[21] *Pierrehumbert* [2004] and *Pierrehumbert* [2005] found that the FOAM v1.5 GCM [*Jacob*, 1997] (run at R15 horizontal resolution with 18 vertical levels) is far from deglaciation when it is run with a large idealized tropical supercontinent, put in a Snowball state, and pCO_2 is increased as high as 0.2 bar. We argue that the development of a tropical dust layer in the planetary net ablation zone over the lifetime of a Snowball event could significantly alter this conclusion. In order to test this idea, we run the same version of FOAM with the same continental configuration as *Pierrehumbert* [2004], but with a parameterization of a tropical dust layer added (Figure 4), to see whether this dust layer leads to deglaciation.

[22] We parameterize the dust layer by setting the sea ice albedo within an equatorial strip (defined as the region where the absolute value of the latitude is less than λ_0, region II in Figure 4) to a value of α_0 in both the visible and near infrared. We define a deglaciation as occurring if the tropical sea ice melts through to ocean, which in FOAM always leads to the loss of all sea ice globally. This definition of a deglaciation event is a reasonable way to proceed, but we note that there are many assumptions in the sea ice parameterizations of FOAM, and other GCMs, that may limit their applicability to Snowball deglaciation. We consider these issues in section 6.

[23] The albedo in the tropical dust strip would depend on the dust source and the admixture of dust with snow and ice. The modern surface albedo in continental regions, which reflects soil and vegetative cover, ranges from 0.1 to 0.3 [*Gill*, 1982]. During a Snowball event, when there would be no vegetation, the continental albedo would probably be on the higher end of this range. For comparison, in FOAM the bare continental surface in region I of Figure 4 has an albedo of 0.4 in the near infrared and 0.24 in the visible, for a broadband albedo of 0.34, assuming 40% of the sunlight reaching the surface is visible [*Gill*, 1982]. Additionally, there is observational evidence that volcanic ash can reduce the average surface albedo of glaciers to 0.1–0.2 [*Reijmer et al.*, 1999]. We therefore consider values of α_0 ranging from 0.1 to 0.5.

[24] Inspection of Figure 1a indicates that $\lambda_0 \approx 10°$–$15°$ would provide a reasonable approximation of the planetary net ablation zone. If we run FOAM with $\lambda_0 = 10°$ and reduce the surface albedo somewhat, however, we find that the planetary net ablation zone expands greatly so that $\lambda_0 = 30°$ would be more reasonable (Figure 5). This results mainly from the increased ability of eddies to dry the subtropical region when the air is warmer, which overcomes the ability of low-level convergence associated with tropical circulation to wet this region as discussed in section 2.1. Consequently we consider values of λ_0 ranging from 10° to 30°.

[25] The addition of a tropical dust strip to FOAM leads to Snowball deglaciation at pCO_2 = 0.1 bar for a wide range

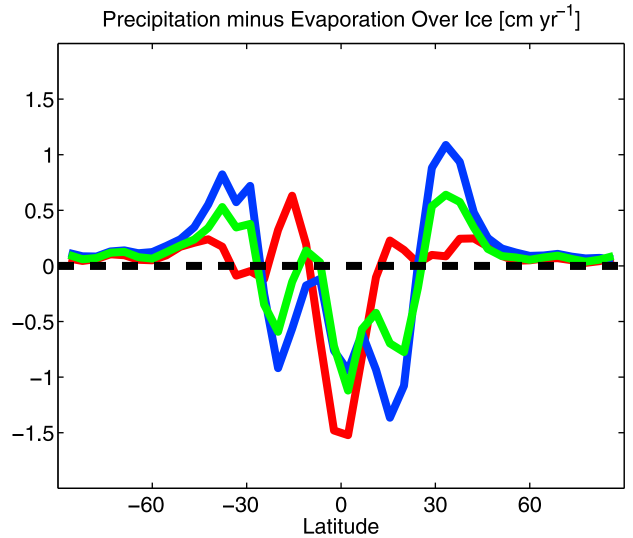

Figure 5. Expansion of the "planetary net ablation zone" as the albedo of the tropical dust strip is lowered. Precipitation minus evaporation is annually and zonally averaged over ice and plotted as a function of latitude in the Snowball state with CO_2 = 0.1 bar and no tropical dust strip (red), a tropical dust strip with λ_0 = 10° and α_0 = 0.4 (green), and a tropical dust strip with λ_0 = 10° and α_0 = 0.3 (blue). The red curve is the same as the black dotted curve in Figure 1a.

of dust strip properties which, on the basis of the discussion above, we believe could reasonably be expected to develop (Figure 6). We view these results as a demonstration of the plausibility of the idea that the development of a tropical dust strip in the planetary net ablation zone could lead to

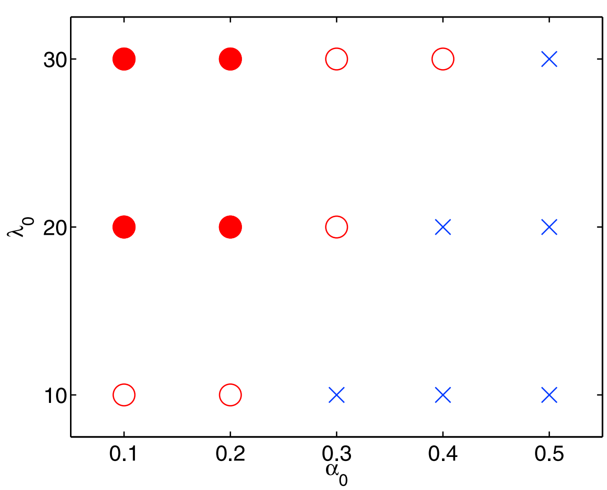

Figure 6. Snowball deglaciation as a function of the properties of the tropical dust strip. The dust strip is defined have an albedo of α_0 and to exist where the absolute value of the latitude is less than λ_0. An open red circle denotes deglaciation in FOAM for the particular dust strip properties at pCO_2 = 0.1 bar, a blue cross denotes the absence of deglaciation at pCO_2 = 0.1 bar, and a filled red circle denotes deglaciation even at pCO_2 = 0.01 bar.

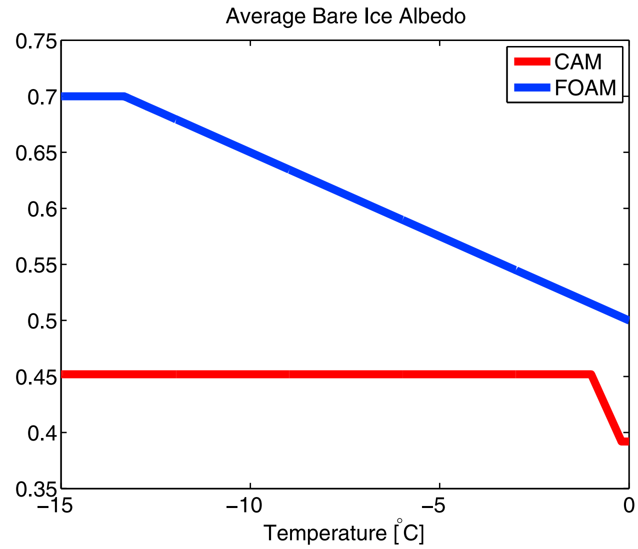

Figure 7. Temperature dependence of broadband bare sea ice albedo in FOAM and CAM for sea ice thicker than 1 m.

Snowball deglaciation. Deglaciation is even possible in FOAM at pCO_2 = 0.01 bar for more extreme values of λ_0 and α_0.

[26] Thus far our investigation of the effects of a tropical dust strip has been limited to changing the surface albedo in the sea ice parameterization. An O(10 m) thick layer of dust on top of sea ice might have surface heat capacity and evaporative properties more like land than darkened ice. To test how important such effects might be, we alternatively parameterized the dust strip as continent, so that region II in Figure 4 was exactly the same as region I, chose a dust strip width of λ_0 = 15°, and ran the model for a range of CO_2 values. Defining deglaciation as occurring if this continent heated up enough to melt surrounding sea ice and cause global sea ice loss, deglaciation occurred for pCO_2 = 0.1 bar in this case, but not for pCO_2 = 0.01 bar. Given that the continental broadband albedo is 0.34, this change in surface parameterization made deglaciation somewhat easier than when we only changed the ice albedo (Figure 6). Therefore, we can conclude that, at least in FOAM, a tropical dust strip with continental surface properties should promote Snowball deglaciation at least as well as a tropical dust strip with surface properties that resemble darkened ice.

4.2. Comparison With Previous FOAM Results

[27] *Pierrehumbert* [2005] found almost no effect on the Snowball climate for large changes in the bare sea ice albedo. Therefore, some discussion of the reason we found such a significant effect is necessary. We show the temperature dependence of the bare sea ice broadband albedo for sea ice thicker than 1 m in FOAM in Figure 7, as well as a comparison with the parameterization used in CAM. The decrease in albedo with temperature near the freezing point of water is meant to simulate changes in albedo due to the formation of melt pools on the ice surface. This effect is larger and extends over a larger temperature range in FOAM than in CAM. Additionally, FOAM has a higher sea ice albedo for all temperatures than does CAM, which we will return to in section 5. The reason we found that changing

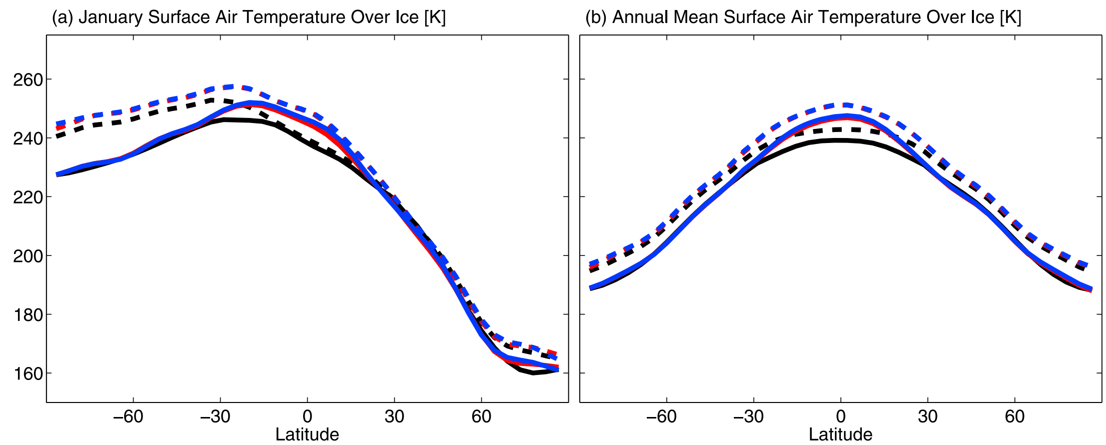

Figure 8. Zonally averaged surface air temperature over ice (a) in January and (b) in the annual mean for FOAM simulations with different bare sea ice albedo parameterizations. Here $p\mathrm{CO}_2 = 10^{-4}$ bar for solid lines and $p\mathrm{CO}_2 = 0.1$ bar for dashed lines. The albedo values are standard FOAM (black), visible albedo of 0.5 and near-infrared albedo of 0.5 (red), and visible albedo of 0.68 and near-infrared albedo of 0.3 (blue) (these are the values used in CAM). In the latter two cases we have disabled the temperature dependence of the sea ice albedo to more closely approximate the CAM sea ice albedo scheme, which has only a very weak temperature dependence (Figure 7).

the sea ice albedo can have a large effect on deglaciation whereas *Pierrehumbert* [2005] did not is that we changed the low-temperature albedo (0.7) while *Pierrehumbert* [2005] changed the near-freezing albedo (0.5). Since the temperature only rarely or never reaches −13.3°C in the Snowball climate, the temperature at which the sea ice albedo begins to ramp downward, the change that *Pierrehumbert* [2005] made has almost no effect on the Snowball climate.

[28] Given the difference between the CAM and FOAM bare sea ice albedo scheme, we feel it is important to repeat the simulations presented by *Pierrehumbert* [2004] and *Pierrehumbert* [2005] with the FOAM sea ice albedo scheme replaced with one more similar to that in CAM. We present the results of these simulations in Figure 8. The temperature in the planetary net ablation zone increases somewhat, but the temperature is hardly changed poleward of 30° because this region is covered with snow. Most importantly, the main result of *Pierrehumbert* [2004] and *Pierrehumbert* [2005], that large increases in the CO_2 (up to 0.1 bar) are not sufficient to cause Snowball deglaciation in FOAM, still holds.

5. CAM Simulations and Comparison With FOAM

[29] Here we describe simulations we have performed using NCAR's CAM v3.1 (T42 horizontal resolution, 26 vertical levels) that are similar to the FOAM simulations presented in section 4 in order to show that a tropical dust layer can promote Snowball deglaciation in a model with more sophisticated physical parameterizations (FOAM's physical parameterizations are based on Community Climate Model, Versions 2 and 3, earlier versions of CAM) run at higher resolution. Because of the high computational cost of running CAM, we only repeat a sampling of the simulations presented in section 4.

[30] We run CAM in three different configurations: (1) with a tropical supercontinent of the same size and in the same position as that in the FOAM runs surrounded by sea ice, (2) with a glaciated tropical supercontinent surrounded by sea ice, and (3) with a tropical supercontinent surrounded by glacial ice rather than sea ice. For the unglaciated continental region we assume that the soil is 50% clay and 50% sand, that there is no vegetation, and that the continental albedo is 0.4 in the near infrared and 0.24 in the visible, as in FOAM. Glacial ice is a land surface boundary condition in CAM whereas sea ice is calculated by the slab ocean scheme. These two types of ice have different properties, the most important of which is that glacial ice has a broadband albedo of 0.65 as opposed to the broadband albedo of 0.45 for sea ice. We might associate CAM's sea ice with the"marine ice" and CAM's glacial ice with the "meteoric ice" discussed by *Goodman* [2006].

[31] CAM produces vastly different climates when run in these different configurations (Figure 9). When we run CAM with a tropical supercontinent similar to that which we used in the FOAM simulations and $p\mathrm{CO}_2 = 10^{-4}$ bar, the equilibrated surface temperatures are significantly higher than in FOAM (Figure 9a). In fact, the tropical surface temperatures in CAM with $p\mathrm{CO}_2 = 10^{-4}$ bar are 5–10°C warmer than the FOAM surface temperatures at $p\mathrm{CO}_2 = 0.1$ bar in the same configuration. This is due to the much lower bare sea ice albedo (Figure 7) and to a stronger cloud radiative forcing (Figure 9b) in CAM than FOAM. When $p\mathrm{CO}_2$ is increased to 0.1 bar in CAM in this configuration, the sea ice retreats completely (the model deglaciates).

[32] When CAM is run with a glaciated supercontinent the resulting surface air temperatures and the effect of increasing CO_2 to 0.1 bar are roughly similar to those in the standard FOAM run (Figure 9a). Warming due to increased cloud radiative forcing relative to FOAM (Figure 9b) is offset by the high albedo of the glaciated continent. The

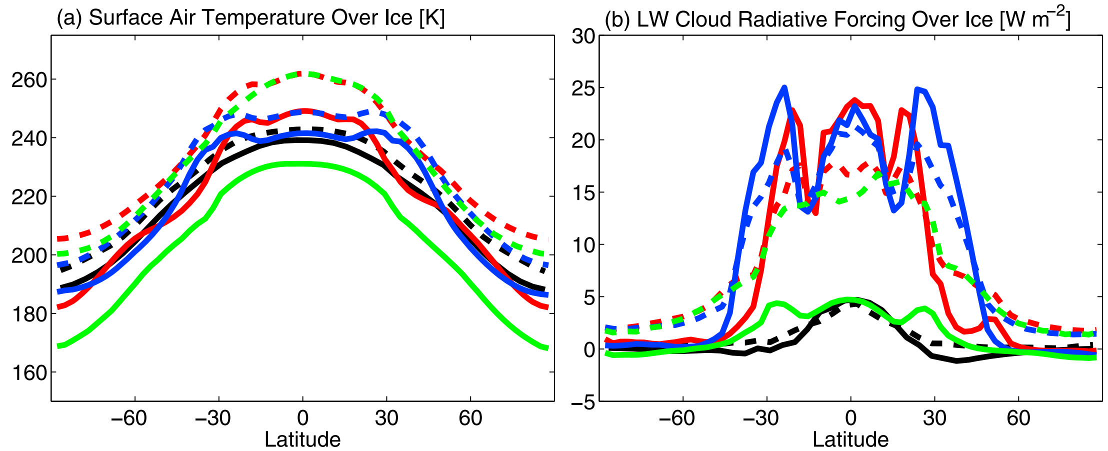

Figure 9. (a) Zonal and annual mean surface air temperature over ice and (b) zonal and annual mean longwave (LW) cloud radiative forcing over ice for FOAM and CAM simulations. Here $p\mathrm{CO_2} = 10^{-4}$ bar for solid lines and $p\mathrm{CO_2} = 0.1$ bar for dashed lines. The simulations are: standard FOAM (black), CAM run with similar boundary conditions as the standard FOAM simulation (red), CAM run with a glaciated tropical supercontinent (blue), and CAM run with a tropical supercontinent similar to that in FOAM that is surrounded by glacial ice rather than sea ice (green). CAM results from the tropical supercontinent surrounded by sea ice run with $p\mathrm{CO_2} = 0.1$ bar (red dashed line) are averaged over the first 3 years, before significant sea ice loss occurs. We plot only the longwave component of the cloud radiative forcing here because net cloud radiative forcing is strongly affected by the surface albedo and therefore not as useful for comparison between these simulations.

subtropical reversals of the meridional temperature gradient in this case are caused by a change in the $P - E$ pattern (Figure 10), with additional net ablation zones in the subtropics (leading to exposure of low-albedo bare sea ice). These appear to develop because increased atmospheric temperature increases the specific humidity so that eddies can transport more moisture from the subtropics to higher latitudes. At the same time, the higher subtropical atmospheric temperatures are due to the fact that eddies are transporting more moisture poleward from the subtropics and exposing bare sea ice. This feedback appears to result from the high contrast between the broadband snow (0.79) and bare ice (0.45) albedos in CAM, which compares to albedos of 0.72 for snow and 0.70 for bare ice in FOAM. In any case, we would still expect dust to accumulate in the surface of any of the net ablation zones. A simple test of whether such a development could lead to deglaciation is to halve the broadband bare sea ice from 0.45 to 0.23, which only affects the net surface albedo in net ablation regions where there is no snow cover. This modification does result in deglaciation at $\mathrm{CO_2} = 0.1$ bar.

[33] When we run CAM in the final configuration, with a tropical supercontinent surrounded by glacial ice rather than sea ice, the resulting $P - E$ pattern contains a distinct and broad tropical net ablation zone (Figure 10). This result is consistent with that found in previous studies [*Pierrehumbert*, 2005; *Pollard and Kasting*, 2005; *Goodman and Pierrehumbert*, 2003] and with FOAM results here. Using glacial ice rather than sea ice results in a significantly cooler climate (Figure 9a) because of the lower albedo of glacial ice and lower cloud radiative forcing (Figure 9b). Additionally, increasing $\mathrm{CO_2}$ causes a much stronger increase in surface temperature than in FOAM or in the other CAM configurations. This is due, at least in part, to a positive cloud radiative forcing feedback (an increase in cloud radiative forcing as the surface temperature increases) (Figure 9b). Such a feedback does not occur in the other model configurations (Figure 9b). Despite the strong warming in this configuration when $p\mathrm{CO_2}$ is increased to 0.1 bar, the surface air temperature is still below freezing over all the ice regions (Figures 11c and 11d). Since we use glacial ice, which cannot melt away in CAM, rather than sea ice in this configuration, we can no longer use the same deglaciation test. If the surface air temperature were above freezing over

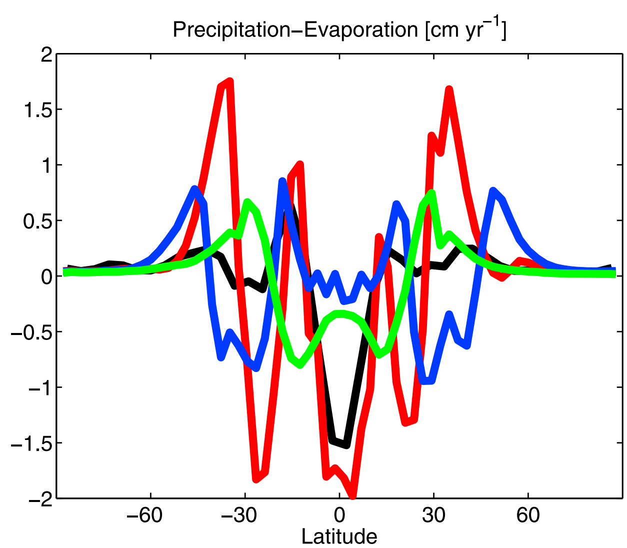

Figure 10. Zonal and annual average precipitation minus evaporation ($P - E$) over ice in FOAM and CAM with $p\mathrm{CO_2} = 10^{-4}$ bar. Colors are as in Figure 9.

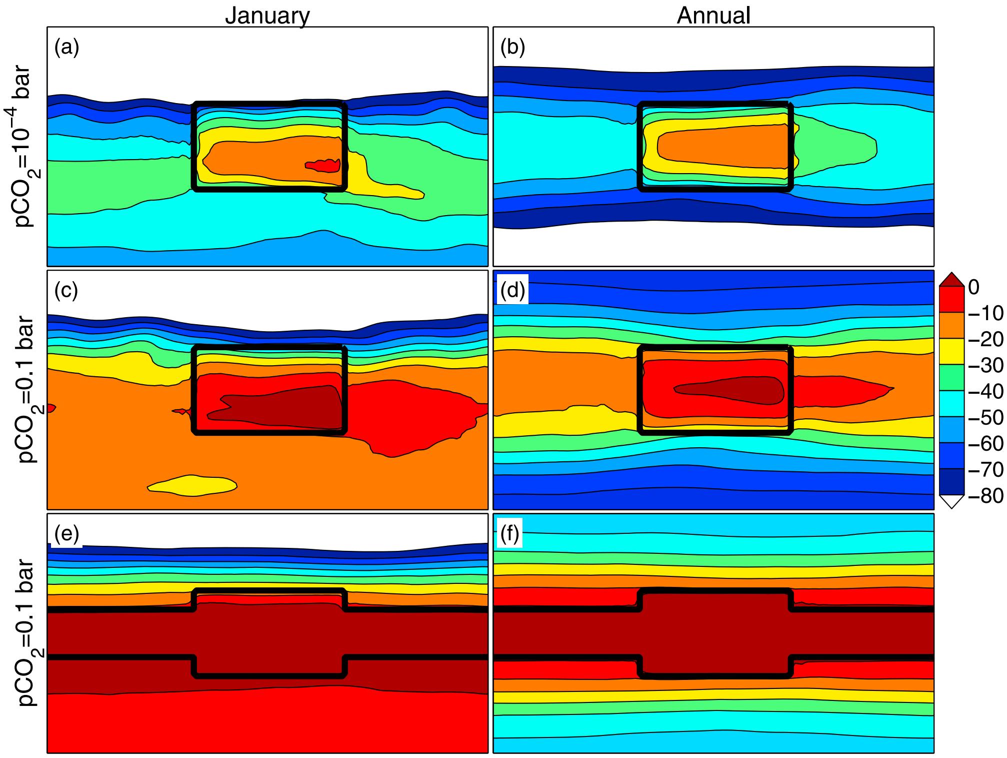

Figure 11. Surface air temperature in degrees Celsius (left) in January and (right) in the annual mean as a function of longitude (horizontal axis) and latitude (vertical axis, which is linear and stretches from the South Pole to the North Pole) in CAM simulations with an idealized tropical supercontinent that is surrounded by glacial ice rather than sea ice. Here pCO_2 is (a and b) 10^{-4} bar and (c and d) 0.1 bar. (e and f) The continental region is extended to include all the area within 15° of the equator to simulate a tropical dust strip and pCO_2 = 0.1 bar. In Figures 11a–11f the continental outline is drawn as a thick black line.

a region of ice in the annual average, this would be a good indication that deglaciation could occur, so we will use this as a working deglaciation test for this configuration. By this definition, increasing the CO_2 to 0.1 bar does not cause deglaciation (Figure 11d). We note, however, that the surface air temperature over ice in this case is roughly as high as when pCO_2 = 0.1 bar when the tropical supercontinent is surrounded by sea ice (Figure 9a), which causes the sea ice to retreat. We discuss the implications of this in section 6.

[34] We add a tropical dust strip to the glacial ice simulation in the following simple way: we expand the continent so that it stretches around the globe within 15° of the equator (Figures 11e and 11f). As discussed in section 4.1, this involves assuming the ice and dust mixture in the dust strip would take on properties more characteristic of continent than ice. The result of this change is that the annual mean surface air temperature exceeds freezing throughout the dust strip region (Figure 11f), implying that the dust strip could enable deglaciation.

[35] To summarize, although the results from CAM are more nuanced than those from FOAM, it appears that the main conclusions from our work with FOAM hold: (1) when CAM is in a Snowball, it is difficult to cause deglaciation, though in some configurations possible, simply by increasing the pCO_2 to 0.1 bar and (2) the formation of a dust strip in the planetary net ablation zone can produce warming sufficient to lead to deglaciation when it is not otherwise possible.

6. Discussion

[36] A continuous clay drape has been found at the top of the 635 Ma glacial deposits, directly below the cap dolostone, at sites in the Mackenzie mountains of northwest Canada, the Yangtze Gorges area of South China, and in northeast Svalbard (P. Hoffman, personal communication, 2009). By the arguments of section 2, we would expect such a layer to form preferentially in tropical regions if the hypothesis advanced in this paper were operating during this Snowball event. This suggests a geological test for our hypothesis: the thickness or presence/absence of such clay drapes could be documented as a function of paleolatitude for the Snowball events. We predict that the clay drapes should be thicker in tropical than in extratropical regions, or not present in extratropical regions at all.

[37] It is somewhat disturbing how different simulations of the Snowball in FOAM and the different configurations of CAM are from each other. The major causes of these differences are different surface albedos, all of which are within the range that could potentially be relevant for Snowball sea glaciers [*Warren et al.*, 2002], and different cloud parameterizations, which should be viewed extremely skeptically when applied to the Snowball climate. All of this

reflects the fact that we should imbue no climate model with undue respect and reverence and must be particularly careful when using such models to simulate a climate so vastly different from our own.

[38] The deglaciation test we used when we parameterized the Snowball ice as sea ice was that a deglaciation occurred if the sea ice melts through to ocean in the tropics. In CAM and FOAM this generally occurs when the annual-mean surface air temperature reaches roughly $-10°C$. There are two important reasons for this. First, meltwater, even meltwater produced during diurnal cycles, is assumed to disappear down moulins or into remote ponds rather than remaining to refreeze. This allows the sea ice to "ratchet" away in a manner that might not be appropriate for thick sea glaciers. Second, the solar penetration parameterization allows the ice interior temperature to be warmer than the surface temperature at times, which helps melting. Given the challenges of accurately representing solar penetration effects [*Warren et al.*, 2002], we cannot be sure that the parameterizations in CAM and FOAM accurately represent the behavior of Snowball ice. In contrast, when we parameterize Snowball ice as glacial ice, we assume deglaciation occurs when the annual-mean surface air temperature reaches 0°C, which we can view as a hard limit at which deglaciation must occur. More study is needed to determine what tropical surface air temperatures would be required to trigger a Snowball deglaciation. Finally, we note that in both ice cases we neglect the glacial dynamical effects considered by *Goodman and Pierrehumbert* [2003] that could potentially be important for the thick sea glaciers of a Snowball Earth.

[39] Some of the arguments presented here may appear to be at odds with *Pierrehumbert* [2005], who argued that the flow of sea glaciers into the tropics could raise the albedo rather than lower it. The argument given there, however, is based on the fact that sea glaciers should have a higher albedo than tropical "bare" ice and does not factor in the slow accumulation of dust over the lifetime of the Snowball.

[40] It might be suggested that local "recycling" of snow, sublimated ice that falls as snow at a similar latitude, might cover the accumulated dust of the planetary terminal moraine. Indeed, roughly 90% of the ice that sublimates in the tropics would fall as snow there rather than be exported to higher latitudes (Figure 1). This would be a reasonable argument for a small amount of dust; however, it seems that meters of dust would likely overwhelm the ability of this process to simply cover it up. Additionally, surface sublimation would likely be significantly reduced or eliminated altogether once the admixture of dust with ice in the dust strip reached a certain fraction, which would prevent this process.

[41] Because of pressure broadening of CO_2 absorption lines, atmospheric surface pressure is an unknown and potentially important quantity when investigating Snowball deglaciation. We used modern surface pressure in our simulations, which is reasonable given that atmospheric oxygen probably had a partial pressure $\sim$0.15 bar lower than today during the global glaciations [*Holland et al.*, 2006; *Canfield et al.*, 2007], and we increased $p\mathrm{CO}_2$, for which self-broadening is 30% stronger than foreign broadening, by only up to 0.1 bar. Additionally, calculations using the *Haqq-Misra et al.* [2008] radiative-convective model suggest that pressure broadening is not a significant effect for O(10%) variations in surface pressure (J. Kasting, personal communication, 2009). However, if current estimates of $p\mathrm{CO}_2$ at the end of Neoproterozoic Snowball events prove to be too low, pressure broadening could be an important mechanism that could drastically increase the radiative forcing associated with increasing $p\mathrm{CO}_2$ and help allow deglaciation.

[42] One objection to the thin-ice Snowball model of *Pollard and Kasting* [2005] is that it might require an ice albedo so low (0.2–0.4) that the tropical ice would simply melt, no matter the $p\mathrm{CO}_2$ [*Warren and Brandt*, 2006]. We have shown that consistent with the energy balance model results of *Pollard and Kasting* [2005] and *Pollard and Kasting* [2006], in the FOAM GCM a Snowball can be stable for certain dust strip widths with an ice surface albedo of 0.2–0.4 and $p\mathrm{CO}_2$ = 0.01–0.1 bar. This implies that if we included dynamic sea ice in our simulations, the lower albedo cases might have thin tropical sea ice until they deglaciated. Since a dirty surface rather than bubbly ice would cause the low albedo in our model, solar radiation might not pass through the thin ice as it did in the model of *Pollard and Kasting* [2005]. Thin ice would, however, be much more likely to develop cracks and leads in which life could survive.

[43] In modern glaciers, dust flux onto the surface often leads to the development of "cryoconite holes" [e.g., *Wharton et al.*, 1985] when the reduced albedo caused by the dust melts some of the surface ice. The resulting effect on albedo is complicated, as during this process some of the dust can be lost into the interior of the ice if the water refreezes on top of it, and the water itself has a significantly lower albedo than surrounding ice. Additionally, the algae that grows in the holes can have a significantly lower albedo than the dust itself [*Takeuchi et al.*, 2001; *Takeuchi*, 2002]. We intend to investigate this process and the effect it could have on the ideas presented in this paper more thoroughly in the future; however, we feel that the model results presented here represent a strong argument that the development of a tropical dust strip during a Snowball Earth event is a plausible mechanism for promoting Snowball deglaciation.

[44] We did not explicitly consider the potential for wind transport of dust in our analysis. Wind could transport dust into continental valleys, ice crevasses, and other protected hollows, or could concentrate dust in loess deposits covering a small percentage of the tropical surface. Furthermore, if global-scale dust transport occurred, as it does on Mars, dust could be blown to higher latitudes and covered with snow quicker than it flows back to lower latitudes with the sea glaciers. To a certain extent we have dealt with this issue by establishing the parameter range of dust strip widths and albedos for which deglaciation is possible. Additionally, we are considering the radiative effects of dust aerosols on Snowball Earth in ongoing work. That said, the adhesiveness of the dust to the ice surface in the tropics under Snowball conditions is an issue that deserves further thought.

7. Conclusion

[45] In this work we have hypothesized that the Snowball hydrological cycle would likely drive glacial flow in such a

way that over the lifetime of the Snowball a dust layer composed of volcanic and continental dust would develop at the surface in the tropics that would significantly lower the tropical albedo and enable Snowball deglaciation, which has heretofore been difficult to produce in global climate models, for pCO_2 = 0.01–0.1 bar.

[46] We predict that if this hypothesis is true, clay drapes found at the top of the 635 Ma glacial deposits should be thicker in tropical than extratropical regions.

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References

Allen, P. A., and J. I. Etienne (2008), Sedimentary challenge to Snowball Earth, *Nat. Geosci.*, *1*, 817–825.
Bao, H., J. R. Lyons, and C. M. Zhou (2008), Triple oxygen isotope evidence for elevated CO_2 levels after a Neoproterozoic glaciation, *Nature*, *453*(7194), 504–506.
Bao, H., I. Fairchild, P. Wynn, and C. Spotl (2009), Stretching the envelope of past surface environments: Neoproterozoic glacial lakes from Svalbard, *Science*, *323*, 112–119.
Bordoni, S., and T. Schneider (2008), Monsoons as eddy-mediated regime transitions of the tropical overturning circulation, *Nat. Geosci.*, *1*(8), 515–519.
Caldeira, K., and J. F. Kasting (1992), Susceptibility of the early Earth to irreversible glaciation caused by carbon-dioxide clouds, *Nature*, *359*(6392), 226–228.
Canfield, D. E., S. W. Poulton, and G. M. Narbonne (2007), Late-Neoproterozoic deep-ocean oxygenation and the rise of animal life, *Science*, *315*(5808), 92–95.
Ceplecha, Z. (1996), Luminous efficiency based on photographic observations of the Lost City fireball and implications for the influx of interplanetary bodies onto Earth, *Astron. Astrophys.*, *311*(1), 329–332.
Donnadieu, Y., F. Fluteau, G. Ramstein, C. Ritz, and J. Besse (2003), Is there a conflict between the Neoproterozoic glacial deposits and the Snowball Earth interpretation: An improved understanding with numerical modeling, *Earth Planet. Sci. Lett.*, *208*(1–2), 101–112.
Evans, D. A. D. (2000), Stratigraphic, geochronological, and paleomagnetic constraints upon the neoproterozoic climatic paradox, *Am. J. Sci.*, *300*(5), 347–433.
Gill, A. E. (1982), *Atmosphere-Ocean Dynamics*, 662 pp., Academic, San Diego, Calif.
Goodman, J. C. (2006), Through thick and thin: Marine and meteoric ice in a ''Snowball Earth'' climate, *Geophys. Res. Lett.*, *33*, L16701, doi:10.1029/2006GL026840.
Goodman, J. C., and R. T. Pierrehumbert (2003), Glacial flow of floating marine ice in ''Snowball Earth'', *J. Geophys. Res.*, *108*(C10), 3308, doi:10.1029/2002JC001471.
Haqq-Misra, J. D., S. D. Domagal-Goldman, P. J. Kasting, and J. F. Kasting (2008), A revised, hazy Methane greenhouse for the Archean Earth, *Astrobiology*, *8*(6), 1127–1137.
Harrison, S. P., K. E. Kohfeld, C. Roelandt, and T. Claquin (2001), The role of dust in climate changes today, at the last glacial maximum and in the future, *Earth Sci. Rev.*, *54*(1–3), 43–80.
Held, I. M., and B. J. Soden (2006), Robust responses of the hydrological cycle to global warming, *J. Clim.*, *19*(21), 5686–5699.
Hoffman, P. F., and Z.-X. Li (2009), A palaeogeographic context for Neoproterozoic glaciation, *Palaeogeogr., Palaeoclimatol., Palaeoecol.*, *277*, 158–172.
Hoffman, P. F., and D. P. Schrag (1999), Considering a Neoproterozoic Snowball Earth: Reply, *Science*, *284*, 1087–1088.
Hoffman, P. F., A. J. Kaufman, G. P. Halverson, and D. P. Schrag (1998), A Neoproterozoic Snowball Earth, *Science*, *281*(5381), 1342–1346.
Holland, M. M., C. M. Bitz, and B. Tremblay (2006), Future abrupt reductions in the summer Arctic sea ice, *Geophys. Res. Lett.*, *33*, L23503, doi:10.1029/2006GL028024.
Hyde, W. T., T. J. Crowley, S. K. Baum, and W. R. Peltier (2000), Neoproterozoic ''Snowball Earth'' simulations with a coupled climate/ice-sheet model, *Nature*, *405*, 425–429.
Jacob, R. (1997), Low frequency variability in a simulated atmosphere ocean system, Ph.D. thesis, Univ. of Wisconsin, Madison.
Kirschvink, J. (1992), Late Proterozoic low-latitude global glaciation: The Snowball Earth, in *The Proterozoic Biosphere: A Multidisciplinary Study*, edited by J. Schopf and C. Klein, pp. 51–52, Cambridge Univ. Press, New York.
Konishchev, V. N. (1982), Characteristics of cryogenic weathering in the permafrost zone of the European USSR, *Arct. Antarct. Alp. Res.*, *14*(3), 261–265.
Krinner, G., O. Boucher, and Y. Balkanski (2006), Ice-free glacial northern Asia due to dust deposition on snow, *Clim. Dyn.*, *27*(6), 613–625.
Le Hir, G., G. Ramstein, Y. Donnadieu, and R. T. Pierrehumbert (2007), Investigating plausible mechanisms to trigger a deglaciation from a hard Snowball Earth, *C. R. Geosci.*, *339*(3–4), 274–287.
Le Hir, G., G. Ramstein, Y. Donnadieu, and Y. Godderis (2008), Scenario for the evolution of atmospheric pCO(2) during a Snowball Earth, *Geology*, *36*(1), 47–50.
Lewis, J. P., A. J. Weaver, and M. Eby (2006), Deglaciating the Snowball Earth: Sensitivity to surface albedo, *Geophys. Res. Lett.*, *33*, L23604, doi:10.1029/2006GL027774.
Mason, B. G., D. M. Pyle, and C. Oppenheimer (2004), The size and frequency of the largest explosive eruptions on Earth, *Bull. Volcanol.*, *66*(8), 735–748.
Peixoto, J. P., and A. H. Oort (1992), *Physics of Climate*, 520 pp., Am. Inst. of Phys., New York.
Pierrehumbert, R. T. (2002), The hydrologic cycle in deep-time climate problems, *Nature*, *419*(6903), 191–198.
Pierrehumbert, R. T. (2004), High levels of atmospheric carbon dioxide necessary for the termination of global glaciation, *Nature*, *429*(6992), 646–649.
Pierrehumbert, R. T. (2005), Climate dynamics of a hard Snowball Earth, *J. Geophys. Res.*, *110*, D01111, doi:10.1029/2004JD005162.
Pollard, D., and J. F. Kasting (2005), Snowball Earth: A thin-ice solution with flowing sea glaciers, *J. Geophys. Res.*, *110*, C07010, doi:10.1029/2004JC002525.
Pollard, D., and J. F. Kasting (2006), Reply to comment by Stephen G. Warren and Richard E. Brandt on ''Snowball Earth: A thin-ice solution with flowing sea glaciers'', *J. Geophys. Res.*, *111*, C09017, doi:10.1029/2006JC003488.
Reijmer, C. H., W. H. Knap, and J. Oerlemans (1999), The surface albedo of the Vatnajokull ice cap, Iceland: A comparison between satellite-derived and ground-based measurements, *Boundary Layer Meteorol.*, *92*(1), 125–144.
Schneider, T., and S. Bordoni (2008), Eddy-mediated regime transitions in the seasonal cycle of a Hadley circulation and implications for monsoon dynamics, *J. Atmos. Sci.*, *65*(3), 915–934.
Takeuchi, N. (2002), Optical characteristics of cryoconite (surface dust) on glaciers: The relationship between light absorbency and the property of organic matter contained in the cryoconite, *Ann. Glaciol.*, *34*, 409–414.
Takeuchi, N., S. Kohshima, and K. Seko (2001), Structure, formation, and darkening process of albedo-reducing material (cryoconite) on a Himalayan glacier: A granular algal mat growing on the glacier, *Arct. Antarct. Alp. Res.*, *33*(2), 115–122.
Trindade, R. I. F., and M. Macouin (2007), Palaeolatitude of glacial deposits and palaeogeography of neoproterozoic ice ages, *C. R. Geosci.*, *339*(3–4), 200–211.
Warren, S. G., and R. E. Brandt (2006), Comment on ''Snowball Earth: A thin-ice solution with flowing sea glaciers'' by David Pollard and James F. Kasting, *J. Geophys. Res.*, *111*, C09016, doi:10.1029/2005JC003411.
Warren, S. G., R. E. Brandt, T. C. Grenfell, and C. P. Mckay (2002), Snowball Earth: Ice thickness on the tropical ocean, *J. Geophys. Res.*, *107*(C10), 3167, doi:10.1029/2001JC001123.
Wharton, R. A., C. P. Mckay, G. M. Simmons, and B. C. Parker (1985), Cryoconite holes on glaciers, *Bioscience*, *35*(8), 499–503.

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