Clouds and Snowball Earth deglaciation

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[1] Neoproterozoic, and possibly Paleoproterozoic, glaciations represent the most extreme climate events in post-Hadean Earth, and may link closely with the evolution of the atmosphere and life. According to the Snowball Earth hypothesis, the entire ocean was covered with ice during these events for a few million years, during which time volcanic CO_2 increased enough to cause deglaciation. Geochemical proxy data and model calculations suggest that the maximum CO_2 was 0.01–0.1 by volume, but early climate modeling suggested that deglaciation was not possible at CO_2 = 0.2. We use results from six different general circulation models (GCMs) to show that clouds could warm a Snowball enough to reduce the CO_2 required for deglaciation by a factor of 10–100. Although more work is required to rigorously validate cloud schemes in Snowball-like conditions, our results suggest that Snowball deglaciation is consistent with observations. **Citation:** Abbot, D. S., A. Voigt, M. Branson, R. T. Pierrehumbert, D. Pollard, G. Le Hir, and D. D. B. Koll (2012), Clouds and Snowball Earth deglaciation, *Geophys. Res. Lett.*, *39*, L20711, doi:10.1029/2012GL052861.

[2] Clouds both reduce infrared emission to space, warming a planet, and reflect solar radiation, cooling a planet. Since high surface albedo reduces the cooling effect of cloud reflection of solar radiation, one expects some cloud warming over snow and ice on a Snowball Earth (see *Kirschvink* [1992] and *Hoffman et al.* [1998] for descriptions of the Snowball Earth hypothesis). Early work using a radiative-convective model outlined the phase space of Snowball cloud behavior and established that clouds could significantly decrease the CO_2 threshold for Snowball deglaciation [*Pierrehumbert*, 2002], but clouds provided little warming in the first general circulation model (FOAM) study of Snowball deglaciation [*Pierrehumbert*, 2004, 2005]. As a result, the model was far from deglaciating at CO_2 = 0.2 by volume, whereas geochemical proxy data and model calculations constrain CO_2 to 0.01–0.1 [*Kasemann et al.*, 2005; *Bao et al.*, 2008, 2009; *Sansjofre et al.*, 2011; *Le Hir et al.*, 2008]. Varied climate and cloud behavior was found in subsequent GCM studies inspired by the pioneering FOAM work [*Le Hir et al.*, 2007; *Abbot and Pierrehumbert*, 2010; *Le Hir et al.*, 2010; *Hu et al.*, 2011; *Pierrehumbert et al.*, 2011], although the many discrepancies among simulations made it difficult to unambiguously identify the reasons for the differences in climate. Here we run simulations using a series of GCMs containing sophisticated cloud schemes with consistent boundary conditions, which allows us to constrain the region of cloud phase space appropriate for Snowball deglaciation.

[3] Our GCM suite includes SP-CAM, which contains a two-dimensional cloud resolving scheme within each gridbox, and is therefore the most sophisticated cloud scheme ever used to address Snowball Earth climate. Furthermore, we run CAM, LMDz, and ECHAM, which all contain modern cloud fraction and prognostic cloud condensate parameterizations. We run all models with a uniform surface albedo of 0.6, zero aerosols, zero ozone, all greenhouse gases other than CO_2 and H_2O set to zero, an obliquity of 23.5°, an eccentricity of 0°, and a solar constant of 1285 W m^{-2} (auxiliary material).[1] We set the land surface to "glacial ice," like Greenland and Antarctica in modern simulations, everywhere.

[4] The tropical (20°S to 20°N) and annual mean surface temperature (TS) in FOAM is 7–11 K colder than the other models at CO_2 = 0.1 (Figure 1). TS increases 16–20 K in the models when CO_2 is increased from 10^{-4} to 0.1, implying a climate sensitivity of 1.6–2.0 K per doubling of CO_2 (we get similar values for global mean temperature). Consequently, CO_2 would have to be increased by a factor of ≈10–100 in FOAM for it to be as warm as the other models. We found differences of only a few W m^{-2} when we compared clear-sky outgoing longwave radiation outputted by each model with a single offline radiation scheme forced by the model's zonal mean temperature and humidity (auxiliary material). This is small compared with the ≈40 W m^{-2} associated with increasing CO_2 from 10^{-4} to 0.1, which indicates that differences in clear-sky radiation schemes among models are not strong enough to drive the surface temperature differences. Given that other variables and boundary conditions are uniform among models, clouds are the main potential driver of intermodel temperature differences.

[5] The cloud radiative forcing (CRF) varies widely among the models in Snowball conditions (Figure 1). Differences in longwave CRF drive this variation at CO_2 = 10^{-4}, whereas at CO_2 = 0.1 the shortwave CRF becomes large enough in some models to contribute to the variation, despite the high surface albedo (Table S2 in Text S1).

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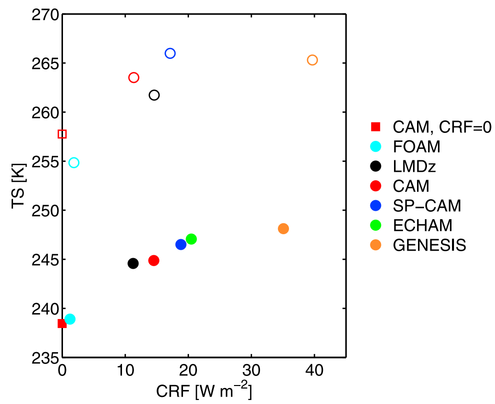

Figure 1. Scatter plot of the relation between tropical and annual mean surface temperature (TS) and top-of-atmosphere cloud radiative forcing (CRF) for the models. Simulations in CAM with CRF artificially set to zero are also included. Filled symbols represent $CO_2 = 10^{-4}$ and empty symbols represent $CO_2 = 0.1$.

Consistent with the idea that clouds drive intermodel temperature differences is the correlation of CRF with TS (Figure 1). However, higher temperatures lead to higher radiative fluxes, which inflate CRF, and could lead to spurious correlation between CRF and TS. We artificially set the CRF to zero in CAM by setting clouds to zero inside both the longwave and shortwave radiation schemes and repeating the simulations. CAM with CRF = 0 produces surface temperatures that are similar to those produced by FOAM (Figure 1). This test demonstrates that the dominant cause of cold temperatures in FOAM is low CRF, and confirms that differences in CRF drive differences in TS.

[6] Cloud fraction differences among models are not clearly related to differences in CRF (Figures S2 and S3 in Text S1). To understand how models can produce non-negligible CRF values in the Snowball (Figure 1), we must consider cloud condensate (cloud water and ice concentration). Cloud condensate, even at relatively low concentrations and elevations, can strongly increase CRF and warm the Snowball [*Pierrehumbert*, 2002, 2004, 2005]. Below 600 mb, the tropical vertical profile of cloud condensate is similar in shape to the modern for all models [*Su et al.*, 2011]. The amount of cloud condensate is clearly related to the resulting CRF, although ECHAM and GENESIS produce a higher CRF relative to the other models than would be expected from their cloud condensate level. This suggests differences in the models' parameterizations of cloud radiative properties. No model produces the distinct maximum in cloud condensate at ≈300 mb observed in the modern climate. This may reflect the lack of truly deep convection in the Snowball, although this deep maximum is not well-produced by CAM even in modern climate simulations [*Su et al.*, 2011].

[7] Cloud condensate is much lower in FOAM than the other models (Figures 2 and S4). The FOAM cloud condensate scheme is borrowed from CCM3, the ancestral version of CAM, and specifies cloud condensate as a simple exponential decay with a scale height diagnosed from total column water vapor [*Hack*, 1998], rather than calculating it prognostically. In cold, dry Snowball conditions this scale height is very small, so the total column cloud condensate (proportional to the scale height) is small. Furthermore, a small scale height concentrates cloud condensate near the surface, where its warming longwave radiative effect is minimal. Our comparison shows that the FOAM cloud condensate values are an order of magnitude lower than those in other models, even with consistent boundary conditions.

[8] As can be seen in SP-CAM, which is illustrative of the other models, vigorous atmospheric circulation is ultimately responsible for low-level cloud condensate approaching modern values (Figure 3). The tropical mean climate shown in Figures 1 and 2 is the average of an annual cycle with large seasonal excursions of the region of atmospheric ascent due to low surface heat capacity [*Pierrehumbert*, 2005; *Abbot and Pierrehumbert*, 2010; *Pierrehumbert et al.*, 2011]. The Snowball Hadley cell is four times stronger than the modern at $CO_2 = 10^{-4}$ (not shown) and six times stronger than modern at $CO_2 = 0.1$ (Figure 3; D.S. Abbot et al. (manuscript in preparation, 2012) will focus on atmospheric circulation in these simulations). Cloud condensate and fraction are mainly associated with low-level convection in the ascent region of the strong Hadley cell. The CRF is near zero in the winter hemisphere, where there is a strong inversion, but rises to ≈40 W m^{-2} in the summer hemisphere (Figure 3). This is similar to the modern longwave CRF over ocean, despite the lack of high-level cloud condensate (Figure 2). As a result of the warming provided by clouds, the surface temperature in the summer subtropics rises above the melting point (Figure 3), and melting occurs.

[9] SP-CAM is the model with the least parameterized cloud scheme. The SP-CAM cloud scheme explicitly resolves cloud processes in each grid box on a two-dimensional grid with parameterization of cloud microphysics. The cloud radiative forcing produced by SP-CAM is broadly similar to that of CAM, LMDz, and ECHAM, which each have a prognostic cloud condensate scheme. The cloud condensate scheme in FOAM causes an up to ≈6 K cold bias in cold

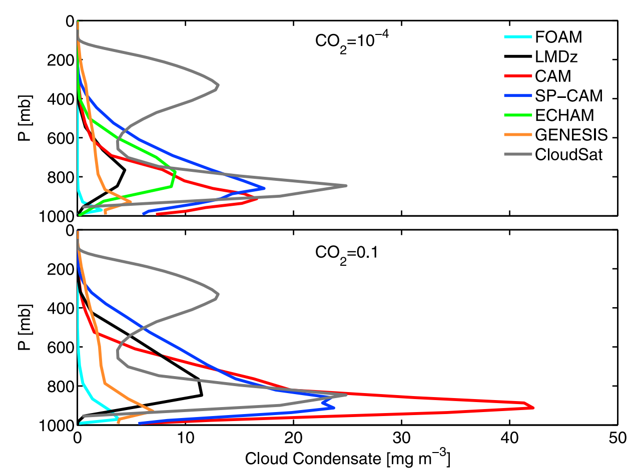

Figure 2. Vertical profiles of tropical and annual mean cloud condensate for the models. For reference, CloudSat observations of the modern climate averaged over the tropical ocean from August 2006 to July 2010 are provided [*Su et al.*, 2011]. Low-level cloud condensate in CAM and SP-CAM is approximately as large as modern, and FOAM's cloud condensate is about an order of magnitude lower.

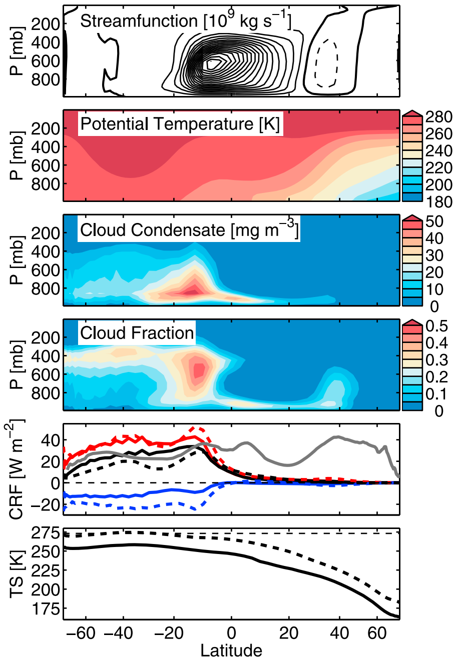

Figure 3. Climate of SP-CAM in January. For Eulerian mass streamfunction, clockwise circulation is depicted by thin solid lines, counterclockwise circulation is depicted by thin dashed lines, the zero contour is thick and solid, and the contour spacing is 50×10^9 kg s^{-1}. Contour plots are for CO_2 = 0.1 only, but show similar patterns at $CO_2 = 10^{-4}$. Longwave (red), shortwave (blue), and total (black) cloud radiative forcing (CRF) are plotted for $CO_2 = 10^{-4}$ (solid) and 0.1 (dashed). For reference, zonal mean CERES [*Loeb et al.*, 2008] satellite measurements of longwave CRF over ocean in the modern climate are plotted in gray. Surface temperature (TS) is plotted for $CO_2 = 10^{-4}$ (solid) and 0.1 (dashed), and the melting point is plotted as a thin dashed line.

regions of the modern climate [*Rasch and Kristjansson*, 1998], which is consistent with it underestimating Snowball cloud warming. It therefore is reasonable to consider a CRF like that in SP-CAM, CAM, LMDz, and ECHAM (Figure 1) as most likely for a Snowball Earth climate, although it is possible that microphysical effects not included in these models could limit the Snowball CRF.

[10] Although we do not explicitly simulate Snowball deglaciation here, our work suggests clouds would make it significantly easier than previously thought, allowing consistency with CO_2 estimates from geochemical proxy data. Given that volcanic and continental dust could also significantly lower albedo [*Schatten and Endal*, 1982; *Abbot and Pierrehumbert*, 2010; *Le Hir et al.*, 2010; *Abbot and Halevy*, 2010], deglaciation no longer appears to pose a serious problem for the Snowball Earth hypothesis. Thin-ice [*Pollard and Kasting*, 2005] or waterbelt [*Hyde et al.*, 2000; *Chandler and Sohl*, 2000; *Peltier et al.*, 2007; *Abbot et al.*, 2011; *Yang et al.*, 2012a, 2012b] models for Neoproterozoic glaciations, could also deglaciate at low enough CO_2. More fundamental work on cloud behavior in cold, Snowball-like conditions is needed to confirm the results described here. Such work is also critical for understanding climate change on modern Earth in the sensitive polar regions.

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