

Clouds and Snowball Earth deglaciation

Dorian S. Abbot,¹ Aiko Voigt,² Mark Branson,³ Raymond T. Pierrehumbert,¹ David Pollard,⁴ Guillaume Le Hir,⁵ and Daniel D. B. Koll¹

Received 22 June 2012; revised 19 September 2012; accepted 20 September 2012; published 25 October 2012.

[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 CO₂ increased enough to cause deglaciation. Geochemical proxy data and model calculations suggest that the maximum CO₂ was 0.01–0.1 by volume, but early climate modeling suggested that deglaciation was not possible at CO₂ = 0.2. We use results from six different general circulation models (GCMs) to show that clouds could warm a Snowball enough to reduce the CO₂ 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 CO₂ 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 CO₂ = 0.2 by volume, whereas geochemical proxy data and model calculations constrain CO₂ 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 grid-box, 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 CO₂ and H₂O 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 CO₂ = 0.1 (Figure 1). TS increases 16–20 K in the models when CO₂ is increased from 10^{−4} to 0.1, implying a climate sensitivity of 1.6–2.0 K per doubling of CO₂ (we get similar values for global mean temperature). Consequently, CO₂ 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 CO₂ 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 CO₂ = 10^{−4}, whereas at CO₂ = 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).

¹Department of Geophysical Sciences, University of Chicago, Chicago, Illinois, USA.

²Max Planck Institute for Meteorology, Hamburg, Germany.

³Department of Atmospheric Science, Colorado State University, Fort Collins, Colorado, USA.

⁴Earth and Environmental Systems Institute, College of Earth and Mineral Sciences, Pennsylvania State University, University Park, Pennsylvania, USA.

⁵Institut de Physique du Globe de Paris, Université Paris 7-Denis Diderot, Paris, France.

Corresponding author: D. S. Abbot, Department of Geophysical Sciences, University of Chicago, 5734 South Ellis Ave., Chicago, IL 60637, USA. (abbot@uchicago.edu)

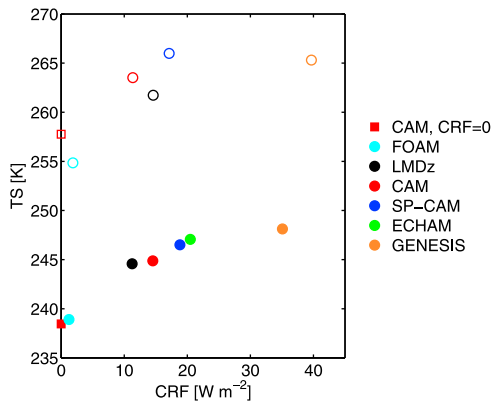


Figure 1. Scatter plot of the relation between tropical and annual mean surface temperature (TS) and top-of-atmosphere cloud radiative forcing (CRF) for the models. Simulations in CAM with CRF artificially set to zero are also included. Filled symbols represent $\text{CO}_2 = 10^{-4}$ and empty symbols represent $\text{CO}_2 = 0.1$.

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 $\text{CRF} = 0$ 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 $\text{CO}_2 = 10^{-4}$ (not shown) and six times stronger than modern at $\text{CO}_2 = 0.1$ (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 $\approx 40 \text{ W m}^{-2}$ 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 $\approx 6 \text{ K}$ cold bias in cold

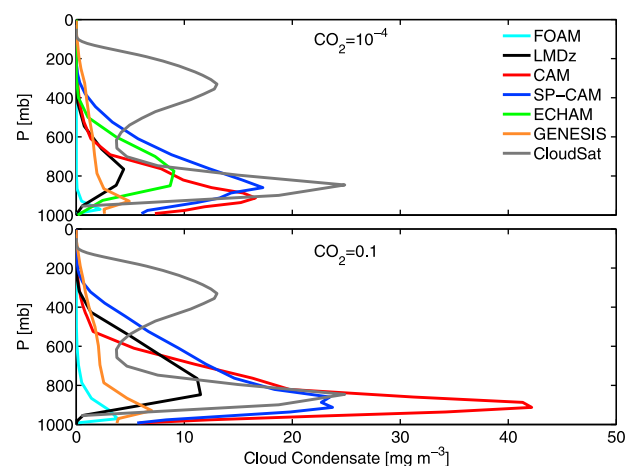


Figure 2. Vertical profiles of tropical and annual mean cloud condensate for the models. For reference, CloudSat observations of the modern climate averaged over the tropical ocean from August 2006 to July 2010 are provided [Su et al., 2011]. Low-level cloud condensate in CAM and SP-CAM is approximately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

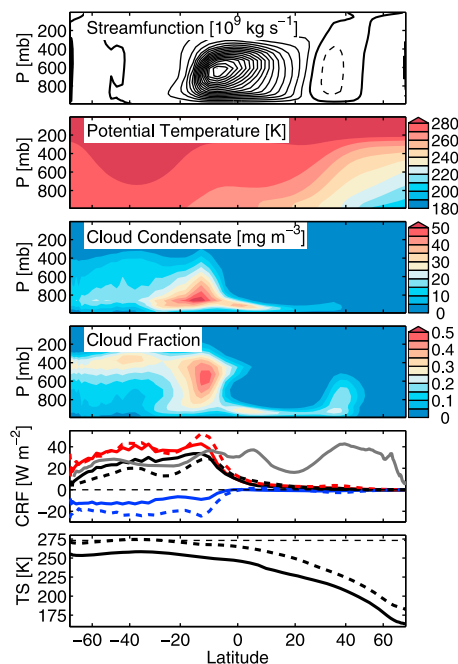


Figure 3. Climate of SP-CAM in January. For Eulerian mass streamfunction, clockwise circulation is depicted by thin solid lines, counterclockwise circulation is depicted by thin dashed lines, the zero contour is thick and solid, and the contour spacing is $50 \times 10^9 \text{ kg s}^{-1}$. Contour plots are for $\text{CO}_2 = 0.1$ only, but show similar patterns at $\text{CO}_2 = 10^{-4}$. Longwave (red), shortwave (blue), and total (black) cloud radiative forcing (CRF) are plotted for $\text{CO}_2 = 10^{-4}$ (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 $\text{CO}_2 = 10^{-4}$ (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.

[11] **Acknowledgments.** We thank Hui Su for providing CloudSat data. We thank Eli Tziperman and Colin Goldblatt for thoughtful reviews. This work was partially supported by NSF DMS-0940261, which is part of the NSF Math Climate Research Network. We acknowledge the German Research Foundation (DFG) program for the initiation and intensification of international collaboration. RTP was supported NSF ATM-0933936. This work was partially supported by the Max Planck Society for the Advancement of Science. The ECHAM6 simulation was performed at the German Climate Computing Center (DKRZ) in Hamburg, Germany. The authors thank GENCI (Grand Equipement National de Calcul Intensif) and the CEA (Commissariat à l'Energie Atomique) for providing the computer power necessary for the LMDz simulation.

[12] The Editor thanks Colin Goldblatt and an anonymous reviewer for their assistance in evaluating this paper.

References

- Abbot, D. S., and I. Halevy (2010), Dust aerosol important for Snowball Earth deglaciation, *J. Clim.*, **23**(15), 4121–4132, doi:10.1175/2010JCLI3378.1.
- Abbot, D. S., and R. T. Pierrehumbert (2010), Mudball: Surface dust and Snowball Earth deglaciation, *J. Geophys. Res.*, **115**, D03104, doi:10.1029/2009JD012007.
- Abbot, D. S., A. Voigt, and D. Koll (2011), The Jormungand global climate state and implications for Neoproterozoic glaciations, *J. Geophys. Res.*, **116**, D18103, doi:10.1029/2011JD015927.
- Bao, H., I. Fairchild, P. Wynn, and Spotl (2009), Stretching the envelope of past surface environments: Neoproterozoic glacial lakes from Svalbard, *Science*, **323**(5910), 119–122, doi:10.1126/science.1165373.
- Bao, H. M., 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, doi:10.1038/nature06959.
- Chandler, M. A., and L. E. Sohl (2000), Climate forcings and the initiation of low-latitude ice sheets during the Neoproterozoic Varanger glacial interval, *J. Geophys. Res.*, **105**(D16), 20,737–20,756.
- Hack, J. J. (1998), Sensitivity of the simulated climate to a diagnostic formulation for cloud liquid water, *J. Clim.*, **11**(7), 1497–1515.
- Hoffman, P. F., A. J. Kaufman, G. P. Halverson, and D. P. Schrag (1998), A Neoproterozoic Snowball Earth, *Science*, **281**(5381), 1342–1346.
- Hu, Y., J. Yang, F. Ding, and W. R. Peltier (2011), Model-dependence of the CO_2 threshold for melting the hard Snowball Earth, *Clim. Past*, **7**(1), 17–25, doi:10.5194/cp-7-17-2011.
- 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**(6785), 425–429.
- Kasemann, S. A., C. J. Hawkesworth, A. R. Prave, A. E. Fallick, and P. N. Pearson (2005), Boron and calcium isotope composition in Neoproterozoic carbonate rocks from Namibia: Evidence for extreme environmental change, *Earth Planet. Sci. Lett.*, **231**(1–2), 73–86, doi:10.1016/j.epsl.2004.12.006.
- 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.
- 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, doi:10.1016/j.crte.2006.09.002.
- 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.
- Le Hir, G., Y. Donnadieu, G. Krinner, and G. Ramstein (2010), Toward the Snowball Earth deglaciation, *Clim. Dyn.*, **35**(2–3), 285–297, doi:10.1007/s00382-010-0748-8.
- Loeb, N. G., B. A. Wielicki, D. R. Doelling, G. L. Smith, D. F. Keyes, S. Kato, N. Manalo-Smith, and T. Wong (2008), Toward optimal closure of the Earth’s top-of-atmosphere radiation budget, *J. Clim.*, **22**(3), 748–766, doi:10.1175/2008JCLI2637.1.
- Peltier, W. R., Y. G. Liu, and J. W. Crowley (2007), Snowball Earth prevention by dissolved organic carbon remineralization, *Nature*, **450**(7171), 813–818.
- 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, doi:10.1038/nature02640.
- Pierrehumbert, R. T. (2005), Climate dynamics of a hard Snowball Earth, *J. Geophys. Res.*, **110**, D01111, doi:10.1029/2004JD005162.
- Pierrehumbert, R. T., D. S. Abbot, A. Voigt, and D. Koll (2011), Climate of the Neoproterozoic, *Annu. Rev. Earth Planet. Sci.*, **39**, 417–460, doi:10.1146/annurev-earth-040809-152447.

- 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.
- Rasch, P. J., and J. E. Kristjansson (1998), A comparison of the CCM3 model climate using diagnosed and predicted condensate parameterizations, *J. Clim.*, *11*(7), 1587–1614.
- Sansjofre, P., M. Ader, R. I. F. Trindade, M. Elie, J. Lyons, P. Cartigny, and A. C. R. Nogueira (2011), A carbon isotope challenge to the Snowball Earth, *Nature*, *478*(7367), 93–96, doi:10.1038/nature10499.
- Schatten, K. H., and A. S. Endal (1982), The faint young Sun-climate paradox: Volcanic influences, *Geophys. Res. Lett.*, *9*(12), 1309–1311.
- Su, H., J. H. Jiang, J. Teixeira, A. Gettelman, X. Huang, G. Stephens, D. Vane, and V. S. Perun (2011), Comparison of regime-sorted tropical cloud profiles observed by CloudSat with GEOS5 analyses and two general circulation model simulations, *J. Geophys. Res.*, *116*, D09104, doi:10.1029/2010JD014971.
- Yang, J., W. Peltier, and Y. Hu (2012a), The initiation of modern “soft Snowball” and “hard Snowball” climates in CCSM3. Part I: The influence of solar luminosity, CO₂ concentration and the sea-ice/snow albedo parameterization, *J. Clim.*, *25*, 2711–2736, doi:10.1175/JCLI-D-11-00189.1.
- Yang, J., W. Peltier, and Y. Hu (2012b), The initiation of modern “soft Snowball” and “hard Snowball” climates in CCSM3. Part II: Climate dynamic feedbacks, *J. Clim.*, *25*, 2737–2754, doi:10.1175/JCLI-D-11-00190.1.