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Present-day and ice-covered equilibrium states in a comprehensive climate model

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[1] We show that in a comprehensive climate model both the current climate and a completely ice-covered Earth are stable states under today's total solar irradiance (TSI) and CO_2 level. We employ the Max Planck Institute for Meteorology coupled atmosphere-ocean general circulation model ECHAM5/MPI-OM, at relatively high resolution (horizontally T63 in the atmosphere and 1.5 degrees in the ocean). Setting TSI to near-zero causes a transition from realistic present-day climate to a completely ice-covered state within 15 years; this state persists even when TSI re-assumes today's value. A break-up of the complete ice cover occurs with today's TSI and 100 times – but not with 10 times – today's atmospheric CO_2 level. While TSI is near-zero, extremely strong meridional overturning ensues in both the Atlantic and the Pacific Oceans. Our results imply that a snowball Earth is possible, in principle, with inception possibly triggered by a brief dark spell. **Citation:** Marotzke, J., and M. Botzet (2007), Present-day and ice-covered equilibrium states in a comprehensive climate model, *Geophys. Res. Lett.*, *34*, L16704, doi:10.1029/2006GL028880.

1. Introduction

[2] Earth's climate might exist in two dramatically different equilibria, either completely ice-covered or nearly ice-free. This was suggested by simple climate models long ago [*Budyko*, 1969; *Sellers*, 1969; *North*, 1975], but it has been unclear whether comprehensive climate models show the same bi-stability [*Langen and Alexeev*, 2004]. As one consequence, attempts to reconstruct the transitions into and out of the "snowball Earth" [*Harland*, 1964; *Kirschvink*, 1992; *Hoffman et al.*, 1998], postulated to have existed during the Neoproterozoic, have been only partially successful [*Poulsen et al.*, 2001; *Donnadieu et al.*, 2004a, 2004b; *Pierrehumbert*, 2004; *Poulsen and Jacob*, 2004]. Demonstrating that an ice-covered Earth is a climate equilibrium is the basic dynamical requirement for a self-consistent snowball Earth scenario; further requirements are convincing mechanisms for snowball inception and demise. Previous work has found ice-covered equilibrium states in palaeo-geographical configurations most appropriate for the snowball Earth, using models of intermediate complexity or atmosphere-ocean mixed layer models [*Donnadieu et al.*, 2004a, 2004b; *Pierrehumbert*, 2004; *Poulsen and Jacob*, 2004]. With palaeo-geographical configurations, TSI and CO_2 appropriate for the Neoproterozoic, dynamic ocean processes prevented a snowball solution in a low-resolution coupled GCM [*Poulsen et al.*, 2001; *Poulsen and Jacob*, 2004]. In contrast to previous work, we apply here a state-of-the-art coupled climate model, the Max Planck Institute for Meteorology atmosphere-ocean general circulation model (AOGCM) ECHAM5/MPI-OM [*Marsland et al.*, 2003; *Roeckner et al.*, 2003], which has been extensively and successfully evaluated against observations [*Hagemann et al.*, 2004; *Jungclaus et al.*, 2006; *Roeckner et al.*, 2006; *Wild and Roeckner*, 2006]. We use the model in the exact configuration that was applied for the scenario runs [*Brasseur and Roeckner*, 2005; *Bengtsson et al.*, 2006; *Landerer et al.*, 2007], submitted to the fourth Assessment Report (AR4) by the Intergovernmental Panel on Climate Change (IPCC). In particular, the model has the relatively high resolution (T63, 31 levels in the atmosphere; nominally 1.5 degrees, 40 levels in the ocean) that makes possible its realistic control climate.

[3] We address here, for the first time with a comprehensive climate model, the fundamental question of whether Earth's climate can exist in both the current and the fully ice-covered states, with the same parameters, especially total solar irradiance (TSI) and atmospheric CO_2 level. By using a fully coupled model, we guarantee internal consistency between atmospheric and oceanic transport processes even under drastic climate changes. Our experimental strategy – we look for bistability in our comprehensive model, where one equilibrium is the current climate – precludes, as one consequence, an attempt to model snowball Earth inception directly. In particular, we cannot offer a realistic pathway into a snowball Earth; for that, palaeo-continental configurations, TSI and CO_2 level would all need to be adjusted for the Neoproterozoic. Instead, we want to provide a reference point with the most sophisticated model available to us. If this model permits a completely ice-covered solution, a clear target is defined for extensive future sensitivity studies, especially with computationally less demanding models.

2. Model Results

[4] Starting from current climate conditions (experiment CTRL, TSI = 1365 Wm^{-2}; 1990 CO_2 levels of 348 parts per million volume, ppmv), we abruptly reduce TSI to 0.1 per mil of its realistic value (run DARKNESS). This extreme strategy all but guarantees that a completely frozen state is indeed attained [*Poulsen et al.*, 2001; *Bendtsen and Bjerrum*, 2002; *Poulsen and Jacob*, 2004], thus providing the bracketing case of the climate response to the most extreme sudden cooling event imaginable (less extreme ones being a giant volcano eruption or an asteroid impact).

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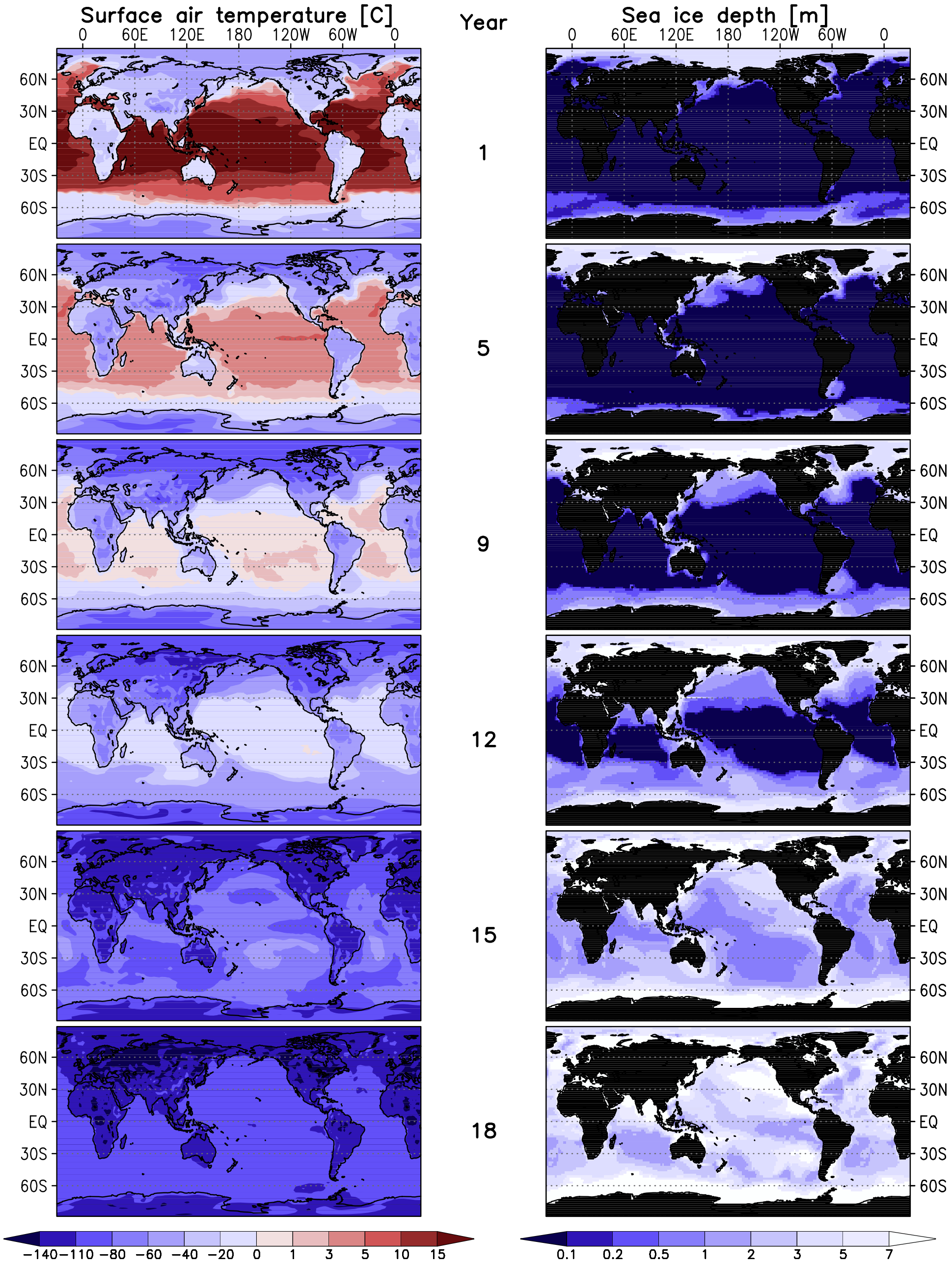

Figure 1. (left) Annual mean surface air temperature and (right) sea ice depth for experiment DARKNESS, for (top to bottom) years 1, 5, 9, 12, 15, and 18. In the right column, the darkest blue indicates absence of sea ice.

Moreover, we can study the transients towards a cold climate.

[5] Cooling first sets in over land (Figure 1, left); within about 5 years, annual-mean surface air temperatures over land plummet to −20°C and below practically everywhere, despite still-open ocean areas. A few months after the "blocking" of the sun, sea ice cover has nearly doubled (Figure 2a). For another ca. five years, however, total ice cover does not change much, presumably because vigorous convective mixing (see below) keeps transporting heat upward in the ocean and thus inhibits sea ice formation. Then, the sea ice edge rapidly progresses equatorward, first in the Northern Hemisphere, but from year 9 on also in the Southern Hemisphere (Figure 1, right). The low latitude ocean, covering nearly half the ocean area, largely remains at temperatures above freezing until about year 10 but then rapidly and nearly uniformly freezes over (Figures 1, 2a, and 2b). The ocean is completely ice-covered by year 15, consistent with the minimum time predicted by a simple coupled model [*Bendtsen and Bjerrum*, 2002]. After ice cover is complete, land surface temperatures fall even more, to below −100°C over almost all areas, in the annual mean (Figure 1). Snow accumulation practically ceases because the frozen-over ocean is no longer a moisture source.

[6] Sea ice thickness and total sea ice volume keep growing even after ice cover is complete (Figures 1, 2a, and 2c). After year 19, we turn the sun back on again, for one year with 50% of current TSI, then with 100% (exper-

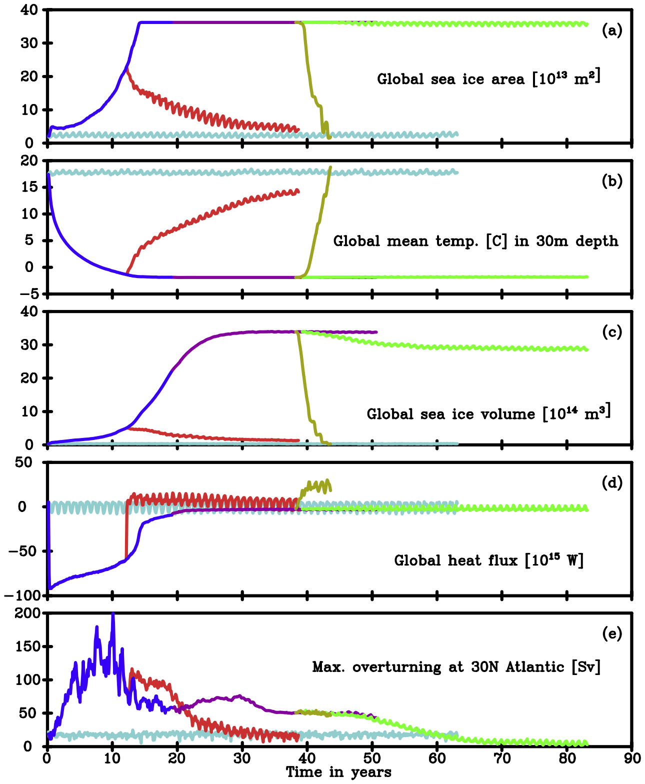

Figure 2. Time series of (a) global sea ice area, (b) global mean ocean temperature at 30 m depth, representing SST, (c) global sea ice volume, (d) globally integrated ocean heat gain, and (e) strength of the Atlantic MOC at 30°N. Cyan: CTRL; blue: DARKNESS (TSI = 0.1365 Wm^{-2}); magenta: LIGHT (TSI = 682.5 Wm^{-2} for the first year, 1365 Wm^{-2} thereafter); red: as DARKNESS but with TSI = 1365 Wm^{-2}); ochre: as LIGHT but with pCO_2 = 34,800 ppmv; green: as LIGHT but with pCO_2 = 3,480 ppmv.

iment LIGHT). Sea ice cover remains complete, however, and global mean sea surface temperature (SST) remains at freezing (Figures 2a and 2b). Total ice volume increases until it reaches roughly 34 × 10^{14} m^3 by about year 30; the value remains stable thereafter (Figure 2c). Right after the start of DARKNESS, the ocean loses heat at a rate of nearly 100 PW (petawatts, 1 PW ≡ 10^{15} W, Figure 2d). At year 50, ocean heat loss has settled to about 2 PW (Figure 2d) or, expressed per unit area of ocean, to 5.5 Wm^{-2}. Within the model's limitations (see below), this characterizes a new equilibrium. The CTRL and LIGHT experiments thus constitute two stable states of our climate model, with partial and complete ice cover, respectively. Just as predicted by the early simple models [*Budyko*, 1969; *Sellers*, 1969; *North*, 1975], the albedo of the ice-covered Earth is so high (global mean 0.78, ocean mean 0.85, almost the maximum sea ice albedo, and land mean 0.61) that too much solar energy is reflected to space, and ice cannot melt.

[7] Twelve years into DARKNESS, we begin an additional simulation and set TSI = 100%, to obtain a feel for the required length of complete darkness before the transition from the current climate to an ice-covered Earth becomes unstoppable. Although at year 12 the surface ocean is at near-freezing everywhere and covered with ice for 2/3 of its area, the remaining open areas absorb sufficient solar energy to lead to a near-complete deglaciation in 25 years.

[8] That a period of complete darkness of only about 15 years is sufficient to cause unstoppable global glaciation, in the sense that setting TSI back to today's value does not lead to deglaciation, may appear surprising but is consistent with a simple coupled climate model [*Bendtsen and Bjerrum*, 2002]. Indeed, it only takes 10 years to cool the entire ocean to below 0°C (Figure 3). Hence, once the surface has frozen over, there is very little heat content left. The ocean loses nearly all heat so rapidly because the cooling from the top, everywhere, leads to ubiquitous convective mixing, which is a very efficient vertical transport process. The heat brought up from the deep is then

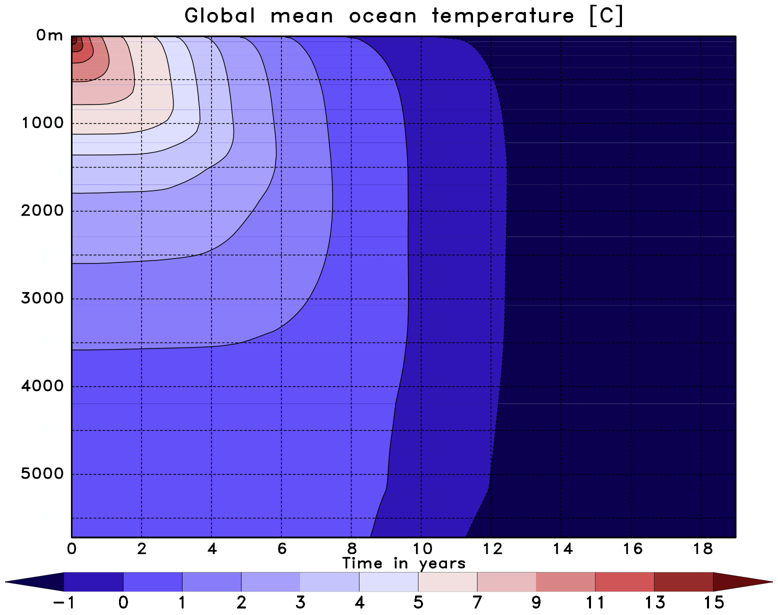

Figure 3. Time-depth plot of horizontally averaged ocean temperature during experiment DARKNESS.

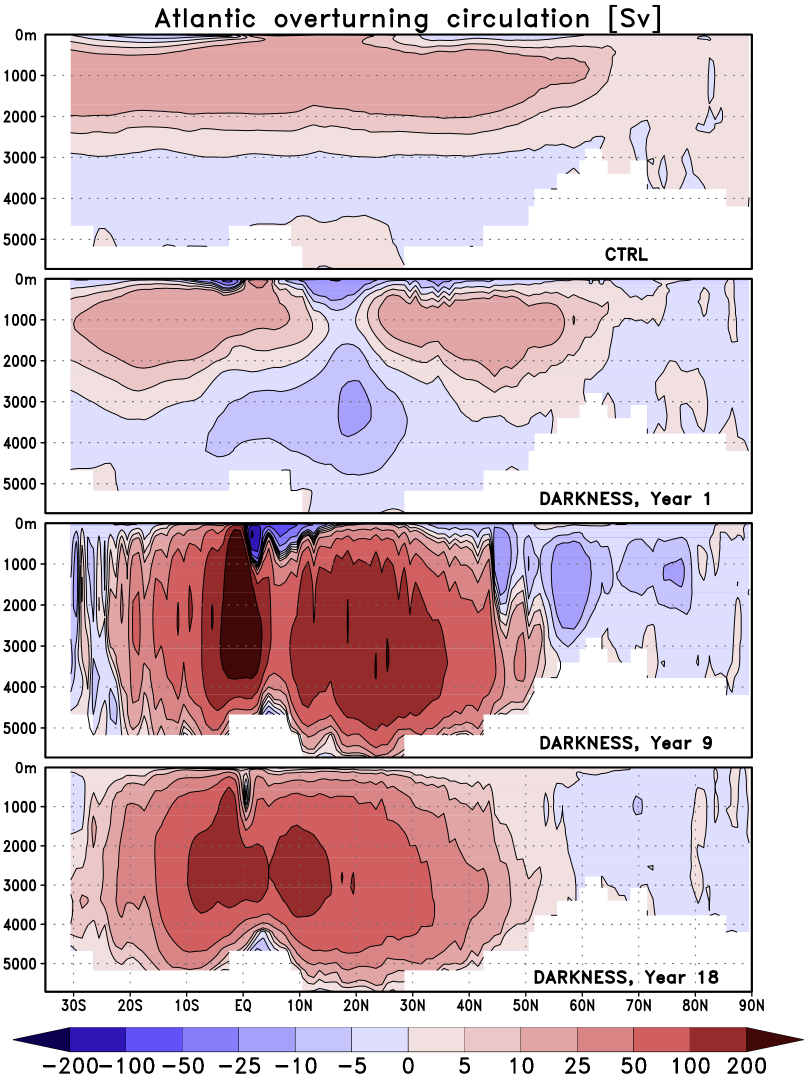

Figure 4. Annual mean meridional overturning stream function, which is defined in a latitude-depth plane, for CTRL, DARKNESS at years (top to bottom) 1, 9, and 18. Positive values indicate clockwise circulation, negative values indicate counterclockwise circulation.

rapidly transferred to the atmosphere by air-sea interaction and finally radiated to space.

[9] The ubiquitous convective mixing causes another surprising aspect of the transition from CTRL to DARKNESS, as shown by the time history of the ocean's meridional overturning circulation (MOC, Figure 4). One might have anticipated a gradual spin-down, owing to the reduction in overall SST gradients under ice. Instead, the convective mixing everywhere causes an extremely strong MOC because horizontal density gradients now occur over the entire depth range, not only in the upper ocean. The Atlantic MOC, which in CTRL has strength of around 20 Sv (Sverdrups, 1 Sv $\equiv 10^6$ m^3 s^{-1}), attains maximum values of around 200 Sv, with interannual fluctuations of several 10s of Sv (Figures 2e and 4). The Pacific also shows much-increased MOC (not shown). These "flushes" were first discovered in relatively simple ocean models [*Marotzke*, 1989; *Weaver and Sarachik*, 1991; *Winton and Sarachik*, 1993] and were hitherto not thought to play a role in comprehensive models. After 30 years into LIGHT, the Atlantic MOC still is around 50 Sv but shows signs of a gradual weakening. We conclude that a cooling event, even a less extreme one than considered here, would lead to significantly enhanced deep ocean circulation. Notice, however, that the ocean heat transport associated with the flush event increases to a much lesser extent than the MOC increase (figure not shown), because vertical temperature difference are very small after year 10 (Figure 3).

[10] A crucial question is how an ice-covered Earth manages the transition back to partial or no ice cover. In previous work, even pCO$_2$ levels of around 150,000 ppmv (0.2 bar) were insufficient to provide a strong enough greenhouse effect to melt the ice, albeit at TSI weaker by 6% compared to the present [*Pierrehumbert*, 2004, 2005]. In our model, we find that 100 times the 1990 pCO$_2$ level (34,800 ppmv) reduces ice cover to below current values within 5 years (Figure 2a). Notice that the rapidity of the deglaciation has no geophysical significance concerning the time it took to melt the hypothetical snowball Earth, which would have had sea ice thicknesses of hundreds of meters.

The rapidity merely indicates that the model has been placed considerably beyond the bifurcation point separating ice-covered and ice-free states. To obtain some idea of where that bifurcation point lies, we start from experiment LIGHT and set pCO_2 to an intermediate value of 10 times the 1990 pCO_2 (3,480 ppmv). The stronger greenhouse effect only leads to small seasonal ice retreat and some reduction in total ice volume. This reduction, however, levels off after about 50 years and does not constitute a departure from the ice-covered state (Figure 2c).

[11] The sea ice component of our coupled model follows the dynamics of *Hibler* [1979] and the very simple thermodynamics of the so-called ‘‘zero-layer model’’ of [*Semtner*, 1976]. Snow cover on sea ice is represented explicitly. Notice that the known major weakness of the zero-layer model, its exaggeration of the seasonal cycle [*Semtner*, 1984], is of minor import here as seasonality plays no role. In our implementation, sea ice thickness is limited to about 8 m, 75% of the thickness of the top ocean model layer, to avoid the problem of dry model levels. While the limitation on sea ice thickness clearly limits the study of the sea ice dynamics in the ice-covered state, its thermodynamic implications are those of an implied heat source at the bottom of the ice, equal to the latent heat of freezing of the sea ice thickening that should but does not occur. A heat source works against maintenance of the total sea ice cover, after the TSI is set back to 100%, and especially in the experiment with 10 times the present pCO_2 level, but even in that case total ice cover is maintained in our simulations. On the other hand, the artificial heat source works in favor of deglaciation, so there is additional uncertainty in our results. However, with CO_2 levels 100 times of the present, the extreme additional greenhouse effect (around 25 Wm^{-2} in the first month of that segment of the simulation) by far outweighs the artificial heating implied at the sea ice base. Thus, we do not expect any change in our conclusions from a more sophisticated model that can handle arbitrary sea ice thicknesses. That said, it needs to be investigated why our model behaves differently from that of *Pierrehumbert* [2004, 2005], at roughly similar radiative forcing. Notice that radiative forcing from CO_2 is approximately logarithmic with concentration, so the relevant characterization of the difference between the two radiative forcings is that *Pierrehumbert* [2004, 2005] employed two further doublings of CO_2, compared to our run with 100 times CO_2. These two doublings approximately make up for his 6% lower TSI, which at the high albedo of the ice-covered state corresponds to a reduction in radiative forcing of 5–6 Wm^{-2}. One possible explanation for the difference in model behavior is the very different distributions of continents in the two models.

3. Conclusions and Outlook

[12] The existence of the fully ice-covered equilibrium in our model demonstrates that a snowball Earth, fundamentally, would be consistent with climate dynamics as represented in a comprehensive AOGCM. Moreover, our modelled Earth freezes over within 15 years, indicating that snowball inception could arise from a brief dark spell, induced perhaps by a giant volcano eruption or an asteroid impact [*Pope et al.*, 1994; *Bendtsen and Bjerrum*, 2002]. Snowball demise could result, later, from very high but not impossible CO_2 levels [*Kirschvink*, 1992]. Before a volcano eruption or an asteroid impact can be confirmed as a valid scenario for snowball inception, however, the effect of dust deposition on snow or sea ice must be investigated with a coupled aerosol-climate model. Dust cover might reduce the albedo to below what is required for the ice-covered equilibrium. But a dark spell of just a few years, even if unable to trigger complete long-term glaciation, would still cause dramatic changes. Among the most surprising is the several-fold strengthening in deep ocean circulation simulated here, with hitherto unexplored consequences.

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References

Bendtsen, J., and C. J. Bjerrum (2002), Vulnerability of climate on Earth to sudden changes in insolation, *Geophys. Res. Lett.*, *29*(15), 1706, doi:10.1029/2002GL014829.

Bengtsson, L., K. I. Hodges, and E. Roeckner (2006), Storm tracks and climate change, *J. Clim.*, *19*, 3518–3543.

Brasseur, G. P., and E. Roeckner (2005), Impact of improved air quality on the future evolution of climate, *Geophys. Res. Lett.*, *32*, L23704, doi:10.1029/2005GL023902.

Budyko, M. I. (1969), The effect of solar radiation variations on the climate of the Earth, *Tellus*, *21*, 611–619.

Donnadieu, Y., Y. Godderis, G. Ramstein, A. Nedelec, and J. Meert (2004a), A ‘snowball Earth’ climate triggered by continental break-up through changes in runoff, *Nature*, *428*, 303–306.

Donnadieu, Y., G. Ramstein, F. Fluteau, D. Roche, and A. Ganopolski (2004b), The impact of atmospheric and oceanic heat transports on the sea-ice-albedo instability during the Neoproterozoic, *Clim. Dyn.*, *22*, 293–306.

Hagemann, S., B. Machenhauer, R. Jones, O. B. Christensen, M. Deque, D. Jacob, and P. L. Vidale (2004), Evaluation of water and energy budgets in regional climate models applied over Europe, *Clim. Dyn.*, *23*, 547–567.

Harland, W. B. (1964), Critical evidence for a great infra-Cambrian glaciation, *Geol. Rundsch.*, *54*, 45–61.

Hibler, W. D. (1979), A dynamic thermodynamic sea ice model, *J. Phys. Oceanogr.*, *9*, 815–846.

Hoffman, P. F., A. J. Kaufman, G. P. Halverson, and D. P. Schrag (1998), A Neoproterozoic snowball Earth, *Science*, *281*, 1342–1346.

Jungclaus, J. H., M. Botzet, H. Haak, N. Keenlyside, J.-J. Luo, M. Latif, J. Marotzke, U. Mikolajewicz, and E. Roeckner (2006), Ocean circulation and tropical variability in the coupled model ECHAM5/MPI-OM, *J. Clim.*, *19*, 3952–3972.

Kirschvink, J. L. (1992), Late Proterozoic low-latitude global glaciation: The snowball Earth, in *The Proterozoic Biosphere*, edited by J. W. Schopf and C. Klein, pp. 51–52, Cambridge Univ. Press, New York.

Landerer, F. W., J. H. Jungclaus, and J. Marotzke (2007), Regional dynamic and steric sea level change in response to the IPCC-A1B scenario, *J. Phys. Oceanogr.*, *37*, 296–312.

Langen, P. L., and V. A. Alexeev (2004), Multiple equilibria and asymmetric climates in the CCM3 coupled to an oceanic mixed layer with thermodynamic sea ice, *Geophys. Res. Lett.*, *31*, L04201, doi:10.1029/2003GL019039.

Marotzke, J. (1989), Instabilities and multiple steady states of the thermohaline circulation, in *Oceanic Circulation Models: Combining Data and Dynamics*, edited by D. L. T. Anderson and J. Willebrand, pp. 501–511, Kluwer, Dordrecht, Netherlands.

Marsland, S. J., H. Haak, J. H. Jungclaus, M. Latif, and F. Röske (2003), The Max-Planck-Institute global ocean/sea ice model with orthogonal curvilinear coordinates, *Ocean Modell.*, *5*, 91–127.

North, G. R. (1975), Theory of energy-balance climate models, *J. Atmos. Sci.*, *32*, 2033–2043.

Pierrehumbert, R. T. (2004), High levels of atmospheric carbon dioxide necessary for the termination of global glaciation, *Nature*, *429*, 646–649.

Pierrehumbert, R. T. (2005), Climate dynamics of a hard snowball Earth, *J. Geophys. Res.*, *110*, D01111, doi:10.1029/2004JD005162.

Pope, K. O., K. H. Baines, A. C. Ocampo, and B. A. Ivanov (1994), Impact winter and the Cretaceous-Tertiary extinctions—Results of a Chicxulub asteroid impact model, *Earth Planet. Sci. Lett.*, *128*, 719–725.

Poulsen, C. J., and R. L. Jacob (2004), Factors that inhibit snowball Earth simulation, *Paleoceanography*, *19*, PA4021, doi:10.1029/2004PA001056.

Poulsen, C. J., R. T. Pierrehumbert, and R. L. Jacob (2001), Impact of ocean dynamics on the simulation of the Neoproterozoic “snowball Earth”, *Geophys. Res. Lett.*, *28*, 1575–1578.

Roeckner, E., et al. (2003), The atmospheric general circulation model ECHAM5. part I: Model description, 127 pp., Max Planck Inst. for Meteorol., Hamburg, Germany.

Roeckner, E., R. Brokopf, M. Esch, M. Giorgetta, S. Hagemann, L. Kornblueh, E. Manzini, U. Schlese, and U. Schulzweida (2006), Sensitivity of simulated climate to horizontal and vertical resolution in the ECHAM5 atmosphere model, *J. Clim.*, *19*, 3771–3791.

Sellers, W. D. (1969), A global climate model based on the energy balance of the Earth-atmosphere system, *J. Appl. Meteorol.*, *8*, 392–400.

Semtner, A. J. (1976), A model for the thermodynamic growth of sea ice in numerical investigations of climate, *J. Phys. Oceanogr.*, *6*, 379–389.

Semtner, A. J. (1984), On modeling the seasonal thermodynamic cycle of sea ice in studies of climatic-change, *Clim. Change*, *6*, 27–37.

Weaver, A. J., and E. S. Sarachik (1991), The role of mixed boundary-conditions in numerical models of the ocean’s climate, *J. Phys. Oceanogr.*, *21*, 1470–1493.

Wild, M., and E. Roeckner (2006), Radiative fluxes in ECHAM5, *J. Clim.*, *19*, 3792–3809.

Winton, M., and E. S. Sarachik (1993), Thermohaline oscillations induced by strong steady salinity forcing of ocean general-circulation models, *J. Phys. Oceanogr.*, *23*, 1389–1410.

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