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CO_2 levels required for deglaciation of a "Near-Snowball" Earth

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Abstract. Geologic evidence suggests that in the Late Neoproterozoic (~600 Ma) almost all land masses were glaciated, with sea-level glaciation existing even at the equator. A recent modeling study has shown that it is possible to simulate an ice-covered Earth glaciation with a coupled climate/ice-sheet model. However, separate general circulation model experiments suggest that a second solution may exist with a substantial area of ice free ocean in the tropics. Although 0.1 to 0.3 of an atmosphere of CO_2 (~300 to 1000 X) is required for deglaciation of a "Snowball Earth," the "exit" CO_2 levels for an open water solution could be significantly less. In this paper we utilize a coupled climate/ice sheet model to demonstrate four points: (1) the open water solution can be simulated in the coupled model if the sea ice parameter is adjusted slightly; (2) a major reduction in ice volume from the open water/equatorial ice solution occurs at a CO_2 level of about 4X present values - about two orders of magnitude less than required for exit from the "hard" snowball initial state; (3) additional CO_2 increases are required to get fuller meltback of the ice; and (4) the open water solution exhibits hysteresis properties, such that climates with the same level of CO_2 may evolve into either the snowball, open water, or a warmer world solution, with the trajectory depending on initial conditions. These results set useful targets for geochemical calculations of CO_2 changes associated with the open-water solution.

Introduction

The possibility of a late Neoproterozoic (~800-600 Ma, million years ago) "Snowball Earth" has stimulated a great deal of interest. Ice sheets apparently reached the equator at this time [e.g, *Hambrey and Harland,* 1985; *Eyles,* 1993], and the first appearance of multi-celled animals (metazoans) may have preceded [e.g, *Hofmann et al.,* 1990; *Bromham et al.,* 1998] the last phase of the glaciation (~600 Ma). Although early modeling work on this problem failed to simulate equatorial glaciation, recent studies with a coupled climate/ice sheet model [*Hyde et al.,* 2000] indicate that a Snowball Earth solution can be parsimoniously obtained if solar luminosity is reduced by the astrophysically reasonable value of 6% and CO_2 levels are near present values. *Hyde et al.* [2000] further demonstrated that the existence of the ice-covered Earth solution is dependent on inclusion of the ice sheet, because the elevation and thermal inertia of the ice-sheet allows it to "weather" Milankovitch interglacial orbital configurations. *Hoffman et al.* [1988] invoked estimates by *Caldiera and Kasting* [1992] that massive levels of atmospheric CO_2 (~0.1 bar) were required to escape from this state, and such values are comparable to those calculated by *Hyde et al.* [2000].

A further result of the *Hyde et al.* [2000] study is that stand-alone general circulation model (GCM) experiments with fixed ice sheets in some cases yielded open water conditions in equatorial regions removed from the ice sheet [cf. *Baum and Crowley,* 2000]. The existence of open water solutions reflects standard length-scale arguments discussed in the climate literature [*North,* 1984; *Manabe and Broccoli,* 1985], in which the thermal influence of an ice sheet (or other perturbation) diminishes by a scale which is determined from characteristic radiative damping and heat transport values for the Earth. GCM sensitivity experiments [*Baum and Crowley,* 2000] indicate that this open water solution is obtained for CO_2 values of about 2.5X present levels. This study also demonstrated that a strong negative cloud feedback plays a role in offsetting the enhanced albedo from expanded ice cover, as the decreased air temperatures are associated with lower water vapor levels [cf. *Jenkins,* 1993; *Chandler and Sohl,* 2000].

The existence of the open water solution has important implications for climate theory and the survival of metazoans, and it would provide a valuable constraint on geochemical models of the late Neoproterozoic. The open water may also lead to significantly lower deglaciation CO_2 threshold levels than are necessary for a "hard" snowball solution. In this paper we use the same coupled climate/ice sheet model as that in *Hyde et al.* [2000] to conduct a range of sensitivity experiments to determine whether open water solutions can be obtained by such coupled models and, if so, the CO_2 levels required for deglaciation from such a state.

Climate Model and Boundary Conditions

The climate/ice sheet model [*Deblonde and Peltier,* 1991; *Deblonde et al.,* 1991; *Tarasov and Peltier,* 1997] consists of four submodels, which respectively predict ice flow, mass balance, temperature, and bedrock sinking. The latter is assumed to occur with a time constant of 4,000 years [*Peltier,* 1998]. Ice flows subject to a temperature-independent rheology based on the *Nye* [1959] formulation, and mass balance is computed according to statistical models [*Reeh,* 1990; *Huybrechts and T'Siobbel,* 1995], which take as input monthly temperatures generated from a nonlinear, two-dimensional diffusive seasonal EBM [*Hyde et al.,* 1990]. The EBM has been validated against many different GCMs [*Crowley et al.,* 1991], while the coupled ice sheet model reproduces many features of both the last glacial cycle and the

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Carboniferous and Neoproterozoic glaciations [*Tarasov and Peltier*, 1999; *Hyde et al.*, 1999, 2000].

Baseline paleogeography for the Neoproterozoic is from *Dalziel* [1997], although sensitivity experiments [*Hyde et al.*, 2000] indicate that the principal results are not strongly dependent on details of geography. We examined a range of elevations for the continents ranging from sea level to 500 m. In the extremely cold climate we study in this paper our equilibrium results are relatively insensitive to this parameter. The model is driven with Milankovitch insolation variations appropriate to the Pleistocene [see for example *Hyde et. al* 1989]. Other boundary conditions include a solar luminosity 6% below present [*Crowley and Baum*, 1993], CO_2 levels varying from ~0.5-10X present levels and a variable sea ice albedo. The sea ice albedo of the EBM in Pleistocene simulations [*Deblonde and Peltier*, 1991] is 0.5. This value is lower than employed in GCMs and reflects the fact that the EBM specification is for the total surface-atmosphere column. Due to atmospheric reflectance and absorption, the column albedo is about 10% lower than the surface [see *Hyde et al.*, 1989]. Because cloudiness decreases in our Neoproterozoic GCM runs [*Baum and Crowley*, 2000], we mimic this effect on absorbed solar radiation by lowering albedo to 0.45 over sea ice. The net changes in radiation receipt from the decreased albedo (~25 W/m^2) are still somewhat less than the changes in surface radiation receipt from a 20% decrease in cloudiness [*Baum and Crowley*, 2000], but given all the uncertainties in the albedo calculations, we consider it a reasonable adjustment for the model. Further justification for the sea ice albedo reduction is based on the supposition that the albedo should be lower at low zenith angles in the tropics [for example, the albedo of water varies from 0.02 to 0.06 as the zenith angle changes from 0 to 60 degrees, *Peixoto and Oort*, 1992, p. 103].

Precipitation is highly parametertized; we employ a spatially uniform value of 0.6 mm/day, which declines with height and temperature. This value is characteristic of present

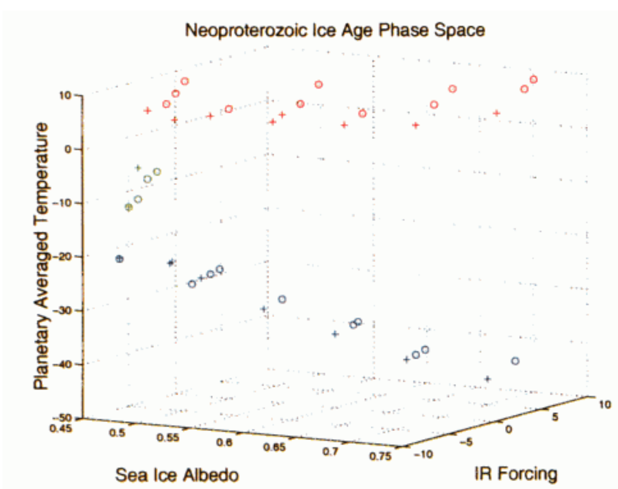

Figure 1. "Phase-space" plot of global sea-level temperature vs. sea ice albedo and infrared radiative forcing perturbations (a proxy for CO_2). Different colors represent the different basic states of the model - "incomplete" glaciation (red), fully ice-covered planet (blue), and ice-covered continents with open water (green). Circles represent experiments in which average continental elevation is 400 m, crosses experiments with elevations at 200 m.

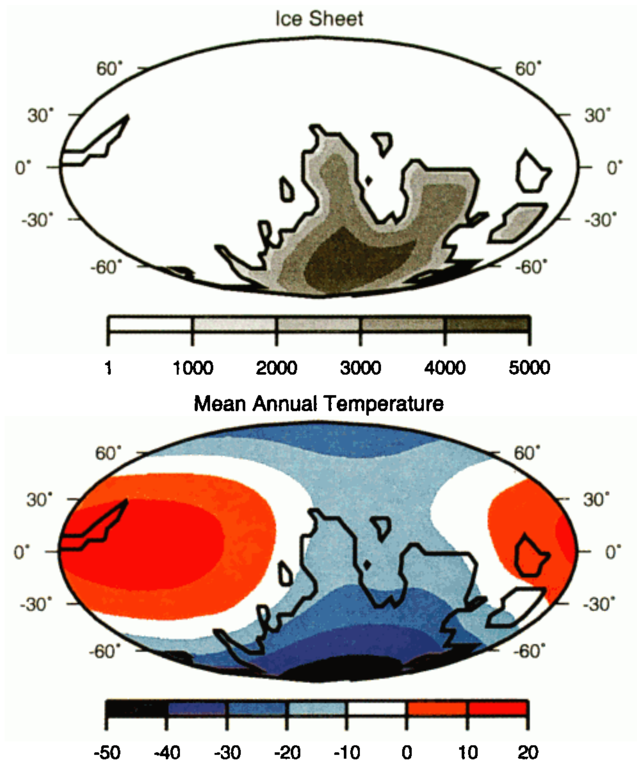

Figure 2. Example of model open water solution: (a) distribution of ice sheet (thickness in m); (b) mean annual temperature.

northern hemisphere mid/high-latitute land areas. Previous work [*Hyde et al.*, 1999] indicates only a modest sensitivity to variations in this term in very cold climates, and for the work presented here we consider the choice satisfactory for an initial examination of the "open water" deglaciation problem.

The numerical experiments were conducted in two ways. In the first class of experiments the model was initialized as ice free, and run until equilibrium under prescribed input parameters. Most such runs came to equilibrium after 100,000 years of integration, though some were extended to 200,000 years to determine the stability of the resulting climate against Milankovitch forcing. A second set of runs was started at the end of a 200,000 year simulation that yielded a stable open water solution. In this set of experiments we linearly increased the CO_2 concentration from years 220,000 to 270,000 and ran the model to equilibrium with the new forcing, generally until about year 350,000. In this way we are able to test the stability of the open water solution to changes in forcing.

Results

A summary of the principal results (Figure 1) illustrates mean global temperature as a function of both sea ice albedo and CO_2 radiative forcing perturbations. With a sea ice albedo of 0.5 or greater, there are two main solutions, one with a totally ice-covered planet (blue symbols), the other with glaciation extending only to 25-30° paleolatitude (red). However, as sea ice albedo is lowered slightly, a third solution emerges (green) that has ice sheets on most of the continents, including ice at the equator, but also a substantial

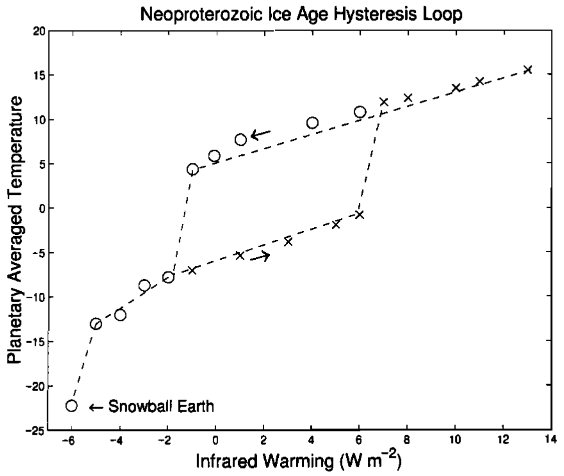

Figure 3. Hysteresis plot of snowball earth runs. Planetary average temperature decreases with infrared forcing. The upper branch represents a climate state without equatorial ice. The full snowball earth solution is shown in the lower right corner. The open water solution follows the lower branch of the hysteresis plot (equivalent to about 0.7-3.0X present CO_2 levels), with deglacation commencing at ~3.5 - 4X CO_2. Circles refer to runs starting from an ice-free initial state, X's runs starting from the open water initial state.

area of open water (Figure 2) in very much the same area as simulated by GCM studies [*Baum and Crowley*, 2000].

Simulations starting from the open water initial state indicate clear hysteresis properties (Figure 3), wherein the model tends to remain in the full ice/open water solution at higher CO_2 levels than would occur if the model run was started from a no-ice initial state. The range of radiative forcing values for the hysteresis effect is equivalent to about 0.7-3.0X the present CO_2 level in the atmosphere. At 3.5-4.0X present levels extensive deglaciation commences and ice volume decreases about an order of magnitude. The duration of the deglaciation (20,000 - 40,000 years) is not nearly as abrupt (Figure 4) as for the full-snowball response (2,000 years - note that the latter did not occur until CO_2 levels were about two orders of magnitude higher than in the present runs). A near-instantaneous CO_2 increase yields a similar time scale for the deglaciation. Even after major deglaciation ice volumes (Figure 4) are greater than Pleistocene ice sheets (about 35 X 10^6 km^3). A progressive increase in radiative forcing to levels as high as 10X CO_2 results in a steady decrease in ice volume, with glacial-interglacial fluctuations due to Milankovitch forcing (as the ice volume decreases this forcing becomes more important).

Discussion and Conclusions

To summarize, a new set of coupled EBM/ice sheet model simulations with variable sea ice albedo provide additional support for a possible open water solution in a Neoproterozoic ice-covered continent scenario. Because of uncertainties in Neoproterozoic continental reconstructions, the particular land masses we simulate as ice free may have been in different locations than specified. If locations of the real landmasses could be more confidently determined, such data would enable a crucial test of our open water solution to be performed. Our results suggest that the most likely area for ice-free land to occur is on isolated continents at some distance from the main ice sheet mass.

If these results can be verified by further work, they have significant implications for understanding Neoproterozoic glaciation and carbon cycling. The CO_2 levels required to obtain open water deglaciation are about two orders of magnitude less than would be required for a "hard" Snowball Earth solution. This result suggests that the planet remained in a near-snowball state for a significantly shorter time than hypothesized in *Hoffman et al.* [1998]. Although it would only require about 100,000 years for degassing to raise atmospheric CO_2 levels to this value, the reactive ocean carbonate reservoir must first be neutralized. This might take on the order of 0.5 million years [*L. Derry*, personal communication, 2000]. As ice sheet calving would still be providing ground carbonate to the ocean, this estimate would be extended by some indeterminate amount. The estimated total time for atmospheric CO_2 buildup is in the lower half of the range of uncertainty for the duration of the Varanger glaciation based on paleomagnetic data [*Sohl et al.*, 1999] and geochemical calculations [*Jacobsen and Kauffman*, 1999]. The required additional rise in CO_2 after equatorial deglaciation might be explainable in terms of CO_2 released from carbonate precipitation in Neoproterozoic postglacial caprocks. The simulated glacial-interglacial fluctuations during this latter time span are consistent with evidence from Neoproterozoic shelf deposits [*Kennedy and Christie-Blick*, 1998].

If the hysteresis response can be verified it may present an additional interesting target for geochemical model studies to determine whether there are any scenarios that might lead to a "CO_2 attractor" in the ~0.7-3.0X CO_2 range, thereby preventing the Earth from reaching the hard snowball state, with attendant implications for the survival of metazoans and CO_2 levels required for deglaciation.

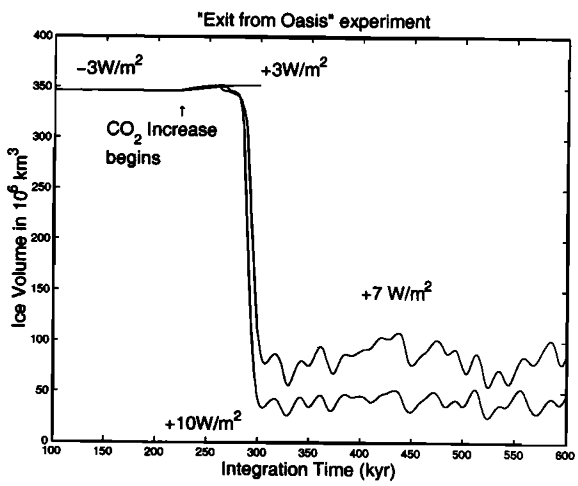

Figure 4. Ice volumes for different levels of CO_2 values associated with the warming of the open water solution. More than 20,000 - 40,000 model years are required to get deglaciation with a ramp increase in CO_2 lasting 50,000 years. The CO_2 changes equivalent to the radiative forcing perturbations are approximately ~1.7X (3 W/m^2), ~3.5X (~7 W/m^2), ~6X (10 W/m^2), and ~10X (12 W/m^2). The first 100,000 years of the model run has been truncated because it represents spinup time.

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