Neoproterozoic 'snowball Earth' simulations with a coupled climate/ice-sheet model

William T. Hyde*, Thomas J. Crowley*, Steven K. Baum* & W. Richard Peltier†

* *Department of Oceanography, Texas A&M University, College Station, Texas 77843-3146, USA*
† *Department of Physics, University of Toronto, Toronto, Ontario, M5S 1A7, Canada*

Ice sheets may have reached the Equator in the late Proterozoic era (600–800 Myr ago), according to geological and palaeomagnetic studies, possibly resulting in a 'snowball Earth'. But this period was a critical time in the evolution of multicellular animals, posing the question of how early life survived under such environmental stress. Here we present computer simulations of this unusual climate stage with a coupled climate/ice-sheet model. To simulate a snowball Earth, we use only a reduction in the solar constant compared to present-day conditions and we keep atmospheric CO_2 concentrations near present levels. We find rapid transitions into and out of full glaciation that are consistent with the geological evidence. When we combine these results with a general circulation model, some of the simulations result in an equatorial belt of open water that may have provided a refugium for multicellular animals.

Some of the most dramatic events in the Earth's history occurred at the end of the Proterozoic era (this era was about 1,000–540 Myr ago). This epoch was characterized by formation of the supercontinent of Rodinia from about 1,000 to 800 Myr ago, the later breakup of this landmass and eventual reassembly into a different configuration by ~550 Myr ago, during the final phase of the pan-African orogeny[1,2]. There were also major changes in strontium, sulphur and carbon isotopes[3], together with the most extensive glaciation of the past billion years. At least two main phases of ice advance occurred, with glaciers apparently extending to the Equator at sea level[4–8]. The first phase from ~760 to 700 Myr is generally termed the Sturtian ice age, while the second, which occurred in the interval ~620–580 Myr ago, is often termed the Varanger (or Marinoan) ice age.

The late Proterozoic also marked the first appearance of metazoans (multicelled animals), perhaps as early as 700–1,000 Myr ago[9–12], while the Varanger ice age was almost immediately followed by the time interval (Vendian) featuring the first fossil remains of multicelled animals—the Ediacaran fauna. If metazoans are pre-Varanger, extensive glaciation may have exerted a significant stress on biota during a critical interval in their evolution.

Recent work[4] has focused attention on the Neoproterozoic by interpreting new carbon-isotope data to indicate that biological productivity of the oceans virtually ceased for perhaps millions of years during the glacial era. These authors concluded from this and other evidence that the planet entered a snowball Earth state, in which it was completely covered by ice until CO_2 outgassing produced a sufficiently large greenhouse effect to melt the ice. In this scenario the sudden warming caused a rapid precipitation of calcium carbonate, producing the cap carbonate rocks often observed in strata of this era. The repetition of such formations suggests that this sequence of events occurred at least twice in the Neoproterozoic.

Although this provocative hypothesis has stimulated considerable interest, it remains to be shown that the physical climate system can respond in the proposed manner. For example, although the concepts of the snowball Earth 'attractor', and rapid exits therefrom, are a feature of energy balance models (EBMs)[13], it is not at all clear to what degree the slow dynamics of ice-sheet physics would modify this model. Previous simulations with EBMs[14] and general circulation models (GCMs)[15,16] indicated that plausible changes in solar luminosity, the Earth's rotation rate and ocean heat transport were

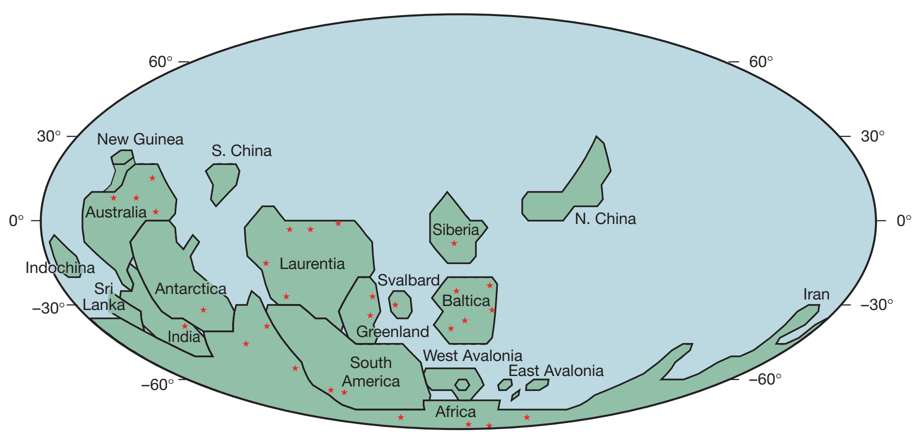

Figure 1 Late Precambrian geography. The figure shows the input geography[2] (Mollweide projection) used in our late Precambrian simulation. Red asterisks indicate the location of Neoproterozoic glacial deposits[7].

insufficient to produce a snowball Earth unless CO_2 is lowered to less than half the present value. Simulations with a radically increased planetary obliquity—which has been suggested as an explanation for equatorial glaciation[17]—preferentially produce snow in low but not high latitudes[18], and their seasonally extremely warm polar areas are in direct conflict with evidence for glaciation in these regions (see, for example, Fig. 1).

None of the above simulations included an ice-sheet model, and few evaluated the effects of orbital insolation variations or used 'realistic' geography. Here we demonstrate that a coupled EBM/ice-sheet model using realistic boundary conditions can successfully predict land ice at the Equator. We further demonstrate that a stand-alone general circulation model experiment with ice-covered continents predicts open water in equatorial regions, thereby creating a refugium for metazoans in this extreme climate.

Ice-sheet model and boundary conditions

The climate/ice-sheet model that we use was introduced by Deblonde and Peltier[19,20]; here, we will use the version described fully by Tarasov and Peltier[21]. Its physical basis consists of four submodels, which predict ice flow, mass balance, temperature and bedrock sinking. In brief, ice is assumed to flow subject to a temperature-independent rheology based on the Nye[22] formulation. Precipitation/ablation is computed according to statistical models[23,24] which themselves take as input monthly temperatures from a nonlinear, two-dimensional (latitude–longitude) diffusive seasonal EBM[25]. Bedrock sinking is assumed to occur with a time constant of 4,000 years. The EBM has been validated against many different GCMs[26], while the coupled EBM/ice-sheet model reproduces many features of climate change over the past 120,000 years and predicts ice for about 80% of the known glacial deposits for the radically different geography of the Carboniferous ice age[27]. The flexibility of the model lends itself to extensive sensitivity testing and long runs; for example, here we have run approximately 20 million years of model simulations.

Model inputs include palaeogeography, atmosphere CO_2 concentrations, precipitation, Milankovitch forcing, and the solar constant. Our baseline palaeogeography for the ~590-Myr Varanger glaciation is taken from Dalziel's reconstruction for 545 Myr ago[2]. Although data suggest that at 590 Myr ago amalgamation was not as great as it was at 545 Myr, the latitudes of important glacial sites in Namibia and Australia are close to their Varanger positions. However, interior seaways existed before the final joining of Africa and South America, and Baltica was closer to Laurentia before the opening of the Iapetus Ocean. We will later discuss a number of experiments that examine the sensitivity of our conclusions to uncertainties in palaeogeography. For convenience we initially assume that the continents are at sea level; adding freeboard is nearly equivalent to reducing the solar constant. We employ a solar constant appropriate to the era in question, that is, about 6% below present[14]. Precipitation and CO_2 values are varied to test the snowball Earth hypothesis. In our initial simulations, ice sheets were driven by a uniform precipitation of 0.6 mm d^{-1}, which declines with height and decreasing temperature. This value is characteristic of current Northern Hemisphere mid-latitude land areas. Given the important role which the tropics play in our simulations, a lower value cannot be justified for our (initially ice-free) Neoproterozoic simulations. The model is relatively insensitive to higher precipitation rates[27]. The model is driven with Milankovitch variations appropriate to the Pleistocene epoch[28], although the very cold climates we investigate are relatively uninfluenced by this forcing. The model makes predictions of temperature, sea ice, and ice volume that can be tested for consistency against the snowball Earth hypothesis and other geological evidence.

Neoproterozoic simulations

Our first experiment, with the above inputs and present-day atmospheric CO_2 concentration resulted in a large ice sheet (not shown) that in coastal areas approaches 40° latitude, but which is restricted by summer warming on the supercontinent to about 50° S in the interior. Sensitivity experiments with the alternative reconstruction of Bond *et al.*[29] yield similar results, but the smaller amount of land area in middle to high latitudes leads to ice growth to 30° palaeolatitude (not shown).

Neoproterozoic CO_2 levels are not known, but may have been very variable due to large carbon reservoir shifts as indicated by $\delta^{13}C$ changes[3,4,30]. We therefore conducted a series of sensitivity tests in which CO_2 levels were varied by adding or subtracting infrared forcing. For CO_2 levels approximately half that of the Holocene epoch (that is, ~130 p.p.m. or a cooling of roughly 5 W m^{-2}) the entire land surface is glaciated (Fig. 2), while for higher levels (slightly over twice the Holocene value) the ice volume approaches that of the Last Glacial Maximum. The transition from a cold state to a fully snow-covered one (snowball Earth) is abrupt (~2,000 years; Fig. 3), as conjectured by Hoffman *et al.*[4] and consistent with other EBM[31], GCM[32] and EBM/ice-sheet studies[27]. It occurs at a CO_2 level equivalent to an infrared cooling of 4.75 to 5.0 W m^{-2}. The exact location of this discontinuity is not particularly important, given the known uncertainty in solar constant and—for example—the different land albedo of a world without vascular plants. The

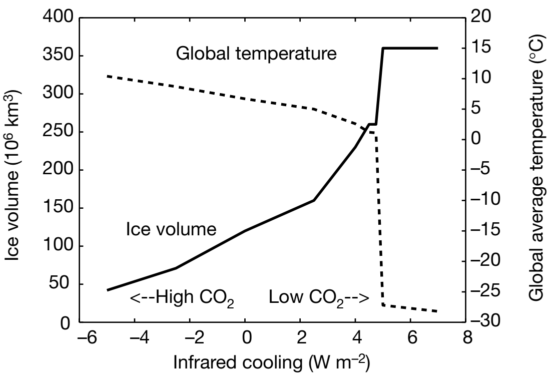

Figure 2 Operating curve for the climate/ice-sheet model. Ice volume and global sea level temperature are shown as a function of CO_2 level and of infrared forcing. A bifurcation occurs at a cooling of ~5 W m^{-2}, beyond which all land surfaces are ice-covered and global temperature drops considerably: the snowball Earth.

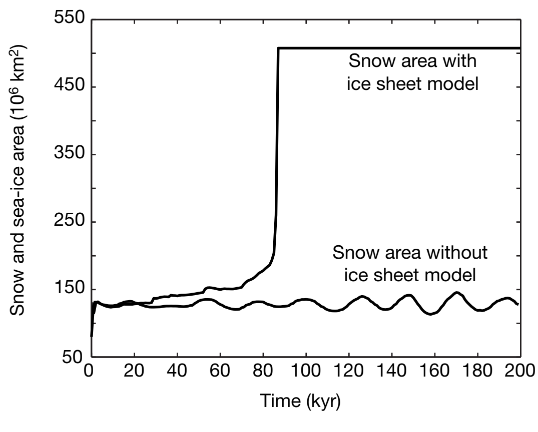

Figure 3 The effect of ice dynamics on model results for an experiment with −5 W m^{-2} of infrared forcing. Without the ice sheet, the climate model predicts a climate in which only about one-quarter of the planetary surface is covered in snow or sea ice. If the ice sheets are allowed to grow, however, they flow into areas originally too warm for year-round snow, but the ice-sheet's thermal inertia and elevation prevent it from melting.

transition is associated with an abrupt drop in globally averaged surface temperature, from approximately zero to about $-27\,°C$.

Continental freeboard and atmosphere CO_2 concentrations (or luminosity) play complementary roles in determining ice volume—the higher the freeboard, the greater the level of CO_2 consistent with ice-covered continents. Sensitivity experiments (not shown), in which we raised baseline continental topography by 100-m increments, indicate that for a 500-m baseline, the snowball Earth solution occurs with CO_2 levels approximately 50–100% greater than present. These are reasonable freeboards for the interval (800–550 Myr ago) associated with the pan-African orogeny.

A number of additional experiments (not shown) were run to determine the sensitivity of our results to continental configuration. For example, most of the Neoproterozoic glaciation occurred in an era of dispersed geography[4]. In the absence of a reliable 590-Myr geography, we tested the effect of continental dispersion by using a Cenomanian (94 Myr ago) geography[33]. No qualitative difference was observed—although total ice volume was approximately one-third less than in our supercontinent runs, and the bifurcation point for the transition between cold and 'snowball' conditions occurred at a higher CO_2 level, equivalent to about $1\,W\,m^{-2}$ more infrared forcing. Runs with alternative Neoproterozoic reconstructions[2,29] yielded similar results, although the critical CO_2 level again differed slightly ($<1\,W\,m^{-2}$) for different geographies.

To determine what difference the ice sheet makes to our solution, we ran the EBM with the same luminosity and CO_2 values used to obtain the ice-covered land-mass solution (note that snow and sea ice are still predicted). A comparison (Fig. 3) of the EBM alone and EBM/ice-model solutions indicates that an interactive ice sheet leads to a qualitatively different result. Ice elevation, and the long time constant associated with melting thick ice, allows the ice to 'weather' warm summer orbital configurations[14,34] without deglaciating. Whereas ice growth is restricted by the time constants inherent in ice flow and precipitation, snow area can grow rapidly. A CO_2 increase equivalent to a $69–70\,W\,m^{-2}$ warming is required to melt the ice (not shown). This is approximately equivalent to one-third of a bar of CO_2. The deglaciation is very rapid, requiring less than 2,000 years to complete. Both of the above results are in agreement with ref. 4. This rapid and large warming would also explain the unusual configuration of tropical carbonates resting on diamictites—a characteristic feature of Neoproterozoic sediment sequences.

Although the coupled model generates a snowball Earth solution with sea-ice albedos greater than 0.5, the highly idealized sea-ice parameterization[25] raises questions about the validity of the oceanic component of the model simulation. In addition, the EBM's diffusive heat transport mechanism is not as good a representation of tropical dynamics as it is of mid-latitude eddies. We therefore ran an additional set of stand-alone GCM mixed-layer ocean experiments (that is, without oceanic dynamics) with the fixed ice sheet obtained from our coupled run. We employed the GENESIS 2 GCM[35] which has diurnal forcing, interactive clouds, semi-lagrangian water vapour transport, and a six-layer sea-ice model. The sea-ice albedos vary between 0.7 and 0.8 (at visible wavelengths) and 0.4 to 0.5 (infrared) depending on whether temperatures are at the melting point, 5 °C below the melting point, or in between. The total albedo at each ocean point is based on sea-ice fraction, itself dependent on the sea-ice thickness. The model was run at T31 resolution (approximately $3.75° \times 3.75°$). The initial conditions involved a zonally averaged state with an average temperature of 0 °C. Simulations[36] used CO_2 values ranging from 0.5 to 2.5 times present levels, and the same luminosity decrease as in our coupled runs.

Simulations[36] with 0.5 and 1.0 times CO_2 (not shown) quickly yielded ice-covered oceans whereas a doubled CO_2 experiment resulted in ice-free tropical oceans but with a very slow drift to lower temperatures. We illustrate the last three years of the more stable $2.5 \times CO_2$ run (Fig. 4). An important feature of this experiment is that sea ice extends to only about 25° palaeolatitude (Fig. 4b), with sea-ice thickness varying from ~1 m near the ice

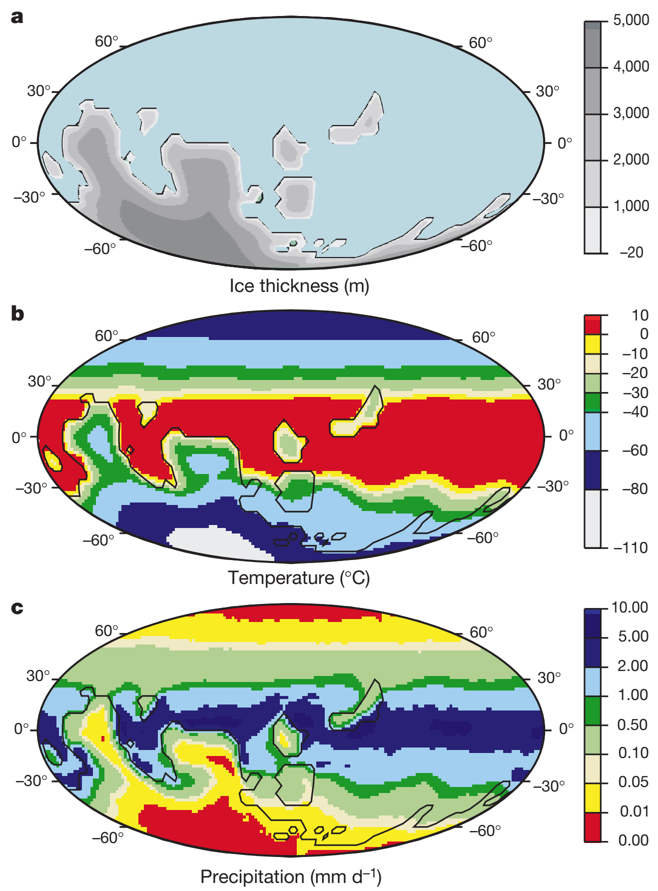

Figure 4 Late Precambrian (Neoproterozoic) climate simulations. **a**, Ice extent at equilibrium (Mollweide projection) for a 'snowball Earth' simulation. **b**, Annually averaged temperature for a GCM Neoproterozoic simulation using the above ice sheets as a lower boundary condition. A considerable area of open water exists in the tropics, though the ice sheets are almost everywhere below the melting point. **c**, Annually averaged precipitation in the GCM experiment. The cold temperatures and restricted area of open water result in a precipitation field which is less than $0.06\,mm\,d^{-1}$ over the ice-sheet accumulation zone.

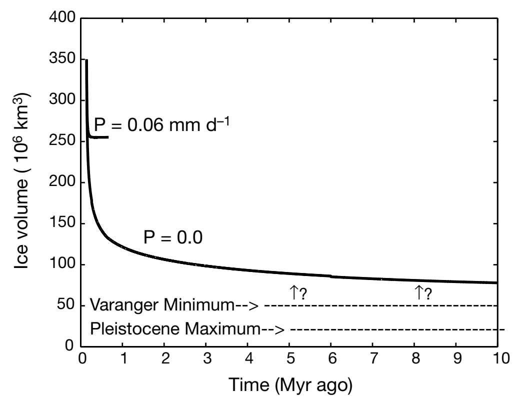

Figure 5 The effect of precipitation on global ice volume. When precipitation is reduced by a factor of ten, the ice sheets quickly achieve a slightly thinner equilibrium state. When precipitation is reduced to zero, the decline in ice volume is continuous. After 10 Myr the volume is about double that of the Pleistocene maximum, but comparable to some sea-level estimates[43] for the late Precambrian (Varanger) glaciation.

edge to 10 m in higher latitudes (the model cut-off limit). Open water can exist despite extensive ice sheets for two reasons—the ~1,500-km length scale of the influence of ice sheets on temperatures[37–39], and an important negative feedback in the GCM that is not included in the EBM. Reduced absolute humidity in a colder world reduces cloudiness by about 35%, and hence the reduction in insolation due to the younger Sun is somewhat compensated by a lower equatorial albedo (compare ref. 40). Dynamical feedback between sea-ice formation and ocean circulation might further affect the equatorial extent of sea ice. Additional model simulations with the fully coupled climate/ice-sheet model (W. T. Hyde *et al.*, manuscript in preparation) mimicking this negative feedback yield open water in the tropics. However, in this open-water case isolated land masses in equatorial regions are ice-free. This problem will be examined further in subsequent work.

Even with this large area of open water zonally averaged precipitation over much of the ice sheet is less than 0.05 mm d^{-1}; this is because most of the precipitation rains out over the open water, with very little transport into continental interiors. Temperatures below zero over most of the ice sheet indicate that open water is consistent with a mainly ice-covered land area (except for those ice sheets confined to small land masses in the equatorial ocean). Such open-water conditions near the ice are consistent with peritidal deposits[41] for the south Australia (Fig. 1) glacial sequence.

The near-snowball climate, which was first conjectured by Kirschvink[42], is qualitatively different from that of a true snowball Earth. This result is also not consistent with the interpretation[4] of $\delta^{13}C$ records as indicating a near-abiotic ocean. However, if metazoans are found to have evolved by that time, and if the $\delta^{13}C$ record is shown to be open to alternative interpretation (see further comments below), then the open-water solution represents a straightforward reconciliation of these possibilities with equatorial glaciation.

Although the model can rather easily generate ice-covered continents with plausible changes in boundary conditions, the model-generated ice volume (nearly 10 times Pleistocene values) is in excess of estimated sea-level changes due to Precambrian glaciations[4,7]. Such a change would also increase mean ocean salinities by about 20%. The very large ice sheet raises the question of whether the model produces reasonable ice volumes for a given ice area. Several lines of reasoning lead us to believe that our ice sheets are not unrealistically thick. For example, our ice sheets are considerably thinner than predicted by one published area/volume relation[44]. The model's predicted Pleistocene ice volumes are in line with observations[21]. Although our single-domed Neoproterozoic simulation tends to exaggerate ice volume as compared with a more realistic multidomed reconstruction, a comparison between this model's single-domed Laurentide simulation[21] and a multidomed simulation[45,46] implies that this exaggeration is not large. The latter effect would also to some degree be offset by the fact that we do not allow ice to accumulate on continental shelves or epicontinental seas.

As our GCM results indicate (Fig. 4c), a radical reduction in precipitation should occur[4] as a result of decreased temperatures and increased sea ice extent. However, the continued ablation of the ice sheet (largely due to calving of icebergs) would decrease ice volume. This would in turn allow sea level to rise, partially reversing the initial large sea-level drop. We tested this possibility through a series of runs in which the precipitation was lowered from 0.6 to 0.06 mm d^{-1}, and further to a limiting case of 0.0 mm d^{-1}, starting from the glaciated state of Fig. 4. The 10-Myr duration of our zero-precipitation run should be regarded as a minimum estimate for the time required to produce this sea-level drop, as the model uses a temperature- and pressure-independent rheology, and ice velocities would be much slower in the very cold snowball Earth, particularly as the ice sheets thinned, resulting in a slower rate of mass loss at the coast. When precipitation is reduced the ice sheets initially decay quite rapidly, slowing as the ice surface slopes decrease (Fig. 5). A drop in precipitation by a factor of 10 decreases the equilibrium ice amount by about 30%, while a complete shut-off of moisture reduces the ice volume by 70%, but only after ~5–10 million years. Although our 0.06 mm d^{-1} run predicts a sea-level decline far greater than a recent estimate[43] of 160 m for the Neoproterozoic, the magnitude of the observational constraint is still open to question. For example, it took more than a century to narrow down the sea-level decrease for the Last Glacial Maximum[47]. The larger ice volume we simulate could explain the thickness of postglacial carbonates (>300 m) precipitated on continental shelves. Such an amount could be accommodated by the degree of isostatic subsidence (~300 m) accompanying a 1,000-m ice sheet at the coast (Fig. 2a); overdeepening of continental shelves due to ice-sheet scouring would further enhance such accommodation (for reference, shelf depth is several hundred metres in places around Antarctica).

Discussion

An ice-covered Neoproterozoic land mass is predicted by our coupled climate/ice-sheet model with only two significant changes in boundary conditions from present values—a solar luminosity decrease consistent with estimates from solar physics, and a CO_2 level within about 50% of the present value. The final phase of the glacial advance, and the subsequent deglaciation, are consistent with the hypotheses proposed by Hoffman *et al.*[4]. These conclusions are relatively robust, in that they hold true for different land–sea configurations and continental freeboard. The results were obtained without any retuning of a model developed to explain the last glacial cycle.

A major finding of this study is that an area of open water in the equatorial oceans—which could have allowed for the survival of metazoans—is consistent with the evidence for equatorial glaciation at sea level. Although this result requires more model testing, it does not seem to be completely at variance with $\delta^{13}C$ data. Several studies suggest $\delta^{13}C$ values of −3‰ to −4‰ in deposits flanking the Varanger glaciation[4,48,49]; such values are not consistent with a near-abiotic ocean, but would be consistent with the low level of ocean productivity that might have occurred in an ocean refugium of the size we simulate. Because of the need by early metazoans for a shallow water substrate, such an area could only be a true refugium if it contained ice-free continental shelves during the otherwise near-total glaciation. Identification of such sites would be a critical test of our open-water result (some of the equatorial land masses illustrated in Fig. 1 could have been in different locations 590 Myr ago). Length-scale arguments discussed above suggest that these sites would probably be isolated equatorial land masses at some distance from the main land mass (Fig. 1). If such sites are found, they should provide information about metazoa during one of the most significant 'bottlenecks' in the evolution of life. For example, selection pressure exerted on biota by the extreme climates may have led to rapid development of new forms during glaciation and afterwards, with the sea-level rise and expansion of biota into the large number of new and unpopulated habitats. Although there is clearly a need for more climate modelling and additional geological data (for example, clarification of the $\delta^{13}C$ and sea-level evidence, and more high-quality palaeomagnetic sites), our results indicate that inclusion of explicit ice-sheet physics significantly closes the gap between models and data for the largest glaciation of the past billion years and for one of the most critical intervals of evolution in Earth history. □

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Correspondence and requests for materials should be addressed to W.T.H. (e-mail: hyde@rossby.tamu.edu).