Absence of a runaway ice-albedo feedback in the Neoproterozoic

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ABSTRACT

A runaway ice-albedo feedback at low latitudes has been proposed as an essential mechanism for achieving an ice-covered Earth. Previous simulations of the Neoproterozoic climate using a coupled ocean-atmosphere model have failed to trigger this feedback, but have also not simulated sea ice equatorward of 30°. To directly test for a runaway ice-albedo feedback, a 1000 yr simulation of the Neoproterozoic using a coupled ocean-atmosphere general circulation model was completed with sea ice specified to 10° latitude. In this experiment, low-latitude sea-surface temperatures remain above freezing, inhibiting sea-ice advance toward the equator. An ice-free ocean is maintained by a low latent-heat flux to the atmosphere and low ocean-heat transport from low latitudes. A Neoproterozoic climate with low-latitude sea ice is not stable. Once the sea-ice specification was released, allowing sea ice to melt, the sea-ice margin retreated poleward of 45° latitude. These climate simulations demonstrate the resilience of Earth's climate and emphasize the extreme conditions required to initiate a ''hard'' snowball Earth.

Keywords: paleoclimate, general circulation models, Neoproterozoic, ice ages, oceans.

INTRODUCTION

Low-latitude glaciogenic sequences during the Neoproterozoic likely mark Earth's climate nadir. To explain these deposits, Kirschvink (1992) proposed a snowball Earth, a Neoproterozoic world with extensive continental glaciation and pack ice over most of the middle and high latitudes. On the basis of $\delta^{13}C$ data from carbonate rocks capping the low-latitude Neoproterozoic glacial deposits, Hoffman et al. (1998) suggested that the entire ocean was covered by ice (also known as a ''hard'' snowball). In the snowball Earth hypothesis of Hoffman et al. (1998) and Hoffman and Schrag (2002), an ice-covered Earth was initiated by the burial of atmospheric CO_2 as marine sediments through silicate weathering of tropical continents. The reduction in atmospheric CO_2 caused global cooling, triggering a runaway ice-albedo feedback (Hoffman et al., 1998; Hoffman and Schrag, 2002), which has a positive feedback on climate.

The runaway ice-albedo feedback represents a climate instability in which the surface radiation over open-ocean regions does not balance the heat loss to ice-covered regions and to space, causing cooling and ice growth. This instability has been demonstrated in a number of energy-balance models (EBMs) (e.g., Budyko, 1969; Held and Suarez, 1974) and atmospheric general circulation models (AGCMs) (Wetherald and Manabe, 1975). The runaway ice-albedo feedback also occurs in many AGCM simulations of the Neoproterozoic (e.g., Baum and Crowley, 2001; Jenkins and Smith, 1999; Poulsen et al., 2001). However, a consistent feature of the EBMs and AGCMs is their simplified, nondynamical representation of the oceans. This simplified treatment of the oceans may affect the initiation point of a runaway ice-albedo feedback. By using an energy-balance model, Held and Suarez (1974) illustrated that the latitude at which the ice-albedo feedback occurs is sensitive to the poleward heat transport in these models. Poulsen et al. (2001) demonstrated that ocean dynamics play an important role in halting the advance of the sea-ice margin in Neoproterozoic simulations.

The initiation point of a runaway ice-albedo feedback is critical to identifying the mechanism that triggered a hard snowball Earth. Our previous Neoproterozoic simulations (Poulsen et al., 2001, 2002) have tested a large range of model parameter space, yet a runaway ice-albedo feedback has not been identified in coupled ocean-atmosphere models. Here, the presence of a runaway ice-albedo feedback in coupled ocean-atmosphere models of the Neoproterozoic is specifically explored.

MODEL DESCRIPTION

The model experiments were completed using the Fast Ocean-Atmosphere Model (FOAM) version 1.4, a fully coupled ocean-atmosphere GCM. The atmospheric component of FOAM is a parallelized version of the National Center for Atmospheric Research's Community Climate Model 2 (CCM2) with the upgraded CCM3 physics from version 3.6 (described in Kiehl et al., 1996). The atmospheric model contains 18 vertical levels and a spectral resolution of R15 ($4.5° \times 7.5°$). The ocean component (OM3) contains 16 vertical layers and uses a 128×128 point grid ($1.4° \times 2.8°$). In this version of FOAM, sea ice forms when an ocean grid cell is cooler than -1.9 °C and disappears when the grid cell is warmer than -1.9 °C. Sea ice has a constant thickness and albedos of 0.70 and 0.50 in the visual and near-infrared wavelengths. FOAM does not yet include sea-ice dynamics.

FOAM successfully simulates many aspects of the present-day climate and compares well with other contemporary medium-resolution climate models (Jacob, 1997). The climate sensitivity of FOAM is similar to other contemporary models. For example, in a simulation of the Last Glacial Maximum, tropical sea-surface temperatures (SSTs) are ~2–3 °C less than simulated modern SSTs.

The Neoproterozoic experiment includes an idealized supercontinent centered on the equator, similar to the modeled supercontinents of Crowley and Baum (1993), Jenkins and Frakes (1998), and Jenkins and Smith (1999), with two 500 m N-S mountain ranges on its western and eastern margins. Elsewhere, the relief is 50 m. Our purpose in implementing the idealized paleogeography was to maximize tropical continental area, an important factor if the ice-cap instability was triggered by silicate weathering (Marshall et al., 1988). Because land plants had yet to evolve by the Neoproterozoic, the land surface in the model has the radiative characteristics of a desert (albedo of 0.35 and 0.51 in the visible and near-infrared wavelengths, respectively). At 600 Ma, the solar luminosity was between 4.7% and 6.3% lower than at present (Crowley and Baum, 1993). To facilitate snowball conditions, a 7.0% reduction in the solar constant and a CO_2 value of 140 (ppmv) was used. The CH_4 concentration was set to the preindustrial value (700 ppbv [billion in this use is 10^9]). The model eccentricity, obliquity, precession, rotation rate, and ozone concentrations were defined as modern values. The ocean has a uniform depth of 5000 m. To further facilitate snowball conditions, this experiment was initialized with the following ocean temperature profile: 0.0 °C (10 m), -0.5 °C (30 m), -1.0 °C (75 m), -1.5 °C (125 m), and -1.8 °C (200–5000 m). The ocean was initialized with a uniform salinity of 34.9 psu.

NUMERICAL EXPERIMENTS

Two Neoproterozoic experiments were completed. In the first experiment (the prescribed-ice experiment), sea ice was prescribed from the poles to 10° latitude. In this run, sea ice could grow and advance to the equator, but could not melt and retreat to the poles. If SST at latitudes >10° exceeded the freezing point of seawater, SST was reset to the freezing point. The prescribed-ice experiment was run for 1000 model years to allow the deep ocean to equilibrate.

A second experiment (the free-ice experiment) was initiated from the final year of the

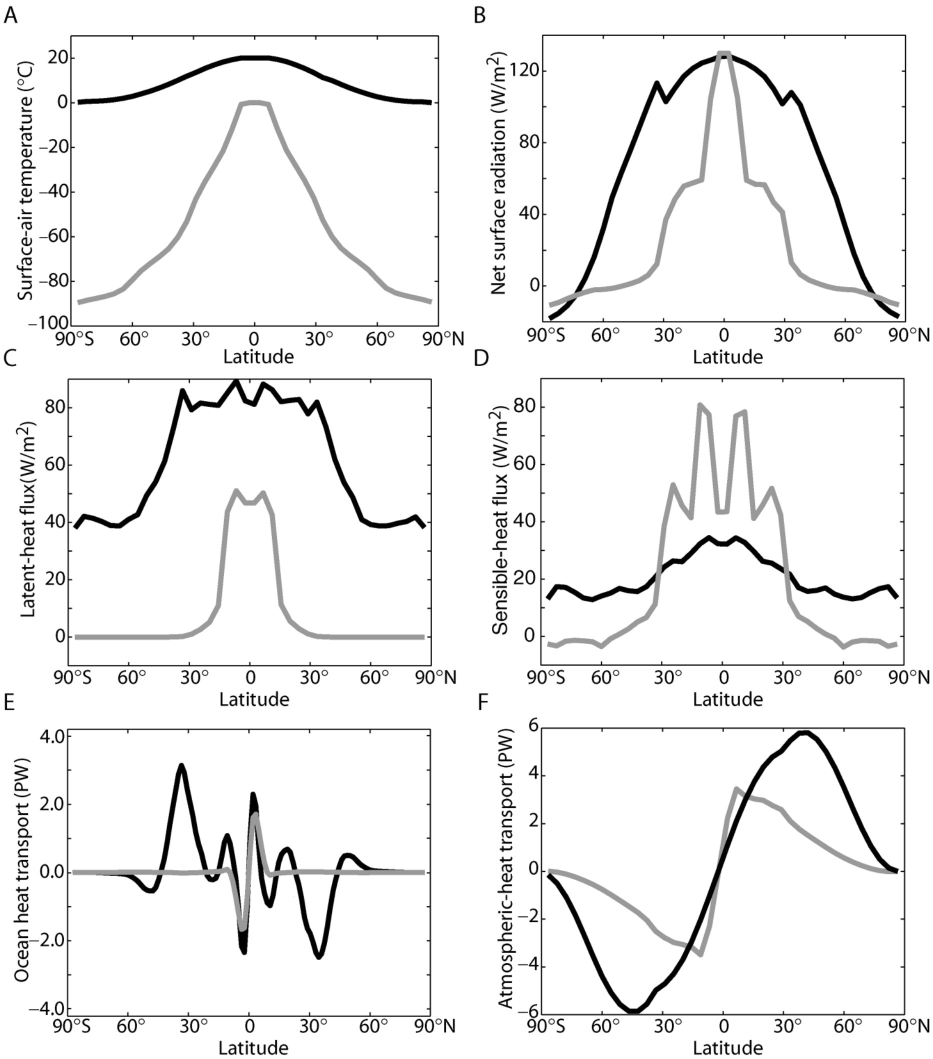

Figure 1. A–D: Annual (5 yr mean) zonally averaged surface-air temperature over ocean (in degrees Celsius), surface radiation, latent- and sensible-heat fluxes (in watts per square meter). E–F: Ocean-heat and atmospheric-heat transport (in petawatts). Results from prescribed-ice and free-ice experiments are shown in gray and black, respectively.

first run, and integrated for 50 model years. In this branch run, the sea-ice specification (from the poles to 10°) was released. In other words, sea ice was allowed to melt, and the sea-ice margin could retreat poleward.

The simple sea-ice model in FOAM version 1.4 does not allow for sea-ice thickness to change. In nature, sea ice will thicken until an energy balance is achieved between the air temperatures above the sea ice and seawater below the sea ice. Because this process does not occur in the model, a uniform sea-ice thickness of >1000 m is prescribed in both experiments, which is consistent with simple thermodynamic calculations of sea-ice thickness (e.g., Goodman and Pierrehumbert, 2003). In FOAM, the sea-ice thickness has no effect on the ocean circulation. In the presence of sea ice, the ocean model does not ''see'' the atmosphere.

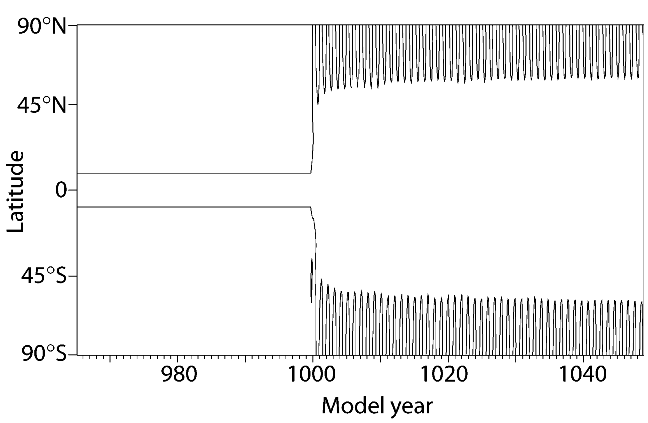

Figure 2. Latitudinal position of sea-ice line during past 85 yr of model run. Sea-ice line represents complete circumglobal sea-ice coverage. Until model year 1000 (prescribed-ice experiment), sea ice was not allowed to melt. Note abrupt retreat and seasonal variation of sea-ice line after model year 1000 (free-ice experiment).

CLIMATE MODEL RESULTS

The specification of low-latitude sea ice to 10° latitude represents a significant climatic forcing. In comparison to the ocean surface, sea ice has a higher albedo and reflects more incoming solar radiation to space. As expected, this forcing has a significant effect on Neoproterozoic surface temperatures (Fig. 1A). The global average surface-air temperature in this experiment is −45.6 °C. Nonetheless, after 1000 model years, sea ice has not formed equatorward of 10°, indicating the absence of a runaway ice-albedo feedback (Fig. 2). The average open-water SST for the prescribed-ice experiment is −1.73 °C.

An important consideration is whether the prescribed-ice experiment is in equilibrium, or whether it is cooling, leaving the possibility that sea ice will eventually advance to the equator. In fact, this experiment is warming. Over 1000 yr, the global mean annual ocean-temperature time rate of change is 5.0×10^{-5} $°C \cdot yr^{-1}$. Over the final 100 yr, the global mean annual ocean temperature increased at a rate of 2.3×10^{-5} $°C \cdot yr^{-1}$. The global mean annual air temperature shows essentially no change over the final 50 yr of model integration.

With prescribed sea ice, it is questionable whether the climate is steady state. To evaluate the stability of the climate, the sea-ice forcing was released in the free-ice experiment, allowing the sea ice to melt. In the absence of the sea-ice forcing, the global average surface temperature increased to 10.8 °C (Fig. 1A), and the sea-ice margin retreated poleward of 45° latitude (Fig. 2). The sea-ice extent and seasonal fluctuations are similar to previous Neoproterozoic simulations using FOAM (Poulsen et al., 2001, 2002). This experiment is also warming. Over 50 yr of model integration, the global ocean temperature increased at a rate of 9.98×10^{-3} $°C \cdot yr^{-1}$.

The tropical climate is influenced heavily by dynamics that redistribute energy through the transport of moisture, heat, and momentum. Yet, in the broadest terms, the maintenance of an ice-free tropical region can be considered an energy-balance problem. The heat exported from the low-latitude sea surface must be balanced by the global net surface radiation (i.e., the difference between the net incoming solar radiation and the net outgoing longwave radiation), or cooling will prevail. A comparison of these two Neoproterozoic experiments, with and without pre-

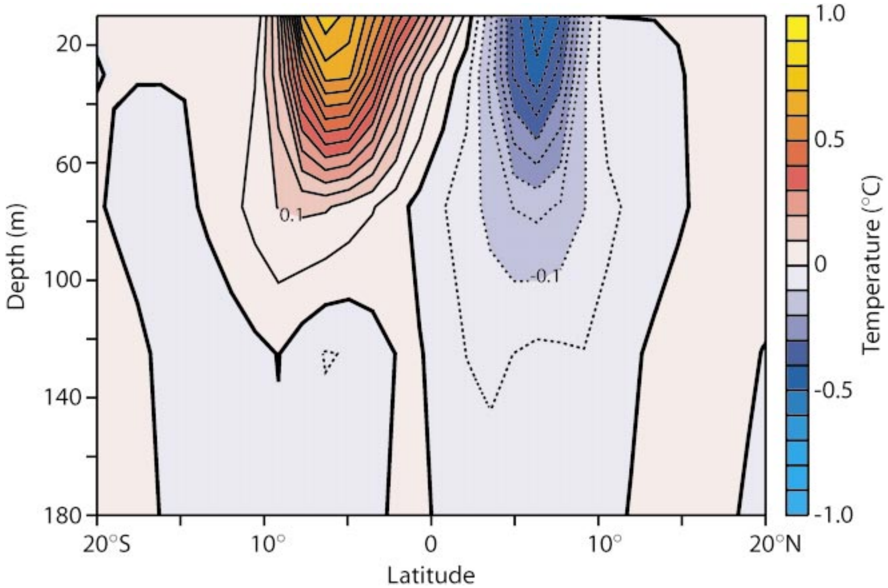

Figure 3. Zonally averaged, upper-ocean temperature difference (color scale in degrees Celsius) between January and July in prescribed-ice experiment. Note that seasonal temperature change occurs over 150 m depth. Difference represents 5 yr average.

scribed sea ice, demonstrates that changes in the low-latitude surface-energy balance will maintain an ice-free low-latitude ocean and prevent a runaway ice-albedo feedback. Between the Neoproterozoic experiments, the global net surface radiation decreases from 90.5 to 37.8 W/m^2 in the prescribed-ice experiment. In the low-latitude, ice-free regions, however, the net surface radiation decreases by only 9.8 W/m^2 (from 127.2 to 117.4 W/m^2) between experiments and actually increases near the equator (Fig. 1B). Moreover, the ice-free ocean region of the prescribed-ice experiment loses 36.4 W/m^2 less heat through latent-heat flux to the atmosphere (Fig. 1C). The decrease in latent-heat flux is the result of lower atmospheric temperatures and saturation vapor pressures, causing evaporation rates to decrease. In addition to the reduction in latent-heat flux, ocean-heat transport from the low latitudes is also diminished in the prescribed-ice experiment (Fig. 1E). In both experiments, ocean-heat transport occurs primarily through the wind-blown upper-ocean circulation. Over sea-ice–covered regions, the upper ocean is isolated from the wind-forcing and ocean heat transport is reduced. However, because of the extensive sea-ice distribution, ocean heat transport is least in the prescribed-ice experiment.

Increases in sensible-heat flux of 26.8 W/m^2 offset the gains made by the reduction in latent-heat flux and ocean-heat transport (Fig. 1D). The principal source of sensible heat is the seasonal (i.e., winter) heat loss from the ocean to the atmosphere in the ice-free regions. (The supercontinent is a smaller, but important, source of sensible heat. In Figure 1D, note that the poleward edges of the supercontinent at 30° latitude are marked by a sharp decline in zonally averaged sensible-heat flux.) Over most of the open ocean, the heat flux to the ocean is positive during all seasons. However, near the sea-ice margin, the heat flux is negative, reaching values <-190 W/m^2 during the winter in the prescribed-ice experiment (not shown). Figure 3 illustrates the seasonal temperature change (January–July) in the upper ocean. Heat stored in the upper ocean (i.e., the upper ~150 m) during the summer months provides the heat to maintain above-freezing winter SSTs.

In addition to a decrease in ocean-heat transport, atmospheric heat transport is also diminished in the prescribed-ice experiment (Fig. 1F). At low latitudes, the decrease in atmospheric heat transport is largely due to the reduction in the latent-heat loss from the surface (Fig. 1C). The peak atmospheric heat transport occurs at the sea-ice margin, the location of maximum sensible-heat flux.

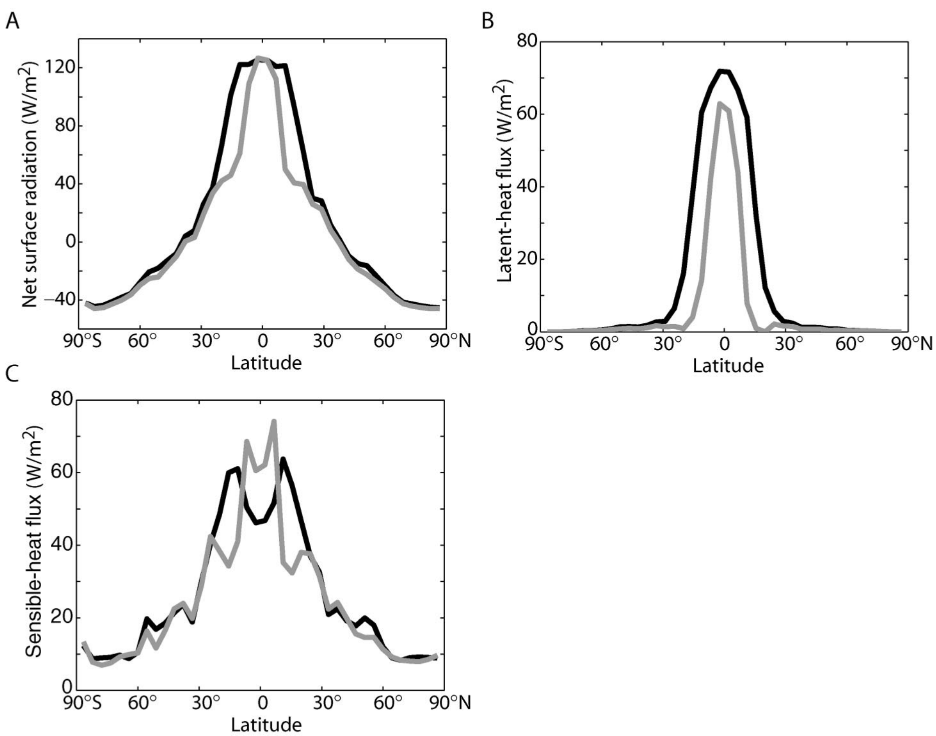

Figure 4. Zonally averaged net surface radiation and latent and sensible heat fluxes (in watts per square meter) calculated from uncoupled atmospheric general circulation model (AGCM) simulation of Neoproterozoic conditions. Black lines represent fluxes during model year 33 of simulation when sea-ice line was positioned at ~15° latitude. Gray lines represent heat transport during model year 38, when sea-ice margin was situated at <10° latitude. In this uncoupled AGCM experiment, oceans were completely ice covered by model year 43 (Poulsen et al., 2001).

DISCUSSION AND CONCLUSIONS

Previous studies (e.g., Baum and Crowley, 2001; Jenkins and Smith, 1999; Poulsen et al., 2001) have demonstrated a runaway ice-albedo feedback in AGCM simulations of the Neoproterozoic. However, these AGCMs exhibit changes in surface radiation and heat fluxes similar to those shown here with the incursion of sea ice into low latitudes. For example, Figure 4 illustrates the changes in zonally averaged net surface radiation, latent-heat flux, and sensible-heat flux for years 33 and 38 of a Neoproterozoic experiment run completed with an AGCM coupled to a mixed-layer ocean model (see Poulsen et al., 2001, for details). A runaway ice-albedo feedback was exhibited in this experiment. As the sea-ice line moved toward the equator, the low-latitude (i.e., ice-free regions) net surface radiation was largely unchanged (Fig. 4A); the latent-heat flux declined (Fig. 4B); and the sensible-heat flux increased (Fig. 4C). In contrast to the coupled experiment, the ocean-heat transport increased (not shown). This model response is the result of using a diffusive mixed-layer ocean; as the meridional temperature gradient increases, the heat transport increases. However, it is unlikely that the increase in ocean-heat transport is the only cause of the sea-ice advance, because a similar experiment with no ocean-heat transport also predicted ice-covered oceans (Poulsen et al., 2001). Moreover, many AGCMs, including the GENESIS model used by Baum and Crowley (2001) and Jenkins and Smith (1999), parameterize ocean-heat transport in different ways.

Another factor that may encourage a runaway ice-albedo feedback in AGCMs is the fact that mixed-layer ocean models underrepresent the heat capacity of the Neoproterozoic ocean. A typical mixed-layer ocean model depth is 50 m, significantly less than the nearly 150 m of water that accounts for the seasonal flux of heat from the ocean in the coupled experiments (Fig. 3). The heat capacity of a 150-m-deep water body is three times that of a 50-m-deep water body. As a result, the release of sensible heat to the atmosphere during the winter will cause more surface cooling in the 50-m-deep mixed layer than in the 150-m-deep mixed layer. Neoproterozoic experiments using the Goddard Institute for Space Studies (GISS) AGCM, which implements a mixed-layer ocean with a seasonally varying depth, support this conclusion. A hard snowball Earth has not been simulated with the GISS model (Chandler and Sohl, 2000).

The atmospheric CO_2 values used in this study are much lower than the values used in previous studies (e.g., Jenkins and Smith, 1999; Hyde et al., 2000; Baum and Crowley, 2001) that simulate a snowball Earth. It is possible that a further reduction of atmospheric pCO_2 in the model could drive the sea-ice line equatorward of its mid-latitude position in the free-ice experiment. However, in the absence of a runaway ice-albedo feedback, the silicate weather mechanism is not an effective trigger for initiating an ice-covered Earth. Lower pCO_2 levels cannot be justified in the prescribed-ice experiment because tropical continental temperatures are well below freezing throughout the year, preventing silicate weathering.

The geological jury on a hard snowball Earth is still out. For example, the duration of the Neoproterozic glaciations may have been much shorter than previously thought (Bowring et al., 2003). Moreover, the interpretation of the geological evidence is contentious. For example, Kennedy et al. (2001) offered an alternative to the Hoffman et al. (1998) snowball Earth hypothesis—that the negative $\delta^{13}C$ excursion in the cap carbonates represents the destabilization of gas hydrate in terrestrial permafrost—that does not require the oceans to be completely frozen (see Hoffman and Schrag [2002] for possible shortcomings of the gas hydrate hypothesis). The Neoproterozoic experiments described here side with a less extreme Neoproterozoic climate, but cannot rule out an ice-covered Earth. Rather, the model results suggest that an ice-covered Earth must have been the product of an extreme climate event that substantially reduced the receipt of solar radiation at low latitudes. This implication is all the more compelling because we offer no justification for the sea-ice forcing used in this study.

In summary, a runaway ice-albedo feedback is not exhibited in coupled ocean-atmosphere experiments of the Neoproterozoic. Moreover, the Neoproterozoic climate experiments indicate that a low-latitude sea-ice distribution cannot be sustained. The Neoproterozoic ocean-atmosphere experiments demonstrate the resilience of Earth's climate and its suitability for life. Under extreme forcing, the climate system maintains ice-free regions that might have served as refuge for early life (Runnegar, 2000).

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