A snowball Earth versus a slushball Earth: Results from Neoproterozoic climate modeling sensitivity experiments

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ABSTRACT

The Neoproterozoic was characterized by an extreme glaciation, but until now there has been no consensus as to whether it was a complete glaciation (snowball Earth) or a less severe glaciation (slushball Earth). We performed sensitivity experiments with an Earth model of intermediate complexity for this period of dramatic global cooling. Our simulations focus on the climate response on a cool versus a cold ocean, on a desert versus a glacier land surface, and on a lower versus a higher CO_2 concentration. All Neoproterozoic model experiments represent much colder conditions than today and widespread glaciation. In case of an initial forcing representing a snowball Earth, the model maintains its complete glaciation, and temperatures are as low as –45 °C in equatorial latitudes. At the poles, the snowball experiments demonstrate annual average temperatures of <–70 °C. If the initial model forcing is more moderate (slushball Earth), polar temperatures are <–50 °C, but temperatures in low latitudes stay well above the freezing point of water, and therefore ice-free ocean areas remain. Based on our simulations, we are able to observe that global climate reacts less sensitively to reductions of atmospheric CO_2 during times with increased glaciations. Our results suggest that the development of glaciers on land contributes significantly to intense ice coverage of the oceans. Because simulations initialized without complete ice cover do not reach the global glaciation condition, we conclude that our simulations support the rather moderate scenario of a slushball Earth than the extreme snowball Earth hypothesis. The experimental design and the model might, however, limit the interpretation of our results.

Keywords: Neoproterozoic, snowball Earth, climate modeling, climate modeling sensitivity experiment.

INTRODUCTION

The Neoproterozoic (1000–542 Ma; Gradstein et al., 2004) was a rather cold epoch in Earth's history. Neoproterozoic sedimentary deposits can be found across the world and they document an intense degree of glaciation for that time (e.g., Young and Gostin, 1989; Hoffman et al., 1998b; Evans, 2000; Evans et al., 2000). Based on lithostratigraphic and chemostratigraphic evidence from Australia, Canada, Congo, Namibia, and Spitzbergen (Meert et al., 1995; Schmidt and Williams, 1995; Kaufman et al., 1997, and references therein; Sohl et al., 1999), as many as five successive ice ages have been proposed for the entire Neoproterozoic (e.g., Hoffman et al., 1998a; Saylor et al., 1998). However, analyzing the available data, Kennedy et al. (1998) concluded that only two major glaciations can be identified with confidence. The existence of a third, possibly Marinoan-aged, major glaciation, evidenced by glacigenic deposits at the Varanger Peninsula in northern Norway (Fairchild and Hambrey, 1984; Knoll et al., 1986), is debated (e.g., Kennedy et al., 1998, Gradstein et al., 2004). Based on paleomagnetic reconstructions, glacial conditions in the Neoproterozoic advanced even into equatorial latitudes (e.g., Evans, 2000, and references therein). Owing to the widespread occurrence of glacial deposits in the Neoproterozoic, some studies proposed a rather extreme and provocative scenario, which is now well known as the so-called Neoproterozoic snowball Earth hypothesis (e.g., Kirschvink, 1992; Hoffman et al., 1998b; Schrag et al., 2002). However, serious doubts arose as to whether the Neoproterozoic Earth was really completely glaciated (e.g., Christie-Blick et al., 1999; Leather et al., 2002). For example, some massive geological deposits indicate an actively working hydrological cycle during the Neoproterozoic (e.g., Christie-Blick et al., 1999; Leather et al., 2002). In addition, the organic geochemical analysis of several chemofossils (so-called biomarkers) extracted from low-latitude Neoproterozoic synglacial black shales of the São Francisco craton in southeastern Brazil document the existence of a widespread, photosynthetic, active, and therefore light-dependant, complex microbial ecosystem (Olcott et al., 2005). Likewise, silicified microfossils from preglacial and synglacial deposits in the Death Valley area, California, and acritarchs from preglacial and postglacial successions of Australia show no evidence for a prominent extinction phase during that time (Corsetti et al., 2003, 2006). These facts contradict the scenario of a snowball Earth, which implies that the oceans froze to a depth of ~1 km (Hoffman et al., 1998b; Runnegar, 2000), thereby forming a dense to non-light-transmissive layer, and support a modified version of a less severe Neoproterozoic glaciation (i.e., a slushball Earth).

It is a matter of an ongoing discussion whether it was a Neoproterozoic snowball or slushball Earth (e.g., Christie-Blick et al., 1999; Hoffman and Schrag, 1999; Leather et al., 2002; Allen and Hoffman, 2005). Not at least because of this controversial debate on the degree of the glaciation, the Neoproterozoic paradox has attracted the interest of paleoclimate modelers. Using sensitivity experiments with climate models of different complexity, modelers analyzed the various processes that might have initiated (e.g., Chandler and Sohl, 2000; Donnadieu et al., 2004a) or terminated (e.g., Pierrehumbert, 2004) the Neoproterozoic glaciation. These climate modeling studies focused on the role of a reduced solar luminosity and a different configuration of the Earth's orbit, the effects of different

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concentrations of atmospheric carbon dioxide, the relevance of the paleogeography and/or paleo-orography and changed land surface cover, and the ocean (e.g., Chandler and Sohl, 2000; Hyde et al., 2000; Poulsen et al., 2001, 2002; Donnadieu et al., 2003, 2004a, 2004b; Poulsen, 2003; Pierrehumbert, 2004; Poulsen and Jacob, 2004; Romanova et al., 2006). One could think that models give a consistent view on the Neoproterozoic because they are based on (more or less) unalterable physical principles. With some limitations, different climate models produce quite consistent results when applied to the present situation and future climate change (e.g., Cubasch et al., 2001). It is, however, interesting that the different climate model experiments do not give consistent answers to the Neoproterozoic climate paradox. For example, Poulsen et al. (2001) used an atmosphere model, which was coupled to an ocean circulation model and to a slab ocean model. The results of both modeling approaches differ significantly. The fully coupled model was not able to get into full glaciation, while the other, with its slab ocean, was. In another model study with a fully coupled atmosphere-ocean model, Poulsen and Jacob (2004) obtained full snowball conditions, but only if the crucial parameters of wind-driven circulation and cloud-radiative forcing were absent. What is wrong if model simulations (even of the same study) contradict each other that much (e.g., Chandler and Sohl, 2000; Hyde et al., 2000; Poulsen et al., 2001; Donnadieu et al., 2003, 2004a, 2004b)?

The ocean circulation and ocean heat transport in particular play prominent roles (e.g., Covey and Thompson, 1989; Rind and Chandler, 1991; Cohen-Solal and Le Treut, 1997), not only for the present-day climate (e.g., Mikolajewicz and Voss, 2000) but also for past time intervals (e.g., Maier-Reimer et al., 1990; Rind and Chandler, 1991; Mikolajewicz and Crowley, 1997; Brady et al., 1998; Bice et al., 2000; Steppuhn et al., 2006). Poulsen et al. (2001) studied the role of the ocean with respect to the onset of a Neoproterozoic glaciation with a slab ocean and a fully coupled climate model: the atmosphere model coupled to a mixed-layer ocean model produced a snowball Earth, but even for present-day conditions, sea ice extends too far toward the equator with the used model configuration. Therefore, the reliability of the solution of a snowball Earth in that study is questionable. Other Neoproterozoic model experiments also include some simplifications, such as an idealized continent (e.g., Poulsen et al., 2002) or a uniform paleo-orography (e.g., Donnadieu et al., 2004a), or used even modern boundary conditions (e.g., Romanova et al., 2006). This is justified partly because detailed information is lacking and partly because of the experimental design as sensitivity simulations, but at least it also documents the problems in explaining the Neoproterozoic icehouse phase.

The inconsistencies and shortcomings of previous Neoproterozoic simulations and the contradictory interpretations of the geological record demonstrate that the Neoproterozoic glacial phase is not well understood. Also, there are still many uncertainties with respect to various parameters such as the paleogeography (e.g., Meert and Powell, 2001; Torsvik, 2003); this motivates our study. We define some climate modeling experiments to investigate some fundamental processes with an Earth system model of intermediate complexity, i.e., an atmospheric general circulation model coupled to a slab ocean model. The Neoproterozoic simulations are designed as sensitivity experiments and consider basically two different ocean settings, a cool and a cold situation. In addition, we analyzed the climate response on different global land surface cover (desert versus glacier), and we investigated the relevance of variations of atmospheric pCO_2 (510 ppm versus 280 ppm). On the one hand, we analyzed the climate response to these changed initial conditions, and on the other hand, we checked how consistent our results are compared to other modeling studies and the fossil record. We could not analyze the climate response to all parameters relevant for past climate changes within our present study; however, we established a set of boundary conditions that can be used as a basis for further model experiments.

EXPERIMENTAL DESIGN

EMIC Planet Simulator

We performed climate modeling sensitivity experiments for the Neoproterozoic. For these purposes, we used the Earth system model of intermediate complexity (EMIC) Planet Simulator (e.g., Fraedrich et al., 2005a, 2005b). The core module is the simple atmospheric general circulation model (AGCM) PUMA-2, which is coupled to a slab ocean and thermodynamic sea ice model. PUMA-2 is based on the primitive equations representing, for example, the conservation of mass and momentum, and the first principle of thermodynamics. As compared to its previous dry-dynamics version (e.g., Fraedrich et al., 1998; Frisius et al., 1998), the atmosphere model was improved and includes now schemes for various physical processes such as radiation transfer, large-scale and convective precipitation, and cloud formation. The AGCM is coupled to a thermodynamic sea ice and to a mixed-layer ocean model (i.e., the model does not simulate the ocean circulation, but it accounts for the heat exchange between atmosphere and ocean). It is also possible to force the model with observational climatological fields such as present-day sea surface temperatures. As compared to the highly complex circulation models, the EMIC Planet Simulator consists of simpler parameterization schemes, e.g., for radiation transfer. For further details about the model physics, see Fraedrich et al. (2005a, 2005b). At the expense of some uncertainties, intermediate complexity models such as the Planet Simulator or CLIMBER-2 (e.g., Petoukhov et al., 2000) are generally computationally much more efficient than the more realistic complex general circulation models. For some specific applications, EMICs are useful tools as they bridge the gap from the rather simple to highly complex models (Claussen et al., 2002). The Planet Simulator was used for the present-day climate (e.g., Fraedrich et al., 2005b; Junge et al., 2005; Grosfeld et al., 2007; Kleidon et al., 2007), where it proved its reliability. Micheels et al. (2006) also used the model for a late Miocene sensitivity study. This Miocene experiment gives quite consistent results as compared to a late Miocene simulation with the highly complex AGCM ECHAM4 coupled to a slab ocean model (e.g., Steppuhn et al., 2006; Micheels et al., 2007). We emphasize that Romanova et al. (2006) applied the model for some Neoproterozoic sensitivity experiments, even though they used largely modern boundary conditions.

The horizontal resolution of the spectral atmospheric model is T21, which corresponds to a Gaussian grid of ~5.6° × 5.6° for the longitude and latitude. The vertical domain is represented with five layers using terrain-following σ-coordinates. With the Planet Simulator, we perform a present-day control experiment (referred to as CTRL). The control run uses the same set of boundary conditions as the highly complex AGCM ECHAM5 (e.g., Roeckner et al., 2003, 2004). For CTRL, we force the model with present-day sea surface temperatures (SSTs) and sea ice cover (SIC), the modern vegetation, and atmospheric CO_2 set to the preindustrial concentration of 280 ppm.

Neoproterozoic Boundary Conditions

We performed Neoproterozoic sensitivity experiments, which use an adapted paleogeography and paleo-orography, and consider a lower solar luminosity by –6% as compared to today (e.g., Gough, 1981). The configuration of orbital parameters refers to present-day conditions. In contrast, SSTs, SIC, land surface cover, and atmospheric pCO_2 differ in the simulations. We performed eight Neoproterozoic sensitivity experiments, the experimental design details of which are described in the following (for a summary of the specifications, see Table 1).

Paleogeography and Paleo-orography

Figure 1 illustrates the Neoproterozoic paleogeography and paleo-orography. The land-sea distribution is based on the software Plate Tracker(http://www.scotese.com/software.htm). It represents almost one supercontinent known as Rodinia (e.g., Dalziel, 1991; Moores, 1991; Hoffman, 1991). Within the paleogeographical concept of Rodinia, Laurentia is placed in the core of the supercontinent, which is surrounded by the landmasses of Baltica–Amazonia–Rio Plata–Congo to the south and the east as well as the continents Siberia–south China–Australia–East Antarctica–India to the north and west (Torsvik, 2003). This supercontinent configuration seems to be generally supported by paleomagnetic data (e.g., Powell et al., 1993; Torsvik et al., 1996, 2001; Weil et al., 1998). From model experiments, it is assumed that the breakup of Rodinia initiated the transition into the Neoproterozoic glaciation (e.g., Donnadieu et al., 2004a) and that the supercontinent configuration with its positioning in lower latitudes is needed for an extreme degree of glaciation during the Neoproterozoic (e.g., Donnadieu et al., 2003). Even though there is debate about the exact continent configuration (e.g., Meert and Powell, 2001; Torsvik, 2003), the land-sea distribution in our simulations is largely consistent with other studies (e.g., Donnadieu et al., 2003, 2004a, 2004b).

We also adapt the paleo-orography (Fig. 1). We assume that mountains at the continental plate margins were formerly as much as 1000 m high. For interior parts of the landmasses, we used a paleoelevation of 50 m. Previous Neoproterozoic model simulations often considered a uniform altitude of ~100 m (e.g., Donnadieu et al., 2004a) because precise information about the paleo-orography is lacking. Donnadieu et al. (2003) first introduced a topographic reconstruction for the Neoproterozoic, estimating that mountains had peak elevations of 2000 m. Poulsen et al. (2001) described mountain chains of maximum 500 m height. In our Neoproterozoic sensitivity simulations, the paleo-orography is not necessarily fully realistic, but it is between previous estimates (Donnadieu et al., 2003; Poulsen et al., 2001). Also, it is known that Neoproterozoic orography was not the driving factor responsible for the initiation of snowball conditions (Romanova et al., 2006).

Land Surface Cover: Desert versus Glacier

For the land surface, we define two scenarios: desert and glacier. The glacier simulations (NEO-1 to NEO-4) include a set of surface parameters corresponding to fully ice-covered continents. For the desert experiment (NEO-5), surface parameters refer to a completely ice-free situation with desert covering the landmasses. For both scenarios, we adapted the set of surface parameters (vegetation cover, albedo, roughness length) based on Claussen (1994). The soil water capacity was modified after Hagemann et al. (1999). For the desert scenario, we specified conditions of a sand desert because it has a higher albedo ($\alpha = 0.35$) as compared to a normal desert ($\alpha = 0.20$). This setting should contribute to force a rather cold state in the model experiments.

CO_2: Higher versus Lower

The present-day control experiment CTRL uses a carbon dioxide concentration of 280 ppm. The Neoproterozoic experiments also require a specification of atmospheric CO_2. However, this is not an easy task because knowledge about the concentration of carbon dioxide in the atmosphere is less well known the further one looks back in time (e.g., Berner, 1997; Ridgwell et al., 2003). There is debate on CO_2 concentrations even for the well-known late Tertiary (e.g., Kürschner et al., 1996; Pagani et al., 2005; MacFadden, 2005), and for the Neoproterozoic (Hoffman et al., 1998b; Ridgwell et al., 2003). A climate modeling experiment for the Neoproterozoic suggests that the breakup of the supercontinent Rodinia lead to a decrease of atmospheric CO_2, resulting in values of ~510 ppm (or even lower) because of increased continental weathering rates (e.g., Donnadieu et al., 2004a). Following this study, we prescribe an atmospheric carbon dioxide concentration of 510 ppm in a first set of sensitivity experiments (NEO-1 to NEO-5). Owing to uncertainties of reconstructing past concentrations of CO_2, a second set of simulations (NEO-3-280 to NEO-5-280) uses the preindustrial CO_2 level of 280 ppm. For the snowball Earth simulations NEO-1 and NEO-2, we did not perform simulations with reduced pCO_2 because both simulations were able to maintain their full glaciation.

Initial Ocean Conditions: Cool Ocean versus Cold Ocean

For the ocean setup, we adjusted the deep-sea temperature (DST), the SST, and the depth of the mixed-layer (MLD). The initialization of the deep ocean considers a global constant temperature close to the freezing point ($DST_{global} = 273$ K). The depth of the mixed-layer is also initialized with a globally constant value ($MLD_{global} = 50$ m). With respect to the SSTs and SIC, we defined cold (NEO-1 and NEO-2) and cool (NEO-3 to NEO-5) ocean scenarios. The cold ocean scenario uses global constant SSTs of 271 K, which correspond to the freezing point of seawater ($T_{freeze} = 271.15$ K) in the Planet Simulator. Contrarily, SSTs of the cool scenario vary as a function of the cosine of the latitude [$SST(\varphi) = f(\cos \varphi)$]. At the equator, initial SSTs are set to 280 K and decline to 265 K at the poles.

TABLE 1. THE SPECIFICATIONS OF THE PRESENT-DAY CONTROL RUN AND THE NEOPROTEROZOIC SENSITIVITY EXPERIMENTS

Experiment identification	Boundary conditions	
CTRL	CO_2 = 280 ppm, S_0 = 100%, present-day geography, orography, vegetation, SSTs, and SIC	
NEO-1	CO_2 = 510 ppm, S_0 = 94%, paleogeography, paleo-orography, no explicit flux correction	+ desert land + cold ocean + global sea ice
NEO-2		+ glaciated land + cold ocean + global sea ice
NEO-3		+ glaciated land + cool ocean + polar sea ice + sea ice around continent
NEO-4		+ glaciated land + cool ocean + polar sea ice
NEO-5		+ desert land + cool ocean + polar sea ice
NEO-3-280	as NEO-3, but CO_2 = 280 ppm	
NEO-4-280	as NEO-4, but CO_2 = 280 ppm	
NEO-5-280	as NEO-5, but CO_2 = 280 ppm	

Note: CTRL—present-day control experiment; SST—sea surface temperature; SIC—sea ice cover.

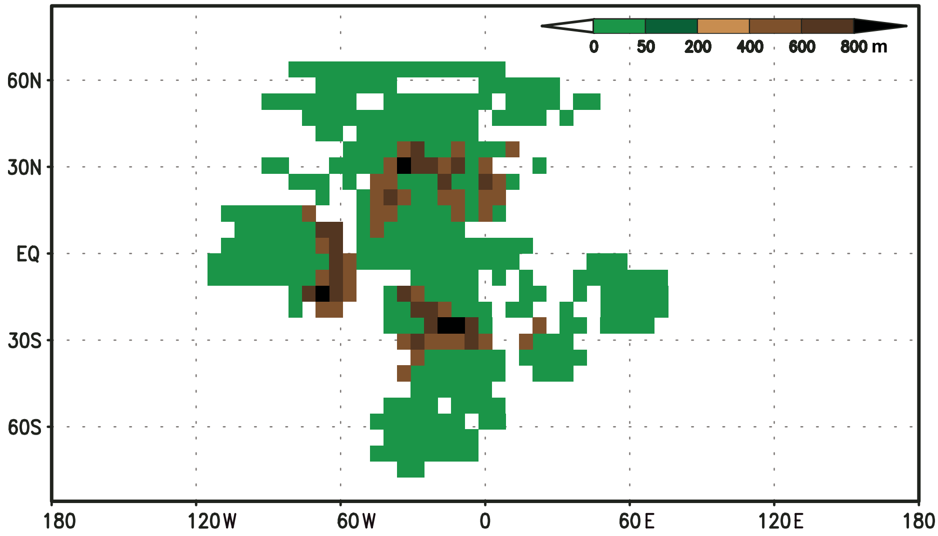

Figure 1. The paleogeography and paleo-orography (in meters) as used for all Neoproterozoic sensitivity experiments.

The described SSTs and MLDs are required for the initialization of the slab ocean model, but ocean properties change according to atmospheric conditions during the model integrations. In order to realistically represent the heat exchange between the atmosphere and the ocean, slab ocean models usually use a flux correction, but it is usually difficult to obtain a good estimate for the flux correction for past climate situations (e.g., Steppuhn et al., 2006). In particular, if the paleogeography differs from the present-day situation, there are some problems in adapting ocean properties. A changed land-sea configuration can make it even impossible to use the present-day flux correction. Present-day ocean grid cells could have been land surface points in the past (and vice versa). In that case, the model can drift toward numerically unstable conditions. The information about ocean properties in Neoproterozoic time is quite poor, and no explicit flux correction was used; i.e., it is not zero but is restored to the previous year's climatology calculated during our model simulations.

Consistent with the setup of the ocean surface conditions, we initialized the sea ice of the thermodynamic sea ice model. The cold ocean scenario is initialized with a global ice cover of a depth of 1 m. For the cool ocean, we allow ice cover where SSTs are below the freezing point and set ice depth to 1 m. Based on the cool ocean, we defined an additional scenario, which has sea ice around the landmasses. For any sea grid point, which is adjacent to land, the initial sea ice depth is set to 1 m and SSTs are corrected to the freezing point.

With the set of boundary conditions, we performed eight Neoproterozoic sensitivity experiments (Table 1), where NEO-1 and NEO-2 refer to a snowball Earth and the others to a slushball Earth. In order to achieve equilibrium for the initially fully glaciated scenarios NEO-1 and NEO-2, we ran the Planet Simulator for 4 k.y. (Fig. 2). The other simulations reacted less sluggish and were integrated over 1 k.y. Figure 2A illustrates that the global temperature immediately drops when starting the simulations, but the model needs to be integrated over several decades more as temperatures decrease further. The sea ice (Fig. 2B) strongly increases in the first one or two simulation years, but it continuously declines after reaching this peak. The stronger the model forcing toward snowball conditions, the longer the sea ice needs to get into equilibrium. This is caused by the development of the ice depth (Fig. 2B), whereas the ice cover is relatively fast in its equilibrium. For our further analysis, we use the averages of the past 10 yr of each simulation.

RESULTS AND DISCUSSION

Global Temperature and Sea Ice Cover

The energy budget is the simplest version of a climate model. As a first approximation, it is possible to calculate the global average temperature from the global energy balance of the incoming (shortwave) and the outgoing (longwave) radiation fluxes:

$$S^{\downarrow} = S^{\uparrow}$$
$$(1-\alpha)S_0 \cdot \pi R_E^2 = \sigma T^4 \cdot 4\pi R_E^2$$
$$T = \sqrt[4]{\frac{(1-\alpha)S_0}{4\sigma}}, \qquad (1)$$

where T (K) is the global average temperature, $S^{\downarrow}$ and $S^{\uparrow}$ (W/m^2) are the solar and terrestrial radiation flux, S_0 (W/m^2) is the solar constant, α (fractional) is the planetary albedo, $\sigma = 5.67 \cdot 10^{-8}$ W m^{-2} K^{-4} is the Stefan-Boltzmann constant, and R_E = 6378 km, the radius of the Earth.

For the present-day planetary albedo of α = 0.3, this simple energy budget results in a global average T = –18.2 °C. The global temperature of CTRL is T = 14.6 °C (Fig. 3). The control simulation is warmer than the theoretical value (ΔT = +32.4 °C) because equation 1 neglects, for example, the greenhouse effect. If we assume a Neoproterozoic snowball Earth with a planetary albedo corresponding to ice (α = 0.7) in equation 1, the resulting T = –70.5 °C. In the coldest Neoproterozoic simulation NEO-2, the global average temperature (T = –68.2 °C) is close to the hypothetical value (ΔT = +2.3 °C). NEO-2 represents a fully glaciated snowball Earth (Fig. 3). In a snowball situation compared to today, processes affecting the atmospheric radiation transfer are almost negligible. This is because much of the incoming shortwave radiation energy is directly reflected at the icy surface; little energy is converted into the longer wave spectrum and trapped in the Earth system by the greenhouse effect, which is most efficient in the longwave spectrum.

The global temperatures of the other Neoproterozoic experiments are also well below the freezing point (Fig. 3). NEO-5 (T = –25.7 °C) is the warmest Neoproterozoic experiment. As compared to CTRL, NEO-2 shows the maximum temperature difference of ΔT = –82.8 °C, but even for the relatively warm simulation NEO-5, the difference (ΔT = –40.3 °C) is significant. According to the cold conditions, all Neoproterozoic model simulations demonstrate a strongly increased degree of glaciation as compared to CTRL (Fig. 3). NEO-1 and NEO-2 maintain their initial full global sea ice cover, but the other experiments do not achieve full glaciation. Their global average SIC is between 66% (NEO-5) and 86% (NEO-3-280).

The snowball experiments NEO-1 and NEO-2 demonstrate much colder global temperatures as compared to other modeling studies (e.g., Donnadieu et al., 2003, 2004b). Achieving a full glaciation, the AGCM LMD represented a global temperature of –40 °C for the Neoproterozoic (Donnadieu et al., 2003), which was consistently found from another study with the EMIC CLIMBER-2 (Donnadieu et al., 2003). As in our study, a Neoproterozoic modeling study with the Planet Simulator, which prescribed full initial glaciation, indicated an average of ~–51 °C (Romanova et al., 2006). This is much warmer than NEO-1 and NEO-2 (cf. Fig. 3), but Romanova et al. (2006) used largely modern

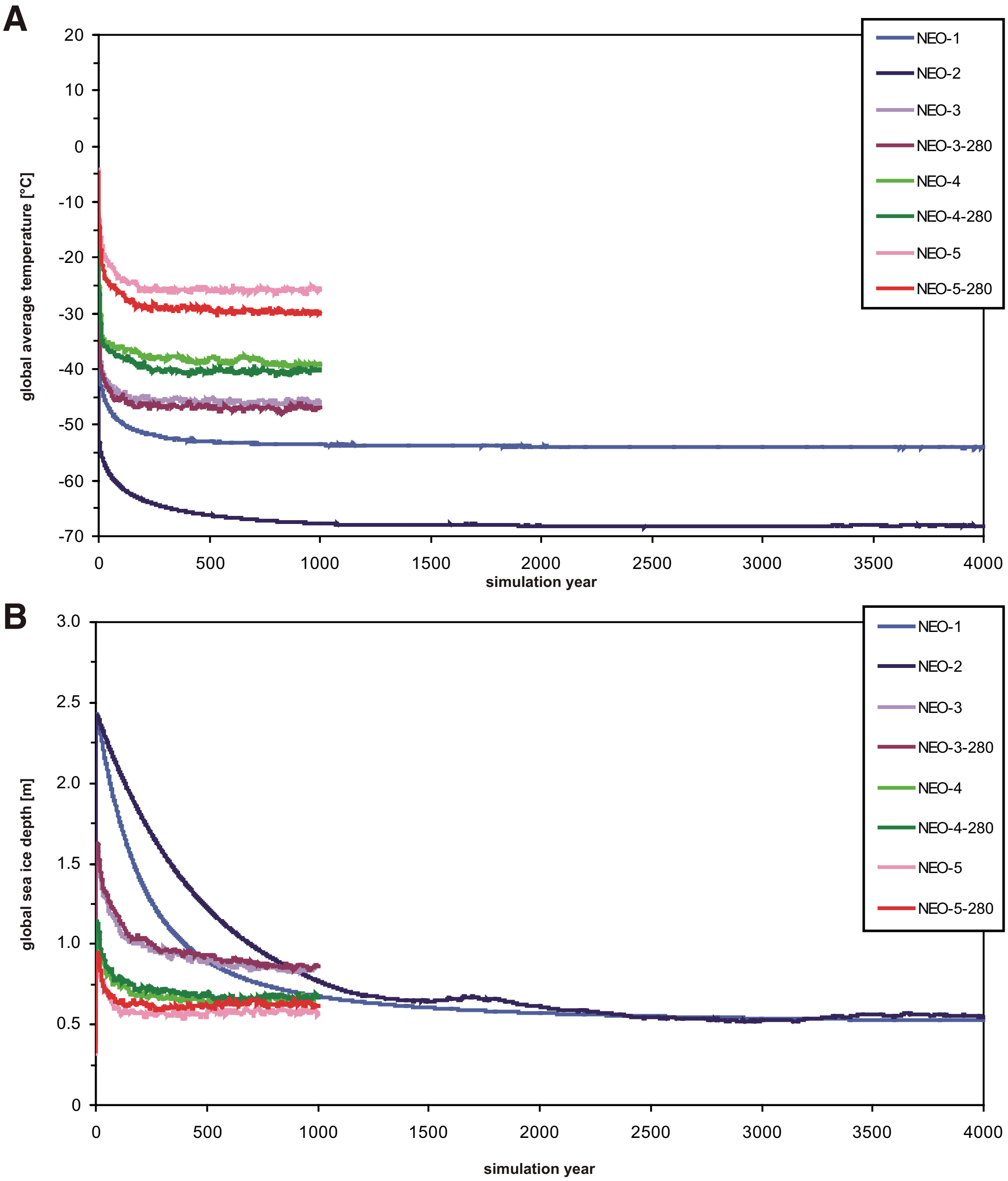

Figure 2. The time series of (A) the global average temperature, and (B) the global sea ice depth of the Neoproterozoic experiments.

boundary conditions, such as the present-day geography and orography. Global average temperatures of NEO-3 to NEO-5-280 are closer to other Neoproterozoic model studies (e.g., Donnadieu et al., 2003; Romanova et al., 2006).

With respect to the degree of sea ice cover, we observe that our snowball scenarios (NEO-1 and NEO-2) maintain their global ice cover. An AGCM coupled to a slab ocean model demonstrated a drift into snowball conditions, while a fully coupled atmosphere-ocean model was not able to produce a globally ice-covered planet (Poulsen et al., 2001; Poulsen, 2003). Our simulations NEO-3 to NEO-5-280 are consistent with the fully coupled model, but not to the slab ocean model of Poulsen et al. (2001). This could indicate that the initialization of the slab ocean is weaker than the forcing of the mixed-layer ocean model of Poulsen et al. (2001). Chandler and Sohl (2000) triggered an AGCM coupled to a slab ocean model with a strong forcing (e.g., low CO_2 and decreased ocean heat transport) toward cold conditions, but they did not achieve global ice cover. This is consistent with our results. With the same EMIC as used in our study, Romanova et al. (2006) showed that their simulations did not get into the snowball state. NEO-3 to NEO-5-280, however, represent a larger extension of sea ice reaching farther toward equatorial latitudes than the study of Romanova et al. (2006), who largely used present-day boundary conditions. We obtain a snowball Earth (NEO-1 and NEO-2) only, if the setup strongly pushes the model into this situation.

Effects of CO_2

Focusing on the effects of CO_2 (NEO-3, NEO-4, NEO-5 minus NEO-3-280, NEO-4-280, NEO-5-280), Figure 3 illustrates that the reduction

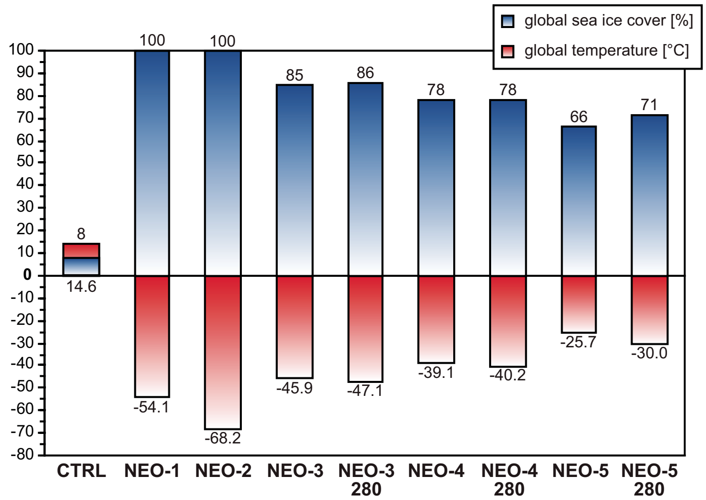

Figure 3. The global average temperatures (red bars) and the global average sea ice cover (blue bars) of the present-day control run and the Neoproterozoic experiments.

of atmospheric CO_2 from 510 ppm to 280 ppm triggers the climate toward cooler conditions. The temperature difference between NEO-5 and NEO-5-280 is –4.3 °C and the SIC differs by +5%. NEO-3 and NEO-3-280 demonstrate smaller differences (ΔT = –1.1 °C, ΔSIC = +1%). On a global scale, experiments NEO-3 versus NEO-3-280 and NEO-4 versus NEO-4-280 represent almost the same climatic conditions. Thus, the stronger the degree of the Earth's glaciation in our Neoproterozoic simulations, the less sensitive is the climate system reaction to variations of greenhouse gas concentrations. This is consistent with the comparison of NEO-2 with the simple energy budget equation (see above). These results might indicate that an escape from an extreme glaciation should require a strongly enhanced CO_2 concentration, which finally should have resulted in a supergreenhouse environment.

Effects of Land Surface Cover

From the comparison of the scenarios with different land surface covers, we observe that, due to their lower albedo, the desert simulations (NEO-1, NEO-4, and NEO-4-280) represent globally warmer conditions and less sea ice than the runs with continental glaciers (NEO-2, NEO-5, and NEO-5-280). This indicates that the formation of continental glaciers via the positive ice-albedo feedback might have contributed significantly to a widespread freezing of the Neoproterozoic Earth. With an atmosphere and ice-sheet model, Hyde et al. (2000) demonstrated that ice-sheet expansion contributed to force the Neoproterozoic climate into a snowball situation. Romanova et al. (2006) found an intense response of the global temperature on variations of the surface albedo. This is consistent with our results, but the cooling (–18 °C) in the Romanova et al. (2006) experiments due to a decreased albedo was more pronounced than in our runs. The weaker response in our study can be explained by different sets of boundary conditions. For example, Romanova et al. (2006) assumed a lower albedo for desert (α = 0.20) as compared to our study (α = 0.35).

Spatial Temperature Patterns and Sea Ice Margins

Figure 4 shows the mean annual temperatures (MAT) and the sea ice margin of the Neoproterozoic simulations. NEO-3-280 and NEO-4-280 are not illustrated because their differences to NEO-3 and NEO-4 are rather small (see above). In NEO-5 (Fig. 4E), polar temperatures are –40 °C in the Southern Hemisphere and –50 °C in the Northern Hemisphere. In equatorial regions, MATs are above 0 °C in NEO-5 and exceed +5 °C over continental areas. Correspondingly, the ocean between 30°S and 15°N remains ice free in NEO-5. The pronounced asymmetry between the Northern Hemisphere and the Southern Hemisphere is due to the usage of the modern orbital parameters and the land-sea configuration. NEO-5-280 represents slightly cooler conditions than NEO-5 (Fig. 4F). The Southern Hemisphere and especially high southern latitude temperatures (<–45 °C) are colder in NEO-5-280 than in NEO-5 (Fig. 5). In the Northern Hemisphere, the temperature distribution and, hence, ice cover are not noticeably affected in NEO-5-280 as compared to NEO-5 (Figs. 4F and 5). The ice margin migrates toward 15° to 20° in both hemispheres, but still large ocean areas are not ice covered in NEO-5-280.

If glaciers cover the entire land surface (NEO-4), the temperatures further decrease as compared to NEO-5 and NEO-5-280 (Figs. 4D and 5). The poles of both hemispheres are colder than –50 °C in NEO-4. For the tropical oceans, temperatures are between –5 °C and 0 °C. Over the equatorial landmasses, temperatures even fall below –15 °C in NEO-4 (Fig. 4D). Despite cool tropical conditions in NEO-4, oceans of latitudes between 15°S and 15°N remain ice free. If the Planet Simulator is initialized with glaciers on land and with sea ice along the coastline (NEO-3), the conditions in the lower latitudes are cold (Fig. 4C), but warmer than in NEO-1 and NEO-2 (Fig. 5). Farther toward middle to high latitudes, the slushball simulation NEO-3 is as cold as the snowball scenarios NEO-1 and NEO-2 (Fig. 5). NEO-3 maintains the initial ice cover adjacent to the landmasses (also in the tropics), but there still remain ice-free ocean areas between 15° in both hemispheres. The freezing line in NEO-3 does not noticeably move farther toward the equator than in NEO-4. The experiments NEO-3 to NEO-5 do not represent a Neoproterozoic snowball Earth, but they do correspond to a slushball Earth.

NEO-1 and NEO-2 demonstrate extremely cold conditions, which are sufficient to maintain the initial global glaciation (Fig. 3). In NEO-1 (Figs. 4A and 5), polar temperatures are comparable with NEO-3 (<–55 °C), but NEO-1 is also rather cold in the low latitudes. Over ocean surfaces the temperatures are below –45 °C, and the desert land surfaces are the warmest regions in NEO-1 with maximums of –25 °C. In NEO-2 (Fig. 4B), the landmasses with glaciers indicate temperatures of generally <–60 °C. Tropical oceans tend to be slightly warmer, but MATs are still <–55 °C. In NEO-2, the low latitudes are even colder than the poles in NEO-1 (Fig. 5). The high latitudes in NEO-2 demonstrate mean annual temperatures of <–70 °C. The strong cooling due to the ice-albedo feedback mechanism in NEO-2 leads to a rather weak meridional temperature gradient (Fig. 5), but there is an extreme seasonal temperature contrast. Monthly average temperatures in low latitudes vary between –55 °C in the warmest month and –60 °C in the coldest month, and at the poles the span ranges from –95 °C to –40 °C (Fig. 5).

Hydrological Cycle and the Fossil Record

The geological record supports a widespread glaciation in the Neoproterozoic (e.g., Hambrey and Harland, 1985; Evans, 2000). In particular, Hoffman et al. (1998b) found evidence for glaciation at around 12°S during the Sturtian

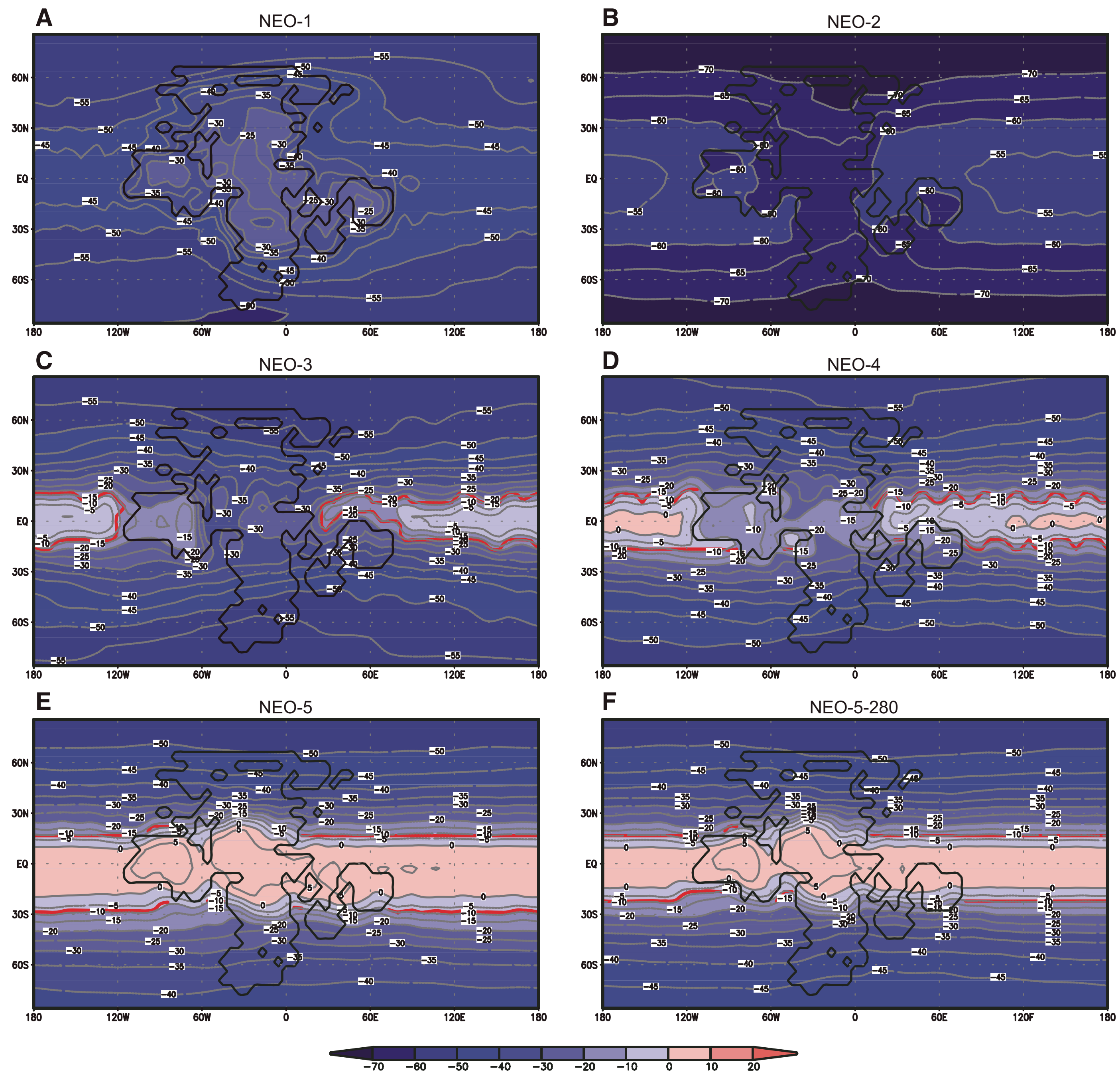

Figure 4. The mean annual temperatures (°C) and sea ice margin (red line) of the Neoproterozoic experiments.

Ice Age. Of course, any indication of glacial deposits (e.g., Evans, 2000; Schrag et al., 2002) fits with our snowball simulations NEO-1 and NEO-2. However, the sea ice margin of the slushball experiments and especially in NEO-3 is relatively close to the equator (~15°). Based on Evans (2000), Schrag et al. (2002, Fig. 2 therein) illustrated reliable occurrences of Neoproterozoic glacial sediments at paleolatitudes between 30° and 40°. Any of our sensitivity simulations NEO-3 to NEO-5-280 supports that sea ice advanced that close into low to middle latitudes. Thick layers from some formations of the Neoproterozoic glacial phase (e.g., Prave, 1999) support that there was an actively working hydrological cycle (e.g., Christie-Blick et al., 1999; Leather et al., 2002). If the evidence contradicts the snowball hypothesis (e.g., Christie-Blick et al., 1999), it agrees fully with the ice-free ocean belt in our slushball scenarios.

The scenario of a more or less completely frozen snowball Earth includes serious consequences for the biosphere. A massive extinction of global proportions should especially affect the light-dependent organisms, such as photoautotrophic prokaryotes and eukaryotes. We note, however, that there is no reliable evidence for a global extinction event. In contrast to an extinction event, microfossil assemblages show basically no corresponding response to

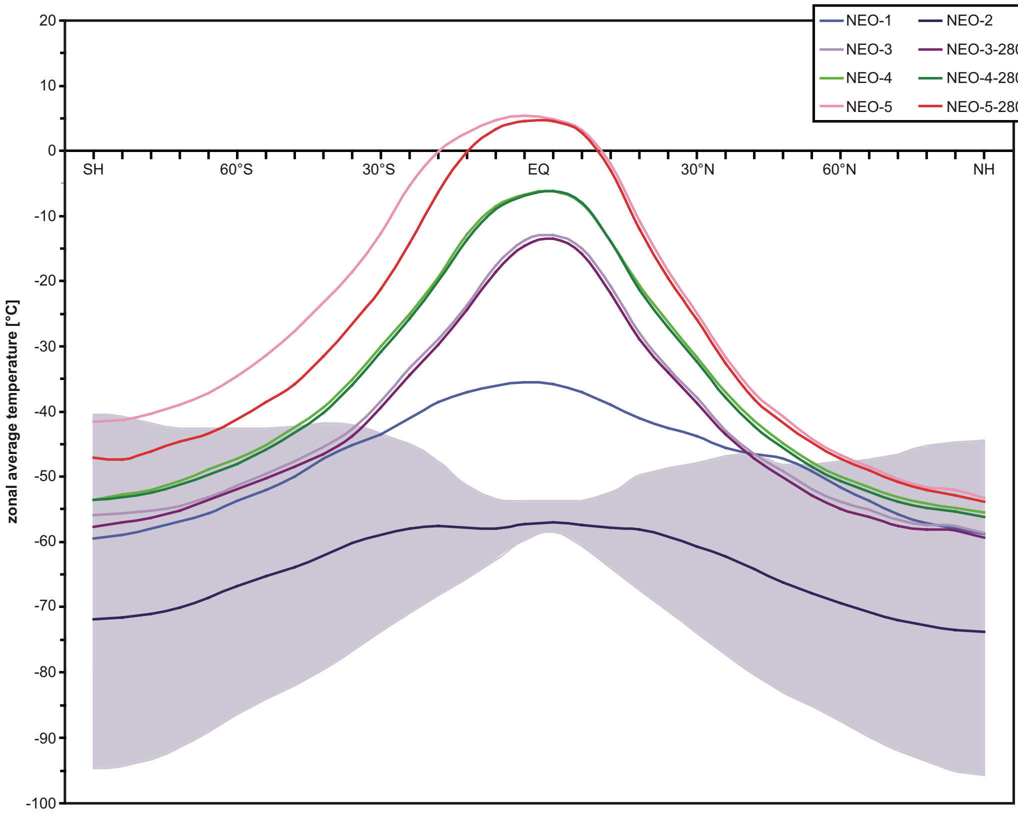

Figure 5. The zonal averages of the mean annual temperatures of the Neoproterozoic experiments. For NEO-2, the seasonal temperature range is indicated with the light blue shading.

the glaciation events (Corsetti et al., 2003). Furthermore, Corsetti et al. (2006) were able to document a slight increase in the diversity of synglacial microbiota, which clearly contradicts the assumed effects of a mass extinction event on these organisms.

Low Temperatures and CO_2

Carbon dioxide has a sublimation and/or deposition point of T = –78.45 °C under standard atmospheric pressure (p_0 = 101.325 kPa), where CO_2 changes between the gaseous and solid phase (so-called dry ice). Because of thermodynamic reasons, liquid CO_2 does not naturally occur under surface conditions. Minimum temperatures in NEO-2 fall much below the sublimation and/or deposition point of CO_2 (Fig. 5). Other model experiments demonstrated that temperatures can be extremely low for snowball situations; Donnadieu et al. (2003) emphasized that polar temperatures reach down to –110 °C and tropics never exceed –25 °C. As compared to results of Donnadieu et al. (2003), NEO-1 and NEO-2 are not so cold in polar regions and not so warm at the equator (Fig. 5), but the trends are comparable. If the Neoproterozoic was really so cold in the high latitudes such as represented in NEO-2 or even colder (Donnadieu et al., 2003), the question arises whether it is possible and reasonable that CO_2 could have changed from its gas into its solid phase in winter. CO_2 in its solid phase would mean that the greenhouse effect of increasing concentrations of CO_2 could be nullified. From that point of view, the Neoproterozoic Earth could be an analogue to other planets (or vice versa) such as Mars, for which recent results indicate the occurrence of frozen carbon dioxide (e.g., Bibring et al., 2004; Colaprete et al., 2005). However, solid CO_2 should have only been possible during winter, because summer temperatures are well above the sublimation and/or deposition point of CO_2. Moreover, phase changes of gases depend not only on temperature but also on pressure. Thus, carbon dioxide does not necessarily go into its solid phase, even though temperatures are low. In addition, it usually needs some degree of saturation, for example, with water vapor, to produce rainfall and/or snowfall. In order to have fallout of carbon dioxide, it would also need a certain concentration. If the escape mechanism out of a snowball situation is a strongly enhanced CO_2 concentration, this would mean that the possibility of the occurrence of carbon dioxide ice increases. Conditions on Mars demonstrate that the occurrence of water and carbon dioxide ice is generally possible (e.g., Bibring et al., 2004; Colaprete et al., 2005). However, the escape out of a snowball Earth becomes really difficult if atmospheric carbon dioxide and, therefore, the greenhouse effect are reduced due to phase changes of CO_2. In addition, photoautotrophic organisms are dependent on gaseous or dissolved CO_2 in order to carry out the photosynthesis via the carbon-fixating enzyme ribulose-1,5-bisphosphate-carboxylase-oxygenase (RuBISCO; e.g., Lawlor, 2001, and references therein). Because of the inability of photosynthetic organisms to fixate CO_2 as dry ice, a change of CO_2 from the gaseous phase into the solid state would have terminated the photosynthesis, and therefore inevitably led to a mass extinction among the photosynthetic active primary producers. As mentioned above, no extinction can be observed. Nevertheless, we suggest that the possibility of freezing carbon dioxide could be tested with atmospheric models including a sophisticated scheme for atmospheric chemistry.

Weak Points of the Model Runs

As none of the simulations without initial global glaciation (i.e., NEO-3 to NEO-5-280) can step into the snowball state, we conclude that the slushball Earth is the more realistic scenario. However, our sensitivity experiments include some assumptions and simplifications (as well as the EMIC model concept). For example, we use no explicit flux correction and do not consider that ocean currents east of Rodinia should transport warmer water masses toward middle and high latitudes, while western ones should bring cooler water into low latitudes. This might modify but should not fundamentally change the climate response in our simulations. Romanova et al. (2006) demonstrated that climatic differences due to variations of the ocean heat transport are small. Also, our sensitivity studies might include some uncertainties with respect to the paleogeography and paleo-orography (e.g., Meert and Powell, 2001; Torsvik, 2003). However, if a supercontinent configuration (such as realized in our simulations) is a major prerequisite for initiating snowball conditions, this should modify but not substantially change our results. Further model experiments could focus on the sensitivity with respect to the paleogeography and paleo-orography.

SUMMARY

With the Earth system model of intermediate complexity, Planet Simulator, we performed a series of sensitivity experiments for the Neoproterozoic, which considered a cool versus a cold ocean, a desert versus a glacier land surface, and a higher versus a lower atmospheric concentration of carbon dioxide. The results of the sensitivity simulations can be summarized as follows.

NEO-1 and NEO-2 represent a Neoproterozoic snowball Earth with extremely cold conditions. This is not surprising because the initial glaciation of both runs provides a strong model forcing via the ice-albedo feedback mechanism. Global average temperatures of NEO-1 and NEO-2 are close to the hypothetical value, which results from the energy balance model in its simplest version (i.e., $S^{\downarrow} = S^{\uparrow}$). Depending on the latitude, minimum temperatures in the extremely cold experiment NEO-2 are close to or even fall below the sublimation and/or deposition point of CO_2 ($T = -78.45$ °C).

If we use a less extreme model forcing, the simulations NEO-3 to NEO-5 achieve moderately cold slushball conditions with a more or less distinct ice-free ocean belt in lower latitudes. The climate response to variations of atmospheric $p$$CO_2$ is minor in our Neoproterozoic simulations. This might indicate that the exit out of a frozen situation via the greenhouse effect should require a strongly enhanced CO_2 concentration, which should finally result in an extreme greenhouse world.

With respect to the land surface cover, our sensitivity experiments demonstrate that those runs with the deserted land surface are relatively warmer than those with glaciers. The glaciation of the continents contributes to an increased degree of glaciation of the oceans. Hence, it provides a mechanism to explain a rather strong Neoproterozoic glaciation.

Our sensitivity experiments with the Planet Simulator support the Neoproterozoic slushball Earth more than the snowball Earth, but the specified sensitivity experiments include some simplifications (e.g., the ocean setup) that can limit the reliability of the model results. Since the escape mechanism out of a snowball Earth is an enhanced greenhouse effect, further model studies should address the question of how much CO_2 is necessary to melt a snowball Earth.

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REFERENCES CITED

Allen, P.A., and Hoffman, P.F., 2005, Extreme winds and waves in the aftermath of a Neoproterozoic glaciation: Nature, v. 433, p. 123–127, doi: 10.1038/nature03176.

Berner, R.A., 1997, The rise of plants and their effect on weathering and atmospheric CO_2: Science, v. 276, p. 544–546, doi: 10.1126/science.276.5312.544.

Bibring, J.-P., Langevin, Y., Poulet, F., Gendrin, A., Gondet, B., Berthe, M., Soufflot, A., Drossart, P., Combes, M., Bellucci, G., Moroz, V., Mangold, N., Schmitt, B., and the OMEGA team, 2004, Perennial water ice identified in the south polar cap of Mars: Nature, v. 428, p. 627–630.

Bice, K.L., Scotese, C.R., Seidov, D., and Barron, E.J., 2000, Quantifying the role of geographic change in Cenozoic ocean heat transport using uncoupled atmosphere and ocean models: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 161, p. 295–310, doi: 10.1016/S0031-0182(00)00072-9.

Brady, E.C., DeConto, R.M., and Thompson, S.L., 1998, Deep water formation and poleward ocean heat transport in the warm climate extreme of the Cretaceous (80 Ma): Geophysical Research Letters, v. 25, no. 22, p. 4205–4208, doi: 10.1029/1998GL900072.

Chandler, M.A., and Sohl, L.E., 2000, Climate forcings and the initiation of low-latitude ice sheets during the Noproterozoic Varanger glacial interval: Journal of Geophysical Research, v. 105, p. 20,737–20,756, doi: 10.1029/2000JD900221.

Christie-Blick, N., Sohl, L.E., and Kennedy, M.J., 1999, Considering a Neoproterozoic Snowball Earth: Comment: Science, v. 284, p. 1087, doi: 10.1126/science.284.5417.1087a.

Claussen, M., 1994, On coupling global biome models with climate models: Climate Research, v. 4, p. 203–221.

Claussen, M., Mysak, L., Weaver, A., Crucifix, M., Fichefet, T., Loutre, M.-F., Weber, S., Alcamo, J., Alexeev, V., Berger, A., Calov, R., Ganopolski, A., Goosse, H., Lohmann, G., Lunkeit, F., Mokhov, I., Petoukhov, V., Stone, P., and Wang, Z., 2002, Earth system models of intermediate complexity: Closing the gap in the spectrum of climate system: Climate Dynamics, v. 18, p. 579–586, doi: 10.1007/s00382-001-0200-1.

Cohen-Solal, E., and Le Treut, H., 1997, Role of the oceanic heat transport in climate dynamics. A sensitivity with an atmospheric general circulation model ocean heat transport GCM sensitivity studies: Tellus, v. 49A, p. 371–387.

Colaprete, A., Barnes, J.R., Haberle, R.M., Hollingsworth, J.L., Kieffer, H.H., and Titus, T.N., 2005, Albedo of the south pole on Mars determined by topographic forcing of atmosphere dynamics: Nature, v. 435, p. 184–188, doi: 10.1038/nature03561.

Corsetti, F.A., Awramik, S.M., and Pierce, D., 2003, A complex microbiota from snowball Earth times: Microfossils from the Neoproterozoic Kingston Peak Formation, Death Valley, USA: National Academy of Sciences Proceedings, v. 100, p. 4399–4404, doi: 10.1073/pnas.0730560100.

Corsetti, F.A., Olcott, A.N., and Bakermans, C., 2006, The biotic response to Neoproterozoic snowball Earth: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 232, p. 114–130, doi: 10.1016/j.palaeo.2005.10.030.

Covey, C., and Thompson, S.L., 1989, Testing the effects of oceanic heat transport on climate: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 75, p. 331–341, doi: 10.1016/0031-0182(89)90193-4.

Cubasch, U., Meehl, G.A., Boer, G.J., Stouffer, R.J., Dix, M., Noda, A., Senior, C.A., Raper, S., and Yap, K.S., 2001, Projections of future climate change, *in* Houghton, J.T., et al., eds., Climate change 2001: The scientific basis. Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change: Cambridge, Cambridge University Press, p. 525–582.

Dalziel, I.W.D., 1991, Pacific margins of Laurentia and East Antarctica–Australia as a conjugate rift pair: Evidence and implications for an Eocambrian supercontinent: Geology, v. 19, p. 598–601, doi: 10.1130/0091-7613(1991)019<0598:PMOLAE>2.3.CO;2.

Donnadieu, Y., Fluteau, F., Ramstein, G., Ritz, C., and Besse, J., 2003, Is there a conflct between the Neoproterozoic glacial deposits and the snowball Earth interpretation: An improved understanding with numerical modelling: Earth and Planetary Science Letters, v. 208, p. 101–112, doi: 10.1016/S0012-821X(02)01152-4.

Donnadieu, Y., Godderis, Y., Ramstein, G., Nedelec, A., and Meert, J., 2004a, A 'snowball Earth' climate triggered by continental break-up through changes in runoff: Nature, v. 428, p. 303–306, doi: 10.1038/nature02408.

Donnadieu, Y., Ramstein, G., Fluteau, F., Roche, D., and Ganopolski, A., 2004b, The impact of atmospheric and oceanic heat transports on the sea-ice-albedo instability during the Neoproterozoic: Climate Dynamics, v. 22, p. 293–306, doi: 10.1007/s00382-003-0378-5.

Evans, D.A., 2000, Stratigraphic, geochronological, and paleomagnetic constraints upon the Neoproterozoic climatic paradox: American Journal of Science, v. 300, p. 347–433, doi: 10.2475/ajs.300.5.347.

Evans, D.A.D., Li, Z.X., Kirschvink, J.L., and Wingate, M., T.D., 2000, A high-quality mid-Neoproterozoic paleomagnetic pole from South China, with implications for ice ages and the break-up configuration of Rodinia: Precambrian Research, v. 100, p. 313–334.

Fairchild, I.J., and Hambrey, M.J., 1984, The Vendian succession of northeastern Spitsbergen: Petrogenesis of a dolomite-tillite association: Precambrian Research, v. 26, p. 111–167, doi: 10.1016/0301-9268(84)90042-1.

Fraedrich, K., Kirk, E., and Lunkeit, F., 1998, PUMA: Portable University Model of the atmosphere: DKRZ Technical Report 16, 24 p.

Fraedrich, K., Jansen, H., Kirk, E., Luksch, U., and Lunkeit, F., 2005a, The planet simulator: Towards a user friendly model: Meteorologische Zeitschrift, v. 14, p. 299–304, doi: 10.1127/0941-2948/2005/0043.

Fraedrich, K., Jansen, H., Kirk, E., and Lunkeit, F., 2005b, The planet simulator: Green planet and desert world:

Meteorologische Zeitschrift, v. 14, p. 305–314, doi: 10.1127/0941-2948/2005/0044.

Frisius, T., Lunkeit, F., Fraedrich, K., and James, I.A., 1998, Storm-track organization and variability in a simplified atmospheric global circulation model: Royal Meteorological Society Quarterly Journal, v. 124, p. 1019–1043, doi: 10.1002/qj.49712454802.

Gough, D.O., 1981, Solar interior structure and luminosity variations: Solar Physics, v. 74, no. 1, p. 21–34, doi: 10.1007/BF00151270.

Gradstein, F., Ogg, J., and Smith, A., 2004, A geologic time scale 2004: Cambridge, Cambridge University Press, 589 p.

Grosfeld, K., Lohmann, G., Rimbu, N., Fraedrich, K., and Lunkeit, F., 2007, Atmospheric multidecadal variations in the North Atlantic realm: Proxy data, observations, and atmospheric circulation model studies: Climate of the Past, v. 3, p. 39–50.

Hagemann, S., Botzet, M., Dümenil, L., and Machenhauer, B., 1999, Derivation of global GCM boundary conditions from 1 km land use satellite data: Max-Planck-Institut für Meteorologie Report no. 289, 34 p.

Hambrey, M.J., and Harland, W.B., 1985, The late Proterozoic glacial era: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 51, p. 255–272, doi: 10.1016/0031-0182(85)90088-4.

Hoffman, P.F., 1991, Did the breakout of Laurentia turn Gondwanaland inside-out?: Science, v. 252, p. 1409–1412, doi: 10.1126/science.252.5011.1409.

Hoffman, P.F., Kaufman, A.J., and Halverson, G.P., 1998a, Comings and goings of global glaciation on a Neoproterozoic tropical platform in Namibia: GSA Today, v. 8, no. 5, p. 1–9.

Hoffman, P.F., Kaufman, A.J., Halverson, G.P., and Schrag, D.P., 1998b, A Neoproterozoic Snowball Earth: Science, v. 281, p. 1342–1346, doi: 10.1126/science.281.5381.1342.

Hoffman, P.F., and Schrag, D.P., 1999, Considering a Neoproterozoic Snowball Earth: Reply: Science, v. 284, p. 1087.

Hyde, W.T., Crowley, T.J., Baum, S.K., and Peltier, W.R., 2000, Neoproterozoic 'snowball Earth' simulations with a coupled climate/ice-sheet model: Nature, v. 405, p. 425–429, doi: 10.1038/35013005.

Junge, M., Blender, R., Fraedrich, K., Gayler, V., Luksch, U., and Lunkeit, F., 2005, A world without Greenland: Impacts on the Northern Hemisphere winter circulation in low- and high-resolution models: Climate Dynamics, v. 24, p. 297–307, doi: 10.1007/s00382-004-0501-2.

Kaufman, A.J., Knoll, A.H., and Narbonne, G.M., 1997, Isotopes, ice ages, and terminal Proterozoic earth history: National Academy of Sciences Proceedings, v. 94, p. 6600–6605, doi: 10.1073/pnas.94.13.6600.

Kennedy, M.J., Runnegar, B., Prave, A.R., Hoffmann, K.H., and Arthur, M.A., 1998, Two or four Neoproterozoic glaciations: Geology, v. 26, p. 1059–1063, doi: 10.1130/0091-7613(1998)026<1059:TOFNG>2.3.CO;2.

Kirschvink, J.L., 1992, Late Proterozoic low-latitude glaciation, *in* Schopf, J.W., ed., The Proterozoic biosphere: New York, Columbia University Press, p. 51–52.

Kleidon, A., Fraedrich, K., and Low, C., 2007, Multiple steady-states in the terrestrial atmosphere-biosphere system: A result of a discrete vegetation classification?: Biogeosciences Discussions, v. 4, p. 687–705.

Knoll, A.H., Hayes, J.M., Kaufman, A.J., Sweet, K., and Lambert, I.B., 1986, Secular variation in carbon isotope ratios from the upper Proterozoic succession of Svalbard and east Greenland: Nature, v. 321, p. 832–839, doi: 10.1038/321832a0.

Kürschner, W.M., van der Burgh, J., Visscher, H., and Dilcher, D., 1996, Oak leaves as biosensor of late Neogene and early Pleistocene paleoatmospheric CO_2 concentrations: Marine Micropaleontology, v. 27, p. 299–312, doi: 10.1016/0377-8398(95)00067-4.

Lawlor, D.W., 2001, Photosynthesis (third edition): Oxford, UK, Longman, 386 p.

Leather, J., Allen, P.A., Brasier, M.D., and Cozzi, A., 2002, Neoproterozoic snowball Earth under scrutiny: Evidence from the Fiq glaciation of Oman: Geology, v. 30, p. 891–894, doi: 10.1130/0091-7613(2002)030<0891:NSEUSE>2.0.CO;2.

MacFadden, B.J., 2005, Terrestrial mammalian herbivore response to declining levels of atmospheric CO_2 during the Cenozoic: Evidence from North American fossil horses (Family Equidae), *in* Ehleringer, J.R., et al., eds., A history of atmospheric CO_2 and its effects on plants, animals, and ecosystems. Ecological Studies, v. 177, p. 273–292.

Maier-Reimer, E., Mikolajewicz, U., and Crowley, T.J., 1990, Ocean general circulation model sensitivity experiment with an open Central American Isthmus: Palaeoceanography, v. 5, p. 349–366.

Meert, J.G., and Powell, C.M., 2001, Assembly and break-up of Rodinia: Introduction to the special volume: Precambrian Research, v. 110, p. 1–8, doi: 10.1016/S0301-9268(01)00177-2.

Meert, J.G., Van der Voo, R., and Ayub, S., 1995, Paleomagnetic investigation of the Neoproterozoic Gagwe lavas and Mbozi complex, Tanzania and the assembly of Gondwana: Precambrian Research, v. 74, p. 225–244, doi: 10.1016/0301-9268(95)00012-T.

Micheels, A., Bruch, A.A., Mosbrugger, V., Uhl, D., and Utescher, T., 2006, Simulating climatic effects of palaeovegetation changes in the late Miocene using different climate models, *in* 7th European Palaeobotany-Palynology Conference, Prague, Program and Abstracts: Prague, Czech Republic, National Museum, p. 93.

Micheels, A., Bruch, A.A., Uhl, D., Utescher, T., and Mosbrugger, V., 2007, A late Miocene climate model simulation with ECHAM4/ML and its quantitative validation with terrestrial proxy data: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 253, p. 267–286, doi: 10.1016/j.palaeo.2007.03.042.

Mikolajewicz, U., and Crowley, T.J., 1997, Response of a coupled ocean/energy balance model to restricted flow through the Central American isthmus: Palaeoceanography, v. 12, p. 429–441, doi: 10.1029/96PA03542.

Mikolajewicz, U., and Voss, R., 2000, The role of the individual air-sea flux components in CO_2-induced changes of the ocean's circulation and climate: Climate Dynamics, v. 16, p. 627–642, doi: 10.1007/s003820000066.

Moores, E.M., 1991, Southwest U.S.–East Antarctica (SWEAT) connection: A hypothesis: Geology, v. 19, p. 425–428, doi: 10.1130/0091-7613(1991)019<0425:SUSEAS>2.3.CO;2.

Olcott, A.N., Sessions, A.L., Corsetti, F.A., Kaufman, A.J., and de Oliviera, T.F., 2005, Biomarker evidence for photosynthesis during Neoproterozoic glaciation: Science, v. 310, p. 471–474, doi: 10.1126/science.1115769.

Pagani, M., Zachos, J.C., Freeman, K.H., Tipple, B., and Bohaty, S., 2005, Marked decline in atmospheric carbon dioxide concentrations during the Paleogene: Science, v. 309, p. 600–603, doi: 10.1126/science.1110063.

Petoukhov, V., Ganopolski, A., Brovkin, V., Claussen, M., Eliseev, A., Kubatzki, C., and Rahmstof, S., 2000, CLIMBER-2: A climate system model of intermediate complexity. Part I: Model description and performance for present climate: Climate Dynamics, v. 16, p. 1–17, doi: 10.1007/PL00007919.

Pierrehumbert, R.T., 2004, High levels of atmospheric carbon dioxide necessary for the termination of global glaciation: Nature, v. 429, p. 646–649, doi: 10.1038/nature02640.

Poulsen, C.J., 2003, Absence of a runaway ice-albedo feedback in the Neoproterozoic: Geology, v. 31, p. 473–476, doi: 10.1130/0091-7613(2003)031<0473:AOARIF>2.0.CO;2.

Poulsen, C.J., and Jacob, R., 2004, Factors that inhibit Snowball Earth simulation: Paleoceanography, v. 19, PA4021, doi: 10.1029/2004PA001056

Poulsen, C.J., Pierrehumbert, R.T., and Jacob, R.L., 2001, Impact of ocean dynamics on the simulation of the Neoproterozoic "snowball Earth": Geophysical Research Letters, v. 28, p. 1575–1578, doi: 10.1029/2000GL012058.

Poulsen, C.J., Jacob, R.L., Pierrehumbert, R.T., and Huynh, T.T., 2002, Testing palaeographic controls on a Neoproterozoic snowball Earth: Geophysical Research Letters, v. 29, p. 10–1-10–4.

Powell, C.M., Li, Z.X., McElhinny, M.W., Meert, J.G., and Park, J.K., 1993, Paleomagnetic constraints on timing of the Neoproterozoic breakup of Rodinia and the Cambrian formation of Gondwana: Geology, v. 21, p. 889–893, doi: 10.1130/0091-7613(1993)021<0889:PCOTOT>2.3.CO;2.

Prave, A.R., 1999, Two diamictites, two cap carbonates, two $\delta^{13}C$ excursions, two rifts: The Neoproterozoic Kingston Peak Formation, Death Valley, California: Geology, v. 27, p. 339–342, doi: 10.1130/0091-7613(1999)027<0339:TDTCCT>2.3.CO;2.

Ridgwell, A.J., Kennedy, M.J., and Caldeira, K., 2003, Carbonate deposition, climate stability, and Neoproterozoic ice ages: Science, v. 302, p. 859–862, doi: 10.1126/science.1088342.

Rind, D., and Chandler, M., 1991, Increased ocean heat transports and warmer climate: Journal of Geophysical Research, v. 96, p. 7437–7461.

Roeckner, E., Bäuml, G., Bonaventura, L., Brokopf, R., Esch, M., Giorgetta, M., Hagemann, S., Kirchner, I., Kornblueh, L., Manzini, E., Rhodin, A., Schlese, U., Schulzweida, U., and Tompkins, A., 2003, The atmospheric general circulation model ECHAM5—Model description: Max-Planck-Institut für Meteorologie Report no. 349, 140 p.

Roeckner, E., Brokopf, R., Esch, M., Giorgetta, M., Hagemann, S., Kornblueh, L., Manzini, E., Schlese, U., and Schulzweida, U., 2004, The atmospheric general circulation model ECHAM5—Sensitivity of simulated climate to horizontal and vertical resolution: Max-Planck-Institut für Meteorologie Report no. 354, 64 p.

Romanova, V., Lohmann, G., and Grosfeld, K., 2006, Effect of land albedo, CO_2, orography, and oceanic heat transport on extreme climates: Climate of the Past, v. 2, p. 31–42.

Runnegar, B., 2000, Loophole for snowball Earth: Nature, v. 405, p. 403–404.

Saylor, B.Z., Kaufman, A.J., Grotzinger, J.P., and Urban, F., 1998, A composite reference section for the terminal Proterozoic strata of southern Namibia: Journal of Sedimentary Research, v. 68, no. 6, p. 1223–1235.

Schmidt, P.W., and Williams, G.E., 1995, The Neoproterozoic climatic paradox: Equatorial palaeolatitude for Marinoan glaciation near sea level in South Australia: Earth and Planetary Science Letters, v. 134, p. 107–124, doi: 10.1016/0012-821X(95)00106-M.

Schrag, D.P., Berner, R.A., Hoffman, P.F., and Halverson, G.P., 2002, On the initiation of a snowball Earth: Geochemistry, Geophysics, Geosystems, v. 3, no. 6, doi: 10.1029/2001GC000219.

Sohl, L.E., Christie-Blick, N., and Kent, D.V., 1999, Paleomagnetic polarity reversals in Marinoan (ca. 600 Ma) glacial deposits of Australia: Implications for the duration of low-latitude glaciation in Neoproterozoic time: Geological Society of America Bulletin, v. 111, p. 1120–1139, doi: 10.1130/0016-7606(1999)111<1120:PPRIMC>2.3.CO;2.

Steppuhn, A., Micheels, A., Geiger, G., and Mosbrugger, V., 2006, Reconstructing the late Miocene climate and oceanic heat flux using the AGCM ECHAM4 coupled to a mixed-layer ocean model with adjusted flux correction: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 238, p. 399–423, doi: 10.1016/j.palaeo.2006.03.037.

Torsvik, T.H., 2003, The Rodinia jigsaw puzzle: Science, v. 300, p. 1379–1381, doi: 10.1126/science.1083469.

Torsvik, T.H., Smethurst, M.A., Meert, J.G., Van der Voo, R., McKerrow, W.S., Brasier, M.D., Sturt, B.A., and Walderhaug, H.J., 1996, Continental break-up and collision in the Neoproterozoic and Palaeozoic—A tale of Baltica and Laurentia: Earth-Science Reviews, v. 40, p. 229–258, doi: 10.1016/0012-8252(96)00008-6.

Torsvik, T.H., Carter, L.M., Ashwal, L.D., Bhushan, S.K., Pandit, M.K., and Jamtveit, B., 2001, Rodinia refined or obscured: Paleomagnetism of the Malani Igneous Suite (NW India): Precambrian Research, v. 108, p. 319–333, doi: 10.1016/S0301-9268(01)00139-5.

Weil, A.B., Van der Voo, R., Niocaill, C.M., and Meert, J.G., 1998, The Proterozoic supercontinent Rodinia: Paleomagnetically derived reconstructions for 1100 to 800 Ma: Earth and Planetary Science Letters, v. 154, p. 13–24, doi: 10.1016/S0012-821X(97)00127-1.

Young, G.M., and Gostin, V.A., 1989, An exceptionally thick upper Proterozoic (Sturtian) glacial succession in the Mount Painter area, South Australia: Geological Society of America Bulletin, v. 101, p. 834–845, doi: 10.1130/0016-7606(1989)101<0834:AETUPS>2.3.CO;2.

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