Reconstructing Neoproterozoic palaeoclimates using a combined data/modelling approach

L. E. SOHL & M. A. CHANDLER

Center for Climate Systems Research at Columbia University, 2880 Broadway, New York, New York 10025, USA. (e-mail: les14@columbia.edu; tel: (212) 678-5550/fax: (212) 678-5648)

Abstract: Climate reconstructions of the Neoproterozoic Era (1000–542 Ma) face special challenges because many proxies used to constrain younger palaeoclimates are not available/applicable in Precambrian time. Given the few available proxies, deep time climate simulations are best viewed as a means to address more fundamental questions about the nature of climate change and to address disparities in data interpretation by examining phenomena from a process-related perspective. The Global Climate Model (GCM) simulations presented here were aimed at determining what combination of forcings might have permitted the initiation of low-to mid-latitude continental ice-sheets during the Sturtian glacial interval, c. 750 Ma. However, despite the formation of extensive extratropical ice cover, tropical regions in these experiments remain too warm for the initiation of large ice-sheets. The enhanced precipitation along the leading edge of icy regions suggests that the addition of topographic relief and dynamic ice flow could make ice-sheets viable into subtropical regions. However, these simulations suggest that ‘hard’ snowball Earth solutions are only likely for much earlier intervals in Earth history, and are certainly not viable in combination with large accumulations of greenhouse gases.

The Neoproterozoic Era (1000 to 542 Ma) is a remarkable interval in Earth history from the standpoint both of climatic change and biological innovation. On the climate front, we have evidence for tremendous swings between conditions for which we have as yet no adequate explanation. Parts of the Neoproterozoic Era are marked by an equable climate that is perhaps warmer than we would have expected, given a less-luminous Sun (a smaller-scale version of the ‘faint young Sun’ paradox afflicting early Earth palaeoclimate; see e.g. Tajika 2003; Chumakov 2004). However, we also have two of the most severe glaciations in Earth history, the so-called ‘snowball Earth’ events, in which continental-scale ice-sheets were able to exist, at sea level, within 20° of the equator (Park 1997; Schmidt & Williams 1995; Sohl et al. 1999). On the biological front, we have the first appearance of macroscopic multicellular organisms (the Mistaken Point and Ediacaran faunal assemblages) toward the end of the Neoproterozoic Era (Narbonne 2005), a prelude to the great expansion in biodiversity known as the Cambrian explosion. Perhaps more importantly, there is the evolution of the first shelly marine animals such as Cloudina (e.g. Hofmann & Mountjoy 2001), whose ability to create their own shells may have forever altered ocean chemistry and helped established the marine carbon cycle as we now know it (Ridgwell et al. 2003; Bartley & Kah 2004).

As the macroscopic fossil record appears almost entirely limited to a time some 60 million years after the snowball Earth glaciations occurred (Narbonne 2005), there has been considerable speculation about the extent to which the extreme climate changes could have driven mass extinctions and/or biological innovations. Naturally, we would like to explore these potential impacts by better quantifying characteristics of Neoproterozoic climate, such as surface air temperature, snow cover, and sea-ice extent through the use of proxies and computer models. However, reconstructing the palaeoclimate of an interval so far in Earth’s past presents special challenges not faced by geoscientists and climate modellers who work on younger time intervals. The palaeoclimate proxies available to constrain climate forcings, such as greenhouse gas levels and ocean circulation, are not as quantitative or readily interpreted as we might like, while boundary conditions for Global Climate Models, such as continent/ocean distribution, topography, ground cover, and even solar luminosity, are not well constrained.

The purpose of Neoproterozoic climate modelling is therefore largely limited to addressing more fundamental questions about the nature of climate change and its potential relationship to biological and ecosystem changes. Since our understanding of Neoproterozoic environments is limited compared to those of the Phanerozoic Era, the need to address the big picture questions first is not a handicap. In fact, Neoproterozoic climate change has the potential to shed light on the large-scale forcings involved in creating the broadest range of natural...
climate variability that the Earth has ever displayed. It is also clear that with such extreme environmental ranges, many first-order physical processes – such as moisture fluxes, dynamic transports of mass and momentum, detailed parameterizations of hydrologic processes, and to the extent possible, ocean processes – are critical to an accurate portrayal and understanding of the climates of this time. The Neoproterozoic Era thus presents us with astounding, non-hypothetical climate scenarios that both captivate geoscientists and modellers, and require a high degree of communication and cooperation amongst a wide range of research specializations.

One cautionary note must be kept in mind when modelling deep time palaeoclimates such as the snowball Earth intervals, for which boundary constraints are scarce: it is very easy to fall into the trap of setting up simulations that explore interesting theoretical considerations, but do not address any questions regarding actual geological events in a manner that is truly useful for interpreting the past. In planning and evaluating deep time palaeoclimate simulations, climate modellers need to be aware not only of their own model limitations, but also the need to be careful about trying too hard to achieve a specific goal rather than seeking to understand the results that models deliver based on available geological inputs. Our approach to climate modelling has focused on creating an ensemble of experiments that define a range of possible climate forcings and boundary conditions, and allowing the GCM to tell us how the forcings, feedbacks and other processes interact to yield various climate solutions.

Overview of climate trends and chronology of events for the Neoproterozoic Era

The Neoproterozoic Era can be divided into three broad climate intervals, roughly equivalent to chronostratigraphic divisions and defined by the inferred presence of little or no glacial deposits in the geological record (the Tonian and Ediacaran Periods) vs. the presence of widely distributed glacial deposits that suggest that an ice age was in progress (the Cryogenian Period) (see Fig. 1). Firm age constraints are hard to come by for any climatically significant changes reflected in sedimentation patterns; indeed, this is a general problem for the entire era. For the sake of convenience, we define the boundaries of the glacial intervals by the approximate age ranges of the glacial deposits, although global cooling trends probably began at an earlier date and proceeded gradually into the more extreme state. The Neoproterozoic climatic intervals can be described as follows:

The pre-Cryogenian warm interval (informal name, Tonian Period; 1000–850 Ma) is defined on the basis of a general abundance of substantial, presumably warm-water carbonate deposits that are sometimes associated with evaporite rocks (e.g. the Mackenzie Mountains Supergroup and Coates Lake Group, northwestern Canada, Jefferson & Parrish 1989; the Callana and Burra Groups, southern Australia, Preiss 1987). This interval is assumed to represent a continuation of the warm, equable conditions believed to have existed for the previous 1200 million years (Hambrey & Harland 1981, 1985; Lowe 1992; Buick et al. 1995).

The Cryogenian interval (850–635 Ma) is marked by glacial deposition on nearly every modern landmass except Antarctica (Hambrey & Harland 1981, 1985) during two episodes commonly referred to as the Sturtian and Marinoan (or Varanger) glaciations. What makes these two glaciations unusual is their severity compared with the Pleistocene glaciation, leading to the nickname ‘snowball Earth’ glaciations. Reliable palaeomagnetic evidence for ice-sheets having existed at low latitudes near sea level is strongest for the Marinoan glacial deposits of Australia (e.g. Sohl et al. 1999). Less well constrained are the positions of other glaciated continents, but existing palaeomagnetic data suggest that concurrent glaciation may have also existed at mid- to high latitudes (Torsvik et al. 1995; Meert et al. 1994). There is also evidence of low-latitude glaciation in northwestern Canada during the Sturtian glaciation (Park 1997), with other possible Sturtian glacial deposits at low latitudes in Australia (McWilliams & McElhinny 1980), India (e.g. Unrug 1992), and Namibia (Meert et al. 1995). Between these two glaciations is an interglacial interval of unknown duration, marked by the presence of largely siliciclastic successions (e.g. sandstones and shales) that contain varying proportions of carbonate rocks, such as in Australia (Preiss 1987) and northwestern Canada (Narbonne & Aitken 1995). The sedimentology and mineralogy of the interglacial carbonate rocks in south Australia suggest that subtropical to tropical conditions existed within 15–20° of the equator, and imply that latitudinal temperature gradients may have been comparable to the present day (Sohl 2000; cf. James et al. 2005).

Until recently, the commonly accepted timing of Cryogenian glaciations assumed a separation of over 100 million years between the Sturtian and Marinoan episodes, with the approximate age of the Sturtian set at 750–725 Ma and the age of the Marinoan at c. 600 Ma. New radiometric age constraints (Kendall et al. 2006) now suggest that glacial deposits associated with the Sturtian...
Glaciation encompass such a broad timespan—perhaps 750 Ma to 650 Ma—that our view of Cryogenian glacial episodes is in serious need of revision. For the purposes of this paper, we assume a ‘classic’ Sturtian glacial scenario set at c. 750 Ma.

The post-Cryogenian warm interval (635–542 Ma, essentially coincident with the recently ratified Ediacaran Period; Knoll et al. 2004) is defined based upon the development of several substantial shallow-water carbonate platform successions, implying warm-water depositional conditions in a number of widespread locations, e.g. India (Shankar et al. 1993; Jiang et al. 2002), Siberia (Pelechaty et al. 1996), and south China (Jiang et al. 2003). The Gaskiers glaciation has recently been shown to occur during this otherwise warm interval (Thompson & Bowring 2000; Condon et al. 2005), although it does not appear to have been as severe or as long-lived as the previous snowball Earth-type glaciations. There have been suggestions that additional minor glacial events occurred based in part upon sedimentological observations and fluctuations in the global δ¹³C isotopic curve (e.g. DiBona 1991; Kaufman et al. 1997), but debate continues over these purported events (Zhang et al. 2005).

**Challenges in deep-time palaeoclimate reconstruction**

As amply illustrated by the other chapters in this volume, geochemical and biological proxies are

### Fig. 1. Earth history context for climatic intervals of the Neoproterozoic Era (adapted from Chandler & Sohl 2000). Stratigraphic subdivisions follow Cowie et al. (1989) and Knoll et al. (2004). Ages and age ranges for geological and climatic events are approximate, given sparse radiometric age constraints in most cases. Tectonic events are after Hoffman (1991); Powell et al. (1993); Meert & van der Voo (1997); Dalziel (1997). Biological events follow Hofmann et al. (1990); Bromham et al. (1998); Grotzinger et al. (1995); Narbonne (2005). Environmental events according to Schmidt & Williams (1995); Canfield & Teske (1996); Park (1997); Sohl et al. (1999); Thompson & Bowring (2000); Kendall et al. (2006).
among the most common tools for indirectly extracting climate and primary productivity information from the geological record. These proxies include stable isotope values primarily measured in sediments and faunal assemblages in both the marine and terrestrial realms. The proxies can be related to climate system variables such as ocean temperature and circulation patterns, ocean productivity, seawater alkalinity, and atmospheric CO₂ levels. The utility of proxies has been demonstrated repeatedly in Pleistocene and younger sediments, while their use in older intervals of the Phanerozoic Era has met with varying levels of success (see reviews in Bradley 1994; Henderson 2002).

However, a review of the previous descriptions of the Neoproterozoic climate intervals will show that mention of geochemical and biological proxies is conspicuously lacking. That is because the deeper one goes into Earth’s past, the more problematic the use of these proxies becomes, since the sources of the proxy data are either unavailable or not readily interpreted owing to various factors. A principal problem is that many of the common organic sources for the geochemical proxies, and indeed all the climate-sensitive faunal assemblages typically used for the Cenozoic Era (e.g. incidence of *Pachyderma* left- vs. right-coiling foraminifera; occurrences of bryozoans vs. corals), simply do not exist in the Neoproterozoic Era: the organisms had not yet evolved. Organic-walled microfossils called acritarchs did exist throughout the Neoproterozoic Era, but their fossil record is remarkably dull for much of that time. There are few distinctive index forms until after the Marinoan glaciation, so there are limited biostratigraphic constraints to associate with the proxy data measured. Small shelly fossils (SSFs), the first organisms with calcite shells, only make their first appearance near the close of the Neoproterozoic Era (c. 550 Ma; Narbonne 2005), and so they too are of little help as sources of isotopic proxy data for palaeoclimatic reconstructions. In any case, we have insufficient data to determine whether the SSFs interacted with their ecosystem in a manner similar enough to foraminifera, such that we could draw the same environmental conclusions from the proxies measured. The TEX86 proxy for deriving sea surface temperatures (SSTs) from organic lipids in marine crenarchaeota (Schouten et al. 2003) may hold some promise, but the proxy is fairly new and remains untested for this time period.

There are alternative sources to purely organic sources of geochemical proxy data – the Neoproterozoic rock record does include a fair proportion of limestones and dolomites – but these sources also have severe limitations. Diagenetic (post-depositional chemical) changes can alter the values of geochemical proxies substantially. Oxygen isotope (δ¹⁸O) values are perhaps the most vulnerable to alteration (Killingley 1983; Schrag et al. 1995), but none of the other geochemical proxies is completely immune to alteration after burial (e.g. Lehmann et al. 2002). Even with great care in sample collection, it is not always possible to determine whether the values measured in the laboratory reflect the original geochemical conditions in the samples’ depositional environment (Jiang 2002; Kaufman et al. 2006). Mineralogical compositions and textures are also frequently altered through diagenesis (e.g. aragonite dissolves and is replaced by calcite; carbonate mud recrystallizes during neomorphism; Tucker 1990), and it may not be possible to determine the original state of the sample.

Despite these difficulties, we are not left without any means to get a sense of Neoproterozoic climatic conditions. The occurrence and distribution of ‘climate-sensitive’ sediments becomes the main proxy for palaeoclimate, through simple analogy to modern occurrences of the same types of rocks (Briden & Irving 1964; Parrish 1998). Glacial sediments require the action of glaciers, and thus a cold climate. Shallow-water limestones require warm water and supersaturation of carbonate in marine waters, which is typical of tropical marine shelf environments (assuming, perhaps erroneously, an ocean composition similar to present). Evaporite deposits require a hot, dry climate to evaporate seawater in restricted basins. The global climate picture must then be pieced together by making correlations between stratigraphic successions in different regions, and available palaeolatitude data, to determine the palaeogeographic extent of approximately age- and climatically equivalent rocks.

As a prime example of this methodology, the arguments in favour of the two Neoproterozoic snowball Earth glaciations depend principally on the sheer number and widespread palaeogeographical extent of presumed age-equivalent glacial deposits, which in some cases lay well down into low latitudes (Fig. 2; Hambrey & Harland 1981, 1985). There are additional, smaller-scale sedimentary features that also argue for cold climates across a range of latitudes during the glacial intervals. For example, there are examples of fossil permafrost features called sand wedges (e.g. Nystuen 1976; Deynoux 1982; Williams & Tonkin 1985; Deynoux et al. 1989; Zhang 1994), vertically foliated sand-filled cracks in frozen ground that are known to have formed during the Pleistocene ice age under arid conditions, when the annual average air temperature was below freezing but seasonally above freezing (Black 1976). There is localized disruption of sediments underlying glacial diamicites, which can be interpreted as ice-contact deformation at a shoreline. There is also the probable occurrence of glendonites, metastable
Hoffman & Schrag 2002) have taken the geological snowball Earth hypothesis (Hoffman et al.) to macroscopic life. Previous iterations of the possible climatic influence on the evolution of attracted most of the attention in recent years, in the record, the snowball Earth glaciations have we interpret from the Neoproterozoic geological described for the Mesozoic Era. Neoproterozoic Era is one of an equable climate, not unlike that today with tropical regions. The overall portrait of the non-glacial intervals of the Neoproterozoic Era comes from geological terrestrial/shallow marine proxy sources. Outside of the glacial intervals, there are widespread deposits of shallow marine carbonate rocks (limestones and dolomites) that are deposited today mainly in tropical settings, where the warmth of the seawater encourages the chemical reactions that precipitate carbonate out of the water (e.g. Preiss 1987; Jefferson & Parrish 1989; Jiang et al. 2002). Prior to the Sturtian glaciation, there are also regionally extensive mid-latitude deposits of evaporitic rocks (salts such as gypsum and halite) that must have accumulated in hot, arid environments (e.g. Hill et al. 2000), a depositional setting also more commonly associated today with tropical regions. The overall portrait of the non-glacial intervals of the Neoproterozoic Era is one of an equable climate, not unlike that described for the Mesozoic Era.

Of these two ‘end-member’ climate conditions we interpret from the Neoproterozoic geological record, the snowball Earth glaciations have attracted most of the attention in recent years, in large part because of the discussion over the possible climatic influence on the evolution of macroscopic life. Previous iterations of the snowball Earth hypothesis (Hoffman et al. 1998; Hoffman & Schrag 2002) have taken the geological evidence for widespread cold climates and extrapolated a vision of a world practically entombed in ice, with oceans frozen over or nearly so for millions of years. Such a condition would obviously have had an enormous impact on the ability of life to survive in abundance, had the freeze-over actually happened – and therein lays a point of contention. Proponents of the ‘hard’ snowball Earth have suggested that total or near-total sea-ice cover is necessary to explain both an interpreted rapid transition (a few hundreds to thousands of years) from the glacial to non-glacial state, as well as unusual $^{13}$C signatures, as low as $-5\%e$, in carbonate rocks (‘cap carbonates’) directly overlying the glacial deposits (Hoffman et al. 1998; Hoffman & Schrag 2002). Those in favour of a slightly less extreme scenario, the ‘slushball’ Earth, point to sedimentary deposits (ice-rafted debris in deep marine settings) that can be used to argue in favour of more open ocean rather than less (McMechan 2000; Condon et al. 2002; Kelllerhals & Matter 2003). Icebergs carrying the debris would have needed room to drift and melt as they dropped their sediment load, and active, wet-based continental glaciers probably needed a significant open-water source of moisture for precipitation, in order to be maintained. The problem here is that there are no data available that can support or refute the existence of total or near-total global sea-ice cover – there is no geological proxy in the Neoproterozoic Era for sea-ice cover at all.

The only tool we then have available to evaluate the likelihood of nearly, or totally, frozen oceans during either of the Neoproterozoic snowball Earth glaciations is computer climate simulation. To illustrate one possible modelling approach, we use here a version of the GISS Global Climate Model (GCM) to simulate the ‘classic’ Sturtian glaciation, c. 750 Ma, which has slightly better geological constraints available for the model boundary conditions. Our principal goal is to attempt to duplicate, as closely as possible, the surface conditions that would have permitted the existence of large-scale ice-sheets in mid- to low latitudes as indicated by the distribution of glacial deposits in the geological record. We emphasize that the GCM is used here to assess whether known forcings are consistent with direct geological evidence for the extreme glacial episodes and, additionally, to supply the physical process information required to unravel the mechanisms that led to climate change. Tests of the hypotheses for the exact nature of these snowball Earth glaciations (‘hard’ snowball vs. ‘slushball’) must ultimately take into account a combination of geochemical data, sedimentary features, depositional settings, and numerical model behaviour.
Choosing a climate model for palaeoclimate simulations

There are several classes of climate models that are employed in exploring deep-time palaeoclimate. Energy balance models (EBMs) are well suited for calculating the radiative effects of changes in greenhouse gas concentrations, especially when those concentrations are quite large with respect to the modern atmosphere. This circumstance typically arises in simulations of early Earth climate, when carbon dioxide and/or methane concentrations are thought to have been several orders of magnitude higher than at present (e.g. Caldeira & Kasting 1992). Earth models of intermediate complexity (EMICs) come in a wide variety of forms and typically incorporate a much broader range of parameters than EBMs, including a hydrologic cycle (Claussen et al. 2002). EMIC simulations can be run longer than GCM simulations, but computing efficiency is achieved in part by lowering the grid resolution of the model, as well as prescribing, to varying degrees, atmospheric and oceanic physics. Global climate models (GCMs) provide the most detailed representation of the climate system and its dynamics. GCMs include not only explicit calculation of energy balance, but also incorporate the conservation of mass, moisture and momentum. Although GCMs are much more computationally expensive than the simpler models, their higher geographic resolution makes them most appropriate for examining the effects of altered boundary conditions, and they are most directly compared with spatially delimited palaeoclimate proxy data (e.g. the distribution of glacial deposits). Large-scale climate change scenarios that directly related to tropical feedback mechanisms (such as snowball Earth climates) would be difficult to simulate accurately without inclusion of moisture and momentum physics.

For the study presented here, we employ an atmospheric GCM that incorporates an approximation of the Neoproterozoic palaeogeography (e. c. 750 Ma) in addition to other boundary conditions drawn from palaeoclimate proxy data.

Global climate model description

The experimental results presented here were produced using a newer version of the GISS Model II global climate model (GCM). The GCM is a three-dimensional Cartesian grid-point model, which solves numerically the equations of conservation for energy, mass, momentum and moisture, as well as the equation of state. It uses a horizontal grid resolution of 8° latitude by 10° longitude, with nine layers in the atmosphere, and has two ground layers for hydrological storage (which is minimal in our Neoproterozoic simulations, since all land is defined as desert). The model accounts for seasonal and diurnal solar variations and contains parameterizations for large-scale and convective clouds, background aerosols, and all radiatively important trace gases. Gases explicitly incorporated into the radiation scheme include carbon dioxide, methane and nitrous oxides (anthropogenic chlorofluorocarbons, or CFCs, were of course zeroed out in these simulations). The model generates precipitation whenever supersaturated conditions occur, and snow depth is based on the balance between snowfall, melting and sublimation.

Sea surface temperatures (SSTs) are calculated using model-derived surface energy fluxes and specified ocean heat convergences. The ocean heat convergences vary both seasonally and regionally, but are otherwise fixed. This is the primary mixed-layer ‘Q-flux’ ocean model developed for use with the GISS GCM, full details of which are described in Russell et al. (1985), and in appendix A of Hansen et al. (1988). We modified the original method of Russell et al. (1985) by using five harmonics instead of two in defining the seasonally varying energy flux and upper ocean energy storage, which improves accuracy in regions of seasonal sea ice formation. By deriving vertical fluxes and upper ocean heat storage from a run with appropriate palaeogeography for the Sturtian interval, the model provides a more self-consistent method for obtaining ocean heat transports.

The role of the ice/albedo feedback will be critical in experiments that produce significant changes in sea ice and snow cover. Newly fallen snow in the model has an albedo of 0.85 and ages within 50 days to a lower limit of 0.5. The sea-ice parameterization is thermodynamic; below −1.6 °C, ice of 0.5 m thickness forms over a fractional area of the grid box and henceforth grows horizontally as needed to maintain energy balance. Surface fluxes change the ocean water and sea-ice temperature in proportion to the area of a grid cell they cover. Conducive cooling occurs at the ocean/ice interface, with sea ice thickening if the water temperature remains at −1.6 °C. Sea ice melts when the ocean warms to 0 °C, and the SST in a grid box remains at 0 °C until all ice has melted in that cell. The albedo of snow-free sea ice is independent of thickness; it is assigned a value of 0.55 in the visible spectrum and 0.3 in the near infrared, for a spectrally weighted value of 0.45.

Setting up boundary conditions and forcings for a snowball Earth simulation

The scarcity of any palaeoclimate proxy data for the Neoproterozoic Era beyond the distribution and
characteristics of climate-sensitive sediments leaves rather more degrees of freedom in setting up simulations than would be the case for more recent palaeoclimates. For certain boundary conditions and initial forcings, we have no choice but to make educated guesses as to appropriate values based upon the information we do have. At the same time, the considerable age of the Neoproterozoic Era also introduces new considerations, such as significant changes in solar luminosity and palaeogeography, which would typically not be of interest to Pleistocene ice age or future climate modellers.

The key boundary conditions and forcings we altered for this series of Sturtian simulations, and the rationale behind the values selected, are as follows:

Solar luminosity

According to current astrophysical theory, a G2-type yellow star such as the Sun grows brighter as it ages; various standard stellar evolution models suggest that the Sun would have been 25 to 30% less luminous when the Solar System formed. Gough (1981) gives the luminosity change as

\[ L(t) = \left[ 1 + \frac{2}{3}(1 - t/t_\odot) \right]^{-1} L_\odot \]  

where \( t \leq t_\odot \), and \( t_\odot \) is the age of the Earth. Using Gough's equation, with \( t_\odot = 4550 \) Ma (Dalrymple 1991), the solar luminosity value for 750 Ma is 6.19% less than the modern value of 1366.619 W m\(^{-2}\), or 1282.026 W m\(^{-2}\). Incoming solar radiation is reduced in GISS Model II by decreasing the total amount of shortwave radiation entering the top of the atmosphere, and it is proportionally reduced at all wavelengths.

Palaeogeography/topography

The supercontinent of Rodinia, which had formed by the beginning of the Neoproterozoic Era (Hoffman 1991; Dalziel 1995), began to break apart again c. 800–780 Ma (e.g. Preiss 1987; Narbonne & Aitken 1995), although widespread rifting leading to the final break-up apparently did not occur until approximately 750 Ma, commensurate with the approximate onset of global cooling and the Sturtian glacial interval (e.g. Powell et al. 1993; Borg & DePaolo 1994). For our Sturtian experiments, we developed a palaeocontinental reconstruction based upon the available palaeomagnetic data (see Evans 2000 for a summary of palaeomagnetic constraints on glacial deposits) and other geological constraints (e.g. Dalziel 1997) used to define what Rodinia may have looked like just prior to break-up. Such reconstructions are necessarily tentative, as reliable palaeomagnetic data are not abundant and age constraints on the relevant rocks are not well defined (see e.g. Wingate et al. 2002 for a discussion of alternative reconstructions). However, a radically different continental configuration for the Sturtian interval, as compared with modern geography, does provide an opportunity to examine the possible effects of varying land distribution on climate, and specifically whether the low to mid-latitude landmass distribution could have played a key role in triggering glaciation by increasing the surface albedo of tropical regions (as per Kirschvink 1992).

Since the true relief for the Sturtian interval is unknown, we set the topography in all Sturtian simulations at a uniform 50-metre height for all land areas. Some of our previous unpublished Marinoan experiments have suggested that orography may be an important influence on the initiation of low-latitude glaciation. Topographic estimates can be made for presumed elevated regions based upon the age and distribution of orogenic belts and rift zones (Ziegler et al. 1985), but we do not pursue that issue in this study.

Vegetation

The GCM requires the assignment of vegetation categories to each grid cell containing land; these assignments are integral to the ground hydrology parameterization of our GCM (Hansen et al. 1983). Since the earliest-known fossils of vascular land plants are Ordovician in age (Wellman et al. 2003), it is generally assumed that Neoproterozoic continents were not vegetated in the modern sense (cf. Heckman et al. 2001; Kennedy et al. 2006). We mimic the lack of land vegetation by assigning desert conditions to all land area (Matthews 1983).

Greenhouse gases

The level of atmospheric CO\(_2\) during the Neoproterozoic Era is of prime importance for understanding palaeoclimate conditions, but proxy data are few, so varying viewpoints exist as to appropriate levels. Some researchers suggest that atmospheric CO\(_2\) levels needed to be high, perhaps 10\(\times\) modern, in order to counteract the decreased solar luminosity (e.g. Kasting 1987). Others have suggested that CO\(_2\) should have been at levels close to the modern value, given that the size of the dissolved inorganic carbon (DIC) reservoir in the oceans, which can be correlated with atmospheric p\(\text{CO}_2\), was not significantly different from the modern DIC reservoir (Bartley & Kah 2004). Our previous simulations run for the younger Marinoan
(Varanger) glaciation (Chandler & Sohl 2000) showed that modern and 4 × modern levels of CO₂ produced climatic conditions much too warm to support low-latitude glaciation, despite a less-luminous Sun. Even when we reduced CO₂ to a level of just 40 ppmv, less than a quarter of the minimum value during the late Pleistocene glaciation (Bartola et al. 1987), it was still not quite sufficient to make the Earth cold enough to support continental-scale ice-sheets in tropical latitudes.

For this series of Sturtian simulations then, we opted to modify combinations of both CO₂ and methane (CH₄). Although methane is present in the modern atmosphere in much lower concentrations than CO₂, it is a considerably more powerful greenhouse gas (as much as 32 × as effective as CO₂ in trapping heat), and we anticipated an increased cooling through the reduction of both gases in our simulations. Greenhouse gas levels employed for CO₂ included 315 ppmv (the ‘modern’ value measured in 1958), 140 ppmv (half the accepted pre-industrial value) and 40 ppmv (an extreme value we carry over from previous experiments). For methane, gas levels included 1.224 ppmv (again, the ‘modern’ value measured in 1958), and 0.612 (half the modern value) and 0.306 ppmv (25% of the modern level, and slightly less than the Late Pleistocene [145 ka] minimum value of c. 0.325 ppmv; Chappellaz et al. 1990).

Ocean heat transports

The transport of heat by ocean circulation is critical to the distribution of temperatures on the planet. Today, the oceans transport heat, on an annually averaged global basis, away from tropical and subtropical areas and into the middle and high latitudes. Previous simulations (e.g. Chandler & Sohl 2000) have shown that reducing ocean heat transports allows the polar regions to cool to a greater extent but sequesters heat in tropical regions. Although these changes are a simple redistribution of energy, the global climate impact is not negligible because altered distributions of ocean–atmosphere energy exchange directly affect certain key climate feedback mechanisms (e.g. ice albedo feedback, tropical moisture feedback). Thus, the ultimate impact of changes in ocean heat transport can be to alter the global climate. For this series of Sturtian experiments, we simulated the potential effects of poleward ocean heat transports that were 100%, 50% and 10% of the modern value. The geographic distribution of the ocean transports are necessarily modified since the Neoproterozoic ocean basin configurations are considerably different from modern configurations.

Ocean heat storage

As described above, the GCM uses a mixed-layer ocean model with a Q-flux parameterization to simulate horizontal transport of energy in the oceans. For these experiments, we adjusted the maximum mixed-layer depth (MLD) to be the full depth of the ocean (up to 4000 m) so the ocean heat storage capacity is equal to that of the full ocean. The GISS GCM does have an option to use a simple diffusion into the deeper ocean to mimic the sequestration of heat below the relatively shallow modern mixed layer. However, the diffusion rates are parameterized based on the geography of modern deep water production and are not appropriate for use with an altered ocean basin configuration. A fully dynamic coupled ocean is a desirable option, and some researchers have already employed one in their simulations of a snowball Earth-type ice age (e.g. Poulsen et al. 2001); however, with no Neoproterozoic bathymetry information available, even a dynamic ocean model does not provide definitive answers regarding the state of ocean circulation.

Experimental design

In order to determine just how much change we are introducing to a simulation by modifying the forcings, we have run two types of experimental control runs. In these, the forcings such as solar luminosity, atmospheric greenhouse gases and ocean heat transports are set at their modern levels (specifically, our control runs use 1958 forcing values, since that was the first year direct, continuous measurements of greenhouse gas levels were made). Our baseline reference for the model is a control run with modern geography, appropriately labelled ‘Modern’ for this series of experiments. For palaeoclimate experiments, we also conducted a control run (labelled S001R) using the Sturtian-age Rodinian continental distribution, but with all other forcings identical to the modern control. This ‘palaeo-control’ run allows us to make more meaningful assessments of the effects of solar and greenhouse gas forcings on the latitude by longitude climate differences for the Sturtian interval, which are otherwise dominated by the presence/absence of land at any given geographic location. The effects of the palaeogeographic changes on climate are shown separately as well.

All of the Sturtian experiments and control runs were run for 100 simulated years, and results analysed based on averages from the final five years of each simulation. A list of the specific setup of each simulation is shown in Table 1, and key
climate feedback quantities for each simulation are listed in Table 2.

There are ramifications to using a very deep maximum mixed-layer depth, and these are covered in the discussion section below. Our value of 4000 m is far greater than the average mixed-layer depth of 127 m in the modern ocean, but modern seasonal values in the cold polar regions are in fact substantially deeper (Kara et al. 2001) and are not out of the question for a very cold Neoproterozoic ocean during extreme glaciation events (D. Martinson, pers. comm. 2006). Additional examination of the effects of mixed-layer depth assignments in GCMs is certainly warranted.

**Simulation results**

**Individual forcings**

**Palaeogeography.** Our first comparison, between Modern and Sturtian (S001R) control runs, is a simple test of the effects of a change in landmass distribution. Kirschvink (1992) and Hoffman et al. (1998) had previously suggested that a concentration of landmass in mid- to low-latitude regions would lead to increased global cooling, on the assumption that land’s generally higher albedo would reflect more incoming solar radiation back out to space than the equivalent area of (initially) ice-free ocean. However, we found that the Sturtian control run was notably warmer than the Modern control run, in both tropical areas (by 1.6 °C) and in terms of global average temperature (by 2.0 °C) (see Table 3), clearly precluding the possibility that concentrating land at low latitudes would, on its own, induce a snowball Earth glaciation (see also Fig. 3a, b). It appears that while the centring of the Rodinia supercontinent on the equator does increase ground albedo, the concentration of land in one area (rather than the broken land distribution in the Modern) also permits higher net absorption of radiation over a broader area of the ocean, primarily in tropical and mid-latitude regions. At the same time, ocean ice cover was also higher for S001R (Table 3); this is not a contradiction of the warmer global average Sturtian temperature, but rather a reflection of having less land and more ocean area available for sea-ice growth in the polar regions.

**Solar luminosity.** A comparison between S002 and the control run S001R reveals the significant cooling impact of a 6.19% reduction in incident solar radiation, as per Gough’s (1981) equation. Simulation S002 is nearly 4 °C colder than the Sturtian control run, both in the global average and in the tropical average temperatures (Table 3). Total snow and ice cover, snow depth and ocean ice cover are all nearly double that of S001R, although the increase is fairly tightly focused in zonal bands between 60° and 75° latitude in both hemispheres, along the edge of the maximum annual sea-ice extent in S002. The increase in snow and ice cover is coincident with the greatest decreases in temperature, with the equator-to-pole temperature gradient steepening in S002 as anticipated. However, despite the dramatic decrease in global and tropical average temperatures, a significant portion of the planet – from the mid-latitudes to tropical regions – remains above 0 °C on an annual average basis, with zonal average temperatures in excess of 10 °C extending as far as 38° north and south of the equator. Such conditions rule out the possibility of annually persistent snow and ice cover that could develop into ice sheets at low latitudes.
Table 2. Values for key climate feedback quantities*

<table>
<thead>
<tr>
<th>Run ID</th>
<th>Altered forcings</th>
<th>Surface air temperature</th>
<th>Snow and ice</th>
<th>Cloud cover</th>
<th>Albedo</th>
<th>Atm. water vapour</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>CO₂ (ppm)</td>
<td>CH₄ (ppm)</td>
<td>OHT (% mod)</td>
<td>Global mean (°C)</td>
<td>Tropical (°C at 4° N)</td>
<td>Total cover (%)</td>
</tr>
<tr>
<td>Modem</td>
<td>315</td>
<td>1.224</td>
<td>100</td>
<td>13.11</td>
<td>24.8</td>
<td>12.24</td>
</tr>
<tr>
<td>S001R</td>
<td>315</td>
<td>1.224</td>
<td>100</td>
<td>15.11</td>
<td>26.4</td>
<td>7.52</td>
</tr>
<tr>
<td>S002</td>
<td>315</td>
<td>1.224</td>
<td>100</td>
<td>11.03</td>
<td>22.6</td>
<td>13.35</td>
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<td>S003</td>
<td>140</td>
<td>1.224</td>
<td>100</td>
<td>9.11</td>
<td>21.5</td>
<td>16.63</td>
</tr>
<tr>
<td>S004</td>
<td>140</td>
<td>1.224</td>
<td>50</td>
<td>7.98</td>
<td>22.9</td>
<td>19.07</td>
</tr>
<tr>
<td>S005</td>
<td>140</td>
<td>0.612</td>
<td>100</td>
<td>8.77</td>
<td>21.4</td>
<td>17.53</td>
</tr>
<tr>
<td>S006</td>
<td>140</td>
<td>0.612</td>
<td>50</td>
<td>7.76</td>
<td>22.8</td>
<td>19.26</td>
</tr>
<tr>
<td>S007</td>
<td>40</td>
<td>1.224</td>
<td>100</td>
<td>6.29</td>
<td>19.8</td>
<td>20.70</td>
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<tr>
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<td>0.612</td>
<td>50</td>
<td>4.68</td>
<td>21.1</td>
<td>23.85</td>
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<tr>
<td>S009</td>
<td>40</td>
<td>0.306</td>
<td>50</td>
<td>4.45</td>
<td>21.0</td>
<td>24.16</td>
</tr>
<tr>
<td>S010</td>
<td>40</td>
<td>0.306</td>
<td>10</td>
<td>3.36</td>
<td>22.1</td>
<td>26.50</td>
</tr>
</tbody>
</table>

*Values shown are global annual averages unless otherwise specified. All simulations except the Modern control run use Sturtian palaeogeography, and all runs except the Modern run and S001R use a 6.19% reduction in solar luminosity.
Table 3. Differences in pairs of simulations for changes to a single forcing*

<table>
<thead>
<tr>
<th>Run ID</th>
<th>Altered forcings</th>
<th>Surface air temperature</th>
<th>Snow and ice</th>
<th>Cloud cover</th>
<th>Albedo</th>
<th>Atm. water vapour</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Global mean (°C)</td>
<td>Tropical (°C at 4°N)</td>
<td>Total cover (%)</td>
<td>Snow depth (mm)</td>
<td>Ocean ice cover (%)</td>
</tr>
<tr>
<td><strong>Geography only</strong></td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Modern control–S001R</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>2.00</td>
<td>1.6</td>
</tr>
<tr>
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<td></td>
<td></td>
<td></td>
<td>-4.08</td>
<td>-3.8</td>
</tr>
<tr>
<td>Atmospheric CO₂</td>
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<td></td>
<td></td>
<td></td>
<td>315 &gt; 140 ppm</td>
<td>-1.92</td>
</tr>
<tr>
<td>S002–S003 (CH₄ = 1.224 ppm, OHT = 100)</td>
<td>315 &gt; 40 ppm</td>
<td>-4.74</td>
<td>-2.8</td>
<td>7.35</td>
<td>5.5</td>
<td>6.0</td>
</tr>
<tr>
<td>Atmospheric CH₄</td>
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<td></td>
<td></td>
<td></td>
<td>140 &gt; 40 ppm</td>
<td>-2.82</td>
</tr>
<tr>
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<td>140 &gt; 40 ppm</td>
<td>-3.08</td>
<td>-1.7</td>
<td>4.59</td>
<td>5.8</td>
<td>3.6</td>
</tr>
<tr>
<td>Atmospheric CH₄</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>140 &gt; 40 ppm</td>
<td>-3.08</td>
</tr>
<tr>
<td>S006–S008 (CH₄ = 0.612 ppm, OHT = 50)</td>
<td>140 &gt; 40 ppm</td>
<td>-3.08</td>
<td>-1.7</td>
<td>4.59</td>
<td>5.8</td>
<td>3.6</td>
</tr>
<tr>
<td>Ocean heat transports</td>
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<td></td>
<td></td>
<td></td>
<td>1.224 &gt; 0.612 ppm</td>
<td>-0.34</td>
</tr>
<tr>
<td>S003–S005 (CO₂ = 140 ppm, OHT = 100)</td>
<td>601.2 &gt; 0.306 ppm</td>
<td>-0.23</td>
<td>-0.1</td>
<td>0.31</td>
<td>2</td>
<td>0.1</td>
</tr>
<tr>
<td>Atmospheric CH₄</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>601.2 &gt; 0.306 ppm</td>
<td>-0.23</td>
</tr>
<tr>
<td>S009–S010 (CO₂ = 40 ppm, OHT = 50)</td>
<td>601.2 &gt; 0.306 ppm</td>
<td>-0.23</td>
<td>-0.1</td>
<td>0.31</td>
<td>2</td>
<td>0.1</td>
</tr>
</tbody>
</table>

* For comparisons based on geography and solar radiation, modern forcings for greenhouse gases and ocean heat transports apply. For other comparisons, values of ‘unchanged’ forcings may deviate from modern but are consistent within the pairs of simulations examined; specific forcings are indicated for each pair listed. Increases in climate feedback quantities are indicated in bold face; decreases are shown in italics.
Fig. 3. Comparison of the Modern control run and Sturtian control run S001R with simulation S010, the coldest of the series of experiments discussed in this chapter. The locations of known Sturtian-age glacial deposits are indicated by stars in the plots for S001R and S010, as per Hambrey & Harland (1981) and Evans (2000). Annual average surface air temperatures are shown for the Modern control run, S001R and S010 in (a), (b), and (c), respectively. The significant decrease in surface temperatures in S010 compared to S001R is highlighted in (d). Similarly, annual average snowfall is shown for the Modern control run, S001R and S010 in (e), (f), and (g), respectively, with the contrast in snowfall between S010 and S001R illustrated in (h).
Carbon dioxide. There are four pairs of simulations in which CO₂ is changed while other forcings were held constant at various values (see Table 1) under a less-luminous Sun: for S002/S003, CO₂ is decreased from 315 to 140 ppmv; for S002/S007, CO₂ is decreased from 315 to 40 ppmv; and for S003/S007 and S006/S008, CO₂ is decreased from 140 to 40 ppmv (the difference between the latter pairs being the CH₄ and OHT levels; see also Table 3). As expected, each successive decrease in CO₂ leads to cooling, both in tropical regions and on the global average; increases in total snow and ice cover and albedo; and decreases in atmospheric water vapour as the atmosphere becomes progressively colder and less capable of retaining moisture. The greatest cooling among these simulation pairs is, not surprisingly, induced by the maximum (275 ppmv) decrease in CO₂ (illustrated through a differencing of S002 and S007), is ~4.74 °C for the global annual average and ~2.8 °C in tropical regions (Table 3). The extent of cooling is comparable to that induced by decreased solar radiation alone, although the pattern of cooling is different: decreasing CO₂ creates a larger drop in global average temperatures than in tropical regions, reflecting the ‘blanketing’ effect of an atmospheric gas concentration versus the ‘spotlight’ effect of incident solar radiation.

Methane. There are three pairs of simulations in which CH₄ is changed while other forcings were held constant at various values (see Table 1) under a less-luminous Sun: for S003/S005 and S004/S006, CH₄ is decreased from 1.224 to 0.612 ppmv (half of the modern level); and for S008/S009, CH₄ is dropped from 0.612 to 0.306 ppmv. Given methane’s greenhouse warming potential, we had anticipated a noticeable enhancement to cooling effects produced by reductions in the other forcings. However, the reductions in atmospheric methane produced relatively little change across the board for the variables examined (Table 3). Temperatures did not drop more than a few tenths of a degree C; snow and ice cover, cloud cover, and albedo all increased by less than 1%; and atmospheric water vapour was virtually unchanged. The reason for this relative lack of response to methane forcing lies in the atmospheric levels used for these simulations. Although the warming potential of methane increases rapidly with concentration, the changes to the low concentrations we employed simply do not evoke significant impacts.

Altered ocean heat transports. There were three pairs of simulations in which OHT was reduced while other forcings were held constant at various values (see Table 1) under a less-luminous Sun: for S003/S004 and S005/S006, OHT was reduced to 50% of modern, and for S009/S010, OHT was reduced to just 10% of the modern level. As anticipated from previous experiments (Chandler & Sohl 2000), reducing OHT increased the average annual temperature of tropical regions, since less heat was shifted poleward; the average temperatures here never dipped below a warm 20 °C (see Table 3). The concentration of heat in tropical areas did not, however, prevent a reduction in the global average temperature and, in fact, total snow and ice cover as well as snow depth increased with each decrease in OHT (Table 3). Much of the increase in snowfall was over areas newly covered by ocean ice, which acted as a platform for snow accumulation, although some of the increase did occur over portions of land in the Northern Hemisphere.

Combined forcings

As with our previous snowball Earth experiments (Chandler & Sohl 2000), the combined effects of alterations to multiple forcings were more effective in both reducing surface temperatures and increasing snow and ice cover than individual forcings alone. Since the pattern of climatic change is similar for runs S002 through S010, varying chiefly by degree, we focus here on the results of our coldest simulation, S010. This simulation had the greatest reductions in CO₂, CH₄ and OHT, in addition to solar luminosity (see Tables 2 and 3), compared to the Sturtian control run S001.

With a global annual average surface temperature of just 3.36 °C, S010 is 11.75 °C colder than S001, and 9.75 °C colder than the Modern control run. Snow and ice coverage in S010 expands to more than a quarter of the planet as well (26.5%), with a notable increase in both planetary and ground albedo as a consequence (6.42% and 8.91% over S001R, respectively). The decrease of atmospheric water vapour by more than a third from S001R is both a reflection of the increasingly cold and dry atmosphere as well as a feedback in and of itself, since water vapour is the most powerful greenhouse gas of all. However, despite the profound overall cooling observed in S010, the reduced OHT sequestered sufficient heat in equatorial regions to make annual average tropical temperatures warmer in S010 than in experiments with greater ocean heat transports (S007, S008 and S009) (see Tables 2 and 3). The tropical temperature in S010 is still colder than the Modern control, but only by 2.7 °C, while it was 4.3 °C cooler than the Sturtian control run S001R.

Perhaps the best way to appreciate the extent of cooling in S010 is to compare it visually to both the Modern and Sturtian control runs. Maps of the annual average surface air temperature show that the contrast between the modern and palneo
control runs is small (Fig. 3a and Fig. 3b), which is reasonable given that the only difference between the two is geography. The annual average surface air temperature for S010 (Fig. 3c), however, is clearly colder than its palaeo control, S001R (Fig. 3b), with above-freezing temperatures confined to a narrower band in tropical and lower mid-latitude regions. The anomaly plot (Fig. 3d) illustrates the temperature differential in S010 compared to S001R and underscores the considerable extent of cooling associated with the maximum decreases in solar luminosity, greenhouse gases and ocean heat transport. A similar pattern is evident in maps of the annual average snowfall, expressed here as millimetres of water equivalent per day (1 mm H₂O equivalent is roughly equal to 10 mm freshly fallen snow). The Modern (Fig. 3e) and S001R (Fig. 3f) runs are similar to each other, while S010 shows snowfall rates that are considerably increased (Fig. 3g) in middle latitudes (Fig. 3h). Note that the maximum snowfall (as well as differential increase) is highest along the leading edges of the sea-ice cover of both hemispheres in the two Sturtian simulations (Fig. 4b–c), relating directly to the availability of moisture for precipitation along the sea ice–open ocean boundary, despite the low atmospheric water vapour content for S010 in particular (see Table 2). Note also that our choice of low topography everywhere over the continents has a substantial influence on possible sites of low-latitude snow accumulation. There is no equivalent in our Sturtian simulations to the modern Tibetan Plateau or Western Cordillera, which today can shift annual average snowfall contours equatorward by 10 to 20 degrees of latitude.

Another series of maps that emphasize the extent of global cooling in S010 are shown in Figure 4, which contrasts the Sturtian vs. Modern annual snow and ice coverage in the Northern and Southern Hemispheres. The polar ice caps in S001R (Fig. 4b) are even smaller than those of the modern climate (Fig. 4a), especially in the south polar region, illustrating the regional warming effect of the open polar oceans in the Sturtian as contrasted with the land-covered (Antarctica) and land-locked (Arctic) poles in present-day geography. Both northern and southern polar ice caps in S010 are vast (Fig. 4c), with the 50% snow and ice cover contour extending to roughly 45° latitude and impinging upon the mid-latitude landmasses in both hemispheres. The zonal average temperatures for each of these simulations (Fig. 5) illustrates the temperature control on the snow and ice cover extent quite well; the upper ‘shoulders’ of each profile coincide with the latitude at which the annual average surface air temperature reaches the freezing mark. For the Modern control run, this point is around 60° latitude; for S001R, the freezing mark coincides with the Modern control run in the Northern Hemisphere, but is at 70° latitude in the Southern Hemisphere (reflecting the warming produced by the removal of Antarctica from a polar position); and for S010, the freezing mark is at a remarkably low 45° latitude.

Discussion

Having identified S010 as our coldest simulation, we need to assess whether the surface conditions described by S010 are favourable to what we know of the distribution of Sturtian glacial deposits. The surface air temperature and snowfall maps for S010 (Fig. 3c, g), which include the marked positions of known Sturtian-age glacial rocks, show
significant inconsistencies with the geographic locations of low-latitude glacial deposits. Most of those deposits fall within a tropical zone that is actually quite warm in this simulation, with an annual average temperature in excess of 20 °C (Fig. 3c). Not surprisingly, such warmth is not conducive to precipitation as snowfall, much less the existence of annually persistent snow and ice cover that could lead to ice sheet formation. This enhanced tropical warmth despite an overall much colder planet is the direct result of reducing ocean heat transports to just 10% of the modern level, which sequesters heat in the tropical oceans. The zonal temperature gradient for S010 thus ends up fairly steep compared to the Modern and Sturtian control runs (Fig. 5). This steep gradient acts as a barrier to land and sea-ice encroachment into tropical latitudes, in contrast to other climate studies which use diffusive rather than dynamic means to transport heat out of tropical regions (e.g. Donnadieu et al. 2004). While the lack of tropical sea ice is not a problem for ‘soft’ snowball Earth scenarios, the lack of tropical land ice is, as these simulations do not reproduce surface conditions necessary for the onset and persistence of low-latitude ice-sheet formation.

Our previous simulations of the younger Varanger (Marinoan) glaciation (Chandler & Sohl 2000) did produce snow and ice coverage on land in tropical regions. The Sturtian simulations are considerably warmer than our original Varanger simulations, by as much as 15 °C and more in tropical regions, despite similar combinations of solar and greenhouse gas forcings. This contrast is probably the result of two key factors. First, our previous Varanger palaeogeographic reconstruction was dominated by a landmass configuration that covered the south polar region and extended northward across all latitudes of the Southern Hemisphere. Relative to the Sturtian interval, there was little land at tropical latitudes. The higher concentration of mid- to high-latitude landmass would allow Varanger ice-sheets to initiate more readily, and the ice-sheets themselves could then contribute to climate change through elevated surface albedo and topographic influence on wind and precipitation. Such local climate effects would allow for some equatorward expansion of snow cover. In the real world, where ice dynamics play an important role, we surmise that stable accumulations of snow in the mid-latitudes would probably lead to flow of glaciers into lower latitudes. In this regard, Sturtian palaeogeography is somewhat ‘handicapped’ by the difficulty of initiating as well as maintaining ice-sheets at lower latitudes.

Secondly, the maximum mixed-layer depth for our previous Varanger simulations was only set to 127 m, compared to the 4000 m used in the Sturtian experiments. As a result, the Varanger mixed-layer ocean had less heat storage available than the Sturtian ocean and was more susceptible to rapid cooling. Indeed, this rapid cooling sometimes
causes instabilities in the model as the ocean freezes through the entire depth of the shallower mixed layer. Our use of a 4000 m maximum for the simulations discussed here allowed us to avoid instabilities, but the greater heat capacity of such a thick mixed layer depth means that the simulations do not achieve equilibrium within 100 years. We can assume that the simulations would cool further, but the final results are likely to be no cooler than our initial Varanger experiments; it would simply have taken longer to reach that point. The role of the maximum mixed-layer depth and the correct assignment of geographic and seasonal values do ‘however’ need to be explored in much greater detail, and additional study with fully coupled ocean–atmosphere GCMs is warranted.

Still, our results from simulation S010, the coldest of the Sturtian runs completed, bring to light some additional interesting points for future investigations. One point is that although overall atmospheric water vapour levels are considerably reduced, the relatively warm tropical regions act as a heat engine driving enhanced snowfall along the margins of the annually persistent snow and ice cover (Fig. 3g). The contrast between S001R and S010 in these marginal areas is in excess of 4 mm H₂O per day, a difference not unlike the pre- and S010 in these marginal areas is in excess of ice cover (Fig. 3g). The contrast between S001R and S010 in these marginal areas is in excess of 4 mm H₂O per day, a difference not unlike the precipitation contrast between the modern Sahara Desert and northeastern United States. It may be that a tropical heat engine is required, at least initially, for there to be sufficient snow and ice build-up in mid-latitude regions, thus overcoming the Sturtian palaeogeographic handicap. Ice-sheet growth on land at these locations, rather than sea ice growth, may be needed to provoke the additional cooling that would ultimately allow continent-scale ice-sheets to spread across the fragments of Rodinia. Recall also that the reconstruction for these simulations did not include topographic relief. The uniform 50-metre height of our Sturtian continents is low compared to the 236-metre average elevation for our Modern control run. In the Modern control run, the extension of snow and ice cover into mid-latitude regions has also clearly been influenced by the presence of the Tibetan Plateau, the Western Cordillera and the Andes Mountains (Fig. 3e). The possibility that orographic effects influenced climate during the Sturtian glaciation cannot be excluded.

Irrespective of such additional influences, which may indeed have made low-latitude continental icesheets viable, the expansion of sea ice across all latitudes does not seem to be consistent with any understanding of the likely range of climate forcings that existed during the Neoproterozoic Era. Earlier in Earth history, extreme glacial episodes may very well have resulted in ‘hard’ snowball Earth scenarios. For example, the Huronian glaciation of the Palaeoproterozoic Era (c. 2.1 Ga) would have occurred under solar luminosities that were as much as 15% below modern – or more than twice the decrease experienced by the Sturtian glaciation. Three-dimensional climate models, like the GISS GCM, have sensitivities that suggest a runaway icehouse can be achieved at modern levels of greenhouse gases if solar luminosities are only 85% to 90% of modern (J. Hansen, pers. comm, 2001). Thus the palaeoclimates of early Earth are generally thought to require higher amounts of atmospheric greenhouse gases to explain why the planet was not continuously in a ‘hard snowball’ state.

Conclusions

In this effort to replicate climatic conditions amenable to the development of large ice-sheets in low to mid-latitude continental regions during the Neoproterozoic Sturtian glaciation, no single climate forcing is sufficient to cool the planet adequately. Reductions in solar luminosity, carbon dioxide, methane and ocean heat transports all induce some cooling, albeit in different ways, with solar luminosity and carbon dioxide having the greatest impact on temperature. The effects of the solar luminosity reduction and maximum decrease in carbon dioxide to 40 ppmv are similar in magnitude, although CO₂ reduction has a greater global impact while solar luminosity reduction has a greater impact on tropical regions. Decreases in methane lead to only minor enhancement of cooling brought on by other forcing changes; this result perhaps runs counter to what might be expected, given methane’s high greenhouse warming potential, but is in fact in line with methane’s nonlinear response to changes in concentration. Reductions in ocean heat transports led to global cooling, with dramatic cooling at higher latitudes, but slight warming in some tropical regions, as heat is less efficiently moved away from low latitude areas as ocean heat transport declines.

Combinations of altered forcings did increase the amount of global cooling in the model, with the most extreme combination of forcings in simulation S010 producing a world with sea ice extending to 40° latitude and a global average surface temperature of just 3.36 °C. Despite this considerable cooling with respect to the Modern and Sturtian control runs, most continental areas never became cold enough to support annually persistent snow and ice cover, which is required for the large ice-sheets inferred from the geological record of glacial deposits to develop. Taken together, however, the tropical warming induced by reduced ocean heat transports, coincident
increase in precipitation, and the steep zonal average temperature gradient, suggests that the model need not produce much more cooling before larger portions of land would be capable of supporting year-round snow and ice cover. Coupled dynamic ice-sheet modelling (a field that still needs to see considerable advances even for modern ice-sheets) may be important to accurately portraying the land-ice masses of the Neoproterozoic Era.

While this series of experiments was not intended to address specifically any of the tenets of the ‘hard’ snowball Earth hypothesis (Hoffman et al. 1998; Hoffman & Schrag 2002), our results do have some implications for two key assumptions not directly based upon data in the geological record. The first assumption is that concentration of landmass in low to mid-latitude regions would increase the planet’s albedo and lead to global cooling. We find that a low- to mid-latitude landmass concentration does increase ground albedo, but planetary albedo itself decreases and the planet is warmer than anticipated. A contributing factor appears to be the absence of a landmass like modern Antarctica in the polar regions, which is isolated from heat exchange with lower latitudes by the oceanic Circumpolar Current.

A second assumption of the ‘hard’ snowball Earth hypothesis is that global cooling would have lead to sea ice encroachment into ever-lower latitudes, until an ice-albedo feedback effect set in and rapidly produced total or near-total sea ice cover in tropical regions. We find that reducing ocean heat transports to as little as 10% of the modern value does permit sea ice to extend further equatorward, but incoming solar radiation is then sequestered within a smaller area of the tropical ocean. The resulting tropical temperatures are even higher than the Modern control run, and act as an absolute barrier to sea ice development over a sizeable swath of ocean. Our results in this case highlight the utility of computer models in ‘checking’ conceptual models of climatic change, even if the computer model itself does not fully reproduce the features preserved in the geological record.

The authors thank Jeff Jonas and Michael Shopsin for programming assistance. This work has been supported by NSF grants ATM-9907640 (to N. Christie-Blick, LS & MC) and ATM-0214400, ATM-0231400 and ATM-03 23516 (to MC).

References


Kennedy, M. J., Droscher, M., Mayer, L. M., Pevear, D. & Mrofka, D. 2006. Late Precambrian


