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Earth and Planetary Science Letters 211 (2003) 1–12

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The Sturtian 'snowball' glaciation: fire and ice

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Received 31 January 2003; received in revised form 31 March 2003; accepted 3 April 2003

Abstract

The Sturtian 'snowball' glaciation (730 Ma) is contemporary with the dislocation of the Rodinia supercontinent. This dislocation is heralded and accompanied by intense magmatic events, including the onset of large basaltic provinces between 825 and 755 Ma. Among these magmatic events, the most important one is the onset of a Laurentian magmatic province at 780 Ma around a latitude of 30°N. The presence of these fresh basaltic provinces increases the weatherability of the continental surfaces, resulting in an enhanced consumption of atmospheric CO_2 through weathering, inducing a global long-term climatic cooling. Based on recent weathering laws for basaltic lithology and on climatic model results, we show that the weathering of a 6×10^6 km^2 basaltic province located within the equatorial region (where weathering of the province and consumption of CO_2 are boosted by optimal climatic conditions) is sufficient to trigger a snowball glaciation, assuming a pre-perturbation PCO_2 value of 280 ppmv. We show that the Laurentian magmatic province might be the main culprit for the initiation of the Sturtian 'snowball' glaciation, since the Laurentian magmatic province had drifted within the equatorial region by the time of the glaciation.
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Keywords: snowball Earth; glaciation; basaltic traps; carbon cycle; modeling

1. Introduction

The Earth underwent at least two episodes of severe glaciation at the termination of the Proterozoic era, one around 730 Ma (Sturtian/Rapitan episode), and a second around 600 Ma (Marinoan/Varanginian episode). In 1992, Kirschvink [1] suggested that both episodes might correspond to a complete glaciation event (the 'snowball Earth' hypothesis), with total cover-up of continental and oceanic surfaces by ice. This hypothesis seems to account for most field observations [2], and especially for the low paleolatitudes inferred from paleomagnetic results derived from some glacigenic deposits [3,4] as well as for the observed large negative excursions in carbonate $\delta^{13}C$ [5,6]. However, the causes that drove the

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doi:10.1016/S0012-821X(03)00197-3

Earth into a snowball state remain obscure. Since the simple hypothesis invoking higher obliquity cannot explain the Neoproterozoic snowballs [7–9], cooling that occurred during these episodes must be related to changes in atmospheric greenhouse gas concentrations. To date, only two qualitative scenarios have been proposed, linking the onset of a snowball glaciation with a long-term decrease in the partial pressure of atmospheric CO_2. The first hypothesis invokes breakup of the Rodinia supercontinent, trapping atmospheric CO_2 into organic carbon deposits in the newly created continental margins [5]. A second hypothesis relies on the equatorial location of continental masses and invokes methane release from organically rich sediments into a low oxygen atmosphere, ensuring a rather long residence time of CH_4 within the atmosphere, boosting consumption of atmospheric CO_2 by weathering of continental silicates through enhanced greenhouse conditions [10]. Once methane release ends, atmospheric methane collapses through oxidation and the remaining atmospheric CO_2 might be low enough to result in the onset of a snowball glaciation. Although appealing (it might explain the pre-snowball $\delta^{13}C$ decrease recorded in some carbonates, especially beneath glaciogenic Marinoan formations [11]), the methane hypothesis requires an ad hoc process able to initiate and stop methane release from sediments into the atmosphere that has yet to be identified.

Here we further explore the impact of continental large igneous provinces preceding the breakup of a supercontinent and the subsequent rift-to-drift transition on the global carbon cycle. A causative link is suggested between emplacement of basaltic traps and nearly coeval breakup of Rodinia with the onset of the first snowball episode: the Sturtian glaciation.

2. Continental flood basalts, Rodinia breakup and the carbon cycle

The impact of continental flood basalts on the global carbon cycle and climate has been quantitatively described by Dessert et al. [12] in the case of the Deccan trap event occurring at the Cretaceous–Tertiary (K–T) boundary [13]. Coeval with the beginning of trap eruption, atmospheric CO_2 first rises due to degassing of the hot basaltic lavas. The subsequent global warming is then slowly counteracted by the increasing consumption of atmospheric CO_2 due to continental silicate weathering (including the weathering of the newly exposed basaltic surface). Indeed, the weathering of continental silicates is enhanced under a warmer and (assumed) wetter climate [14]. This dependence on climatic conditions introduces a negative feedback loop which stabilizes the partial pressure of atmospheric CO_2 [15], increasing the sink as the source increases. Atmospheric CO_2 thus reaches a new steady state several million years after the end of trap emplacement and degassing. However, because basaltic rocks weather about eight times faster than granitic rocks [16], the global weatherability of the continental surface drastically increases, and the new steady-state atmospheric CO_2 and mean global air temperature are lower than the corresponding pre-perturbation values. Would it be possible that basaltic traps initiated the Sturtian glaciation?

The Rodinia supercontinent starts to break up at ≥ 750 Ma [17]. This event is heralded by pulses of tholeiitic magmatism spanning the 825–755 Ma time interval, especially in Australia [18,19] and northwest Laurentia [20], but also in South China [21] and Congo [22] cratons. The age, location and likely areal extension of these basaltic provinces is presented in Fig. 1. We stress that, in the field, these mafic rocks underlie the formations containing the Sturtian/Rapitan glacigenic deposits. Collectively, these magmatic events appear as a possible plume time-cluster in the database of Ernst and Buchan [23]. The 825 and 780 Ma old events can be confidently regarded as plume-related large continental basaltic provinces, due to the length of dyke swarms, the thickness of associated volcanics and the geochemical nature of the magmatic rocks [18,20,24]. The 780 Ma old Laurentian plume is especially well characterized in minimum size due to the giant radiating dyke swarm, which can be traced in the Mackenzie Mountains of northwestern Canada, as well as in the Wyoming province of the western USA.

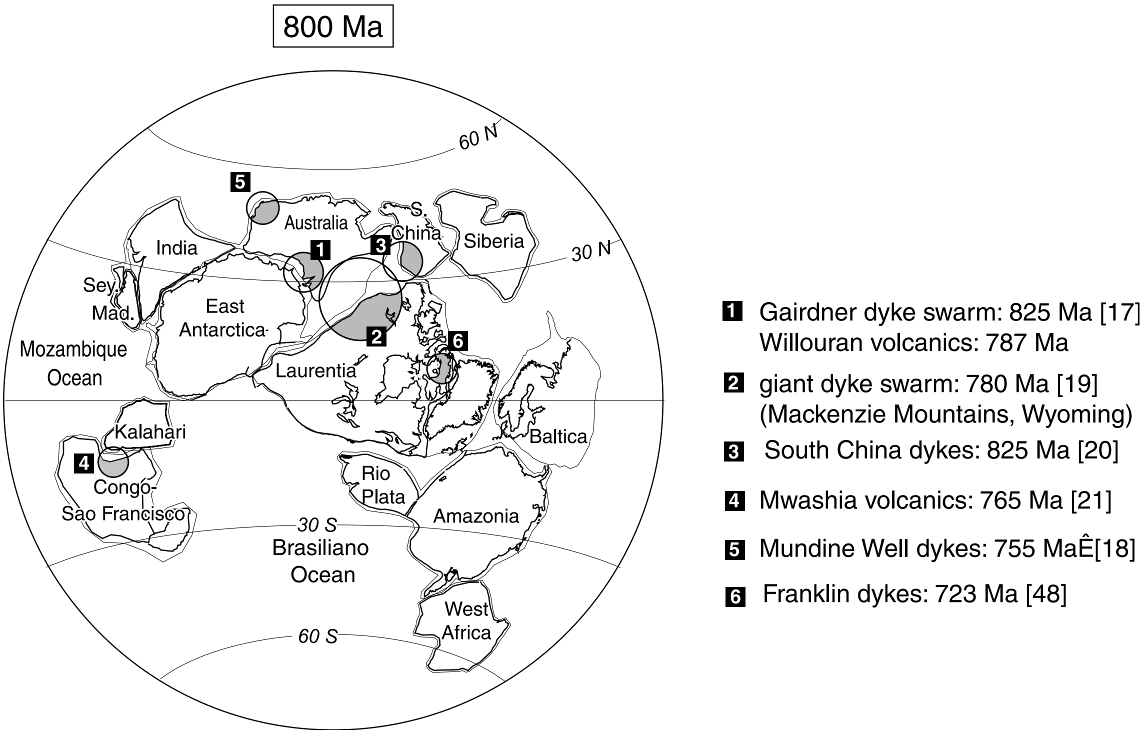

Fig. 1. Continental reconstruction at around 800 Ma (adapted from Weil et al. [50] and Torsvik et al. [28]), i.e. prior to the Rodinia breakup. Numbers refer to basaltic igneous provinces, which will form during the 825–725 Ma time interval; gray area represents attested geological remnants (such as flood basalts, dyke swarm or gabbroic layered intrusives), circles are in proportion to the inferred areal extents of the magmatic provinces.

The directions of outcropping dykes converge towards a focal point located near the present Queen Charlotte Islands (British Columbia), hence a *minimum* areal extent for the related floods equals 3×10^6 km^2, comparable with the estimated initial size of the Siberian traps (5×10^6 km^2) [25]. The focal point marks the arrival of a plume head that will further initiate the opening of the Proto-Pacific Ocean (Fig. 2), and accelerate the southward drift of the Laurentia plate.

Given the north-tropical location of these new continental basaltic surfaces, there is a delay of about 40–50 Myr before the northwestern Laurentia traps will be potentially exposed to the warmest and wettest equatorial climate as a consequence of the rifting and southward drifting of the Laurentia plate. That drift was first probably relatively slow. From the reconstructed continental configuration (Figs. 1 and 2), we estimate that the southward drift was not bigger that 2 cm/yr between 780 and 750 Ma. Then, the drift might have increased with the opening of the Proto-Pacific Ocean, as suggested by the southern location of the continental masses during the Marinoan age. A reasonable drift speed of 8 cm/yr is enough to bring the Laurentian traps to the equatorial region at 730 Ma. The equatorial warm climate will facilitate weathering of these fresh basaltic surfaces, thus maximizing atmospheric CO_2 pumping. Furthermore, continental runoff is most probably maximum within the equatorial region. The global cooling initiated by the chemical weathering of a continental basaltic trap will thus be most effective at a time that will coincide with the Sturtian ice age.

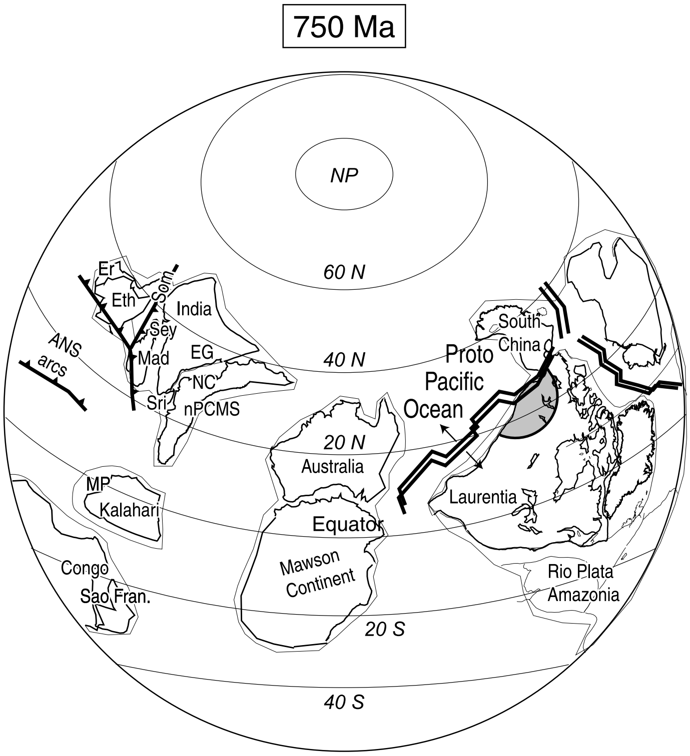

Fig. 2. Continental reconstruction at around 750 Ma, with location of the Proto-Pacific Ocean after Meert et al. [39]. ANS, Arabian–Nubian Shield; EG, Eastern Ghats; Eth, Ethiopia; Mad, Madagascar; Mawson continent, part of East Antarctica; NP, Napier Complex. Gray area represents the location of the Laurentia magmatic province at 750 Ma.

3. Quantification of the trap effect

3.1. Critical CO_2 level

The estimation of the size of a trap able to produce a snowball Earth first requires the knowledge of the atmospheric CO_2 threshold below which the Earth undergoes a global glaciation. To estimate the numerical value of this threshold, we chose to use a 1D energy balance model (EBM) that calculates annual mean air temperature as a function of atmospheric CO_2 pressure in 18 latitude bands [26]. The original infrared scheme was modified in agreement with Caldeira and Kasting [27]. Assuming a Sturtian equatorial paleogeography (total continental surface = 90×10^6 km^2 [28]) and a solar constant reduced by 6%, the snowball glaciation is initiated once atmospheric PCO_2 falls below 135 ppmv. This result falls within the range of previous studies based on AGCM simulations (100 ppmv in Jenkins and Smith [29]) as well as on EBM coupled to the ice sheet model (130 ppmv in Hyde et al. [30]). However, the threshold value is highly dependent on total continental surface and on the albedo adopted for unvegetated surfaces (taken to be 0.30 in this work). Adopting a Sturtian continental surface equivalent to the present-day surface (150×10^6 km^2) increases the threshold PCO_2 up to 770 ppmv. This would not strongly modify the present hypothesis, since the most important parameter is the amount of CO_2 that can be removed through trap weathering, hence the difference between the pre-trap and threshold PCO_2.

3.2. Weathering laws and basaltic province area

The quantification of the trap hypothesis requires global weathering laws. The consumption of atmospheric CO_2 can be written as [16,31]:

1. (i) for a granitic lithology:

$$f_{w,i}^{gra} = k_{gra} \cdot \text{runoff}_i \cdot \text{area}_i \cdot \exp\left(-\frac{48200}{R}\left(\frac{1}{(T_i+273.15)}-\frac{1}{(T_0+273.15)}\right)\right) \quad (1)$$

2. (ii) for a basaltic lithology:

$$f_{w,i}^{bas} = \text{area}_i^{bas} \cdot \text{runoff}_i \cdot 3234.4\exp(0.0652T_i) \quad (2)$$

T_i and runoff$_i$ are air temperature in °C and continental runoff in cm/yr within latitude band i. Area$_i$ represents the continental area within latitude band i, and area$_i^{bas}$ the area of exposed basaltic surface in latitude band i in square kilometers. k_{gra} is a constant calibrated so that, under present-day climatic and paleogeographic conditions, $F_{w,i}^{gra}$ equals 9.5×10^{12} mol/yr [32], and T_0 is a reference temperature in °C. These laws were determined through the compilation of the studies performed on small granitic and basaltic catchments. They integrate the physical erosion term. Transposing these weathering laws into the Neoproterozoic environment implies that the contribution from physical erosion to chemical weathering was the same as at present for the same temperature and runoff conditions, an assumption that appears conservative but might be wrong during intense orogenic phases.

The pre-trap steady state can be described by the following equation [15]:

$$F_{vol} = \sum_{i=1}^{18} F_{w,i}^{gra} = \sum_{i=1}^{18} k_{gra} \cdot \text{runoff}_i \cdot \text{area}_i \cdot \exp\left(-\frac{48200}{R}\left(\frac{1}{(T_i+273.15)}-\frac{1}{(T_0+273.15)}\right)\right) \quad (3)$$

where F_{vol} accounts for all CO_2 degassing sources (MOR and arc volcanism). This equation expresses that the amount of carbon released within the ocean–atmosphere system must be equal to the total consumption of carbon through continental silicate weathering. Note that to each value of F_{vol} corresponds a given set of zonal temperatures T_i, hence a given steady-state atmospheric PCO_2. Assuming that the trap spreads over a 'pre-existing' granitic lithology within the latitude band j, the steady state reached by the global ocean–atmosphere carbon cycle several million years after the end of the eruptive phase will be described by:

$$F_{vol} = k_{gra} \cdot \text{runoff}_j (\text{area}_j - \text{area}_j^{bas}) \exp\left(-\frac{48200}{R}\left(\frac{1}{(T_j+273.15)}-\frac{1}{(T_0+273.15)}\right)\right) + \text{area}_j^{bas} \cdot \text{runoff}_j \cdot 3234.4\exp(0.0642T_j) + \sum_{i=1,\, i\neq j}^{18} F_{w,i}^{gra} \quad (4)$$

This critical equation is used to estimate the size of the trap area$_j^{bas}$ required to plunge the Earth into a snowball, as a function of the latitude band j of trap location and the background CO_2 degassing F_{vol} that can be translated in terms of pre-trap steady-state PCO_2. For the sake of simplicity, we will assume no split of the trap surface between different latitude bands, i.e. the trap is assumed to be completely included within a single latitude band. The temperatures (T_i and T_j) in Eq. 4 are given by the 1D EBM simulation with a threshold PCO_2 of 135 ppmv. The continental runoff terms runoff$_i$ and runoff$_j$ are calculated as a parametric function of zonal air temperature, continental area and latitude [26].

Fig. 3 presents the minimum trap size for differ-

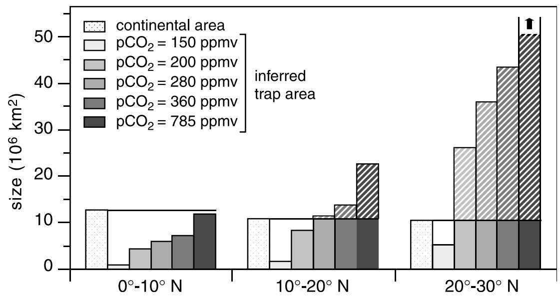

Fig. 3. Minimum size of continental basaltic traps required to force the climatic system into a snowball state (in million km^2), as a function of pre-trap steady-state atmospheric PCO_2 and of the latitudinal location of the traps (along the Proto-Pacific rift). The trap areas are compared to the available continental surface in the considered latitude band.

ent pre-perturbation PCO_2, and for three different locations along the Proto-Pacific Ridge (between 0° and 30°N). For instance, if the PCO_2 prior to the trap-weathering event was in excess of 785 ppmv, no acceptable solutions are found, since the required trap size equals or exceeds the available continental area in each latitude band. On the other hand, if the pre-perturbation weathering PCO_2 ranges from 150 to 360 ppmv, we found acceptable solutions when the traps are located and weathered between 0° and 10°N, the trap areal extent required to freeze completely the Earth ranging from 0.9×10^6 to 7.2×10^6 km^2. The hypothesis is highly plausible, as this size range encompasses the reconstructed minimum initial size of the northwestern Laurentia traps.

3.3. Modeling the consequences of the Laurentian traps (780 Ma) on the global carbon cycle

As an illustration, the onset of the Sturtian snowball glaciation following a trap-weathering event is simulated using a modified version of the Dessert et al. model [12]. This three-box model (atmosphere, surface and deep ocean) describes the global inorganic carbon and alkalinity cycles, including weathering and precipitation of carbonates. The organic carbon deposition and oxidation of kerogen carbon on continents is not calculated. The original model has been improved as follows: the annual mean air temperature is calculated at each time step within 18 latitude bands as a function of PCO_2 by a coupled energy balance climate model [26,27], while the continental zonal runoff is calculated as a function of the zonal air temperature, continental area and latitude [26]. Weathering laws for silicate lithologies are given by Eqs. 1 and 2. Pre-perturbation PCO_2 is assumed to be 280 ppmv. We choose to illustrate the sequence of events following the onset of the 780 Ma Laurentian basaltic province. The eruptive phase is assumed to start at 780 Ma and to last an arbitrary 500 000 yr, with a total degassing of 1.6×10^{18} mol (equivalent to the Deccan trap degassing [33]). Global mean air temperature rises by about 6°C. Then CO_2 is rapidly consumed mainly through the weathering of the granitic lithologies (Fig. 4). In order to finally produce a snowball glaciation, we fix the area of the basaltic floods at a constant 6.2×10^6 km^2, which is slightly larger than the required 6.0×10^6 km^2 (Fig. 3). The traps are then assumed to drift southward at a speed of 2 cm/yr, based on paleogeographic reconstructions (Figs. 1 and 2) during 30 Myr. Then the southward drift is accelerated to a speed of 8 cm/yr to simulate the opening of the Proto-Pacific Ocean beginning at 750 Ma. All along the course of the drift, the consumption of CO_2 by the weathering of the Laurentian basaltic province increases, until it becomes predominant (Fig. 4). The 'snowball' glaciation is suddenly initiated once the threshold PCO_2 of 135 ppmv is reached through consumption by basalt weathering (145 ppmv consumed, 1.8 times the difference between the last glacial maximum and the pre-industrial PCO_2), coeval global air temperature collapses about 50 Myr after the trap eruption, and CO_2 consumption through weathering stops (Fig. 4).

The assumption of the constancy in the size of the basaltic province during 50 Myr might appear excessive, since weathering will slowly destroy the province. This argument can be bypassed assuming that the original size of the province was somewhat greater. The important point is the remaining size of the province when entering the equatorial area. Second, the Laurentian basaltic province is not the only magmatic province contributing to CO_2 consumption prior to the snowball event (Fig. 1). Due to the lack of constraints, these other contributors are not included in the model simulation. Finally, weathering of the Laurentian basaltic province might be gradually reduced through the development of thick soils at the multi-million-year timescale. However, the weathering law we are using for basalts [16] was established on a large variety of basaltic catchments, with age ranging from 65 Myr to present-day (including Deccan, Columbia River, Kamchatka, La Réunion Island, ...). This law thus integrates old terrains, as well as fresh basaltic surfaces. Furthermore, physical erosion is sustaining chemical erosion through the removal of soils, and is also integrated in the Dessert et al. law, as previously mentioned.

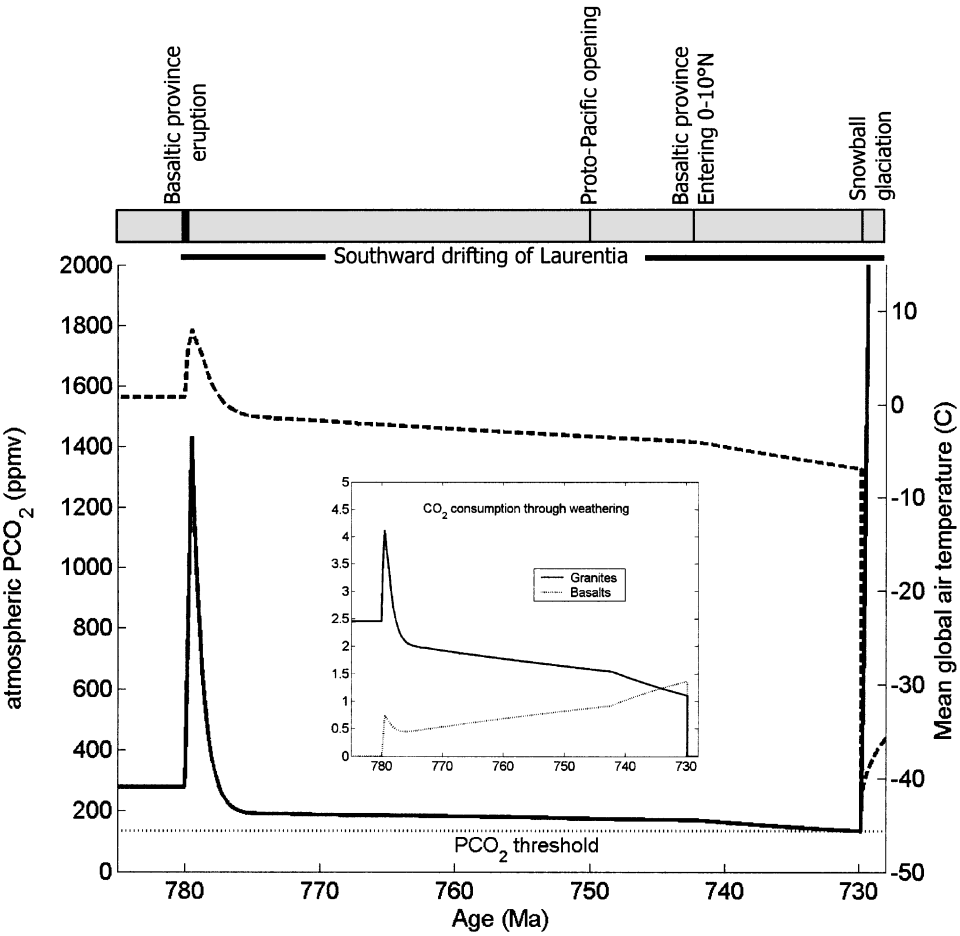

Fig. 4. Illustration of the time evolution of PCO_2 (solid line, scale on the left) and mean global air temperature (dashed line, scale on the right) in the aftermath of a trap formation event occurring at 780 Ma, as simulated by a global carbon cycle model coupled to a 1D climatic model. The 'snowball' glaciation is suddenly initiated once the threshold PCO_2 of 135 ppmv is reached through consumption by fresh basalt weathering, and coeval global air temperature collapses about 50 Myr after the eruption of the flood basalt. The included graph displays the evolution of the consumption of CO_2 (10^{12} mol/yr) through weathering of the granitic lithologies (solid line), and of the Laurentian magmatic province (dotted line).

4. Discussion

4.1. Neoproterozoic paleoenvironment

Weathering of basaltic traps might have contributed to the onset of the Sturtian snowball Earth, under the two following conditions: (1) the flood basalt province must be located close to the equator, a plausible location after the breakup of Rodinia, rifting and southward drift of the Laurentia plate, due to the formation of the Proto-Pacific Ocean; (2) the pre-weathering climate was already rather cool. For instance, the mean global air temperature corresponding to a PCO_2 of 280 ppmv is only equal to +0.8°C in the Neoproterozoic conditions.

Regarding the second point, it is highly probable that the snowball glaciation occurred in an already favorable climatic setting, independent of any proposed scenario (the methane hypothesis relies on a pre-perturbation PCO_2 of 300 ppmv [10]). We suggest two processes able to consume atmospheric CO_2 and to cool the global climate on the long term, prior to the trap onset, degassing and weathering that drove the Earth into the snowball glaciation. A first process able to cool the global climate through several steps might be the successive onset and weathering of the various basaltic provinces erupted from 825 to 755 Ma, as soon as these basalts are exposed to the appropriate climatic conditions. Secondly, the breakup of the Rodinia supercontinent has strongly reduced the continentality above the land masses. The corresponding increase in continental runoff has resulted in an enhanced consumption of atmospheric CO_2 and a long-term cooling of the global climate during the breakup process. From coupled 2.5D climatic–geochemical model simulations, we expect the partial pressure of CO_2 in the atmosphere to be reduced by possibly more than 1000 ppmv during the Rodinia breakup (Donnadieu et al., in preparation), through decreased continentality.

Once the sea ice cover reaches 30° latitude [30] as a consequence of the global cooling triggered by the weathering of the basaltic trap, the rapid development of a complete ice cover will reduce and then stop the weathering process ('snowball' glaciation). Thereafter, CO_2 slowly accumulates in the atmosphere due to volcanic degassing until the ensuing greenhouse effect will be able to trigger a catastrophic thawing event. At that moment (when PCO_2 reaches about 10^5 ppmv [5]), the incompletely consumed basaltic traps will still be exposed at the surface, and might potentially drive the Earth's climate into an almost endless succession of glaciation–deglaciation cycles (until the trap will be fully consumed by weathering). However, this potential cyclicity is broken by the southward drifting of the Laurentia plate (Fig. 5). Assuming a conservative number of 5–8 cm/yr for the Laurentia drift rate, the basaltic traps will reach the southern dry tropical divergence zone within about 25–40 Myr, after the onset of the snowball glaciation. If the background CO_2 degassing does not change after the onset of the glaciation, it will take 9–20 Myr to build up the 10^5 ppmv of CO_2 within the atmosphere required to melt the ice cover [5] (assuming the various pre-trap PCO_2 and background degassing rates of this study). By that time, the basaltic trap responsible for the glaciation will be approaching the southern tropical divergence zone, where precipitation regime and hence runoff strongly decrease. The southward drift of the trap will reduce the CO_2 consumption by weathering, preventing the Earth from cyclic snowball glaciations.

4.2. Carbonate and seawater isotope signatures

The onset of several basaltic provinces between 825 and 750 Ma accompanied by massive degassing of mantle carbon ($\delta^{13}C = -5$‰) into the exospheric system will force the seawater $\delta^{13}C$ to decrease prior to the onset of the snowball. This pre-snowball decrease is observed in many carbonate sections all over the world for the younger Marinoan glaciation (the amplitude of the negative excursion reaches 6–10‰ [34]). However, a smaller excursion is reported prior to the Sturtian glaciation (Chuos glaciation, in Namibia [6], displaying a negative excursion of about 3‰ only). A rough estimation of the $\delta^{13}C$ decrease linked to a Sturtian basaltic province onset might be performed assuming a trap degassing similar to the estimated degassing for the Deccan traps (reconstructions are not available for other trap onsets). The Deccan traps degassed about 1.6×10^{18} mol of CO_2 [33], which represents 50% of the present-day ocean–atmosphere carbon content. Since the pre-trap seawater $\delta^{13}C$ was close to 5‰ Pee Dee Belemnite (PDB) [6], it might have been reduced by about 3.3‰ in the direct vicinity of a basaltic province onset and degassing (if the degassing was fast at the geological timescale), in agreement with the sparse available data [6].

Finally, the persisting low seawater $^{87}Sr/^{86}Sr$ observed for the 800–700 Ma period [35,36] might be the result of the intense weathering of the basaltic provinces heralding the Rodinia breakup, since the basalts most probably displayed $^{87}Sr/^{86}Sr$ close to the mantle value.

4.3. The Marinoan 'snowball' glaciation

The trap-weathering scenario is dependent on the distribution of the continental masses and on the global tectonic regime. For this reason, the mechanism leading to the Marinoan/Varan-

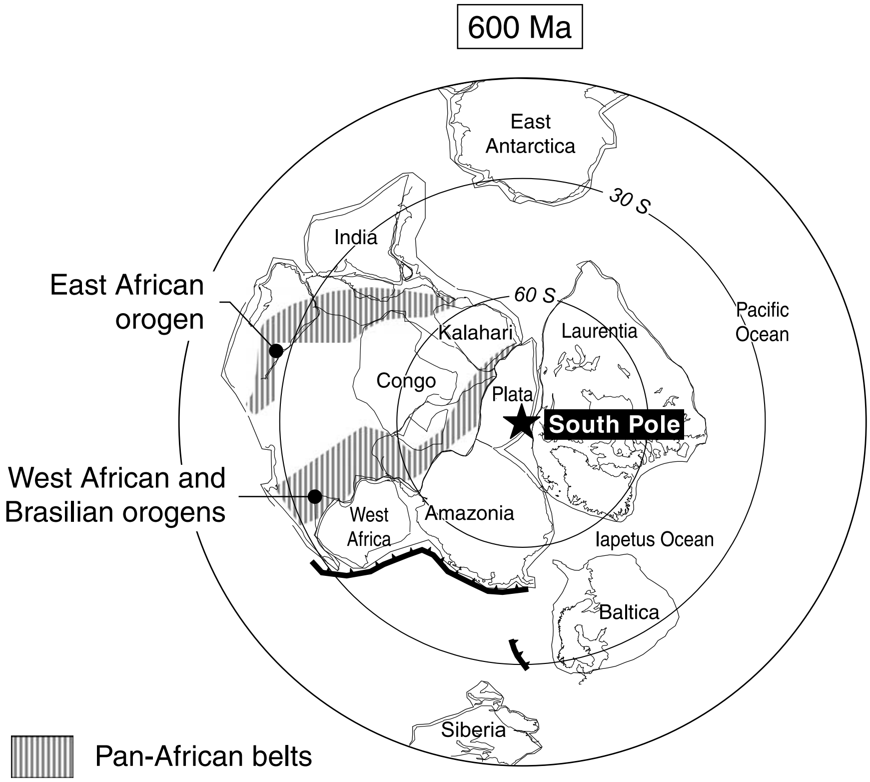

Fig. 5. Continental reconstruction at around 600 Ma after Meert et al. [39] and Cawood et al. [51], with location of Proto-Iapetus rift and main Pan-African orogenic belts (dashed areas).

gian ice age cannot be the same. At around 600 Ma, most continental masses were moving in the southern hemisphere to be assemblied in another supercontinent (Pannotia [37] or Gondwana [38, 39]) and continental collisions were building the two Panafrican orogenic belts of western and eastern Africa [40,41] (Fig. 5). There is no material to compare the geodynamic settings corresponding to the older (Sturtian/Rapitan) glaciogenic deposits on the western side of Laurentia and the younger ones (Marinoan/Varanger/Ice brook) on the eastern side (present-day coordinates), as proposed by Young [42], at least for the following two reasons. First, the older glaciation follows the breakup of Rodinia and formation of the Proto-Pacific Ocean by a few tens of Myr, whereas the formation of the Proto-Iapetus Ocean is roughly synchronous [43] or slightly postdates [44] the younger glaciation. Second, and more important, the Iapetus rifting occurs at high latitudes (Fig. 5), hence in low-weathering conditions. Models of the Marinoan glaciation should discuss the influence of orogeny [45] and polar continentality [46], and are likely to be more akin to the cases of the Phanerozoic glaciations. For instance, and for the analog case of the Himalayan orogeny, recent studies [45,47] suggest an increased organic carbon burial related to enhanced sedimentation rate driven by enhanced erosion, storing exospheric carbon within the sediments and cooling the global climate. Remarkably, this enhanced organic carbon burial is coeval with a decrease in Cenozoic seawater $\delta^{13}C$ due to the rapid decline in the carbon isotopic fractionation by marine photosynthesis triggered by decreasing CO_2 partial pressure. A scenario relying on enhanced organic carbon burial is thus not contradictory to the observed decrease in seawater $\delta^{13}C$ prior to the Marinoan glacial event.

4.4. The opening of the Atlantic

A comparison of the Sturtian event with the climatic consequences of the opening of the Atlantic Ocean is indeed more appropriate, as this rifting began with the formation of the Central Atlantic Magmatic Province (CAMP) [48,49], erupted 200 Myr ago with a reconstructed initial size of 7.0×10^6 km^2, and occurred at low to moderate paleolatitudes. However, other conditions had changed and were against the comeback of a new snowball Earth: the solar constant was near its present-day value (98.5%). We calculate that the surface of an equatorial magmatic province required to drive the Earth into a snowball glaciation at the Triassic–Jurassic boundary would be in excess of 20×10^6 km^2, far beyond the estimated size of the CAMP. Furthermore, the plants had spread all over the continental realm at that time, thereby reducing its albedo. Conversely, traces of older ice ages (older than Neoproterozoic) are rare, because of a CO_2-rich atmosphere largely compensating for the faint young sun, with the exception of the Huronian glaciation.

5. Conclusion

The proposed scenario remains rather speculative. However, the trap-weathering scenario links 'naturally' the breakup of the Rodinia supercontinent with the onset of the snowball glaciation. The Sturtian snowball Earth is due to a unique conjuction of internal forces (breakup of an equatorial supercontinent) and external conditions (lower solar constant and higher continental albedo with respect to present-day values). Neither before nor after did this specific conjunction happen again. Furthermore, it allows a simple explanation for the observed pre-snowball decrease in seawater δ^{13}C, without invoking ad hoc methane emission from the sediments. Further modeling studies (including the organic carbon cycle) are required to fix the amplitude of the negative δ^{13}C excursion linked to trap degassing.

However, a similar trap scenario cannot explain the onset of the Marinoan snowball glaciation, essentially since most of the continental masses are located close to the South Pole, where temperature and runoff are amongst the lowest, strongly inhibiting the consumption of CO_2 by weathering. A causal link between the Pan-African orogenesis and the global cooling of the climate should be explored instead.

'Fire' would then have triggered the Sturtian glaciation. Thus the weathering of the fresh basaltic surface might have consumed enough atmospheric CO_2 to plunge the Earth into a snowball event. As a conclusion, internal forces that drove the Earth into a snowball (plume-related eruption) might also have been responsible for termination of the glaciation, through CO_2 accumulated by long-term degassing and the drifting of tectonic plates.

Acknowledgements

The thorough review of an earlier version of the manuscript by V. Courtillot is greatly acknowledged. This study has been financially supported by the CNRS-ECLIPSE program. L.M.F. is supported by the Belgian Fund for Scientific Research (FNRS). C. Cavaré-Hester is thanked for the drawings.***[VC]***

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