GEOLOGY | THE GEOLOGICAL SOCIETY OF AMERICA®

https://doi.org/10.1130/G51669.1

Manuscript received 22 August 2023
Revised manuscript received 8 January 2024
Manuscript accepted 17 January 2024

Published online 7 February 2024

Duration of Sturtian "Snowball Earth" glaciation linked to exceptionally low mid-ocean ridge outgassing

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ABSTRACT

The Sturtian "Snowball Earth" glaciation (ca. 717–661 Ma) is regarded as the most extreme interval of icehouse climate in Earth's history. The exact trigger and sustention mechanisms for this long-lived global glaciation remain obscure. The most widely debated causes are silicate weathering of the ca. 718 Ma Franklin large igneous province (LIP) and changes in the length and degassing of continental arcs. A new generation of two independent Neoproterozoic full-plate tectonic models now allows us to quantify the role of tectonics in initiating and sustaining the Sturtian glaciation. We find that continental arc length remains relatively constant from 850 Ma until the end of the glaciation in both models and is unlikely to play a role. The two plate motion models diverge in their predictions of the timing and progression of Rodinia break-up, ocean-basin age, ocean-basement depth, sea-level evolution, and mid-ocean ridge (MOR) carbon outflux. One model predicts MOR outflux and ocean basin volume–driven sea level lower than during the Late Cenozoic glaciation, while the other predicts outgassing and sea level exceeding those of the Late Cretaceous hothouse climate. The second model would preclude a major glaciation, while the first model implies that the trigger for the Sturtian glaciation could have been a combination of an extremely low MOR outflux (∼9 Mt C/yr) and Franklin LIP weathering. Such minimal outflux could have maintained an icehouse state for 57 m.y. when silicate weathering was markedly reduced, with a gradual build-up of MOR CO_2 in the atmosphere paired with terrestrial volcanism leading to its termination.

INTRODUCTION

The low-paleolatitude Sturtian glaciation (ca. 717–661 Ma) represents the longest and most extreme period of icehouse climate in Earth's history, and the mainstay of the "Snowball Earth" hypothesis in which the entire global ocean surface becomes frozen via a runaway ice-albedo feedback (Hoffman et al., 2017). Various factors have been implicated in initiating the glaciation (Walzer and Hendel, 2023), including an increase in planetary albedo caused by volcanic aerosols (Macdonald and Wordsworth, 2017) and high obliquity of the ecliptic (Williams, 2008). However, enhanced continental silicate weathering and organic burial linked to the breakup of Rodinia are viewed as the main mechanisms drawing down atmospheric CO_2 and driving planetary cooling (e.g., Goddéris et al., 2003; Donnadieu et al., 2004; Cox et al., 2016; Hoffman et al., 2017). The emplacement of the Franklin large igneous province (LIP; Ernst et al., 2021; Fig. 1) has been of particular interest because its rapid weathering at tropical latitudes may have triggered the Sturtian glaciation (Goddéris et al., 2003; Cox et al., 2016). Precise dating of the Franklin LIP at 718 Ma now places it immediately prior to the Sturtian glaciation (Dufour et al., 2023); however, its duration of ∼2 m.y. (Pu et al., 2022; Dufour et al., 2023) likely limited the supply of fresh mafic rock surface for weathering due to regolith development (Park et al., 2021). The role of the Franklin LIP as the sole trigger of the Sturtian glaciation has been further questioned as the majority of tropical Phanerozoic LIPs did not drive icehouse climates (McKenzie et al., 2016; Park et al., 2021), and there is no significant LIP associated with the Marinoan glaciation (Ernst et al., 2021). In addition, numerical box modeling by Defliese (2021) suggests that the continental and seafloor weathering feedback mechanisms become insignificant at extremely low temperatures, rendering the drawdown of CO_2 insufficient to maintain a "Snowball Earth" glaciation. Defliese (2021) hypothesized that the long duration of the Sturtian glaciation was sustained by a consistently low crustal production and mid-ocean ridge (MOR) CO_2 outgassing rate. This variable provides a critical input parameter for any model designed to understand the mechanisms driving Cryogenian glaciations but is unconstrained. We are now able to place quantitative bounds on MOR outgassing rates using a new generation of two independent full-plate models (Me21 [Merdith et al., 2021] and Li23 [Li et al., 2023b]; Fig. 1) that capture Neoproterozoic plate boundary evolution. We also evaluate both plate models in terms of their compatibility with the Sturtian glaciation and show that crustal production and MOR outflux played a key role in Cryogenian cooling of Earth.

METHOD

We use GPlately (https://github.com/GPlates/gplately; Mather et al., 2023) to compute MOR length, mean global spreading rates, and crustal production (Figs. 2A–2C) for plate motion models Me21 (Merdith et al., 2021) and Li23, the preferred model of Li et al. (2023b). Melt beneath MORs dissolves mantle carbon, which partly degasses during seafloor spreading. The MOR carbon outflux (Fig. 2D) is computed following Keller et al. (2017). Sea level is computed by producing grids of the age-area distribution of ocean crust through time based on the plate topologies and rotations in each plate model using the method of Williams et al. (2021). The paleo-age grids are then translated

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CITATION: Dutkiewicz, A., et al., 2024, Duration of Sturtian "Snowball Earth" glaciation linked to exceptionally low mid-ocean ridge outgassing: Geology, v. XX, p. , https://doi.org/10.1130/G51669.1

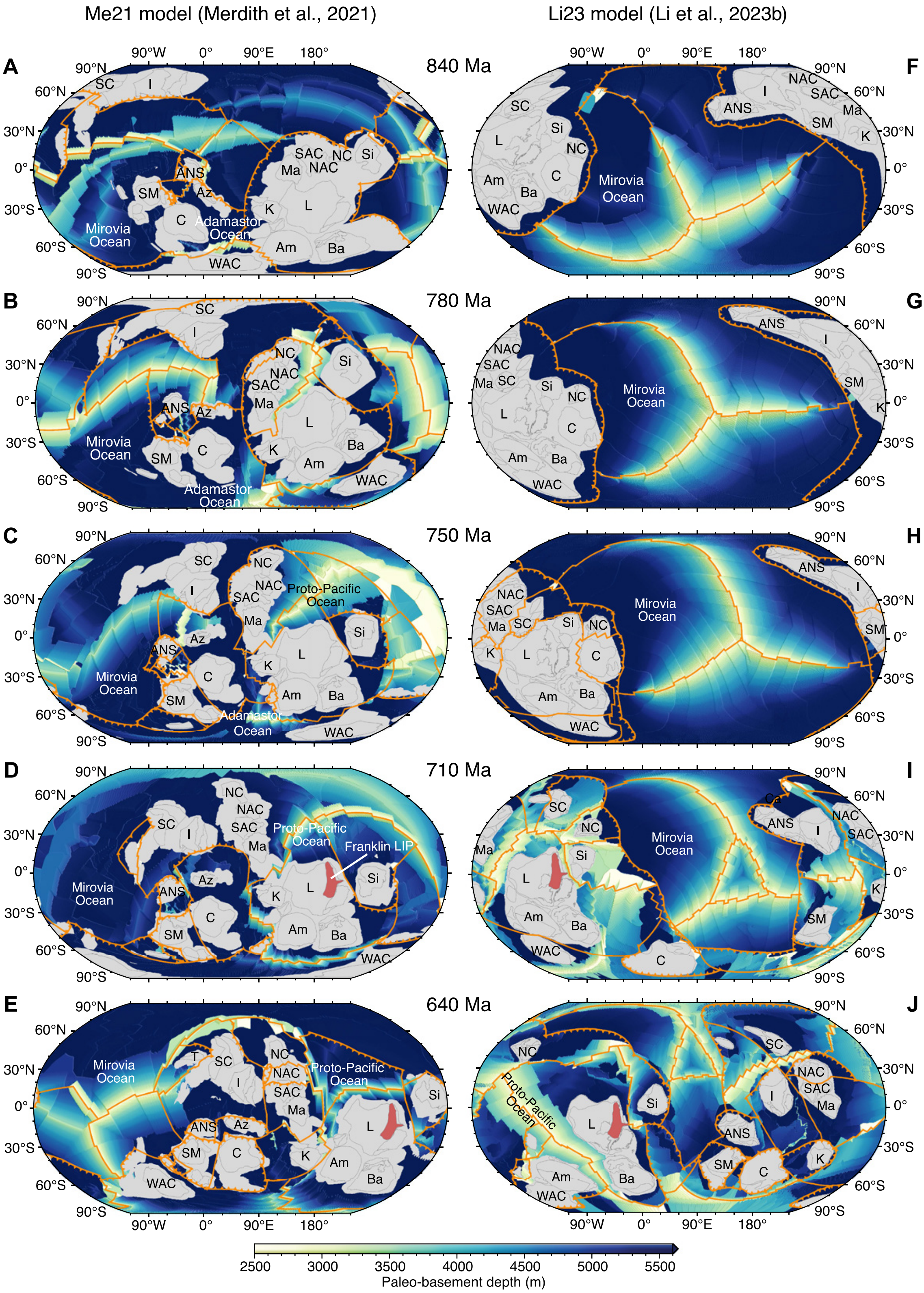

Figure 1. Rodinia reconstructions using the plate motion models Me21 (Merdith et al., 2021) (A–E) and Li23 (Li et al., 2023b) (F–J) overlying basement depth (see text). Orange lines are mid-ocean ridges, orange lines with teeth are subduction zones, and the red polygon is the Franklin large igneous province (LIP; Ernst et al., 2021). Cratonic crust (dark gray outlines) is annotated as: Am—Amazonia; ANS—Arabian-Nubian Shield; Az—Azania; Ba—Baltica; C—Congo; I—India; K—Kalahari; L—Laurentia; Ma—Mawson; NAC—North Australian Craton; NC—North China; SAC—South Australian Craton; SC—South China; Si—Siberia; SM—Sahara Metacraton; WAC—West African Craton. Light gray areas outside of the dark gray outlines are approximate extents of continental crust.

to paleo-basement depth grids using the age-depth relationship of Richards et al. (2018). We disregard other processes contributing to global sea level to focus on a first-order comparison of the implication of two alternative plate models and implied oceanic paleo-depth distributions on global sea-level change (Fig. 2E).

PLATE MOTION MODEL COMPARISON

The two available full-plate models for the Neoproterozoic are based on diverse data; however, Me21 (Merdith et al., 2021) emphasizes geological data while Li23 (Li et al., 2023b) favors paleomagnetic data, resulting in starkly different plate motion and plate boundary evolution histories (Fig. 1; Table S1 and Videos S1 and S2 in the Supplemental Material[1]). The

[1]Supplemental Material. Figure S1, Table S1, and Videos S1 and S2. Please visit https://doi.org/10.1130/GEOL.S.25016264 to access the supplemental material; contact editing@geosociety.org with any questions.

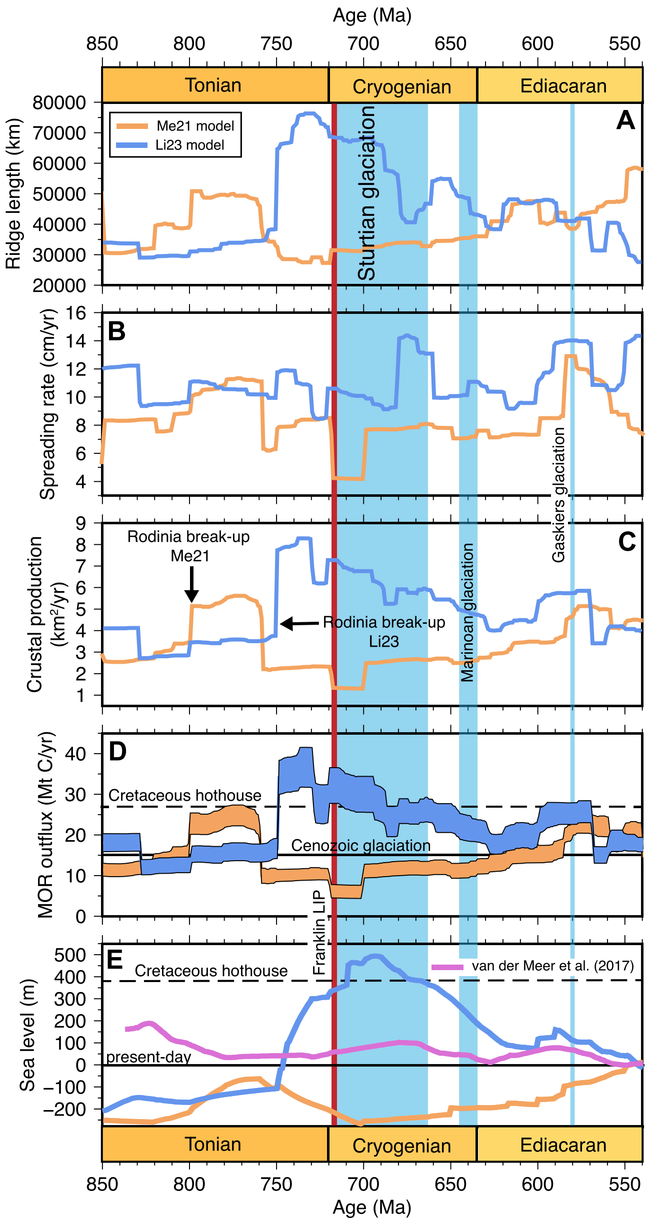

Figure 2. Tectonic parameters for plate motion models Me21 (Merdith et al., 2021) and Li23 (Li et al., 2023b). Blue bars indicate glacial intervals with ages from Walzer and Hendel (2023). Red bar is the Franklin large igneous province (LIP) with emplacement age from Dufour et al. (2023). (A) Ridge length. (B) Spreading rate. (C) Crustal production rate (ridge length × spreading rate). (D) Mid-ocean ridge (MOR) (outflux of CO_2 expressed as weight/yr of carbon). The error envelope primarily reflects the uncertainty in upper mantle CO_2 content (Müller et al., 2022). Mean MOR outflux for the onset of the last glaciation at ca. 34 Ma (black line) and for the Cretaceous hothouse climate at 120 Ma (black dashed line) are from Müller et al. (2022). (E) Eustatic sea level computed as a function of the difference in mean ocean-basin depth relative to the present-day (black line), including the effect of isostasy (see text). Black dashed line indicates sea level at 120 Ma based on the same method (see Fig. S1 in the Supplemental Material [see text footnote 1]). The van der Meer et al. (2017) curve is derived from seawater $^{87}Sr/^{86}Sr$ ratios.

models are especially different in the configuration and dispersal of Rodinia, which plays a key role in understanding the causes of Cryogenian glaciations (e.g., Cox et al., 2016; Hoffman et al., 2017; Li et al., 2023b). In Me21, Rodinia excludes several continents separated by ocean basins (Fig. 1A), with an initial break up at ca. 800 Ma (Merdith et al., 2021). This results in a modest lengthening of the global MOR system from ~40,000 km to ~50,000 km at 800 Ma (Fig. 2A) and an increase in crustal production rate from ~3.5 km^2/yr to ~5 km^2/yr (Fig. 2C). A global plate reorganization at ca. 760 Ma (Fig. 1C) causes a shortening of the global plate boundary system, which is reflected in the subduction and/or cessation of some spreading ridges and the transition of several subduction zones to passive margins (see Merdith et al., 2021, for details). Subduction along the South China block and northern India and along much of the western side of Rodinia stops at this time (Fig. 1E). Another major tectonic reorganization occurs at 720 Ma, when both subduction and spreading systems surrounding the North China block, as well as the North and South Australian cratons, become extinct, resulting in a shortening of the Mirovia MORs (Fig. 1D) and a reduction in crustal production (Fig. 1C; Collins et al., 2021).

In contrast, in Li23, all continents are assembled into Rodinia, which is encircled by the Mirovia Ocean (Figs. 1F–1H). The initiation of Rodinia's dispersal at ca. 750 Ma (Fig. 1H) results in a doubling of the MOR length from 40,000 km to 80,000 km (Fig. 2A), as well as the crustal production rate from ~4 km^2/yr at 750 Ma to ~8 km^2/yr at 740 Ma (Fig. 2C). Plate reorganization, including the separation of Siberia from Laurentia at ca. 720 Ma (Li et al., 2023b), leads to a modest decrease in ridge length to ~65,000 km (Fig. 1A) and a high rate of crustal production of ~7 km^2/yr (Fig. 1C) at the onset of the Sturtian glaciation. Crustal production remains relatively high at 6–7 km^2/yr in Li23 throughout the Sturtian glaciation until ca. 620 Ma, while in Me21, crustal production remains low at ~2.5 km^2/yr until ca. 620 Ma (Figs. 1A and 1C), when the fragmentation of plates reaches a post-Rodinia break-up maximum.

SOLID EARTH OUTGASSING

The "Snowball Earth" model–driven hypothesis proposes a shutdown of the hydrological cycle during the Cryogenian glaciations (Hoffman et al., 2017). However, abundant and diverse geological evidence, such as hummocky cross-stratification generated by storm waves (Le Heron, 2015; Qi et al., 2023), complex communities of microbiota (Moczydłowska, 2008), and non-glacial sediments and sedimentary structures (Allen and Etienne, 2008) in Sturtian glacial successions, indicates the presence of open marine water and a functioning hydrological cycle (Le Heron, 2015; Spence et al., 2016; Lloyd et al., 2023). The presence of open water facilitates a slow and continuous exchange of CO_2 between the ocean and the atmosphere, modulating climate and weathering reactions as part of the long-term carbon cycle (Berner, 2004). Carbon dioxide is constantly outfluxed from the solid Earth, chiefly from degassing of MORs during seafloor spreading and outgassing from volcanic arcs (Müller et al., 2022). The build-up of atmospheric CO_2 is, in part, regulated by chemical weathering of silicates on continents and in seafloor basalt (Brantley et al., 2023). This temperature-dependent negative feedback mechanism draws down CO_2 and buffers Earth's climate over geological timescales (Berner, 2004). During a global glaciation, when silicate weathering becomes negligible at extremely low temperatures (Defliese, 2021) and high aridity (Brantley et al., 2023), the uptake of CO_2 by weathering is severely reduced, allowing that greenhouse gas to build up in the atmosphere, leading to a warmer climate. We propose that in order for the Sturtian glaciation to be sustained for 57 m.y., the reduction in the drawdown of atmospheric CO_2 must have been balanced by an exceptionally low solid Earth carbon outflux for the duration of the glaciation.

The MOR outflux (Fig. 2D) mirrors the rates of crustal production (Fig. 2C), with the Me21 and Li23 models diverging significantly during the Sturtian glaciation when the rates of MOR outflux are high in Li23 and low in Me21 (Fig. 1D). The outflux in Me21 shows a stepwise decrease from a maximum of ~25 Mt C/yr at ca. 770 Ma, to ~12 Mt C/yr between 760 and 720 Ma, reaching a Neoproterozoic minimum of ~9 Mt C/yr at the onset of the Sturtian glaciation. This minimum persists for 20 m.y., with a slight increase to ~12 Mt C/yr after 700 Ma, which remains relatively constant until the end of the Sturtian glaciation, gradually increasing to a maximum of ~23 Mt C/yr at ca. 570 Ma (Fig. 1D). In contrast, the Li23 model shows a much earlier MOR outflux minimum of ~16 Mt C/yr at ca. 770 Ma. This value increases dramatically to a Neoproterozoic maximum of ~37 Mt C/yr at ca. 740, remaining very high (~32–28 Mt C/yr) throughout the duration of the Sturtian glaciation and decreasing toward a minimum of ~15 Mt C/yr at ca. 565 Ma (Fig. 1D). The peak MOR outflux in Li23 during the Sturtian far exceeds the mean of ~12 Mt C/yr computed for the Cenozoic glaciation and even the Cretaceous hothouse maximum of ~27 Mt C/yr following the break-up of Pangea (Müller et al., 2022; Fig. 1D). This comparative analysis suggests that the MOR outgassing history implied by the Li23 model is inconsistent with a Cryogenian icehouse climate and would preclude a widespread glaciation. We propose that the MOR outflux from Me21, which is lower than Cenozoic icehouse estimates, would be sufficiently small to help trigger and sustain a global glaciation.

Outgassing along continental arcs is another potentially important driver of long-term climate (McKenzie et al., 2016). Continental arc CO_2 emissions depend on the outflux of carbon from the subducting plate (in the lithospheric mantle, crust, and sediments) into the sub-arc mantle (Müller et al., 2022), and metamorphic decarbonation of carbon-bearing rocks (e.g., carbonate platforms) in the overriding plate (Mason et al., 2017). These components have evaded quantification for pre-Phanerozoic time because relicts of these reservoirs are rarely preserved in the geological record. McKenzie et al. (2016) suggested that Cryogenian glaciations were triggered by a major drop in CO_2 fluxes from continental volcanic arcs associated with the assembly of Rodinia, with arc length and CO_2 increasing during Rodinia break-up. A similar result by Mills et al. (2017) was based on total global subduction length. Neither approach considered changes in convergence rates. However, analogous to MOR outflux, we expect arc outgassing to roughly track crustal destruction rates, which are the product of the arc length and convergence rates (Fig. 3). We find that the continental arc length in both Me21 and Li23 models remains relatively constant from 850 Ma until the end of the Sturtian glaciation at 661 Ma (Fig. 3A). The maximum arc length increases post–660 Ma in both models, doubling from ~20,000 km to 40,000 km in Li23, with a transient increase from 20,000 km to 27,000 km in Me21 (Fig. 3A). However, at the same time, plate convergence and subduction rates decrease dramatically in Li23 and remain low in Me21 (Figs. 3B and 3C). This indicates that despite an increase in continental volcanic arc length, subduction-related CO_2 outflux postdating 660 Ma was likely modest because crustal destruction rates are low in both models (Fig. 3C). In summary, neither the onset nor the termination of Cryogenian glaciations appear to have been induced by changing continental arc length.

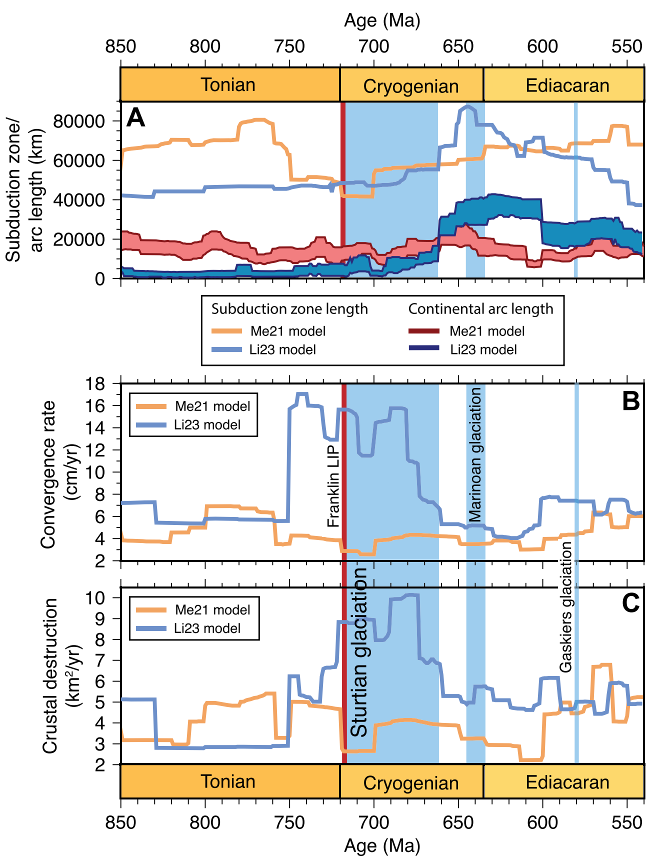

Figure 3. Global subduction parameters for the plate motion models Me21 (Merdith et al., 2021) and Li23 (Li et al., 2023b). (A) Subduction zone length and continental arc length. Minimum and maximum continental arc lengths are based on arc-trench distances of 250 km and 450 km, respectively, following Pall et al. (2018). (B) Plate convergence rate. LIP—large igneous province. (C) Crustal destruction rate. Labeling as in Figure 2.

NEOPROTEROZOIC EUSTATIC SEA LEVEL

Our modeled sea-level curves, which are solely based on changes in mean oceanic basement depth (Fig. 2E), are starkly different between Me21 and Li23. Sea level is similar for both plate models for much of the Tonian but diverges markedly at ca. 750 Ma when it drops by ~150 m in Me21 due to a shortening of the global plate boundary system (Fig. 1) and rises by ~550 m in Li23 due to the formation of new MORs. The Sturtian glaciation is marked by a sea-level minimum in Me21 and a sea-level maximum in Li23. The sea-level curves slowly converge in the Ediacaran, following a major decrease in sea level in Li23 and a small increase in Me21 (Fig. 1E). The Sturtian sea-level peak in Li23 is ~100 m higher than the mid-Cretaceous hothouse sea level computed with the same method (Fig. 2E; Fig. S1) when solid-Earth CO_2 degassing was at a maximum following Pangea break-up (Müller et al., 2022). Such a climate would be the consequence of the Li23 plate model evolution, inhibiting a global glaciation. A sea-level curve derived from the seawater $^{87}Sr/^{86}Sr$ record (van der Meer et al., 2017), largely based on pre/post glacial strata (Spence et al., 2016), is broadly similar in shape to the Me21 sea level. It lacks the Sturtian maximum evident in Li23, providing an independent, but not well-constrained, check of the models.

CONCLUSIONS

Our computations suggest that the low solid-Earth outflux during the Sturtian glaciation based on the Me21 model would be sufficient to sustain a global glaciation in the presence of open water and significantly reduced silicate weathering. We envisage that the initial driving mechanism for the Sturtian glaciation was a combination of Franklin LIP weathering and extremely low MOR outflux. The termination of the glaciation may have been triggered by a gradual build-up of MOR CO_2 in the atmosphere combined with deglacial terrestrial volcanism (Li et al., 2023a). Solid-Earth degassing alone does not explain the drivers or the termination of the Marinoan glaciation (Fig. 2); however, multiple poorly constrained carbon sinks (e.g., seafloor alteration, organic carbon flux), sources (e.g., rift volcanism; Lan et al., 2022), and climate-modulating orbital forcing (Benn et al., 2015) may also be important.

ACKNOWLEDGMENTS

We thank three anonymous reviewers for their constructive comments. Funding for this work includes Australian Research Council (ARC) Future Fellowship FT190100829 (A. Dutkiewicz), ARC DECRA

Fellowship DE230101642 (A. Merdith), ARC Linkage grants LP210200822 and LP200301457 (A. Collins), ARC Discovery grant DP200100966 (B. Mather), ARC DECRA Fellowship DE210100084 (S. Zahirovic), and the National Collaborative Research Infrastructure Strategy (NCRIS) via AuScope (L. Ilano and R.D. Müller).

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Printed in the USA