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The sea-level fingerprint of a Snowball Earth deglaciation

Jessica R. Creveling [a,*], Jerry X. Mitrovica [b]

[a] Geological and Planetary Sciences, California Institute of Technology, 1200 E. California Blvd., Pasadena, CA 91125, USA
[b] Earth and Planetary Sciences, Harvard University, 20 Oxford Street, Cambridge, MA 02138, USA

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ABSTRACT

Cap dolostones are thought to represent deposition from seas transgressing over formerly glaciated continental margins during Marinoan Snowball deglaciation. Nevertheless, facies associations within some cap dolostones indicate that an episode of regional regression punctuated these transgressive sequence tracts. To date, inferences of sea-level change during and after the Marinoan Snowball deglaciation have been interpreted using simple, qualitative arguments. In the present study, we explore the full spatio-temporal variability of sea-level change during Snowball deglaciation and its aftermath using a gravitationally self-consistent theory that accounts for the deformational, gravitational and rotational perturbations to sea level on a viscoelastic Earth model. The theory is applied to model Marinoan Snowball deglaciation on a generalized Ediacaran paleogeography with a synthetic continental ice-sheet distribution. We find that sea-level change following a synchronous, rapid (2 kyr) collapse of Snowball ice cover will exhibit significant geographic variability, including site-specific histories that are characterized by syn-deglacial sea-level fall or stillstand. Moreover, some sites that experience syn-deglacial transgression will continue to experience transgression in the post-deglacial phase. Taken together, these results suggest that sea-level change recorded by strata capping Snowball glaciogenic units may reflect a more complicated trajectory than previously thought, including deposition that was not limited to the deglaciation phase. These simulations, as well as others that explore the response to asynchronous melting and deglaciation phases of longer duration, demonstrate that an episode of regional regression interrupting a cap dolostone transgressive sequence tract may reflect one of several processes (or their combination): (1) near field adjustment associated with rapid local melting during an otherwise global hiatus in deglaciation; (2) post-glacial uplift of sites during a period of slowing deglaciation, and (3) a transition, at some sites, from a sea-level fall dominated by post-glacial uplift to a phase of sea-level rise due to eustasy and peripheral bulge subsidence throughout an extended (order 50 kyr or greater) Snowball deglaciation.

1. Introduction

Glaciogenic diamictites interpose low-latitude marine strata on nearly every Cryogenian–Ediacaran paleocontinent (Evans, 2000; Hoffman and Li, 2009; Hoffman, 2010; Li et al., 2013). 'Cap carbonates' sharply overlie most late Neoproterozoic Marinoan Snowball glacial deposits (Hoffman et al., 1998). Traditionally, a cap carbonate encompasses a basal dolostone unit and an overlying limestone unit (Hoffman et al., 1998); however, cap carbonates may also include siliciclastics within stratigraphic successions recording post-glacial sea-level change (e.g., the Brachina Formation, Australia, to name just one example; Rose and Maloof, 2010). Stratigraphic relationships indicate that cap dolostones represent the transgressive systems tract of a broader cap depositional sequence (Hoffman and Schrag, 2002). Further, the consistent vertical succession of sedimentary structures and composite $\delta^{13}C_{carb}$ chemostratigraphic data across reconstructed slope-to-platform paleoenvironments suggest a time-transgressive (diachronous) model for cap dolostone deposition (Hoffman et al., 2007; Rose and Maloof, 2010). Under the assumption that sea-level transgression requires contemporaneous deglacial melting, cap dolostones have been interpreted to record deposition over a time-scale confined to the Snowball deglaciation (Hoffman et al., 2007). Theoretical (e.g., Hoffman et al., 1998) and climate model-based (Hyde et al., 2000) predictions of Snowball deglaciation posit a melt timescale of 2–10 kyr; in contrast, paleomagnetic polarity reversals preserved within some cap dolostones imply deposition that lasted >100 kyr (Trindade et al., 2003; Kilner et al., 2005; Hoffman et al., 2007). By comparison to the bathymetric profile of modern carbonate platforms, Hoffman (2010) estimated an $\sim$1–1.5 km Marinoan glacioeustatic rise on the Namibian margin.

* Corresponding author. Tel.: +1 (626) 395 1783.
E-mail address: jcrevel@gps.caltech.edu (J.R. Creveling).

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Despite broad consensus that cap dolostones record post-glacial transgression, episodes of an early sea-level fall interrupting this transgression have been inferred from the stratigraphic transition from below wave-base limestone turbidites to low-angle, swaley and wave-ripple cross-stratified peloidal dolograinstone in cap dolostones on the Congo and Kalahari cratons (and, perhaps, Australia based on a reinterpretation of sedimentological descriptions of Kennedy (1996); Hoffman et al., 2007; Hoffman and Macdonald, 2010). In contrast (or, possibly, in addition) to sea-level fall in the cap dolostone of Australia, Rose et al. (2013) concluded that regression punctuated the transgressive syn-deglacial siliciclastic lithofacies of the underlying Elatina Formation, Flinders Range, Australia.

These studies attributed regression to the loss of gravitational attraction of the sea surface to local, waning ice sheets, although Rose et al. (2013) acknowledged the possibility of, and discussed the temporal implications for, isostatic rebound in contributing to a local sea-level fall. Moreover, Hoffman and Macdonald (2010) argued that a rapid and early sea-level fall would have decreased lithostatic pressure, contributing to pore-fluid over-pressurization and the subsequent formation of bed-parallel sheet-cracks filled with isopachous cements. Sheet-crack cements have been identified in the basal-meters of cap dolostones on multiple cratons (Hoffman, 2011), indicating, based on the model of Hoffman and Macdonald (2010), that regionalized early melt (and localized sea-level fall) preceded the eustatic transgression at each of the localities where this sedimentary structure appears. The logical corollary of these assumptions is that Marinoan ice-sheets vanished asynchronously, not in unison (Hoffman and Macdonald, 2010). However, many well-studied cap dolostone successions do not exhibit sedimentary evidence for sea-level fall at the base of cap dolostones, including some that host sheet-crack cements.

A number of questions arise from the above stratigraphic studies that have relevance for the interpretation of the geological record of Marinoan Snowball deglaciation. What is the plausible range of geographic variability in regional sea-level change driven by the deglaciation? Is this geographic variability a strong function of the duration of the deglaciation? Can a local geological inference of the magnitude of transgression provide a robust estimate of the globally averaged (eustatic) sea-level rise associated with the deglaciation? Finally, what specific circumstances could lead to a pronounced regional regression prior to, and perhaps also during, glacioeustatic transgression?

In this study, we explore the spatio-temporal variability of sea-level change driven by Marinoan Snowball deglaciation using a gravitationally self-consistent theory and numerical algorithm that accounts for the deformational, gravitational and rotational perturbations to sea level on a viscoelastic Earth model and time-dependent shoreline migration (Mitrovica and Milne, 2003; Kendall et al., 2005). Our numerical model of the Marinoan Snowball deglaciation is configured with an Ediacaran paleogeography and a synthetic continental ice-sheet distribution. Using the model, we explore the sensitivity of the predictions to variations in both the relative synchronicity of regional ice melting and the duration of the global deglaciation phase. In the discussion below, we frame our predictions in terms of the physics of sea-level change at 'near-field' versus 'far-field' localities. Sites in the near field of a specific region of ancient ice cover are located within ~1000–2000 km of the margin of the ice sheet, and far-field sites are located beyond this zone. (Within a pan-continental Snowball glaciation, sites in the far-field of all ice complexes would be located in oceans; nevertheless, this designation is useful when examining both broad-scale sea-level trends and melt-water volume balance.) Furthermore, in discussing time-evolving sea-level trends, we apply the terms 'syn-deglacial' when referencing time during a melting event and 'post-deglacial' to signify the time after the complete melt of global ice sheets; together, these define the 'post-glacial' interval.

Liu and Peltier (2013) describe a preliminary study of sea-level change associated with Snowball glaciation. Their analysis focused on the net sea-level fall across a multi-million year glaciation, i.e., after isostatic equilibrium is achieved in the glaciated state, and, as a consequence, the spatial variability they predict is muted (see their Fig. 12). In addition, they adopted a model of paleogeography at 570 Ma for the Marinoan (~635 Ma) Snowball event and assumed that the position of shorelines was fixed in time (i.e., coastlines were characterized by vertical cliffs). Notwithstanding these approximations, their analysis provides a (slow) glaciation-phase complement to the detailed spatio-temporal predictions of syn- and post-deglacial sea-level change described herein. The time-dependent crustal deformations, perturbations to the gravity field and rotational state, and eustatic sea-level variations that drive syn- and post-deglacial sea-level changes dictate the sequence stratigraphic architecture of the abundant and widely distributed cap successions and thus our numerical predictions provide an important framework for interpreting the geological record.

2. Modeling the Marinoan Snowball Earth deglaciation

The configuration of Ediacaran paleocontinents remains uncertain (compare, for instance, Torsvik, 2003; Meert and Torsvik, 2004; Hoffman and Li, 2009; and Li et al., 2013). Our simulations of sea-level change in response to Snowball deglaciation were guided by the 635 Ma paleogeographic reconstruction of Li et al. (2013), but for model simplicity we assume 12 paleocontinents that, in some cases, represent an amalgamation of multiple paleocratons (Fig. 1A). Each shaded region in Fig. 1A represents the areal extent of a continent. (Adopting a paleogeography in which the continents were fully assembled into a supercontinent shows relatively small differences in the calculations described below.) Global topography prior to the Snowball glaciation is prescribed as follows: the elevation of continental interiors is set to 850 m and decreases linearly to 0 m within 350 km of the shoreline; offshore, the bathymetry drops to −150 m over a distance of 80 km (the modeled continental shelf), to −2000 m over the next 30 km (the continental slope) and to −3800 m in the next 300 km (the continental rise) where it is maintained as the depth of the abyssal plain. These values are consistent with mean values in modern topography.

We model a pan-glacial scenario (Hoffman, 2009) in which ice covers all continents. Our ice model at maximum glaciation assumes an equilibrium (parabolic) thickness profile with maximum elevation proportional to the minimum width of the continent as determined from lines passing through the centroid (Fig. 1B). We adopt a proportionality constant between ice elevation and continental width such that the equivalent globally averaged (eustatic) sea-level (ESL) change associated with the maximum ice cover is 1 km, as estimated by Hoffman (2011). We assume that a single ice dome covers the following pairs of continents: B and C, E and F, G and H, and J and K (Fig. 1B). Guided by arguments in Hoffman (2011), the grounding line of each ice sheet extends to the edge of the continental shelf for the smallest ice sheets (as on continents A and L in Fig. 1A) and to 12 km down the continental slope (where local water depths are ~1000 m) for the largest ice sheets (as on continents E/F and J/K). An equilibrium ice thickness profile is maintained during all modeled deglaciation events.

In Fig. 1A, we show the approximate paleo-location of 20 cap dolostones (Li et al., 2013; Hoffman and Li, 2009; Table 1); while not exhaustive, these sites have a rich geological literature documenting the Marinoan Snowball deglaciation and are globally distributed, thus providing reference points for further discussion of sea-level change on a suite of paleo-continental margins. The paleogeography and site locations (Fig. 1A), as well as the maximum ice cover (Fig. 1B), pattern and duration of deglaciation, are subject