Spatial variation in Late Ordovician glacioeustatic sea-level change

Jessica R. Creveling a, *, Seth Finnegang, Jerry X. Mitrovica c, Kristin D. Bergmann d

a College of Earth, Ocean, and Atmospheric Sciences, Oregon State University, Corvallis, OR 97330, USA
b Department of Integrative Biology, University of California, Berkeley, CA 94720, USA
c Department of Earth and Planetary Sciences, Harvard University, Cambridge, MA 02138, USA
d Department of Earth, Atmospheric and Planetary Sciences, Massachusetts Institute of Technology, Cambridge, MA 02139, USA

A R T I C L E   I N F O
Article history:
Received 26 September 2017
Received in revised form 1 May 2018
Accepted 6 May 2018
Available online 24 May 2018
Editor: B. Buffett

Keywords:
Ordovician glaciation sea level glacial isostatic adjustment stratigraphy

A B S T R A C T
Mass extinction of Late Ordovician marine fauna closely coincided with southern hemisphere glaciation. The sequence stratigraphic architecture of shallow marine deposits informs estimates of glacioeustatic sea-level change at sites both proximal and distal to the reconstructed Ordovician ice sheet(s) and contemporaneous changes in ice volume. A recent correlation framework for the stratigraphic architectures of one near and one far field Late Ordovician margin concluded that the Late Ordovician glaciation encompassed multiple long-term cycles of ice volume growth and retreat with superimposed higher frequency cycles. Here we posit that—similar to Cenozoic glacial cycles—glacial isostatic adjustment can preclude synchronous and similar magnitude (or directional) changes in Late Ordovician sea level between ice proximal and ice distal locations and, hence, distort a globally correlative sequence stratigraphy. We explored whether long-duration (i.e., million year) Late Ordovician glacial cycles should produce a globally coherent, eustatic record of sea-level change between ice proximal and ice distal margins using a gravitationally self-consistent theory that accounts for the deformational, gravitational and rotational perturbations to sea level on a viscoelastic Earth model. We adopted a Late Ordovician paleogeography and a synthetic continental ice-sheet distribution and volume informed by the areal extent of glaciogenic deposits and geochemical records, respectively. We demonstrate that modeled million year Late Ordovician glacial cycles produce sea-level histories on near and far field margins that differ from eustasy, and from one another, due primarily to elastic flexure and associated gravitational effects. While predicted far-field sea-level histories faithfully preserve the temporal structure of modeled glacioeustasy, their amplitude may differ from eustasy by as much as 30–40%. The impact of glacial isostatic adjustment is largest at the margins of glaciated continents, and these effects can be of the same order of magnitude as the eustatic, and even induce a local sea-level rise during an episode of ice growth and eustatic sea-level fall, and vice versa. In this regard, stratal surfaces of maximum regression and flooding expressed at near-field margins need not reflect global (eustatic) trends in ice sheet growth and decay, respectively, and thus may not provide chronostratigraphic horizons for correlation with far-field sequence stratigraphic architectures.

© 2018 Elsevier B.V. All rights reserved.

1. Introduction

Continental ice sheets expanded across southern Gondwana during the Late Ordovician, and the attendant ocean–atmosphere cooling and glacioeustatic sea-level fall likely initiated the Late Ordovician Mass Extinction (Hallam and Wignall, 1999; Sheehan, 2001; Finnegang et al., 2012). Compelling sedimentological, stratigraphic and geochemical evidence for glaciation appears in sedimentary basins both proximal and distal to reconstructed Ordovician ice centers, yet these local records produce contrasting and, at times, conflicting reconstructions of the duration, areal extent, and volume of the Ordovician glaciation (Brenchley, 1994). The absence of pre-Hirnantian glaciogenic deposits contributes to the view that a short-lived glaciation began and ended within the Hirnantian Stage (445.2 ± 1.4–443.8 ± 1.5 Ma; Brenchley, 1994; Finney et al., 1999; Sutcliffe et al., 2000, 2001; Loi et al., 2010), or persisted locally into the Silurian Period (Landoyve Stage: 443.8 ± 1.5–433.4 ± 0.8 Ma; Hambrey, 1985; Grahn and Caputo, 1992; Díaz-Martínez and Grahn, 2007; Le Heron and Craig, 2008). Nested surfaces of glacial erosion intercalated with peri-Gondwana marine strata evidence dynamic intra-glacial ice advance and retreat and delimit the maximum areal extent of Gondwanan ice, reveal-
ing that, locally, the grounding line reached the continental shelf–slope break (Hambrey, 1985; Sutcliffe et al., 2000; Ghiemne, 2003; Le Heron et al., 2007; Le Heron and Craig, 2008).

In contrast to the short-lived glaciation deduced from ice-proximal, glaciogenic strata, hierarchical stratigraphic cycles that extend across non-glaciated, ice proximal and far-field continental margins of Darrwillian–Hirnantian age (467.3 ± 1.1–443.8 ± 1.5 Ma) most parsimoniously reflect glacioeustatic sea-level change induced by frequent expansions and contractions of Gondwanan ice sheets with inferred Milankovitch frequencies (e.g., Saltzman and Young, 2005; Desrochers et al., 2010; Loi et al., 2010; Turner et al., 2012; Elrick et al., 2013; Ghiemne et al., 2014; Dabard et al., 2015). The characteristically condensed Hirnantian stratigraphic records from far-field successions typically encompass an erosion surface (sequence boundary) onlapped by transgressive lithofacies; ostensibly, these features reflect a single episode of glacial advance, maximum, and retreat, though these surfaces could conceal a more nuanced glacioeustatic history during glacial maximum (Ghiemne et al., 2014).

Recently revised, high-resolution sequence stratigraphic frameworks for archetypal near-field (glaciated) and far-field (non-glaciated) locations in the Anti-Atlas Mountains of Morocco and Anticosti Island, Canada, respectively, appear to resolve discrepancies between ice proximal and ice distal stratigraphies and instead reveal a coherent, global glacioeustatic sea-level history (Desrochers et al., 2010; Ghiemne et al., 2014). Ghiemne et al. (2014) sequentially correlated the bounding surfaces of three latest Katian–Hirnantian low-order genetic stratigraphic sequences—and a handful of subordinate, higher order and lower significance maximum regressive and maximum flooding surfaces—between the two locations. While this method allows for diachronity on surfaces at a temporal resolution below that afforded by biostratigraphy, it assumes that fluctuations in global ice volume produce glacioeustatic sea-level rise or fall everywhere the same on the Ordovician paleoglobe, unless modulated by tectonic/geodynamic changes in base level and/or local sediment supply. For sites unaffected or similarly affected by these confounding factors, or for sites in which backstripping procedures removed these signals, the (glacio)eustatic conceptual model only considers correlation of equivalent stratal surfaces valid; for instance, under no circumstance would a maximum flooding surface at a near field location correspond temporally to the maximum regressive surface at a far field site because that would imply that global ice-volume change could simultaneously produce glacioeustatic sea-level rise at one location and glacioeustatic sea-level fall at the other.

However, a variety of physical markers for past sea level during Cenozoic glaciations and ongoing geodetic observations demonstrate that isostatic deformation and local ice gravitation—physical processes collectively described as glacial isostatic adjustment (GIA)—produce site-specific time-histories of sea-level change that may differ substantially from the eustatic, or global mean, sea-level curve produced by the melting of ice sheets (e.g., Milne and Mitrovica, 2008; Raymo et al., 2011; Stocchi et al., 2013). For instance, the post-glacial uplift and emergence of shelves surrounding formerly glaciated regions, like Hudson Bay (Andrews, 1970), contrasts with the prolonged submergence of far-field sites like Barbados (Peltier and Fairbanks, 2006), though both occurred during late Pleistocene–Holocene ‘glacioeustatic’ sea-level rise. Hence, local differences in relative sea level, and concomitant opposing trends in the creation and destruction of accommodation space, can characterize glacioeustasy. Thus, for certain Cenozoic glacial scenarios, corresponding stratal surfaces at near and far field locations—for instance, sequence boundaries or maximum flooding surfaces—neither correlate globally nor carry chronostratigraphic significance. Could this also be true for the Late Ordovician? We sought to explore the circumstances in which ice proximal and ice distal stratigraphies provide a unified and accurate reconstruction of Late Ordovician glacioeustasy.

Determining whether and how stratigraphic inferences of Late Ordovician local sea level differ from eustasy depends on robust constraints on numerous factors, including the temporal history of the volume and location of Ordovician ice; the rheological properties of the solid Earth; the location of paleocontinents; the shelf bathymetry, shoreline geometry, and topography of these continents; and the temporal changes in sediment supply to each margin. Uncertainties in the Ordovician-relevant values for these parameters preclude numerical predictions of site-specific sea-level change of sufficient accuracy to fit inferences determined from careful sequence stratigraphic analysis. Such models can, however, provide qualitative, physical insight into questions about ancient ice configuration, volume, and melt history (e.g., Creveling and Mitrovica, 2014). To this end, this study explores the spatial and temporal variation in sea level induced by the glaciation and deglaciation of a modeled Late Ordovician ice sheet covering southern Gondwana using a theory that accounts for the deformational, gravitational and rotational perturbations to sea level on a viscoelastic Earth model (Mitrovica and Milne, 2003; Kendall et al., 2005). Here we demonstrate that (i) the histories of sea-level change along continental margins both near- and far-afield of the Late Ordovician ice sheet can differ substantially between locations, as well as deviate from the glacioeustatic curve; (ii) reconstructions of glacioeustasy from far-field margins may not faithfully reflect the magnitude of glacioeustasy; and, (iii) by inference, the stratal surfaces of transgression and regression developed at near-field margins may, counter-intuitively, result from ice sheet growth and melt, respectively, and thus may not serve as chronostratigraphic horizons for correlation to far-field stratigraphies.

2. A model Late Ordovician glaciation

2.1. Paleogeography and ice volume history

Paleomagnetic reconstructions of Late Ordovician paleogeography similarly place Gondwana in a south polar position and distribute three large landmasses—Laurentia, Baltica, and Siberia—across the equatorial latitudinal band (Killian et al., 2016; Torsvik and Cocks, 2017; Blakely, 2008), though models differ in the exact paleolatitude and the position of minor landmasses and island arc terranes. Here we adapt the 450 Ma paleogeography of Torsvik and Cocks (2017) with a revised ~10° northeastward shift of Laurentia following Swanson-Hysell and Macdonald (2017) (Fig. 1A).

In detail, the paleogeography and paleobathymetry of Ordovician continents are not well resolved. Coupled climate–ice sheet models parameterized for the Ordovician reveal an insensitivity of the areal extent of ice to modeled continental topography (Pohl et al., 2016). Thus, we followed Creveling and Mitrovica (2014) and prescribed an initial paleogeography consistent with modern mean elevations. Modeled continental interiors have an elevation of 850 m that linearly tapers to sea level (0 m) within 350 km of any shoreline; oceanward of the shoreline, the modeled continental shelf deepens to ~150 m across a distance of 80 km, then to ~2000 m over the next 30 km (the continental slope), and to ~3800 m over the next 300 km (the continental rise) to the depth of the modeled abyssal plain (Fig. 1B).

The time history of Ordovician–Silurian ice volume adopted here follows the reconstruction of Finnegan et al. (2011) who leveraged clumped isotope paleothermometry to deconvolve the trend in Ordovician δ18O_carb imparted by continental ice growth from the signal of coeval seawater temperature change. This ice history, expressed in units of globally uniform sea level equivalent (SLE), extends from ice free conditions (0 m SLE) at 449 Ma to a Hirnantian glacial maximum (henceforth, HGM) at 444.5 Ma.
Given these areal and volumetric constraints, modeled ice height at HGM peaked at 2780 m (Fig. 3B).

2.2. Sea level theory

Our model predictions adopt the gravitationally self-consistent ice age sea-level theory developed by Mitrovica and Milne (2003) and Kendall et al. (2005) to account for the migration of shorelines due to local transgression/regression and the advance or retreat of ice from marine sectors. The latter occurs when the modeled Gondwanan ice sheet transits the continental shelf as ice volume expands and contracts around the HGM. The sea level theory also incorporates the rotational stability theory of Mitrovica et al. (2005) that accounts for the feedback into sea level of surface ice and ocean water load-induced perturbations to the Earth’s rotation vector. We note that in Pleistocene ice age calculations, the figure of the Earth adopted in computing perturbations in rotation generally incorporates the known present-day excess ellipticity of the Earth (Mitrovica et al., 2005; Mitrovica and Wahr, 2011). In the absence of constraints on the figure of the Late Ordovician Earth, we set excess ellipticity to zero; we note that rotational feedback is a minor contributor to these sea level predictions and the associated uncertainty does not impact our conclusions.

In the following, we adopt a one-dimensional, Maxwell viscoelastic Earth model as is standard in the Pleistocene ice-age literature (Peltier, 2004). We base our sea level predictions on a pseudo-spectral algorithm (Kendall et al., 2005) with a truncation at spherical harmonic degree and order 256. The elastic structure of the Earth model is taken from the seismic model PREM (Dziewonski and Anderson, 1981). We consider a ‘standard’ viscosity profile defined by an elastic (essentially infinite viscosity) lithosphere of 120 km thickness, an upper mantle viscosity of $5 \times 10^{20}$ Pa·s, and a lower mantle viscosity of $5 \times 10^{21}$ Pa·s. The boundary between the latter regions is set to 670 km depth. This profile falls within the class of viscosity models preferred in most Pleistocene ice age calculations (e.g., Lambeck et al., 1998; Mitrovica and Forte, 2004). However, in the results below, we test the sensitivity of the results to changes in the Earth model.

We adopt the definition in ice-age sea level literature of ‘sea level’ as the difference between the height of the equipotential that defines the sea surface (which varies globally) and the height of the solid Earth surface. Thus, at any location on the Earth’s surface, sea level may change due to a perturbation to the elevation of either (or both) of these bounding surfaces. We express symbolically the decomposition of sea level ($SL$) at a site of colatitude $\psi$ and east longitude $\Phi$ at time $t$, as:

$$SL(\theta, \varphi, t) = G(\theta, \varphi, t) - R(\theta, \varphi, t),$$

(1)

where $G$ refers to the sea surface height and $R$ to the elevation of the solid Earth surface, and we write the change ($\Delta$) in sea level since the initiation of glaciation as:

$$\Delta SL(\theta, \varphi, t) = \Delta G(\theta, \varphi, t) - \Delta R(\theta, \varphi, t).$$

(2)

In this expression, $\Delta G$ is the change in sea surface height and $\Delta R$ is the radial displacement of the crust.

To elucidate the physics underlying predictions for Late Ordovician sea-level change at the various sites described below, we further decompose the sea surface height change, $\Delta G$, into globally uniform ($\Delta G_{\psi}$) and geographically variable ($\Delta G_{\theta}$) terms (e.g., Kendall et al., 2005):

$$\Delta SL(\theta, \varphi, t) = \Delta G_{\theta}(\theta, \varphi, t) + \Delta G_{\phi}(\theta, \varphi, t) - \Delta R(\theta, \varphi, t).$$

(3)

The term $\Delta G_{\phi}$ differs from the eustatic sea level change largely because the mean crustal elevation of the ocean changes with
time in a GIA simulation. Finally, we refer below to predictions of relative sea level (RSL), which we define as the elevation of any shoreline marker of age \( t \) relative to sea level at the end of the simulation (\( t_f, 437 \) Ma; see Fig. 2):

\[
RSL(\theta, \varphi, t) = \Delta SL(\theta, \varphi, t) - \Delta SL(\theta, \varphi, t_f).
\]  
(4)

The question remains how numerical predictions of sea level relate to inferences of glacioeustatic determined from shelf stratigraphic architectures. Since the aforementioned definition of sea level is synonymous with the stratigraphic concept of ‘accommodation space’ (e.g., Posamentier et al., 1981), and sequence stratigraphic architectural units and their bounding surfaces reflect the interplay between accommodation and sedimentation (Catuneanu et al., 2009), our numerical predictions accord with sequence stratigraphic architectural units that: (i) are deemed to reflect glacioeustatic sea-level change, and do not, or only minimally, reflect accommodation change related to sedimentation and/or tectonism, or (ii) have been corrected for the mechanisms that increase or decrease accommodation via a method such as backstripping.

3. Results

We begin with a global prediction of RSL (equation (4)) at HGM, though to isolate the predicted departures from eustasy arising from glacial isostatic adjustment, we have removed from this RSL prediction the modeled eustatic sea-level rise between the HGM and the end of glaciation (i.e., \( \Delta SL_{\text{HGM}} - \Delta SL_{\text{E}27_{\text{Ma}}} = 130 \) m; Fig. 4). The global map (Fig. 4A) and the stereographic southern hemisphere projection (Fig. 4B) allow us to explore the distinct magnitudes of departure from eustasy for various regions on the Late Ordovician paleo-globe.

Predicted post-glacial rebound of the formerly ice-covered region of southern Gondwana produces a net sea-level fall that peaks at \( \sim 700 \) m (relative to the eustatic curve) at the center of the model ice sheet (Fig. 4B). The predicted departures from eustasy distant from the perimeter of the ice sheet, and throughout the global ocean, are of smaller amplitude, typically varying from \( \sim 50 \) to \( \sim 25 \) m (Fig. 4A). The spatial pattern of these departures reveals that they arise largely from flexure of the elastic lithosphere, which results because the modeled duration of ice sheet volume changes (i.e., Myr) exceeds the dominant decay times that govern GIA (i.e., kyrs). (Henceforth we will use the term ‘flexure’ to describe the ice and ocean load-induced crustal deformation although, even on the protracted time scale considered here, viscous effects, while relatively small, are non-zero.) The contribution of lithospheric flexure is evident on these maps in two ways. First, in marine areas surrounding the region of model ice cover, a localized (short wavelength) peripheral zone of subsidence predicted across the interval from HGM to 437 Ma leads to sea-level rise \( \sim 25 \) m in excess of the eustatic rise. By comparison, in predictions of Pleistocene relative sea level, viscous effects produce a far more extensive zone of peripheral subsidence (e.g., Creveling et al., 2017). Elastic flexure, however, has a spatial scale governed by the thickness of the elastic lithosphere which, for the model prediction in Fig. 4, is 120 km. The hinge point of the ice-load-induced tilting extends slightly oceanward of the shoreline (Fig. 4B, where the 0–90 m yellow contour extends into the ocean surrounding the ice sheet.) Second, far-afield from the ice sheet, ocean loading leads to continental uplift (and sea-level fall) and subsidence of the oceanic crust (sea-level rise). The interface between these regions is characterized by a flexural tilting of the crust over a spatial scale that is once again a function of the lithospheric thickness. This flexural tilting is manifest as a gradient in eustatic departures (Fig. 4A) where, closer to the continent, uplift-induced sea-level fall attenuates the eustatic rise (cool colors) and, more oceanward,
The predicted relative sea level (RSL; see equation (4)) at the model Hirnantian Glacial Maximum (444.5 Ma; see Figs. 2 and 3) as (A) a global map and (B) a stereographic projection centered at the South Pole and extending to the equator. In both maps the eustatic sea-level change from HGM to the end of the simulation (130 m; see Fig. 2) has been removed from the prediction to emphasize the spatial deviations from eustasy. Note that the right scale bar—RSL in meters, from −360 to 720 m—corresponds to the amplitude of RSL change at near-field locations emphasized in frame (B) and the left scale bar—RSL from −24 to 48 m—corresponds to the amplitude of RSL at far-field locations emphasized in frame (A). The entirety of the far-field (left) scale is encompassed within the yellow and orange colorbar divisions of the near-field (right) scale; for this reason, for instance, the red-colored, negative RSL values around southern Gondwana (Frame A, −6 to −24 m) appear in orange in Frame B.

This modeling exercise demonstrates that GIA induces spatially variable departures from eustasy that are largest in regions surrounding far-field continental shorelines and in the near field of the model Gondwana ice sheet, which at HGM extended to the edge of the continental shelf. To explore these regional departures from eustasy in greater detail, we computed time-histories of RSL at 20 sites along three profiles representing: (1) a near-field region, from the Armorica Terrane to the southern margin of Gondwana; (2) a transect across the inundated, shallow marine Arabian crustal block, distal from the southern Gondwana ice sheet; and (3) a far-field transect orthogonal to the southern margin of Laurentia (Fig. 3). Comparisons of the time histories of predicted RSL at all 20 sites for the full glacial–deglacial simulation (black lines) to the eustatic sea-level change associated with the model ice volume (red lines) reveal local departures from eustasy along a profile, and differences in sea-level histories between profiles (Fig. 5).

The predicted RSL histories along the Laurentian profile, in the far field of the Late Ordovician ice sheet, show the same temporal structure and directional trend as the eustatic curve; however, the amplitude of the departure from eustasy increases landward along the profile (Fig. 5C). The largest departure coincides with the HGM, and it reaches ~40 m at the northernmost site along the profile. This trend can be seen spatially in Fig. 4. During the model deglaciation, for example, the eustatic sea-level rise increases the ocean load (on oceanic crust and continental shelves), but does not load the remaining continent. This causes a tilting of the crust downward towards the ocean and upward towards the continent (as discussed in the context of the far-field flexure signal in Fig. 4A). Therefore, as one moves landward along the profile, GIA will drive a progressively more pronounced crustal uplift during the deglaciation that acts in opposition to—and effectively attenuates—the glacioeustatic rise (Fig. 5C). In this region, the RSL prediction at the most oceanward site (site 1) most closely matches the eustatic.

We next explore the two profiles whose constituent sites sat close to and under the model ice sheet at its maximum extent during the HGM (Figs. 5A, B). (Note that dashed sections of the RSL histories in Figs. 5A and 5B indicate the interval over which ice covered the site.) The general trend in the Armorican (sites 2–8) and Arabian (sites 1–7) profiles shows the dramatic impact of GIA on predictions of relative sea level as one moves landward.

**Fig. 4.** Predicted relative sea level (RSL; see equation (4)) at the model Hirnantian Glacial Maximum (444.5 Ma; see Figs. 2 and 3) as (A) a global map and (B) a stereographic projection centered at the South Pole and extending to the equator. In both maps the eustatic sea-level change from HGM to the end of the simulation (130 m; see Fig. 2) has been removed from the prediction to emphasize the spatial deviations from eustasy. Note that the right scale bar—RSL in meters, from −360 to 720 m—corresponds to the amplitude of RSL change at near-field locations emphasized in frame (B) and the left scale bar—RSL from −24 to 48 m—corresponds to the amplitude of RSL at far-field locations emphasized in frame (A). The entirety of the far-field (left) scale is encompassed within the yellow and orange colorbar divisions of the near-field (right) scale; for this reason, for instance, the red-colored, negative RSL values around southern Gondwana (Frame A, −6 to −24 m) appear in orange in Frame B.

**Fig. 5.** RSL predictions for the model simulation at sites along (A) the Armorican profile, 8 sites; (B) the Arabian profile, 7 sites; and (C) the Laurentian profile, 5 sites. See Fig. 3 for the location of these profiles. Note that dashed sections of the RSL histories in frames (A) and (B) depict the interval over which ice covered the site and will not have a marine sedimentary record that preserves relative sea level. The red line on each frame is the modeled eustatic (globally averaged) sea-level change for the adopted ice history (see Fig. 2). The profiles in frames (A)–(C) extend 1250 km, 2250 km, and 1100 km, respectively. The numbered sites are not evenly distributed along these profiles.
along the profiles toward the model Late Ordovician ice sheet. For instance, while we modeled 65 m of eustatic sea-level rise in the 1 Myr after the HGM, we predict a coeval 35 m sea-level fall at Armorica site 7, a differential amplitude of ~100 m and one of opposite trend to the eustatic history (Fig. 5A). We note that the non-monotonicity in the sea level histories along the Armorican profile results from this profile initiating near the eastern margin of the Armorican Terrane (site 1), transiting the shallow sea to the east (site 2), and continuing a landward transect up southern Gondwana (sites 3–8). Thus, the profile trends landward as one moves from site 2 to site 1 and from site 2 to site 8, and in this regard the trend in RSL predictions from sites 2–8 (Fig. 5A) closely matches that of the Arabian profile (Fig. 5B).

Clearly post-glacial rebound of near-field sites contributes to these departures from eustasy and, as discussed above, this effect extends beyond the margin of the ice sheet at HGM (e.g., see Armorican profile, sites 4 and 5; Fig. 5A). But is the GIA signal along the Armorican and Arabian profiles entirely a consequence of crustal rebound? To answer this question, we decomposed the sea level prediction at Armorica site 6 into the various contributing terms summarized in equations (2) and (3) (Fig. 6A). We first focus on the net deflection of the sea surface ($\Delta G(\theta, \phi, t)$, solid blue line), which is a sum of the contributions from geographically uniform ($\Delta G_{\phi}(t)$, dotted blue line) and geographically variable ($\Delta G_{\phi}(\theta, \phi, t)$, dashed blue line) perturbations to sea surface height. The latter represents the local gravitational effect of the GIA process on the sea surface elevation. As the model ice sheet grew toward HGM (prior to 444.5 Ma), it exerted a direct gravitational pull on the ocean and raised the sea surface. This was countered by subsidence beneath the ice sheet, which acted to weaken the gravitational pull and lower the local sea surface equipotential. If the system were in perfect isostatic equilibrium, then these two signals would have canceled; however, strength in the elastic lithosphere ensures that, even over such long time scales (order Myr), the gravitational effect of crustal displacement does not cancel the direct gravitational effect, such that the system remains in isostatic disequilibrium. In this case, the sum of the two signals (the geographically variable component of the sea surface height) leads to an elevation of the sea surface during glaciation and a lowering of this surface during deglaciation (Fig. 6A, dashed blue line).

Next, we examine the temporal history of crustal displacement ($\Delta R(\theta, \phi, t)$; Fig. 6A, dotted green line). Following equations (2) and (3), the total sea-level change (Fig. 6A, solid black line) equals $\Delta G(\theta, \phi, t)$ (solid blue line) minus $\Delta R(\theta, \phi, t)$ (the dotted green line). Over most of the simulation, this site remained at the periphery of the ice sheet and the crust was deflected upward as the ice sheet grew and downward as the ice sheet waned. The exception is the ~1 Myr interval around the HGM, when the ice sheet was sufficiently close to the site that it experienced a rapid subsidence during the last phase of glaciation and uplift during the initial phase of deglaciation. Note that this trend occurs even when the site is outside the ice sheet perimeter because, as we noted above, flexure acts to smooth the crustal displacement field (i.e., the hinge point on the crustal displacement is pushed outward from the ice sheet margin). In summary, while one would anticipate that the large departures from eustasy seen at near-field sites result from crustal displacements alone (Fig. 6A), these results demonstrate that the relative sea level histories at sites close to the ice sheet margin have significant contributions from both crustal elevation changes and perturbations to the gravity field.

We performed an analogous decomposition of the sea level history at Laurentia site 4 (Fig. 6B; see also Figs. 3 and 5C). The site
results
The factor and first modeled (total) general and lithospheric result (aged) 3
Fig. 4A
The viscosity 1022 Pas. ×
viscosity eustatic RSL in we predictions of the history of Late Ordovician ice-volume change, but flexure arising from these changes can progressively attenuate the amplitude of glacioeustasy landward along a continental shelf (Figs. 4A, 5C). For this reason, inferences of glacioeustasy at sites that represent shallow environments on far-field paleo-margins could misrepresent the amplitude of glacioeustatic fluctuation. For far field localities that preserve a transverse cross-section of a continental shelf or ramp, the application of backstripping methods (e.g., Loi et al., 2010) to facies-based inferences of water depth from the site(s) situated furthest down depositional dip will most closely approximate the magnitude of glacioeustasy, but the site-specific departure from eustasy will depend on how the geometry of the coastline maps on to the relatively smooth pattern of ocean-load-induced flexural deformation. However, the most distal depositional environment preserved (or exposed) at a given field area need not have been situated outboard of the zone of flexure, and thus stratigraphic-based inferences of local sea level could differ from eustasy by up to 30–40%.

As we have noted, the spatial-scale of flexure depends largely on the thickness of the underlying elastic lithosphere, although gravitational effects associated with the departure from isostatic equilibrium will have a larger spatial scale; while lithospheric thickness can be estimated for paleo-continents (Grotzinger and Royden, 1990), it is not known precisely for the Late Ordovician. Indeed, quantitative constraints on the down-dip attenuation of the amplitude of relative sea-level change arising from a single cycle of glacioeustatic change could provide a novel estimate of the flexural rigidity (or, equivalently, the effective elastic lithospheric thickness) of Late Ordovician paleocontinents. Moreover, the higher fidelity of the glacioeustatic signal at down-dip locations presents a possible paradox for stratigraphic-based inferences of glacioeustatic sea-level change in that a site located sufficiently offshore so as to minimize elastic flexure may also be characterized by facies belts rather insensitive to the depth change of the typical glacioeustatic amplitude (i.e., order 101–102 m).

Our model simulations also demonstrate that Late Ordovician ice sheet growth and decay can simultaneously produce divergent and even opposing sea-level histories across ice proximal continental shelves (Figs. 5A, B) and between near- and far-field locations (Fig. 5A, B and C). For long-duration glacial cycles, those sites proximal to, and overlain by, the ice sheet will display a sea-level rise leading into and during the glacial maximum as a result of lithospheric deformation and gravitational attraction in response to the ice load; these same sites will experience a sea-level fall in the initial stages of the deglaciation as the crust rebounds elastically and the gravitational attraction relaxes in response to a diminished ice volume. In this regard, ice proximal settings experience a direc-

sits on the landward side of the ocean-load induced tilting evident in Fig. 4A and so it experiences local subsidence during glaciation and uplift during deglaciation. This deformation is accompanied by a similar trend in the geographically variable (i.e., local) component of the sea surface elevation. Since these trends follow the general time history of the eustatic curve, the net effect is that the (total) relative sea level change (Fig. 6B solid black line) also tracks the eustatic curve, albeit at a reduced amplitude. For instance the HGM low stand of 100 m at Laurentia site 4 is not as large as the modeled eustatic value of –130 m (Fig. 5C).

The results presented in Figs. 4 and 5 are based on a single viscoelastic Earth model with a viscosity profile consistent with analyses of various Holocene geologic and geodetic observations related to the GIA process. We evaluated the sensitivity of the results to this choice by running two additional simulations. In the first, we reduced the lithospheric thickness from 120 km to 72 km, and in the second we increased the lower mantle viscosity by a factor of six, from 5 × 1021 Pas to 3 × 1022 Pas. Fig. 7 shows the predicted RSL curves for an illustrative site on all three profiles. The RSL predictions are relatively insensitive to the adopted mantle viscosity (Fig. 7, solid versus dotted lines) because the modeled loading history has a time scale that is much longer than the dominant viscous decay times associated with any realistic Earth model. That is, viscous effects have largely relaxed in calculations of sea-level change over this time scale. (We note, in this regard, our singular focus on the background state—or long-term history—of the ice sheet; any more rapid changes in ice sheet volume, such as those produced by Milankovitch forcing, would be superimposed on this background state and would show a more significant sensitivity to mantle viscosity.) In contrast, the near field (Armoric and Arabian profiles) predictions do show sensitivity to lithospheric thickness (Fig. 7, solid versus dashed lines). As indicated above, sea level predictions near the margin of the ice sheet at HGM will show significant discrepancies with the eustatic sea-level curve due to crustal deformation, and the associated local gravitational perturbation, resulting from flexure. The geometry and magnitude of this flexure is a strong function of the thickness of the lithosphere.

4. Discussion

Robust inferences of Late Ordovician glacioeustasy require a deliberate selection of field sites. For the case of the Myr- duration Late Ordovician glacial cycles modeled here, far-field localities preserve a robust temporal history of Late Ordovician ice-volume change, but flexure arising from these changes can progressively attenuate the amplitude of glacioeustasy landward along a continental shelf (Figs. 4A, 5C). For this reason, inferences of glacioeustasy at sites that represent shallow environments on far-field paleo-margins could misrepresent the amplitude of glacioeustatic fluctuation. For far field localities that preserve a transverse cross-section of a continental shelf or ramp, the application of backstripping methods (e.g., Loi et al., 2010) to facies-based inferences of water depth from the site(s) situated furthest down depositional dip will most closely approximate the magnitude of glacioeustasy, but the site-specific departure from eustasy will depend on how the geometry of the coastline maps on to the relatively smooth pattern of ocean-load-induced flexural deformation. However, the most distal depositional environment preserved (or exposed) at a given field area need not have been situated outboard of the zone of flexure, and thus stratigraphic-based inferences of local sea level could differ from eustasy by up to 30–40%.

As we have noted, the spatial-scale of flexure depends largely on the thickness of the underlying elastic lithosphere, although gravitational effects associated with the departure from isostatic equilibrium will have a larger spatial scale; while lithospheric thickness can be estimated for paleo-continents (Grotzinger and Royden, 1990), it is not known precisely for the Late Ordovician. Indeed, quantitative constraints on the down-dip attenuation of the amplitude of relative sea-level change arising from a single cycle of glacioeustatic change could provide a novel estimate of the flexural rigidity (or, equivalently, the effective elastic lithospheric thickness) of Late Ordovician paleocontinents. Moreover, the higher fidelity of the glacioeustatic signal at down-dip locations presents a possible paradox for stratigraphic-based inferences of glacioeustatic sea-level change in that a site located sufficiently offshore so as to minimize elastic flexure may also be characterized by facies belts rather insensitive to the depth change of the typical glacioeustatic amplitude (i.e., order 101–102 m).

Our model simulations also demonstrate that Late Ordovician ice sheet growth and decay can simultaneously produce divergent and even opposing sea-level histories across ice proximal continental shelves (Figs. 5A, B) and between near- and far-field locations (Fig. 5A, B and C). For long-duration glacial cycles, those sites proximal to, and overlain by, the ice sheet will display a sea-level rise leading into and during the glacial maximum as a result of lithospheric deformation and gravitational attraction in response to the ice load; these same sites will experience a sea-level fall in the initial stages of the deglaciation as the crust rebounds elastically and the gravitational attraction relaxes in response to a diminished ice volume. In this regard, ice proximal settings experience a direc-
tional change in the creation and destruction of accommodation space to that predicted by contemporaneous, ‘glacioeustatic’ ice volume growth; for the model duration of ice volume fluctuation considered here, these changes may persist for 100–500 kyr (Fig. 5A, B). Thus, an episode of glacioeustatic sea-level fall associated with ice volume growth may appear as a transgressive surface (or retrogradational stratigraphal package) at ice-proximal sites (e.g., Fig. 5A, Armorican profile, sites 4–8)—including those sites not directly overlain by the ice sheet (e.g., Fig. 5A, Armorican profile, sites 4 and 5)—and temporally correlate to a regressive surface (or progradational/degradational stratigraphal package) at ice-distal sites located further down depositional dip of the glaciated margin (e.g., Fig. 5A, Armorican profile, sites 1–3) or at far-field margins (e.g., Fig. 5C, Laurentian profile, sites 1–5), assuming, of course, that the latter sequence architectures reflect glacioeustasy and not dynamics related to tectonism or local sedimentation. Likewise, at ice-proximal margins, regressive surfaces may evidence and coincide with global ‘glacioeustatic’ rise and episodes of ice volume melt, and hence correlate with transgressive surfaces developed across far-field continental margins.

Here we modeled a long-duration Late Ordovician cycle of ice volume growth and decay (Fig. 2), but numerous stratigraphic studies have concluded that the Late Ordovician glaciation was comprised of a hierarchy of higher-order, lower-significance depositional cycles analogous to Cenozoic glaciation (e.g., Ghienne et al., 2014), and some ascribe Milankovitch periodicities—specifically a 400-kyr eccentricity cycle and suborbital ~100 and ~20 kyr orbital cycles—to these constituent depositional sequences (e.g., Loi et al., 2010; Dabard et al., 2015). If robust, then future numerical modeling should incorporate these superimposed, smaller-amplitude fluctuations in global ice volume. While stratigraphic inferences need necessarily inform the amplitude of these higher-order fluctuations, we caution that the viscous deformation arising from more rapid glacial-interglacial cycles will combine with the aforementioned effects dominated by elastic deformation to produce local relative sea-level histories that, like the Plio-Pleistocene, substantially differ from glacioeustasy (e.g., Raymo et al., 2011).

In this regard, our model predictions serve to caution stratigraphers of the potential pitfalls of correlating near- and far-field stratigraphic successions in attempts to refine chronostratigraphic frameworks of Late Ordovician (or other) glaciation. Our approach has been intentionally broad, and we have not undertaken a rigorous analysis to predict site-specific histories of relative sea level around the Late Ordovician paleoglobe. Future efforts to do so could leverage any contemporaneous and contrasting records of transgression and regression (that is, directional changes in accommodation) at globally distributed sites around the Ordovician paleoglobe to constrain the suite of eustatic histories that could accommodate all relative sea-level records. In practice, however, the coarse temporal constraints on these histories will challenge this approach, as will uncertainties in the rheology of the Ordovician Earth.

5. Conclusions

We demonstrate that sea level histories at continental margins in both the near- and far field of the Ordovician ice sheet can differ significantly from the eustatic sea level curve intrinsic to the adopted ice history. At continental margins proximal to the ice sheet at glacial maximum, the impact of glacial isostatic adjustment can approach and even dominate the eustatic signal. This large signal results from a combination of crustal deflection due to lithospheric flexure and the gravitational perturbation due to incomplete isostatic compensation associated with the presence of the lithosphere. As a result of these processes, such sites may even exhibit a sea-level rise during a period in which the eustatic sea level is rapidly falling, and vice versa, when the site is close to the evolving ice margin. In this regard, near-field stratral surfaces of maximum regression and flooding need not reflect global (‘eustatic’) trends in ice sheet growth and decay, respectively, and thus may not provide chronostratigraphic horizons for correlation with far-field sequence architectures. At continental margins in the far field of the glaciation, local sea level variations preserve the temporal structure of the eustatic trend, but the amplitude can depart from this trend by as much as 30–40%.

Our predictions adopt a synthetic Hirnantian glacial maximum (444.5 Ma) ice volume that is equivalent to a 130 m SLE drawdown relative to the end of the glaciation (437 Ma) (Fig. 2). However, the HGM ice volume may have been larger or smaller (Pohl et al., 2016). Importantly, our modeling predictions scale quasi-linearly with the excess ice volume between HGM and end glaciation. If the excess ice volume at HGM were, for example, 260 m, then the amplitudes in Fig. 4 would scale up by a factor of 2 and the spatial patterns would remain relatively unaltered. For instance, while the details of site-specific sea level histories near coastlines would change (e.g., Fig. 5)—since a larger ice volume change causes a larger migration of shorelines, and this impacts predicted departures from eustasy—the percent departure in the amplitude relative to the eustatic trend would be relatively robust.

The predictions presented here are sensitive to lithospheric thickness, but not to mantle viscosity; the latter insensitivity is a consequence of the long-duration modeled Late Ordovician glaciation adopted here, and reflects the fact that on the time scale associated with this ice/ocean loading, relaxation associated with viscous effects in the mantle will be complete. Thus, any departure from isostatic equilibrium largely results from the presence of the elastic (high viscosity) lithosphere, and the gravitational effects associated with this disequilibrium. The same would not be true for the sea level response to shorter time-scale ice mass fluctuations superimposed onto the long time scale trends, where viscous effects would drive departures from eustasy with a distinct geometry and amplitude. Future modeling efforts should focus on these oscillations.

Acknowledgements

We gratefully acknowledge the constructive review by Steven Holland and the editorial work of Bruce Buffett. We thank Jacqueline Austermann for discussion and Francis J. Sousa for assistance with figures. This research was supported in part by a donation from the G. Unger Vetlesen Foundation to Oregon State University (J.R.C.).

References


Brenchley, P.J., 1994. Bathymetric and isotopic evidence for a short-lived late Ordovi-
1130/0091-7613(1994)022<0295:BAWFE>2.3.CO.


Sutcliffe, O.E., Sciences, E., Sy, A., 2000. Calibrating the Late Ordovician glaciation and mass extinction by the eucratities of ‘Earth’ orbit. Geology 28 (11), 967–970.

