**Postglacial Relative Sea-Level Changes in the Gulf of Maine, USA: Database Compilation, Assessment and Modelling**

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**ABSTRACT**

Relative sea level (RSL) reconstructions from paleo records are often the most valuable data set for testing and constraining glacial isostatic adjustment (GIA) models. In some regions, the amplitude and rate of the reconstructed RSL changes are large making the data difficult to fit. Arguably, the reconstructed RSL curve from Maine, USA, is the classic example of such a data set with peak values at about 120 m dropping to a low stand of around -40 m within a few thousand years. To our knowledge, no GIA model has captured these extreme variations and the record has been somewhat neglected by the GIA modelling community. Here we critically assess and present a revised pre-10 ka RSL data base for this region and combine it with two recent Holocene compilations. Based on our assessment, we conclude that the large and rapid changes, inferred in previous studies, are robust. To determine if a successful model fit can be found, we consider a GIA parameter set comprising five ice models and 440 spherically-symmetric, Earth viscosity models assuming a Maxwell rheology. Results show that none of the model parameter sets produce a good fit to the data set. The rapid RSL fall and lowstand can be simulated by Earth models with low values of upper mantle viscosity (~5 × 1019 Pas). However, such low viscosity values result in a large mid-Holocene highstand (order 10 m) while the observations indicate a monotonic RSL rise throughout this interval. Our results indicate that Earth models able to simulate time-dependent viscosity (i.e., those that include transient and/or non-Newtonian deformation) might be required to fit the Maine RSL data set.

Keywords: Holocene, sea level changes, North America, glacial isostatic adjustment, Earth rheology

**1. INTRODUCTION**

As the present ice sheets continue to melt with ongoing global warming, understanding and quantifying the associated sea-level changes becomes increasingly important. One route towards this goal involves considering records of ice-sheet and sea-level changes since the Last Glacial Maximum to better understand the deglaciation process and its sea-level response. In addition to seeking a better understanding of the underlying processes, the influence of the most recent deglaciation continues to impact global changes in sea level through the process of glacial isostatic adjustment (GIA): the response of the solid Earth to past changes in land ice (e.g., Peltier, 1999; Riva et al., 2009), which is particularly strong in regions where the major Quaternary ice sheets formed, such as North America and Eurasia (e.g., Slangen et al., 2012). GIA models, which solve the sea-level equation (Farrell and Clark, 1976) using information on ice history and Earth rheology (Whitehouse, 2018), have been tested and refined using sea-level reconstructions from geological records (e.g., Peltier and Andrews, 1976; Lambeck et al., 1998; Simpson et al., 2009; Tarasov et al., 2012; Roy and Peltier, 2015). These calibrated models play an important role in understanding past and projecting future sea-level changes (e.g., Love et al., 2016).

Data from various regions can be used to test GIA models and constrain parameters. For example, as the influence of GIA is less significant in far-field regions, observations from these regions are typically used to constrain changes in global ice volume (Milne, 2015 and references therein). On the other hand, in regions in the near-field of past ice sheets, where the GIA signal dominates, the observed sea-level response is used to infer information on the regional ice history and Earth rheology (Milne, 2015 and references therein). These records are typically more difficult to fit given the large spatio-temporal variations and sensitivity to parameters relating to both the forcing (ice history) and response (solid Earth). This is particularly true for ice-marginal records where the relative sea level (RSL) response is highly non-monotonic due the local isostatic response being of similar amplitude to the global barystatic signal (e.g., Khan et al., 2015; Gregory et al., 2019). These records are typically characterised by an early and rapid RSL fall followed by a lowstand and rise through the Holocene (e.g., James et al., 2005; Kelley et al., 2010) and thus present a challenging target for GIA models (James et al., 2009; Yousefi et al., 2018). The rigorous tests provided by these data are useful given emerging evidence for an isostatic response that departs from the Maxwell rheology commonly assumed in GIA modelling (e.g., Adhikari et al., 2021; Lau et al., 2021) and the importance of the isostatic response for simulating ice-margin evolution (e.g., Gomez et al., 2012; Whitehouse, 2018; Han et al., 2021).

In this study, we compile, assess and model a RSL database for the Gulf of Maine, USA, where RSL displays a complex and large magnitude signal (e.g., Kelley et al., 2010). RSL reconstructions from this region are compatible with regional deglaciation preceding a large RSL fall around 15 ka which ended in a lowstand near 12 ka followed by a RSL rise that includes a period of relatively low rates of rise (termed a “slowstand”) from 10.5 ka to 7.5 ka (Kelley et al., 2010). From the data presented in Kelley et al. (2010), the RSL fall is estimated to be around 160 m from 15 ka to 12 ka averaging a rate of -53.3 m/kyr. To our knowledge, this large and complex signal has not been successfully reproduced by a GIA model.

Previous GIA modelling studies have considered RSL data from Maine (e.g., Engelhart et al., 2011; Roy and Peltier, 2015; Love et al., 2016). However, they focussed on Holocene data only which can, in general, be fit well by GIA models. However, the pre-Holocene data which capture the rapid RSL fall and lowstand have largely been ignored by the modelling community. Our primary aims in this study are: (1) to compile and assess a pre-Holocene database for this region using agreed community protocols (Khan et al., 2019) and (2) seek to fit this challenging dataset using a GIA model.

**2. METHODS**

2.1 DATA COMPILATION AND ASSESSMENT

The study area ranges from Nova Scotia, Canada, in the north, to Massachusetts, USA, in the south (Fig. 1). For this region, we compile a database of sea-level indicators from the literature, focussing primarily on the period before 10 ka that is not extensively covered by the two existing databases (Engelhart and Horton, 2012; Vacchi et al., 2018). We follow the database format set out by Hijma et al. (2015) and widely implemented in the compilation of Holocene sea-level databases from around the world (Khan et al., 2019). We classify each sea-level indicator as either a sea-level index point (SLIP), which provides the elevation of RSL at a specific location and point in time, or a limiting point, which provides a minimum or maximum constraint on RSL. The new database is provided in Appendix 1.

For each SLIP, we assess the *indicative meaning*, the relationship between the indicator and contemporaneous tidal levels (Van de Plassche, 1986; Shennan, 2015). The indicative meaning, incorporating a midpoint, or *reference water level*, and uncertainty, or *indicative range*, varies between different sea-level indicators (Table 1). The majority of SLIPs in our database stem from subaerially exposed glaciomarine delta topset-foreset contacts (Thompson et al., 1989). Sequence stratigraphic studies commonly ascribe a reference water level of approximately mean sea level (Chappell, 1974; Gobo et al., 2015), although few studies have explicitly validated this with modern observations. We tentatively assign an indicative meaning of 3 m below mean tide level (MTL) to MTL, following recommendations from Fernlund (1988) and Berglund (1992). We do not derive SLIPs from submerged palaeodeltas as the topset-foreset contact cannot typically be confidently discerned in seismic reflection datasets and may be eroded by the subsequent transgression (Barnhardt et al., 1995; 1997). Nevertheless, we use these geomorphic features and the elevation of fluvial incision (Knebel and Scanlon, 1985) to corroborate quantitative reconstructions of the depth of the lowstand obtained from other data sources. Fine-grained estuarine facies containing well-preserved intertidal or shallow subtidal bivalves provide a small number of SLIPs; however, several of these are of critical importance in defining the depth of the lowstand (Lee, 2006). For these samples, we assign indicative meanings with reference to literature on the depth distributions of the particular species of bivalves encountered, as discussed in Section 3.

The majority of marine limiting points in our database relate to the glaciomarine sediments of the Presumpscot Formation (e.g., Bloom, 1963; Stuiver and Borns, 1975; Hyland, 1978; Dorion, 2001; Retelle and Weddle, 2001). Marine microfossils, shells, macroalgae and other macrofossils indicate deposition in a marine environment; consequently, we assign an indicative meaning of below MTL. A small number of terrestrial limiting points relate to organic-rich sediments or minerogenic sediments containing freshwater microfossils or macrofossils (Stuiver and Borns, 1975; Belknap et al., 1987; Anderson et al., 1992). For these samples, we assign an indicative meaning of above highest astronomical tide.

We estimate relevant tidal datums for each sea-level indicator using the TPXO8‐Atlas global tidal model (Egbert and Erofeeva, 2010). Middle and late Holocene SLIPs have been corrected for changes in tidal range (Gehrels et al., 1995, 1996). We do not consider changes in tidal range over time for data older than 7ka; any such changes are likely to be small in comparison to the magnitude of the changes in relative sea level following the deglaciation of the region (Gehrels et al., 1995; Hill et al., 2011). Also, many SLIPs, including those that constrain the lowstand, are related to MTL and so a time-varying tidal range would not affect the inferred RSL value.

For all sea-level indicators, we account for uncertainties in the vertical position, including uncertainties related to the absolute elevation of the top of a core or section and the subsurface depth of the dated sample (Hijma et al., 2015; Khan et al., 2019). Many of the source publications provide elevations with respect to an unspecified datum. We assume that this is mean sea level and do not apply any corrections to account for datum differences. Where possible, we derive estimates of each component of the vertical position uncertainty from the original publications and follow the recommendations of Hijma et al. (2015) where these are not stated.

Radiocarbon dating provides constraints on the age of many of the sea-level indicators in the database. We recalibrate all ages, including those in the databases of Engelhart and Horton (2012) and Vacchi et al. (2018), using OxCal v.4.2 (Bronk Ramsey, 2009) and the IntCal20 and Marine20 calibration curves (Heaton et al., 2020; Reimer et al., 2020). To account for the marine reservoir effect, we use the *deltar* application (Reimer and Reimer, 2017) to recalculate the reservoir correction from paired terrestrial and marine samples from the Mercy Hospital landslide site (Thompson et al., 2011). We apply the resulting correction (ΔR) of 617 ± 149 years to all marine samples older than 12 ka. Marine shells collected in the 19th and 20th centuries suggest a substantially smaller correction is required in the late Holocene (McNeely, 2006). We consequently use ΔR = -70 ± 44 years from the calib 8.2 application (Stuiver et al., 2022) for data younger than 8 ka and linearly interpolate between these two ΔR values (-70 and 617 yr) and their uncertainties to define ΔR values for ages 8 to 12 ka (to avoid a large and abrupt change in this correction).

The subaerially exposed deltas from which we derive SLIPs are poorly constrained by direct age control (Thompson et al., 1989; Hall 2017; Koestler et al., 2017). Hall et al. (2017) and Koestler et al. (2017) have provided new surface exposure age dates for the Pineo Ridge System of deltas and associated recessional moraines that average 15.3 ka and 14.5 ka, respectively. The Hall et al. dates are more landward and likely better reflect the time of the highstand in eastern Maine. Consequently, we are able to assign an age range with a conservative temporal uncertainty based on the timing of deglaciation. Ice-sheet recession occurred in coastal Maine starting from approximately ~15.3 ka-14.5 ka (Borns et al., 2004; Dalton et al., 2020; Hall et al., 2017; Koestler et al., 2017); we have, therefore, assigned an age of 15.3 ± 1.5 ka for each of these SLIPs. While falling relative sea level means that lower elevation deltas very likely postdate higher elevation deltas, uncertainties over the precise pattern and timing of the rapid retreat of ice from the region currently prevent the development of a more detailed chronology (Dalton et al., 2020).

We divide the study region into 10 subregions (also referred to as “sites”) following the spatial partitioning adopted by Vacchi et al. (2018) for Canada and Engelhart and Horton (2012) for the United States, with further subdivision of two subregions that incorporate data from greater distances inland from the present coastline (Fig. 1). We plot SLIPs as boxes with width and height proportional to the vertical and 2σ age uncertainties. We plot marine limiting points as blue T-shaped symbols and terrestrial limiting points as green ⊥-shaped symbols. For limiting points, the horizontal line lies at the midpoint of the height measurement with a one-sided 2σ uncertainty range (lower range for marine limiting data and upper range for terrestrial limiting data) and the width reflects the 2σ age uncertainty.

2.2 GIA MODEL

Assuming a spherically symmetric Earth with a Maxwell (viscoelastic) rheology, our GIA model computes the sea-level response to Late Pleistocene ice-ocean mass redistribution. This response is computed using an algorithm to solve the sea-level equation (Kendall et al., 2005; Mitrovica and Milne, 2003). The model incorporates time-varying shorelines as well as the influence of GIA-induced changes in Earth rotation (Milne and Mitrovica, 1998; Mitrovica et al., 2005). The viscoelastic Love numbers (Peltier, 1974) used to compute Earth deformation include material compressibility (Wu and Peltier, 1982). The two primary model inputs are an ice loading history and an Earth model that defines density and rheological parameters as a function of depth.

As is common in GIA studies, the viscosity of the Earth model is partitioned into three shells: the lithosphere being the outermost region of variable thickness and high viscosity as well as the upper mantle and the lower mantle being deeper regions with uniform viscosity. The upper mantle extends from the base of the lithosphere to 670 km depth and the lower mantle extends from 670km depth to the mantle-core boundary at 2891 km.

Our viscosity models consider multiple combinations of 11 upper mantle viscosity (UMV) values [0.05, 0.08, 0.1, 0.2, 0.3, 0.5, 0.8, 1, 2, 3, 5 x 1021 Pas] and 10 lower mantle viscosity (LMV) values [1, 2, 3, 5, 10, 20, 30, 50, 70, 90 x 1021 Pas] with lithosphere thicknesses of either 46, 71, 96 or 120 km, giving a total of 440 combinations for a single ice model. The Earth’s viscosity structure is not well constrained, and this is reflected in the large range of parameters considered in our GIA modelling. In the following, we abbreviate the viscosity structure of an Earth model to LT-UMV-LMV with LT defined in km and UMV/LMV as 1021 Pas (e.g., 96-0.5-10 corresponds to a model with a 96 km lithosphere, an UMV of 0.5 x 1021 Pas, and a LMV of 10 x 1021 Pas).

For the ice-history input, we selected three published models: one developed by colleagues at the Australian National University (referred to as the ANU model hereafter) (e.g., Lambeck et al., 2014; Lambeck et al., 2017), ICE-6G (Peltier et al., 2015) as well as one North American Ice Complex history from Tarasov et al. (2012), GLAC1D-na9927 (abbreviated to 9927 in the following). This regional model is embedded in the global ICE-6G model. Ice extent for these models in our study region is shown at 21 ka, 17 ka and 13 ka in Figure 2. At 21 ka, the 9927 model has a lower ice thickness compared to ICE-6G and ANU by more than 1000 m in some areas. From 21 ka to 17 ka, all models show a significant amount of thinning, but ice remains present near the Gulf of Maine coastline. Margin retreat is greatest in the 9927 model, followed by ICE-6G with the ANU model showing little margin retreat by 17 ka. By 13 ka, ice has retreated from the region of study in all three models. Finally, we note that the ICE-6G and 9927 models were developed assuming the viscosity model VM5a (Tarasov et al., 2012; Peltier, 2015). The ANU model was developed via an iterative procedure to jointly optimise the Earth viscosity model and ice history, with optimum Earth model viscosity structure beneath North America defined by LT=102 km, UMV=5.1 x 1020 Pas, LMV=1.3 x 1022 Pas (Lambeck et al., 2017).

Cosmogenic evidence on the ice thickness changes in the Gulf of Maine suggests a rapid thinning around 15.2 ± 0.7 ka (Koester et al., 2017). Given this, we generated some relatively extreme pre-15 ka deglaciation scenarios in an attempt to better fit the RSL observations. To do this, we partitioned the ICE-6G model into a regional and global component. The ICE-6G model gave some of the best results in fitting the RSL data and so it was chosen for this aspect of the model sensitivity analysis. We isolated the regional component by multiplying the ICE-6G thickness values by a disc of thickness unity which tapers to zero beyond a specified radius (see Fig. S1). The chronology of ice in this regional component (Fig. S1, inset) was varied to delay ice thinning after 21 ka and thus produce more rapid retreat near the time of rapid RSL fall around 15 ka (Kelley et al., 2010). The standard ICE-6G chronology was applied to ice thickness values beyond the region covered by the disc (Fig. S1, main plot). This procedure allowed us to vary the poorly known ice thickness chronology within the study region without affecting ice extent/chronology in other areas. RSL was computed for the global and regional ice models separately and these results were summed to determine the total RSL signal. The error incurred in this procedure is small (a few metres) as the RSL response is linear in the ice loading component (which dominates the RSL response in this region, see Fig. 7). The error is related to changes in the computed ocean loading signal which is of second order.

2.3 DATA-MODEL COMPARISON

Prior to comparison of the database with GIA model output, redundant limiting data (i.e., those that do not add to RSL constraints provided by other data) were removed using a visual analysis. Limiting points were kept in the time intervals which were sparsely covered by SLIPs. The highest lower (marine) limiting points as well as the lowest upper (terrestrial) limiting points were prioritized to constrain the model curves as much as possible. Limiting points with a smaller time uncertainty were also prioritized due to the better constraint they provide. Figure S2 illustrates the result of this cleaning process at site 8b. Removing redundant data affects the data-model comparison as the misfit criteria (Equations 1 & 2 below) are normalised by the number of data points and so redundant limiting data bias the calculated misfit to lower values in cases where the model curve is compatible. To give greater weight to data that have a relatively low density in time or space, uncertainties of data from large spatial and temporal clusters were increased (Briggs and Tarasov, 2013). In terms of spatial weighting, the RSL elevation uncertainty was scaled by the square root of the number of RSL data points in each site divided by the square root of the total number of RSL data points in the entire dataset. For temporal weighting, we considered four time bins based on the temporal distribution of RSL data points, encompassing 0-3, 3-6, 6-9, and older than 9 ka. Similar to the spatial weighting, the RSL age uncertainty was scaled based on the square root of the number of RSL data points falling into each time bin divided by the square root of the total number of RSL data points for that site.

The quality of data-model fit for a given parameter set (i.e., set of 3 Earth viscosity parameters and specified ice model) was quantified by calculating a misfit value. In total, over 1300 parameter sets were considered.

For a given parameter combination and individual SLIP location, the closest point on the model curve (in height and time) is determined. The misfit ($δ\_{SLIP}$) is then computed using:

$$\begin{array}{c}δ\_{SLIP}=\frac{\sqrt{\sum\_{n=1}^{N}\left(∆\_{RSL,n}^{2}/σ\_{RSL,n}^{2}+∆\_{t,n}^{2}/σ\_{t,n}^{2}\right)}}{N}\#\left(1\right)\end{array}$$

With $Δ\_{RSL,n}$ and $Δ\_{t,n}$ being, respectively, the RSL and age difference between the nth observation point and the model prediction closest to this point. $σ\_{RSL,n}$ and $σ\_{t,n}$ are, respectively, the 2-σ observational uncertainty of the RSL and age, and N is the total number of observational data for a given subregion.

Data-model misfits for the limiting data ($δ\_{Lim}$) were calculated differently as these data provide only a one-sided height constraint on RSL. While the SLIPs indicate that the RSL must be a specific value at a given time, the limiting data accept any value above (for lower limiting) or below (for upper limiting) the data. Thus, finding the closest point on the model RSL curve to determine a misfit is less appropriate in this case. Therefore, model values were determined for the age of the limiting data point (including uncertainty) and a misfit calculated if the model curve sat on the ‘wrong’ side of the point (e.g., above an upper limiting point) using:

$$\begin{array}{c}δ\_{Lim}=\frac{\sqrt{\sum\_{n=1}^{N}\left(∆\_{RSL,n}^{2}/σ\_{RSL,n}^{2}\right)}}{N}\#\left(2\right)\end{array}$$

If the model value is on the ‘correct’ side of the limiting point, no penalty is incurred. Since the limiting data are one-sided RSL constraints, the computed misfit is weighted by 0.5 when determining the total data-model misfit:

$$\begin{array}{c}δ\_{Total}=δ\_{SLIP}+0.5 δ\_{Lim}\#\left(3\right)\end{array}$$

The Earth viscosity parameter combination giving the lowest value of $δ\_{Total}$ is selected as the best fitting viscosity model for each of the ice histories used.

**3. RESULTS AND DISCUSSION**

Our new database includes 223 accepted sea-level indicators, of which 90 are SLIPs, 119 are marine limiting points, and 14 are terrestrial limiting points. The oldest sea-level indicators have mean ages predating 16 ka, while 93% predate 10 ka (see Fig. S3). Of the sea-level index points, 69 are from subaerially exposed glaciomarine delta topset-foreset contacts that lack direct chronological control but must closely postdate deglaciation (Thompson et al., 1989). Our database complements existing, predominantly post-10 ka databases compiled by Engelhart and Horton (2012) and Vacchi et al. (2018), which include 221 and 215 sea-level indicators from the region of this study, respectively. The entire data set (SLIPs and limiting data) used to reconstruct RSL for this region comprises a total of 659 data points from Vacchi et al. (2018), Engelhart and Horton (2012) and the new database. Of these, 167 limiting data were removed in the cleaning process (Section 2.3) leaving 492 datapoints to be used for data-model comparison (Fig. 3).

As indicated in Fig. 1, the data at sites 7 and 8 were sub-divided into areas a and b. This was done because the spatial RSL gradient across these sites is large, resulting in a considerable error when including all the data on a single RSL-time plot (see Fig. S4). Partitioning the data as indicated in Fig. 1 reduces this error considerably. Note, that this spatial variation in the data has no impact on the misfit calculations as these values are computed at the spatial co-ordinates of each data point. It only affects the accuracy of compiling the data onto a RSL-time plot and comparing these data with a model curve (which is computed at the average of all the data points in a given area – diamonds in Fig. 1).

The most complete RSL records exist at sites 7 and 8 which capture a rapid RSL fall around 14 ka that ends sharply around 12 ka, after which the data suggest a RSL rise to present. Sites 7b and 8b provide the best constraints on the timing and amplitude of the lowstand, with maximum depths around -30 m and -70 m, respectively. Other sites lack data to define the time and height of the lowstand so precisely. However, data from many sites are compatible with the lowstand timing and amplitude at sites 7b and 8b, indicating that it is a robust feature of the regional RSL response. Data from several sites (e.g., 2, 3, 7 & 9) suggest a decrease in the rate of RSL rise during the Holocene. A RSL “slowstand” from ~9-6 ka, when sea levels were relatively stable, has been inferred from data at site 7b (Kelley et al., 2010). This interpretation requires a rapid rise in RSL of order 10 m from 7-6 ka. There are insufficient data from other locations to support this RSL scenario. While data from site 2 indicate a relatively rapid and continuous RSL rise during 9-6 ka, Stea et al. (1996) indicate that the ice mass remained longer in this subregion compared to the Gulf of Maine, hence the RSL “slowstand” cannot be confirmed nor rejected. Furthermore, a rapid RSL rise of ~10 m at 7-6 ka is not evident in any far-field records. No evidence of glacially-triggered faulting (Steffen et al., 2021), which could explain this scenario, has been found in the Gulf of Maine (Anderson et al. 1989).

In subregion 8b, two SLIPs constrain the depth of the lowstand (Fig. 3). These index points relate to articulated *Mya arenaria* and *Mytilus edulis* shells encountered in fine-grained sediments from Saco Bay, near Portland, Maine (Lee, 2006). Tyler-Walters (2008) considers *M. edulis* as intertidal to 5 m depth based on a global synthesis of records, while Suchanek (1978) reports the species to 10m below MTL along the US West coast. Similarly, *M. arenaria* is widely reported as an intertidal to shallow subtidal species (Rasmussen, 1973; Newell & Hidu, 1986; Powilleit et al., 1995). While it has been reported from deeper depths (Theroux and Wigley, 1983), it is not clear that these specimens were articulated and in living position at the time of collection rather than having been reworked from their original living depths. We consequently assign larger indicative ranges of MTL to 10 m below MTL to both species (Fig. S2). We reject two further samples from Lee (2006) that are of similar ages and elevations due to poor preservation of the dated shells and the possibility of reworking. We also treat a poorly preserved articulated *Modiolis modiolis* specimen as a marine limiting point due to the wide depth distribution of this species (Tyler-Walters, 2007). The two accepted SLIPs suggest relative sea level fell from -32.5 ± 5.2 m at 12.5 – 13.1 ka to -59.9 ± 5.2 m at 11.9 – 12.7 ka. This lowstand depth is consistent with undated evidence from the submerged Kennebec palaeodelta, which features possible shorelines at elevations of -20 – -30 m, -30 – -40 m and, in its central/southern portion, -50 – -60 m (Shipp et al, 1991; Barnhardt et al., 1997). The depth is also consistent with a fluvially incised basal unconformity observed down to an elevation of -40 m in Penobscot Bay, where it disappears from seismic records in a natural gas deposit (Knebel and Scanlon, 1985). The basal unconformity grades into a conformity at elevations of ~ -70 – -90 m (Kelley and Belknap, 1991), suggesting that the wave base at the lowstand was above this depth.

We seek to determine a model parameter set that provides a good fit to the majority of the RSL data (Fig. 3). In practice, this involves determining the parameters that minimise the misfit function (Eqns 1-3). Figure 4 provides a representative summary of this analysis, in which the darker areas identify the model parameters that produce the best data-model fits. The middle range of UMV values and mid to lower range of LMV values result in the best fits to the full sea-level dataset. While the misfit function is broadly similar in pattern for all three ice models, there are significant differences. For example, the 9927 misfit values tend to be higher than those for ANU and ICE-6G indicating that the model fits are poorer in general for 9927. The best fits for this ice model are found for mid to low LMV values and mid to high UMV values. In comparison, the misfit functions for the ANU and ICE-6G models show a more ‘u-shaped’ pattern of low values that results in a relatively well-constrained range of good-fitting UMV and LMV values. As evident in Fig. 4, optimal parameter values for the ANU and ICE-6G models produce significantly better fits compared to 9927 (minimum misfit values and the Earth viscosity parameters that result in these are given in Fig. 4 caption). One common result for all ice models is the poor fits obtained using UMV values below 1 x 1020 Pas. We selected the set of Earth parameters with the lowest misfit value to visually compare the data to modelled RSL for all 10 sites (Fig. 5).

The generally higher misfit values for the 9927 ice model are evident in Fig. 5. For example, at site 1, the 9927 curve fits most of the data, but does not reach to the SLIP near -30m, in contrast to the ANU and ICE-6G curves. At sites 6 to 8, the 9927 model struggles to match the steep RSL fall as well as the observed lowstand. The ANU and ICE-6G models are also unable to fit the RSL lowstand at sites 7b and 8b. At sites 8 to 10, the ICE6G model most accurately captures SLIPs in the mid-to-late Holocene. The optimal 9927 curve more closely predicts the depth of the RSL lowstand at site 9. The 9927 model does fairly well at sites 2 to 6, but the fits elsewhere are of poorer quality. The ANU and ICE-6G models provide good fits at sites 1 to 6 and at site 10, hence the lower misfit values overall (Fig. 4). None of the models can fit the data at sites 7 to 9 where the observed RSL signal is captured most completely; a particular challenge is fitting the timing and depth of the RSL lowstand.

We also computed RSL curves using the Earth viscosity models that were used in the development of ICE-6G and 9927 (VM5a Earth model; Tarasov et al., 2012; Peltier, 2015) or, for the ANU model, the outcome of a joint inversion (Lambeck et al., 2017). For the ANU case we used parameters that fall within the provided uncertainty ranges, specifically: LT=96 km, UMV=5 × 1020 Pas, LMV=1022 Pas. As expected, the parameters determined to optimise fits to the Maine dataset (Fig. 4) produce better quality fits at most locations (compare results in Figs 5 and S5).

Data-model fits were also calculated for specific sites, notably 7 and 8, to determine if significantly improved fits can be achieved at these most challenging sites. The resulting misfit plots (not shown) are similar to those obtained for the entire data set (Fig. 4) and so the changes in optimal parameter values for each ice model were minor. As a result, there was no significant improvement in the quality of RSL fits at sites 7 and 8 and so we adopt the optimal Earth model parameters inferred using the entire data set.

In the remainder of this section, we present a sensitivity analysis to determine the feasibility of improving the model fits to the RSL lowstand without deteriorating the fit quality elsewhere. For this analysis we focus on four sites: 6, 7b, 8b and 9. These sites were chosen because they capture, to some degree, both the late glacial and Holocene aspects of the RSL response, and so are the most challenging to fit, and they span the majority of the Maine coastline.

We first consider the impact of varying the three Earth model parameter values. We define a reference parameter set that has intermediate values: LT=71 km, UMV=5x1020 Pas, LMV=1022 Pas (black line in Fig. 6) and vary each of the three Earth model parameters sequentially to isolate the sensitivity of modelled RSL to each one. RSL sensitivity to variation in LT indicates that thicker values result in a deeper RSL lowstand. While this is counterintuitive, it can be explained by the dominance of the ice-loading signal in this region (see Fig. 7 and related discussion): the thicker the lithosphere, the smaller the magnitude of ice-induced uplift (RSL fall) which leads to the barystatic signal having a larger effect and, therefore, increasing the depth of the early Holocene lowstand. Model sensitivity to LMV is also significant, with variations on the order of 10s of metres. In this case, however, the intermediate curve (black line) is above both the higher and lower LMV values instead of being between them.

While both LT and LMV changes have a significant effect on the local RSL variations, the model output is most sensitive to changes in upper mantle viscosity (especially within the lower half of the range considered). A high UMV gives a similar curve to the intermediate value, but with a slightly smaller rate of change while the low UMV increases the rate of RSL change significantly. Of all the Earth model parameter variations considered in Fig. 6, adopting a low UMV value is the only case that produces a local RSL lowstand that is broadly compatible with the data (although we note that the timing is earlier than indicated by the observations). However, the low UMV value also results in a RSL highstand during the Holocene which is not observed. Based on our sensitivity analysis results for variations in LMV, we also considered model runs with low UMV values and high and low LMV values in an attempt to remove the Holocene highstand. However, similar results were obtained which we interpret as being due to a more muted sensitivity to LMV when very low UMV values are adopted (associated with deformation being focused in the upper mantle region). This is compatible with the misfit results (Fig. 4), which indicate relatively poor-quality fits for very low UMV values regardless of the adopted LMV (reflecting the large misfits in the Holocene). Thus, it appears that there is a trade-off between fitting RSL data in the late glacial and early Holocene versus fitting those in the mid-to-late Holocene.

Fig. 7 illustrates components of the RSL model curve in the Gulf of Maine for ICE-6G and two Earth viscosity models to assist in our interpretation of the results in Fig. 6. Specifically, the signal associated with the ice load, ocean load, global mean changes, and due to GIA-related changes in Earth rotation are isolated at site 8b (results for other sites are similar). The global mean sea-level (GMSL) change includes the barystatic signal caused by the addition of melt water, and syphoning (Mitrovica and Milne, 2002). As expected, the ice loading signal dominates the RSL response in the Gulf of Maine. This is the case regardless of the adopted ice and Earth model parameters. While considerably lower in magnitude than the ice component shortly after ice retreat, the GMSL and ocean component signals are significant during the late glacial and Holocene when the ice signal has reduced in amplitude. The GMSL curve shows a large increase during deglaciation until around 7 ka, after which global melting is much reduced (e.g., Dutton et al., 2015) and so the syphoning process dominates, leading to a small RSL fall.

The ocean-loading signal contributes a RSL fall of almost 20 m since 15 ka which is dominated by subtle land uplift associated with continental levering (Clark et al., 1978). The rotation signal does not exceed more than a few metres during deglaciation and so is not a significant contributor to the total modelled RSL. It is notable that the ocean, rotation, and GMSL signals are not strongly affected by the change in UMV (compare left and right plots in Fig. 7). Therefore, the large sensitivity to low UMV values found in Fig. 6 is strongly associated with the ice component signal. This signal includes a large RSL fall which is dominated by land uplift during deglaciation with smaller contributions from gravitational changes affecting the sea surface height. Using the optimal Earth model parameters for the ICE-6G model (left plot), the ice signal changes from a fall to a rise around 7 ka as the hinge line migrates towards Hudson Bay resulting in a marked transition in the rate of land uplift (Fig. S6, left plot).

The ice signal for a low UMV (right plot) shows more rapid uplift and an earlier transition from (ice-induced) uplift to subsidence (~12 ka). This explains the deep and early RSL lowstand in Fig. 6 for UMV=5 x 1019 Pas. As the hinge line recedes towards Hudson Bay, this low UMV case also shows a second change in slope of the curve during the mid-Holocene which leads to the highstand around 6 ka in Fig. 6 for this choice of UMV. This indicates a transition from subsidence back to uplift at this time (see Fig. S6, right plot). As the Gulf of Maine is first situated in the near field of the ice sheet and then becomes a peripheral region during ice retreat, this could explain the occurrence of a RSL Holocene highstand for the case of a low UMV. Previous work (Officer et al.,1988) has postulated that models with a high LMV/UMV ratio exhibit channel flow in the upper mantle which is characterised by subsidence followed by uplift of peripheral regions.

The above results indicate that changing the Earth viscosity parameters alone cannot provide good quality fits to the RSL data, particularly at sites 7-9. To complete the sensitivity analysis, we turn our attention to the ice model to try and better fit the timing of the RSL fall and lowstand. As described in Section 2.2, we isolated a regional component of the ICE-6G model (Fig. S1, inset) and varied the chronology of this component only. Specifically, we considered two scenarios where the ice thickness is kept constant from 25 ka until 18 ka (Fig. 8) or 15.5 ka (Fig. S7) such that the deglaciation prior to 15 ka is later and more rapid. Rapid thinning of the ice is compatible with cosmogenic exposure age dating in the region (e.g., Hall et al., 2017; Koestler et al., 2017). Of these two scenarios, the earlier case (18 ka) matched the timing of the observed RSL fall most accurately. The delayed and more rapid deglaciation generally improves the data-model fits for the older data, particularly at sites 7b and 8b. The results in Fig. 8 (and Fig. S7) indicate that varying the timing and rate of deglaciation could result in improved fits for the older data. However, regardless of the ice history used, low values of UMV (<1020 Pas) are required to capture the rate of fall and amplitude of the lowstand and so the issue of fitting the Holocene data remains (see pink and blue curves in Figs 8 & S7).

Our efforts to fit the RSL observations using three different published ice histories, 440 combinations of viscosity parameters for a 1D (spherically symmetric) Maxwell Earth model as well as two extreme scenarios of regional deglaciation have been unsuccessful. Our results indicate that it is not possible to fit the rapid RSL fall immediately following deglaciation and the steady RSL rise in the mid-to-late Holocene. Matching the rapid rate of RSL fall requires low values of UMV but this leads to the prediction of a highstand in the mid-Holocene due to the large UMV-LMV contrast. Our results are based on a typical, three-layer viscosity parametrisation of the lithosphere, upper mantle and lower mantle. A different depth parametrisation of this structure could produce better fits. For example, previous studies have considered Earth models that include an additional layer to represent a low viscosity asthenosphere (e.g., Fjeldskaar and Cathles, 1991; James et al., 2009; Yousefi et al. 2018). We considered the case of adding such a layer to the 3-layer Earth model that produced the optimal fits using ICE-6G (i.e., LT=96 km, UMV=5 x 1020 Pas, LMV=5 x 1021 Pas). The results (Fig. S8) indicate that, while the addition of this layer can have a significant influence on the modelled RSL, none of the parameter sets considered are able to produce a fit that is clearly better that those produced by the simpler 3-layer viscosity models. This is because the addition of this layer results in a model curve that plots between the curves of the bounding 3-layer viscosity models (i.e., the ICE-6G optimal model defined above and a model with the same LT and LMV values but with UMV equal to that in the asthenosphere layer (5 x 1019 Pas). That is, the thin (56 km) asthenosphere model produces an RSL curve similar to that for the 3-layer model 96-p5-5 (compare black and blue lines in Fig. S8) and the model with a thick asthenosphere (346 km) produces an RSL curve that approaches that of the 3-layer model 96-p05-5 (compare dashed black and violet lines in Fig. S8). We repeated this sensitivity test for an asthenosphere viscosity of 5 x 1018 Pas and obtained similar results.

Our results indicate that the RSL data from Maine might only be fit by an Earth model with time dependent viscosity, such that the viscosity is weaker at shorter timescales (transient effects, e.g., Sabadini, 1985; Lau et al., 2021; Ivins et al., 2022) and/or during large loading (stress) changes (non-linear effects, e.g., Wu & Wang, 2008; Kang et al., 2022). A recent study (Simon et al., 2022) has shown that an Earth model based on a Burgers rheology can produce improved fits to RSL data, particularly in regions proximal to the ice sheet margin. Furthermore, we note that the ice models considered here (and the majority of GIA studies), are developed under the assumption of a Maxwell Earth model. Therefore, if departures from a linear Maxwell model are significant, it is important to develop ice models that are compatible with these more complex, time-dependent, viscosity models (Huang et al., 2019). The Maine RSL data set is arguably the best paleo record available to test and constrain the Earth rheology component of GIA models and so we encourage future efforts that consider applications of these more complex rheological models to make use of dataset provided in this publication (Appendix 1).

**4. CONCLUSIONS**

In this study, we present an extended and critically assessed RSL dataset for the Gulf of Maine. Based on our assessment of the data, we conclude that the complex and large-amplitude RSL reconstruction, including a lowstand of ~60 m along the central Maine coast, is robust. All samples that were either poorly preserved or subject to possible reworking were removed from the database. Two SLIP samples (articulated *Mya arenaria* and *Mytilus edulis* shells) remain and support a deep and brief lowstand in this region. Our interpretation of these SLIPs is supported by a host of geomorphological, sedimentological and seismic evidence. However, since our database comprises only two SLIPs to precisely constrain the amplitude and timing of this regional lowstand, we encourage future efforts to collect new data and thus help consolidate this key feature of post-glacial RSL change in this region.

Model results were generated based on five ice histories and 440 Earth models assuming a 1D (spherically symmetric) Earth with a Maxwell rheology. Results show that the models are able to bound the data at most sites relatively well, except for locations 7 and 8 where no set of model parameters is able to accurately capture the amplitude and timing of the observed RSL change, which is due mostly to the large influence of the ice-loading signal. The high rate of RSL fall can be modelled only using low values of upper mantle viscosity (~1019 Pas). However, such low viscosity values result in a mid-Holocene RSL highstand which is not supported by the observations. We considered an extreme and delayed regional deglaciation scenario to test the sensitivity of our GIA simulations to this aspect of the ice model but were still unable to match the RSL observations with a single set of model parameters. Our results suggest that good fits to the RSL data may only be possible using more complex models of Earth rheology that simulate changes in Earth viscosity through time, via transient or non-Newtonian deformation.

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