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Global ice volume during MIS 3 inferred from a sea-level analysis of sedimentary core records in the Yellow River Delta

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A B S T R A C T

Estimates of global ice volume during the glacial phase of the most recent ice age cycle are characterized by significant uncertainty, reflecting the relative paucity of geological constraints on sea level relevant to this time interval. For example, during the middle stages of Marine Isotope Stage 3, published estimates of peak global mean sea level (GMSL) relative to the present range from −25 m to −87 m. The large uncertainty in GMSL at MIS 3 has significant implications for estimates of the rate of ice growth in the period leading to the Last Glacial Maximum (~26 ka). We refine estimates of global ice volume during MIS 3 by employing sediment cores in the Bohai and Yellow Sea that record a migration of the paleo-shoreline at ~50–37 ka through a transition from marine to brackish conditions. In particular, we correct relative sea level at these sites for contamination due to glacial isostatic adjustment using a sea-level calculation that includes a gravitationally self-consistent treatment of sediment redistribution and compaction, and estimate a peak global mean sea level of −38 ± 7 m during the interval 50–37 ka. With suitable sedimentary core records, the approach described herein can be extended to refine existing constraints on global ice volume across the entire glacial period.

1. Introduction

Ice volume variations through the last glacial cycle are a direct and sensitive measure of ice age climate change, and a key input into models of glacial isostatic adjustment (GIA). These variations have been constrained using oxygen isotope records of benthic and planktic foraminifera from deep-sea sedimentary cores (Siddall et al., 2008) and a wide range of geological markers of sea level, including erosional and constructional terraces, sedimentary and biological facies, and coral reefs (Lambeck and Chappell, 2001; Yokoyama et al., 2000; Muhs et al., 2012; Hanebuth et al., 2006). However, the accuracy of ice volume inferences based on oxygen isotope records is limited by regional variability, uncertainties in the conversion from $\delta^{18}O$ related to temperature, and the mean isotopic concentration of continental ice (Siddall et al., 2008; Waelbroeck et al., 2002). Moreover, geological markers of sea-level change are spatially and temporally sparse, and estimates of ice volume based upon them must account for a variety of contaminating signals, most notably GIA (Lambeck and Chappell, 2001; Lambeck et al., 2014; Milne and Mitrovica, 2008). The sparsity of the record is particularly problematic for the period prior to the Last Glacial Maximum (LGM) since many markers of sea level, created during more extensive ice cover, are now submerged. For these reasons, estimates of ice volumes during the bulk of the last glaciation phase, extending from Marine Isotope Stage (MIS) 5c (~100 ka; i. e. Muhs et al., 2012) through MIS 3 (60–25 ka; Siddall et al., 2008), are uncertain to within tens of meters of equivalent global mean sea level (GMSL), where GMSL is defined as the globally averaged sea-level change associated with a given change in total ice mass inventory (i.e., the volume of meltwater divided by the area of the ocean).

Constraints on ice volumes and GMSL during MIS 3 provide an illustrative case in point. Siddall et al. (2008) summarized and compared individual (e.g., Shackleton, 2000) and stacked benthic records (Lisiecki and Raymo, 2005), in addition to planktic records (Dannenmann et al., 2003), during this stage. As an example, over the time period 50–37 ka, peak GMSL estimates can range from −25 to −87 m, relative to present (Siddall et al., 2008). After correcting the coral record at Huon Peninsula for the signal due to GIA and tectonic uplift, Lambeck and Chappell (2001) concluded that GMSL fell from −60 m to −80 m during the same period, while

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the global ice history model ICE-5G is characterized by a GMSL value within the range −87 m to −100 m across this time interval (Peltier and Fairbanks, 2006). The sea-level lowstand at the Last Glacial Maximum (LGM; 26 ka; Clark et al., 2009) reached ~ −130 m (Yokoyama et al., 2000; Austermann et al., 2013), and therefore, within current uncertainty, global ice volume may have increased by more than a factor of 3 or by less than one-third in the 15 kyr period leading up to the LGM.

In this paper our goal is to refine estimates of global ice volume in the middle of MIS 3, from ~50 to 37 ka, using sedimentary cores from the Yellow River Delta in the Bohai Sea of China. These cores record a transition from marine to freshwater conditions at this time that reflects a migration of the ocean margin across the area and they provide an important constraint on local sea level. We correct the inferred local sea-level history in the region for GIA-induced sea-level change using a numerical model that includes a gravitationally self-consistent treatment of the impact of sediment redistribution (Dalca et al., 2013), and reconstruct GMSL during this time interval. The analysis will explore the sensitivity of the estimate of GMSL to various inputs adopted in the GIA calculation, including the models for sediment redistribution (which incorporates erosion, deposition, and compaction), ice history, and Earth structure.

2. Methods

2.1. GIA modeling

Local sea-level changes are not simply related to fluctuations in global ice volume. Ice sheet growth and melting on a viscoelastic Earth produces a complex spatio-temporal pattern of sea level change that is dependent on the full history of the surface mass (ice, water and sediment) load. The redistribution of surface loads over glacial cycles perturbs the Earth's gravitational field through crustal deformation and direct self-attraction, but the redistribution of water is, in turn, governed by this perturbation since the sea surface must remain a gravitational equipotential in a static sea-level theory. Farrell and Clark (1976) were the first to derive a gravitationally self-consistent sea-level theory – the so-called sea-level equation – under the assumption of a non-rotating Earth with fixed shoreline geometry. Their canonical work has been extended to include the effects of rotation (Milne and Mitrovica, 1996), evolving shorelines associated with local sea level changes and/or the migration of grounded, marine-based ice (Johnston, 1993; Milne et al., 1999; Lambeck et al., 2003; Kendall et al., 2005), and, most recently, sediment redistribution (Dalca et al., 2013). In the present study, we adopt the Dalca et al. (2013) sea-level theory, modified to incorporate sediment compaction, and solve it using the pseudo-spectral algorithm described in that paper. For this purpose we use a spherical harmonic truncation at degree and order 512, which represents a surface spatial resolution of ~40 km.

Relative sea level (SL) is defined as the height of the equipotential that coincides with the ocean surface (G) relative to the elevation of the solid surface:

$$SL = G - (R + H + I), \tag{1}$$

where R is the elevation of the crust, not including sediments and grounded ice, I is the thickness of grounded ice, and H is the thickness of sediment. We will henceforth use the terms "sea level" and "relative sea level" interchangeably. We will be concerned here with perturbations in sea level and each of the components in equation (1) from an initial time t_0 to a time t_j. If we denote this perturbation by the symbol Δ, then we can write:

$$\Delta SL_j = \Delta G_j - (\Delta R_j + \Delta H_j + \Delta I_j). \tag{2}$$

Our sea-level predictions solve for ΔSL_j, ΔG_j and ΔR_j, given time-varying input fields ΔH_j and ΔI_j. We prescribe the sediment redistribution, ΔH_j, as an input field computed from a database of dated sediment cores. Because these sediments have undergone compaction, we define the decompacted sediment thickness H_j at time t_j as

$$H_j = H_{present} - (h_j - \delta_j) \tag{3}$$

where $H_{present}$ is the sediment thickness at present to bedrock, H_j is the sediment thickness at t_j, h_j is the compacted sediment thickness deposited from time t_j to present day, and δ_j is the amount that H_j compacted from t_j to the present (see Fig. 1 for illustration).

$H_{present}$ is obtained from a map of isopach sediment thickness to bedrock, and h_j is determined from dated sedimentary cores. We calculate the elevation difference due to decompaction, δ_j, by using the input fields h_j and $H_{present}$, and by assuming an exponential porosity depth relationship (i.e. Athy, 1930; Guillocheau et al., 2012),

$$\Phi(z) = \Phi_0 e^{-\frac{z}{z_0}} \tag{4}$$

where Φ_0 is the surface porosity, z is depth, and z_0 is a lithology-dependent constant. By equating the sediment grain mass in the sediment column before and after compaction we may derive an expression for δ_j (details included in Appendix A):

$$\delta_j = \frac{\Phi_0 h_j \left(1 - e^{-\frac{H_{present}}{z_0}}\right)}{\left(1 - \Phi_0 e^{-\frac{H_{present}}{z_0}}\right)} \tag{5}$$

As noted above, the thickness of sediment deposited since the initial time step ($j = 0$) to time t_j is:

$$\Delta H_j = H_j - H_0. \tag{6}$$

This value of ΔH_j is the input into Equation (2).

Next, we turn to prescribing the surface mass load. In Dalca et al. (2013), the history of loading is written as

$$\Delta L_j = \rho_w \Delta S_j + \rho_I \Delta I_j + \rho_H \Delta H_j \tag{7}$$

where ρ_w, ρ_I, ρ_H are the (assumed constant) densities of water, ice and sediment, respectively, and ΔS_j is the change in ocean

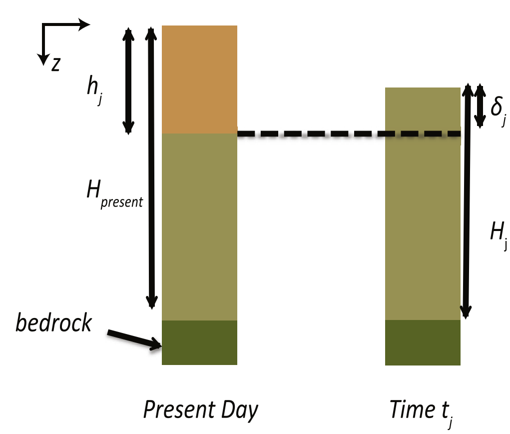

Fig. 1. Schematic illustration of sediment compaction and various parameters discussed in the text.