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Oxygen isotope mass-balance constraints on Pliocene sea level and East Antarctic Ice Sheet stability

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ABSTRACT

The mid-Pliocene warm period (MPWP, 3.3–2.9 Ma), with reconstructed atmospheric *p*CO_2 of 350–450 ppm, represents a potential analogue for climate change in the near future. Current highly cited estimates place MPWP maximum global mean sea level (GMSL) at 21 ± 10 m above modern, requiring total loss of the Greenland and marine West Antarctic Ice Sheets and a substantial loss of the East Antarctic Ice Sheet, with only a concurrent 2–3 °C rise in global temperature. Many estimates of Pliocene GMSL are based on the partitioning of oxygen isotope records from benthic foraminifera ($\delta^{18}O_b$) into changes in deep-sea temperatures and terrestrial ice sheets. These isotopic budgets are underpinned by the assumption that the $\delta^{18}O$ of Antarctic ice ($\delta^{18}O_i$) was the same in the Pliocene as it is today, and while the sensitivity of $\delta^{18}O_b$ to changing meltwater $\delta^{18}O$ has been previously considered, these analyses neglect conservation of $^{18}O/^{16}O$ in the ocean-ice system. Using well-calibrated $\delta^{18}O$-temperature relationships for Antarctic precipitation along with estimates of Pliocene Antarctic surface temperatures, we argue that the $\delta^{18}O_i$ of the Pliocene Antarctic ice sheet was at minimum 1‰–4‰ higher than present. Assuming conservation of $^{18}O/^{16}O$ in the ocean-ice system, this requires lower Pliocene seawater $\delta^{18}O$ without a corresponding change in ice sheet mass. This effect alone accounts for 5%–20% of the $\delta^{18}O_b$ difference between the MPWP interglacials and the modern. With this amended isotope budget, we present a new Pliocene GMSL estimate of 9–13.5 m above modern, which suggests that the East Antarctic Ice Sheet is less sensitive to radiative forcing than previously inferred from the geologic record.

INTRODUCTION

The magnitude of sea-level rise to which anthropogenic global emissions of CO_2 commit us over the coming centuries remains a pressing problem in the Earth sciences. The mid-Pliocene warm period (MPWP) offers a unique window into this problem, as it represents an Earth system equilibrated with modern to near-future radiative forcing (Pagani et al., 2009). Estimates of MPWP global mean sea level (GMSL) range between 10 m and 70 m above modern, with a value of 25 m often assumed for general circulation models (GCMs) configured with Pliocene boundary conditions (Haywood et al., 2010). Historically well-cited estimates come primarily from field-mapped elevations of Pliocene shoreline deposits and depth paleoecology of benthic mollusk and foraminifera assemblages (e.g., Dowsett and Cronin, 1990; Kaufman and Brigham-Grette, 1993; Naish and Wilson, 2009; Miller et al., 2012); however, recent work has shown that these estimates are confounded by the effects of glacial isostatic adjustment (Raymo et al., 2011) and dynamic topography (Rowley et al., 2013).

Alternatively, a number of studies have used oxygen isotope records from benthic foraminifera ($\delta^{18}O_b$) to independently constrain total ice sheet volume and Pliocene GMSL. In these studies, temperature controls on equilibrium fractionation during calcification are deconvolved from the evolution of seawater $\delta^{18}O$ ($\delta^{18}O_{sw}$) through several methods: (1) signal partitioning (Miller et al., 2012); (2) independent Mg/Ca temperature reconstructions (e.g., Dwyer and Chandler, 2009; Woodard et al., 2014); or (3) models of water exchange into restricted basins (Rohling et al., 2014). Pliocene GMSL is then calculated assuming a relationship between $\delta^{18}O_{sw}$ and GMSL of 0.1‰ ± 0.02‰/10 m. Using this method, estimates of peak MPWP GMSL are up to 30 m above modern, necessitating the full deglaciation of the Greenland Ice Sheet (GIS), the West Antarctic Ice Sheet (WAIS), and as much as 30% of the East Antarctic Ice Sheet (EAIS). These estimates conflict with both cosmogenic nuclide data of EAIS thickness (Yamane et al., 2015) and physical ice sheet models forced with MPWP boundary conditions that are only able to simulate 1–29 m GMSL rise, even with the recent inclusion of dynamic ice-sheet processes (Dolan et al., 2011; Pollard and DeConto, 2009; De Boer et al., 2015; Pollard et al., 2015).

The widely used 0.1‰/10 m relationship between $\delta^{18}O_{sw}$ and sea level is based on deglaciation following the Last Glacial Maximum (LGM), calibrated using estimates of volume and ice $\delta^{18}O$ ($\delta^{18}O_i$) of the LGM ice sheets (Olausson, 1963; Dansgaard and Tauber, 1969), temperature-corrected change in $\delta^{18}O_b$ between the LGM and Holocene (Emiliani, 1958; Shackleton and Opdyke, 1973), and independent estimates of deglacial sea-level rise (Fairbanks and Matthews, 1978). Implicit in this calibration is the assumption that $\delta^{18}O_i$ does not vary with climate. The Last Glacial Maximum (LGM)–Holocene is characterized by the full-scale deglaciation of the Laurentide and Fenno-Scandinavian Ice Sheets. Consequently, while the temporal evolution of $\delta^{18}O_i$ during LGM ice sheet growth and deglaciation amplifies $\delta^{18}O_{sw}$ signals (Mix and Ruddiman, 1984), comparisons between LGM and modern $\delta^{18}O_{sw}$ are still dominated by the large volumetric changes in terrestrial ice. In contrast, higher GMSL during the MPWP involves melting of currently extant ice sheets (i.e., GIS, WAIS, EAIS) and smaller changes in ice sheet volume compared to the Pleistocene. The transfer of ^{16}O into these ice sheets as climate cooled from the MPWP to the modern increases $\delta^{18}O_{sw}$ without corresponding changes in ice sheet volume. Therefore, late Cenozoic cooling has the potential to amplify $\delta^{18}O_{sw}$ signals of glaciation, rendering the LGM-calibrated $\delta^{18}O_{sw}$–sea level relationship inappropriate for estimating Pliocene GMSL. Herein, we argue that the inclusion of this previously neglected process into the isotope mass balance of the ocean-ice system substantially reduces estimates of MPWP GMSL and reconciles $\delta^{18}O_b$-based sea-level estimates with ice sheet models.

ANTARCTIC PLIOCENE TEMPERATURES AND $\delta^{18}O$

The fact that $\delta^{18}O_i$ and temperature co-vary at high latitudes is well established and has been exploited to reconstruct Greenland and Antarctic surface temperatures from ice cores (Dansgaard, 1964; Jouzel et al., 1987; EPICA Community Members, 2004). Physically, this co-variation is the result of temperature-dependent changes in the saturation vapor pressure of water in the atmosphere that cause changes in rainout as moisture is transported poleward, toward and over the ice sheets. The precise isotopic composition is further moderated by vapor source conditions, kinetic effects of snow formation, inversion layer dynamics, and transport type (Jouzel and Merlivat, 1984; Hendricks et al., 2000).

Globally, Pliocene temperatures were 2–3 °C warmer than pre-industrial, and ensemble estimates from the Pliocene Model Intercomparison Project (PlioMIP) experiments indicate higher Antarctic temperatures, ranging from 2.5 to 12.5 °C above modern (Haywood et al., 2013); however, higher-end simulated temperatures are due to ice-albedo feedbacks caused by imposed deglaciated WAIS and reduced EAIS boundary conditions (Lunt et al., 2012). Proxy

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estimates of Pliocene Antarctic warmth from the Sirius Group in East Antarctica and reconstructed Ross Sea sea-surface temperatures range from 2 to 20 °C above modern (Retallack et al., 2001; Francis et al., 2007; McKay et al., 2012), though the chronology of the Sirius Group deposits is controversial (Barrett, 2013). Coupled Model Intercomparison Project Phase 5 (CMIP5) projections of summertime Antarctic warming in A.D. 2100 range from ~1 to 4 °C (Representative Concentration Pathways [RCPs] 2.6–8.5) even with no change in the areal distribution of the Antarctic Ice Sheet (IPCC, 2013). Modern observations, combined with reanalysis and GCM projections, show that the WAIS is one of the fastest-warming regions on Earth (Bromwich et al., 2012), while EAIS accumulation is increasing at ~5%/°C due to the greater moisture-holding capacity of warmer air (Frieler et al., 2015). In short, though significant uncertainty surrounds Pliocene Antarctic temperature estimates, available evidence suggests that with near-modern pCO_2, the climate was warmer and wetter resulting in higher $\delta^{18}O_i$. Given that the average residence time for ice in the WAIS and EAIS is 44 k.y. and 125 k.y., respectively (Lhomme et al., 2005), $\delta^{18}O_i$–temperature (T) co-variability must be considered when comparing modern Antarctic ice to Pliocene Antarctic ice.

TABLE 1. MODERN PARAMETERS USED IN EQUATIONS 1 AND 2

Variable	Modern volume (10^6 km^3)	Modern mass (10^{18} kg)	Sea-level equivalent (m)	Modern $\delta^{18}O$ (‰)
Ocean	N.A.	1358	N.A.	0
GIS	2.9*	2.66†	7.3*	–35**
WAIS marine-based	3§	2.43†,#	3.4§	–41**
WAIS non–marine-based	N.A.	0.322†,#	0.9§	–41**
EAIS	23.5§	21.55†	53.3§	–56.5**

Note: GIS—Greenland Ice Sheet; WAIS—West Antarctic Ice Sheet; EAIS—East Antarctic Ice Sheet.
*Bamber et al. (2001).
†Masses are calculated assuming an ice density of 917 kg/m^3 (Fretwell et al., 2013).
§Fretwell et al. (2013).
#Calculation of the separate masses of the WAIS marine-based and non–marine-based sectors is shown in the Data Repository (see text footnote 1).
**Lhomme et al. (2005).

METHODS

To test the sensitivity of the $\delta^{18}O_b$ record and GMSL estimates to changing $\delta^{18}O_i$, we use equations that describe the isotope mass balance of $^{18}O/^{16}O$ in the ocean-ice system between modern and MPWP interglacial end members:

$$M_O + M_{GIS} + M_{WAIS} + M_{EAIS} = M_{pO} + M_{pGIS} + M_{pWAIS} + M_{pEAIS}, \quad (1)$$

and

$$M_O R_O + M_{GIS} R_{GIS} + M_{WAIS} R_{WAIS} + M_{EAIS} R_{EAIS} = M_{pO} R_{pO} + M_{pGIS} R_{pGIS} + M_{pWAIS} R_{pWAIS} + M_{pEAIS} R_{pEAIS}, \quad (2)$$

where M_x and R_x represent total mass and $^{18}O/^{16}O$ ratio, respectively, of the modern (O) and Pliocene (pO) ocean and the modern and Pliocene GIS, WAIS, and EAIS. Modern M_x and R_x values are listed in Table 1. We assume R_{pO} based on the 0.3‰ offset between modern and MPWP interglacial $\delta^{18}O_b$, calculated from the benthic stack of Lisiecki and Raymo (2005), and a bottom-water temperature effect that accounts for 0.1‰ of this offset (i.e., 67:33 signal partitioning ratio of $\Delta\delta^{18}O_{sw}$:ΔT; Miller et al., 2012). To calculate R_{pWAIS} and R_{pEAIS}, we assume a linear temperature dependence of WAIS and EAIS $\delta^{18}O_i$ with possible end-member slopes of 0.42‰/°C and 0.8‰/°C, representing average modeled temporal $\delta^{18}O_i$-T based on GCM simulations of modern and LGM climate (Lee et al., 2008) and the modern spatial relationship (Masson-Delmotte et al., 2008), respectively. For R_{pGIS}, we use the averaged modeled temporal $\delta^{18}O_p$-T slope (0.37‰/°C) at the three ice-core localities on GIS examined by Lee et al. (2008).

Because $\delta^{18}O_{sw}$ records an integrated signal from all terrestrial ice, we present two end-member scenarios of M_{pGIS} and M_{pWAIS} to examine the sensitivity of the $\delta^{18}O_b$ record and sea-level estimates to variable $\delta^{18}O_i$: we treat both the WAIS and GIS as entirely deglaciated in the Pliocene (scenario 1); and we deglaciate only the marine-based sectors of the WAIS (Pollard and DeConto, 2009) and half of the GIS (Dolan et al., 2015) in the Pliocene (scenario 2). In both scenarios, we then solve for the two remaining variables (M_{pO} and M_{pEAIS}) which provides the required additional melt from the EAIS to account for the full $\delta^{18}O_{sw}$ change across the range of estimated Pliocene Antarctic temperatures (2–20 °C). Sensitivity analyses to differential Pliocene warming of the WAIS and GIS along with assumed $\Delta\delta^{18}O_{sw}$:ΔT partitioning ratios of 80:20 and 50:50 are presented in the GSA Data Repository[1]. We note that our analysis is restricted to comparisons of end-member values of MPWP interglacials and the modern, though future work will aim to incorporate isotope mass balance into the temporal evolution of $\delta^{18}O_i$ within the Pliocene to investigate orbital-scale changes in sea level.

RESULTS AND DISCUSSION

The temperature-dependent increases in $\delta^{18}O_i$ act to amplify signals of terrestrial ice melt in $\delta^{18}O_b$ records under warmer conditions; consequently, partitioning of the 0.3‰ $\delta^{18}O_b$ signal into EAIS mass loss and increased $\delta^{18}O_i$ becomes a function of Pliocene high-latitude temperature change. Deglaciation of half of the GIS and the marine-based portion of the WAIS (scenario 2) combined with bottom-water temperature change accounts for 0.2‰ of the full MPWP–present offset (Fig. 1). As estimated Pliocene Antarctic temperatures increase, less EAIS mass loss must be invoked to account for the full 0.3‰ offset. Given the $\delta^{18}O_i$-T relationship of Masson-Delmotte et al. (2008), only a 7 °C increase is required to completely eliminate the need for any Pliocene EAIS mass loss. Under scenario 1, even smaller temperature increases are needed to explain the $\delta^{18}O_b$ record without invoking EAIS mass loss (Fig. DR1 in the Data Repository).

These deglaciation scenarios can be used to calculate the associated GMSL rise needed to reproduce the 0.3‰ $\delta^{18}O_b$ offset using the sea-level equivalent of each modern ice sheet (Fig. 2; Table 1). The total GMSL rise encapsulated in the $\delta^{18}O_b$ record becomes a function of Pliocene Antarctic temperatures, with higher temperatures resulting in lower GMSL. Under scenario 2, less total melt, and subsequently lower GMSL, is required to account for the $\delta^{18}O_b$ record than under scenario 1. In scenario 2, the EAIS accounts for a greater proportion of the total meltwater signal, and as the EAIS has the lowest $\delta^{18}O_i$ (–56.5‰), less total mass loss is needed to account for the $\delta^{18}O_b$ record.

We note that even without changes in $\delta^{18}O_i$ (i.e., $\Delta T = 0$), our maximum Pliocene GMSL estimates are only 15 m above modern, compared to an estimate of 21 m from Miller et al. (2012) using similar constraints. This estimate is lower due to two key differences. First, our mass balance budget allows for distinct meltwater signals from each of the ice sheets based on their modern $\delta^{18}O_i$. The generalized 0.1‰/10 m sea level relationship derived from LGM–Holocene records assumes negligible contributions from the isotopically light EAIS to the total meltwater signal, and therefore leads to an overestimate of Pliocene sea level. Second, by considering independent estimates of mass changes

[1]GSA Data Repository item 2015295, sensitivity analyses, calculation of WAIS marine and terrestrial sector masses, supplementary references, Table DR1 (MPWP sea-level estimates compilation), and Figures DR1–DR4, is available online at www.geosociety.org/pubs/ft2015.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

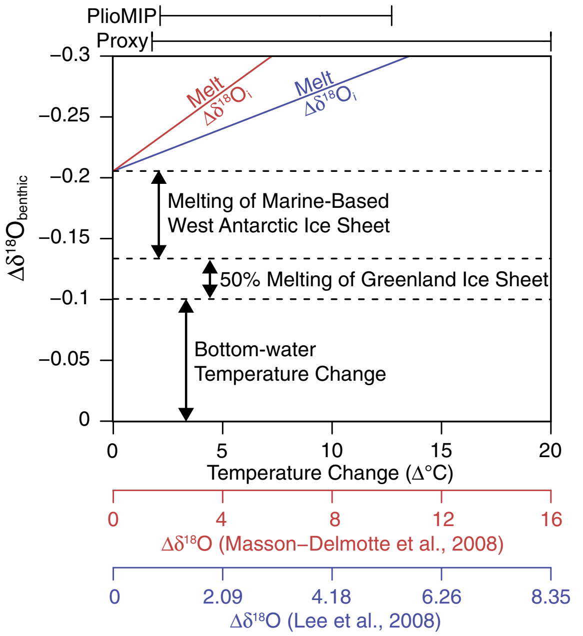

Figure 1. Attribution of changes in benthic foraminiferal $\delta^{18}O$ between mid-Pliocene warm period and present assuming 50% melting of Greenland Ice Sheet and melting of marine-based sectors of West Antarctic Ice Sheet as function of estimated Pliocene temperature increase. Blue and red lines correspond to end-member ice $\delta^{18}O$ ($\delta^{18}O_i$)–temperature (*T*) relationships over Antarctica (see *x*-axis). Area below red or blue line corresponds to signal explained by changes in $\delta^{18}O_i$ composition; area above is signal attributable to ice melt. Error bars above plot are full range of estimated Antarctic ΔT in the Pliocene Model Intercomparison Project (PlioMIP) experiment and proxy-based estimates.

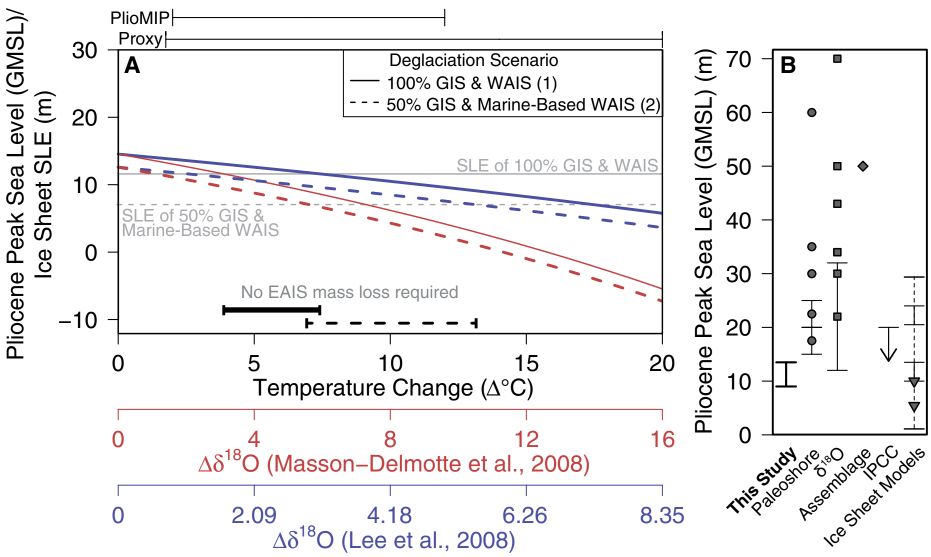

Figure 2. A: Total estimated global mean sea level (GMSL) due to melting of terrestrial ice at mid-Pliocene warm period (MPWP) as function of estimated Pliocene temperature increase. Solid lines correspond to scenario 1, and dashed lines correspond to scenario 2 (see text). Horizontal gray lines are total Greenland Ice Sheet (GIS) and West Antarctic Ice Sheet (WAIS) sea-level equivalents for these scenarios. Error bars above plot are full range of estimated Antarctic temperature change as in Figure 1. B: Previously published estimates of MPWP GMSL (Table DR1 [see footnote 1]), organized by method. Our estimate is indicated at left. Dashed error bar at right indicates full range of estimates from Dolan et al. (2011). PlioMIP—Pliocene Model Intercomparison Project; EAIS—East Antarctic Ice Sheet; SLE—sea-level equivalent; IPCC—Intergovernmental Panel on Climate Change.

and sea-level rise due to deglaciation, we avoid the assumption that all mass loss contributes to global sea level implicit in the 0.1‰/10 m sea level relationship. This is most significant for the WAIS where a significant mass proportion of the marine-based sector will not contribute to global sea level (Fretwell et al., 2013).

Figure 2 also shows the increase in Pliocene $\delta^{18}O_i$ and Antarctic temperature needed to account for the $\delta^{18}O_b$ record without invoking additional mass loss from the EAIS. This point occurs where the GMSL rise is equal to the combined GMSL rise from GIS and WAIS melting in each of our deglaciation scenarios and occurs at Pliocene Antarctic temperatures of 3–13.1 °C above modern (marked by solid and dashed lines along the *x*-axis in Fig. 2). Any additional temperature increase above this point implies an increase in EAIS mass. While some modeling studies suggest that EAIS growth may occur up to pCO_2 levels of 400—560 ppm (Ligtenberg et al., 2013; De Boer et al., 2015), the increase in temperature needed to sufficiently alter $\delta^{18}O_i$ is likely physically incompatible with EAIS mass growth.

Additionally, changes in $\delta^{18}O$ of other terrestrial water reservoirs such as global groundwater may amplify $\delta^{18}O_i$ signals and further reduce Pliocene sea level estimates. However, we neglect these changes from our analysis as there are no estimates as to how the mass of these global reservoirs changed in magnitude or distribution in the Pliocene, and because $\delta^{18}O$ of precipitation is less sensitive to changes in global temperature at lower latitudes.

IMPLICATIONS

Though estimates of Pliocene Antarctic temperature have considerable uncertainty, we view a 2.5–5 °C increase as conservative, given that globally averaged Pliocene temperatures were 2–3 °C above modern. This translates into a 1‰–4‰ increase in the average $\delta^{18}O_i$ of Antarctica. With these assumptions, we estimate that GMSL was ~9–13.5 m above modern, with a 2–4.5 m contribution from the EAIS. This estimate, with a maximum 8.5% loss of the EAIS, is significantly less than previous estimates based on paleoshorelines and benthic mollusk and foraminifera assemblages, and reconciles the $\delta^{18}O_b$ record with ice-sheet models of the Pliocene WAIS and EAIS as well as recent cosmogenic nuclide data suggesting negligible Pliocene EAIS mass loss (Pollard and DeConto, 2009; De Boer et al., 2015; Yamane et al., 2015). Assuming the full uncertainty in $\Delta\delta^{18}O_{sw}$:ΔT partitioning from 80:20 to 50:50, the full range of calculated maximum GMSL becomes 5–17 m above modern (see the Data Repository). These estimates suggest that the EAIS is substantially less sensitive to radiative forcing than previously inferred from the MPWP, and that dramatic deglaciation of the EAIS under modern pCO_2 is not supported by the geologic record.

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Oxygen isotope mass-balance constraints on Pliocene sea level and East Antarctic Ice Sheet stability

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