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Notes
High tide of the warm Pliocene: Implications of global sea level for Antarctic deglaciation

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ABSTRACT

We obtained global sea-level (eustatic) estimates with a peak of ~22 m higher than present for the Pliocene interval 2.7–3.2 Ma from backstripping in Virginia (United States), New Zealand, and Enewetak Atoll (north Pacific Ocean), benthic foraminiferal δ¹⁸O values, and Mg/Ca-δ¹⁸O estimates. Statistical analysis indicates that it is likely (68% confidence interval) that peak sea level was 22 ± 5 m higher than modern, and extremely likely (95%) that it was 22 ± 10 m higher than modern. Benthic foraminiferal δ¹⁸O values appear to require that the peak was ~20–21 m. Our estimates imply loss of the equivalent of the Greenland and West Antarctic ice sheets, and some volume loss from the East Antarctic Ice Sheet, and address the longstanding controversy concerning the Pliocene stability of the East Antarctic Ice Sheet.

INTRODUCTION

Pliocene studies allow evaluation of relationships among global climate, atmospheric CO₂, and sea-level changes under conditions significantly warmer than today, but with a similar paleogeographic configuration (Raymo et al., 2009, 2011; Rohling et al., 2009). Paleotemperature proxies indicate that average global surface temperatures ca. 3 Ma were 2–3 °C warmer than present (Dowsett, 2007). Atmospheric CO₂ estimates for the warm Pliocene are not well constrained (330–415 ppmv; e.g., Pagani et al., 2010), but appear comparable to 390 ppmv measured in 2011 (Common Era, CE) and higher than preanthropogenic levels (280 ppmv).

Published estimates of the peak Pliocene sea level have a wide range, though a ~25 m peak is widely cited (e.g., Raymo et al., 2009; Rohling et al., 2009). A peak of 35 m was obtained by estimating uplift rates for the Orangeburg scarp in North and South Carolina (southeastern United States; +35 ± 18 m; Dowsett and Cronin, 1990); a similar estimate was obtained from uplifted deposits in Alaska (+40 m; Brigham-Grette and Carter, 1992) (Fig. 1). The ~25 m estimate for the highstand generally cited is based on Dowsett and Cronin, 1990, as updated by Dowsett et al. (1999) to be consistent with a lesser ice inventory. The 25 m estimate was not independently derived. Melting of all modern ice sheets would raise sea level by 64 ± 4 m, with 7 m from Greenland and 5 m from the West Antarctic Ice Sheet (WAIS) (Lythe et al., 2001). Thus, an estimate of a 25–35 m peak implies full deglaciation of Greenland and the WAIS, and significant removal (~25%–~45%) of the East Antarctic Ice Sheet (EAIS).

Pliocene global sea-level changes have been reconstructed using records from atolls (Wardlaw and Quinn, 1991), benthic foraminiferal δ¹⁸O (Kennett and Hodell, 1995; Miller et al., 2005, 2011), Mg/Ca (Sosdian and Rosenthal, 2009), and continental margins (Naish and Wilson, 2009). Each method has its limitations. Sea-level changes recorded in coral atolls provide precise water-depth changes, but modeling subsidence rates and dating can be challenging. The δ¹⁸O method is complicated by separating deep water temperature from δ¹⁸O seawater changes due to ice volume variations. Mg/Ca analyses provide a temperature proxy that, combined with δ¹⁸O records, can isolate δ¹⁸O seawater though it is complicated by uncertainties in species calibrations and carbonate ion effects (e.g., Lear et al., 2004). Sea-level changes recorded in passive continental margin sequences include the effects of subsidence and uplift, sediment loading, compaction, and uncertainties in paleowater depth. Backstripping, a technique that progressively removes the effects of compaction, loading,
and thermal subsidence, provides a means of estimating global sea-level changes (see the summary in Miller et al., 2005). Here we use the following data sets to estimate the peak in Pliocene sea level (Figs. 1 and 2) with its implications for ice volume history: (1) backstripped records from Virginia (United States) and New Zealand; (2) δ18O constraints; (3) two different Mg/Ca-δ18O based estimates; and (4) a backstripped estimate from Enewetak Atoll. These results address the dynamics and stability of the mid-Pliocene ice sheets (placing bounds on likely volume loss) under elevated atmospheric CO2 conditions comparable to anthropogenic levels.

**EUSTATIC ESTIMATES**

New estimates for Pliocene sea level are derived from backstripping of the Eyreville, Virginia, corehole drilled in the moat of the Late Eocene (35.4 Ma) Chesapeake Bay impact structure (Fig. 2). Geochronologic resolution of ~0.5–1.0 m.y. was obtained by integration of Sr isotope data and biostratigraphy. Water depth estimates are based on changes in quantitative grain size and benthic foraminiferal biofacies with well-constrained water depth estimates (±10 m in inner neritic and slightly worse in middle-outer neritic environments). Simple one-dimensional backstripping was used to account for the effects of subsidence and loading in the New Zealand sections. Some eustatic lowstands correlate with erosional unconformities, and their estimates of eustatic amplitude must be considered a minimum; however, many of the lowstands are associated with a relative conformity, indicating preservation of the entire 41 k.y. cycle. Eustatic errors are ±10–15 m.

Peak eustatic estimates from Enewetak Atoll, Virginia, and New Zealand are similar (Figs. 1 and 2). Enewetak and Virginia backstripped records do not record sea-level lowstands due to hiatuses, and therefore do not record full amplitudes of eustatic changes. Nevertheless, the peak sea-level values among the three backstripped records are similar in the interval between 2.7 and 3.2 Ma (10–18 m in Virginia, 15–20 m in New Zealand, 20–25 m in Enewetak; Table DR1 in the Data Repository; Fig. 2).

Benthic foraminiferal δ18O records provide constraints on ice volume, subject to certain assumptions. We used the benthic foraminiferal δ18O record of Lisiecki and Raymo (2005) (Figs. 1 and 2), differencing Pliocene benthic foraminifera δ18O from zero age δ18O values. We calculate sea level by attributing 67% to ice and 33% to temperature on glacial-interglacial scales; 80:20 and 50:50 ice:temperature attributions provide end-member assumptions for computing errors (Fig. DR1; Table DR2). We assumed –40‰ for δ18Oice implying a 0.1‰/10 m sea-level δ18O calibration, consistent with previous calibrations (Fairbanks and Matthews, 1978). The –40‰ value is bracketed by δ18O values of –35‰ for Greenland and –42‰ for West Antarctic ice (Lhomme et al., 2005); end members for polar ice sheets are –30‰ to –50‰ (see the Data Repository). Making these assumptions, the eustatic peak was ~21 ± 10 m ca. 2.95 Ma (Fig. 2; Table DR1). Though minimum deep-sea δ18O values are well constrained (2.92‰ ± 0.05‰), our error estimate is ±10 m (Table DR1), given the uncertainties in apportioning temperature and ice volume effects and uncertainties in the δ18Oice (see the Data Repository).

Both foraminiferal (Sosdian and Rosenthal, 2009) and ostracod (Dwyer and Chandler, 2009) Mg/Ca-δ18O records show larger-amplitude eustatic variations compared with the New Zealand and scaled δ18O records.
values are 3.2‰. If melting of ice sheets raised sea level by >20–21 m, the peak to <20–21 m. Minimum deep-sea implications

suggest a peak of <20–21 m and a <1 °C warming in much warming throughout the deep sea. Thus, we conclude that benthic foraminiferal values (Sosdian and Rosenthal, 2009); the extent of warming throughout the deep sea represented by the warming of the deep North Atlantic by 2–3 °C (Sosdian and Rosenthal, 2009) record reduces the uncertainty, but uncertainties on peak estimates for the Mg/Ca-δ18O records are still ±15–25 m (1σ; Table DR1). The sea-level peaks in the Dwyer and Chandler (2009) record are consistent with sea-level estimates from other methods (Figs. 1 and 2; Fig. DR1), and we use these estimates without smoothing.

Comparison of all records suggests that the eustatic peak in the Pliocene was 22 m (Tables DR1 and DR2; Fig. DR3), lower than the 35–40 m obtained from North and South Carolina and Alaska (Dowsett and Cronin, 1990; Brigham-Grette and Carter, 1992; Fig. 2). Comparison of the various estimates considered here show that virtually all are below 25 m except for single points based upon the Mg/Ca method. Though each method has relatively large assumed errors, pooling the data gives an empirical estimate of actual uncertainty for individual sea-level estimates of ±8.6 m (1 standard deviation = 68% confidence; see the Data Repository, Figs. DR2 and DR3; Table DR2). Averaging individual estimates for each highstand yields uncertainties of ±4–5 m and ±8–10 at the 68% and 95% confidence intervals, respectively (see the Data Repository, Table DR2). Thus, in Intergovernmental Panel on Climate Change (2007) parlance, the eustatic peak from 2.7 to 3.2 Ma was likely (68%) in the range of 22 ± 5 m relative to present sea level (ca. 3.16 Ma), and was extremely likely (95%) to be in the range of 22 ± 10 m above present (Table DR2).

Though our statistical analysis of the multiple data sets yields an estimate of 22 ± 10 m, benthic foraminiferal δ18O values appear to constrain the peak to <20–21 m. Minimum deep-sea δ18O values from 2.7 to 3.0 Ma in the Lisiecki and Raymo (2005) stack are 2.9‰, whereas zero age δ18O values are 3.2‰. If melting of ice sheets raised sea level by >20–21 m, then δ18O_ocean would have changed by >0.2‰ ± 0.4‰ (see the Data Repository); with <0.14‰ ascribable to temperature, bottom waters would have been <0.5 °C warmer than modern. Mg/Ca data indicate warming of the deep North Atlantic by 2–3 °C (Sosdian and Rosenthal, 2009); the extent of warming throughout the deep sea represented by the Lisiecki and Raymo (2005) stack is unclear. PRISM (Pliocene Research, Interpretation and Synoptic Mapping) temperature anomalies for the Antarctic Bottom Water source region are ~1 °C, versus >5 °C in the surface waters of the North Atlantic (Dowsett, 2007). This explains the warmer deep-water temperatures in the North Atlantic, but also suggests moderate warming throughout the deep sea. Thus, we conclude that benthic foraminiferal values suggest a peak of <20–21 m and a <1 °C warming in much of the deep sea.

IMPLICATIONS

Our lower peak sea level has implications for ice inventory. The EIS has great thermal inertia and displays significant hysteresis in models, requiring >800 ppm pCO2 levels to cause major surface ablation (DeConto and Pollard, 2003) of its 20.5 × 106 km3 of grounded ice. These physical constraints and other geological data have prompted one school to argue for minimal Pliocene melting of the EIS (stability hypothesis; Marchant et al., 1993; Kennett and Hodell, 1995), while acknowledging peripheral melting of the EIS. In contrast, the “dynamists” school has argued for severe reduction of the EIS to as much as two-thirds of its present size (e.g., Webb and Harwood, 1991). A 12 m increase in sea level (the lower bound of our extremely likely, i.e., 95%, range) requires the loss of the equivalent of the Greenland ice sheet (7 m; 2.9 × 106 km3 of grounded ice above sea level) and the WAIS (5 m; 2.1 × 106 km3 grounded ice above sea level) (Lythe et al., 2001). Our statistical best estimate of 22 m also suggests ~10 m sea-level equivalent loss of the relatively stable EIS, implying a volume of ~80% of modern. Such a Pliocene ice mass loss from Antarctica is consistent with a coupled ice-ocean-atmosphere model (DeConto and Pollard, 2003) and a model capable of simulating marine grounded ice sheet dynamics (Pollard and DeConto, 2009) showing +8 m Pliocene eustatic contribution from Antarctica. In the model, East Antarctica remains largely glaciated ca. 3 Ma, with very thick (>4 km) nodes in the Dronning Maud Land and Gambutsev Plateau, and thick coverage (>2 km) of the Transantarctic Mountains. New Ross Sea drilling data show a dynamic Pliocene ice sheet or ice shelf, with periodic collapse and warm open-water conditions during Pliocene interglacials (Naish et al., 2009). The models are also in accord with geological constraints from terrestrial fossil material and glacial deposits in the Transantarctic Mountains that imply a relatively stable, cold, polar EIS at higher elevations (above +1500 m) since 13.8 Ma (Lewis et al., 2007). Our far-field sea-level data reconcile sea-level, temperature, and ice sheet records and support the relative stability of the interior EIS under atmospheric CO2 levels similar to today. However, it also suggests that the equilibrium condition for sea level under today’s atmospheric CO2 levels requires the nearly total deglaciation of both Greenland and the WAIS, with a contribution of ~10 m from the low-lying, marine-based coastal margins of the EIS.

Our sea-level estimates highlight the limitations of reconstructing global sea level from continental margin records due to the effects of tectonism (thermal and nonthermal), isostatic response, and other errors. Peltier (1998) demonstrated that the whole Earth response to removal of large ice sheets results in major regional differences in relative sea-level history, due to spatial and temporal variations in the viscoelastic response to unloading and to the changes in Earth rotation termed glacial isostatic adjustment (GIA). This results in differences in regional GIA-induced sea-level effects of ~5–10 m during the late Pleistocene to Holocene that may affect the reference level for the Pliocene. Raymo et al. (2011) evaluated potential GIA effects on Pliocene sea-level reconstructions for a range of meltwater scenarios, and showed that they can influence relative sea level on a scale of ~10 m (similar to the errors in our backstripping estimate); they concluded that reconstructing eustasy can only be done by modeling GIA effects in combination with numerous regionally distributed relative sea-level estimates. Although the regional variations in relative sea-level rise that are induced by the rapid melting of a polar ice sheet may be large initially, these variations are rapidly reduced by the subsequent rebound of the crust that occurs in the regions in which ice has been eliminated. A more uniform rise in the rise of sea level everywhere is therefore expected within a period of ~10 k.y., thus supporting our inferences for several well-separated sites of a very similar rise of sea level in the mid-Pliocene.

We acknowledge that reconstructing regional sea-level variations for the Pliocene and older records is complicated by both the influence of Pliocene and more recent GIA effects and those of regional and local tectonics. Though backstripping models the effects of thermal subsidence, compaction, and loading, it does not account for nonthermal subsidence including GIA effects, and the errors in backstripping are ~±10 m, about the same amplitude as the geoidal signal. Despite these uncertainties, we note that peak estimates are similar in three different areas of potential geoidal effects (Eniwetok, Virginia, and New Zealand) and are remarkably similar to the estimates from oxygen isotopes and Mg/Ca. Taken together, our data provide empirical evidence that the precision of sea-level estimates is ~±10 m. We also argue that the deep-sea oxygen isotopic records place similar constraints on sea level, though we conclude that our estimate of a 22 m peak has necessarily large errors (±10 m at 95% confidence). Nevertheless, even considering the large errors, it is clear that
the contribution of polar ice sheet melt to mean global sea-level rise during the Pliocene encompassed at least the equivalent of the present-day Greenland and West Antarctic ice sheets, and we regard it very likely that several meters of eustatic rise can be attributed to ice loss from the marine margins of East Antarctica.

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