Geology

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Geology 2012;40;407-410
doi: 10.1130/G32869.1

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THE GEOLOGICAL SOCIETY OF AMERICA

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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 $\delta^{18}O$ values, and Mg/Ca-$\delta^{18}O$ 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 $\delta^{18}O$ 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 long-standing controversy concerning the Pliocene stability of the East Antarctic Ice Sheet.

INTRODUCTION

Pliocene studies allow evaluation of relationships among global climate, atmospheric CO_2, 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 CO_2 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 $\delta^{18}O$ (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 $\delta^{18}O$ method is complicated by separating deep water temperature from $\delta^{18}O_{seawater}$ changes due to ice volume variations. Mg/Ca analyses provide a temperature proxy that, combined with $\delta^{18}O$ records, can isolate $\delta^{18}O_{seawater}$, 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/or uplift, sediment loading, compaction, and uncertainties in paleowater depth. Backstripping, a technique that progressively removes the effects of compaction, loading,

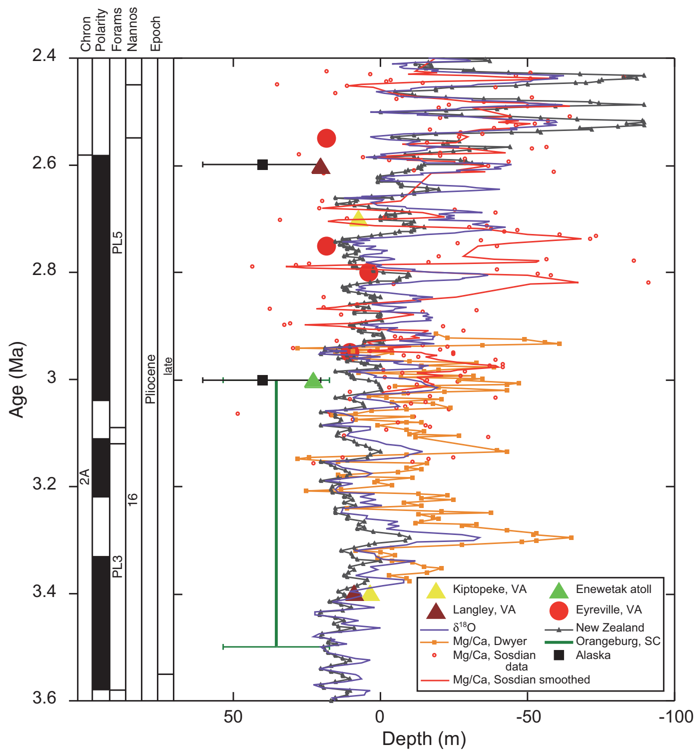

Figure 1. Comparison of late Pliocene sea-level estimates: $\delta^{18}O$ scaled to sea level using assumptions outlined here (see also Miller et al., 2005); Mg/Ca-based estimates (Sosdian and Rosenthal, 2009; Dwyer and Chandler, 2009); Enewetak Atoll (Pacific Ocean; Wardlaw and Quinn, 1991); Eyreville, Virginia (United States; this study); Kiptopeke and Langley, Virginia (Hayden et al., 2008); New Zealand (Naish, 1997; Naish and Wilson, 2009; this study); Orangeburg scarp, South Carolina (Dowsett and Cronin, 1990); and Alaskan terraces (Brigham-Grette and Carter, 1992). Green error bars indicate error estimate for uplift for Orangeburg scarp (Dowsett and Cronin, 1990). PL—Pliocene zone.

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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) $\delta^{18}O$ constraints; (3) two different Mg/Ca-$\delta^{18}O$ 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 CO_2 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 detailed lithofacies, ichnologic, and benthic foraminiferal analyses and are relatively precise (±10 m) due to the shallow-water environments, where proxies are most sensitive (shoreface to inner neritic; 5–25 m paleodepth) (Browning et al., 2009). A time-dependent compaction model for impact-generated materials coupled with one-dimensional backstripping quantifies the effects of impact, regional tectonics, and eustatic change (Kulpecz et al., 2009; Kulpecz, 2008). The Eocene postimpact section was strongly influenced by the initial impact and subsequent time-dependent compaction of impact materials, whereas the Oligocene to early Miocene was affected by regional nonthermal tectonism (Kulpecz et al., 2009; see Hayden et al., 2008, for similar results from other Chesapeake Bay impact structure coreholes, i.e., Kiptopeke and Langley, Virginia, Fig. 1). There were few tectonic or impact-related effects on deposition during the Pliocene in this region (Kulpecz et al., 2009; Hayden et al., 2008), and thus backstripping potentially provides eustatic amplitude estimates for this period with an uncertainty of ~±10–15 m. Nevertheless, the Virginia Pliocene sections only capture a few Pliocene highstands (Fig. 1).

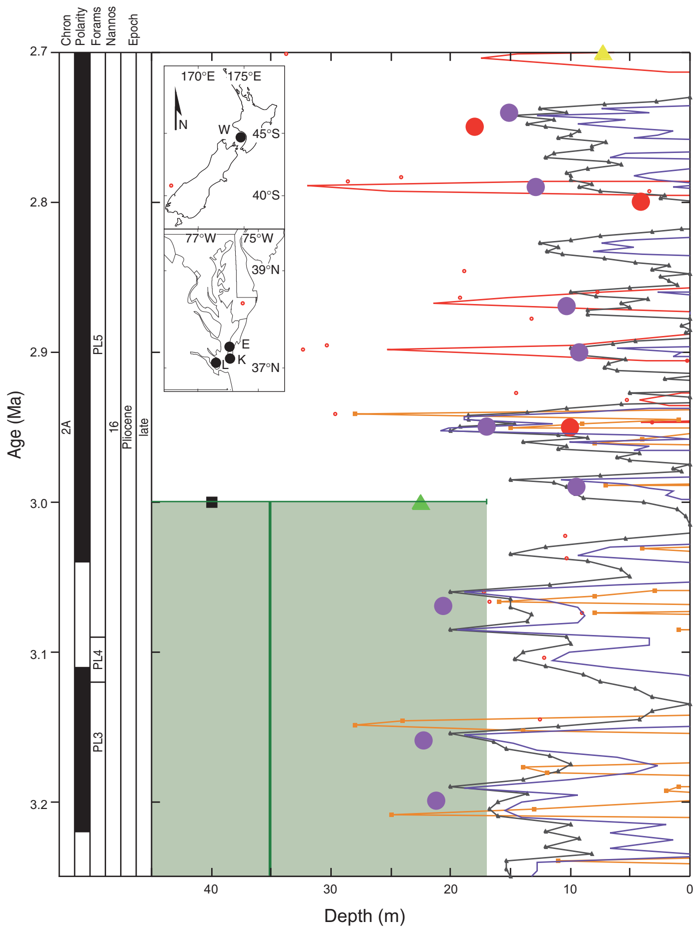

Figure 2. Enlargement of Figure 1, focusing on time interval of highest sea level; location maps are inset. Purple circles are mean values (Tables DR2 and DR3; see footnote 1). Age of Enewetak Atoll is poorly constrained (3.0 ± 0.5 Ma) and is averaged with 2.99 Ma peak observed in astronomically dated proxies. Error estimates are provided in Figure DR1. PL—Pliocene zone. Green shaded zone indicates error estimate for Orangeburg scarp.

The Wanganui Basin, New Zealand, contains a relatively complete Pliocene–Pleistocene record (Naish, 1997; Naish and Wilson, 2009). Uplift has exposed ~1500 m of Pliocene sediments in the Rangitikei Valley in the eastern Wanganui Basin, spanning 3.6–1.8 Ma. Biostratigraphy and magnetostratigraphy were used to constrain correlation of sedimentary cycles to astronomical (41 k.y.) cycles reflected in global benthic foraminiferal $\delta^{18}O$ records (e.g., Lisiecki and Raymo, 2005). Here we provide a revised eustatic estimate for the Wanganui Basin using a new astronomical age model (see the GSA Data Repository[1]). Sequence stratigraphic and lithofacies analysis indicates deposition in neritic environments, and 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 correlative 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 $\delta^{18}O$ records provide constraints on ice volume, subject to certain assumptions. We used the benthic foraminiferal $\delta^{18}O$ record of Lisiecki and Raymo (2005) (Figs. 1 and 2), differencing Pliocene benthic foraminifera $\delta^{18}O$ from zero age $\delta^{18}O$ 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 $\delta^{18}O_{ice}$, implying a 0.1‰/10 m sea-level $\delta^{18}O$ calibration, consistent with previous calibrations (Fairbanks and Matthews, 1978). The –40‰ value is bracketed by $\delta^{18}O$ 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 $\delta^{18}O$ 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 $\delta^{18}O_{ice}$ (see the Data Repository).

Both foraminiferal (Sosdian and Rosenthal, 2009) and ostracod (Dwyer and Chandler, 2009) Mg/Ca-$\delta^{18}O$ records show larger-amplitude eustatic variations compared with the New Zealand and scaled $\delta^{18}O$ records

[1]GSA Data Repository item 2012112, New Zealand age control, oxygen isotope assumptions, and uncertainty analysis, is available online at www.geosociety.org/pubs/ft2012.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

(Figs. 1 and 2). Lear et al. (2004) showed that Mg/Ca-based eustatic estimates may have a higher amplitude signal because of error propagation of Mg/Ca and benthic $\delta^{18}O$ measurements, and suggested using a weighted fit to smooth records and reduce the uncertainties. The Sosdian and Rosenthal (2009) record shows many points >10 m higher than peak estimates from other methods; a 3-point average smoothing reduces the incompatibility (Figs. 1 and 2). We use peak estimates based on averaging points in 12 k.y. windows (2–3 points in each window) (Table DR1; Fig. DR1); these peak estimates still show a higher variability than that other data sets (Fig. DR2), but we are reluctant to use a larger window given the 40 k.y. characteristic Pliocene sea-level cycle. Averaging of points in the Sosdian and Rosenthal (2009) record reduces the uncertainty, but uncertainties on peak estimates for the Mg/Ca-$\delta^{18}O$ 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 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 $\delta^{18}O$ values appear to constrain the peak to <20–21 m. Minimum deep-sea $\delta^{18}O$ values from 2.7 to 3.0 Ma in the Lisiecki and Raymo (2005) stack are 2.9‰, whereas zero age $\delta^{18}O$ values are 3.2‰. If melting of ice sheets raised sea level by >20–21 m, then $\delta^{18}O_{seawater}$ would have changed by >0.2‰ ± 0.04‰ (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 EAIS has great thermal inertia and displays significant hysteresis in models, requiring >800 ppm pCO_2 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 EAIS (stability hypothesis; Marchant et al., 1993; Kennett and Hodell, 1995), while acknowledging peripheral melting of the EAIS. In contrast, the "dynamicists" school has argued for severe reduction of the EAIS 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; 2.9×10^6 km^3 of grounded ice above sea level) and the WAIS (5 m; 2.1×10^6 km^3 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 EAIS, 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 EAIS 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 EAIS under atmospheric CO_2 levels similar to today. However, it also suggests that the equilibrium condition for sea level under today's atmospheric CO_2 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 EAIS.

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 (Enewetak, 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.

ACKNOWLEDGMENTS

We thank M. Siddall, J-B. Clavaud (Chevron), and an anonymous reviewer for comments and the U.S. Geological Survey Eastern Region Mapping Team drillers for collecting the Virginia cores. Funded by National Science Foundation grants EAR-070778 and EAR-1052257.

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Manuscript received 28 September 2011
Revised manuscript received 1 December 2011
Manuscript accepted 5 December 2011

Printed in USA