Article:

Gasson, E. orcid.org/0000-0003-4653-6217, DeConto, R. and Pollard, D. (2016) Modeling the oxygen isotope composition of the Antarctic ice sheet and its significance to Pliocene sea leve. Geology. ISSN 0091-7613

https://doi.org/10.1130/G38104.1

Publisher: GSA
Journal: GEOL: Geology
DOI:10.1130/G38104.1

Modeling the oxygen isotope composition of the Antarctic ice sheet and its significance to Pliocene sea level

Edward Gasson[1,2], Robert M. DeConto[1], and David Pollard[3]

[1]Department of Geosciences, University of Massachusetts, Amherst, Massachusetts, 01003, USA

[2]Department of Geography, The University of Sheffield, Sheffield, S10 2TN, UK

[3]Earth and Environmental Systems Institute, Pennsylvania State University, University Park, Pennsylvania, 16802, USA

ABSTRACT

Recent estimates of global mean sea level based on the oxygen isotope composition of mid-Pliocene benthic foraminifera vary from 9 to 21 m above present, which has differing implications for the past stability of the Antarctic ice sheet during an interval with atmospheric CO_2 comparable to present. Here we simulate the oxygen isotope composition of the Antarctic ice sheet for a range of configurations using isotope-enabled climate and ice sheet models. We identify which ice-sheet configurations are consistent with the oxygen isotope record and suggest a maximum contribution from Antarctica to the mid-Pliocene sea level highstand of ~13 m. We also highlight that the relationship between the oxygen isotope record and sea level is not constant when ice is lost from deep marine basins, which has important implications for the use of oxygen isotopes as a sea level proxy.

INTRODUCTION

There is significant interest in constraining the mid-Pliocene warm period (MPWP) sea level highstand as it provides an opportunity to determine the magnitude of maximum ice-sheet retreat during an interval with atmospheric CO_2 comparable to present day (Raymo et al., 2009; Balco, 2015; Martínez-Botí et al., 2015). The MPWP has also been used to evaluate climate and ice sheet model performance (e.g., De Boer et al., 2015; DeConto and Pollard, 2016; Haywood et al., 2016), however these efforts have been limited by poorly constrained sea-level estimates (Raymo et al., 2011). Paleoshorelines can be used to reconstruct local relative sea level, however interpreting a global eustatic signal from these records is complicated by glacio-isostatic adjustment, dynamic topography, and vertical tectonics (Moucha et al., 2008; Raymo et al., 2011; Rovere et al., 2014). Efforts to correct for dynamic topography using models currently have large uncertainties (Dutton et al., 2015).

Alternative approaches to estimating past sea level make use of the change in the oxygen isotope composition of benthic foraminifera relative to present values ($\Delta\delta^{18}O_{benthic}$) and either independent estimates of deep-sea temperature (Dwyer and Chandler, 2009) or simple assumptions based on the partitioning of $\Delta\delta^{18}O_{benthic}$ between ice volume ($\Delta\delta^{18}O_{seawater}$) and temperature components ($\Delta\delta^{18}O_{temperature}$) (Miller et al., 2012). The definition of $\Delta\delta^{18}O_{benthic}$ for the MPWP is important and will impact sea level estimates based on this approach. $\Delta\delta^{18}O_{benthic}$ is variously cited as either 0.3‰ (e.g., Miller et al., 2012; Winnick and Caves, 2015) or 0.4‰ (e.g., Dutton et al., 2015), depending on whether the differences are for Modern:MPWP (0.3‰) or Holocene:MPWP (0.4‰), where the MPWP is the interglacial MIS G17 and $\delta^{18}O_{benthic}$ is from the multi-site stack of Lisiecki and Raymo (2005). There are additional uncertainties from analytical error in measuring $\delta^{18}O_{benthic}$ and potential hydrographic effects (e.g. Dutton et al., 2015; Woodard et al., 2014; see Data Repository for more discussion on calculation of $\Delta\delta^{18}O_{benthic}$).

In addition to changes in deep-sea temperature and ice mass, a number of recent studies have also considered how changes in the mean oxygen isotope composition of the ice sheets ($\delta^{18}O_{ice}$) may affect interpretation of $\Delta\delta^{18}O_{benthic}$ (Langebroek et al., 2010; Winnick and Caves, 2015; Gasson et al., 2016). Changes in $\delta^{18}O_{ice}$ may be expected due to changes in surface climate, with mean $\delta^{18}O_{ice}$ typically increasing in warmer climates (Langebroek et al., 2010). Studies which have considered changes in $\delta^{18}O_{ice}$ include those which have simulated changes in $\delta^{18}O$ of precipitation ($\delta^{18}O_{precip.}$) falling on the ice sheet surface which is then tracked within the ice sheet based on internal ice velocities to determine the mean oxygen isotopic composition of the ice sheet (DeConto et al., 2008; Langebroek et al., 2010; Wilson et al., 2013; Gasson et al., 2016) and a novel isotope mass-balance approach that estimated changes in $\delta^{18}O_{ice}$ based on modern $\delta^{18}O_{precip.}$ to surface temperature relationships (Winnick and Caves, 2015). Here we follow the former approach and calculate $\delta^{18}O_{ice}$ for a range of proposed MPWP Antarctic ice sheet configurations (Pollard and DeConto, 2009; Mengel and Levermann, 2014; Austermann et al., 2015; Pollard et al., 2015; DeConto and Pollard, 2016; Dowsett et al., 2016) and ultimately calculate the Antarctic contribution to $\Delta\delta^{18}O_{benthic}$. This approach captures the relative influences of climate and ice sheet geometry on its isotope composition, and has the added advantage of quantifying the relative contributions of marine-based ice lying above and below sea-level, to both global mean sea-level and $\delta^{18}O_{seawater}$.

DOI:10.1130/G38104.1

Pliocene Antarctic ice sheet simulations

Previous simulations of the Antarctic ice sheets during the MPWP have shown various stages of ice sheet retreat compared with modern – some of these simulations are included in Table 1 and Figure DR1. These range from configurations with a collapsed West Antarctic ice sheet and limited retreat of the East Antarctic ice sheet (e.g., Pollard and DeConto, 2009; De Boer et al., 2015) to those with retreat deep into the East Antarctic subglacial basins (e.g., Hill et al., 2007; Dolan et al., 2011; Pollard et al., 2015) and various states in-between (e.g., Mengel and Levermann, 2014; Austermann et al., 2015). The contribution to MPWP sea-level rise from the various Antarctic ice sheets shown in Table 1 ranges from 3.7 to 17.8 m. Although these configurations can be compared with sea level estimates and ice proximal evidence, they have not been assessed in terms of compatibility with the $\delta^{18}O_{benthic}$ record. Here we repeat or approximate these simulations using an oxygen isotope enabled climate and ice sheet model (Mathieu et al., 2002; Wilson et al., 2013) to determine the Antarctic ice sheet contribution to $\Delta\delta^{18}O_{benthic}$. The climate model used to calculate $\delta^{18}O_{precip}$ is setup as per DeConto et al., (2012) for all simulations, with modifications to the Antarctic ice sheet topography and extent.

The magnitude of $\Delta\delta^{18}O_{benthic}$ that can be attributed to changes in Antarctic ice-sheet mass varies depending on assumptions about the ratio of $\Delta\delta^{18}O_{seawater}$:$\Delta\delta^{18}O_{temperature}$ and retreat of the Greenland ice sheet during the MPWP. Here we calculate an Antarctic contribution to $\Delta\delta^{18}O_{benthic}$ of 0.18 ±0.13 ‰ (2σ). This is calculated using Monte Carlo error propagation (e.g. Anderson, 1976) and follows previous assumptions for the $\Delta\delta^{18}O_{seawater}$:$\Delta\delta^{18}O_{temperature}$ ratio (50:50 to 80:20), retreat of the Greenland ice sheet during the MPWP (50 – 100 %) (Miller et al., 2012; Winnick and Caves, 2015) and also considering different magnitudes of $\Delta\delta^{18}O_{benthic}$ (0.3–0.4 ‰). We apply an additional uncertainty of ±0.1 ‰ to $\Delta\delta^{18}O_{benthic}$ to account for analytical errors in measuring $\delta^{18}O$ of benthic foraminifera. Note that assuming a larger $\Delta\delta^{18}O_{benthic}$ of 0.4‰ and following the isotope mass-balance approach of Winnick and Caves (2015) we calculate a total Antarctic (East and West Antarctic ice sheets) contribution to MPWP sea level of 8–12 m (3–8 m assuming $\Delta\delta^{18}O_{benthic}$ of 0.3‰) with a total MPWP sea level rise of 13–18 m[1].

RESULTS AND DISCUSSION

Our calculations assume that each ice-sheet state in Table 1 and its internal $\delta^{18}O$ distribution has fully equilibrated with the corresponding steady-state climate. The $\delta^{18}O$ at each internal point within the ice is determined by Lagrangian tracing of steady-state velocities back to the surface and assigning the $\delta^{18}O$ to that of $\delta^{18}O_{precip}$ at the surface location. The $\delta^{18}O_{precip}$ values are taken from a corresponding GCM simulation. The $\delta^{18}O$ values are then integrated over the whole ice sheet to obtain the average $\delta^{18}O_{ice}$ values (Wilson et al., 2013). We do the same for our pre-industrial control climate and ice sheet. The $\Delta\delta^{18}O_{ice}$ values in Table 1 are the differences from this control. We then calculate $\Delta\delta^{18}O_{seawater}$ following a mass-balance approach similar to that of Winnick and Caves (2015); see Data Repository for equations.

[1] [1]Assuming a $\Delta\delta^{18}O_{seawater}$:$\Delta\delta^{18}O_{temperature}$ ratio of 67:33 and a 1–4‰ increase in mean $\delta^{18}O_{ice}$, as per Winnick and Caves (2015) scenarios 1 and 2.

DOI:10.1130/G38104.1

As a result of changes in surface climate, our MPWP GCM simulations show an increase in mean $\delta^{18}O_{precip}$ falling over the Antarctic ice sheet of 6.0–7.5‰. However this does not lead to an equivalent change in mean $\delta^{18}O_{ice}$, which has a smaller increase of between 2.7 and 3.9‰ in our ice sheet simulations. This is due to changes in ice sheet geometry and ice flow (see Figure 1). As the MPWP ice sheet retreats inland and away from the ocean it retreats into regions that have lower $\delta^{18}O_{precip}$ values. Additionally, the ice lost in warm climate simulations is often the isotopically heavier ice from the ice margins as illustrated by comparing the simulated $\Delta\delta^{18}O_{ice}$ and $\Delta\delta^{18}O_{precip}$ values for Austermann et al., (2015) with PRISM4 in Table 1. Therefore in MPWP simulations, if there is no change in $\delta^{18}O_{precip}$ the Antarctic ice sheet would become isotopically lighter (a decrease in $\delta^{18}O_{ice}$), rather than heavier. These effects should be considered when using the modern relationship between surface temperature and $\delta^{18}O_{precip}$ to infer changes in $\delta^{18}O_{ice}$ with changing surface climate (Winnick and Caves, 2015).

We next calculate how much the simulated MPWP Antarctic ice sheets would alter $\delta^{18}O_{seawater}$ and hence the proportion of $\Delta\delta^{18}O_{benthic}$ that can be attributed to changes in Antarctic ice mass. The calculated Antarctic contribution to $\delta^{18}O_{seawater}$ varies across simulations from 0.16 to 0.39‰ because of differences in ice mass and changing mean $\delta^{18}O_{ice}$. All of the simulated MPWP Antarctic ice sheets are within the expected range of the $\delta^{18}O_{benthic}$ record (0.18 ±0.13‰), with the exception of the simulation of Pollard et al. (2015), which has greater retreat than can be inferred from the $\delta^{18}O_{benthic}$ record based on our prior assumptions. This is in part because of a 3.9‰ increase of mean $\delta^{18}O_{ice}$ for the Pollard et al. (2015) ice sheet in the warm climate of the MPWP, however even with no

DOI:10.1130/G38104.1

change in $\delta^{18}O_{ice}$, the $\delta^{18}O_{seawater}$ contribution (0.35‰) is still outside the range of the $\delta^{18}O_{benthic}$ record.

These simulations have a sea level range of 3.7–17.8 m, highlighting that the often-used calibration of $\delta^{18}O_{seawater}$ 0.01‰ per m of sea level change is not appropriate for Antarctica (Figure 2; Miller et al., 2012; Winnick and Caves, 2015). This is in part due to the generally lighter values of mean $\delta^{18}O_{ice}$ for the Antarctic ice sheet compared with Northern Hemisphere ice sheets and because of changes in $\delta^{18}O_{ice}$ with changing climate (Winnick and Caves, 2015). However another difference between $\delta^{18}O_{seawater}$ and sea level records occurs when there is ice loss from subglacial basins. The sea level change calculated here takes into account the infilling with seawater of subglacial basins once they have been evacuated of ice. However $\delta^{18}O_{seawater}$ is a total ice mass signal and this effect is not relevant to $\delta^{18}O_{seawater}$. There is therefore a decoupling between $\delta^{18}O_{seawater}$ and sea level when there is significant change in marine based ice sheets, such as the West Antarctic and large portions of the East Antarctic Ice Sheet. When ice is lost from marine basins there will be a larger change in $\delta^{18}O_{seawater}$ than expected for the equivalent change in sea level (see Figure 2). Note that this effect is accounted for by Winnick and Caves (2015) for the Antarctic ice sheet on average, but that study does not account for geographical locations of ice loss, which means that this effect is more pronounced here (see Figure DR2). Repeating our isotope mass-balance calculations with our simulated range for $\Delta\delta^{18}O_{ice}$ and accounting for the nonlinear relationship between ice sheet mass change and sea level we calculate a total range for the Antarctic contribution to MPWP sea level of -1 to 13 m (see Data Repository).

The nonlinear relationship between $\delta^{18}O_{seawater}$ and sea level partially causes the large $\delta^{18}O_{seawater}$ contribution for the Pollard et al. (2015) Antarctic ice sheet, which has significant loss of ice from deep subglacial basins, such as the Aurora and Wilkes subglacial basins. A number of the other MPWP ice sheet simulations have partial retreat into the East Antarctic subglacial basins, in particular the Wilkes Subglacial Basin (Mengel and Levermann, 2014; Austermann et al., 2015; DeConto and Pollard, 2016). Retreat into the Wilkes Subglacial Basin during the MPWP is also supported by a sediment provenance study (Cook et al., 2013). The simulations with retreat into the Wilkes but not the Aurora Subglacial Basin are within the expected range of the $\delta^{18}O_{benthic}$ record. It can therefore be inferred that the constraints provided by the $\delta^{18}O_{benthic}$ record indicates that there was not complete retreat into all of the deep Antarctic subglacial basins during the MPWP. It should be noted that our simulations account for increased precipitation in the continental interior under a warmer MPWP climate. This increased precipitation generates interior ice sheet thickening that offsets some of the ice lost from subglacial basins (Yamane et al., 2015).

CONCLUSIONS AND IMPLICATIONS

A recent reinterpretation of the $\delta^{18}O_{benthic}$ record suggested that there was negligible mass loss from the East Antarctic ice sheet during the MPWP (Winnick and Caves, 2015). Our results do not exclude this scenario; indeed the simulations that have a collapsed West Antarctic but stable East Antarctic ice sheet (Pollard and DeConto, 2009; De Boer et al., 2015) are within the uncertainty of the $\delta^{18}O_{benthic}$ record – as is the possibility of a small negative (-1 m) Antarctic contribution to MPWP sea level. However, these simulations are not consistent with other geological evidence for retreat of portions of the East Antarctic ice sheet during the MPWP (Cook et al., 2013). A fuller consideration of the uncertainties in the $\delta^{18}O_{benthic}$ record, in particular in calculations of $\Delta\delta^{18}O_{benthic}$, does allow for greater mass loss from Antarctica during the MPWP. Ultimately, the $\delta^{18}O_{benthic}$ record cannot exclude the possibility of large-scale retreat of the Antarctic ice sheet during the mid-Pliocene warm period, but it does imply that the Antarctic ice sheet contribution to sea level rise was at most ~13 m.

SUPPLEMENTARY INFORMATION

Isotope budget calculations:

Calculations shown in Table 1 follow the mass-balance approach used by Winnick and Caves (2015):

$$M_{\mathrm{O}} + M_{\mathrm{GIS}} + M_{\mathrm{WAIS}} + M_{\mathrm{EAIS}} = M_{\mathrm{pO}} + M_{\mathrm{pGIS}} + M_{\mathrm{pWAIS}} + M_{\mathrm{pEAIS}}$$

(1)

$$\begin{aligned} &M_{\mathrm{O}}\delta^{18}\mathrm{O}_{\mathrm{SW}} + M_{\mathrm{GIS}}\delta^{18}\mathrm{O}_{\mathrm{GIS}} + M_{\mathrm{WAIS}}\delta^{18}\mathrm{O}_{\mathrm{WAIS}} + M_{\mathrm{EAIS}}\delta^{18}\mathrm{O}_{\mathrm{EAIS}} = \\ &M_{\mathrm{pO}}\delta^{18}\mathrm{O}_{\mathrm{pSW}} + M_{\mathrm{pGIS}}\delta^{18}\mathrm{O}_{\mathrm{pGIS}} + M_{\mathrm{pWAIS}}\delta^{18}\mathrm{O}_{\mathrm{pWAIS}} + M_{\mathrm{pEAIS}}\delta^{18}\mathrm{O}_{\mathrm{pEAIS}} \end{aligned}$$

(2)

Where M_x is the total mass of the modern (M_O) and Pliocene (M_{pO}) oceans and Greenland (GIS), West Antarctic (WAIS) and East Antarctic (EAIS) ice sheets. Winnick and Caves (2015) solved these equations for M_{pO} and M_{pEAIS}. In this study we are interested in the Antarctic contribution to $\Delta\delta^{18}O_{seawater}$ and are using Pliocene Antarctic ice sheet model simulations for M_{pEAIS}. We therefore ignore changes in the Greenland ice sheet and treat the West and East Antarctic ice sheets together:

$$M_{\mathrm{O}} + M_{\mathrm{AIS}} = M_{\mathrm{pO}} + M_{\mathrm{pAIS}} \quad (3)$$

$$M_{\mathrm{O}}\delta^{18}\mathrm{O}_{\mathrm{SW}} + M_{\mathrm{AIS}}\delta^{18}\mathrm{O}_{\mathrm{AIS}} = M_{\mathrm{pO}}\delta^{18}\mathrm{O}_{\mathrm{pSW}} + M_{\mathrm{pAIS}}\delta^{18}\mathrm{O}_{\mathrm{pAIS}} \quad (4)$$

We take M_{pAIS} and $\delta^{18}\mathrm{O}_{\mathrm{pAIS}}$ from our simulations and then calculate $\delta^{18}\mathrm{O}_{\mathrm{pSW}}$ (the Antarctic contribution to $\Delta\delta^{18}\mathrm{O}_{\mathrm{benthic}}$):

$$\delta^{18}\mathrm{O}_{\mathrm{pSW}} = \frac{M_{\mathrm{O}}\delta^{18}\mathrm{O}_{\mathrm{SW}} + M_{\mathrm{AIS}}\delta^{18}\mathrm{O}_{\mathrm{AIS}} - M_{\mathrm{pAIS}}\delta^{18}\mathrm{O}_{\mathrm{pAIS}}}{M_{\mathrm{O}} + M_{\mathrm{AIS}} - M_{\mathrm{pAIS}}} \quad (5)$$

Our simulated value for the modern mean oxygen isotopic composition of the Antarctic ice sheet is −33.8‰. This is considerably higher than the calculated whole Antarctic value of −54.7‰ of Lhomme et al. (2005). The reason for this discrepancy is due in part to a GCM bias of ~10‰ in modern values for $\delta^{18}\mathrm{O}_{\mathrm{precip.}}$ when compared with observations over the Antarctic interior, caused by modeled surface temperatures that are too warm (Mathieu et al., 2002). Additionally, our values are in equilibrium with the modern surface climate, in contrast to the values from Lhomme et al. (2005) which account for change in $\delta^{18}\mathrm{O}_{\mathrm{ice}}$ through successive glacial periods with predominantly lower $\delta^{18}\mathrm{O}_{\mathrm{precip}}$. We therefore use the Lhomme et al. (2005) values (−54.7‰) throughout for modern $\delta^{18}\mathrm{O}_{\mathrm{AIS.}}$ For Pliocene values of $\delta^{18}\mathrm{O}_{\mathrm{pAIS}}$ we use the anomaly between our Pliocene ($\delta^{18}\mathrm{O}_{\mathrm{PLIOCENE}}$) and pre-industrial control ($\delta^{18}\mathrm{O}_{\mathrm{CONTROL}}$) simulations:

$$\delta^{18}\mathrm{O}_{\mathrm{pAIS}} = -54.7 + \left(\delta^{18}\mathrm{O}_{\mathrm{PLIOCENE}} - \delta^{18}\mathrm{O}_{\mathrm{CONTROL}}\right) \quad (6)$$

Note that we do not change mean ocean $\delta^{18}\mathrm{O}$ in our GCM simulations. The GCM used is an isotope-enabled version of the GENESIS GCM.

DOI:10.1130/G38104.1

Calculation of $\Delta\delta^{18}O_{benthic}$:

In the main paper we highlight two different values for $\Delta\delta^{18}O_{benthic}$ in the literature, 0.31‰ for Modern:G17 and 0.40‰ for Holocene:G17. Although we do not suggest a preference for either value, here we discuss reasons for these differences. The modern value for $\delta^{18}O_{benthic}$ in the LR04 stack (Lisiecki and Raymo, 2005) is 3.23‰, compared with 3.32‰ when averaged over the last 10 kyr and 2.92‰ during MIS G17 (2.95 Ma). The higher value for Holocene $\delta^{18}O_{benthic}$ may be a result of the ice sheets having lower mean $\delta^{18}O_{ice}$ as they would be less equilibrated to modern $\delta^{18}O_{precip}$. Additionally, remnant glacial ice in the early Holocene would also lead to higher values for $\delta^{18}O_{benthic}$ when averaged over the last 10 kyr. Both of these arguments would suggest that Modern:MPWP should be used for $\Delta\delta^{18}O_{benthic}$ over Holocene:MPWP. However, these effects could also lead to higher $\delta^{18}O_{benthic}$ during MPWP interglacials, which are time-averaged due to poor temporal resolution (2.5–5 kyr). A potential solution would be to use a high-resolution $\delta^{18}O_{benthic}$ record. The recently published deep-Pacific ODP Site 1208 shows a Modern:MPWP $\Delta\delta^{18}O_{benthic}$ of 0.49‰ (Woodard et al., 2015). However individual sites may also be affected by ocean circulation changes (this could equally affect the LR04 stack which is weighted towards the Atlantic). Indeed other high-resolution sites show a Modern:MPWP $\Delta\delta^{18}O_{benthic}$ <0.3‰ (M. Patterson, personal communication). A more detailed statistical analysis of $\Delta\delta^{18}O_{benthic}$ is required for individual sites (e.g. Mudelsee et al., 2014), which is beyond the scope of this paper. Here we use the range of 0.3–0.4‰ for $\Delta\delta^{18}O_{benthic}$ and add a ±0.1‰ uncertainty for analytical error, giving a total range of 0.2–0.5‰.

Calculation of Antarctic contribution to MPWP sea level:

We calculate maximum and minimum contributions from the Antarctic ice sheets to MPWP sea level. From $\Delta\delta^{18}O_{benthic}$ we calculate an Antarctic ice sheet component of 0.18 ±0.13 ‰ using Monte-Carlo error propagation to account for uncertainty in the $\Delta\delta^{18}O_{seawater}$:$\Delta\delta^{18}O_{temperature}$ ratio, retreat of the Greenland ice sheet and analytical error in $\delta^{18}O_{benthic}$. Rearranging equation (5) we determine the Antarctic ice sheet mass change required for $\Delta\delta^{18}O_{seawater}$ at the upper (0.31‰) and lower ends (0.05‰) of our error estimate. For $\Delta\delta^{18}O_{ice}$ we use the upper and lower estimates from our simulations (+2.7 to +3.9‰). This results in an Antarctic mass change of +0.53 to -6.61 $\times10^{18}$kg. To account for the nonlinear relationship between ice sheet mass change and sea level due to the effect of marine-subglacial basins we use the calibration generated from an Antarctic ice sheet deglaciation simulation (Figure DR2) to convert from mass change to sea level change. This leads to a lower estimate for the Antarctic contribution to MPWP sea level of -1.4 m (a sea level fall) and an upper estimate of +13.1 m.

ACKNOWLEDGMENTS

We thank Ken Miller, Matthew Winnick and one anonymous reviewer for their constructive comments and suggestions. We also thank Molly Patterson, Ben Keisling and Kate Hibbert for discussions that helped improve the manuscript. This study was supported by US National Science Foundation awards OCE-1202632 (“PLIOMAX”) and AGS-1203910.

REFERENCES CITED

Anderson, G.M., 1976, Error propagation by the Monte Carlo method in geochemical calculations: Geochimica et Cosmochimica Acta, v. 40, no. 12, p. 1533–1538.

Austermann, J., Pollard, D., Mitrovica, J.X., Moucha, R., Forte, A.M., DeConto, R.M., Rowley, D.B., and Raymo, M.E., 2015, The impact of dynamic topography change on Antarctic ice sheet stability during the mid-Pliocene warm period: Geology, v. 43, p. 927–930, doi:10.1130/G36988.1.

Balco, G., 2015, The absence of evidence of absence of the East Antarctic Ice Sheet: Geology, v. 43, p. 943–944, doi:10.1130/focus102015.1.

Cook, C.P., et al., 2013, Dynamic behaviour of the East Antarctic ice sheet during Pliocene warmth: Nature Geoscience, v. 6, p. 765–769, doi:10.1038/ngeo1889.

De Boer, B., et al., 2015, Simulating the Antarctic ice sheet in the late-Pliocene warm period: PLISMIP-ANT, an ice-sheet model intercomparison project: The Cryosphere, v. 9, no. 3, p. 881–903, doi:10.5194/tc-9-881-2015.

DeConto, R.M., and Pollard, D., 2016, Contribution of Antarctica to past and future sea-level rise: Nature, v. 531, p. 591–597, doi:10.1038/nature17145.

DeConto, R.M., Pollard, D., Wilson, P.A., Pälike, H., Lear, C.H., Pagani, M., 2008, Thresholds for Cenozoic bipolar glaciation: Nature, v. 455, p.652–656, doi:10.1038/nature07337

DeConto R.M., Pollard, D., Kowalewski, D., 2012, Modeling Antarctic ice sheet and climate variations during Marine Isotope Stage 31: Global and Planetary Change, v. 88, p 45–52, doi:10.1016/j.glopacha.2012.03.003

Dolan, A.M., Haywood, A.M., Hill, D.J., Dowsett, H.J., Hunter, S.J., Lunt, D.J., and Pickering, S.J., 2011, Sensitivity of Pliocene ice sheets to orbital forcing: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 309, p. 98–110, doi:10.1016/j.palaeo.2011.03.030.

Dowsett, H., Dolan, A., Rowley, D., Pound, M., Salzmann, U., Robinson, M., Chandler, M., Foley, K., and Haywood, A., 2016, The PRISM4 (mid-Piacenzian) palaeoenvironmental reconstruction: Climate of the Past Discussions, v. 4, p. 1–39, doi:10.5194/cp-2016-33.

Dutton, A., Carlson, A.E., Long, A.J., Milne, G.A., Clark, P.U., DeConto, R., Horton, B.P., Rahmstorf, S., and Raymo, M.E., 2015, Sea-level rise due to polar ice-sheet mass loss during past warm periods: Science, v. 349, doi:10.1126/science.aaa4019.

Dwyer, G.S., and Chandler, M.A., 2009, Mid-Pliocene sea level and continental ice volume based on coupled benthic Mg/Ca palaeotemperatures and oxygen isotopes: Philosophical transactions. Series A, Mathematical, physical, and engineering sciences, v. 367, no. p. 157–168, doi:10.1098/rsta.2008.0222.

Gasson, E., DeConto, R.M., Pollard, D., and Levy, R.H., 2016, Dynamic Antarctic ice sheet during the early to mid-Miocene: Proceedings of the National Academy of Sciences of the United States of America, p. 201516130, doi:10.1073/pnas.1516130113.

Haywood, A.M., Dowsett, H.J., and Dolan, A.M., 2016, Integrating geological archives and climate models for the mid-Pliocene warm period: Nature Communications, v. 7, doi:10.1038/ncomms10646.

Hill, D.J., Haywood, A.M., Hindmarsh, R.C.A., and Valdes, P.J., 2007, Characterizing ice sheets during the Pliocene: Evidence from data and models, *in* Williams, M., and Haywood, A.M., eds., Deep-Time Perspectives on Climate Change: Marrying the signal from Computer Models and Biological Proxies., p. 517–538.

Langebroek, P.M., Paul, A., and Schulz, M., 2010, Simulating the sea level imprint on marine oxygen isotope records during the middle Miocene using an ice sheet–climate model: Paleoceanography, v. 25, PA4203, doi:10.1029/2008PA001704.

Lhomme, N., Clarke, G.K.C., and Ritz, C., 2005, Global budget of water isotopes inferred from polar ice sheets: Geophysical Research Letters, v. 32, L20502, doi:10.1029/2005GL023774.

Lisiecki, L.E., and Raymo, M.E., 2005, A Pliocene-Pleistocene stack of 57 globally distributed benthic δ18O records: Paleoceanography, v. 20, doi:10.1029/2004PA001071.

Martínez-Botí, M.A., Foster, G.L., Chalk, T.B., Rohling, E.J., Sexton, P.F., Lunt, D.J., Pancost, R.D., Badger, M.P.S., and Schmidt, D.N., 2015, Plio-Pleistocene climate sensitivity evaluated using high-resolution CO2 records: Nature, v. 518, p. 49–54, doi:10.1038/nature14145.

Mathieu, R., Pollard, D., Cole, J.E., White, J.W.C., Webb, R.S., and Thompson, S.L., 2002, Simulation of stable water isotope variations by the GENESIS GCM for modern conditions: Journal of Geophysical Research, v. 107, p. 1–18, doi:10.1029/2001JD900255.

Mengel, M., and Levermann, A., 2014, Ice plug prevents irreversible discharge from East Antarctica: Nature Climate Change, v. 4, p. 451–455, doi:10.1038/nclimate2226.

Miller, K.G., Wright, J.D., Browning, J.V., Kulpecz, A., Kominz, M., Naish, T.R., Cramer, B.S., Rosenthal, Y., Peltier, W.R., and Sosdian, S., 2012, High tide of the warm Pliocene: Implications of global sea level for Antarctic deglaciation: Geology, v. 40, p. 407–410, doi:10.1130/G32869.1.

Moucha, R., Forte, A.M., Mitrovica, J.X., Rowley, D.B., Quéré, S., Simmons, N.a., and Grand, S.P., 2008, Dynamic topography and long-term sea-level variations: There is no such thing as a stable continental platform: Earth and Planetary Science Letters, v. 271, p. 101–108, doi:10.1016/j.epsl.2008.03.056.

Mudelsee, M., Bickert, T., Lear, C.H., Lohmann, G., 2014, Cenozoic climate changes: A review based on time series analysis of marine benthic $\delta^{18}O$ records: Reviews of Geophysics, v. 52, no. 3, p. 333–374, doi:10.1002/2013RG000440

Pollard, D., and DeConto, R.M., 2009, Modelling West Antarctic ice sheet growth and collapse through the past five million years: Nature, v. 458, p. 329–332, doi:10.1038/nature07809.

Pollard, D., and DeConto, R.M., 2012, Description of a hybrid ice sheet-shelf model, and application to Antarctica: Geoscientific Model Development, v. 5, p. 1273–1295, doi:10.5194/gmd-5-1273-2012.

Pollard, D., DeConto, R.M., and Alley, R.B., 2015, Potential Antarctic Ice Sheet retreat driven by hydrofracturing and ice cliff failure: Earth and Planetary Science Letters, v. 412, p. 112–121, doi:10.1016/j.epsl.2014.12.035.

Raymo, M.E., Hearty, P., De Conto, R., O'Leary, M., Dowsett, H.J., Robinson, M.H., and Mitrovica, J.X., 2009, PLIOMAX: Pliocene maximum sea level project: PAGES News, v. 17, p. 58–59.

Raymo, M.E., Mitrovica, J.X., O'Leary, M.J., DeConto, R.M., and Hearty, P.J., 2011, Departures from eustasy in Pliocene sea-level records: Nature Geoscience, v. 4, p. 328–332, doi:10.1038/ngeo1118.

Rovere, A., Raymo, M.E., Mitrovica, J.X., Hearty, P.J., O'Leary, M.J., and Inglis, J.D., 2014, The Mid-Pliocene sea-level conundrum: Glacial isostasy, eustasy and dynamic topography: Earth and Planetary Science Letters, v. 387, p. 27–33, doi:10.1016/j.epsl.2013.10.030.

Wilson, D.S., Pollard, D., Deconto, R.M., Jamieson, S.S.R., and Luyendyk, B.P., 2013, Initiation of the West Antarctic Ice Sheet and estimates of total Antarctic ice volume in the earliest Oligocene: Geophysical Research Letters, v. 40, p. 4305–4309, doi:10.1002/grl.50797.

Winnick, M.J., and Caves, J.K., 2015, Oxygen isotope mass-balance constraints on Pliocene sea level and East Antarctic Ice Sheet stability: Geology, v. 43, p. 879–882, doi:10.1130/G36999.1.

Woodard, S.C., Rosenthal, Y., Miller, K.G., Wright, J.D., Chiu, B.K., Lawrence, K.T., 2014, Antarctic role in Northern Hemisphere glaciation: Science, v. 346, p. 847–851, doi:10.1126/science.1255586.

Yamane, M., Yokoyama, Y., Abe-Ouchi, A., Obrochta, S., Saito, F., Moriwaki, K., and Matsuzaki, H., 2015, Exposure age and ice-sheet model constraints on Pliocene East Antarctic ice sheet dynamics: Nature Communications, v. 6, doi:10.1038/ncomms8016.

DOI:10.1130/G38104.1

TABLE 1. ANTARCTIC ICE SHEET CONTRIBUTION TO SEA LEVEL AND $\Delta\delta^{18}O_{seawater}$

Study	Δsea level (m)	Δmass (10^{18} kg)	$\Delta\delta^{18}O_{precip}$ (‰)	$\Delta\delta^{18}O_{ice}$ (‰)	$\Delta\delta^{18}O_{seawater}$ (‰)	$\Delta\delta^{18}O_{seawater}$ (‰)
Pollard and DeConto, 2009*	3.7	-2.8	6.0	2.9	0.16	0.11#
Mengel and Levermann, 2014†	5.4	-3.6	6.4	3.6	0.20	0.14#
Austermann et al., 2015*	8.8	-4.9	6.6	3.6	0.25	0.20#
PRISM4§	9.1	-5.0	6.6	2.7	0.24	0.20#
DeConto and Pollard, 2016*	11.3	-6.1	7.5	3.8	0.29	0.25#
Pollard et al., 2015*	17.8	-8.8	7.2	3.9	0.39	0.35#

Note: Calculations assume a modern mean $\delta^{18}O_{ice}$ of −54.7‰ for the Antarctic ice sheet (Lhomme et al. 2005). Values for $\delta^{18}O_{ice}$ are equilibrium values following 300 kyr of simulation, shown relative to a pre-industrial control simulation

*reproduced with addition of isotopes

†approximate whole ice sheet simulation, showing similar response to Wilkes Basin simulations of Mengel and Levermann (2014). This simulation is created by imposing high ocean melt rates (a factor of 10 increase in the value of K in Equation 17 of (Pollard and DeConto, 2012)) and uses the same ice sheet physics as the Pollard and DeConto (2009) simulation

§PRISM4 ice sheet is maintained by imposing a surface mass balance mask which prevents areal expansion of the ice sheet from its starting position, ice thicknesses are maintained by adjusting basal sliding coefficients

#no change in $\delta^{18}O_{ice}$, values fixed at modern

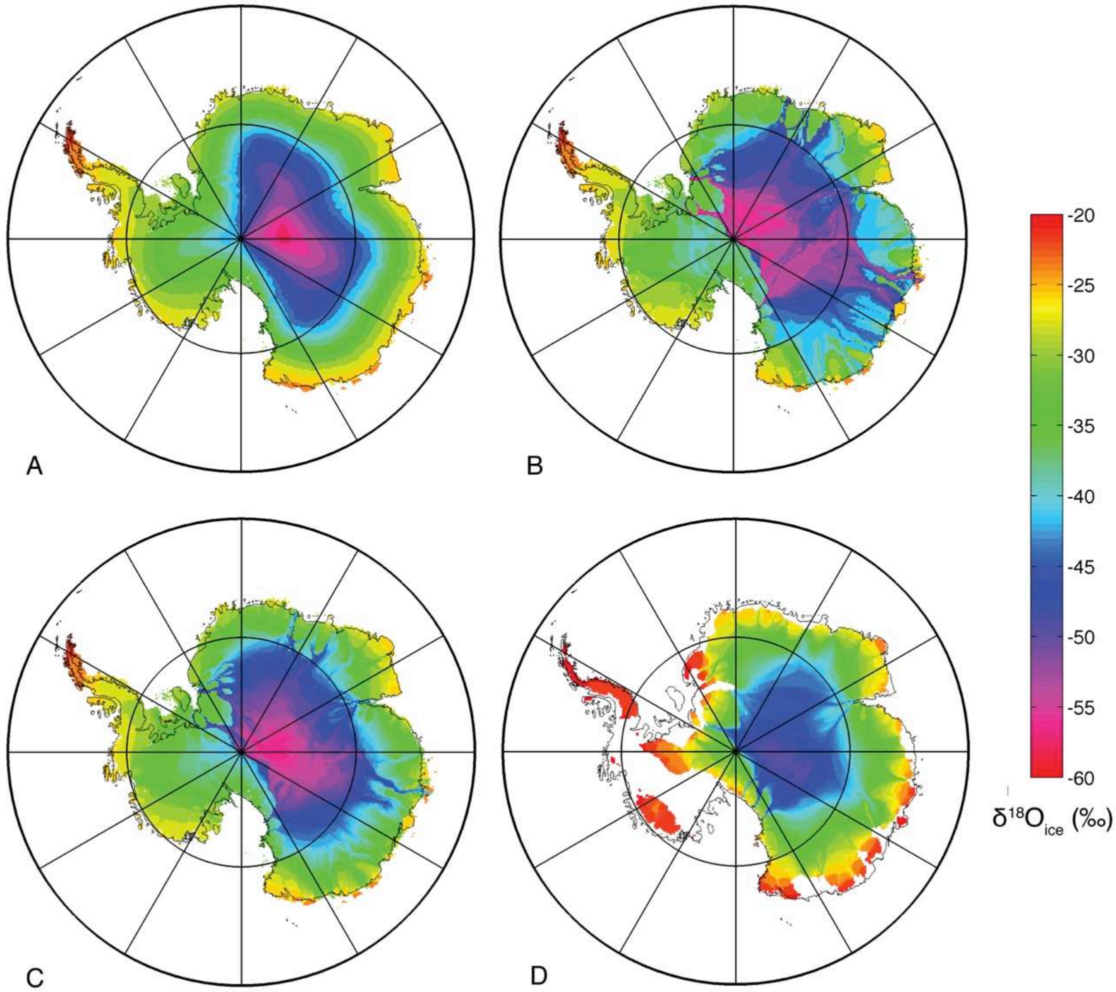

Figure 1. Simulated $\delta^{18}O_{ice}$ at different ice depths. A: Modern ice surface layer. B: Modern ice basal layer. C: Modern depth averaged. D: Pliocene depth averaged (DeConto and Pollard, 2016).

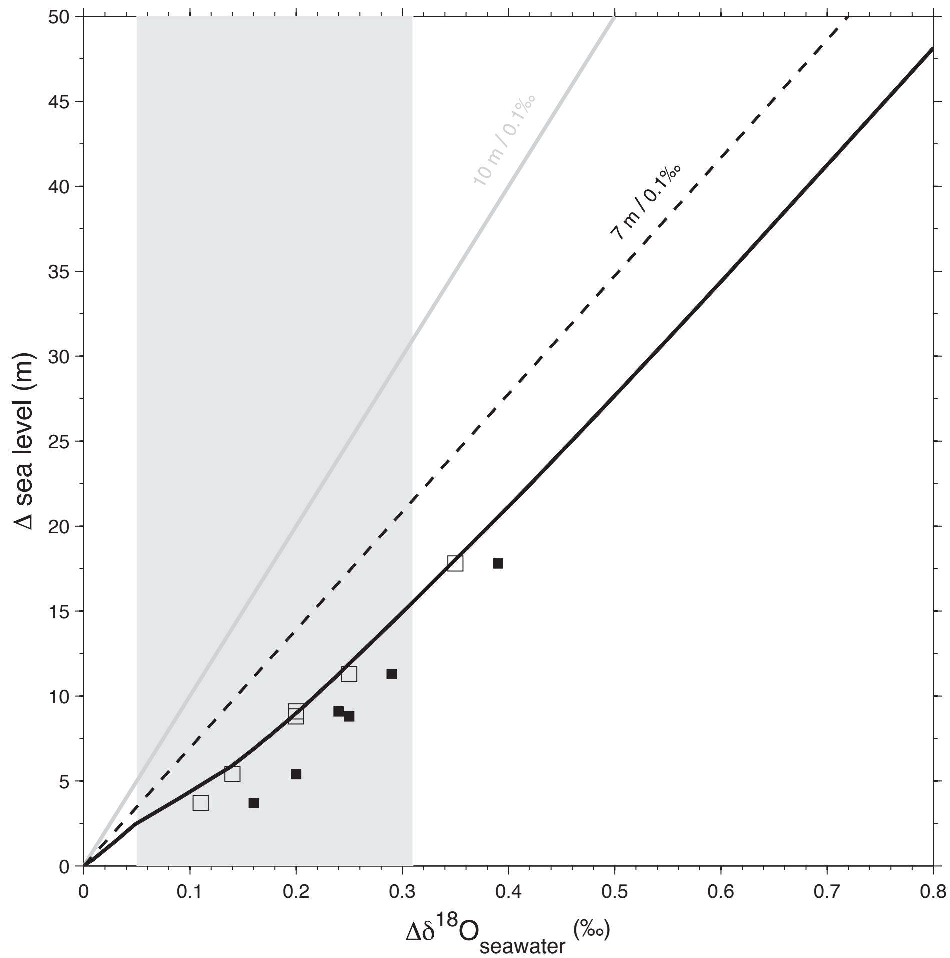

Figure 2. Relationship between $\delta^{18}O_{seawater}$ and sea level. Gray line shows the commonly used 10 m / 0.1 ‰ calibration (e.g. Woodard et al., 2014). Black dashed line shows the calibration for Antarctica for a mean $\delta^{18}O_{ice}$ of -54.7 ‰ (Lhomme et al., 2005), ignoring the impact of marine-based ice on sea level. Black line is an Antarctic ice sheet deglacial simulation, for a mean $\delta^{18}O_{ice}$ of -54.7 ‰, the change in gradient of the black line is caused by the initial loss of marine-based ice. White and black squares are Antarctic ice sheet simulations from Table 1, with fixed $\delta^{18}O_{ice}$ and changing $\delta^{18}O_{ice}$, respectively. Gray band shows the Antarctic contribution to $\Delta\delta^{18}O_{seawater}$ (0.18±0.13 ‰).

Publisher: GSA
Journal: GEOL: Geology
DOI:10.1130/G38104.1

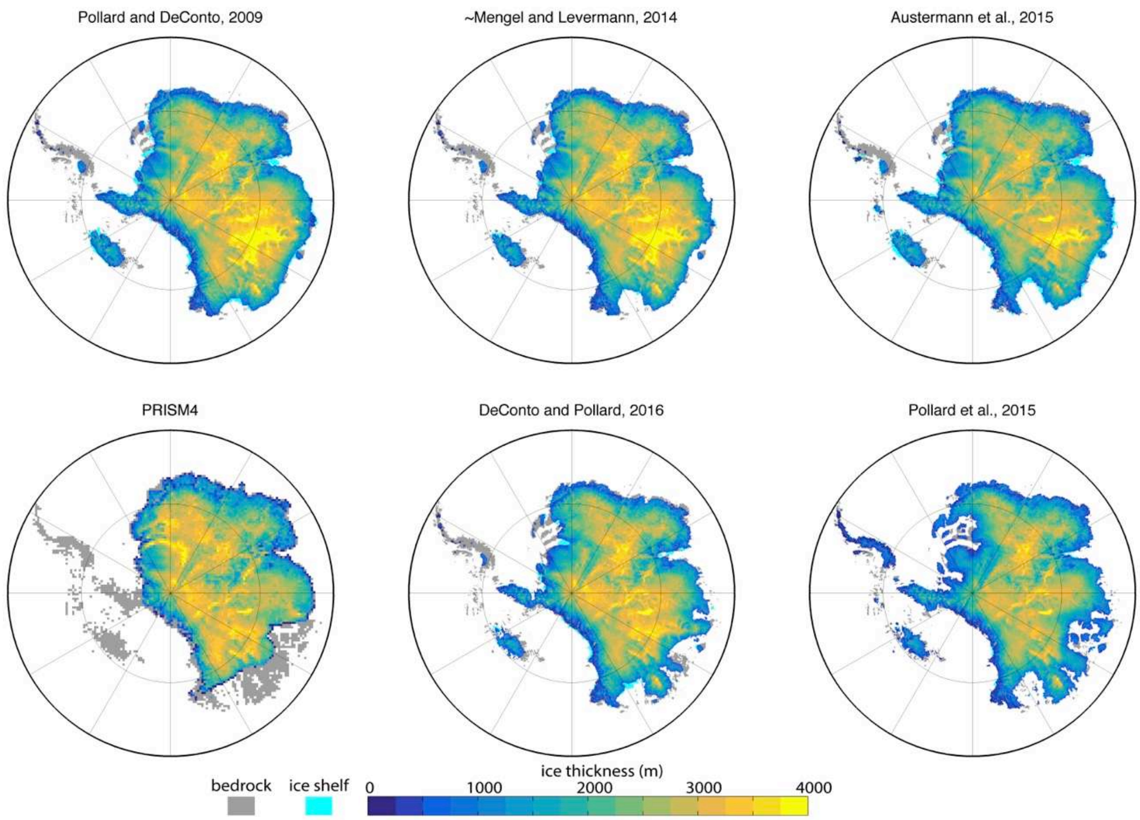

Figure DR1. Previously published Antarctic ice sheet simulations for the mid-Pliocene warm period, repeated or approximated here using isotope enabled climate and ice sheet models.

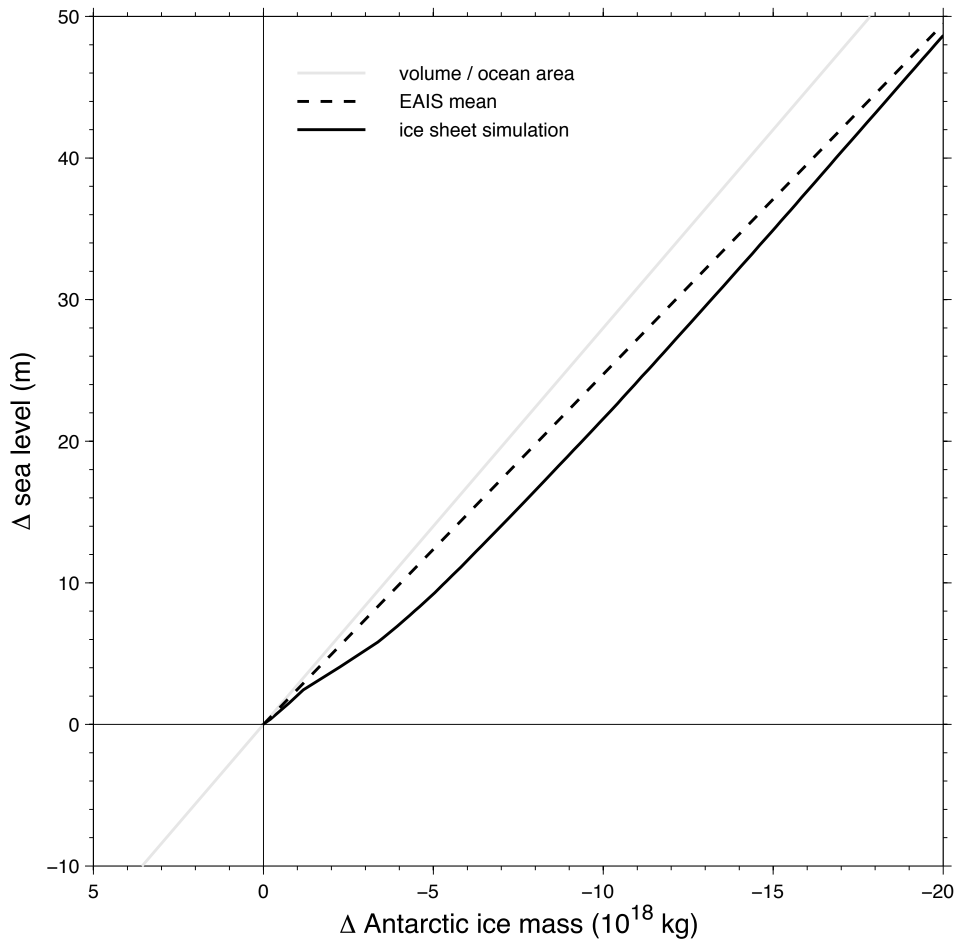

Figure DR2. Relationship between Antarctic ice sheet mass change and sea level change. Gray line shows ice volume divided by ocean area after accounting for the change in state from ice to seawater. Dashed black line is the mean relationship for the East Antarctic ice sheet for a total ice mass of 21.55 $\times 10^{18}$kg and sea level rise of 53.3 m (Fretwell et al., 2013; Winnick and Caves, 2015). Black line is from an Antarctic ice sheet deglacial simulation with initial loss of ice predominantly from marine subglacial basins.