Title

Orbital forcing of the East Antarctic ice sheet during the Pliocene and Early Pleistocene

Journal

Nature Geoscience, 7(11)

ISSN

1752-0894 1752-0908

Authors

Patterson, M. O
McKay, R.
Naish, T.
et al.

Publication Date

2014-10-26

DOI

10.1038/NGEO2273

Peer reviewed

Response of the East Antarctic Ice Sheet to orbital forcing during the Pliocene and Early Pleistocene

Patterson, M. O.[1], McKay, R.[1*], Naish, T.[1], Escutia, C.[2], Jimenez-Espejo, F. J.[3, 4], Raymo, M.E.[5]., Meyers, S.R.[6], Tauxe, L.[7], Brinkhuis, H.[8], and IODP Expedition 318 Scientists[‡]

Affiliations:

1) Antarctic Research Centre, Victoria University of Wellington, Wellington, New Zealand
2) Instituto Andaluz de Ciencias de la Tierra, CSIC – University of Granada, Granada, Spain
3) Department of Earth and Planetary Sciences, Graduate School of Environmental Studies, Nagoya University, Nagoya, Japan.
4) Institute of Biogeosciences, Japan Agency for Marine-Earth Science and Technology, Yokosuka, Japan.
5) Lamont-Doherty Earth Observatory of Columbia University
P.O. Box 1000, 61 Route 9W, Palisades, NY, 10964, USA
6) Department of Geoscience, University of Wisconsin-Madison,1215 W. Dayton St. Madison, WI 53706, USA
7) Scripps Institution of Oceanography, La Jolla, CA 92093-0220, USA
8) Marine Palynology and Paleoceanography, Laboratory of Paleobotany and Palynology, Utrecht University, Budapestlaan 4, 3584 CD Utrecht, NL

‡) a full list of authors and their affiliations appears at the end of the paper.

.

*-corresponding author.

The absence of a strong ~20-kyr precession signal in Early-Pleistocene to Pliocene geological reconstructions of global ice volume and sea-level challenges our fundamental understanding of how ice sheets respond to orbital forcing. Here, we present an orbital-scale time-series of ice-berg rafted debris from a well-dated sediment core (Integrated Ocean Drilling Program (IODP) site U1361) adjacent to the Wilkes Land margin of the East Antarctic Ice Sheet (EAIS). Our data reveal that under a warmer climate state, mean annual insolation, paced by the 41-kyr obliquity cycle, had more influence on Antarctic ice volume than summer insolation intensity modulated by the precession cycle. A transition to precession dominance after ~3.5 Ma reflects a declining influence of oceanic forcing as the high latitude southern ocean cooled and a perennial summer sea-ice field developed. Our data show that precession-paced variability of the EAIS marine margin occurs during cold climate states, even when the obliquity signal dominates globally-integrated proxy records.

Geological reconstructions of global ice volume[1] and sea-level[2] during the Pliocene and Early Pleistocene (5 to 2 Ma) display regular glacial-interglacial cycles occurring every 41-kyrs, paced by variations in Earth's axial tilt (obliquity). The absence of a strong ~20-kyr precession signal prior is confounding, because precession should impart the greatest influence on high-latitude summer insolation intensity, and therefore polar ice volume[3]. Indeed, precession is a major control on the orbital pacing of Late Pleistocene cycles[1], and while a number of hypotheses have been proposed to explain its absence prior to this time[4, 5, 6], reconciliation of this conundrum remains hampered by direct observational evidence of orbital-scale Plio-Pleistocene variance of ice sheets, in particular from Antarctica [7, 8, 9].

IODP core U1361 was recovered from ~3000 m water depth on the continental rise adjacent to the Wilkes Land sector of Antarctica, one of the largest marine based sectors of the EAIS, where recent geophysical surveys show the presence of landward deepening reverse slope troughs penetrating beneath the EAIS reaching depths of up 2 km below sea-level[11], heightening the vulnerability of this sector of the EAIS to marine ice sheet instabilities (Fig. 1). U1361 provides a well-dated and continuous geological archive of Pliocene and Early Pleistocene orbital scale variability of the marine margin of the EAIS. Sediment deposition at this site is controlled by the interplay between: (i) downslope marine sediment gravity flows triggered by the buildup of sediment on the edge of the continental shelf during glacial advance; (ii) the rainout of biogenic detritus from surface water plankton; (iii) iceberg rafting of terrigenous sediments; and (iv) low energy bottom currents (Supplementary Information).

The core consists of eighteen sedimentary cycles spanning an age range of 4.3-2.2 Ma, and comprising alternating terrigenous massive to laminated muds and diatom-rich/bearing silty-mud units (cycles 1-18 Fig. 2; Supplementary Figs. S2 and S3). In places the muds contain packages of well-defined laminae and are consistent with established models of non-erosive overbank hemipelagic deposition onto a channel levee setting via turbidites on the lowermost Antarctic continental rise[10]. Steeply dipping, seaward prograding wedge sediments are evident in seismic reflection profiles across the continental shelf and extend onto the upper continental slope above seismic unconformity WL-U8 (~4.2 Ma; Supplementary Fig. S1)[12, 13]. The geometry of these strata is characteristic of grounding zone deposition by repetitive advances of a marine-based ice sheet to the shelf edge during glacial periods[13]. Sediment overloading near the shelf break at submarine canyons heads, in turn triggers turbidity currents down slope channels and leads to overbank deposition at the core site. Low density turbidity currents in overbank "distal" channel levee environments on the Antarctic continental rise are typically non-erosive[10] (Supplementary Information). Thus, turbidite units are associated with periods of glacial advance to the Wilkes Land continental shelf edge, whereas bioturbated, diatom-rich/bearing facies represent warm interglacial periods of relatively ice-free ocean and increased primary productivity when the grounding line had migrated landward away from the shelf edge. Increased productivity during interglacial warm climates may be associated with enhanced upwelling of nutrient-rich Circumpolar Deepwater (CDW)[14] (Supplementary Information), which has been linked to southward expansion of the westerly wind field in response to a reduced pole-equator temperature gradient during past warm periods[15]. Presently this relatively warm nutrient-rich CDW upwells to the surface north of the Southern Boundary Front of the Antarctic Circumpolar Current and is marked by areas of enhanced productivity to the north of Site U1361 (Fig. 1).

We have developed a high-resolution record (~3-4 kyr sample spacing) of ice-berg rafted debris (IBRD) mass accumulation rates (MAR) for the U1361 core (Supplementary Table S1; Supplementary Fig. S4). Our age model is based on biostratigraphy used to constrain the interpretation of a magnetic polarity stratigraphy[12] (Supplementary Fig. S5), and is not orbitally-tuned. Thus, the age model for our primary spectral analyses (Fig. 3) assumes, long-term (millennial-scale) sedimentation rates between polarity reversal tie points (Supplementary Information).

In general the highest intensity of IBRD occurs during transitions from glacial terrigenous mud facies to interglacial diatom-rich/bearing muds up-core until ~47 mbsf, with most IBRD peaks immediately preceding diatom-rich facies and Ba/Al peaks (Fig. 2). Isotopic Nd and Sr provenance indicators suggest that the terrigenous components in these diatom-rich/bearing muds are associated with periods of deglacial retreat of the ice margin back into the Wilkes Land subglacial basin during the Early Pliocene[16]. The Antarctic ice sheet loses the majority of its mass via iceberg calving and sub-ice shelf melting[17]. However, these processes are intimately linked as enhanced ice shelf melt leads to reduced buttressing, that in turn acts to enhance flow of glacial tributaries in turn enhancing iceberg discharge[18]. Ice sheet reconstructions based on geological data[19]and modeling[8] suggest that the above processes dominated Antarctic mass-loss even during the warmest climates of the last 5 Ma. Thus, we interpret the maxima in IBRD MAR to be the consequence of accelerated calving during glacial retreat from marine terminating outlet glaciers along the Wilkes Land coastline as well as a contribution from EAIS outlet glaciers entering the western Ross Sea (Supplementary Information). This interpretation is consistent with models and paleo-data, which imply the most rapid mass loss of the EAIS margin during the last glacial termination occurred between 17-7 ka, and was primarily the consequence of oceanic warming[20,21].

Spectral analysis of the un-tuned IBRD MAR time-series displays a dominant period of ~40-kyr between ~4.2-3.5 Ma that transitions to strong and significant variance at ~100, and 20-kyr periods after ~3.5 Ma, with a corresponding decrease in power of the ~40-kyr cycle (Figs. 3, 4). On the basis of the strong orbital relationship displayed in frequency spectra of the un-tuned IBRD MAR time series, the paleomagnetic age model, and the apparent near continuous long-term sedimentation (Supplementary Fig. S5), we established a nominal one-to-one correlation between cycles in ice margin variability expressed by our IBRD data and orbitally-paced climatic time series (Fig. 2). Between 4.3-3.4 Ma a strong visual correlation can be observed between 40-kyr cycles in IBRD, mean annual insolation and the benthic δ^{18}O global ice volume record[1], whereas between 3.3-2.2 Ma, IBRD cycles correlate with the ~20-kyr cycles of summer insolation at 65°S (Fig. 2). We acknowledge that our visual correlations, although constrained by 7 precisely-dated paleomagnetic reversals, do not represent a unique solution but, as noted above, they are entirely consistent with the variance in orbital frequencies implied by our spectral estimations.

Using these correlative relationships (Fig. 2) and our evolutionary spectral analyses on un-tuned IBRD data (Fig. 3), we then explore the role of longer-period orbital influences on the pattern of iceberg calving. The top of the ~40-kyr-dominated interval is marked by an ~200 kyr-long condensed section between ~3.5-3.3 Ma (Supplementary Fig. S5) and corresponds to a long-term +1‰ glacial δ^{18}O excursion punctuated by smaller interglacials spanning Marine Isotope Stage (MIS) MG5 and MIS M2 (indicated by arrow in Fig. 2). The stratigraphic condensation, or possible erosion, at Site U1361 is associated with this glacial excursion. Indeed, this glacial excursion has also been associated with southern high-latitude climate cooling and the re-establishment of grounded ice on the middle to outer continental shelf in the Ross Sea following a ~200-kyr period of warm open ocean conditions[7, 9]. Previous studies of older Oligocene and Miocene δ^{18}O glacial excursions have proposed a relationship between intervals of increased glacial amplitude in the δ^{18}O record with a coincidence of 1.2 Ma nodes in obliquity and 400-kyr minima in long period eccentricity[22, 23]. This orbital configuration, which favors extended periods of cold summers and low seasonality is considered optimal for Antarctic ice sheet expansion, and occurs at ~3.3 Ma (Fig. 3c) - the time of the transition from obliquity to precession dominance in the IBRD MAR time series from U1361 (Figs. 2, 3a).

Observed ~20-kyr-duration IBRD cycles correlate with summer insolation calculated for 65°S for the interval of the core between 3.3-2.2 Ma (Fig. 2; Fig. 3a) and are embedded within 100-kyr-duration IBRD cycles with an additional peak at 400 kyr nearing the 90% confidence level (Fig 3d). Broad peaks of IBRD maxima are associated with transitions between laminated mudstones to diatom-rich/bearing muds (Fig. 2). A dramatic decrease in the amplitude of ~20-kyr IBRD peaks, and a change to lithofacies associated with non-erosive low-energy bottom currents, is observed at the core site above ~2.5 Ma, broadly coincident with southern high latitude cooling[9] and the onset of major Northern Hemisphere glaciations[24]. We attribute the progressive reduction in calving intensity to cooling and a relative stabilization of the EAIS ice margin (discussed below). Homogenization of the turbidite sediments during glacial maxima by enhanced bioturbation and bottom current activity is observed and likely reflects increased Antarctic sea ice and polynya-style mixing at this time producing oxygenated high salinity shelf water[9] that is transferred downslope over Site U1361 to form Antarctic Bottom Water[25] (Supplementary Information).

In summary, our correlations of IBRD variations with the benthic δ^{18}O stack and orbital time series identify up to seventeen ~40-kyr-duration cycles (orange lines in Fig. 2) within six major lithological cycles (cycles 13-18, Fig. 2) during the Early Pliocene (4.3-3.5 Ma). This is followed by forty ~20-kyr-duration cycles (blue lines in Fig. 2) modulated by eccentricity, that occur within twelve lithological cycles (cycles 1-12 Fig. 2).

Although the marine sediment core recovered by the ANDRILL Program from the Ross Sea region provided the first direct evidence that advance and retreat of the WAIS margin across the continental shelf was paced by obliquity during the Pliocene prior to ~3 Ma[7], sub-glacial erosion surfaces in the ANDRILL core associated with ice advance have raised the possibility of missing cycles, particularly after 3.1 Ma - e.g., the response of WAIS to orbital forcing is more ambiguous after this time. The U1361 and ANDRILL records together confirm the dynamic response of both the WAIS and EAIS marine margins to obliquity forcing during the warm Pliocene prior to the onset of southern high-latitude cooling at 3.3 Ma[9].

Geological records[7, 9, 16] and model simulations[8] of past warm climates both highlight the sensitivity of the marine-based portions of the Antarctic ice sheets to ocean warming. However, the mechanism by which the coastal ocean warms and destabilizes marine grounding lines, particularly in response to obliquity forcing, remains elusive. It has been proposed that changes in the intensity and the meridional distribution of mean annual insolation controlled by obliquity may have a profound influence on the position and strength of the Southern Hemisphere zonal westerly winds, with implications for marine ice sheet instability[7]. Indeed, an aerosolic dust record from the Southern Ocean is dominated by ~40-kyr cycles in iron and leaf-wax biomarkers prior to ~0.8 Ma[26]. Moreover, prior to ~3.3 Ma the southward expansion of the westerly wind-field over the Antarctic circumpolar convergence zone under a reduced meridional temperature gradient has been associated with a reduced sea-ice field[9]. This may have caused the upwelling of warm, CO_2-rich Circumpolar Deep Water (CDW) onto the continental shelf with consequences for the migration of marine grounding-lines[15, 27, 28].

The dominance of precession-paced variability and the corresponding reduction in obliquity influence revealed by our data after ~3.5 Ma is interpreted to reflect a declining sensitivity of the EAIS to oceanic forcing, as the southern high latitudes cooled. Both model and geological reconstructions imply that past Antarctic ice sheet expansion is closely linked with development of the sea-ice field[29] potentially resulting in northward migration of westerly winds and Southern Ocean fronts[9]. In addition, sea ice expansion after 3.5 Ma[9] likely restricted upwelling and ventilation of warm CO_2 rich CDW at the Antarctic margin acting to further enhance climate cooling. Under such a scenario, a warmer climate state during the Early to mid-Pliocene with higher atmospheric CO_2 concentration[30, 31], the increased duration and intensity of austral summer surface warming, produces a pattern approaching mean annual insolation regulated by obliquity (Fig. 2d), rather than precession that cancels out over the course of a seasonal cycle[5]. Late Pliocene cooling raised the melt threshold such that the duration of the melt season was restricted to times of austral summer insolation maxima controlled by precession (Fig. 2c), and conversely extends the winter sea ice growth season[32]. We propose this decrease in radiative forcing resulted in extensive sea-ice cover extending into much of the summer season limiting the influence of upwelling CDW on marine grounding line stability.

The significant variance at ~20 kyr precession and coincident increase in the ~100-kyr eccentricity frequency bands between 3.5-2.5 Ma is intriguing (Figs. 3a, d, and Fig. 4) and is similar to the pattern of orbital response expressed in ~100-kyr-duration glacial-interglacial cycles in the benthic $\delta^{18}O$ and greenhouse gas ice core records of the last 600-krys[33, 34]. Thus the relative dominance of eccentricity in EAIS at the Wilkes Land marine margin variability after 3.5 Ma may reflect a threshold-response of the ice sheet to orbital forcing under a colder climate regime, as has been proposed to explain the relative dominance of ~100-kyr cycles in global Late Pleistocene ice volume, albeit with different mass balance forcings[35, 36, 32]. Here we propose that the precession response of perennial sea-ice after 3.3 Ma was modulated by eccentricity (Fig. 4), such that aerial extent of sea ice was significantly reduced every ~100 kyr enhancing the influence of oceanic warming at those times.

Our data are also consistent with a reduction of obliquity variance in the benthic $\delta^{18}O$[1, 37] at ~3.5 Ma (Fig 3b), coincident with an obliquity node in the astronomical solution[38]. The gradual re-emergence of a 40-kyr signal at ~3 Ma in benthic δ^{18}O record[37] (Figs. 3b and 4c) most likely reflects a similar reemergence of strong obliquity forcing in the orbital records and possibly also a direct response of the developing northern hemisphere continental ice sheets to obliquity forcing. It is also possible that precession-driven, anti-phase oscillations in both hemispheric ice volume histories may have cancelled out in globally integrated proxy records after 3-2.8 Ma (e.g. ref. 4).

A recent reconstruction of Pliocene Arctic climate[39], ice sheet models[40], and records of iceberg rafted debris from the north Pacific and north Atlantic oceans[24] preclude a significant northern hemisphere contribution to the δ^{18}O signal of global ice volume prior to 2.8 Ma. Thus, it is unlikely that an anti-phase argument can be used to explain the dominance of the obliquity signal seen in our record prior to 3.0 Ma. The strong obliquity signal seen in our IBRD record also suggests that the intensity of summer insolation was a not a direct control on the mass balance of the EAIS prior to 3.0 Ma. The geometry of strata on the Wilkes Land continental shelf indicate that the EAIS in this region periodically expanded towards the continental shelf edge during glacial maxima in the Pliocene[12] and suggests most Antarctic ice volume variance at this time was growth and retreat of the marine-based ice sheets (Supplementary Fig. S1). During the Early Pliocene, when the sea ice field was reduced and the Wilkes Land margin of the EAIS was in more direct contact with oceanic influences, iceberg calving occurred more regularly within both glacial and interglacial facies. Based on the significant decrease in IBRD after 2.5 Ma (Fig. 2; 3) and Southern Ocean records inferring decreased SSTs[9, 41], we also infer the marine margins of the EAIS became less sensitive to ocean-induced melting compared to the WAIS[28]. After this time, the fully-glaciated East Antarctic ice volume (excluding drainage into marine embayments of the Ross and Weddell Sea) likely fluctuated at a magnitude similar to that of Late Pleistocene cycles, with a minimal contribution to the δ^{18}O signal[8, 20].

Notwithstanding this relative stability of the marine-based sectors of EAIS, model results suggest that ~20-kyr-duration fluctuations in the total Antarctic ice volume (i.e., including WAIS) may have contributed up to 15m of the total amplitude during Late Pleistocene glacial-interglacial cycles[8] and, given the δ^{18}O composition of Antarctic ice, this could have offset a larger out-of-phase precessional change in Northern Hemisphere ice volume (e.g. 20-40m) resulting in an enhanced obliquity signal in globally integrated sea-level and ice volume proxy records after 3.0-2.8 Ma (e.g., ref 4, Fig. 3b, Fig. 4). Alternatively, direct obliquity forcing of the Northern Hemisphere (NH) ice sheet is supported by proxy evidence including ice rafted debris records[24] and a recent dust flux record[42], suggesting that NH ice sheet variability (marine-based margins) and climate primarily responded to obliquity under a relatively warm NH climate state. NH cooling and ice sheet growth across the mid-Pleistocene transition ~0.8 Ma has been implicated in a similar switch to precession and eccentricity-paced glaciations, albeit by different mass balance forcing mechanisms[43]. In contrast, however, our results imply that Southern Ocean sea-ice feedbacks caused a fundamentally different response of the marine-based sectors of the EAIS under a cooler Late Pliocene/Early Pleistocene climate state, characterized by a dominance of precession-paced variability.

We conclude that prior to 3.5 Ma, under a warm climate state, EAIS demonstrates high sensitivity on orbital timescales to a relatively small increase in atmospheric CO_2 concentration and mean global surface temperature. With atmospheric CO_2 concentrations and global surface temperatures projected to remain above 400 ppm and >+2°C beyond 2100[44], our results have implications for the equilibrium response of the Antarctic ice sheets, and suggest that the marine margins of EAIS ice sheet as well as the marine-based WAIS, may become increasingly susceptible to ocean warming with the potential for widespread mass loss raising sea-level by meters over the coming centuries to millennia.

Methodology

20cc samples were treated with H_2O_2 to remove organic material and 2M NaOH to remove biogenic opal for grain size analysis. The MAR of the coarse sand fraction was then estimated using the following equation:

$$\text{IBRD MAR} = \text{CS\%} * \text{DBD} * \text{LSR}$$

where IBRD MAR is the mass accumulation rate ($g/cm^2/k.y.$), CS% is the terrigenous coarse-sand weight percent, DBD is the dry-bulk density of the nearest value (g/cm^3) and LSR is the interval average linear sedimentation rate (cm/k.y.). Visual examination of every individual sample for authigenic minerals and volcanic glass was conducted and these were absent, indicating that the IBRD volume percent was directly equivalent to the terrigenous CS%[45].

Following application of the biostratigraphic/magnetostratigraphic age model[12] and the astrochronologic age model derived in this study, we performed time series analysis using "astro: An R package for Astrochronology[46]. Prior to analysis, all data series were conservatively resampled on a 5 kyr grid using piecewise linear interpolation (the median sampling intervals for the IBRD-MAR data and benthic $\delta^{18}O$ stack are 3 kyr and 5 kyr, respectively). Time-frequency analyses utilized the multitaper method (MTM)[47], with a 500 kyr window (5 kyr step) and three 2π prolate tapers; each window was linearly detrended prior to analysis.

Author Contributions

MOP, RM, TN designed the study, conducted sedimentological and time series analyses and wrote the paper. SRM carried performed time-frequency analysis. CE and HB led IODP Expedition 318 and provide seismic reflection data interpretation. FJ and CE analysed XRF geochemical data. LT led development of the age model. MR contributed to writing the manuscript. All authors contributed to the interpretations.

[‡]Integrated Ocean Drilling Program Expedition 318 Scientists

Klaus, A[1], Fehr, A[2], Bendle, J. A. P.[3], Bijl, P. K.[4] , Bohaty., S.M.[5], Carr, S. A.[6], Dunbar, R. B.[7], Flores, J.A.[8], Gonzalez, J.J.[9], Hayden, T. G.[10], Iwai, M.[11], Katsuki, K.[12], Kong, G. S.[13], Nakai, M.[14], Olney, M. P.[15], Passchier, S.[16], Pekar, S. F.[17], Pross, J.[18], Riesselman, C.R.[19], Röhl, U.[20], Sakai, T.[21], Shrivastava, P. K.[22], Stickley, C. E.[23], Sugasaki, S.[24,25], Tuo, S.[26], van de Flierdt, T.,[27] Welsh, K.[28] , Williams, T.,[29] & Yamane, M.[30]

[1]United States Implementing Organization, Integrated Ocean Drilling Program, Texas A&M University, 1000 Discovery Drive, College Station, TX 77845, USA. [2]RWTH Aachen University, Institute for Applied Geophysics and Geothermal Energy, Mathieustrasse 6, D-52074 Aachen, Germany. [3]Geographical and Earth Sciences, University of Glasgow, Gregory Building, Lilybank Gardens, G128QQ Glasgow, UK. [4]Department of Chemistry and Geochemistry, Colorado School of Mines, 1500 Illinois Street, Golden, CO 80401, USA. [5]Ocean and Earth Science, National Oceanography Centre Southampton, University of Southampton, European Way, SO14 3ZH, Southampton, UK. [6]Department of Geological and Environmental

Sciences, Stanford University, 325 Braun Hall, Building 320, Stanford, CA 94305-2115, USA. [7]Department of Environmental Earth System Science, Stanford University, 325 Braun Hall, Building 320, Stanford, CA 94305-2115, USA. [8]Department of Geology, Universidad de Salamanca, 37008, Salamanca, Spain. [9]Instituto Andaluz de Ciencias de la Tierra, CSIC-UGR, 18100 Armilla, Spain.[10]Department of Geology, Western Michigan University, 1187 Rood Hall, 1903West Michigan Avenue, Kalamazoo, MI 49008, USA. [11]Department of Natural Science, Kochi University, 2-5-1 Akebono-cho, Kochi 780-8520, Japan. [12]Geological Research Division, Korea Institute of Geoscience and Mineral Resources, 30 Gajeong-dong, Yuseong-gu, Daejeon 305-350, Republic of Korea. [13]Petroleum and Marine Research Division, Korea Institute of Geoscience and Mineral Resources, 30 Gajeong-dong, Yuseong-gu, Daejeon 305-350, Republic of Korea. [14]Education Department, Daito Bunka University, 1-9-1 Takashima-daira, Itabashi-ku, Tokyo 175-8571, Japan. [15]Department of Geology, University of South Florida, Tampa, 4202 East Fowler Avenue, SCA 528, Tampa, FL 33620, USA. [16]Earth and Environmental Studies, Montclair State University, 252 Mallory Hall, 1 Normal Avenue, Montclair, New Jersey 07043, USA. [17]School of Earth and Environmental Sciences, Queens College, 65-30 Kissena Boulevard, Flushing, NY 11367, USA. [18]Paleoenvironmental Dynamics Group, Institute of Geosciences, Goethe-University Frankfurt, Altenhöferallee 1, 60438 Frankfurt, Germany. [19]Departments of Marine Science and Geology, University of Otago, PO Box 56, Dunedin 9054, New Zealand.[20]MARUM – Center for Marine Environmental Sciences, University of Bremen, Leobener Straße, 28359 Bremen, Germany. [21]Department of Geology, Utsunomiya University, 350 Mine-Machi, Utsunomiya 321-8505, Japan. [22]Antarctica Division, Geological Survey of India, NH5P, NIT, Faridabad 121001, Harlyana, India. [23]Department of Geology, Universitet i Tromsø, N-9037 Tromsø, Norway. [24]Scripps Institution of Oceanography, University of

California, San Diego, La Jolla, California 92093-0220, USA, [25]Department of Earth and Planetary Sciences, University of Tokyo, 7-3-1 Hongo, Bunkyo-ku, Tokyo 113-0033, Japan. [26]School 223 of Ocean and Earth Science, Tongji University, 1239 Spring Road, Shanghai 200092, People's Republic of China. [27]Department of Earth Science and Engineering, Imperial College London, South Kensington Campus, Prince Consort Road, London SW7 2AZ, UK.[28]School of Earth Sciences, University of Queensland, St Lucia, Brisbane QLD 4072, Australia. [29]Lamont Doherty Earth Observatory of Columbia University, PO Box 1000, 61 Route 9W, Palisades, New York 10964, USA. [30]Earth and Planetary Science, University of Tokyo, 7-3-1 Hongo, Bunkyo-ku, Tokyo 113-0033, Japan.

REFERENCES

1. L. E. Lisiecki, M. E. Raymo, A Plio-Pleistocene stack of 57 globally distributed benthic δ^{18}O records. *Paleoceanography* **20**, (2005).
2. K. G. Miller *et al.* High tide of the warm Pliocene: Implications of global sea level for Antarctic deglaciation. *Geology* **40**, 407-410 (2012).
3. M. M. Milanković, *Canon of Insolation and Ice Age problem*. Royal Serbian Academy Special Publication (1941).
4. M. E. Raymo, L. E. Lisiecki, K. H. Nisancioglu, Plio-Plesitocene ice volume, Antarctic Climate, and the global δ^{18}O Record. *Science* **28**, 492-495 (2006).
5. P. Huybers, E. Tziperman, Integrated summer insolation forcing and 40,000-year glacial cycles: the perspective from an ice-sheet/energy-balance model. *Paleoceanography* ***23***, PA1208 (2008).

6. M. E. Raymo, P. Huybers, Unlocking the mysteries of the ice ages. *Nature* **451**, 284-285 (2008).

7. T. Naish *et al.,* Obliquity-paced Pliocene West Antarctic ice sheet oscillations. *Nature* **458**(7236), 322-328 (2009).

8. D. Pollard, R. DeConto, Modeling West Antarctic ice sheet growth and collapse through the past five million years. *Nature* **458**, 329-332 (2009).

9. R. McKay *et al.,* Antarctic and Southern Ocean influences on late Pliocene global cooling. *PNAS* **109**, 6423-6428 (2012).

10. R. G. Lucchi *et al.*, Mid-late Pleistocene glacimarine sedimentary processes of a high-latitude, deep-sea sediment drift (Antarctic Peninsula Pacific margin). *Marine Geology* **189**, 343–370 (2002).

11. Fretwell, P. *et al.*, Bedmap2: improved ice bed, surface and thickness datasets for Antarctica. *The Cryosphere* **7**, 375–393 (2013).

12. L. Tauxe *et al.,* Chronostratigraphic framework for the IODP Expedition 318 cores from the Wilkes Land margin: constraints for paleoceanographic reconstruction. *Paleoceanography* **27** (2012).

13. S. L. Eittreim, A. K. Cooper, J. Wannesson, Seismic stratigraphic evidence of ice-sheet advances on the Wilkes Land margin of Antarctica. *Sedimentary Geology* **96**, 131–156 (1995).

14. R. F. Anderson *et al.,* Wind-driven upwelling in the Southern Ocean and the deglacial rise in atmospheric CO2. *Science* **323**, 1443-1448 (2009).

15. J. R. Toggweiler, J. L. Russell, J. L., S. R. Carson, Midlatitude westerlies, atmospheric CO_2, and climate change during the ice ages. *Paleoceanography* **21,** (2006).

16. C. P. Cook *et al.,* Dynamic behaviour of the East Antarctic Ice Sheet during the Pliocene warmth. *Nature Geoscience* **6**, 765-769 (2013).

17. M. A. Depoorter *et al.,* Calving fluxes and basal melt rates of Antarctic ice shelves. *Nature* 502, 82-92 (2013).

18. Joughin, I., Alley, R. B. & Holland, D. M. Ice-Sheet Response to Oceanic Forcing. *Science* **338**, 1172–1176 (2012).

19. Barrett, P. J. Resolving views on Antarctic Neogene glacial history – the Sirius debate. *Earth and Environmental Science Transactions of the Royal Society of Edinburgh* **104**, 31–53 (2013).

20. A. Mackintosh *et al.,* Retreat of the East Antarctic ice sheet during the last glacial termination. *Nature Geoscience* **4**, 195-202 (2011).

21. Weber, M. E. *et al.,* Millennial-scale variability in Antarctic ice-sheet discharge during the last deglaciation. *Nature* **510**, 134–138 (2014).

22. J. Zachos *et al.,* Climate Response to Orbital Forcing Across the Oligocene-Miocene Boundary, *Science* **292**, 274-276 (2001).

23. H. Pälike *et al.,* The heartbeat of the Oligocene climate system. *Science* **314**, 1894-1898 (2006).

24. H. F. Kleiven, E. Jansen, T. Fronval, T. M. Smith, Intensification of Northern Hemisphere glaciations in the circum Atlantic region (3.5-2.4 Ma) - ice-rafted detritus evidence. *Palaeogeo. Palaeoclim. Palaeoecol.* **184**, 213-223 (2002).

25. G. D. Williams, N. L. Bindoff, S. J. Marsland, S. R. Rintoul, Formation and export of dense shelf water from the Adélie Depression, East Antarctica. *J. Geophys. Res.* **113**, C04039 (2008).

26. A. Martínez-Garcia *et al.,* Southern Ocean dust-climate coupling over the past four million years. *Nature* **476**, 312-315 (2011).

27. D. G. Martinson, S. E. Stammerjohn, R. A. Iannuzzi, R. C. Smith, M. Vernet, Western Antarctic Peninsula physical oceanography and spatio-temporal variability. *Deep Sea Res. II* **55**, 1964–1987 (2008).

28. H. D. Pritchard, *et al.,* Antarctic ice-sheet loss driven by basal melting of ice shelves. *Nature* **484**, 502–505 (2012).

29. R. DeConto, D., Pollard, D. Harwood, Sea ice feedback and Cenozoic evolution of Antarctic climate and ice sheets. *Paleoceanography* **22** (3), (2007).

30. M. Pagani, Z. Liu, J. LaRiviere, A. C. Ravelo, High Earth-system climate sensitivity determined from Pliocene carbon dioxide concentrations, *Nature Geosci* 3, 27–30 (2010).

31. O. Seki *et al.*, Alkenone and boron-based Pliocene pCO2 records. *Earth Planet. Sci. Lett.* **292,** 201-211 (2010).

32. Huybers, P. & Denton, G. Antarctic temperature at orbital timescales controlled by local summer duration. *Nature Geosci* **1**, 787–792 (2008).

33. J. Imbrie *et al.*, On the structure and origin of major glaciation cycles 2. The 100,000-year cycle, *Paleoceanography* 8, 699–735 (1993).

34. J. Jouzel *et al.*, Orbital and Millennial Antarctic Climate Variability Over the Past 800,000 Years, *Science* 317, 793–796 (2007).

35. M. E. Raymo, D. W. Oppo, W. Curry, The Mid-Pleistocene climate transition: A deep sea carbon isotopic perspective, *Paleoceanography* 12, 546–559 (1997).

36. Raymo, M.E., The timing of major climate terminations. *Paleoceanography*, v. 12, p. 577-585 (1997).

37. S. R. Meyers, L. A. Hinnov, Northern Hemisphere glaciation and the evolution of Plio-Pleistocene climate noise, *Paleoceanography* 25 (2010).

38. J. Laskar et al., A long-term numerical solution for the insolation quantities of the Earth, *Astronomy and Astrophysics* 428, 261-285 (2004).

39. J. Brigham-Grette *et al.,* Pliocene Warmth, Polar Amplification, and Stepped Pleistocene Cooling Recorded in NE Arctic Russia. *Science* 340, 1421-1427 (2013).

40. R. M. DeConto et al., Thresholds for Cenozoic bipolar glaciation, *Nature* 455, 652–656 (2008).

41. C. Escutia *et al.,* Circum-Antarctic warming events between 4 and 3.5 Ma recorded in marine sediments from the Prydz Bay (ODP Leg 188) and the Antarctic Peninsula (ODP Leg 178) margins. *Global & Planet. Change* **69**(3), 170-184 (2009).

42. B. D. A. Naafs *et al.,* Strengthening of North American dust sources during the late Pliocene (2.7 Ma). *Earth & Planet. Sci. Lett.* **317**, 8-19 (2012).

43. P. Huybers, Early Pleistocene Glacial Cycles and the Integrated Summer Insolation Forcing, *Science* 313, 508 –511 (2006).

44. IPCC, 2013: *Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change* [Stocker, T.F., et al. (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, 1535 pp.

45. L.A. Krissek, Late Cenozoic ice-rafting records from Leg 145 sites in the north Pacific: Late Miocene onset, Late Pliocene intensification and Pliocene-Pleistocene events. *Proc. Ocean Drill. Program, Sci. Results* **145**, 179-194 (1995).

46. Meyers, S.R. Astro: An R Package for Astrochronology (Version 0.3) (2014).

47. Thomson, D. J. (1982), Spectrum estimation and harmonic analysis, *Proc. IEEE* 70, 1055–1096 (1982).

Acknowledgements: This research used samples and data provided by the Integrated Ocean Drilling Program (IODP). The IODP is sponsored by the US National Science Foundation (NSF) and participating countries under the management of Joint Oceanographic Institutions. Financial support for this study was provided to TN and RM from the Royal Society of New Zealand Marsden Fund Contract VUW0903. Additional support was provided by the Ministry of Science and Innovation (Grant CTM-2011-24079) to CE and FJE, and by the US NSF (Grant OCE-1003603) to SRM and US NSF (Grant OCE-1202632) to MER.

Figure 1. **Location of Site U1361 and bathymetry offshore of the Wilkes Land margin, Antarctica**. Also shown is the location of the Miocene-Plesitocene ANDRILL AND-1B core recovered in the northwestern corner of the Ross Ice Shelf, the southern boundary of the Antarctic Circumpolar Current (ACC), the Mertz Glacier tongue and paleo ice sheet drainage path (white arrows) extending off shore into a slope and rise canyon system. Black lines represent seismic reflection profile tracks represented in Supplementary Figure S1.

Figure 2. **Depth series developed for IODP site U1361 sediment core between 4.4-2.2Ma of (a) XRF-based Ba/Al, (b) IBRD MAR correlated with time series of (c) January insolation and total integrated summer energy (where melt threshold [t]=400GJm^{-2}), (d) mean annual insolation and total integrated summer energy (where melt threshold [t]=250GJm^{-2}), (e) eccentricity[38], and (f) the stacked benthic δ^{18}O record[1].** Also shown is the down core distribution of lithofacies, lithological cycles (transitional lithlogies are represented by both

symbols) and magnetic polarity stratigraphy[12]. Maxima in the productivity proxy Ba/Al coincide with bioturbated/diatom-rich mudstone facies. Grey shaded elipse denotes alignment between a 1.2 Ma node in (d) obliquity modulated mean annual insolation and (e) a 400-kyr minimum in eccentricity which favours polar ice sheet growth and corresponds to (f) a 1‰ glacial δ^{18}O excursion culminating with MIS M2 (arrow). A significant increase in (f) δ^{18}O glacial values from 2.7 Ma (arrow) corresponds with a marked decline in the amplitude of (a) IBRD and a 100ppm decrease in (g) reconstructed atmospheric CO_2 concentration[44].

Figure 3. **2π MTM time-frequency analysis results (500 kyr window) displaying (a) normalized MTM power for the U1361 IBRD-MAR data using the (untuned) age-model[12] compared to (b) normalized MTM power for the LR04 δ^{18}O stack *(1)* and (c) MTM amplitude for the eccentricity-tilt-precession (ETP) solution[38].** Also shown are 2π MTM power spectral estimates for (d, f) the U1361 IBRD-MAR data using the untuned age-model[12], and (e, g) for the δ^{18}O stack[1], with red noise confidence levels using three different approaches. In panels a-c, red indicates large values and blue indicates low values. Milankovitch orbital frequencies are labelled. The results highlight the lack of a strong obliquity signal (40-kyr) in 3500-2500 Ma IBRD-MAR data and the relatively strong influence of ~400-kyr eccentricity compared with the LR04 δ^{18}O stack for the same time interval.

Figure 4. **Quantitative assessment of the evolution of power in the precession band (0.04-0.06 cycles/kyr), obliquity band (0.02-0.035 cycles/kyr), and eccentricity band (0-0.015 cycles/kyr), using the MTM time-frequency results in Figure 3.** The cumlative power in each frequency band is determined for (a) the untuned IBRD-MAR data, (b) the astronomically-tuned

time IBRD-MAR data based on the tielines in Figure 2, and (c) for the LR04 δ^{18}O stack[1]. These results indicate a generally dominant obliquity signal > 3500 kyr, with a reduction in obliquity and enhancement of eccentricity < 3500 kyr in the IBRD-MAR record. The IBRD-MAR eccentricity power is atrributable to the clipping of precession, which transfers preccession power to its ecccentrictiy modulator. Clipped-precession power (eccentricity + precession) increases dramatically in all three time series < 3500 ka. Coincident with the intensification of northern hemisphere continental glaciations, obliquity comes to dominate again at ~2.8 Ma in the LR04 δ^{18}O stack, but not in the IBRD-MAR data. These results imply a local insolation control on EAIS variability < 3500 kyr.