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Title: Geology of the Wilkes Land Sub-basin and Stability of the East Antarctic Ice Sheet: Insights from rock magnetism at IODP Site U1361

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Corresponding Author: Prof. Lisa Tauxe,

Corresponding Author's Institution:

First Author: Lisa Tauxe

Order of Authors: Lisa Tauxe; Saiko Sugisaki; Francisco Jimenez-Espejo; Carlota Escutia; Carys P Cook; Tina van de Flierdt; Masao Iwai

Abstract: IODP Expedition 318 drilled Site U1361 on the continental rise offshore of Adelie Land and the Wilkes sub-glacial basin. The objective was to reconstruct the stability of the East Antarctic Ice Sheet (EAIS) during Neogene warm periods, such as the late Miocene and the early Pliocene. The sedimentary record tells a complex story of compaction, and erosion (thus hiatuses). Teasing out the paleoenvironmental implications is essential for understanding the evolution of the EAIS. Anisotropy of magnetic susceptibility (AMS) is sensitive to differential compaction and other rock magnetic parameters like isothermal remanence and anhysteretic remanence are very sensitive to changes in the terrestrial source region. In general, highly anisotropic layers correspond with laminated clay-rich units, while more isotropic layers are bioturbated and have less clay. Layers enriched in diatoms are associated with the latter, which also have higher Ba/Al ratios consistent with higher productivity. Higher anisotropy layers have lower porosity and moisture contents and have fine grained magnetic mineralogy dominated by maghemite, the more oxidized form of iron oxide, while the lower anisotropy layers have magnetic mineralogies dominated by magnetite. The different magnetic mineralogies support the suggestion based on isotopic signatures by Cook et al. (2013) of different source regions during low productivity (cooler) and high productivity (warmer) times. These two facies were tied to the coastal outcrops of the Lower Paleozoic granitic terranes and the Ferrar Large Igneous Province in the more inland Wilkes Subglacial Basin respectively. Here we present evidence for a third geological unit, one eroded at the boundaries between the high and low clay zone with a "hard" (mostly hematite) dominated magnetic mineralogy. This unit likely outcrops in the Wilkes sub-glacial basin and could be hydrothermally altered Beacon sandstone similar to that detected by Craw and Findlay (1984) in Taylor Valley or the equivalent to the Elatina Formation in the Adelaide Geosyncline in Southern Australia (Schmidt and Williams, 2013). Correlation of the "hard" events with global oxygen isotope stacks of Zachos et al. (2001) and Lisiecki and Raymo (2005) suggest that the source region was eroded during times with higher global ice volume.

UNIVERSITY OF CALIFORNIA, SAN DIEGO

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SANTA BARBARA · SANTA CRUZ

LISA TAUXE
DISTINGUISHED PROFESSOR
TEL: (858) 534-6084
FAX: (858) 534-0784

SCRIPPS INSTITUTION OF OCEANOGRAPHY
LA JOLLA, CA 92093-0220
E-MAIL: ltaxe@ucsd.edu
URL: [HTTP://magician.ucsd.edu/~ltaxe/](http://magician.ucsd.edu/~ltaxe/)

December 7, 2014

To whom it may concern:

Please find enclosed our revised manuscript entitled, "Geology of the Wilkes Land Sub-basin and Stability of the East Antarctic Ice Sheet: Insights from rock magnetism at IODP Site U1361" for consideration for publication in Earth and Planetary Science Letters. We have read the reviews carefully and responded as appropriate. Thank you for your consideration of our revised manuscript.

Sincerely,

A handwritten signature in black ink that reads "Lisa Tauxe". The signature is written in a cursive, flowing style.

Lisa Tauxe

Response to Reviews for “Geology of the Wilkes Land Sub-basin and Stability of the East Antarctic Ice Sheet: Insights from rock magnetism at IODP Site U1361”, by Tauxe et al.

Editor:

Alternatively, you may submit the data to a publicly accessible database or repository such as PANGAEA and indicate in the paper where the data can be accessed. If you decide to use a publicly accessible database, please provide a reference to this database in the results section of the revised manuscript.

I have uploaded the paleomagnetic and rock magnetic data to the publicly accessible MagIC database (<http://earthref.org/MagIC>) where it will be available as soon as this manuscript has been given a DOI (only published data can be made publicly available). The ancillary data (like the NGR, diatom valve counts and isotopic data) have all been previously published, but I put a zip file with all the data and the scripts used to create the plots in the ERDA website (<http://earthref.org/ERDA>), where it will be made publically available as soon as this paper gets a DOI and has been accepted for publication. I will complete the URLs in the paper at the page proof stage.

Reviewer #1:

1) there are two sentences in the abstract in which important findings are garbled: "Higher anisotropy layers have lower porosity and moisture contents and have fine grained magnetic mineralogy dominated by magnetite. Higher anisotropy layers are dominated by maghemite, the more oxidized form of iron oxide supporting the suggestion based on their isotopic signatures by Cook et al. (2013) of different source regions during low productivity (cooler) and high productivity (warmer) times." What is intended, I believe, is "Higher anisotropy layers have lower porosity and moisture contents and have fine grained magnetic mineralogy dominated by maghemite, the more oxidized form of iron oxide, whereas lower anisotropy layers are dominated by magnetite, supporting the suggestion based on their isotopic signatures by Cook et al. (2013) of different source regions during low productivity (cooler) and high productivity (warmer) times."

Thanks for catching that – it is fixed in the manuscript now.

I don't think it's accurate to describe the change in slope (maximum second derivative) between 300 and 425C in Fig 4d as an "inflection point" (zero second derivative).

I changed the wording to “a change in slope”.

Also note the error in labeling: there are two panels in Fig 4 marked as 'd' and

none marked as 'f'.

Oops. This is fixed now.

I'm not enthusiastic about the nomenclature H/S for the ratio of IRMs acquired in 1.0 and 0.5 T: it implies that it represents the ratio of hard and soft remanence contributions, when in fact it represents the ratio of the total IRM (additive over the intervals 0-0.5 and 0.5-1.0 T) to the softer part (acquired in the low interval 0-0.5 T).

I take your point. I struggled with this also. "H" is not an sIRM, because some do not saturate. Instead of H/S, I'm now calling it:

$$\text{IRM}_{0.5T}^{1.0T}$$

Finally, the cyan dots in Fig 7 are not easy to distinguish from the black ones on a normal-size printout; I had to zoom in on the pdf to see the difference.

I changed it to a brown color – I think it is more clear now.

Reviewer #2:

The dynamic processes of the Antarctic Ice Sheet play important role in controlling the global climatic Changes. This study aims to determine the response of the Eastern Antarctic Ice sheet (EAIS) to the paleoclimatic changes during Late Miocene and Early Pliocene by integrating rock magnetic and Geochemical proxies. The overall conclusion is acceptable and meaningful. However, the interpretation of the magnetic assemblage and the correlation between magnetic proxy to the global $\delta^{18}\text{O}$ (low-band filtered) curve need more consideration.

The magnetic type is determined majorly by the unblocking temperature, which has great ambiguities between magnetite and e.g., Al-substituted hematite.

1) Type IV is mostly a high-Al substituted hematite rather than a special high-coercivity magnetite.

I don't really know how you can say that so definitively. There certainly are high coercivity, magnetite blocking temperature rocks in the McMurdo Sound area – in the Pliocene volcanics. I've included a figure with an example as another figure.

2) In type II, we need more experiments to discriminate Titanomagnetite and maghemitized magnetite. T_b is not necessarily related to T_c .

Yes that is true that they are different, but T_b can never be higher than T_c and what we have in Type I is perfectly ordinary magnetite peak blocking

temperatures and in Type II, the peak blocking temperature is HIGHER than that of magnetite, so the T_c cannot be magnetite. It is consistent with maghemite (625C).

A high-temperature cycle between the room-T and about 400 degree will be useful. If the k-T curve is reversible between this temperature interval, it will indicate a T_c for e.g., titanomagnetite. A large loss in magnetization will indicate the conversion of maghemite to hematite.

Maghemite can be stable above 400C up to its Curie T of around 625C. This test will not be diagnostic.

3) The low-band filter of the d18O curve is not clear. Usually, there are 100, and 400 kyr cycles for the paleoclimatic changes. I guess that the 400 kyr cycle could control the NGR curve. Without more accurate correlation, it is not easy to judge the correlation between the magnetic and NGR proxy and the d18O curve. Overall, this is a high-quality manuscript and reveals the dynamic process of the detrital input of magnetic minerals from different provenances at different paleoclimatic conditions. However, I suggest that Authors need to add in more thermal analysis to accurately determine the magnetic assemblage and more quantitative analysis of the orbital-tuning process.

A detailed spectral analysis of sedimentological and geochemical data has been performed at the same site (U1361) and recently published (see Patterson et al., 2014 Nat. Geosci.). As Reviewer #2 suspected, that study identified ~100 kyr and 400 kyr power that is consistent with a non-linear clipped climate response (Imbrie et al., 1993), especially after 3.4 Ma. The Patterson result was in agreement with previous studies of Miocene $\delta^{18}O$ glacial excursions that have proposed a relationship between intervals of increased glacial amplitude in the $\delta^{18}O$ record with a coincidence of 1.2 Ma nodes in obliquity and 400-kyr minima in long period eccentricity (Zachos et al., 2001; Pälike et al., 2006). Given the already published work on that subject, a detailed discussion of the cycles and precise correlation with the global stacks of oxygen isotopes from Zachos et al. (2001) and Lisiecki and Raymo (2005) is not the main aim of this study. Here, the filtered version of the Zachos data is meant to show periods of higher and lower ice volume and suggests the magnetically 'hard' samples appear when there is periods of higher ice volume. As the Patterson paper has now appeared in print, I have added a reference to it for their "quantitative analysis of the orbital tuning process".

Geology of the Wilkes Land Sub-basin and Stability of the East Antarctic Ice Sheet: Insights from rock magnetism at IODP Site U1361

L. Tauxe^{a,b}, S. Sugisaki^{a,c}, F. Jiménez-Espejo^d, C. Escutia^e, C.P. Cook^f, T. van de Flierdt^g, M. Iwai^h

^a*Scripps Institution of Oceanography, La Jolla, CA, USA.*

^b*ltauxe@ucsd.edu*

^c*Department of Earth and Planetary Science University of Tokyo 7-3-1 Hongo Bunkyo-ku Tokyo 113-0033 Japan*

^d*Department of Biogeochemistry, Japan Agency for Marine-Earth Science and Technology (JAMSTEC) 2-15 Natsushima, Yokosuka, 237-0061, Japan*

^e*Instituto Andaluz de Ciencias de la Tierra CSIC-Univ Granada Av. de las Palmeras, 4 18100 Armilla (Granada) Spain*

^f*Department of Geological Sciences, University of Florida, Gainesville, FL, USA.*

^g*Department of Earth Science and Engineering Imperial College London SW7 2AZ London United Kingdom*

^h*Department of Natural Environmental Science Kochi University 2-5-1 Akebono-cho Kochi 780-8520 Japan*

Abstract

IODP Expedition 318 drilled Site U1361 on the continental rise offshore of Adélie Land and the Wilkes sub-glacial basin. The objective was to reconstruct the stability of the East Antarctic Ice Sheet (EAIS) during Neogene warm periods, such as the late Miocene and the early Pliocene. The sedimentary record tells a complex story of compaction, and erosion (thus hiatuses). Teasing out the paleoenvironmental implications is essential for understanding the evolution of the EAIS. Anisotropy of magnetic susceptibility (AMS) is sensitive to differential compaction and other rock magnetic parameters like isothermal remanence and anhysteretic remanence are very sensitive to

29 changes in the terrestrial source region. In general, highly anisotropic layers
30 correspond with laminated clay-rich units, while more isotropic layers are
31 bioturbated and have less clay. Layers enriched in diatoms are associated
32 with the latter, which also have higher Ba/Al ratios consistent with higher
33 productivity. Higher anisotropy layers have lower porosity and moisture con-
34 tents and have fine grained magnetic mineralogy dominated by maghemite,
35 the more oxidized form of iron oxide, while the lower anisotropy layers have
36 magnetic mineralogies dominated by magnetite. The different magnetic min-
37 eralogies support the suggestion based on isotopic signatures by Cook et al.
38 (2013) of different source regions during low productivity (cooler) and high
39 productivity (warmer) times. These two facies were tied to the coastal out-
40 crops of the Lower Paleozoic granitic terranes and the Ferrar Large Igneous
41 Province in the more inland Wilkes Subglacial Basin respectively. Here we
42 present evidence for a third geological unit, one eroded at the boundaries be-
43 tween the high and low clay zone with a “hard” (mostly hematite) dominated
44 magnetic mineralogy. This unit likely outcrops in the Wilkes sub-glacial basin
45 and could be hydrothermally altered Beacon sandstone similar to that de-
46 tected by Craw and Findlay (1984) in Taylor Valley or the equivalent to
47 the Elatina Formation in the Adelaide Geosyncline in Southern Australia
48 (Schmidt and Williams, 2013). Correlation of the “hard” events with global
49 oxygen isotope stacks of Zachos et al. (2001) and Lisiecki and Raymo (2005)
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51 ice volume.

52 *Keywords:* East Antarctic Ice Sheet Stability, rock magnetism, Pliocene
53 paleoclimate, Integrated Ocean Drilling Program, Joides Resolution,

55 **1. Introduction**

56 As atmospheric CO₂ levels exceed the 400 ppm mark, our eyes naturally
57 turn to the last time they were that high, the Pliocene, when they were esti-
58 mated to be between 365 and 415 ppm (Pagani et al., 2010). Implications for
59 temperature and sea level rise can perhaps be gleaned by such comparisons.
60 Global sea level is controlled by both temperature and the amount of ice
61 stored on continents. As global climate changes, understanding the response
62 of global ice volume to rising temperatures is urgently needed. For exam-
63 ple, melting of the Greenland Ice sheet would result in a 7 m rise, and full
64 deglaciation of the Western Antarctic Ice Sheet (WAIS) and East Antarc-
65 tic Ice Sheet (EAIS) would contribute another ~5 m and 52 m respectively
66 (Lythe et al., 2001) (see also Fretwell et al. (2013)). Whereas it appears likely
67 that the Greenland Ice Sheet is quite vulnerable to warming climate (Gre-
68 gory et al., 2004), at least some studies (Huybrechts and de Wolde, 1999)
69 predict that the EAIS will grow owing to increased precipitation (see also
70 Alley et al. (2005)). Worryingly, recent reports indicate that while parts of
71 the EAIS are growing, other parts are decreasing (Fig. 1 and Vaughan et al.
72 (2013)). The past response of the EAIS in times with comparable tectonic
73 configurations and atmospheric CO₂ levels to the present day, i.e. in the
74 Pliocene, therefore are needed to inform current discussions of rising sea-
75 levels. While Miller et al. (2012) summarized evidence for global sea-level
76 that was 22 ± 10 m higher than present during the Pliocene, concluding that
77 it was “very likely that several meters of eustatic rise can be attributed to

78 ice loss from the marine margins of East Antarctica.”, the large error bars
79 leave room for considerable doubt.

80 An attractive target for investigating EAIS stability is the eastern sector
81 of the Wilkes Land margin, located at the seaward termination of the largest
82 East Antarctic subglacial basin, the Wilkes subglacial basin on Wilkes Land,
83 Antarctica (Fig. 1). Such an investigation was one of the rationales for the
84 drilling of Site U1361 (64.2457°S, 143.5320°E) during Expedition 318 of the
85 International Ocean Drilling Program (Escutia et al., 2011). In an initial
86 study, Cook et al. (2013) reported strontium and neodymium isotopic ratios
87 from detrital material indicating erosion of two distinctly different source
88 bodies, now mostly under the ice sheet. They interpreted the data as evidence
89 for retreat of the ice sheet margin several hundreds of kilometers inland
90 during the warmer intervals of the Pliocene. Here we report complimentary
91 rock magnetic data which provide additional evidence for their conclusions
92 and point to erosion of a third geological unit accessed during glacial advance
93 and retreat.

94 **2. Material and Methods**

95 Site U1361 was drilled into a submarine levee off the coast of Adélie
96 Land, just to the west of the Wilkes subglacial basin. Hole U1361A was
97 cored using the advanced piston coring system to refusal at 151.5 mbsf be-
98 low which an extended core barrel was used to a depth of 388 mbsf. Tauxe
99 et al. (2012) compiled magneto-, bio-, and lithostratigraphic information for
100 the Expedition 318 cores documenting a sedimentary record from the Mid-
101 dle Miocene to the late Pleistocene with few hiatuses and excellent mag-

102 netostratigraphic control. The complete data set of previously published
103 paleomagnetic and rock magnetic data is available in the MagIC database
104 at: <http://earthref.org/doi/10.1029/2012PA002308>

105 The interval studied here is 40-160 mbsf and spans ~ 2.2 Ma to 6.4 Ma
106 (Fig. 2). Two major lithofacies are represented: laminated clay-rich units
107 and bioturbated units with less clay and more abundant diatoms.

108 Paleomagnetic samples were taken every core section (~ 1.5 m inter-
109 vals) for a total of 80 discrete samples. Anisotropy of magnetic susceptibility
110 (AMS) including bulk susceptibility (χ) was measured on all discrete samples
111 on the Kappabridge KLY4S magnetic susceptibility instrument either on the
112 ship or in the Scripps Paleomagnetic Laboratory. These data, represented as
113 maximum, intermediate and minimum eigenvalues (τ_1, τ_2, τ_3) were reported
114 by Tauxe et al. (2012). Shipboard measurements of moisture content, poros-
115 ity, and natural gamma ray (NGR) were reported by Escutia et al. (2011).
116 As part of the post-cruise geochemical investigations, Cook et al. (2013)
117 measured x-ray fluorescence (XRF), diatom valve concentrations (DVC) and
118 strontium and neodymium isotopes. The XRF data from Cook et al. (2013)
119 are shown as black lines while those reported here are in blue. Here we
120 use the barium/aluminum (Ba/Al) ratio and shipboard NGR (Escutia et al.,
121 2011) as a proxies for primary productivity (Dymond et al., 1992) and the
122 clay fraction (Dunlea et al., 2013) respectively in the cores.

123 For the present study, we measured anhysteretic remanence (ARM) ac-
124 quired in an alternating field of 180 mT in the presence of a 50 μ T DC bias
125 field using an SI-4 alternating field demagnetizer and measured using the 2-G
126 Enterprises magnetometer in the Scripps Paleomagnetic Laboratory. Follow-

127 ing the ARM step, IRMs were imparted to the discrete samples using an
128 ASC impulse demagnetizer in fields increasing up to ~ 1.2 T. We refer here
129 to the ratio of the IRM at 1 T and the 0.5 T IRM steps as the $\text{IRM}_{0.5T}^{1.0T}$ ratio.
130 Following IRM acquisition experiments, small chips were taken from repre-
131 sentative cubes and glued into clean glass vials with KaSil cement. These
132 were exposed again to a field of ~ 1 T along the (new) X direction. A second
133 IRM was imparted in a 0.5 T field along the Y axis and a third IRM in a
134 0.1 T field along the Z axis. The specimens were then thermally demagne-
135 tized in a step-wise fashion to determine the blocking temperatures of the
136 different coercivity fractions in the specimens in an experiment known as the
137 ‘3D-IRM demagnetization experiment’ of Lowrie (1990). All rock and pa-
138 leomagnetic data analyzed for the present study are available for download
139 from the MagIC data base at:

140 <http://earthref.org/doi/10.1029/2012PA002308>

141 Geochemical and physical property data used in the interpretations here are
142 available for download from the ERDA along with the Python scripts used
143 to generate the figures.

144 3. Results

145 3.1. Anisotropy of magnetic susceptibility

146 We re-plot the eigenvalues of the AMS tensors for the interval 40-160
147 mbsf from Hole U1361A as presented by Tauxe et al. (2012) in Fig. 2a. One
148 measure of the degree of anisotropy is the ratio of τ_1/τ_3 , usually termed P
149 (Chapter 13 in Tauxe et al. (2010)), is shown in Fig. 2b. Stratigraphic in-
150 tervals with high values of P (here taken as > 1.03) are shaded in grey. We

151 also plot the (uncalibrated) aluminum and barium data from the XRF mea-
152 surements of Cook et al. (2013) in Figs. 2c and d (black lines) and report
153 new data for the interval below 104 mbsf here (shown in blue). These data
154 suggest that the zones of high anisotropy correspond with zones of high alu-
155 minum which in turn is directly related to the clay content. This contention
156 is supported by the variations in natural gamma radiation (NGR), shown in
157 Fig. 2e, whereby zones of relatively high NGR ($>\sim 34$) with few exceptions
158 correspond to high P and also high aluminum (clay). We assume in the
159 following that the high P and high NGR intervals ($>\sim 34$) can be used as
160 proxies for clay-rich zones in this study.

161 Schwehr et al. (2006) investigated the role of porosity and water con-
162 tent (related to compaction) in controlling the anisotropy fabric and found
163 a strong correlation. They showed that changes in anisotropy degree can
164 result from compaction disequilibria resulting from changes in lithology, for
165 example from alternating between clay-rich and clay-poor layers, or from hia-
166 tuses. Here, we plot moisture content and porosity in Fig. 2f. The grey (high
167 anisotropy) zones do appear to be associated with zones of low moisture and
168 porosity, further supporting the connection between anisotropy, clay content
169 and laminated versus bioturbated layers.

170 The ratio of barium to aluminum is a proxy for primary productivity (see
171 also Fig. 2) is shown in Fig. 3b. Zones of high Ba/Al ratios are closely as-
172 sociated with the low anisotropy zones (white). Using Ba/Al as a proxy for
173 productivity, we infer that the clay rich intervals are associated with lower
174 productivity. To further explore this possibility, we plot diatom valve counts
175 of Cook et al. (2013) in Fig. 3c. We see that the two zones of high diatom

176 valve counts are also associated with high Ba/Al and low AMS anisotropy.
 177 The expanded section from 88 to 104 mbsf in Fig. 3e shows shipboard high
 178 resolution core photos with light and dark bands associated with the low and
 179 high anisotropy values respectively. Note that the contrast on the photos was
 180 increased to highlight the patterns. As there are only two zones with signifi-
 181 cant diatom valve counts, the Ba/Al where measured is a superior lithological
 182 tool for identifying zones of high productivity. While Ba/Al was not mea-
 183 sured for the entire core, P and NGR were, hence these represents important
 184 proxies for primary productivity in this interval of Hole U1361A. We note
 185 here that because the Ba/Al ratio plot is quite similar to the aluminum plot
 186 shown in Fig. 2, that it is possible that the ratio is dominated by variabil-
 187 ity in aluminum rather than barium and merely reflects the clay content as
 188 opposed to productivity.

189 3.2. Isothermal remanence

190 Representative examples of the behavior of the U1361A sediments during
 191 the IRM acquisition and 3D-IRM demagnetization experiments are shown in
 192 Fig. 4. Figs. 4a and b show the behavior of a specimen whose IRM saturated
 193 by about 0.3 T impulse ($\text{IRM}_{0.5T}^{1.0T}$ of ~ 1); it is also virtually completely de-
 194 magnetized by between 575 and 600°C (Fig. 4b) consistent with a magnetite
 195 remanence. All such ‘Type I’ specimens ($N = 21$) belong to the “clay-poor”,
 196 bioturbated, low anisotropy ($P < 1.03$) and high productivity, lithofacies
 197 (e.g., Fig. 5a). Specimens like the one shown in Figs. 4c and d also saturate
 198 in low impulse fields but display an change in slope in the medium and low
 199 coercivity fraction (Y and Z axes) between 300 and 425°C. Frequently, these
 200 did not completely demagnetize until about 625°C (Fig. 4d), suggestive of

201 maghemite. All of these ‘Type II’ specimens ($N = 20$) had P values greater
 202 than 1.03 and belong to the “clay-rich” facies associated with the laminated,
 203 low productivity intervals (e.g., Fig. 5b). The specimen shown in Figs. 4d
 204 and e does not saturate even in an impulse field of over 1 T ($\text{IRM}_{0.5T}^{1.0T} = 1.11$)
 205 and has 57% of the (total) remanence remaining after demagnetization to
 206 600°C. This ‘Type III’ behavior is characteristic of hematite. Eleven of the
 207 13 specimens displaying this behavior were found at the boundaries between
 208 the high and low P zones (e.g., Fig. 5c). The two exceptions were found
 209 bordering zones with high NGR values. One of these is also adjacent to
 210 a sampling gap and hiatus inferred from the magnetostratigraphic pattern.
 211 The fourth type of behavior is that shown in Figs. 4g and h whereby the IRM
 212 acquisition curve is quite ‘hard’ with $\text{IRM}_{0.5T}^{1.0T}$ values of ~ 1.07 , but the speci-
 213 mens are virtually completely demagnetized by between 575 and 600°C. This
 214 ‘Type IV’ behavior is characteristic of multi-axial single domain magnetite
 215 (Tauxe et al., 2002). All of these specimens ($N = 3$) were also found at the
 216 boundaries between the high and low P zones (e.g., Fig. 5d). The average
 217 P value of the Types III and IV specimens was 1.03 ± 0.02 or transitional
 218 between the ‘low’ and ‘high’ anisotropy zones. Therefore, for the purposes of
 219 this study, we classify specimens as being ‘soft’ if their $\text{IRM}_{0.5T}^{1.0T}$ values were
 220 less than 1.03.

221 4. Discussion

222 ARM, IRM and χ are sensitive to magnetic grain size (Maher and Thomp-
 223 son, 1999) and are frequently plotted against one another to detect changes
 224 in grain size or changes in provenance (Banerjee et al., 1981). In Fig. 6a we

225 plot the mass normalized ARM against bulk susceptibility (χ) in and ARM
 226 against IRM acquired in a 1.2 T field in Fig. 6b. All of the magnetically
 227 hard specimens ($\text{IRM}_{0.5T}^{1.0T} > 1.03$, plotted as red dots) cluster near the origin.
 228 The remaining, magnetically soft, specimens can be divided into two groups:
 229 those that are characterized by low P (Type I, or magnetite remanences)
 230 and those with high P (Type II, or maghemite remanences). The two types
 231 have distinct slopes consistent with their different mineralogies and point to
 232 different sources of the magnetic minerals. It is not surprising that the high
 233 P specimens, belonging to the laminated clay facies, appear to have finer
 234 magnetic grain sizes, based on the steeper slope of the ARM versus χ trend
 235 lines. We note however that the interpretation as to grain size of such data
 236 (e.g., King et al. (1983)) is based solely on magnetite and should be used
 237 with caution in this case as there is evidence of significant maghemitization
 238 of the high P specimens. Nonetheless, the data demonstrate that the high
 239 and low clay facies have markedly different magnetic mineralogies and ap-
 240 parently also magnetic grain sizes. Therefore, the two ‘soft’ types must have
 241 different sedimentological histories.

242 The rock magnetic results point to three distinct populations of mag-
 243 netic mineralogies. Fig. 7 shows the clay proxy NGR and $\text{IRM}_{0.5T}^{1.0T}$ plotted
 244 against stratigraphic depth. With few exceptions, the magnetically soft mag-
 245 netite specimens (black dots) in the $\text{IRM}_{0.5T}^{1.0T}$ profile, are associated with the
 246 clay-poor facies (low P , indicated by white zones), while the magnetically
 247 soft maghemite specimens (brown dots), are associated with the clay-rich
 248 (high P , indicated by grey zones) specimens. The horizontal lines mark the
 249 positions of the magnetically hard specimens (red dots), which with only

two exceptions (indicated as dotted black lines) are found at the transitions between high and low clay in the NGR clay proxy data.

The origin of the two ‘soft’ groups can be understood in the light of the ϵNd (the deviation of measured $^{143}\text{Nd}/^{144}\text{Nd}$ ratios from the Chondritic Uniform Reservoir in parts per 10,000) and $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic data of Cook et al. (2013), shown in Fig. 7d. The intervals of low ϵNd (less than the dashed red line plotted at $\epsilon\text{Nd} < -9.25$) and high $^{87}\text{Sr}/^{86}\text{Sr}$ are all associated with clay-rich intervals (NGR >34). These were deposited (and presumably eroded) during the low-productivity (cooler?) intervals. In contrast, the clay-poor intervals with NGR <34 are associated with high ϵNd and low $^{87}\text{Sr}/^{86}\text{Sr}$ values, deposited during the higher productivity (warmer?) intervals during the Pliocene. Cook et al. (2013) inferred different source regions based on the isotopic signatures of the detrital material and tied these two groups to Lower Paleozoic terranes and the Ferrar Large Igneous Province (FLIP) respectively (see Fig. 8). Based on the occurrences of these two rock types on Wilkes Land, they argued that the cooler intervals have isotopic signatures compatible with granitic bedrock in the hinterland of the nearby Ninnis Glacier (NG on Fig.8). The warmer intervals have isotopic signatures like those of the FLIP rocks whose magnetic anomaly signature was detected in the Wilkes subglacial basin by Ferraccioli et al. (2009).

As already mentioned, with two exceptions, the high $\text{IRM}_{0.5T}^{1.0T}$ intervals (Types III and IV) are observed at the transitions between high and low NGR. All but three of these specimens are identifiable as hematite dominated. Hematite is unusual in such a grey-black sediments and is likely of detrital origin. Although glacial to interglacial transitions could be associated with

275 changes in ventilation that can promote diagenetic enrichments involving
276 secondary hematite pigmentation, reddish tinted horizons are not associated
277 with the horizons actually sampled (e.g., Fig. 5c). Therefore, we suspect a
278 third, as yet unidentified, terrane that eroded during glacial advance and
279 retreat. The closest red bed units in outcrop (which could provide hematite
280 rich sedimentary particles) appear to be Permian red beds of the Amery
281 Group, exposed in the Prince Charles Mountains (see Mikhalsky et al. (2001),
282 Keating and Sakai (1991) and references therein); these are quite distant
283 from U1361 and are unlikely to be a source for the detrital hematite found
284 at Site U1361. However, Veevers and Saeed (2011) found references to red
285 sandstones in Mawson (1915) (v. 2, p 294). According to that delightful
286 account,

287 “Stillwell met with a great range of minerals and rocks in the
288 terminal moraine near Winter Quarters, Adelie Land. Amongst
289 them was red sandstone in abundance, suggesting that the Beacon
290 sandstone formation extend also throughout Adelie Land, but is
291 hidden by the ice-cap.”

292 Moreover, Craw and Findlay (1984) found hydrothermally altered granitoids
293 and Beacon sandstone with enrichment of hematite, altered by the intrusion
294 of the Ferrar sills near Taylor Glacier. These units are also likely to occur
295 in the Wilkes sub-glacial basin along with the FLIP units detected by Fer-
296 raccioli et al. (2009). Another likely source, however, is the Neoproterozoic
297 Elatina Formation, exposed on the southern Australian margin in the Ade-
298 laide geosyncline Williams et al. (2008). This formation has recently been
299 studied by Schmidt and Williams (2013) who found a remanence dominated

300 by hematite. Aitken et al. (2014) reconstructed the geological connections
 301 between Australia and Antarctica largely based on magnetic and gravity
 302 anomalies. We show their reconstruction for the time prior to the break up
 303 of Gondwana at 160 Ma, along with the Adelaide basin (from Schmidt and
 304 Williams (2013) and Williams et al. (2008)) and the location of Site U1361
 305 in Fig. 8 in present coordinates with respect to Antarctica. The Adelaide
 306 basin along with its hematite rich red beds are thought to correlate to units
 307 now covered by ice and have therefore not been identified in outcrop on the
 308 Antarctic margin. Notably, Finn et al. (2006) explained lows in the mag-
 309 netic anomalies east of the Mawson block (MB in Fig. 8) as ‘magnetite-poor
 310 upper Neoproterozoic and lower Paleozoic sedimentary rocks and their meta-
 311 morphic equivalents’. It seems likely that there is a small outcrop of either
 312 Beacon red sandstone (as suspected by Mawson) or Elatina equivalent Neo-
 313 proterozoic red beds on the Antarctic margin (Finn et al., 2006), that eroded
 314 during growth and decay of the East Antarctic Ice Sheet.

315 The Type IV behavior suggests a magnetically hard, magnetite remanence
 316 and the source of this phase is more elusive. Nonetheless, magnetically hard
 317 magnetite was frequently observed in the McMurdo volcanic province (see
 318 data of Lawrence et al. (2009) shown in Fig. 9) and McMurdo volcanics
 319 hidden under the ice sheet is also a possible source for such specimens. These
 320 would be difficult to distinguish from the FLIP units in aeromagnetic surveys.

321 The excellent magnetostratigraphic control for Site U1361 allows us to
 322 tie the hard $\text{IRM}_{0.5T}^{1.0T}$ layers to the GPTS with a high degree of confidence
 323 (dashed green lines in Fig. 7). Patterson et al. (2014) performed a detailed
 324 and quantitative correlation of the interval of U1361A between 50 and 100

325 mbsf and we will not duplicate that effort here. Nonetheless, the oxygen
 326 isotopic stacks of Zachos et al. (2001) and Lisiecki and Raymo (2005) (black
 327 and red curves in Fig. 7e respectively with the cyan curve representing a
 328 low pass filtered version of the Zachos et al. (2001) data) are particularly
 329 intriguing here. While it is tempting but perhaps dangerous to tie a particular
 330 high $IRM_{0.5T}^{1.0T}$ event to a particular isotopic event (say M2), it does appear
 331 likely that unusually cold intervals (high global $\delta^{18}O$ data in the filtered
 332 record between ~ 5.6 - 6.2 Ma, 4.6 - 5 Ma and <3.6 Ma), result in erosion of an
 333 elusive hematite bearing lithofacies now hidden under the ice on Antarctic
 334 continent.

335 5. Conclusions

- 336 • Anisotropy of magnetic susceptibility is a sensitive indicator of the
 337 clay fraction at Hole U1361A. The clay-rich, high anisotropy, zones
 338 are apparently lower productivity and are likely associated with colder
 339 intervals while the clay-poor, low anisotropy, zones are associated with
 340 warmer intervals during the Pliocene.
- 341 • There are four distinct categories of behavior during IRM acquisition
 342 and thermal demagnetization. The first two (Types I and II) are mag-
 343 netically 'soft' and are likely to be magnetite and its more oxidized
 344 cousin, maghemite. The second two (Types III and IV) are magneti-
 345 cally 'hard' and are likely to be hematite and a rare, magnetically hard,
 346 form of magnetite.
- 347 • The maghemite (Type II) mineralogies are associated with the clay-rich

facies while the magnetite (Type I) remanences are associated with the clay-poor facies. In turn, these are associated with zones inferred to be lower and higher productivity, based on the Ba/Al ratios in the sediments hence belong to the colder and warmer intervals of the Pliocene respectively. These are also tied to the Paleozoic and Ferrar Large Igneous Province sources respectively according to the ϵNd and strontium isotopic results of Cook et al. (2013).

- The magnetically hard Types III and IV remanences (hematite and a few rare magnetite specimens respectively) are associated with the transition zones between clay poor and clay-rich facies. These were sourced in an unknown lithologic unit that eroded during glacial advance and retreat. It appears likely that the source for the hematite rich layers is either hydrothermally altered Beacon sandstone units or Neoproterozoic red beds, correlative to units in the Adelaide Basin studied by Schmidt and Williams (2013). The ‘hard’ magnetite layers could have a source in McMurdo volcanics. Both of these are likely hidden under the ice.
- The occurrence of the hard $\text{IRM}_{0.5T}^{1.0T}$ layers in periods with more global ice volume is consistent with the contention that they are sourced in a geological unit that gets eroded during ice sheet advance and retreat of glaciers associated with particularly cold intervals.
- This study, in combination with the work of Cook et al. (2013), strongly supports an active response of the East Antarctic Ice Sheet to climatic forcing in the Pliocene. As CO_2 levels approach those last seen in the

372 Pliocene, we can expect a greater role of EAIS melting than is presently
373 envisioned.

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Figure 1: Map showing location of IODP Expedition 318 Site U1361 drilled at 64.2457°S, 143.5320°E, 3466 mbrf. Inset (adapted from the Vaughan et al. (2013)) shows the ice loss determined between 2006 and 2012 from GRACE time-variable gravity data in cm water/year from the East and West Antarctic Ice Sheets (EAIS and WAIS respectively).

Figure 2: a) Eigenvalues (red squares: τ_1 , blue triangles: τ_2 , black circles: τ_3) b) Anisotropy degree, ($P = \tau_1/\tau_3$). c) Aluminum (expressed as arbitrary units from XRF data. d) Barium (arbitrary units), e) Natural gamma radiation (NGR, Escutia et al. (2011)). f) Shipboard moisture content and porosity from (Escutia et al., 2011). g) Inclinations from Tauxe et al. (2012). Small blue dots are data from the archive halves demagnetized to 20 mT. Red (cyan) triangles are acceptable best-fit lines and Fisher means respectively, according to the criteria defined by Tauxe et al. (2012). h) Polarity log. Black intervals are normal (negative inclinations) , white reverse (positive inclinations) and grey are intervals with no data. i) Geomagnetic Polarity Time Scale (GPTS), black (white) intervals are normal (reverse) polarity. Chrons are calibrated by as in the Geological Time Scale of Gradstein et al. (2004), GTS04 for consistency with other work on Expedition 318 material. Intervals with high P ($P > 1.03$) are marked with grey bars. Black lines in barium and aluminum data are from Cook et al. (2013) and blue lines are reported here.

Figure 3: a) P (from Fig. 2). Grey bars as in Fig. 2. d) Ba/Al (data from Fig. 2.) c) Diatom valve concentration (valves/g; black). Valve counts (from Cook et al. (2013)) are divided by 10^7 . d) Expanded section showing core photo and associated P values.

Figure 4: a), c), e), g) Representative isothermal remanent magnetization (IRM) acquisition curves. b), d), f), h) 3-D IRM demagnetization experiments.

Figure 5: Photos of the core sections in which the samples in Figure 4 were taken. Locations of sample horizons indicated by red stars. Exposure of the photos enhanced to 100%, but colors were otherwise not manipulated. Horizontal scale 2x vertical. Section lengths are 74cm in a) and 150cm in b-d.

Figure 6: a) Plot of mass normalized ARM versus bulk susceptibility (χ) for the three types of specimens. Best-fit lines from a linear regression for the high P versus low P groups of low $\text{IRM}_{0.5T}^{1.0T}$ specimens are also shown, suggesting that the high P (clay-rich) sediments have smaller magnetic grain sizes as well as bulk sediment grain sizes. b) Plot of mass normalized ARM versus IRM distinguished by different degrees of anisotropy (P) and magnetic ‘hardness’ ($\text{IRM}_{0.5T}^{1.0T}$). Best-fit line and second order polynomial are shown for the low and high P groups respectively.

Figure 7: Summary of data for U1361A. a) Magnetic polarity zonation (see Fig.1), b) NGR (Fig. 2), c) $IRM_{0.5T}^{1.0T}$ where black dots are ‘magnetite’, brown are ‘maghemite’ and red are ‘hard’, d) ϵNd and $^{87}Sr/^{86}Sr$ (Cook et al., 2013), e) Global stacks of oxygen isotopes of Zachos et al. (2001) (black) and Lisiecki and Raymo (2005) (red). Heavy cyan curve is a low pass filter of the Zachos et al. (2001) curve. f) The GPTS (Gradstein et al., 2004). The grey intervals represent high anisotropy layers from Fig. 1. Horizontal lines are the positions of high $IRM_{0.5T}^{1.0T}$ samples. Blue (red) solid lines are low clay to high clay (high clay to low clay) transitions. Black dashed lines are neither. Green dashed lines are correlations of high $IRM_{0.5T}^{1.0T}$ layers to the oxygen isotopic record.

Figure 8: Magnetic anomaly map (for Australia and Wilkes Land region of Antarctica of Aitken et al. (2014)). The continents are in the ‘Leeuwen’ Gondwana reconstruction for 160 Ma. The location of Site U1361 is the current location with respect to Antarctica. WSB is the Wilkes Subglacial Basin. Geological piercing points drawn as double arrows are suggested geological correlations. GA is the Gawler-Terre Adélie connection. Geological units of Ferrar Large Igneous Province (FLIP) (black) and Lower Paleozoic age (white) were inferred from isotopic analyses of Cook et al. (2013) as sources of detritus in the clay-rich and clay-poor zones. The closest likely source of the Lower Paleozoic age material is just south of the Ninnis Glacier (NG). The closest likely source of FLIP material is in the WSB where it is inferred to exist based on aeromagnetic anomalies of Damaske et al. (2003) and Ferraccioli et al. (2009). It is possible that there are outcrops of hydrothermally altered Beacon sandstones (with hematite enrichment) in association with the FLIP material, as seen in near the Taylor Glacier Craw and Findlay (1984). Alternatively, the Adelaide geosyncline of South Australia from Schmidt and Williams (2013) contains outcrops of the red beds of the Elatina Formation. If present on the Antarctic continent, these units could be the source of the hematite present at many clay-rich/clay-poor transitions.

Figure 9: Typical example of high coercivity behavior coupled with blocking temperatures (maximum of $\sim 580^{\circ}\text{C}$) typical of magnetite for a specimen from the McMurdo Sound volcanics. a) Alternating field demagnetization. b) Thermal demagnetization. [Data from Lawrence et al. (2009) available for download at: <http://earthref.org/MAGIC/9411/>.]

- Climate models predict growth of East Antarctic Ice Sheet with global warming
- Records from IODP Exp. 318 show instability of the EAIS during the Pliocene.
- Rock magnetic techniques detect erosion of three distinct geological units.

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