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# 1 Gradual and abrupt changes during the Mid-Pleistocene Transition

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4 Heather L. Ford<sup>1</sup>, Sindia Sosdian<sup>2</sup>, Yair Rosenthal<sup>3</sup>, Maureen E. Raymo<sup>1</sup>

5  
6 <sup>1</sup>Biology and Paleo Environment, Lamont-Doherty Earth Observatory, Palisades, New York,  
7 USA

8 <sup>2</sup>Earth and Ocean Sciences, Cardiff University, Cardiff, UK

9 <sup>3</sup>Department of Marine and Coastal Sciences and Department of Earth and Atmospheric  
10 Sciences, Rutgers University, New Brunswick, New Jersey, USA

11  
12  
13 \*Corresponding author: [hford@ldeo.columbia.edu](mailto:hford@ldeo.columbia.edu)  
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## 15 Abstract

16 During the Mid-Pleistocene Transition (MPT), the dominant glacial-interglacial cyclicality  
17 as inferred from the marine  $\delta^{18}\text{O}$  records of benthic foraminifera ( $\delta^{18}\text{O}_{\text{benthic}}$ ) changed from 41  
18 kyr to 100 kyr years in the absence of a comparable change in orbital forcing. Currently, only  
19 two Mg/Ca-derived, high-resolution bottom water temperature (BWT) records exist that can be  
20 used with  $\delta^{18}\text{O}_{\text{benthic}}$  records to separate temperature and ice volume signals over the Pleistocene.  
21 However, these two BWT records suggest a different pattern of climate change occurred over the  
22 MPT—a record from North Atlantic DSDP Site 607 suggests BWT decreased with no long-term  
23 trend in ice volume over the MPT, while South Pacific ODP Site 1123 suggests that BWT has  
24 been relatively stable over the last 1.5 Myr but that there was an abrupt increase in ice volume at  
25 ~900 kyr. In this paper we attempt to reconcile these two views of climate change across the  
26 MPT. Specifically, we investigated the suggestion that the secular BWT trend obtained from  
27 Mg/Ca measurements on *Cibicidoides wuellerstorfi* and *Oridorsalis umbonatus* species from N.  
28 Atlantic Site 607 is biased by the possible influence of  $\Delta[\text{CO}_3^{2-}]$  on Mg/Ca values in these  
29 species by generating a low-resolution BWT record using *Uvigerina* spp., a genus whose Mg/Ca  
30 values are not thought to be influenced by  $\Delta[\text{CO}_3^{2-}]$ . We find a long-term BWT cooling of ~2-

31 3°C occurred from 1500 to ~500 kyr in the N. Atlantic, consistent with the previously generated  
32 *C. wuellerstorfi* and *O. umbonatus* BWT record. We also find that changes in ocean circulation  
33 likely influenced  $\delta^{18}\text{O}_{\text{benthic}}$ , BWT, and  $\delta^{18}\text{O}_{\text{seawater}}$  records across the MPT. N. Atlantic BWT  
34 cooling starting at ~1.2 Ma, presumably driven by high-latitude cooling, may have been a  
35 necessary precursor to a threshold response in climate-ice sheet behavior at ~900 ka. At that  
36 point, a modest increase in ice volume and thermohaline reorganization may have caused  
37 enhanced sensitivity to the 100 kyr orbital cycle.

38

### 39 **1. Introduction**

40 Variations in the Earth's orbit, and consequently incoming solar radiation, act as a  
41 pacemaker for glacial-interglacial cycles [*Hays et al.*, 1976]. During the mid-Pleistocene  
42 transition (MPT, ~1.2 to 0.65 Ma) the glacial-interglacial periodicity as recorded by marine  $\delta^{18}\text{O}$   
43 records of benthic foraminifera ( $\delta^{18}\text{O}_{\text{benthic}}$ ) changed from primarily 41 to 100 kyr year cycles  
44 without any obvious change in external orbital forcing [*Shackleton and Opdyke*, 1976; *Pisias and*  
45 *Moore*, 1981; *Imbrie et al.*, 1992; *Raymo and Nisancioglu*, 2003]. As the 100 kyr eccentricity  
46 cycle has a relatively weak effect on incoming solar radiation [*Imbrie et al.*, 1993], feedbacks  
47 within the climate system must have amplified the glacial-interglacial response to the 100 kyr  
48 cycle. However, after nearly 40 years of research, the processes and mechanisms that caused the  
49 MPT remain enigmatic [*Clark et al.*, 2006].

50 Explanations as to what caused the MPT are diverse (summarized in [*McClymont et al.*,  
51 2013]), and they generally invoke changes in climate boundary conditions such as: (1) gradual  
52 global cooling, possibly related to  $p\text{CO}_2$  decrease [*Berger and Jansen*, 1994; *Raymo et al.*,  
53 1997], (2) changes in ice sheet dynamics that allow for the build up of large ice sheets [*Clark et*

54 *al.*, 2006; *Elderfield et al.*, 2012], and/or (3) thermohaline circulation reorganization [*Schmieder*  
55 *et al.*, 2000; *Sexton and Barker*, 2012; *Pena and Goldstein*, 2014]. Much of our understanding of  
56 the MPT comes from  $\delta^{18}\text{O}_{\text{benthic}}$  records, which reflect a combination of bottom water  
57 temperature (BWT) and seawater oxygen isotope composition ( $\delta^{18}\text{O}_{\text{seawater}}$ ) [*Shackleton*, 1967;  
58 *Pisias and Moore*, 1981; *Labeyrie et al.*, 1987; *Maasch*, 1988; *Ruddiman et al.*, 1989; *Saltzman*  
59 *and Maasch*, 1991; *Mudelsee and Schulz*, 1997; *Rutherford and D'Hondt*, 2000; *Lisiecki and*  
60 *Raymo*, 2005; 2007]. Despite years of study and statistical analysis, disagreement still exists as to  
61 whether the MPT reflects a gradual transition that started  $\sim 1.2$  Ma and finished  $\sim 0.65$  Ma [*Pisias*  
62 *and Moore*, 1981; *Ruddiman et al.*, 1989; *Rutherford and D'Hondt*, 2000; *Clark et al.*, 2006;  
63 *Sosdian and Rosenthal*, 2009; *McClymont et al.*, 2013] or represents an abrupt event centered  
64 around 0.9 Ma [*Maasch*, 1988; *Mudelsee and Schulz*, 1997; *Elderfield et al.*, 2012]. The timing  
65 of the MPT is particularly relevant to evaluating the mechanisms of climate change (e.g., a  
66 gradual response of ice to a gradual forcing, an abrupt response of ice to an abrupt forcing event,  
67 or a threshold (abrupt) response of ice to long-term gradual climate forcing). Because  $\delta^{18}\text{O}_{\text{benthic}}$   
68 records a number of environmental parameters (global ice volume, BWT, and local  
69 hydrography), part of the disagreement over the timing and character of the MPT may be related  
70 to the use of different  $\delta^{18}\text{O}_{\text{benthic}}$  records, with local hydrographic effects obscuring global trends  
71 in temperature and ice volume in some records [*Clark et al.*, 2006; *Elderfield et al.*, 2012; *Bates*  
72 *et al.*, 2014].

73 Because  $\delta^{18}\text{O}_{\text{benthic}}$  reflects BWT and  $\delta^{18}\text{O}_{\text{seawater}}$  (e.g. ice volume and local hydrography),  
74 several environmental signals are imprinted on any given  $\delta^{18}\text{O}_{\text{benthic}}$  record. To isolate BWT and  
75  $\delta^{18}\text{O}_{\text{seawater}}$  components of  $\delta^{18}\text{O}_{\text{benthic}}$  records, the Mg/Ca ratio of benthic foraminiferal calcite is

76 widely used as an independent proxy of BWT on orbital and tectonic timescales [*Lear et al.*,  
77 2000; *Billups and Schrag*, 2002; *Sosdian and Rosenthal*, 2009; *Elderfield et al.*, 2010; 2012].

78 Currently, only two high-resolution coupled benthic foraminifera Mg/Ca and  $\delta^{18}\text{O}_{\text{benthic}}$   
79 records exist across the MPT and their implications for how global ice volume changed across  
80 the MPT disagree. The first record is the N. Atlantic (the *Sosdian and Rosenthal* (2009) record  
81 from Deep Sea Drilling Project (DSDP) Site 607, 41.0° N, 33.0° W, 3427 m water depth), and  
82 the second is from the S. Pacific (the *Elderfield et al.* (2012) record from Ocean Drilling Program  
83 (ODP) Site 1123 located on the Chatham Rise, 41.8° S, 171.5° W, 3290 water depth) (Figure 1).  
84 The interpretation of Site 607 record has been challenged based on the observation that, in  
85 addition to temperature, changes in carbonate ion saturation ( $\Delta[\text{CO}_3^{2-}]$ ) could influence Mg/Ca in  
86 benthic species thus biasing BWT estimates in some species (e.g. *Cibicidoides wuellerstorfi*; [*Yu*  
87 *and Broecker*, 2010]; but also see [*Sosdian and Rosenthal*, 2010]). The  $\Delta[\text{CO}_3^{2-}]$  is defined as  
88  $[\text{CO}_3^{2-}]_{\text{in situ}} - [\text{CO}_3^{2-}]_{\text{saturation}}$ .

89 In addition, the distinct differences in implied BWT histories at Sites 607 and 1123 (and  
90 thus climate interpretations) also merit further study. For instance, The N. Atlantic *Sosdian and*  
91 *Rosenthal* (2009) record at Site 607 suggests that significant BWT cooling ( $\sim 2^\circ\text{C}$ ) occurred over  
92 an interval of a million years, from  $\sim 1.5$  to 0.5 Ma. On the other hand, The S. Pacific *Elderfield*  
93 *et al.* (2012) record from Site 1123 shows relatively constant, near-freezing temperatures over  
94 the over the entire 1.5 Myr record. Although a significant volume of global water flows over  
95 Chatham Rise, this area is also sensitive to the relative mixture of northern and southern  
96 component water masses (Figure 1).

97 Given that these sites are in different oceans and at different depths, it is possible that  
98 they could be characterized by different BWT trends; however, when the Mg/Ca-derived BWTs

99 are subtracted from the  $\delta^{18}\text{O}_{\text{benthic}}$  records at each site, two very different patterns of  $\delta^{18}\text{O}_{\text{seawater}}$   
100 are observed across the MPT implying seemingly contradictory global ice volume histories.  
101 Specifically, Site 607 shows no discernible long-term trend in  $\delta^{18}\text{O}_{\text{seawater}}$  across the MPT  
102 [Sosdian and Rosenthal, 2009], whereas Site 1123 record shows an abrupt increase occurring at  
103 ~900 ka, which Elderfield et al. [2012] attributes to an large expansion of Antarctic ice volume.  
104 Instead of global cooling of the deep ocean across the early Pleistocene, the site 1123 record was  
105 interpreted as suggesting that an abrupt change in ice sheet dynamics and/or ice volume increase  
106 may have been the primary cause of the MPT.

107 Here we revisit DSDP Site 607 and generate a Mg/Ca-derived BWT record from 500 to  
108 1500 ka using mixed *Uvigerina* spp.. We compare this record to the previous published *C.*  
109 *wuellerstorfi* and *Oridorsalis umbonatus* record [Sosdian and Rosenthal, 2009] in order to  
110 evaluate long-term changes in N. Atlantic BWT over the MPT. We also evaluate the possible  
111 influence of  $\Delta[\text{CO}_3^{2-}]$  on Mg/Ca BWT reconstructions based on *Uvigerina* spp., *C. wuellerstorfi*  
112 and *Oridorsalis umbonatus*. We compare  $\delta^{18}\text{O}_{\text{benthic}}$ , BWT and  $\delta^{18}\text{O}_{\text{seawater}}$  records from Site 607  
113 with those from Site 1123 in the S. Pacific. Lastly, we consider how changes in ocean circulation  
114 could influence the interpretation of  $\delta^{18}\text{O}_{\text{benthic}}$ , BWT and  $\delta^{18}\text{O}_{\text{seawater}}$  records from these two sites.

115

### 116 3. Materials and methods

117 Mg/Ca were measured on mixed *Uvigerina* spp. from DSDP 607 to estimate bottom  
118 water temperature. Previous calibration studies suggests that the Mg/Ca ratios of infaunal species  
119 such as *Uvigerina* spp. primarily respond to temperature [Elderfield et al., 2010] and show very  
120 little  $\Delta[\text{CO}_3^{2-}]$  effect [Yu and Elderfield, 2007].  $\Delta[\text{CO}_3^{2-}]$  approaches zero at shallow depths  
121 within the sediment because pore waters come to rapid equilibration with calcium carbonate,

122 regardless as to whether the bottom water  $\Delta[\text{CO}_3^{2-}]$  is oversaturated or undersaturated at  
123 particular location [Elderfield et al., 2006]. By contrast, calibration studies suggest that the  
124 Mg/Ca ratios of epifaunal species such as *Cibicidoides wuellerstorfi* respond to both temperature  
125 and changes in  $\Delta[\text{CO}_3^{2-}]$  [Elderfield et al., 2006; Yu and Elderfield, 2007; Elderfield et al.,  
126 2010], possibly resulting in biased temperature reconstructions in areas with markedly large  
127 changes in bottom water  $\Delta[\text{CO}_3^{2-}]$  [Yu and Broecker, 2010]. Thus, it is hypothesized that Mg/Ca  
128 measurements derived from *Uvigerina* spp. would more accurately reflect BWT variations  
129 through time [Elderfield et al., 2010; 2012].

130 *Uvigerina* spp. specimens were picked from the >150  $\mu\text{m}$  size fraction with  
131 approximately 5-15 specimens analyzed per sample, with a median of 7 specimens per sample.  
132 The average sampling resolution is one sample every ~9,000 years. The *Uvigerina* spp.  
133 measurements are relatively uniformly distributed between glacial and interglacial periods  
134 (Supplemental Figure 1). The vast majority (>88%) of the samples are from the same core  
135 samples that were used to generate the Ruddiman et al., [1989]  $\delta^{18}\text{O}_{\text{benthic}}$  record. Samples were  
136 cleaned using standard analytical procedures [Rosenthal et al., 1999; Martin and Lea, 2002] -  
137 samples were repeatedly rinsed in water and methanol to remove clays and cleaned with  
138 reductive and oxidative reagents. Owing to small sample weights (~100  $\mu\text{g}$ ), not all samples  
139 were subjected to a weak acid leaching. Elemental measurements were made either at Lamont-  
140 Doherty Earth Observatory (LDEO) on a Thermo Scientific iCAP-Q inductively coupled plasma  
141 mass spectrometer (ICP-MS) or at Rutgers University on a Finnigan Element-XR ICP-MS.  
142 Al/Ca, Fe/Ca, Mn/Ca and Ti/Ca (Supplemental Figure 2) were used to monitor samples for  
143 contamination due to diagenetic overprinting or to residual detrital material (n = 6 samples  
144 removed) [Rosenthal et al., 1999]. We find no obvious downcore trends in Mg/Ca and Al/Ca,

145 Fe/Ca, and Ti/Ca. Mn/Ca values are generally high ( $\mu = 126.5 \mu\text{mol/mol}$   $\sigma = 53.56 \mu\text{mol/mol}$ )  
146 and correlate with Mg/Ca ( $R^2 = 0.33$ ,  $<0.001$ ). These high Mn/Ca values do not appear to be due  
147 to Mn-Fe oxide and  $\text{MnCO}_3$  contamination because all samples were reductively cleaned and we  
148 found that that after repetitive weak acid leaches, the Mn/Ca values remained relatively high  
149 within a subset of samples. There is a weak correlation downcore between Mn/Ca and Mg/Ca but  
150 we believe this is a coincidence not caused by the same process. Elderfield et al. [2012] see a  
151 similar weak correlation downcore at Site 1123 between Mn/Ca and Mg/Ca that they attribute to  
152 diagenetic reorganization. A similar process may have occurred at Site 607.

153 Instrumental precision at LDEO and Rutgers was 0.5-1%. Long-term precision of a liquid  
154 consistency standard at LDEO and Rutgers University was between 1-2%. At LDEO,  
155 reproducibility of three consistency standards with Mg/Ca of 1.4840, 4.1325, 8.1035 mmol/mol  
156 were  $\pm 1.3\%$ ,  $\pm 1.8\%$ , and  $\pm 1.7\%$ , respectively. At Rutgers University, reproducibility of  
157 consistency standards with Mg/Ca of 1.25, 3.32, and 7.51 mmol/mol were  $\pm 1.2\%$ ,  $\pm 1.2\%$ , and  
158  $\pm 0.6\%$ , respectively. To ensure inter-laboratory accuracy, standards from Rutgers University  
159 were run at LDEO. Inter-laboratory accuracy was found to be 3.0%, within the range of  
160 previously reported inter-laboratory comparisons [Rosenthal et al., 2004].

161 There are several species-specific calibrations to choose from to account for known  
162 differences in temperature sensitivity and offsets for various benthic foraminifera species [Lear  
163 et al., 2002; Healey et al., 2008; Yu and Elderfield, 2008; Sosdian and Rosenthal, 2009;  
164 Elderfield et al., 2012; Cappelli et al., 2015]. For ease of comparison with the site 1123  
165 *Uvigerina* spp. record, we use a modified version of the temperature equation, and the same  $\delta^{18}\text{O}$   
166 equation, as used by Elderfield et al. (2010, 2012). To correct for known Mg loss during the  
167 reductive step [Barker et al., 2003; Rosenthal et al., 2004] used in our cleaning procedure, we



168 adjusted the Elderfield *Uvigerina* spp. paleotemperature equation by 10%:  $Mg/Ca = 0.9 +$   
 169  $0.1 * BWT$ . Previously published *Cibicidoides wuellerstorfi*, and *Oridorsalis umbonatus*  
 170 paleotemperature equations were used for further BWT comparisons between species and to  
 171 account for species offsets [Sosdian and Rosenthal, 2009]. Finally, to estimate  $\delta^{18}O_{seawater}$ , we  
 172 used the published  $\delta^{18}O_{benthic}$  record [Ruddiman et al., 1989], the Mg/Ca-derived BWT, and the  
 173  $\delta^{18}O_{seawater}$  equation:  $\delta^{18}O_{seawater} = \delta^{18}O_{benthic} + 0.27 - 0.25 * (16.9 - BWT)$  [O'Neil et al., 1969;  
 174 Shackleton, 1974; Kim and O'Neil, 1997; Elderfield et al., 2010; 2012].

175 To investigate changes in ocean circulation across the MPT we use previously published  
 176  $\delta^{13}C$  values of benthic foraminifera ( $\delta^{13}C_{benthic}$ ) (Table 1) and calculated the percent Northern  
 177 Component Water (%NCW) [Oppo and Fairbanks, 1987; Raymo et al., 1990; Flower et al.,  
 178 2000; Venz and Hodell, 2002]. The %NCW is used to estimate the relative contribution of deep  
 179 water formed in the N. Atlantic (i.e. North Atlantic Deep Water, NADW) at a particular site  
 180 [Oppo and Fairbanks, 1987] and is calculated using the following equation:

$$181 \quad \%NCW = \frac{\delta^{13}C_{SITE} - \delta^{13}C_{SCW}}{\delta^{13}C_{NCW} - \delta^{13}C_{SCW}} * 100$$

182 where the  $\delta^{13}C_{SITE}$  is the  $\delta^{13}C_{benthic}$  value at a given location of interest and  $\delta^{13}C_{NCW}$  and  $\delta^{13}C_{SCW}$   
 183 are the  $\delta^{13}C_{benthic}$  values at the designated northern and southern (i.e. Antarctic Bottom Water)  
 184 water mass end-member locations. Here, we use ODP Site 982 and ODP 1090 as a North  
 185 Atlantic and South Atlantic end-member, respectively [Venz and Hodell, 2002].

186

## 187 **4. Results and Discussion**

### 188 **4.1 Comparison of *Uvigerina* spp., *Cibicidoides wuellerstorfi*, and *Oridorsalis umbonatus*** 189 **records at DSDP Site 607**

190 The new Mg/Ca record derived from *Uvigerina* spp. at Site 607 shows a decreasing trend  
191 from 1500 to 500 kyrs (Figure 2). The *Uvigerina* spp. Mg/Ca record has a similar long-term  
192 trend and variability to the previously published *Cibicidoides wuellerstorfi*, and *Oridorsalis*  
193 *umbonatus* Mg/Ca records [Sosdian and Rosenthal, 2009]. Uncertainty in Mg/Ca values is based  
194 on the 95% confidence of Monte Carlo error propagation (1000 resampling,  $\sigma = 0.128$   
195 mmol/mol for *Uvigerina* spp., this study and  $\sigma = \sim 0.1$  for *Cibicidoides wuellerstorfi*, and  
196 *Oridorsalis umbonatus* Sosdian and Rosenthal, [2009]. Offsets in Mg/Ca values are likely  
197 related to vital effects.

198 Large changes in deep ocean chemistry occurred over the MPT [Raymo et al., 1997;  
199 Lisiecki, 2014] as evidenced from the  $\text{CaCO}_3$  [Ruddiman et al., 1989],  $\delta^{13}\text{C}_{\text{benthic}}$  [Ruddiman et  
200 al., 1989] and %NCW (Figure 2) at Site 607, changes which may have influenced *C.*  
201 *wuellerstorfi* and *O. umbonatus* Mg/Ca values [e.g. Yu and Broecker, 2010]. Specifically, after  
202  $\sim 900$  kyr a step change in deep ocean chemistry is suggested by an increased amplitude of  
203 variability in  $\text{CaCO}_3$ ,  $\delta^{13}\text{C}_{\text{benthic}}$ , and %NCW, as well as a greater contribution of low  $[\text{CO}_3^{2-}]$   
204 SCW water at Site 607, particularly during glacial periods. However, given the similarity in the  
205 overall long-term Mg/Ca trends in *Uvigerina* spp., *C. wuellerstorfi* and *O. umbonatus*, if  
206 *Uvigerina* spp. Mg/Ca values primarily respond to changes in temperature [Elderfield et al.,  
207 2010; 2012], then much of the long-term trend in *C. wuellerstorfi* and *O. umbonatus* Mg/Ca  
208 values (Figure 2) may also be explained by temperature and not by long-term changes in deep  
209 sea carbon chemistry (i.e. changes in  $[\text{CO}_3^{2-}]$ ) during the 1500 to 500 kyr time interval at Site  
210 607.

211 Additionally, we performed an F test to determine whether the large amplitude changes  
212 in deep ocean chemistry after  $\sim 900$  kyr also influenced the variability in Mg/Ca values of

213 *O. umbonatus*, *C. wuellerstorfi* and *Uvigerina* spp.. We compared the Mg/Ca values between the  
214 600 – 800 ka and 1000 – 1250 ka, intervals of large and small amplitude changes in deep ocean  
215 chemistry, respectively, for each species and found no statistical difference Mg/Ca variability  
216 prior to or after ~900 kyr (Table 1). The lack of coherence and synchronicity in the long-term  
217 trend and variability between Mg/Ca and the deep ocean chemistry parameters suggests changes  
218 in [CO<sub>3</sub><sup>2-</sup>] do not dominate benthic foraminifera Mg/Ca values during the MPT at Site 607  
219 [*Sosdian and Rosenthal*, 2009; 2010].

220 The new BWT record derived from *Uvigerina* Mg/Ca values at Site 607 shows a  
221 decreasing trend in mean temperature of ~2-3°C from 1500 to 500 kyr (Figure 3), regardless of  
222 calibration choice, suggesting a long-term cooling of bottom waters occurred. In Figure 3,  
223 *O. umbonatus*, *C. wuellerstorfi* and *Uvigerina* spp. Mg/Ca values were converted to temperature  
224 using available low-temperature species-specific temperature calibrations [*Lear et al.*, 2002;  
225 *Healey et al.*, 2008; *Yu and Elderfield*, 2008; *Sosdian and Rosenthal*, 2009; *Elderfield et al.*,  
226 2012; *Cappelli et al.*, 2015]. The magnitude of long-term temperature cooling between 1500 to  
227 500 kyr is dependent on calibration choice with *C. wuellerstorfi* broadly showing less cooling  
228 than *O. umbonatus* and *Uvigerina* spp.. These differences in temperature reconstructions are not  
229 related to changes in the presence or absence of different species during interglacial and glacial  
230 periods as the species are generally found throughout the 1500 to 500 kyr interval (see  
231 Supplemental Figure 1 for *Uvigerina* spp. distribution). The previously published Mg/Ca records  
232 of *O. umbonatus* and *C. wuellerstorfi* [*Sosdian and Rosenthal*, 2009] suggest a ~2°C and ~1-2°C  
233 cooling trend from 1500 to 500 kyr respectively, depending on calibration choice, which is in  
234 broad agreement with the *Uvigerina* spp. record (Figure 3). Additionally, all species show a

235 similar magnitude of cooling at the warm and cold range of temperatures which suggests there  
236 was a similar magnitude of cooling in the interglacial and glacial intervals over the MPT  
237 [Sosdian and Rosenthal, 2009]. Likewise, for the late Pleistocene, the previously published  
238 *O. umbonatus* and *C. wuellerstorfi* records are in good agreement with an ostracode Mg/Ca-  
239 BWT record [Dwyer et al., 1995], though there is also disagreement as to whether ostracode  
240 Mg/Ca values also respond to changes  $\Delta[\text{CO}_3^{2-}]$  [Elmore et al., 2012; Farmer et al., 2012].  
241 Regardless of species-specific calibration choice, the *Uvigerina* spp., *O. umbonatus* and *C.*  
242 *wuellerstorfi* records show a similarly decreasing BWT trend over the MPT.

243         Based on the above observations, we combine the *Uvigerina* spp., *C. wuellerstorfi* and *O.*  
244 *umbonatus* data to create a composite BWT record at Site 607 (Figure 4A). Mg/Ca values for  
245 each species were converted to temperature using the species-specific paleotemperature  
246 equations [Sosdian and Rosenthal, 2009; Elderfield et al., 2010; 2012] to account for species  
247 offsets and Mg/Ca-temperature sensitivities. Using the *Uvigerina* spp. BWT generated in this  
248 study and the previously published  $\delta^{18}\text{O}_{\text{benthic}}$  record [Ruddiman et al., 1989], paired  $\delta^{18}\text{O}_{\text{seawater}}$   
249 values were calculated (blue dots, Figure 4B). We then combine our *Uvigerina* spp.-derived  
250  $\delta^{18}\text{O}_{\text{seawater}}$  values with the previously published *C. wuellerstorfi* and *O. umbonatus*-derived  
251  $\delta^{18}\text{O}_{\text{seawater}}$  values [Sosdian and Rosenthal, 2009] to create a composite  $\delta^{18}\text{O}_{\text{seawater}}$  record (Figure  
252 4). We note that during the early Pleistocene some warm BWT estimates result in reconstructed  
253  $\delta^{18}\text{O}_{\text{seawater}}$  values approaching +1.5‰, which seems unlikely considering the early Pleistocene is  
254 thought to have had less ice than the recent glacial period. Part of the issue may be that several  
255 species of benthic foraminifera are used in these reconstructions (e.g. the  $\delta^{18}\text{O}_{\text{benthic}}$  is based on  
256 *C. wuellerstorfi* and the BWT is based on *Uvigerina* spp.) or that BWT and  $\delta^{18}\text{O}_{\text{benthic}}$  records are  
257 not in phase (i.e. temperature leads ice volume, Sosdian and Rosenthal, [2009]). To account for

258 the fact that these records are based on different species with different temperature and  $\delta^{18}\text{O}$   
259 sensitivities, the composite BWT and  $\delta^{18}\text{O}_{\text{seawater}}$  records were smoothed using a three point  
260 moving to reduce the error on the estimates (see Supplemental Text). By smoothing the records,  
261 we can focus on our primary goal of understanding the long-term trends in BWT and  $\delta^{18}\text{O}_{\text{seawater}}$ .  
262 The advantage of combining *Uvigerina* spp., *C. wuellerstorfi* and *O. umbonatus* analyses is that  
263 we are able to achieve a more continuous, high-resolution record because these species often  
264 alter in abundance with *Uvigerina* being generally more abundant during glacial periods.

265 Interestingly, over the same period, the  $\delta^{18}\text{O}_{\text{benthic}}$  record at Site 607 has a decreasing  
266 trend of 0.317‰ (Figure 5), which is equivalent to a  $\sim 1.3^\circ\text{C}$  cooling using a sensitivity of 0.25‰  
267 per  $^\circ\text{C}$ . A combination of possibilities could explain this trend: 1) the  $\delta^{18}\text{O}_{\text{benthic}}$  record is  
268 dominated by BWT cooling and there was little ice volume increase over the MPT, 2) there was  
269 a change in ice sheet end member values and thus mean ocean  $\delta^{18}\text{O}_{\text{seawater}}$  (e.g. [Winnick and  
270 Caves, 2015]), and/or 3) changes in ocean circulation may influence the  $\delta^{18}\text{O}_{\text{seawater}}$  values at Site  
271 607. Due to the variable constraints on  $\delta^{18}\text{O}_{\text{seawater}}$  and the large error on the  $\delta^{18}\text{O}_{\text{seawater}}$  estimate,  
272 we do not intend to reconstruct sea level and instead focus on the long-term trends in BWT and  
273  $\delta^{18}\text{O}_{\text{seawater}}$  at Site 607.

274

#### 275 **4.2. Comparison of Site 607 and Site 1123 BWT records**

276 The BWT,  $\delta^{18}\text{O}_{\text{benthic}}$  and  $\delta^{18}\text{O}_{\text{seawater}}$  records at Sites 607 and 1123 have dissimilar long-  
277 term trends from 1500 to 500 kyr (Figure 5). While Site 607 shows a subtle yet distinct long-  
278 term BWT cooling trend from 1500 to  $\sim 500$  kyr (taking into account uncertainties discussed in  
279 previous section; Figure 4, 5), Site 1123 data show colder, near freezing temperatures

280 characterizing the entire 1500 kyr record (Figure 5) [Elderfield *et al.*, 2012]. The  $\delta^{18}\text{O}_{\text{benthic}}$   
281 records at Sites 607 and 1123 correlate well with the globally averaged benthic oxygen isotope  
282 LR04 stack [Lisiecki and Raymo, 2005] (Figure 5) except during the glacial periods between  
283 MIS 22 and MIS 18 when Site 1123 shows a pronounced  $\sim 0.3$  to  $0.5\text{‰}$  increase during glacial  
284 periods relative to LR04. At MIS 16 the  $\delta^{18}\text{O}_{\text{benthic}}$  records converge again.

285 Site 607  $\delta^{18}\text{O}_{\text{seawater}}$  values show no change in the long-term mean or amplitude from  
286 1500 to 500 kyr (Figure 4, 5). On the other hand, at Site 1123,  $\delta^{18}\text{O}_{\text{seawater}}$  values show an  
287 increase in the mean and the amplitude after MIS 22 (Figure 5). The site 1123  $\delta^{18}\text{O}_{\text{seawater}}$  record  
288 has been interpreted as showing a large build up of ice volume at MIS 22 [Elderfield *et al.*,  
289 2012], relative to the earlier interval, whereas the Site 607  $\delta^{18}\text{O}_{\text{seawater}}$  record shows no obvious,  
290 abrupt increase in ice volume occurred at MIS 22. If there was a large increase in glacial ice  
291 volume at MIS 22, relative to earlier intervals, as Elderfield *et al.* (2012) suggests, then it is  
292 difficult to reconcile the different  $\delta^{18}\text{O}_{\text{seawater}}$  records at Site 1123 and 607 because a large  
293 increase in ice volume should be common to both records. This suggests that hydrographic  
294 changes may have occurred over the MPT at one or both sites, changes that could have  
295 influenced the  $\delta^{18}\text{O}_{\text{seawater}}$ . In the next sections we explore possible hydrographic changes at Sites  
296 1123 and 607.

297

#### 298 ***4.3. Teasing apart local changes in ocean circulation and global climate***

299  $\delta^{18}\text{O}_{\text{seawater}}$  values depend *globally* on the build-up/loss of  $^{16}\text{O}$  on the continents on  
300 glacial-interglacial time scales and *locally* on water that flows over any abyssal location,  
301 integrating both global climate and regional bottom water properties imprinted by high latitude  
302 deep-water formation and ocean circulation. One implication of the Elderfield *et al.*, [2012]

303 record is that if the Site 1123  $\delta^{18}\text{O}_{\text{seawater}}$  is assumed to be the global  $\delta^{18}\text{O}_{\text{seawater}}$  signal (as  
304 suggested by those authors), then the residual  $\delta^{18}\text{O}_{\text{benthic}}$  temperature component in the LR04  
305 stack as well as in the Sites 607, 1020, 1146, 1143, 846, 849  $\delta^{18}\text{O}_{\text{benthic}}$  records would suggest a  
306 warming of the deep Atlantic and Pacific water of nearly 2°C (e.g.,  $\sim 0.5\text{‰}$ , assuming  $0.25\text{‰}/\text{°C}$ )  
307 occurred during glacial MIS 22 (Figure 5). Indeed, if Elderfield et al.'s [2012] conclusions about  
308 global ice volume are correct, then these sites would seemingly reflect bottom water warming  
309 during glacial periods across the MPT, until MIS 16 when the  $\delta^{18}\text{O}_{\text{benthic}}$  converge again (Figure  
310 5). It is difficult to explain why large expanses of the deep Atlantic and Pacific oceans would be  
311 warming during glacial excursions across the MPT. Here we suggest that changes in ocean  
312 circulation may influence the magnitude of the abrupt change observed at Site 1123 and the  
313 record reflects both global and local changes. Numerous studies have documented changes in  
314 deep ocean circulation over the MPT using  $\delta^{13}\text{C}$  of benthic foraminifera [Raymo et al., 1997;  
315 Venz et al., 1999; Venz and Hodell, 2002; Hodell et al., 2003; Ferretti et al., 2010; Poirier and  
316 Billups, 2014] and carbonate preservation [Schmieder et al., 2000; Sexton and Barker, 2012].  
317 Here we examine how these circulation changes may have influence the  $\delta^{18}\text{O}_{\text{benthic}}$  and Mg/Ca-  
318 derived  $\delta^{18}\text{O}_{\text{seawater}}$  records from Site 1123 and 607.

319

#### 320 **4.3.1. Changes in ocean circulation from MIS 24 to 22**

321 The  $\delta^{18}\text{O}_{\text{benthic}}$  record at Site 1123 (and thus also the calculated  $\delta^{18}\text{O}_{\text{seawater}}$  record) stands  
322 out as anomalous in comparison to the LR04 stack (Figure 5). Elderfield et al. [2012] argue that  
323 Site 1123 is representative of global conditions because a large volume of water flows over the  
324 Chatham Rise and into the Pacific (Figure 1) and LR04 is not representative of global conditions  
325 because it is biased to the Atlantic Ocean; however, Site 1123  $\delta^{18}\text{O}_{\text{benthic}}$  values are different from

326 many of the records from the rest of the Pacific basin. In comparison to the previously published  
327 high-resolution mid-to-deep (i.e. >2000 m water depth, Table 2)  $\delta^{18}\text{O}_{\text{benthic}}$  records from the  
328 South China Sea, California Margin, and equatorial Pacific (Figure 6), Site 1123 is  $\sim 0.4\text{‰}$   
329 heavier than most of the other Pacific records. Although Elderfield et al. [2012] point out Site  
330 1123 shows good agreement with Site 677, Site 677 is located in the Panama Basin (sill depth  
331 2300 to 2920 m [Lonsdale, 1977]) and is bathymetrically isolated from the rest of the deep  
332 eastern equatorial Pacific. The other Pacific records, including those from the deep eastern  
333 equatorial Pacific (e.g. Sites 846 and 849, water depth 3307 and 3849 m, respectively), show  
334 good agreement with LR04. The choice of benthic species used in generating the  $\delta^{18}\text{O}_{\text{benthic}}$   
335 records also does not appear to contribute to these discrepancies. Although many of these  
336  $\delta^{18}\text{O}_{\text{benthic}}$  records are based on *C. wuellerstorfi* or a mixture of *C. wuellerstorfi* and *Uvigerina*  
337 spp., as Elderfield et al. [2012] note, at Site 1123 during MIS 22, the abrupt shift in  $\delta^{18}\text{O}_{\text{benthic}}$  is  
338 observed in both *C. wuellerstorfi* and *Uvigerina* spp. samples. The similarity between *C.*  
339 *wuellerstorfi* and *Uvigerina* spp.  $\delta^{18}\text{O}_{\text{benthic}}$  samples appears to hold for the rest of the Pacific.  
340 Consequently, we infer that the very positive  $\delta^{18}\text{O}_{\text{benthic}}$  excursions at MIS 22 at Sites 677 and  
341 1123 may reflect a local hydrographic effects, possibly due to the time-varying mixing of  
342 different water masses, that are not recorded in the rest of the Pacific basin (Figure 1).

343 In the Atlantic, the relative importance of NCW and SCW changed over the MPT  
344 [Raymo et al., 1997; Venz and Hodell, 2002; Hodell et al., 2003; Pena et al., 2008; Lisiecki,  
345 2014]. Significant changes in ocean circulation over the MPT may have created hydrographic  
346 conditions unique to Site 1123 that are not representative of deep Pacific Ocean or global  
347 conditions. This change could have had downstream consequences for the Pacific because a  
348 mixture of SCW and NCW flows into the deep Pacific [Hall et al., 2001; Elderfield et al., 2012].



349 We suggest that Site 1123  $\delta^{18}\text{O}_{\text{seawater}}$  record is likely not representative of mean ocean  
350  $\delta^{18}\text{O}_{\text{seawater}}$  because its location on the Chatham Rise appears to be sensitive to the mixing ratio of  
351 SCW and NCW (Figure 1).

352 In Figure 7,  $\delta^{13}\text{C}_{\text{benthic}}$  and %NCW records show marked changes in glacial deep ocean  
353 circulation in the Atlantic across the MPT (for completeness, the Pacific basin can be found in  
354 Supplemental Figure 3). Prior to MIS 24, NCW filled the Atlantic above ~3500m during glacial  
355 periods [Raymo et al., 1997; Venz and Hodell, 2002; Hodell et al., 2003; Pena et al., 2008;  
356 Lisiecki, 2014]. Beginning at MIS 24 and intensifying at MIS 22, Nd isotopes and  $\delta^{13}\text{C}_{\text{benthic}}$   
357 from intermediate and deep water horizons suggest bottom waters were predominantly ventilated  
358 by SCW [Raymo et al., 1997; Venz and Hodell, 2002; Hodell et al., 2003; Pena et al., 2008;  
359 Lisiecki, 2014; Poirier and Billups, 2014]. These changes in Atlantic circulation may be related  
360 to an observed decrease in the mean flow speed over the Chatham Rise starting at MIS 22 (e.g.  
361 Site 1123; [Hall et al., 2001; Venuti et al., 2007]), a hydrographic change which may have  
362 contributed to the unique  $\delta^{18}\text{O}_{\text{benthic}}$  signal recorded at Site 1123.

363 This change in Atlantic deep ocean circulation likely influenced the Site 607  $\delta^{18}\text{O}_{\text{benthic}}$   
364 record as well as the records throughout the Atlantic. Elderfield et al. [2012] point out Site  
365 U1308 also shows an abrupt increase in  $\delta^{18}\text{O}_{\text{benthic}}$  at MIS 22 which is similar to Site 1123  
366 (Figure 8). However, this abrupt increase in  $\delta^{18}\text{O}_{\text{benthic}}$  is absent in the Site 607 and LR04  
367 records. Although Sites 607 and U1308 are both in the N. Atlantic, they experience different  
368 water masses because they are on different sides of the basin separated by the mid-Atlantic ridge.  
369 Site 607 is in the western Atlantic basin and ventilated by either NCW and SCW whereas U1308  
370 is in the eastern Atlantic basin which is ventilated from the south by a mixture of NCW and  
371 SCW that flows over the Romanche Fracture Zone at the equator [Raymo et al., 1997]. The value

372 of using a stacked record like LR04 is that it changes in hydrography are mostly averaged out.  
373 The LR04 stack, though biased toward the Atlantic, has at least 23 records that on average do not  
374 show an abrupt change in  $\delta^{18}\text{O}_{\text{benthic}}$  at MIS 22. The LR04 stack combined with the observations  
375 in the Pacific this suggests that an abrupt, large magnitude change at MIS 22 is likely related to  
376 changes in ocean circulation ventilating that particular area rather than whole ocean changes in  
377  $\delta^{18}\text{O}$  related to ice volume.

378

#### 379 ***4.3.2 Long-term cooling at Site 607***

380 The cause of the long-term bottom water cooling found in at Site 607 and its relationship  
381 to the emergence of the 100 kyr cycle remains elusive. Long-term cooling that begins or  
382 intensifies at  $\sim 1.2$  Ma is found in multiple mid- to high latitude sea surface temperature records  
383 [Lawrence *et al.*, 2010; McClymont *et al.*, 2013], suggesting that this long-term secular cooling  
384 trend is not limited to the deep ocean. Although this trend is observed in a number of surface  
385 locations in the N. Atlantic, the BWT cooling trend observed at Site 607 could also be due to  
386 cold, southern component waters gradually penetrating further into the N. Atlantic over the MPT.  
387  $\delta^{13}\text{C}_{\text{benthic}}$  records from deep North Atlantic locations show gradually decreasing values starting  
388 at  $\sim 1.1$  Ma [Raymo *et al.*, 1990; 1997; 2004; Ferretti *et al.*, 2005; Lisiecki, 2014; Poirier and  
389 Billups, 2014] consistent with an increasing contribution of an isotopically light, southern-  
390 sourced water mass. It is likely that the BWT cooling observed at Site 607 reflects both cooling  
391 in the deep ocean and changing deep ocean circulation, such as increasing influence of southern  
392 component waters [Sosdian and Rosenthal, 2009] as shown in Figure 7.

393

#### 394 ***4.3.3 Long-term and Threshold Climate Change across the Mid-Pleistocene Transition***

395 It is clear from the gradual BWT cooling at Site 607 as well as the abrupt changes at Site  
396 1123 that the MPT exhibited both gradual and threshold-like responses to climate forcing.  
397 Gradually changing environmental conditions (summarized by *McClymont et al.*, [2013]),  
398 including N. Atlantic BWT and mid- to high-latitude SST cooling [this study, *Sosdian and*  
399 *Rosenthal*, 2009; *Lawrence et al.*, 2010; *McClymont et al.*, 2013], expanded sea-ice coverage  
400 [*Kemp et al.*, 2010], equatorward polar front migration [*McClymont et al.*, 2008; *Lawrence et al.*,  
401 2010; *Martinez-Garcia et al.*, 2010], decreased deep ocean ventilation [*Hall et al.*, 2001; *Venuti*  
402 *et al.*, 2007], intensified ocean/atmospheric circulation [*de Garidel-Thoron et al.*, 2005;  
403 *McClymont and Rosell-Mele*, 2005], and increased ice-sheet size [*Clark et al.*, 2006] may have  
404 altered climate boundary conditions in a way that shifted the cryosphere across a threshold that  
405 permanently altered climate–ice sheet response to orbital forcing. Centered on this long-term  
406 trend, the LR04  $\delta^{18}\text{O}_{\text{benthic}}$  stack shows a  $\sim 0.15\%$  increase at MIS 22 ( $\sim 900$  ka, Figure 5), likely  
407 related to ice volume growth, and a statistically significant emergence of the 100 kyr cycle  
408 [*Maasch*, 1988; *Mudelsee and Statterger*, 1997; *Clark et al.*, 2006; *Raymo et al.*, 2006; *Lisiecki*  
409 *and Raymo*, 2007; *McClymont et al.*, 2013]. As southern Hemisphere summer insolation was  
410 anomalously low during the MIS 24 to 22 interval, this ice volume growth may have occurred in  
411 Antarctica (as suggested by [*Raymo et al.*, 2006] and *Elderfield et al.* [2102]). However, we  
412 suggest the ice volume growth is not to the magnitude implied by record at Site 1123 ( $\sim 0.3\%$   
413  $\delta^{18}\text{O}_{\text{seawater}}$ ) due to possible local hydrographic effects as discussed above. Additional globally  
414 distributed, high-resolution  $\delta^{18}\text{O}_{\text{seawater}}$  records are needed to fully deconvolve the influence of  
415 ice volume and circulation changes during the MIS 24 to MIS 22 transition.

416 As with the glacial cycles that characterize the late Pleistocene [*Shakun et al.*, 2015], ice  
417 sheets may be a largely responding to changes in global temperatures and atmospheric  $\text{CO}_2$  on

418 glacial-interglacial time scales and these ice sheets may have had a threshold response to global  
419 cooling during the MPT. For example, ice volume may have switched from a 41 kyr to a 100 kyr  
420 orbital signal because global SSTs underwent a similar transition during the MPT: SSTs are  
421 dominated by 41 kyr cycles during the early Pleistocene [*Herbert et al.*, 2010; *McClymont et al.*,  
422 2013] and the 100 kyr cycle slowly emerges at ~1.2 Ma [*McClymont et al.*, 2013] and fully  
423 dominates at ~900 ka [*Medina-Elizalde and Lea*, 2005; *McClymont et al.*, 2013]. However,  
424 invoking a 41 kyr to 100 kyr transition in SST change as the mechanism for forcing the 41 kyr to  
425 100 kyr transition in  $\delta^{18}\text{O}_{\text{benthic}}$  records just moves the question of “why” to another proxy.

426 It is possible that thermohaline circulation reorganization [*Pena and Goldstein*, 2014] and  
427 the increased influence of the southern ocean in deep water circulation [*Raymo et al.*, 1997;  
428 2004; *Ferretti et al.*, 2005; *Lisiecki*, 2014; *Poirier and Billups*, 2014] may have altered heat and  
429 salt transport and carbon cycling in the deep ocean. At ~900 ka, the BWT and  $\delta^{18}\text{O}_{\text{seawater}}$  records  
430 at Site 607 and 1123 converge (Figure 5) suggesting enhanced connectivity between the Atlantic  
431 and Pacific ocean basins. High-resolution atmospheric  $\text{CO}_2$  estimates and records of deep ocean  
432 chemistry are necessary to fully understand the links between temperature, ocean circulation,  
433  $\text{CO}_2$ , ice volume during the MPT.

434

## 435 5. Summary

436 We have reconstructed a Mg/Ca-derived BWT record using *Uvigerina* spp. from DSDP  
437 Site 607 that shows an ~2-3 °C cooling from ~1500 to 500 kyrs across the MPT. This gradual  
438 cooling trend is consistent with a previously published BWT record based on *C. wuellerstorfi*  
439 and *O. umbonatus* [*Sosdian and Rosenthal*, 2009] implying that Mg/Ca records based on  
440 epifaunal benthic species at Site 607 during the MPT are robust. We hypothesize that

441 thermohaline reorganization at starting at MIS 24 and intensifying at MIS 22 likely influenced  
 442 the local hydrography at ODP Site 1123, consequently explaining the unusually large decrease in  
 443  $\delta^{18}\text{O}_{\text{seawater}}$  at that site. Gradual N. Atlantic BWT and mid- to high-latitude sea surface  
 444 temperature cooling likely pushed the climate system across a threshold starting at MIS 24. A  
 445 modest increase in ice volume at ~900 ka, possibly in Antarctica, combined with thermohaline  
 446 reorganization, created a new climate regime with enhanced sensitivity to the 100 kyr orbital  
 447 cycle.

448  
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456  
 457  
 458 **Table 1. F-test statistics comparing Mg/Ca values of *C. wuellerstorfi* and *O. umbonatus***  
 459 **and *Uvigerina* spp. between the 600 – 800 ka and 1000 – 1250 ka. No statistical difference in**  
 460 **Mg/Ca variability is observed prior to or after MIS 22.**

	F-test	Degrees of Freedom Numerator	Degrees of Freedom Denominator	P value
<i>C. wuellerstorfi</i>	1.5128	47	31	0.2251
<i>O. umbonatus</i>	0.7160	16	24	0.4948
<i>Uvigerina</i> spp.	1.0393	30	37	0.9027

462 **Table 2. Core Locations of Previously Published Benthic Stable Isotope Values.**

Site	Latitude	Longitude	Water Depth (m)	$\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ Reference	$\delta^{13}\text{C}$ Reference	$\delta^{18}\text{O}$ Reference
<b>Atlantic</b>						
GIK13519	5.7 N	19.9 W	2862	[Sarnthein et al., 1994]		
GeoB1034	21.7 S	5.4 E	3731	Bickert and Wefer [1996]		
GeoB1211	24.5 S	7.5 E	4089	Bickert and Wefer [1996]		
DSDP 552	56 N	23.2 W	2301	[Shackleton and Hall, 1984]		
DSDP 607	41 N	33 W	3427	[Ruddiman et al., 1989]		
ODP 664	0.1 N	23.2 W	3806	[Raymo et al., 1997]		
ODP 925	4.2 N	43.5 W	3040	[Bickert et al., 1997]		
ODP 926	3.7 N	42.9 W	3598	[Lisiecki et al., 2008]		
ODP 927	5.5 N	44.5 W	3326	[Bickert et al., 1997]		
ODP 928	5.5 N	43.7 W	4012	[Lisiecki et al., 2008]		
ODP 929	6 N	43.7 W	4369	[Bickert et al., 1997]		
ODP 980	55.5 N	14.7 W	2179	[Flower et al., 2000]		
ODP 981	55.5 N	14.7 W	2173	[Raymo et al., 2004]		
ODP 982	57.5 N	15.9 W	1146	[Venz et al., 1999; Venz and Hodell, 2002]		
ODP 983	60.4 N	23.6 W	1983	[Raymo et al., 2004]		
ODP 984	61.4 N	24.1 W	1650	[Raymo et al., 2004]		
ODP 1063	33.7 N	57.6 W	4584	[Ferretti et al., 2005; Poirier and Billups, 2014]		
ODP 1088	41.1 S	13.6 E	2081	[Hodell et al., 2003]		
ODP 1089	40.9 S	9.9 E	4621	[Hodell et al., 2001]		
ODP 1090	42.9 S	8.9 E	3699	[Venz and Hodell, 2002]		
ODP 1264	28.5 S	2.8 E	2505	[Bell et al., 2014]		
ODP 1267	28.1 S	1.7 E	4355	[Bell et al., 2014]		
IODP U1308	49.9 N	24.2 W	3871	[Hodell et al., 2008]		
IODP U1313	41 N	33 W	3426	[Ferretti et al., 2010]		
IODP U1314	56.4 N	27.9 W	2820	[Alonso-Garcia et al., 2011]		
<b>Pacific</b>						
GeoB15016	27.5 S	71.1 W	956		[Martínez-Méndez et al., 2013]	
PC72	0.1 N	139.4 W	4298			[Murray et al., 2000]
RC13110	0.1 S	95.7 W	3231	[Mix et al., 1991]		
ODP 677	1.2 N	83.7 W	3461	[Shackleton et al., 1990]		
ODP 806	0.3 N	159.4 E	2520		[Bickert et al., 1993]	
ODP 846	3.1 S	90.8 W	3307	[Mix et al., 1995a]		
ODP 849	0.2 N	110.5 W	3849	[Mix et al., 1995b]		
ODP 1020	41 N	126.4 W	3038			Liu, personal communication to L. Lisiecki [2002]
ODP 1123	41.8 S	171.5 W	3290	[Hall et al., 2001; Harris, 2002; Elderfield et al., 2012]		
ODP 1143	9.4 N	113.3 E	2772		[Cheng et al., 2004]	[Tian et al., 2002]
ODP 1146	19.5 N	116.3 E	2091	[Prell, 2003]		
ODP 1241	5.8 N	86.4 W	2027	[Lalicata and Lea, 2011]		

463

465 **Figure Captions**

466 Figure 1. Locations of sites:  $\delta^{13}\text{C}_{\text{benthic}}$  from the Atlantic (squares, Table 1),  $\delta^{18}\text{O}_{\text{benthic}}$  from the  
 467 Pacific (circles, Table 2), and DSDP Site 607 and ODP Site 1123 with Mg/Ca-derived BWT  
 468 records (stars) (A). Hydrographic transects for Site 607 (B) and Site 1123 (C). Atlantic transect  
 469 of carbonate ion concentration  $[\text{CO}_3^{2-}]$  [Key et al., 2004]. Site 607 is in area sensitive to changes  
 470 in  $[\text{CO}_3^{2-}]$  on glacial-interglacial time scales. Site 1123 is located in an area sensitive to water  
 471 mass mixing (panel modified from [Carter et al., 1999]). Figure generated using Ocean Data  
 472 View [Schlitzer] and [Troupin et al., 2012].

473

474 Figure 2. DSDP Site 607 Mg/Ca values of *Uvigerina spp.* (blue, this study) and previously  
 475 published *O. umbonatus* (pink) and *C. wuellerstorfi* (light green) [Sosdian and Rosenthal, 2009]  
 476 (A). Deep ocean chemistry indicators %CaCO<sub>3</sub> (B),  $\delta^{13}\text{C}_{\text{benthic}}$  (C) and %NCW (D).

477

478 Figure 3. Mg/Ca values converted to BWT using previously published species-specific  
 479 temperature calibrations of L2002 [Lear et al., 2002], SR2009 [Sosdian and Rosenthal, 2009],  
 480 YE2008 [Yu and Elderfield, 2008], C2015 [Cappelli et al., 2015], H2008 [Healey et al., 2008]  
 481 and E2012 [Elderfield et al., 2012].

482

483 Figure 4. All species included for a composite BWT record at Site 607 using the species-specific  
 484 temperature calibrations of [Sosdian and Rosenthal, 2009] and [Elderfield et al., 2012]. (A).  
 485 Previously published (grey) and new *Uvigerina spp.* (blue, this study)  $\delta^{18}\text{O}_{\text{seawater}}$  values (B).  
 486 Error bars on the left hand side indicate the 3 point smoothed combined error. Error for BWT

487 measurements is  $\pm 1.4^{\circ}\text{C}$  and the propagated error for  $\delta^{18}\text{O}_{\text{seawater}}$  estimates is  $\pm 0.33\text{‰}$  (see  
488 Supplemental Text).

489

490 Figure 5. Comparison of LR04 (black), ODP Site 1123 (A, orange), DSDP Site 607 (B, blue), for  
491  $\delta^{18}\text{O}_{\text{benthic}}$ , BWT (C), and  $\delta^{18}\text{O}_{\text{seawater}}$  (D). Site 607 shows a long-term cooling trend over the MPT  
492 while Site 1123 has near freezing temperatures over length of the 1.5 Myr record.

493

494 Figure 6. Previously published, high-resolution, mid to deep ocean depth  $\delta^{18}\text{O}_{\text{benthic}}$  records from  
495 the Pacific plotted in comparison to LR04. ODP Sites 1123 and 677 stand out as anomalous in  
496 comparison to most of the records from the Pacific Ocean basin and LR04.

497

498 Figure 7. Cross-section view of mean glacial  $\delta^{13}\text{C}_{\text{benthic}}$  values and %NCW for MIS 38 and MIS  
499 26 to MIS 16 for sites in the Atlantic Ocean basin. Figure generated using Ocean Data View  
500 [Schlitzer]. For completeness, the Pacific Basin is included in Supplemental Figure 2.

501

502 Figure 8. Previously published  $\delta^{18}\text{O}_{\text{benthic}}$  records from Sites 607, U1308 and 1123 plotted in  
503 comparison to LR04. ODP Sites 1123 and U1308 stand out as anomalous in comparison to Site  
504 607 and LR04.

505

506 Supplemental Figure 1. Distribution of *Uvigerina* spp. measurements on the  $\delta^{18}\text{O}_{\text{benthic}}$  record.  
507 Over 88% of the *Uvigerina* spp. samples are directly paired with  $\delta^{18}\text{O}_{\text{benthic}}$  samples.

508



509 Supplemental Figure 2. Various elemental to calcium ratios that were used to monitor  
510 contamination.

511

512 Supplemental Figure 3. Cross-section view of mean glacial  $\delta^{13}\text{C}_{\text{benthic}}$  values and %NCW for  
513 MIS 38 and MIS 26 to MIS 16 for sites in the Pacific Ocean basin. These transects were not  
514 interpolated as in Figure 5 because the Pacific basin is large and the data is more sparse than in  
515 the Atlantic. Figure generated using Ocean Data View [*Schlitzer*].

516

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