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Abstract

We present new sedimentological and environmental magnetic records spanning ~3.2–2.2 Ma, during the intensification of Northern Hemisphere glaciation, from North Atlantic Integrated Ocean Drilling Program Site U1307 on Eirik Drift. Our new datasets and their high-fidelity age control demonstrate that while inland glaciers – and potentially also at times restricted marine-terminating ice-caps – have likely existed on southern Greenland since at least ~3.2 Ma, persistent and extensive marine-terminating glacial margins were only established in this region at 2.72 Ma, ~300 kyr later than in northeastern and eastern Greenland. Despite a dramatic increase in Greenland-sourced ice-rafted debris deposition on Eirik Drift at this time, contemporaneous changes in the bulk magnetic properties of Site U1307 sediments, and a reduction in sediment accumulation rates, suggest a decrease in the delivery of Greenland-sourced glaciofluvial silt, which we attribute to a shift in depositional regime from bottom-current-dominated to glacial-IRD-dominated between ~2.9–2.7 Ma in response to a change in the depth of the flow path of the Western Boundary Undercurrent relative to our study site.

Keywords Plio-Pleistocene transition; Glaciation; Paleoclimatology; Paleoceanography; Paleomagnetism; Greenland; North Atlantic; Ice-rafted debris; Relative paleointensity

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Blake-Mizen *et al.* Southern Greenland glaciation and Western Boundary Undercurrent evolution recorded on Eirik Drift during the late Pliocene onset of Northern Hemisphere glaciation

Highlights

- New high-fidelity RPI-based age model for Site U1307
- Orbital-resolution history of Greenland IRD deposition on Eirik Drift during iNHG
- Persistent marine-terminating margins established in southern Greenland from 2.7 Ma
- Potentially ~300 ka later than in northeastern and eastern Greenland
- Major reduction in flow speed of bottom currents bathing Site U1307 during iNHG

1 **Southern Greenland glaciation and Western Boundary Undercurrent**
2 **evolution recorded on Eirik Drift during the late Pliocene intensification of**
3 **Northern Hemisphere glaciation**

4
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16
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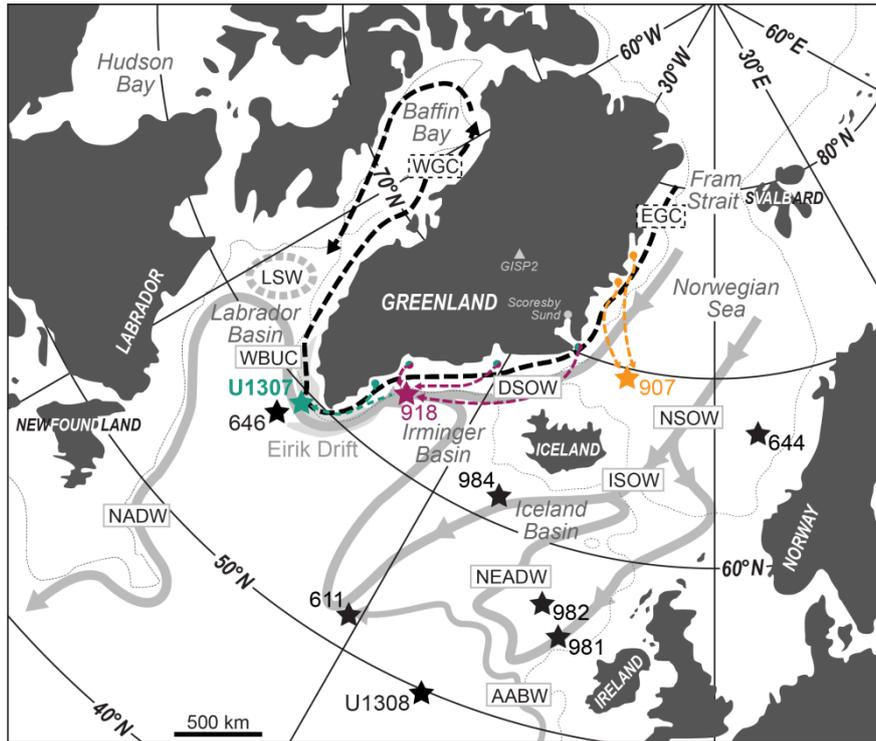
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34

35 **1. Introduction**

36 While glaciers have existed intermittently on Greenland since at least the late Eocene, with its
37 first ice-shelf glaciations occurring possibly as early as the Miocene (Larsen *et al.*, 1994),
38 multiple lines of evidence (e.g., Larsen *et al.*, 1994; Jansen *et al.*, 2000; Thiede *et al.*, 2011;
39 Bierman *et al.*, 2016) suggest that a major ice-sheet was not a persistent feature on Greenland
40 until the late Pliocene to earliest Pleistocene intensification of Northern Hemisphere glaciation
41 (iNHG), ~3.6–2.4 Ma (Mudelsee and Raymo, 2005). Our understanding of the spatial history
42 of ice-sheet expansion on Greenland at this time, however, remains uncertain.

43



44

45 **Figure 1.** Map of Greenland and surrounding landmasses and ocean basins, showing the
 46 location of Site U1307 on Eirik Drift, the position of sites mentioned in this text, and their
 47 relationships to the paths of key modern deep (solid grey) and surface (dashed grey) ocean
 48 currents relevant to this study. ODP Sites 907 and 918, whose ice-rafting records are discussed
 49 in the text, are also highlighted, and main sources of ice-rafted debris to these and U1307 are
 50 schematically represented by colour-coded arrows (based on iceberg trajectory simulations for
 51 the Last Glacial Maximum by Bigg *et al.*, 1998). NSOW = Norwegian Sea Water, ISOW =
 52 Iceland-Scotland Overflow Water, DSOW = Denmark-Scotland Overflow Water, NEADW =
 53 Northeast Atlantic Deep Water, AABW = Antarctic Bottom Water, WBUC = Western
 54 Boundary Undercurrent, LSW = Labrador Sea Water, NADW = North Atlantic Deep Water.
 55 The 1000 m isobath is given by thin dotted lines.

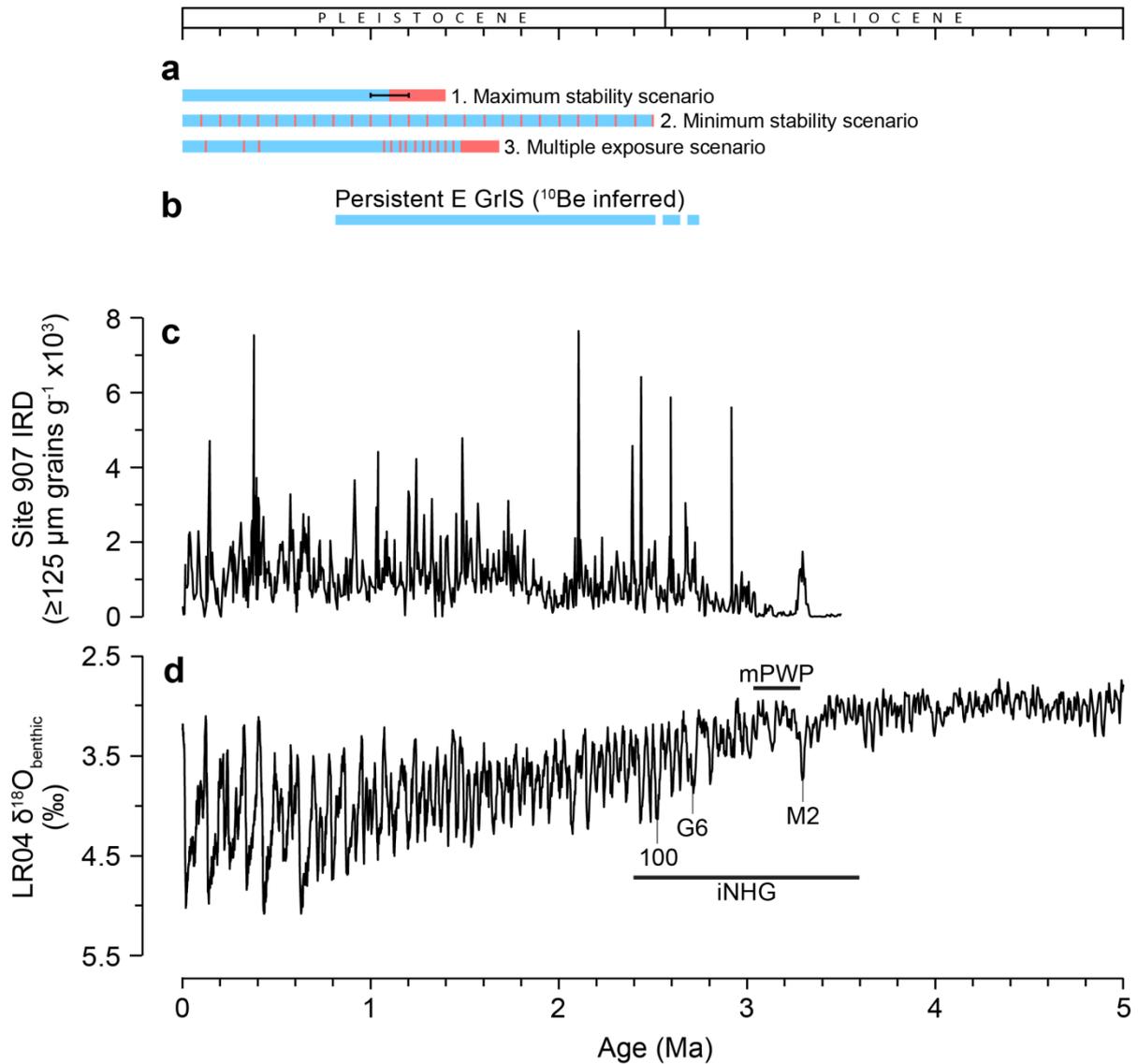
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57 Since terrestrial evidence of Greenland glaciation during iNHG is rare (and its temporal
 58 interpretation non-unique, e.g., Schaefer *et al.*, 2016) due to its removal by subsequent glacial
 59 advances, our understanding of Greenland Ice Sheet (GrIS) evolution during the Plio-
 60 Pleistocene mainly relies on Greenland-proximal marine sediment records of ice-rafted debris
 61 (IRD; e.g., Jansen *et al.*, 2000; Thiede *et al.*, 2011). Greenland-proximal IRD deposition is
 62 dominated by terrigenous sediment transported mainly by the East Greenland Current (EGC)

63 in icebergs derived from multiple GrIS iceberg-calving sources (Fig. 1; Bigg *et al.*, 1998; White
64 *et al.*, 2016). Spatial comparisons of orbitally-resolved paleo-records of marine IRD deposition
65 near Greenland can therefore provide important insights into where and when the GrIS
66 extended to the coast during iNHG. For this time period, IRD records are available from sites
67 where ice-rafted sediments were sourced from northeastern (ODP Site 907; Jansen *et al.*, 2000),
68 eastern (ODP Site 907; Jansen *et al.*, 2000; ODP Sites 914–918; Larsen *et al.*, 1994; St John
69 and Krissek, 2002) and southern (ODP Site 646; Wolf and Thiede, 1991) Greenland (Figs. 1
70 and 2). However, due to low benthic foraminifera abundances, it is not possible to generate
71 independent orbital-resolution benthic $\delta^{18}\text{O}$ stratigraphies for any of these Greenland-proximal
72 records spanning iNHG. Poor core recovery has also prevented the construction of detailed
73 paleomagnetic stratigraphies for most of these sites.

74 The only continuous orbital-resolution Greenland-proximal IRD record with a
75 complete palaeomagnetic reversal stratigraphy for iNHG comes from Site 907 on the Iceland
76 Plateau (Figs. 1 and 2c; Jansen *et al.*, 2000). Based on elevated IRD deposition at this site from
77 ~ 3 Ma (Fig. 2c-d), we can infer that at least isolated marine-calving glaciers occupied coastal
78 northeastern and eastern Greenland on orbital timescales following the end of the mid-
79 Piacenzian warm period (mPWP, 3.264–3.025 Ma; Dolan *et al.*, 2011) (see iceberg trajectories
80 shown in Fig. 1; Bigg *et al.*, 1998). A further and significant increase in IRD deposition at Site
81 907 from ~ 2.7 Ma indicates, however, that extensive marine-calving margins may not have
82 been established in this region of Greenland until Marine Isotope Stage (MIS) G6, 2.72 Ma
83 (Lisiecki and Raymo, 2005) (Fig. 2c–d). This suggestion is broadly supported by seismic
84 evidence offshore Scoresby Sund that confirms the central-eastern GrIS was only frequently
85 grounded below sea-level from ~ 2.6 Ma (Vanneste *et al.*, 1995), and by a recent study of the
86 cosmogenic radionuclide (^{10}Be and ^{26}Al) isotope composition of quartz sands from ODP Sites
87 918 and 987 (Figs. 1 and 2b; Bierman *et al.*, 2016) that infers the existence of a large ice-cap

88 on at least eastern Greenland since the onset of the Quaternary. Our ability to improve our
 89 understanding of the wider regional history of GrIS growth during this time is hampered,
 90 however, by a lack of a well-dated, orbital-resolution continuous IRD record that is well-placed
 91 to receive detritus from southeastern and southern Greenland during iNHG.
 92



93
 94 **Figure 2.** Synthesis of paleoclimatic records relevant to the Plio-Pleistocene history of
 95 Greenland Ice Sheet (GrIS) evolution: (a) three scenarios for GrIS glaciation consistent with
 96 cosmogenic beryllium (^{10}Be) and aluminium (^{26}Al) isotopes in bedrock from the base of the
 97 GISP2 ice core (see Fig. 1 for location) (1. Maximum stability scenario = 280 ± 30 kyr ice-free
 98 conditions followed by 1.1 Ma continuous ice cover, 2. Minimum stability scenario = ice-free
 99 for 8 kyr of each 100-kyr cycle, 3. Multiple exposure scenario = ice-free for several thousand

100 years during numerous major Pleistocene interglacials; Schaefer *et al.*, 2016); **(b)** periods of
101 eastern GrIS growth and stability inferred from seismic profiles offshore Scoresby Sund
102 indicating East (E) GrIS frequently grounded below sea-level (Vanneste *et al.*, 1995) and from
103 cosmogenic ^{10}Be and ^{26}Al isotopes in marine cores offshore eastern Greenland (Bierman *et al.*,
104 2016); alongside **(c)** Iceland Basin ODP Site 907 IRD (Jansen *et al.*, 2000); and **(d)** the LR04
105 $\delta^{18}\text{O}_{\text{benthic}}$ stack for reference (Lisiecki and Raymo, 2005). Labels in (d) are Marine Isotope
106 Stages, and durations of the intensification of Northern Hemisphere glaciation (iNHG) and the
107 mid-Piacenzian warm period (mPWP; Dolan *et al.*, 2011) are also indicated.

108

109 Eirik Drift off southern Greenland is ideally located to monitor the history of the southern GrIS
110 because it lies in the path of the EGC, and in the present day receives IRD from a range of
111 eastern and southern Greenland iceberg-calving sources (Bigg *et al.*, 1996; White *et al.*, 2016).
112 It is also well-positioned to monitor the strength of the Western Boundary Undercurrent
113 (WBUC; Hunter *et al.*, 2007), the behaviour of which during iNHG is not fully understood
114 (Hunter *et al.*, 2007; Müller-Michaelis and Uenzelmann-Neben, 2014; Parnell-Turner *et al.*,
115 2015). Studies that combine analysis of both the sedimentological character and environmental
116 magnetic signature of Pleistocene sediments from Eirik Drift have proved particularly useful
117 in this regard, because they can be utilised to reconstruct relative changes in bottom-current
118 strength and the provenance of glacially-derived terrigenous sediments from southern
119 Greenland that they entrain (e.g., Evans *et al.*, 2007; Stoner *et al.*, 1995; Hatfield *et al.*, 2016;
120 2017; Channell *et al.*, 2014). However, these techniques have yet to be applied to records
121 spanning the iNHG interval.

122 To further our understanding of GrIS and WBUC behaviour during iNHG, we present
123 high-resolution IRD, paleomagnetic and environmental magnetic records spanning ~3.2 to 2.2
124 Ma from Integrated Ocean Drilling Program (IODP) Site U1307, situated on Eirik Drift in the
125 northern North Atlantic Ocean (Fig. 1). Our study benefits from updated chronological control
126 through the generation of a new relative paleointensity (RPI)-based age model – the first of its

127 kind for high-latitude sediments deposited during iNHG. On this high-fidelity age model, our
128 new high-resolution multi-proxy records demonstrate for the first time that glacial maturation
129 of southern Greenland ~2.7 Ma occurred in concert with a change in WBUC behaviour.

130

131 **2. Study Site and Methods**

132 ***2.1 Study site and sampling***

133 Eirik Drift is an extensive, elongate contourite drift located just south of Greenland (Fig. 1)
134 that began to form in the Miocene (Hunter *et al.*, 2007; Müller-Michaelis and Uenzelmann-
135 Neben, 2014). It lies underneath the path of the EGC, a southward-flowing surface current that
136 transports icebergs calved from outlet glaciers along Greenland's eastern coast. Eirik Drift is
137 also sculpted directly by the vigorous and dynamic deep WBUC, the main axis of which shoals
138 during late Pleistocene glacials (Hillaire-Marcel *et al.*, 1994; Hillaire-Marcel and Bilodeau,
139 2000; Mazaud *et al.*, 2015). The EGC and WBUC constitute major transport pathways for
140 delivering detrital sediments glacially eroded on Greenland to the adjacent marine margins,
141 and ultimately to Eirik Drift (Hunter *et al.*, 2007). Icebergs transported along the EGC can
142 contain clay- to boulder-sized IRD from a range of marine-terminating eastern and southern
143 Greenland iceberg-calving sources, which rains out to the seabed as they melt (Fig. 1; White
144 *et al.*, 2016). The WBUC can entrain these sediments and mix them with subglacially-eroded
145 southern Greenland Precambrian terrane bedrock clay- and silt-sized grains, as well as volcanic
146 detritus from East Greenland and Iceland (Stoner *et al.*, 1995; Carlson *et al.*, 2008; Colville *et*
147 *al.*, 2011; Hatfield *et al.*, 2016).

148 IODP Site U1307 is located on the northern flank of Eirik Drift (Fig 1; 58°30.3'N,
149 46°24.0'W, 2575 m water depth). Two holes were drilled at Site U1307 (U1307A and U1307B)
150 during IODP Expedition 303 in 2004, which together recovered a ~175 m composite Plio-
151 Pleistocene section (Expedition 303 Scientists, 2006a). To examine the evolution of the

152 southern GrIS and WBUC during iNHG, Site U1307 cores were sampled with u-channels
153 (typically $2 \times 2 \times 150$ cm continuous samples; Weeks *et al.*, 1993) and discrete 20 cc scoops
154 at 10 cm intervals at the MARUM IODP Core Repository in Bremen, Germany. Sampling was
155 guided by the shipboard-derived paleomagnetic record and the primary splice (Expedition 303
156 Scientists, 2006a) between ~113–148 metres below seafloor (mbsf) in U1307A, and ~110–136
157 mbsf in U1307B.

158

159 **2.2 Methods**

160 *2.2.1 Paleo- and environmental magnetism*

161 The natural remanent magnetisation (NRM) and anhysteretic remanent magnetisation (ARM)
162 of each u-channel sample were measured at 1-cm intervals using a 2G Enterprises™ model
163 755-1.65UC superconducting rock magnetometer at the Paleo- and Environmental Magnetism
164 Laboratory at Oregon State University (OSU). NRM was measured following inline stepwise
165 alternating field (AF) demagnetisation at peak AF from 20 to 80 mT. Additional steps up to
166 peak AF of 100 mT were applied to nine u-channels known from shipboard paleomagnetic
167 measurements to contain a polarity reversal (Expedition 303 Scientists, 2006a). Component
168 paleomagnetic directions used to define the characteristic remanent magnetisation (ChRM)
169 directions, and maximum angular deviation (MAD) used to assess the quality of the component
170 magnetisation estimates, were calculated from principal component analysis of 7 equally-
171 spaced demagnetisation steps over 20–50 mT following Kirschvink (1980), using the UPmag
172 software of Xuan and Channell (2009). Guided by the inclination data, declination values were
173 rotated to a mean of 0° (180°) for periods of normal (reversed) polarity on a core-by-core basis.

174 ARM was acquired inline using a 100 mT peak AF and a 0.05 mT direct current (DC)
175 bias field along each u-channel's long-axis. The ARM was remeasured after each 5–10 mT
176 increment of AF demagnetisation in the 10–60 mT peak field range. Low-field volume-

177 normalised (bulk) magnetic susceptibility (κ) was measured every 1cm for each u-channel at
178 OSU's Marine Geology Repository using a Bartington MS2C 36 mm diameter loop sensor
179 mounted on a software motion-controlled track. ARM and κ both reflect the concentration of
180 ferrimagnetic (titanomagnetite, magnetite) grains in a sample, although ARM is more sensitive
181 to the fine ferrimagnetic fraction (King *et al.*, 1983).

182 The susceptibility of ARM (κ_{ARM}) was determined by normalising ARM by the DC
183 field applied during ARM acquisition. The ratio of κ_{ARM} over the low-field bulk magnetic
184 susceptibility, $\kappa_{\text{ARM}}/\kappa$, is commonly used to track variations in ferrimagnetic grain-size. Low
185 (high) $\kappa_{\text{ARM}}/\kappa$ values imply a coarser (finer) average ferrimagnetic grain-size (King *et al.*, 1983;
186 Bloemendal *et al.* 1992), and this parameter has been used in Pleistocene North Atlantic studies
187 to monitor variations in deep-water circulation and inputs of Greenlandic detritus (e.g., Stoner
188 *et al.*, 1995; Evans *et al.*, 2007; Mazaud *et al.*, 2012; Channell *et al.*, 2014).

189 The magnetic signature of Eirik Drift is sensitive to deposition of glaciofluvial silt-size
190 sediments subglacially eroded from Archean and Paleoproterozoic felsic crystalline bedrock
191 by the GrIS, and Cenozoic volcanics from Iceland and eastern Greenland (e.g., Colville *et al.*,
192 2011). These distinct sources can be discriminated using radiogenic (Colville *et al.*, 2011) and
193 magnetic (Hatfield *et al.*, 2013; 2017) properties measured on a particle-size specific basis.
194 Clay-sized terrestrial fractions (defined here as $\leq 3 \mu\text{m}$) from both Greenland and Iceland are
195 characterised by low concentrations of ferrimagnetic minerals (low κ) and relatively fine
196 ferrimagnetic grain-sizes (high $\kappa_{\text{ARM}}/\kappa$). In contrast, silt (and sand) size fractions from both
197 sources have up to an order of magnitude higher magnetic susceptibility (Hatfield *et al.*, 2017),
198 indicating higher concentrations of ferrimagnetic minerals in the larger size fractions. Silts and
199 sands from Iceland are dominated by magnetically fine-grained titanomagnetite inclusions
200 (yielding higher $\kappa_{\text{ARM}}/\kappa$ values), which can be discriminated from Greenland-derived silts and
201 sands that are dominated by coarser discrete magnetites (yielding lower $\kappa_{\text{ARM}}/\kappa$ values)

202 (Hatfield *et al.*, 2013; 2017). While Eirik Drift bulk sediments are an aggregate of
203 magnetically-fine silts and sands from Iceland and magnetically-fine clays that can originate
204 from multiple sources, significant source-driven coarsening of the bulk magnetic grain-size
205 record can only be driven by accumulation of Greenland-derived silt and sand, which can be
206 linked to changes in GrIS dynamics (Colville *et al.*, 2011; Hatfield *et al.*, 2016).

207

208 *2.2.2 Shipboard splice revision*

209 To improve the continuity of the iNHG record from Site U1307, we used our new high-
210 resolution u-channel κ data to revise the shipboard splice for our target interval. Our revised
211 splice for ~117–176 revised metres composite depth, rmcd (see Tables 1 and S1 and Fig. S1)
212 uses, but refines, tie points utilised in the shipboard splice (Expedition 303 Scientists, 2006a;
213 see also Fig. S2 of Supplementary Material).

214

215 *2.2.3 Relative paleointensity (RPI) based age model*

216 Benthic foraminifera abundances are low in Pliocene-aged Eirik Drift sediments (Expedition
217 303 Scientists, 2006a). It is therefore difficult to generate a benthic $\delta^{18}\text{O}$ -based age model for
218 Site U1307, and planktic $\delta^{18}\text{O}$ -derived records from this region can be influenced by freshwater
219 inputs from ice melt (Hillaire-Marcel *et al.*, 1994). To circumvent these problems, we generated
220 a reversal- and a relative paleointensity (RPI)-based magnetostratigraphy using our u-channel
221 NRM and ARM data. Previous paleomagnetic studies of Eirik Drift sediments (Stoner *et al.*,
222 1995; Evans *et al.*, 2007; Mazaud *et al.*, 2012; Channell *et al.*, 2014) and of Site U1307 cores
223 in particular (Kawamura *et al.*, 2012; Mazaud *et al.*, 2015) show that the magnetic assemblage
224 is dominated by (titano)magnetite with a relatively uniform magnetic grain-size in the pseudo-
225 single domain (PSD) range (Fig. S4b), and that RPI is a useful tool to provide chronological
226 control.

227 Since NRM intensity is sensitive to the strength of Earth's magnetic field upon, or
228 shortly after, sediment deposition, and ferrimagnetic grain concentration, we used ARM to
229 normalise NRM intensity to compensate for variations in magnetic concentration (King *et al.*,
230 1983). RPI was estimated using the slope of NRM and ARM values over the 20–50 mT peak
231 AF demagnetisation interval (Channell *et al.*, 2002). To generate a U1307 RPI-based age
232 model, we initially anchored the stratigraphy at the polarity reversal boundaries, then tuned the
233 RPI record within periods of stable polarity to the RPI record from IODP Site U1308 (Channell
234 *et al.*, 2016). The Site U1308 RPI record was chosen as a tuning target because it spans the past
235 ~3.15 Ma, has a high-quality magnetic record and orbital-resolution benthic $\delta^{18}\text{O}$ -based
236 chronology tied to the LR04 stack, and serves as the anchor record for the PISO stack (Channell
237 *et al.*, 2009; 2016). To provide a magnetic stratigraphy beyond the u-channelled interval, we
238 constructed inclination and RPI proxy records using lower (5 cm)-resolution shipboard-
239 acquired data for both U1307 (down to the maximum drilled depth) and U1308 (to the end of
240 Hole C) – deriving a RPI estimate from NRM intensity (demagnetised in peak AF of 20 mT)
241 normalised by bulk magnetic susceptibility, κ (see Section B of Supplementary Material and
242 Fig. S3). This estimate of RPI is not as robust as our u-channel-based estimate, partly due to
243 imperfect normalisation of NRM by κ , but its use here is justified by the similarity of this record
244 to our u-channel-derived RPI estimates where overlap exists (Fig. S3). Importantly, it allows
245 us to extend our observations through the top of the Mammoth subchron (3.207 Ma), which
246 improves our age model validation.

247

248 2.2.4 Ice-rafted debris estimates

249 To examine the history of iceberg-rafting to Site U1307 during iNHG, we generated a record
250 of weight percent (wt.%) IRD for the $\geq 212 \mu\text{m}$ size fraction. Each discrete sample was dried
251 in an oven to determine its dry bulk weight, then disaggregated in a Calgon solution and washed

252 over a 63 μm sieve to isolate the sand-sized fraction. The percentage of ice-rafted terrigenous
253 material in the $\geq 212 \mu\text{m}$ fraction of each sample (following further sieving) was estimated
254 using a standard method (e.g., St. John and Kriisek, 2002). Sand-sized constituents excluded
255 from our definition of ‘ice-rafted debris’ were volcanic tephra, pyrite, biogenics (mostly
256 diatoms and foraminifers) and burrow casts. Weight percent IRD was then estimated for each
257 sample using the $\geq 212 \mu\text{m}$ fraction weight as a percentage of the dry bulk sediment weight. To
258 compare the history of terrigenous sand inputs at Site U1307 to those previously published
259 from Eirik Drift at Ocean Drilling Program (ODP) Site 646 (Fig. 1; Wolf and Thiede, 1991),
260 we also estimated the wt.% of sand-sized terrigenous constituents in the bulk $\geq 63 \mu\text{m}$ sand
261 fraction of each of our samples.

262 The Iceland Plateau Site 907 IRD record is reported in $\geq 125 \mu\text{m}$ lithic grains per gram
263 of dry sediment (Jansen *et al.*, 2000), whereas our new IRD record from Site U1307 is reported
264 as wt.% $\geq 212 \mu\text{m}$ IRD. To help compare our data to those available from Site 907, we generated
265 twenty-eight additional IRD data for Site U1307 for our study interval, expressed as lithic
266 grains (minus fresh volcanic glass and pyrite) $\geq 125 \mu\text{m}$ per gram of dry sediment, following
267 Bailey *et al.* (2013). To investigate whether changes in the size fraction used to perform IRD
268 counts have the potential to modify inferences made on the magnitude of iceberg rafting to our
269 study site, we recounted the same samples after sieving for the $\geq 150 \mu\text{m}$ size-fraction.

270

271 2.2.5 Grain-size distribution

272 To evaluate the potential influence of changes in the relative abundance of clay, silt and sand
273 deposited at Site U1307 during iNHG on our bulk magnetic records, we estimated physical
274 grain-size distributions for the terrigenous fraction using discrete 1–2 cm^3 samples for 18
275 selected points between 136.9 and 118.3 rmd – an interval containing the largest-amplitude κ

276 and ARM variations in our magnetic records (see Fig. S4a) – using a Malvern Mastersizer 3000
277 laser diffraction particle size analyser at the University of Exeter’s Penryn Campus.

278 Prior to analysis, organics and biogenics were removed from each sample (see Section
279 E of Supplementary Material). Ten repeat grain size measurements were made on a well-mixed
280 aliquot of each sample, and this procedure repeated on a separate subsample. The values
281 reported for each sample are an average of all measurements made. Following Hatfield *et al.*
282 (2013; 2017), size fractions are defined here as: clay ≤ 3 μm , very fine silt 3–10 μm , fine–
283 medium silt 10–32 μm , medium–coarse silt 32–63 μm , and sand >63 μm .

284

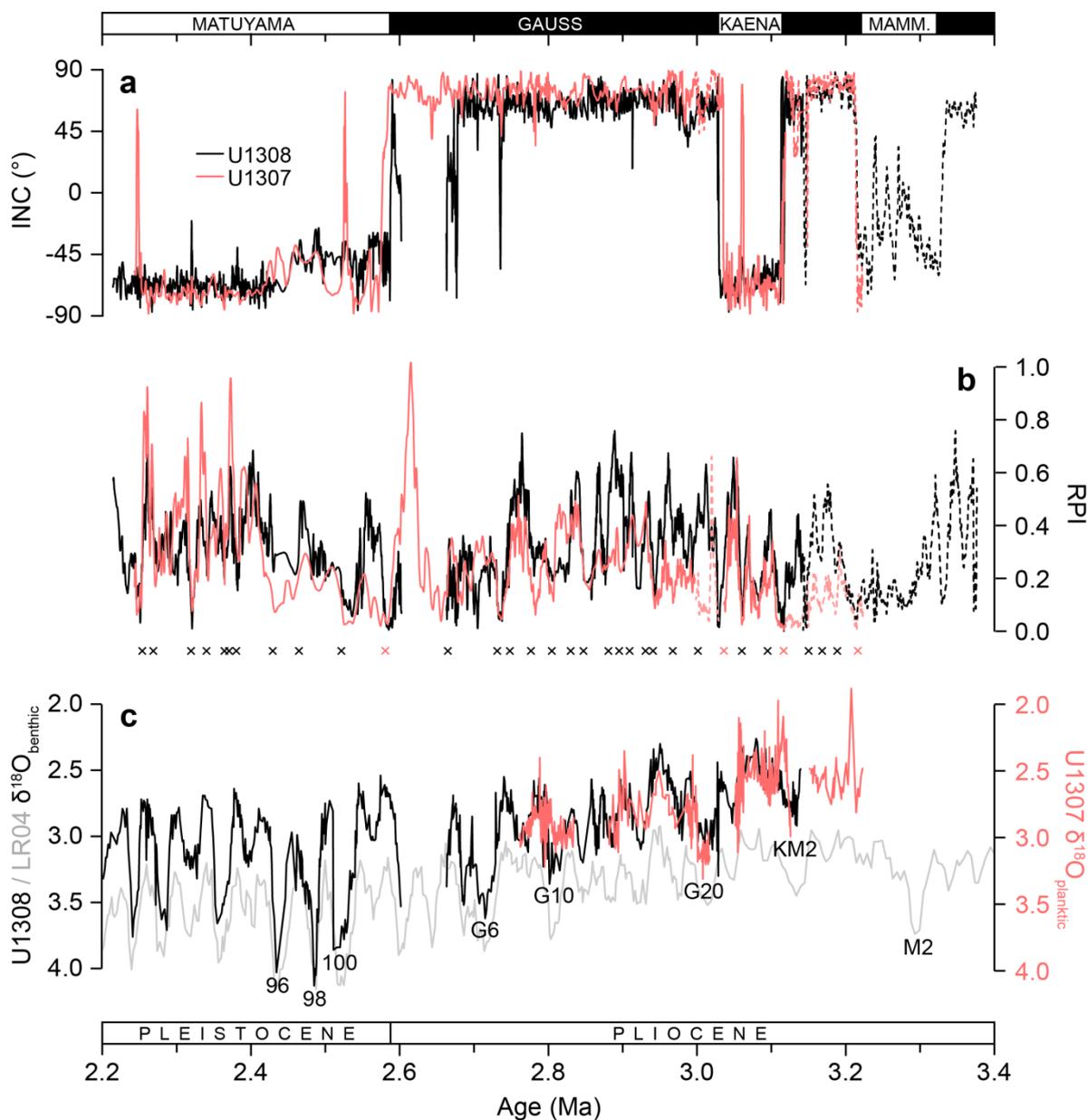
285 **3. Results and Discussion**

286 ***3.1 New RPI-based age model for Eirik Drift sediments deposited during iNHG***

287 The results of our nested magnetostratigraphic and RPI-based tuning exercise between Sites
288 U1307 and U1308 are shown in Figure 3 (see also Section C in Supplementary Information).
289 Using the Gauss-Matuyama (G/M) boundary and the top of the Kaena as initial tie points, the
290 majority of the peaks and troughs in the U1307 and U1308 RPI estimates can be correlated
291 between 3.22–2.24 Ma. The mismatch in RPI between U1307–U1308 during ~ 3.15 – 3.10 Ma
292 (Fig. 3b) can be attributed to the low concentration of ferrimagnetic minerals in U1307
293 sediments deposited during this interval indicated by the magnetic susceptibility low in this
294 site’s stratigraphy between 163.4–156.4 rmcd (Fig. S4a; Expedition 303 Scientists, 2006a).
295 The match between the two RPI records is achieved using thirty-three tie-points (see Tab. S3
296 in Supplementary Material) and an assumption of constant sedimentation rate between adjacent
297 tie-points. The regionally-coherent nature of the RPI high captured in our record immediately
298 prior to the G/M boundary, and of the trends in our U1307 record in general, are confirmed by
299 the strong correspondence between our RPI record and that from nearby Gardar Drift Site
300 U1314 between 2.7–2.2 Ma (Fig. S6; Ohno *et al.*, 2012). Key similarities also exist between

301 our RPI record and those derived from other globally-distributed sites, including ODP Leg 138
 302 (Valet and Meynadier, 1993) and the equatorial Pacific EPAPIS-3000 stack (Yamazaki and
 303 Oda, 2005) (Fig. S6). The limited planktic $\delta^{18}\text{O}$ data available for U1307 for this interval
 304 (Sarnthein *et al.*, 2009) correlate well with the U1308 benthic $\delta^{18}\text{O}$ stratigraphy (and with the
 305 LR04 stack where U1308 data is absent) on this new age model (Fig. 3c), providing
 306 independent confirmation for our RPI tuning.

307



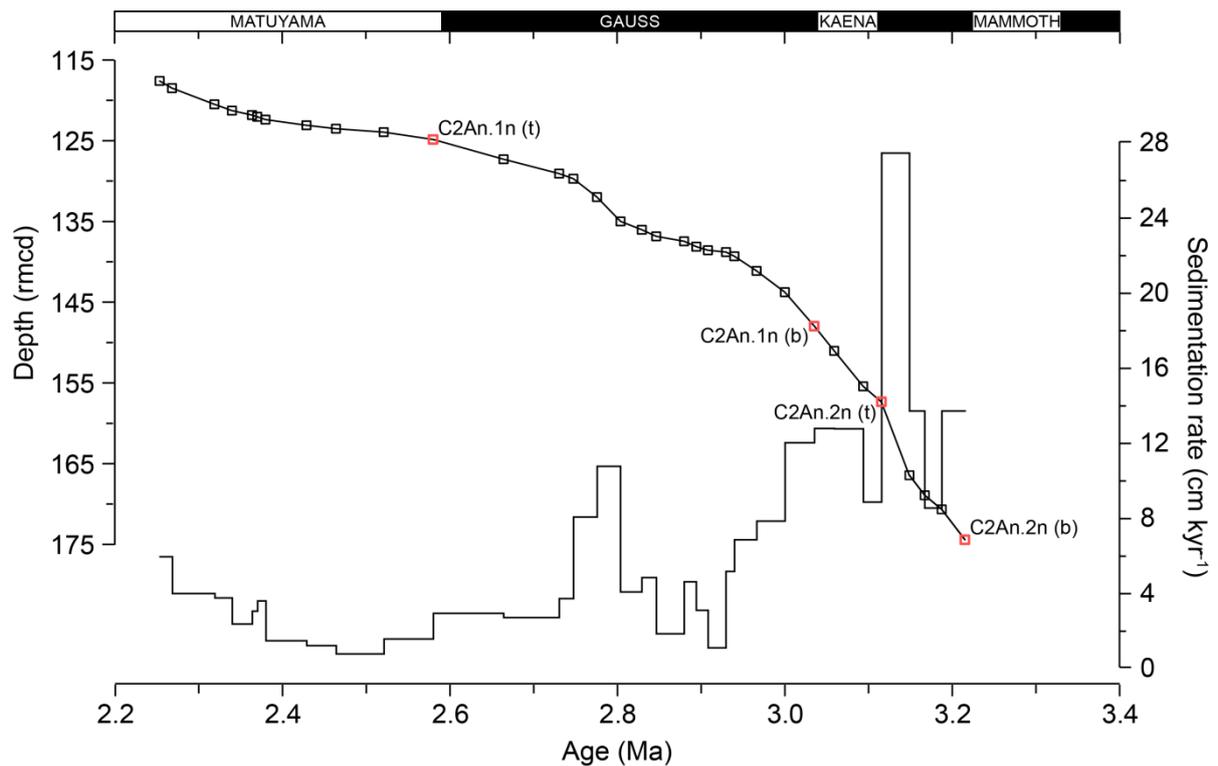
308

309 **Figure 3.** Site U1307 and U1308 records of (a) inclination (INC; red - this study; black –
310 Channell *et al.*, 2016), (b) relative paleointensity (RPI; red – this study; black – Channell *et al.*,
311 2016) and (c) $\delta^{18}\text{O}$ (red – Sarnthein *et al.*, 2009; black – Channell *et al.*, 2016; grey – Lisiecki
312 and Raymo, 2005). Solid lines show u-channel-derived data. Dashed lines show shipboard-
313 derived split core data (see Section B and Fig. S3 of Supplementary Material). Red/black
314 crosses indicate reversal-/RPI based tie-points, also given in Table S3. Labels in (c) are Marine
315 Isotope Stages. The U1308 reference RPI stratigraphy was ‘unhooked’ at 197.40 mcd because
316 a hiatus has been identified at this depth spanning MIS G2–104 (~2.65–2.60 Ma; Channell *et*
317 *al.*, 2016).

318

319 Based on our new age model, the oldest sediments recovered at Site U1307 are ~3.22 Ma and
320 the oldest observed paleomagnetic reversal is the top of the Mammoth (Fig. 3a; C2An.2r, 3.207
321 Ma; Ogg, 2012), significantly revising the shipboard-designated age of ~3.58 Ma. This new
322 chronology suggests that the Site U1307 stratigraphy contains a near-complete record of the
323 mPWP between ~175.5–146.7 rmcd. We tentatively suggest that the short normal polarity
324 excursion captured near the top of our record (Fig. 3a) is equivalent to an unnamed event
325 recorded at ODP Site 982 at ~2.24 Ma (Channell and Guyodo, 2004). Other short normal
326 polarity excursions occur at ~2.52 and ~3.06 Ma in our record (Fig. 3a), but to our knowledge
327 excursion events at these times have not been identified elsewhere.

328



329

330 **Figure 4.** Age-depth relationship for Site U1307 based on tie-points between U1307 and
 331 U1308 shown in Fig. 3 (see also Tab. S3 of Supplementary Material), and linear sedimentation
 332 rates calculated between each tie-point. Black/red squares indicate relative paleointensity-
 333 /reversal-based tie-points used.

334

335 Our new age model reveals that between ~ 3.2 and 2.2 Ma, the average sedimentation
 336 rate at Site U1307 was ~ 6 cm kyr^{-1} (Fig. 4), similar to the average Pleistocene rate of ~ 5.5 cm
 337 kyr^{-1} reported by Mazaud *et al.* (2015). The early part of the record, however, is characterised
 338 by much higher and more variable sedimentation rates than this average, which decreased from
 339 values of ~ 10 – 12 cm kyr^{-1} to ~ 2 – 4 cm kyr^{-1} between ~ 3.0 and 2.7 Ma, reaching a minimum of
 340 just ~ 1 cm kyr^{-1} at ~ 2.5 Ma before quadrupling between ~ 2.4 and ~ 2.2 Ma (Fig. 4).
 341 Sedimentation rates estimated for the deeper-water Eirik Drift Site 646 (~ 3450 m), although
 342 highly averaged over our study interval due to limited age-control, are broadly consistent with
 343 our new highly-resolved U1307 record and also hint that sedimentation rates likely decreased

344 over a wide range of water depths on Eirik Drift around ~2.7 Ma (Fig. S7b; Wolf and Thiede,
345 1991).

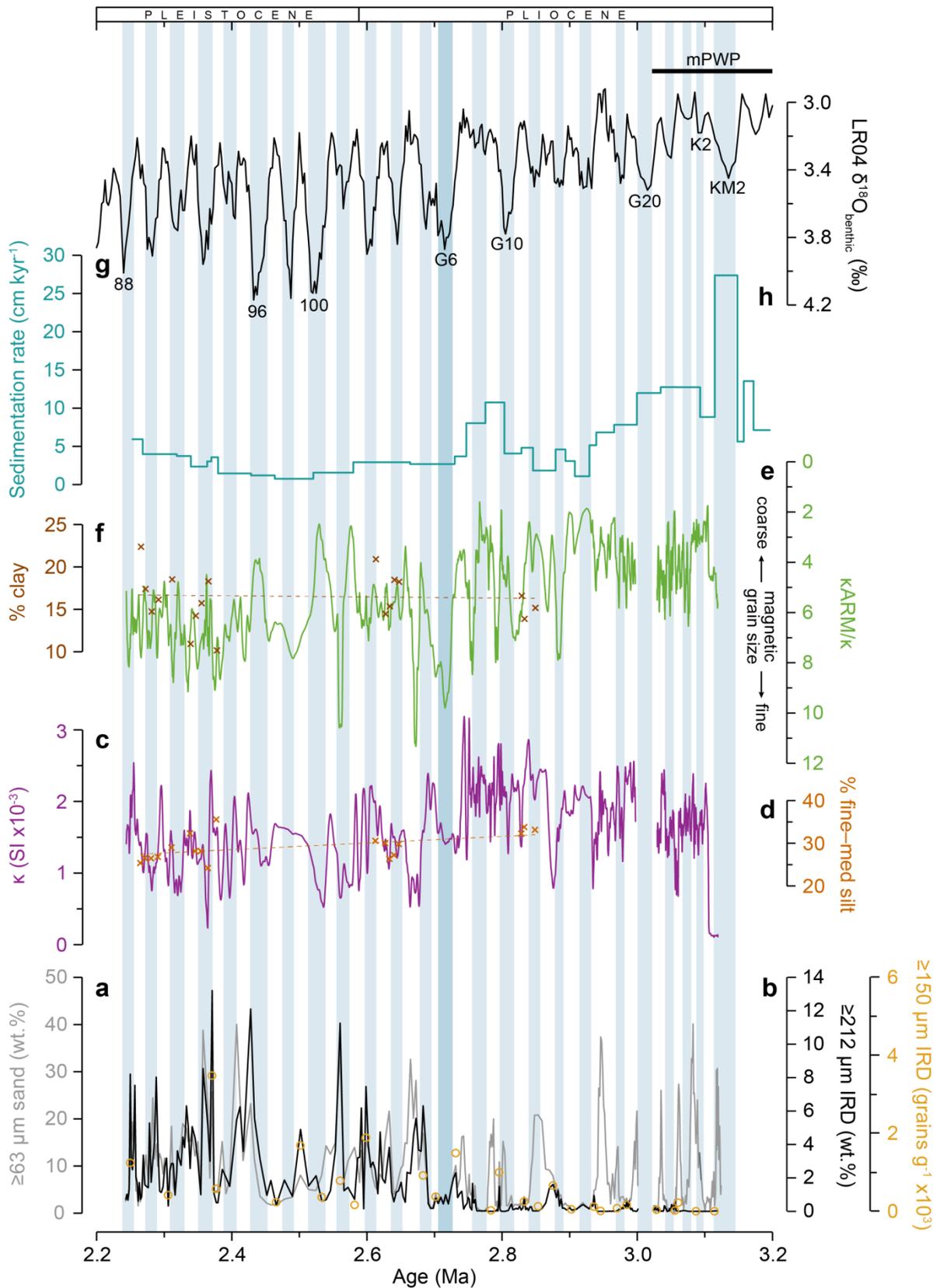
346

347 ***3.2 New records of IRD deposition and environmental magnetism on Eirik Drift during*** 348 ***iNHG***

349 Our new late Pliocene to early Pleistocene records of coarse ($\geq 212 \mu\text{m}$ and $\geq 150 \mu\text{m}$) IRD
350 abundance, wt. % $\geq 63 \mu\text{m}$ terrigenous sand, bulk volume-specific magnetic susceptibility (κ)
351 and magnetic grain-size ($\kappa_{\text{ARM}}/\kappa$) of Site U1307 sediments deposited between ~3.2–2.2 Ma are
352 presented in Figure 5. The coarse IRD deposited at U1307 during our target interval is mainly
353 composed of quartz, feldspar, mica and lithic clasts of granite, gneiss, basalt and
354 troctolite/gabbro, with minor accessory hornblende and epidote. Variability in its abundance
355 follows the same general pattern as that seen in Pleistocene and other late Pliocene records
356 (IRD elevated periodically on ~41-kyr timescales during cold stages and/or glacial
357 terminations; Fig. 5b). Coarse IRD deposition is mainly absent at U1307 during the mPWP,
358 and IRD is only present in small abundances (~1–2 wt. % in the $\geq 212 \mu\text{m}$ fraction; ~200–1000
359 grains $\geq 150 \mu\text{m g}^{-1}$) during (de)glacials between ~3–2.75 Ma. From MIS G8 onwards,
360 however, coarse IRD inputs were persistently elevated on orbital timescales (with peak
361 abundances during glacials of ~10–14 wt. % in the $\geq 212 \mu\text{m}$ fraction; ~1800–3800 grains ≥ 150
362 $\mu\text{m g}^{-1}$).

363 The abundance of $\geq 63 \mu\text{m}$ sand in U1307 sediments deposited during our study interval
364 also varies most strongly on ~41-kyr timescales, ranging from 0–40 wt. % throughout our study
365 interval (Fig. 5a). In contrast to our coarse $\geq 212 \mu\text{m}$ wt.% IRD record, however, $\geq 63 \mu\text{m}$
366 terrigenous sand deposition at Site U1307 is strongly elevated during interglacials prior to ~2.7
367 Ma that in the vast majority of cases are not associated with coarse IRD deposition (Fig. 5a–
368 b). Sand deposited during these intervals is composed of mostly well-sorted, fine, rounded

369 quartz sand grains, and contains a significant but variable (~10–70%) biogenic component of
370 both foraminifera and diatom tests. Following the onset of persistently elevated coarse IRD
371 deposition ~2.7 Ma, however, the $\geq 63 \mu\text{m}$ terrigenous sand fraction is composed of mostly
372 (sub)angular grains that are heterogeneous in composition with at most a rare biogenic
373 component, and variations in its abundance closely follow changes in wt. % $\geq 212 \mu\text{m}$ IRD.



374

375 **Figure 5.** Site U1307 paleoclimate records: (a) wt.% $\geq 63 \mu\text{m}$ terrigenous sediment, (b) wt.%
 376 $\geq 212 \mu\text{m}$ IRD, (c) magnetic susceptibility (κ), (d) % fine-medium silt ($10\text{--}32 \mu\text{m}$), (e) $\kappa_{\text{ARM}}/\kappa$,
 377 (f) % clay ($\leq 3 \mu\text{m}$) and (g) sedimentation rate. The LR04 benthic $\delta^{18}\text{O}$ stack is shown in (h)

378 for reference (Lisiecki and Raymo, 2005). Numbers in (h) are Marine Isotope Stages, with the
379 duration of the mid-Piacenzian warm period (mPWP; Dolan *et al.*, 2011) also indicated.
380 Vertical blue bars highlight cold stages. See also Table S3 of Supplementary Material for
381 values given in (d) and (f).

382

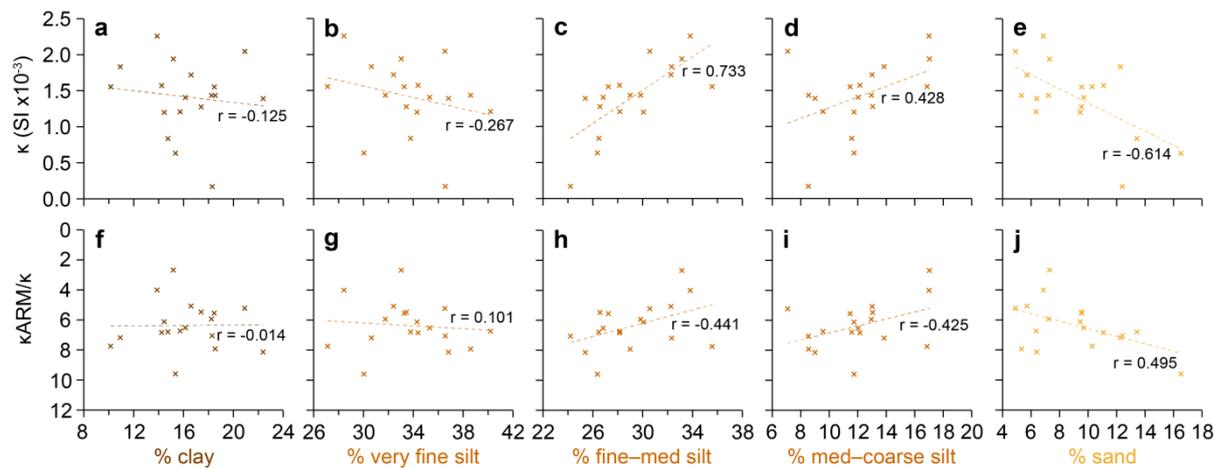
383 Our environmental magnetic records (κ and $\kappa_{\text{ARM}}/\kappa$) exhibit orbital-scale variability throughout
384 most of our study interval, with higher magnetic susceptibility and coarser magnetic grain-size
385 generally characterising (de)glacial intervals containing coarse IRD (compare Fig. 5b to 5c, e).
386 Site U1307 magnetic susceptibility increased dramatically from near-zero values during the
387 termination of MIS KM2, ~ 3.1 Ma, and is associated with relatively low $\kappa_{\text{ARM}}/\kappa$ values.
388 Following the cessation of the mPWP, ~ 3 Ma, magnetic susceptibility remains high and $\kappa_{\text{ARM}}/\kappa$
389 values remain relatively low, but glacial-interglacial variability in these parameters increased
390 between 2.9 and 2.5 Ma, particularly in the $\kappa_{\text{ARM}}/\kappa$ record. Both records show a marked shift
391 between 2.9 and 2.7 Ma, coincident with the onset of persistently elevated coarse IRD inputs
392 to U1307 – κ decreases and the average magnetic grain-size assemblage fines (higher $\kappa_{\text{ARM}}/\kappa$
393 values).

394 The bulk magnetic nature of Eirik Drift sediments can be influenced by changes both
395 in sediment source (Stoner *et al.*, 1995; Hatfield *et al.*, 2016) and potentially sediment texture
396 (Hatfield *et al.*, 2013; 2017). The sensitivity of our magnetic records to both these processes
397 can be established by comparing their relationships to terrigenous grain-size data (Hatfield *et al.*
398 *et al.*, 2016; 2017), and the relationships of κ and $\kappa_{\text{ARM}}/\kappa$ with percent clay, very fine/fine–
399 medium/medium–coarse silt, and sand in U1307 sediments are shown in Figure 6. Terrestrial
400 sources (Hatfield *et al.*, 2017) and sediments from core MD99-2227 (Hatfield *et al.*, 2016)
401 suggest that the magnetic susceptibility of the clay-size fraction is several times lower than the
402 silt-size fractions, and as a result the clay fraction likely exerts restricted influence on bulk κ
403 values (e.g., Hatfield *et al.*, 2019). Relatively low variability in % clay data (Fig. 5f) and the

404 little to no correlation between % clay and κ ($r = -0.125$; Fig. 6a) suggests that any increase in
405 magnetically weak clay-size fractions is unlikely to have driven the secular fining signal seen
406 in our magnetic records between ~ 2.9 and 2.7 Ma in U1307. Of the three silt size fractions, %
407 fine-medium silt ($10\text{--}32\ \mu\text{m}$) has the strongest relationship with bulk magnetic properties, and
408 this size fraction has recently been found to have a strong influence on bulk magnetic
409 susceptibility records (Hatfield *et al.*, 2019). The relatively strong positive correlation that we
410 observe between % fine-medium silt and bulk κ ($r = 0.733$; Fig. 6c) is most likely attributable
411 to the enrichment of ferrimagnetic grains in this terrigenous grain-size fraction (Hatfield *et al.*,
412 2013; 2017).

413 Bulk Mrs/Ms (ratio of remanent saturation moment (Mrs) to saturation moment (Ms))
414 values from sediments deposited at Site U1307 during the Plio-Pleistocene are generally $\sim 0.1\text{--}$
415 0.2 (Mazaud *et al.*, 2015; Kawamura *et al.*, 2012), which implies a relatively restricted coarse
416 PSD-size range of ferrimagnetic grains when viewed on a Day plot (Day *et al.*, 1977) (Fig. S4).
417 Particle-size-specific studies of terrestrial sources relevant for Eirik Drift provenance (outside
418 of Heinrich event intervals) show that only silts and sands from Greenland (and not the
419 Cenozoic volcanics of Iceland or of eastern Greenland) can yield Mrs/Ms values $< \sim 0.15$
420 (Hatfield *et al.*, 2017). Higher fine-medium silt abundance and bulk κ are most strongly
421 associated with lower $\kappa_{\text{ARM}}/\kappa$ values ($r = 0.441$; Fig. 6h) and thus a coarser magnetic grain-size
422 assemblage, which is consistent with increased sourcing of silt from Greenlandic terranes
423 relative to Cenozoic volcanic contributions (Hatfield *et al.*, 2016; 2017). This relationship
424 between bulk magnetic parameters and % silt has previously been observed at Eirik Drift Site
425 MD99-2227, where it has been shown that increases in bulk κ , % silt and coarser magnetic
426 grain sizes reflect increased export of glaciofluvial silt from Greenland (Hatfield *et al.*, 2016;
427 2017).

428



429

430 **Figure 6.** Cross-plots of (a–e) bulk magnetic susceptibility (κ) and (f–j) magnetic grain size
 431 (κ_{ARM}/κ ; note reversed axis since higher values = coarser magnetic grains) with physical grain
 432 size percentage abundance for clay ($\leq 3 \mu\text{m}$), very fine silt (3–10 μm), fine–medium silt (10–
 433 32 μm), medium–coarse silt (32–63 μm) and sand ($\geq 63 \mu\text{m}$) fractions (following Hatfield *et*
 434 *al.*, 2019) derived from discrete sample analyses ($n = 18$; see also Table S4 of Supplementary
 435 Material).

436

437 3.3 Changes in bottom-current strength recorded at Site U1307 during iNHG

438 Using what we know about the magnetic properties of Greenlandic vs. Icelandic sediments
 439 transported to Eirik Drift, the long-term decrease in average ferrimagnetic grain size and
 440 sedimentation rates observed at Site U1307 between ~ 2.9 – 2.7 Ma (Fig. 5) likely reflects a
 441 secular decrease in the abundance of silt of Greenlandic origin transported to the site. This
 442 interpretation is supported by our discrete grain size measurements, which show a long-term
 443 reduction of $\sim 5\%$ fine–medium silt between 2.85–2.25 Ma (Fig. 5d). Since we might logically
 444 expect a greater influx of Greenland-derived material following the onset of significant NHG
 445 ~ 2.7 Ma, as indicated by our IRD records (Fig. 5b), the changes we observe in our magnetics
 446 records, and the contemporaneous long-term decrease in sedimentation rate, are most likely
 447 explained by a shift in the position of the WBUC and delivery of terrigenous sediments to a
 448 different area of the drift. Additional support that the strength of bottom currents bathing our
 449 study site changed significantly during iNHG may be found in our record of wt.% $\geq 63 \mu\text{m}$ sand

450 (Fig. 5a). Peaks in the abundance of well-sorted fine sand, which prior to ~2.7 Ma occur in Site
451 U1307 sediments deposited during interglacials (Fig. 5a), may reflect increased export of
452 glaciofluvial sediment from Greenland to Eirik Drift, most strongly during warm stages. We
453 therefore propose that prior to ~2.9 Ma, the core flow of the WBUC occupied a depth that
454 permitted the delivery of relatively high abundances of Greenland-derived silt and fine sand to
455 Site U1307 during both warm (predominantly) and cold stages on orbital timescales – so our
456 study site was then characterised by a bottom-current-dominated depositional setting – and that
457 between ~2.9–2.7 Ma the volume of Greenland-derived silt delivered to Site U1307 decreased
458 and a glacial IRD-depositional-dominated setting subsequently ensued.

459 Based on the findings of studies that used depth transects of Eirik Drift sediments to
460 infer changes in WBUC vigour during the late Pleistocene (Hillaire-Marcel *et al.*, 1994;
461 Channell *et al.*, 2014; Mazaud *et al.*, 2012, 2015), the decreases in sedimentation rate and
462 magnetic grain-size at U1307 during iNHG that we report may show that WBUC vigour
463 increased and that its core flow deepened relative to our study site between ~2.9–2.7 Ma. A
464 large deepening of the core flow of the WBUC by ~2.7 Ma is not consistent, however, with
465 changes in sedimentation rates reported for Hole 646B on Eirik Drift at ~3450 m water depth,
466 which also appear to have decreased across ~2.7 Ma (Fig. S7b; Wolf and Thiede, 1991).
467 Moreover, a spin-up in the WBUC at this time is also inconsistent with an interpretation based
468 on seismic reflection data that bottom currents not only shallowed, but weakened over Eirik
469 Drift at this time (Müller-Michaelis and Uenzelmann-Neben, 2014), and the observation that
470 iNHG was associated with Last Glacial Maximum-like reductions in the volume of the NADW-
471 overturning cell during cold stages from ~2.7 Ma (Lang *et al.*, 2016).

472 The only mechanism currently proposed that could explain any ‘spin-up’ in the WBUC
473 at this time is the hypothesised late-stage closure history of the Central American Seaway
474 (Bartoli *et al.*, 2005). By contrast, if WBUC vigour actually decreased at this time, this would

475 be more consistent with forcing by sea-ice expansion and increased glacial meltwater input in
476 the Arctic and Nordic Seas, which could have weakened the WBUC by lowering the salinity
477 and density of the water masses that contribute to its formation (see Raymo *et al.*, 2004). Any
478 reduction in WBUC vigour during iNHG could also be explained by increased activity of the
479 Icelandic Hot Spot, which is believed to have uplifted the GSR from ~2.7 Ma to restrict Nordic
480 Seas overflows from this time (e.g., Wright and Miller, 1996; Parnell-Turner *et al.*, 2015).
481 Regardless, our new records provide the first direct geological evidence that the behaviour of
482 the WBUC changed with the expansion of large Northern Hemisphere ice-sheets during the
483 late Pliocene. Ultimately, however, our understanding of whether WBUC vigour increased or
484 decreased during iNHG, and of the potential mechanism(s) and climatological consequences
485 involved, can only be improved by future observations from precisely-dated continuous
486 Pliocene sequences recovered from a depth and spatial transect of drilling sites on Eirik Drift.

487

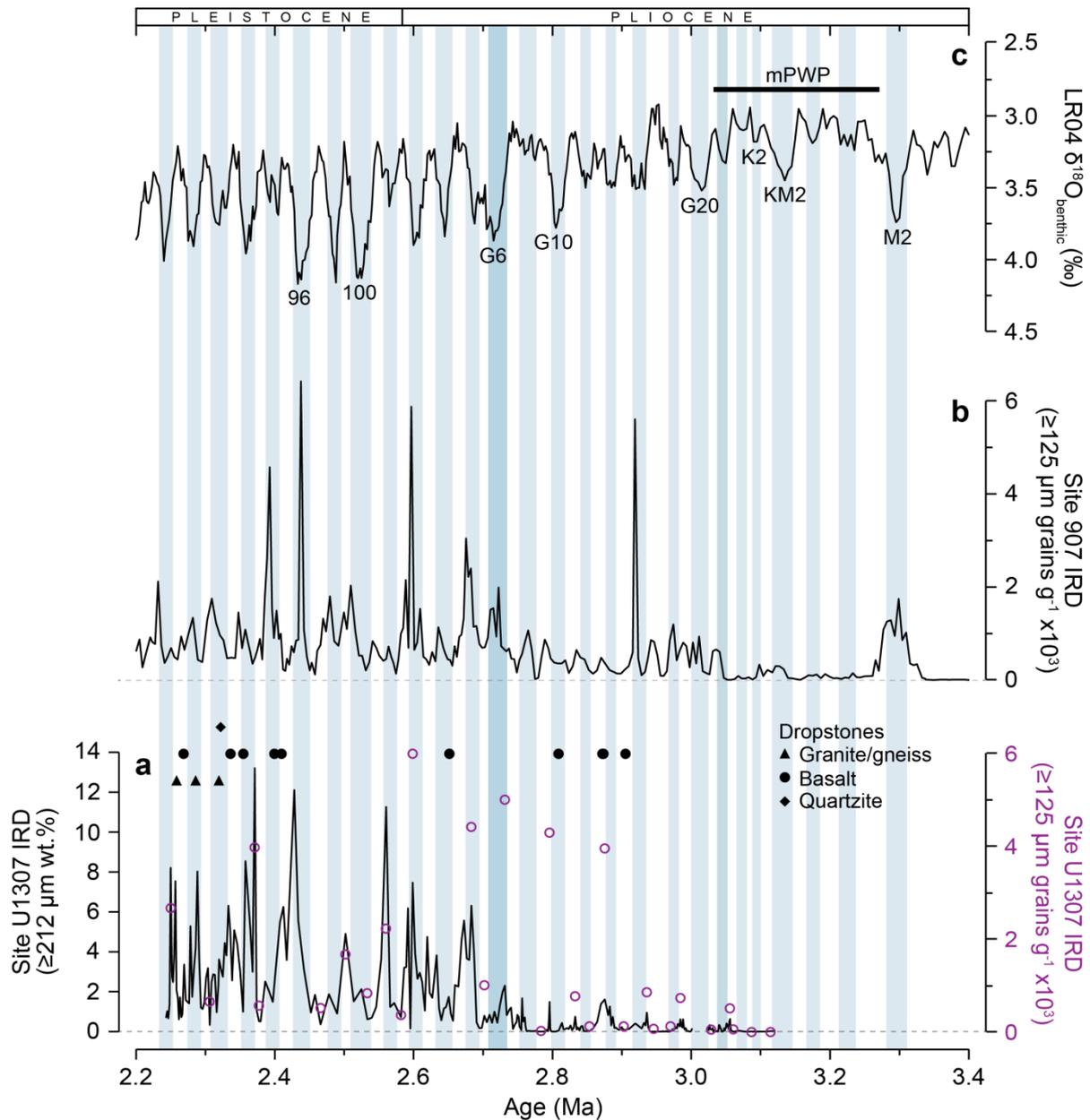
488 **3.4 3.4 Southern GRIS evolution recorded at Site U1307 during iNHG**

489 The general absence of IRD deposition on Eirik Drift (this study and Site 646; Wolf and Thiede,
490 1991; Fig. S7) and the Iceland Plateau (Site 907; Jansen *et al.*, 2000) during the latter half of
491 the mPWP (~3.12–3.03 Ma; Fig. 7a-b) suggests that marine-calving margins were restricted in
492 northeastern, eastern and southern Greenland during this time. Numerical model simulations
493 indicate that continental ice on Greenland may have been mainly restricted to the southern and
494 eastern highlands during the mPWP (e.g., Dolan *et al.*, 2011). The absence of persistent coarse
495 IRD deposition at U1307, yet relatively coarse magnetic grain-sizes (low κ_{ARM}/κ and M_{rs}/M_s
496 values <0.12 (Kawamura *et al.*, 2012)) and high κ , prior to ~2.7 Ma suggests that glaciofluvial
497 silt-producing icecaps existed at least inland on southern Greenland prior to the onset of
498 significant Northern Hemisphere glaciation, including perhaps during the mPWP. The
499 enhanced orbital-scale variability in magnetic grain-size from ~2.9 Ma may also reflect an

500 increase in glacial-interglacial dynamism of previously predominantly inland ice-sheet growth
501 and decay in this region of Greenland.

502 Our $\geq 125 \mu\text{m}$ grains per gram record lacks orbital resolution, but it highlights that IRD
503 inputs to U1307 may have become significantly elevated from $\sim 3 \text{ Ma}$, at least during cold
504 stages (Fig. 7a). The history of ice-rafting inferred from this record is arguably similar to that
505 recorded at Site 907 based on the $\geq 125 \mu\text{m}$ grain-size fraction of IRD (Jansen *et al.*, 2000),
506 which suggests that IRD deposition on the Iceland Plateau was also persistently elevated from
507 $\sim 3 \text{ Ma}$ (Fig. 7b). If correct, this interpretation of these geological data is supported by the
508 finding, based on numerical ice-sheet models, that the mountainous regions of eastern and
509 southern Greenland represent key nucleation points for the Pliocene GrIS (Dolan *et al.*, 2011).
510 Arguably, however, a different story can be drawn from both our coarser IRD $\geq 150 \mu\text{m}$ grains
511 per gram and wt.% $\geq 212 \mu\text{m}$ records, which appear to show that the first sustained episode of
512 significant IRD deposition at U1307 may not have begun until 2.72 Ma, during MIS G6
513 (compare yellow and black data in Fig. 5b). The timing of the onset of continuous major ice-
514 rafting to Eirik Drift during iNHG may therefore actually be most comparable to the history of
515 IRD deposition on the Vøring Plateau in the Nordic Seas, and in the subpolar northeast North
516 Atlantic (e.g., as recorded at ODP Site 644, DSDP Site 611, ODP Site 984, and IODP Site
517 U1308; Fig. 1; Jansen and Sjøholm, 1991; Bailey *et al.*, 2010; Bailey *et al.*, 2013; Bartoli *et*
518 *al.*, 2005). Consequently, the onset of persistent IRD deposition on Eirik Drift $\sim 2.7 \text{ Ma}$ may
519 well post-date the first sustained elevation in ice-rafting on orbital timescales to the more
520 northerly Iceland Plateau (at Site 907), at the cessation of the mPWP $\sim 3 \text{ Ma}$, by $\sim 300 \text{ kyr}$ (Fig.
521 7; also see Section G of Supplementary Information).

522



523

524 **Figure 7.** Records of ice-rafted debris (IRD) abundance from (a) IODP Site U1307 (this study;
 525 black record $\geq 212 \mu\text{m wt.}\%$, purple circles $\geq 150 \mu\text{m grains g}^{-1}$) with shipboard-determined
 526 occurrences of dropstones (Expedition 303 Scientists, 2006a), and (b) ODP Site 907 (Jansen
 527 *et al.*, 2000). The LR04 benthic $\delta^{18}\text{O}$ stack is shown in (c) for reference (Lisiecki and Raymo,
 528 2005). Numbers in (c) are Marine Isotope Stages, with the duration of the mid-Piacenzian warm
 529 period (mPWP; Dolan *et al.*, 2011) also indicated. Vertical blue bars highlight cold stages.

530

531 Any potential delay in the onset of abundant IRD deposition on Eirik Drift relative to the
 532 Iceland Plateau cannot be readily explained by iceberg survivability. This is because the EGC

533 was likely a feature of Nordic Seas surface circulation for the past ~4.5 Myr (e.g., De Schepper
534 *et al.*, 2015), and iceberg trajectory modelling for the warm late Pliocene shows that abundant
535 icebergs only reach the surface waters above Eirik Drift when southern Greenland icebergs-
536 calving sources exist (Smith *et al.*, 2018; their Fig. 4). Instead, it would hint at a regionally
537 diachronous GrIS maturation during iNHG. The implication being, while at least outlet glaciers
538 extended to the coast in northeastern and eastern Greenland following the cessation of the
539 mPWP ~3 Ma, persistent marine-calving margins may not have been established in southern
540 Greenland during cold stages until ~2.7 Ma, when the Fennoscandian and Barents ice-sheets
541 also expanded to their marine-calving margins (Jansen and Sjøholm, 1991; Knies *et al.*, 2014)
542 and glacial expansion occurred at least inland on Arctic Canada (Lang *et al.*, 2014; Bolton *et*
543 *al.*, 2018). This southward expansion may also be echoed in the shipboard-derived dropstone
544 record at U1307 (Fig. 7a; Expedition 303 Scientists, 2006a). Although providing a much less
545 temporally-resolved picture, the dominance of dropstones of basaltic lithology in the U1307
546 stratigraphy prior to ~2.4 Ma contrasts with a mixed suite of basaltic, granitic/gneissic and
547 sandstone dropstones from ~2.4 Ma – consistent with the development of more extensive
548 iceberg-calving sources on Greenland’s southern Precambrian basement terranes during iNHG.

549 The apparent difference in timing of the onset of sustained IRD inputs to U1307 during
550 iNHG that can be determined from the three different grain-size fractions that we examined for
551 evidence of iceberg-rafting highlights the importance of choosing the most appropriate grain-
552 size in sedimentological analysis of IRD. Our wt.% $\geq 212 \mu\text{m}$ and $\geq 150 \mu\text{m}$ grains g^{-1} IRD
553 records can both be used to infer an almost complete absence of IRD deposition on Eirik Drift
554 during warm periods prior to 2.72 Ma. The earlier onset of elevated glacial IRD inputs at U1307
555 that we infer from the $\geq 125 \mu\text{m}$ terrigenous grain-size fraction may be a product of sea-ice
556 rafted sand and/or of the coarsest fine sand also transported to our study site by stronger bottom-
557 currents prior to ~2.7 Ma, complicating our ability to use this grain-size fraction on Eirik Drift

558 as a proxy for iceberg-rafting. Future investigation using, e.g., grain-size end-member mixing
559 may help to determine whether or not the histories of the onset of significant iceberg-rafting to
560 U1307 and 907 were temporally offset during iNHG. Regardless, using our improved RPI-
561 based age model for U1307, the available IRD data indicate that extensive iceberg-calving
562 margins definitely existed in both eastern and southern Greenland from 2.72 Ma (Fig. 7a–b).

563

564 **4. Conclusions**

565 The history of southern Greenland glaciation during the late Pliocene and earliest Pleistocene
566 intensification of Northern Hemisphere glaciation (iNHG) is poorly constrained. Our new
567 sedimentological and paleomagnetic datasets from Site U1307 on Eirik Drift – which receives
568 ice-rafted debris from icebergs transported in the East Greenland Current, and glaciofluvial silt
569 and fine sand via the deep Western Boundary Undercurrent (WBUC) – reveal for the first time
570 that while continental ice existed inland on southern Greenland prior to the onset of significant
571 Northern Hemisphere glaciation, with occasional marine-terminating glaciers during cold
572 intervals from the end of the mid-Piacenzian warm period, marine-calving ice-sheet margins
573 only likely persisted in this region from 2.72 Ma. Our new datasets also highlight for the first
574 time that the depth of the core flow of the WBUC changed relative to our study site during
575 iNHG. This finding underscores the need to redrill the Plio-Quaternary Eirik Drift to obtain a
576 depth transect of drill sites. Only by doing so can we understand the importance of our
577 observations on the WBUC for changes in Atlantic Meridional Overturning Circulation during
578 iNHG, and the role the oceans may have played in driving glaciation at this time.

579

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589

590 All new data presented here are published on www.pangaea.de. Password protection will be
591 lifted upon successful publication of this manuscript, and datasets will be accessible using the
592 following citation:

593 Blake-Mizen, K., Hatfield, R., Stoner, J.S., Carlson, A.E., Xuan, C., Walczak, M.H., Lawrence,
594 K.T., Channell, J.E.T., Bailey, I., 2018. Paleomagnetism from IODP Site 303-U1307.
595 PANGAEA, doi:10.1594/PANGAEA.895029

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598

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772

773 **Table 1.** Site U1307 stratigraphy: comparison of shipboard-derived and revised splice (this study).

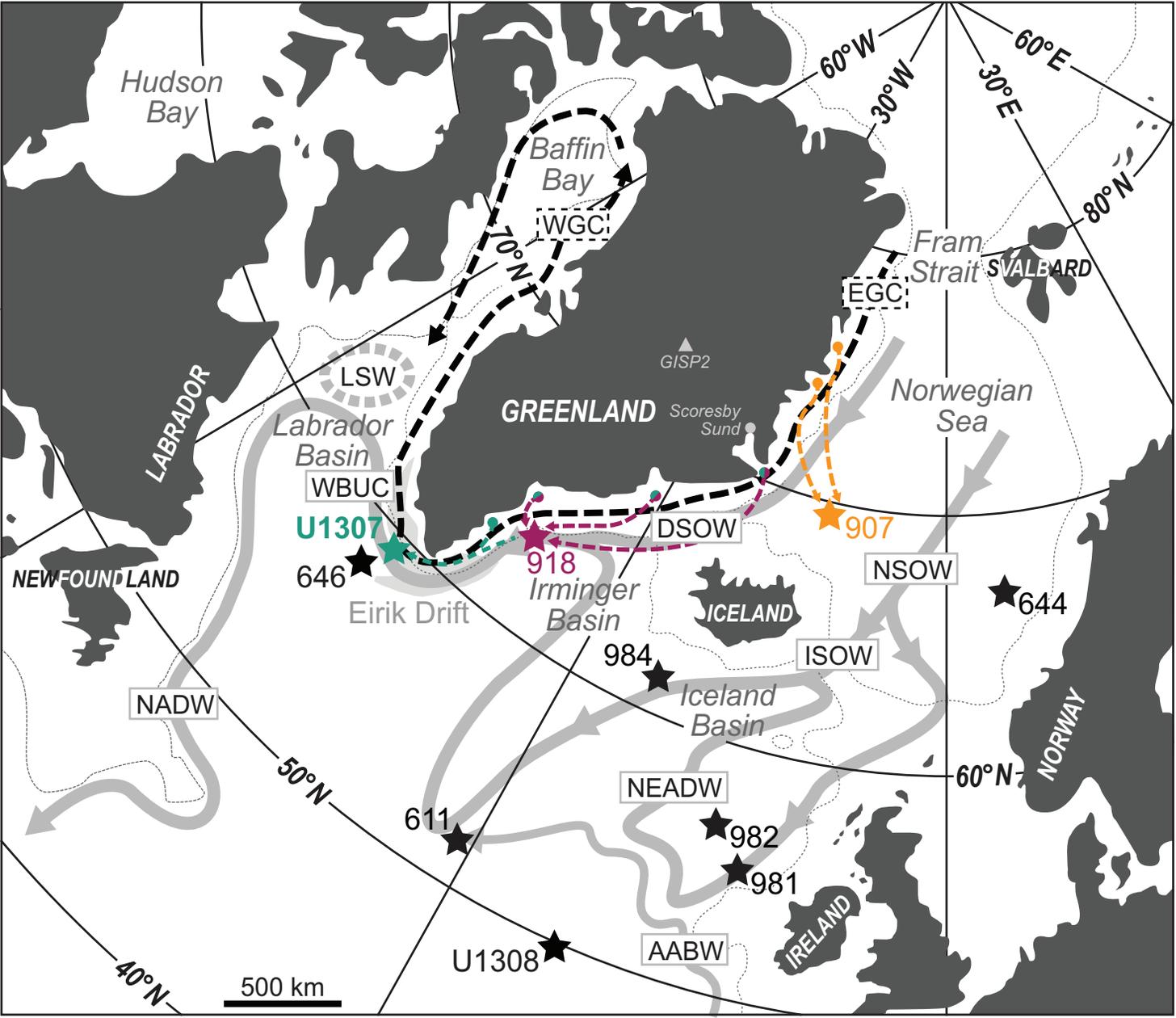
774 See Table S1 of Supplementary Material for complete revised splice.

Hole-Core	Shipboard splice ^a		Revised splice			Hole-Core	Shipboard splice ^a		Revised splice	
	mbsf	mcd	mbsf	rmcd			mbsf	mcd	mbsf	rmcd
U1307B-13H	116.62	123.15	115.82	122.35	tie	U1307A-14H	115.24	123.15	114.44	122.35
U1307A-14H	120.13	128.04	121.73	129.64	tie	U1307B-14H	120.06	128.04	121.74	129.64
U1307B-14H	-	-	127.22	135.13	tie	U1307A-16H	-	-	129.34	135.13
U1307B-14H	128.40	136.38	-	-	append	U1307B-15H	128.30	136.55	-	-
U1307A-16H	-	-	133.62	139.40	tie	U1307B-15H	-	-	131.70	139.40
U1307B-15H	138.00	146.25	-	-	append	U1307B-16H	137.80	147.12	-	-
U1307B-15H	-	-	138.00	145.70	append*	U1307A-17H	-	-	135.20	145.70
U1307A-17H	-	-	140.93	151.43	tie	U1307B-16H	-	-	139.65	151.43
U1307B-16H	145.25	154.57	145.63	157.41	tie	U1307A-18H	146.18	154.47	146.89	157.41
U1307A-18H	152.93	161.32	152.93	163.44	append	U1307A-19H	153.10	163.71	153.10	165.83
						U1307A-19H	162.81	173.42	162.81	175.54

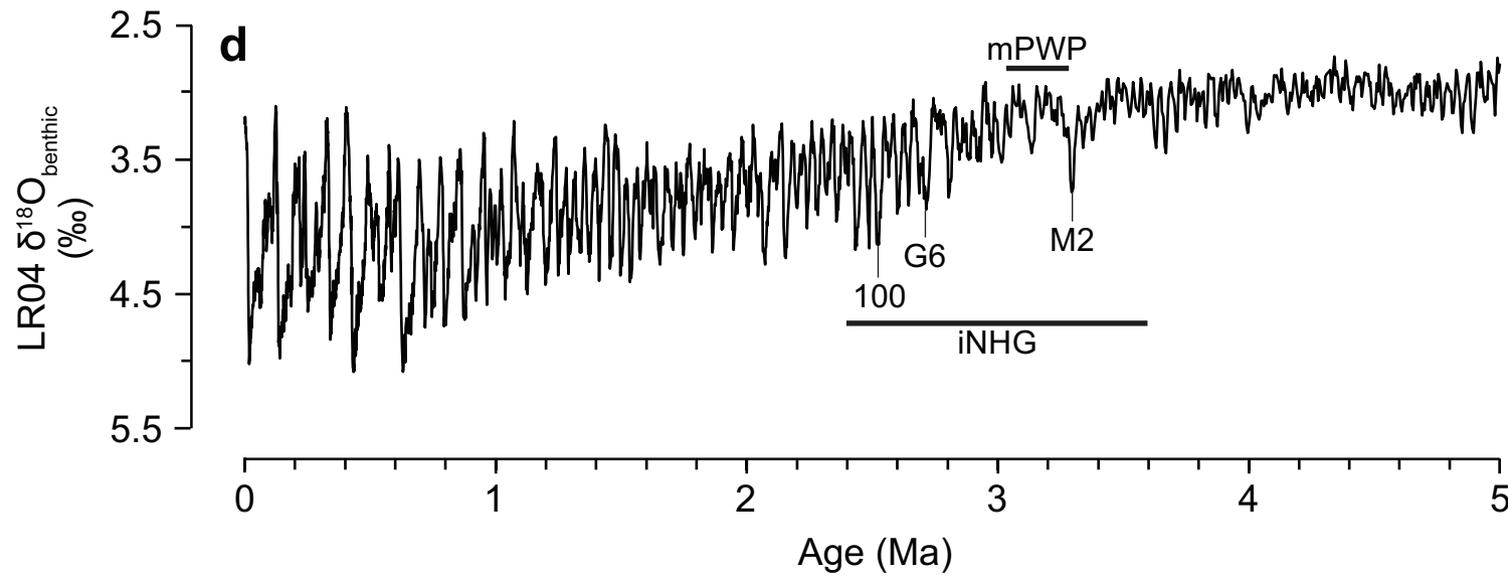
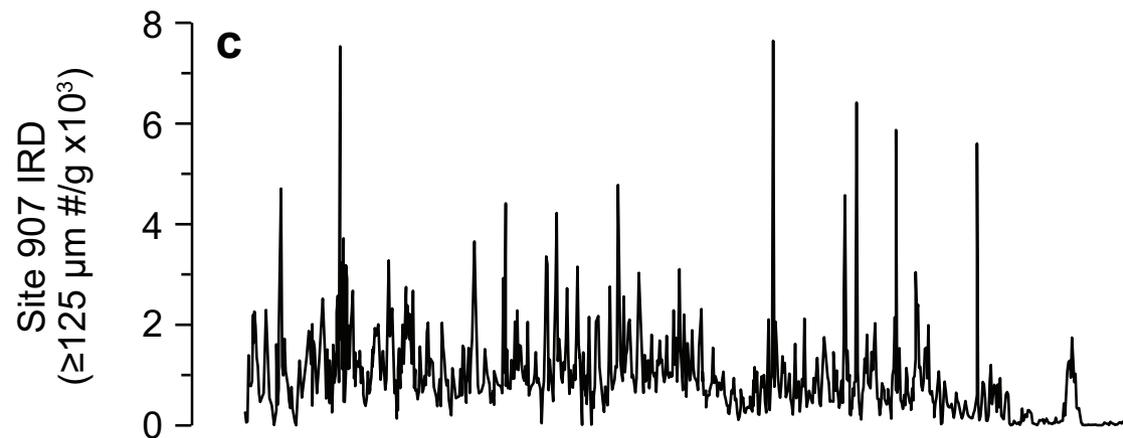
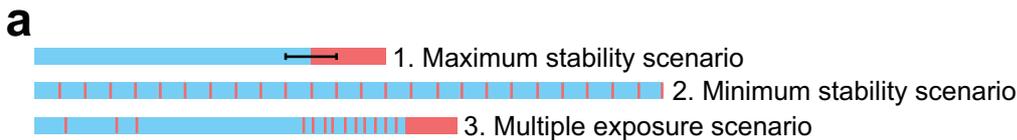
775 mbsf = metres below seafloor; (r)mcd = (revised) metres composite depth.

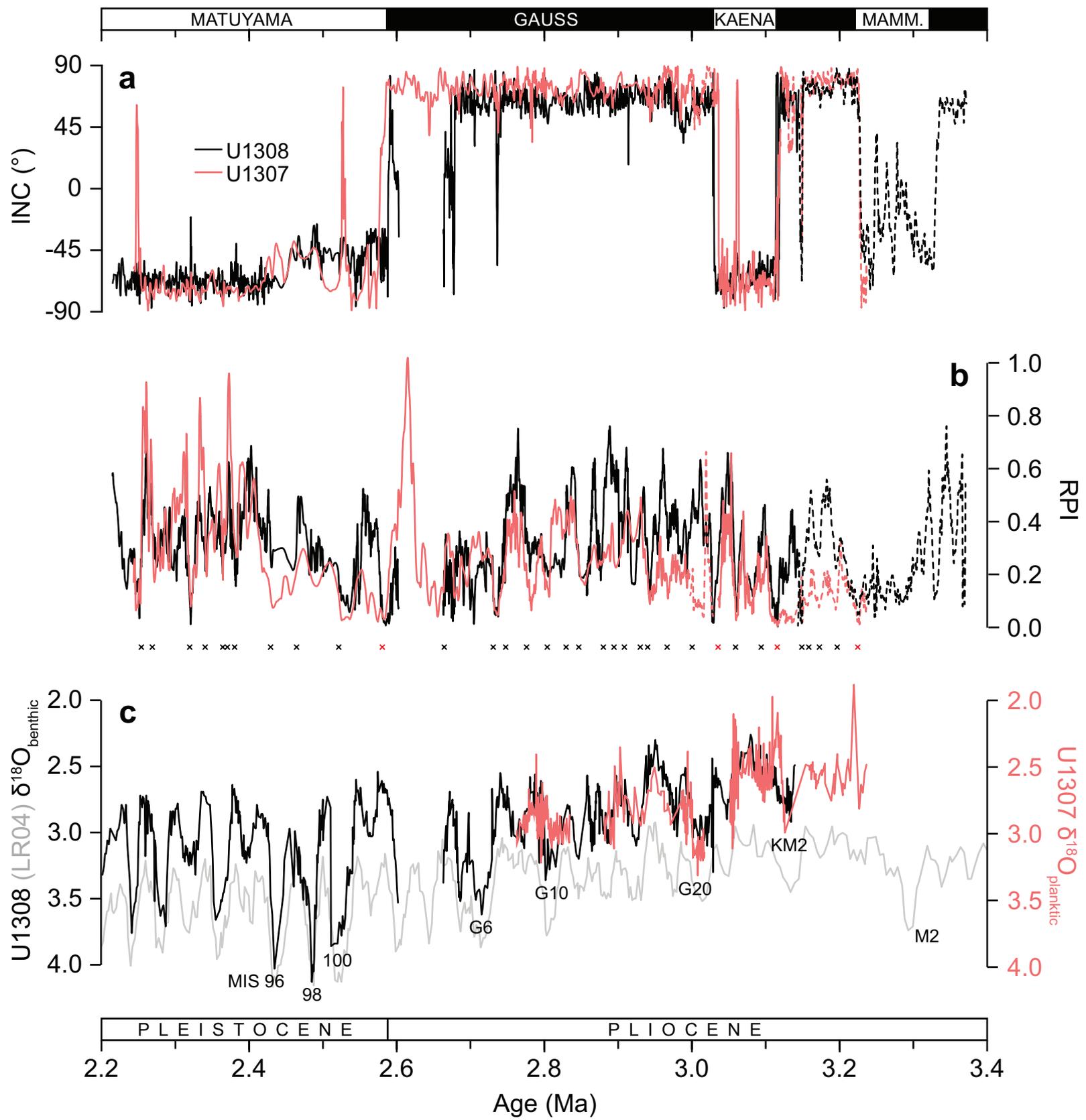
776 ^aExpedition 303 Scientists (2006a).

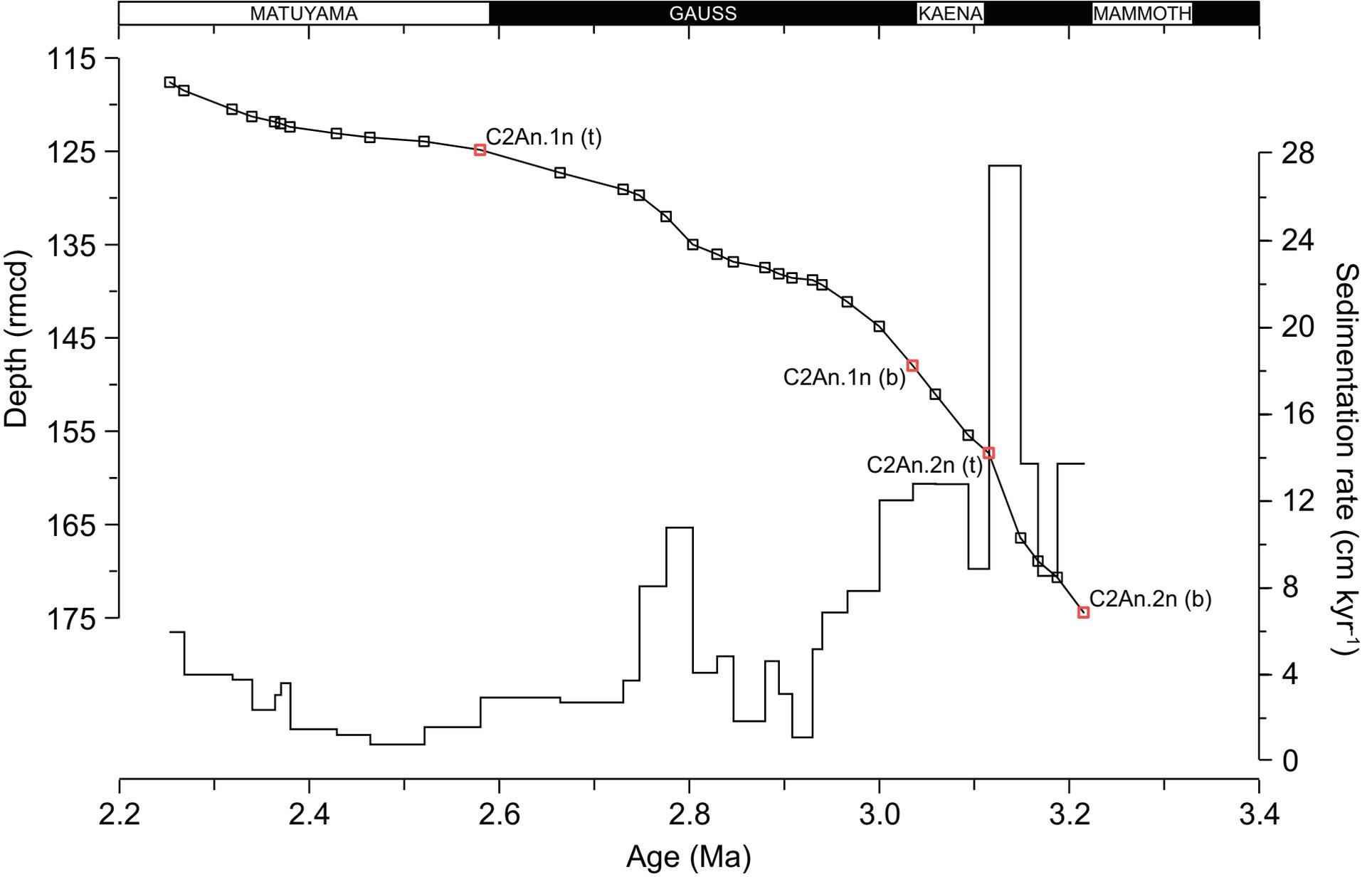
777 *See Fig. S2 of Supplementary Material.

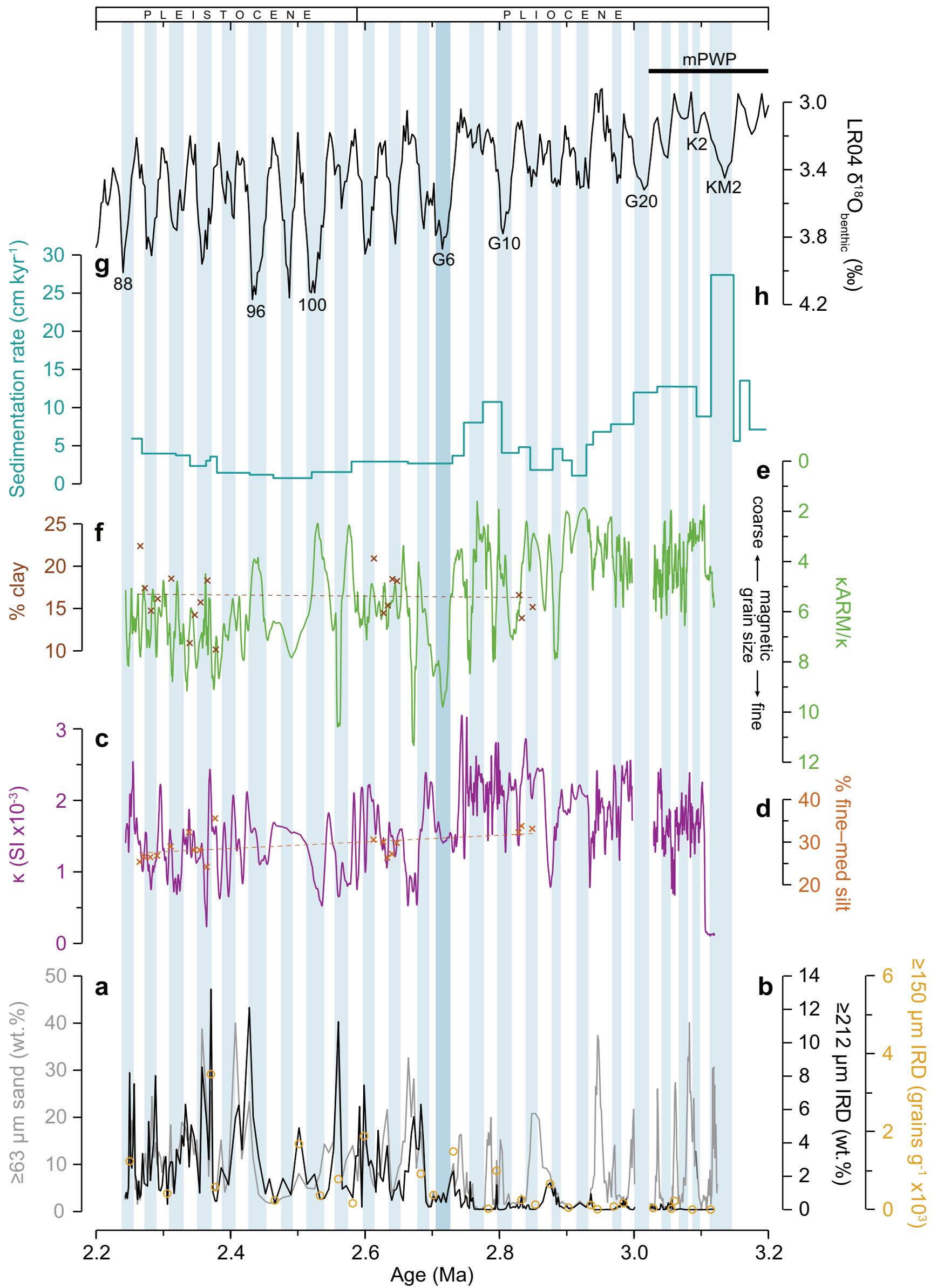


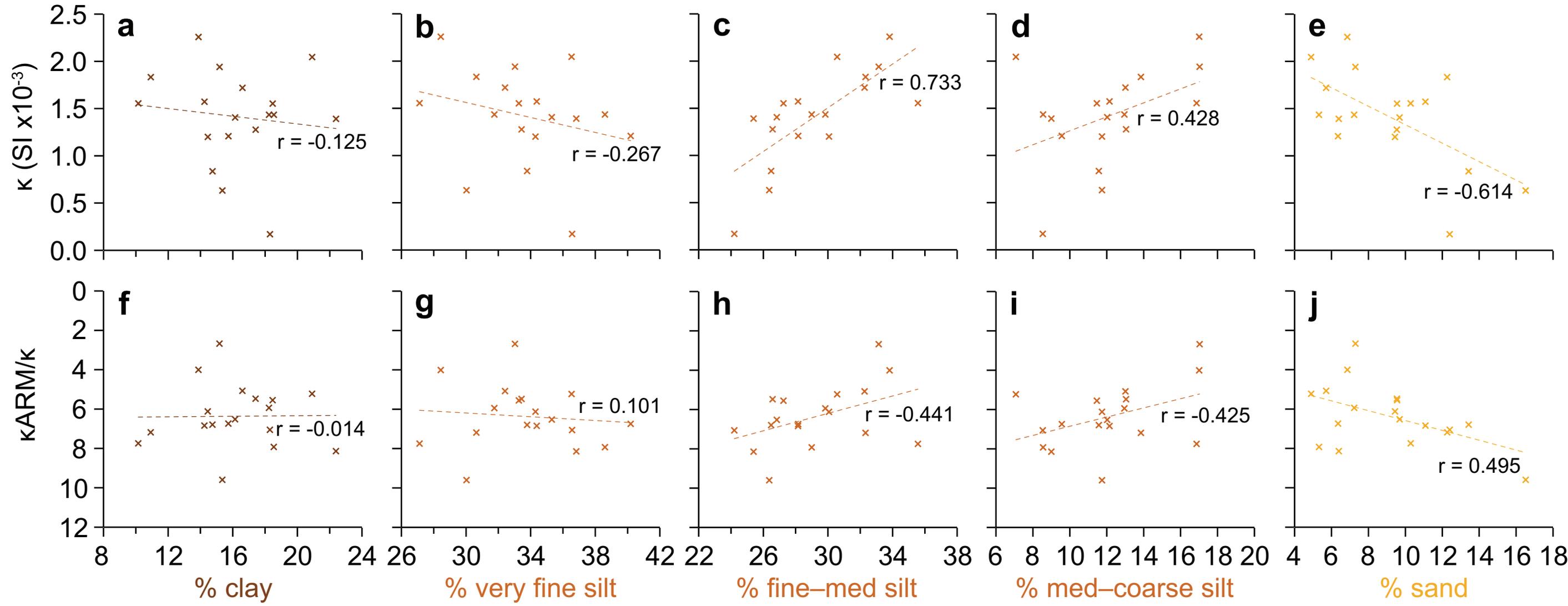
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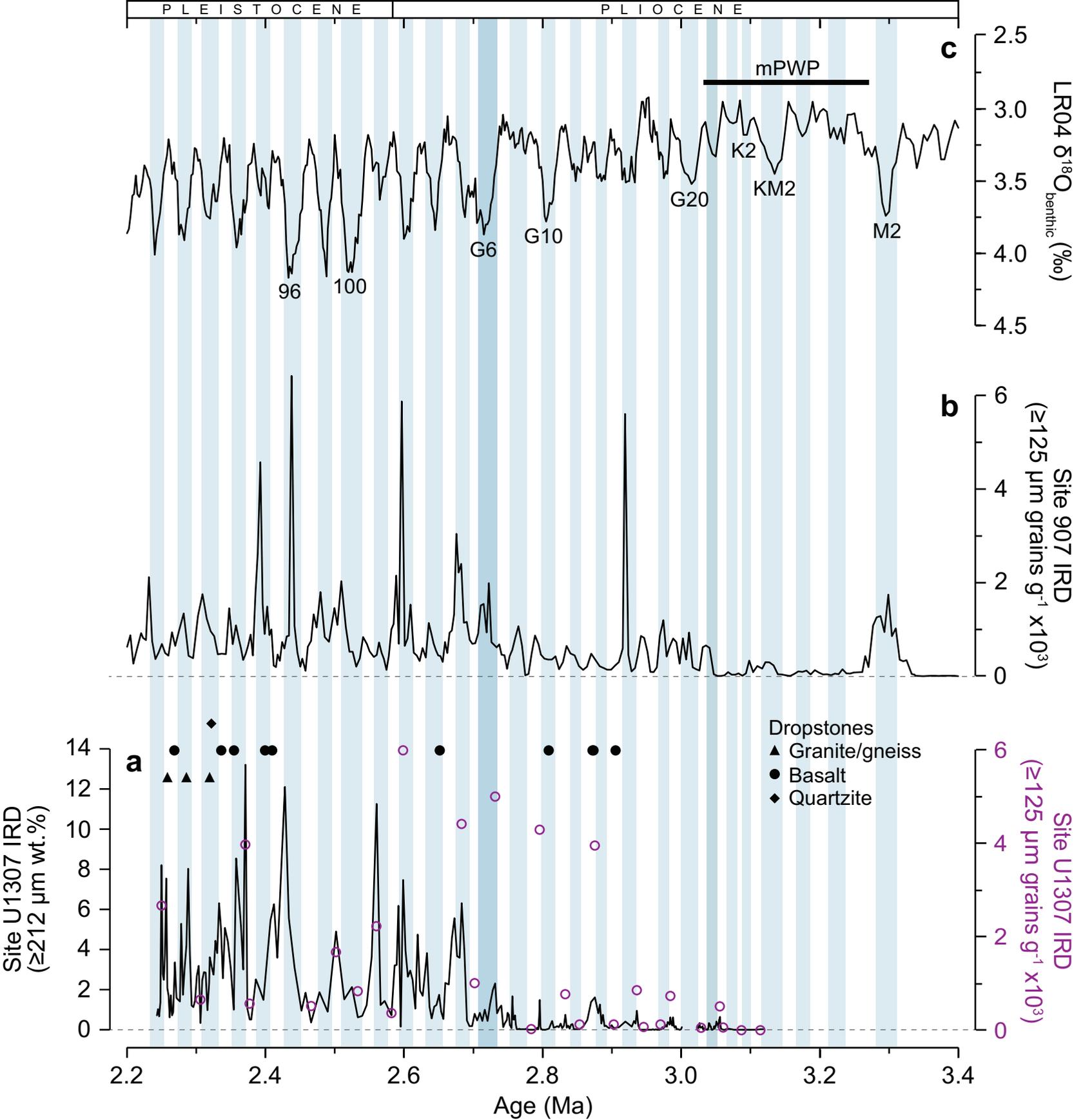


Table 1. Site U1307 stratigraphy: comparison of shipboard-derived and revised splice (this study).

Hole-Core	Shipboard splice ^a		Revised splice			Hole-Core	Shipboard splice ^a		Revised splice	
	mbsf	mcd	mbsf	rmcd			mbsf	mcd	mbsf	rmcd
U1307B-13H	116.62	123.15	115.82	122.35	tie	U1307A-14H	115.24	123.15	114.44	122.35
U1307A-14H	120.13	128.04	121.73	129.64	tie	U1307B-14H	120.06	128.04	121.74	129.64
U1307B-14H	-	-	127.22	135.13	tie	U1307A-16H	-	-	129.34	135.13
U1307B-14H	128.40	136.38	-	-	append	U1307B-15H	128.30	136.55	-	-
U1307A-16H	-	-	133.62	139.40	tie	U1307B-15H	-	-	131.70	139.40
U1307B-15H	138.00	146.25	-	-	append	U1307B-16H	137.80	147.12	-	-
U1307B-15H	-	-	138.00	145.70	append*	U1307A-17H	-	-	135.20	145.70
U1307A-17H	-	-	140.93	151.43	tie	U1307B-16H	-	-	139.65	151.43
U1307B-16H	145.25	154.57	145.63	157.41	tie	U1307A-18H	146.18	154.47	146.89	157.41
U1307A-18H	152.93	161.32	152.93	163.44	append	U1307A-19H	153.10	163.71	153.10	165.83
						U1307A-19H	162.81	173.42	162.81	175.54

See Table S1 of Supplementary Material for complete revised splice.

mbsf = metres below seafloor; (r)mcd = (revised) metres composite depth.

^aExpedition 303 Scientists (2006a).

*See Fig. S2 of Supplementary Material.

CRediT Author Statement

1. Keziah Blake-Mizen: Conceptualization; Methodology; Validation; Formal Analysis; Investigation; Data Curation; Writing – Original Draft; Writing – Review & Editing; Visualization; Funding Acquisition
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9. Ian Bailey: Conceptualization; Methodology; Resources; Writing – Original Draft; Writing – Review & Editing; Supervision; Project Administration; Funding Acquisition

A. New splice for Site U1307

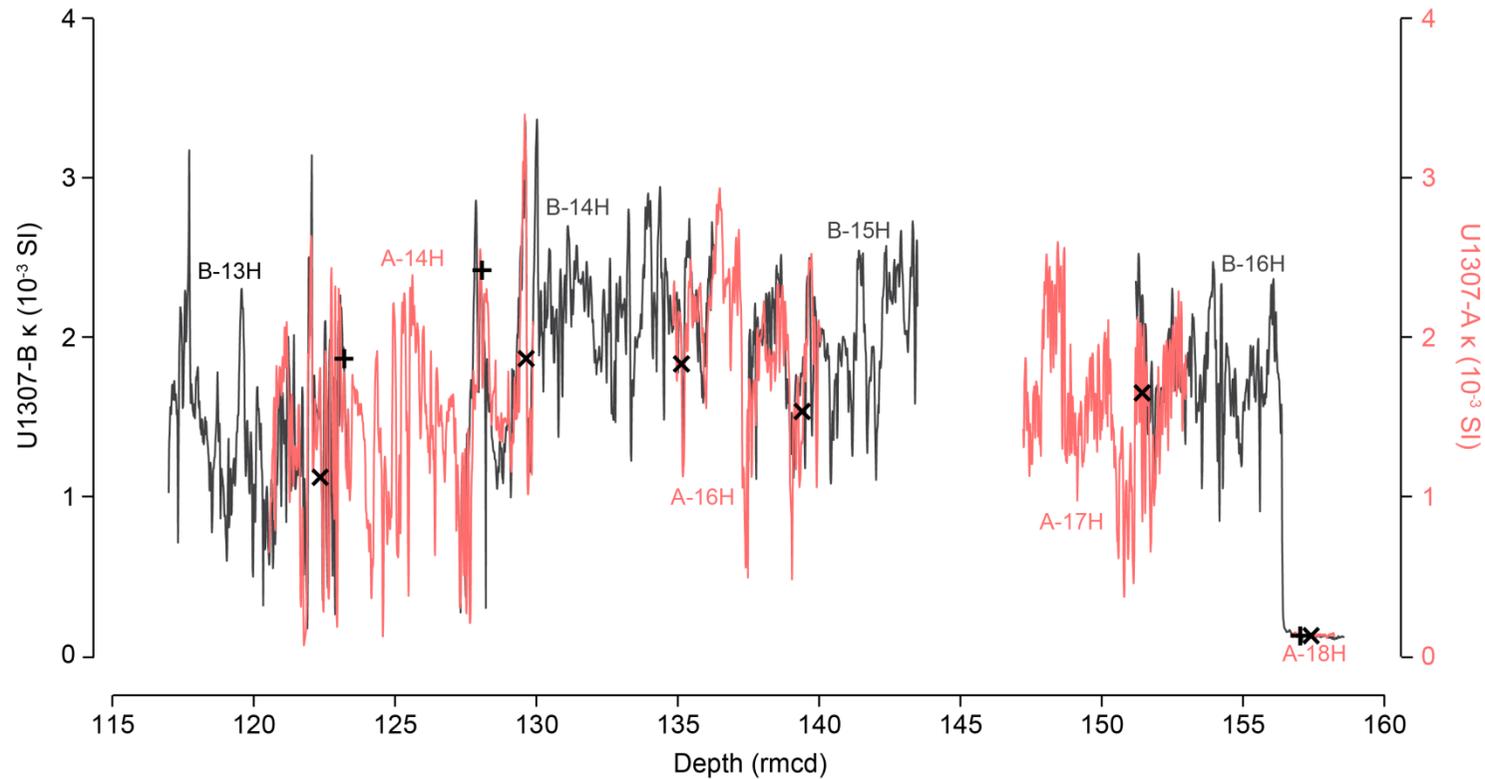


Figure S1. Plot of new high (1 cm)-resolution u-channel-derived magnetic susceptibility (κ) data for IODP Site U1307 Hole A (red) and Hole B (black) between ~117 and 158 revised metres composite depth (rmcd; also see Tab. S1). Newly-determined splice-points denoted by black crosses (\times). Original shipboard splice points are indicated by black plus (+) symbols. The data gap between B-15H and A-17H reflects the absence of u-channel κ data for the bottom 223 cm of B-15H and the top 150 cm of A-17H.

Blake-Mizen *et al.* Southern Greenland glaciation and Western Boundary Undercurrent evolution recorded on Eirik Drift during the late Pliocene onset of Northern Hemisphere glaciation

Table S1. Revised splice for IODP Site U1307.

Core	Affine			Splice								
	Top mbsf	Top rmcd	mbsf-rmcd offset	Section	Interval	mbsf	rmcd	Section	Interval	mbsf	rmcd	Relationship
U1307B-1H	0.00	0.00	0.00	1	0.00	0.00	0.00	2	41.70	1.92*	1.92	tie to
U1307A-1H	0.00	0.19	0.19	2	23.10	1.73*	1.92	5	93.10	6.93	7.12	tie to
U1307B-2H	4.80	5.20	0.40	2	41.90	6.72	7.12	5	40.50	11.20	11.60	tie to
U1307A-2H	9.50	9.69	0.19	2	41.10	11.41	11.60	6	51.20	17.51	17.70	tie to
U1307B-3H	14.30	15.60	1.30	2	59.60	16.40	17.70	5	81.90	21.12	22.42	tie to
U1307A-3H	19.00	20.08	1.08	2	84.10	21.34	22.42	6	73.20	27.23	28.31	tie to
U1307B-4H	23.80	26.99	3.19	1	132.00	25.12	28.31	5	88.10	30.68	33.87	tie to
U1307A-4H	28.50	30.54	2.04	3	32.10	31.82*	33.87	6	64.40	36.64	38.68	tie to
U1307B-5H	33.30	37.24	3.94	1	144.00	34.74	38.68	4	63.40	38.43	42.37	tie to
U1307A-5H	38.00	41.05	3.05	1	132.00	39.32	42.37	5	133.70	45.31	48.36	tie to
U1307B-6H	42.80	47.17	4.37	1	118.50	43.99	48.36	7	77.30	52.11	56.48	append
U1307B-7H	52.30	56.67	4.37	1	0.00	52.30	56.67	7	63.30	61.98	66.35	append
U1307B-8H	61.80	66.71	4.91	1	0.00	61.80	66.71	4	18.60	66.49	71.40	tie to
U1307A-8H	62.00	67.13	5.13	3	126.60	66.27	71.40	7	31.30	71.31	76.44	tie to
U1307B-9H	71.30	75.84	4.54	1	60.00	71.90	76.44	7	25.30	80.55	85.09	tie to
U1307A-10H	77.70	84.07	6.37	1	102.00	78.72	85.09	5	95.50	84.66	91.06	tie to
U1307B-10H	80.80	85.56	4.76	4	96.70	86.27	91.03	7	5.20	89.95	94.61	tie to
U1307A-11H	87.20	93.72	6.52	1	88.50	88.09	64.61	4	22.80	91.93	98.45	tie to

Blake-Mizen *et al.* Southern Greenland glaciation and Western Boundary Undercurrent evolution recorded on Eirik Drift during the late Pliocene onset of Northern Hemisphere glaciation

U1307B-11H	90.30	95.50	5.20	2	144.50	93.25	98.45	7	110.10*	99.48	104.68	tie to
U1307A-12H	96.70	104.38	7.68	1	30.00	97.00	104.68	3	18.10	99.88	107.56	tie to
U1307B-12H	99.80	106.22	6.42	1	134.00*	101.14	107.56	7	60.60	109.41	115.83	tie to
U1307B-13H	109.30	115.83	6.53	1	0.00	109.30	115.83	5	51.90	115.82	122.35	tie to
U1307A-14H	112.40	120.31	7.91	2	54.30	114.44	122.35	7	33.40	121.73	129.64	tie to
U1307B-14H	118.80	126.70	7.90	2	147.30	121.74	129.64	6	92.40	127.22	135.13	tie to
U1307A-16H	125.70	131.49	5.79	3	64.30	129.34	135.13	6	41.80	133.62	139.40	tie to
U1307B-15H	128.30	136.00	7.70	3	10.30	131.70	139.40	7	70.00	138.00	145.70	append**
U1307A-17H	135.20	145.70	10.50	1	0.00	135.20	145.70	4	124.47	140.93	151.43	tie to
U1307B-16H	137.80	149.58	11.78	2	34.78	139.65	151.43	6	32.58	145.63	157.41	tie to
U1307A-18H	144.70	155.22	10.52	2	66.16	146.89	157.41	6	70.00	152.93	163.44	append
U1307A-19H	153.10	165.83	12.73	1	0.00	153.10	165.83	7	71.00	162.81	175.54	

Portion of shipboard splice revised in this study lies between 115.83–175.54 rmd. Original splice 0–115.83 rmd from Expedition 303 Scientists, 2006a.

mbsf = metres below seafloor; rmd = revised metres composite depth.

*corrected for minor typographical errors present in the splice published in the Site U1307 shipboard report (Expedition 303 Scientists, 2006a).

**see Fig. S2.

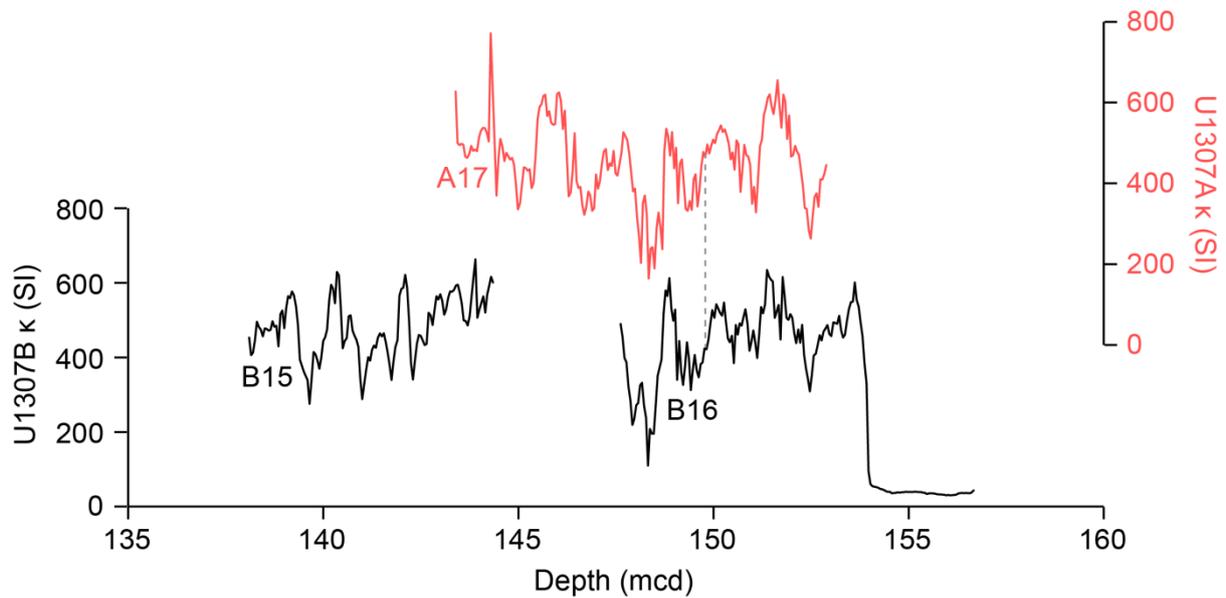


Figure S2. Shipboard magnetic susceptibility core logger (κ) data from Site U1307 on the original shipboard-derived splice metres composite depth (mcd), which appended core B-16H to B-15H to extend the stratigraphy deeper than ~ 145 mcd (Expedition 303 Scientists, 2006a). Core A-17H was not used in the generation of the original shipboard-derived splice but is included in the revised splice presented here (Tab. S1). Data from A-17H plotted here on mcd relative to its splice tie-point with B-16H in the revised splice presented here (see also Fig. S1 and Tab. S1) justify this decision, since the comparison shown suggests that A-17H captures at least part of the stratigraphy lost from the core gap between B-15H and B-16H. Core A-17H cannot be tied to B-15H. To include this core in our revised splice, we assumed that no stratigraphic gap exists between B-15H and A-17H. While a stratigraphic gap of unknown thickness likely exists between these two cores, our inclination- and RPI-based correlations between Sites U1307 and U1308 (Fig. S3) highlight that this gap is likely to be very small.

B. Extended shipboard-derived NRM/ κ record of RPI

The Integrated Ocean Drilling Program (IODP) Site U1308 relative paleointensity (RPI) tuning target for our new IODP Site U1307 record spans 0–3.15 Ma (Channell *et al.*, 2016). In order to assign ages to depths in the Site U1307 stratigraphy >3.15 Ma, we extended the u-channel-based inclination (INC) and RPI records (based on natural remanent magnetisation (NRM)/magnetic susceptibility (κ); e.g., Gorgoza *et al.*, 2006) for both Site U1307 and U1308 using shipboard data. We did this by appending 5-cm resolution shipboard INC, NRM (10 mT) and κ data from cores U1308C 25H–29H to the base of the Site U1308 splice in Hole C (to extend it from 248 metres composite depth (mcd) to 266.5 mcd; note that Hole A data extends further, but the magnetic signal is not well recorded in the deeper portion; Expedition 303 Scientists, 2006b), and cores U1307A 18H–19H to the base of our revised Site U1307 splice (to extend it from 158 revised metres composite depth (rmcd) to 175.5 rmcd, the maximum drilled depth). Where both types of data were available for a given depth, NRM was normalised by κ to give a ‘rough and ready’ measure of RPI, which compares favourably with u-channel-derived RPI data from both sites (Fig. S3).

Blake-Mizen *et al.* Southern Greenland glaciation and Western Boundary Undercurrent evolution recorded on Eirik Drift during the late Pliocene onset of Northern Hemisphere glaciation

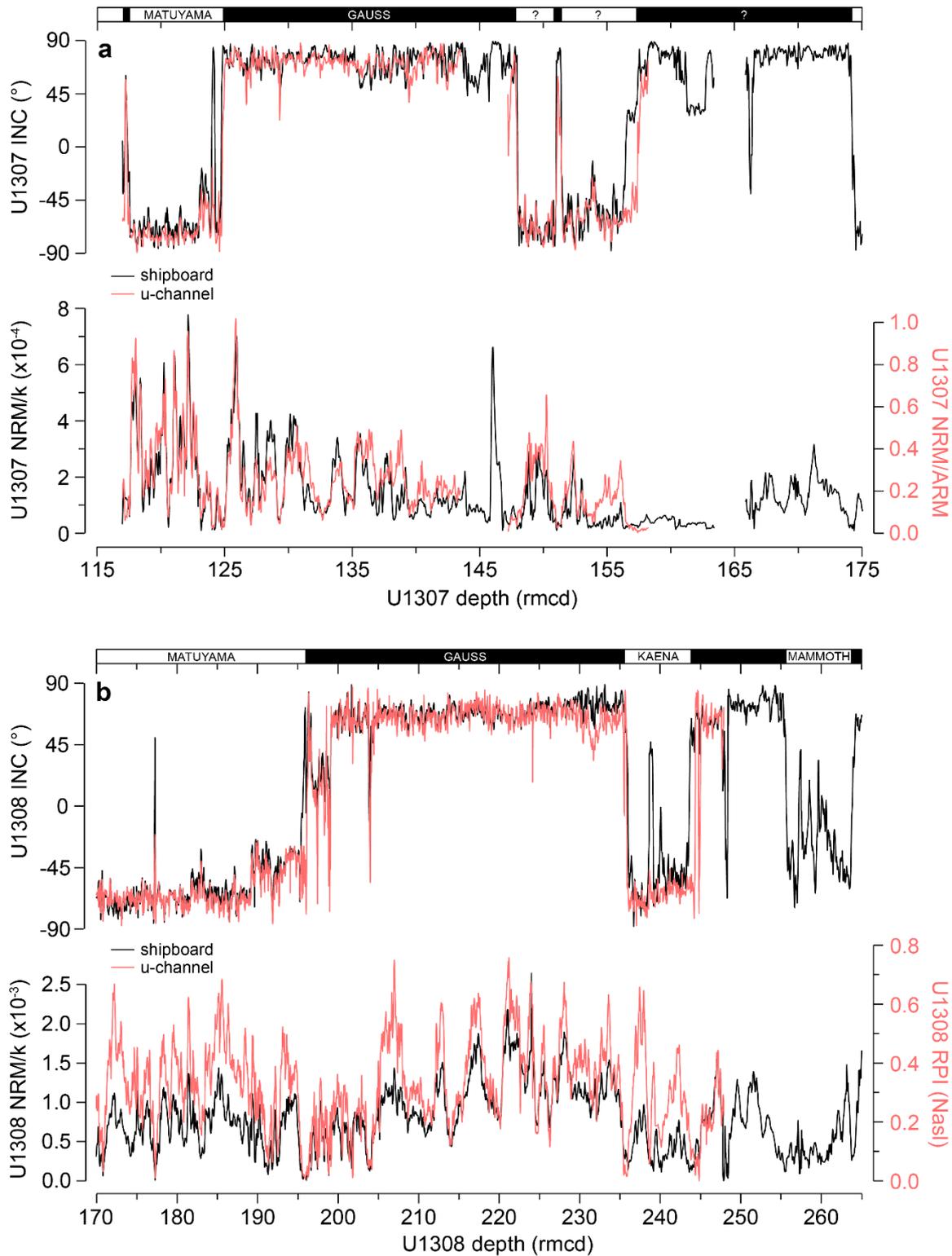


Figure S3. Shipboard- (black lines) and u-channel- (red lines) derived records of inclination (INC) and RPI for IODP Sites (a) U1307 and (b) U1308. The relationships shown demonstrate that shipboard-derived NRM/κ data >3.15 Ma can be used as a ‘rough and ready’ RPI record for these sites to help validate the new age model presented in this study for Site U1307.

C. Construction of U1307 RPI-based age model and comparison to other records

To generate reliable RPI estimates, it is generally accepted that variations in magnetic grain size and magnetic concentration should be minimal (less than an order of magnitude in concentration), and that the magnetic mineralogy remains uniform throughout the interval of interest (e.g., Evans *et al.*, 2007). For our study interval at Site U1307, κ and ARM vary within an order of magnitude (Fig. S4a). A Day plot (Day *et al.*, 1977) of Site U1307 hysteresis ratios (Fig. S4b; Kawamura *et al.*, 2012) shows that, for our study interval, ferrimagnetic grain sizes are fairly well-constrained in the coarse PSD range, and are similar to, and overlap with, hysteresis ratios from shallower depths at U1307 (Mazaud *et al.*, 2015). This suggests that Site U1307 (titano)magnetites generally fall within the size range considered to be most suitable for RPI determination (King *et al.*, 1983). The lack of a clear relationship between NRM/ARM vs. ARM data (Fig. S4c) indicates that NRM/ARM (our RPI estimate) is not dependent on environmental variations in ferrimagnetic concentration, and is thus sensitive to variations in past field intensity.

Component u-channel inclination and declination data for ~117–158 rncd are given in Figure S5a–b, alongside the MAD values (Fig. S5c). MAD values are generally low (<5°), reflecting a well-defined ChRM magnetisation. However, higher (up to ~45°) values are a feature of polarity reversals and other low intensity intervals, indicating complex or poorly constrained ChRM directions (Fig. S5c). Inclination values for both polarities vary close to the expected value ($\pm 72.5^\circ$) assuming a geocentric axial dipole field (Fig. S5b). The declination record reveals reversal horizons coeval with the inclination data (Fig. S5a), which can then be correlated to the geomagnetic polarity time scale. Based on these data, we identify the Gauss-Matuyama (G/M) boundary (C2An.1n top) at 125.10 rncd (~2.581 Ma; Ogg, 2012), the top of the Kaena (C2An.1r) at 147.97 rncd (~3.032 Ma; Ogg, 2012), and short excursions to normal

polarity within the Matuyama chron at ~117 rmcd and within the Kaena subchron at ~151 rmcd (see Tab. S2).

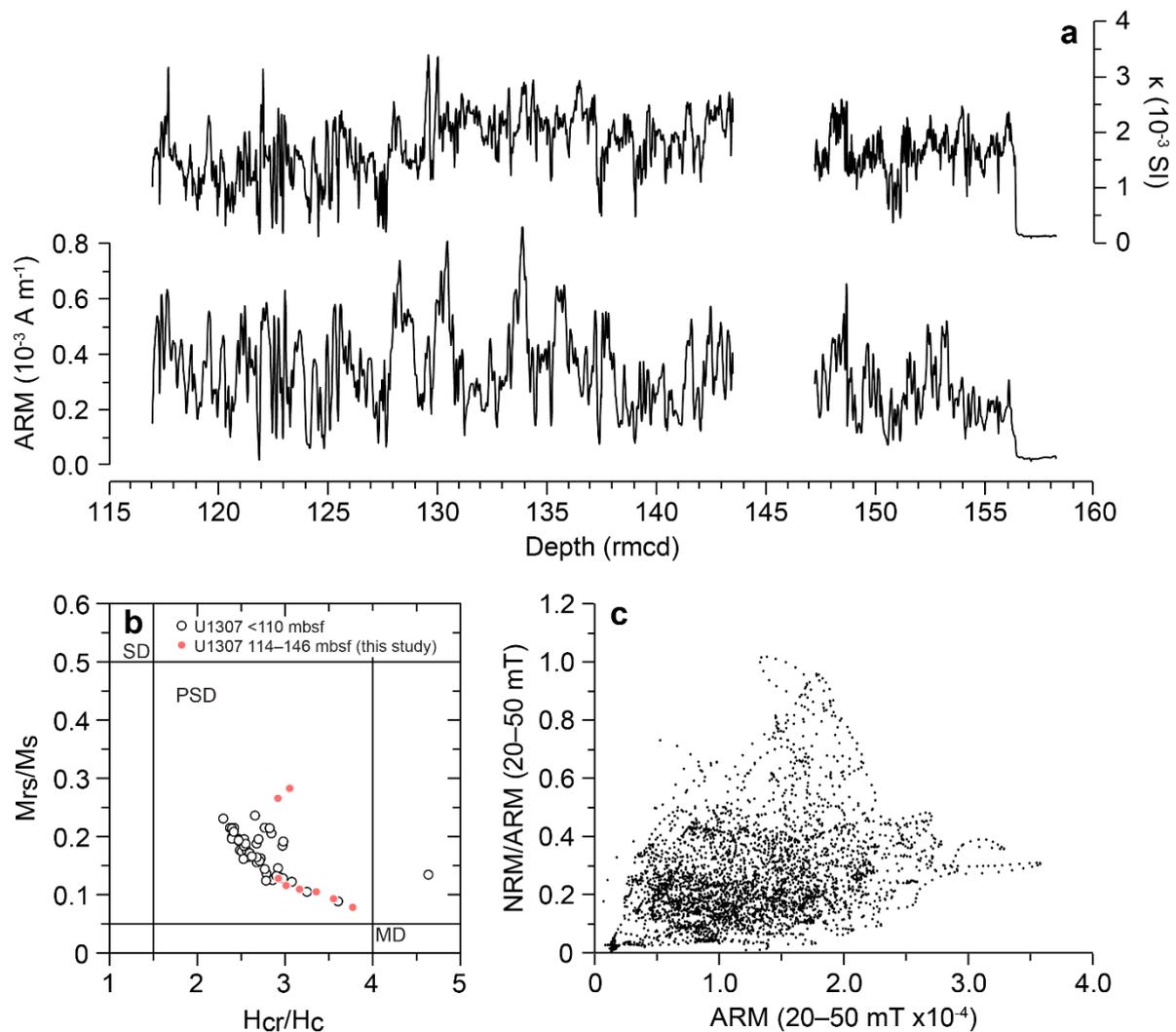


Figure S4. (a) Plot of Site U1307 low-field bulk volume magnetic susceptibility (κ) and u-channel-derived anhysteretic remanent magnetisation (ARM before AF demagnetisation) against revised metres composite depth (rmcd); (b) hysteresis parameters (M_{rs}/M_s , ratio of remanent saturation moment M_{rs} , to saturation moment M_s ; against H_{cr}/H_c , ratio of remanent coercive force, H_{cr} , to coercive force, H_c) for single U1307 samples showing (titano)magnetite grain size distribution, plotted on a Day *et al.* (1977) diagram (MD = multi-domain grains, PSD = pseudo-single domain, SD = single domain), with samples in our study interval indicated by red points (modified from Kawamura *et al.*, 2012); and (c) bivariate plot of natural remanent magnetisation (NRM)/ARM vs ARM for the 20–50 mT peak field interval.

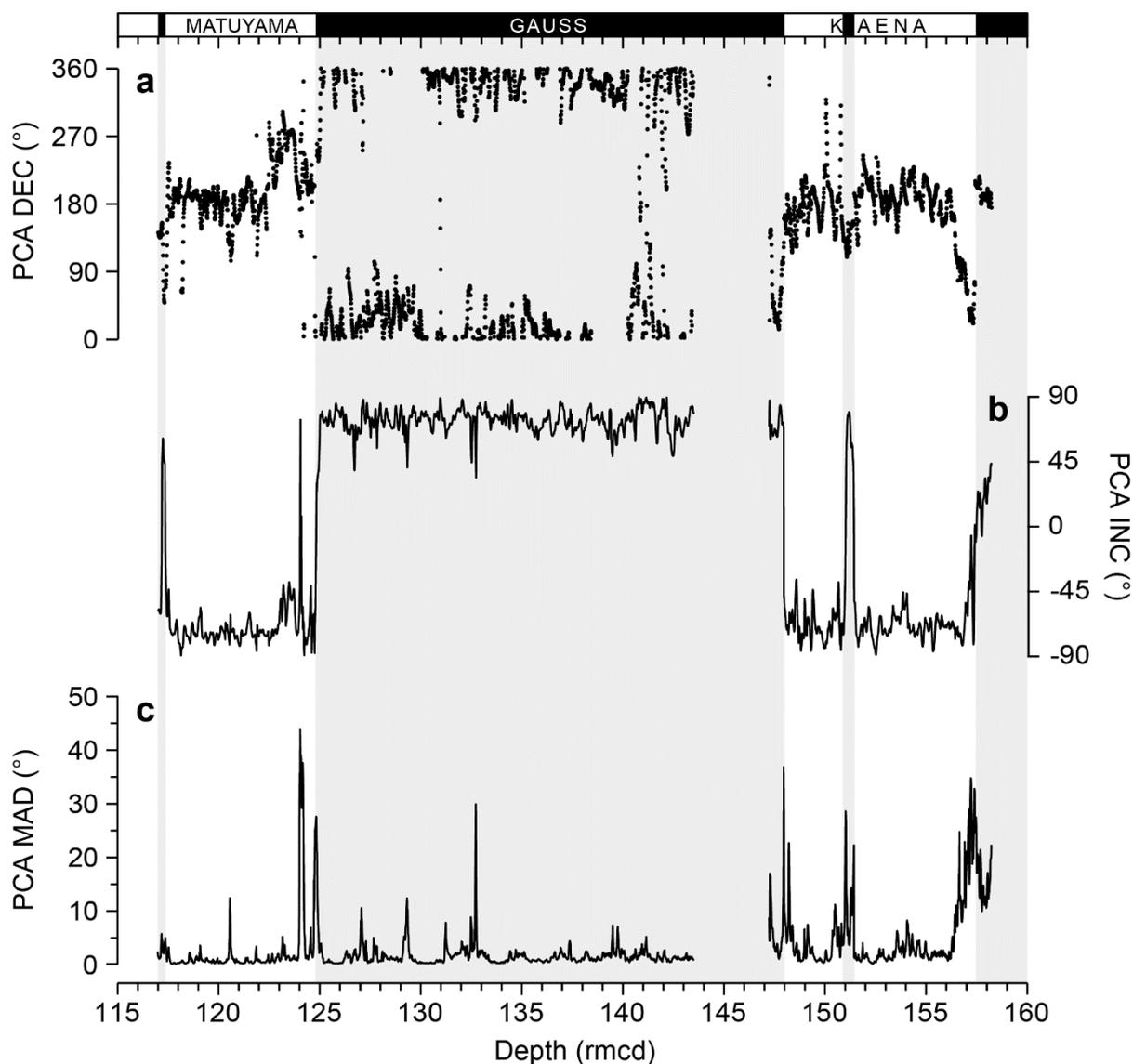


Figure S5. Site U1307 natural remanent magnetisation (NRM) component (a) declination, (b) inclination and (c) maximum angular deviation (MAD) values, calculated for the 20–50 mT peak interval, against revised metres composite depth (rmcd). The absence of data between ~143.5–147.2 rmcd is the result of a sampling gap (the bottom 223 cm of U1307B-15H and the top 150 cm of U1307A-17H could not be u-channelled). Shipboard data do not suggest a reversal occurs during this interval; see Fig. S3). Polarity (normal/reversed) chrons are denoted at the top by black/white horizontal bars/grey vertical bars and labels, based on the age model presented in this study.

Blake-Mizen *et al.* Southern Greenland glaciation and Western Boundary Undercurrent evolution recorded on Eirik Drift during the late Pliocene onset of Northern Hemisphere glaciation

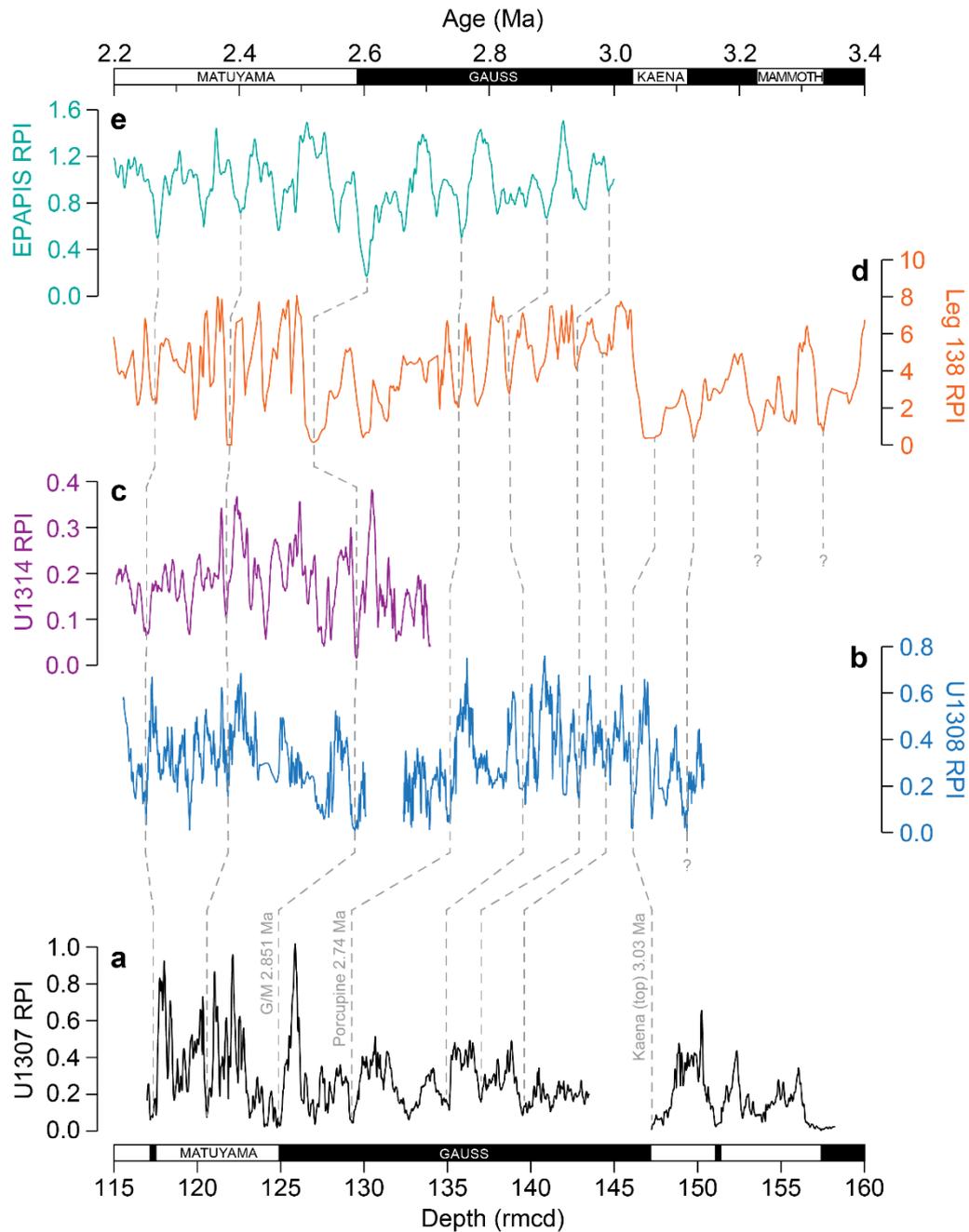


Figure S6. First-order comparison of (a) our new RPI record from IODP Site U1307 on revised metres composite depth (rmcd; this study) with other RPI stratigraphies from the North Atlantic ((b) Site U1308, Channell *et al.*, (2006); (c) Site U1314, Ohno *et al.*, (2012)) and equatorial Pacific ((d) Ocean Drilling Program (ODP) Leg 138, Valet and Meynadier, (1993); (e) EPAPIS-3000 stack, Yamazaki and Oda, (2005)) on their respective published age models. This comparison shows that the U1307 RPI stratigraphy records a regionally coherent signal and shares many similarities with records from further afield.

D. Comparison of U1307 and ODP Site 646 sedimentation rate and $\geq 63 \mu\text{m}$ sand records

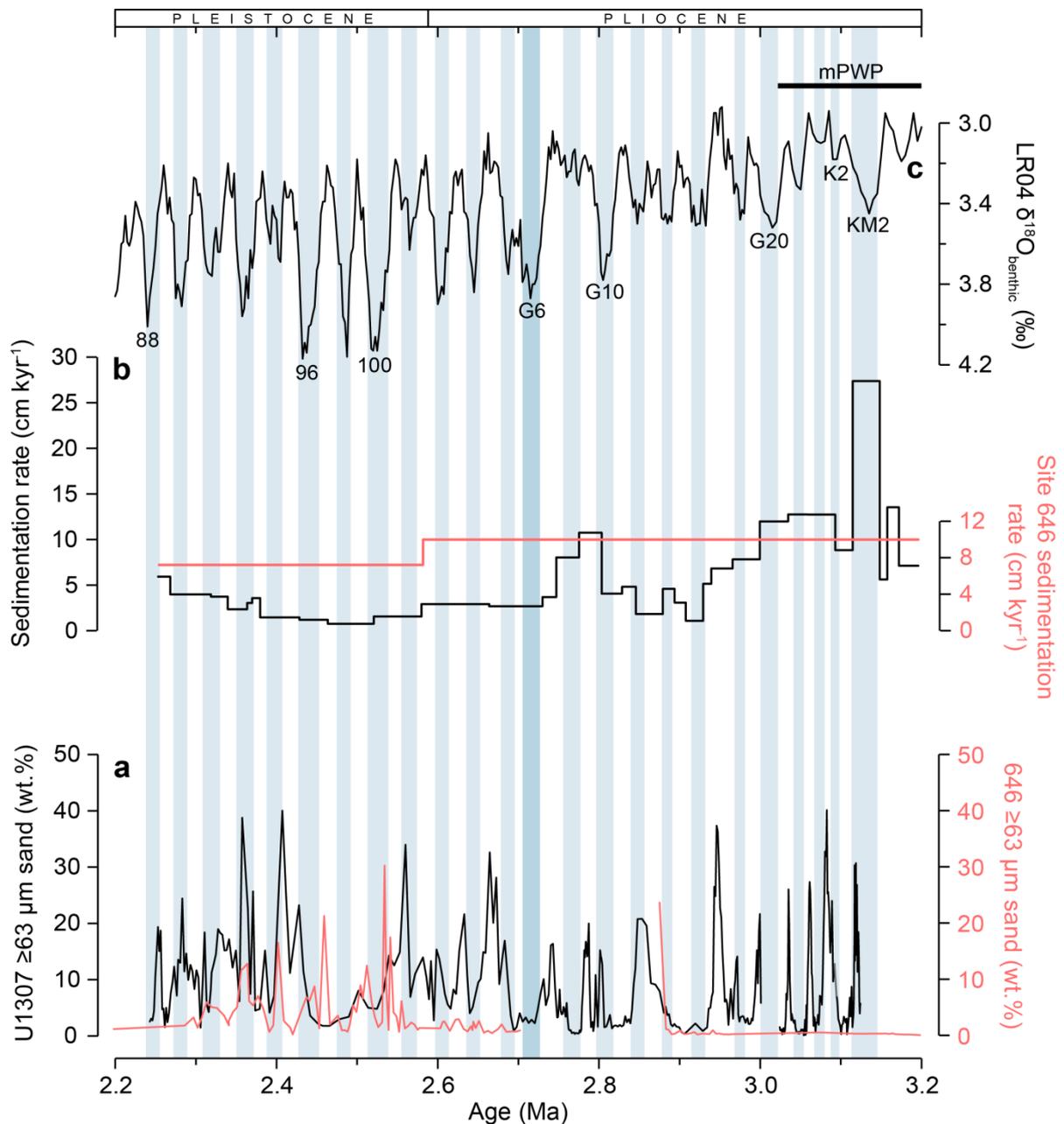


Figure S7. Comparison of Eirik Drift records from IODP Site U1307 and ODP Hole 646B: (a) weight percent (wt. %) $\geq 63 \mu\text{m}$ terrigenous sand; (b) linear sedimentation rates (data from Wolf and Thiede, 1991). The LR04 benthic $\delta^{18}\text{O}$ stack (Lisiecki and Raymo, 2005) is also shown in (c). Numbers in (c) are Marine Isotope Stages, with the duration of the mid-Piacenzian warm period (mPWP; Dolan *et al.*, 2011) also indicated. Vertical blue bars highlight cold stages. The Hole 646B stratigraphy is constrained by only three palaeomagnetostratigraphic

Blake-Mizen *et al.* Southern Greenland glaciation and Western Boundary Undercurrent evolution recorded on Eirik Drift during the late Pliocene onset of Northern Hemisphere glaciation

reversals spanning ~2.5 Myr (the base of the Jaramillo, 1072 ka; the Gauss/Matuyama, 2581 ka; and the Gauss/Gilbert, 3596 ka). Our ability to compare sand deposition and accumulation rates at U1307 and 646B during our study interval is therefore limited by the absence of a high fidelity $\delta^{18}\text{O}$ - or relative paleointensity-based age model for the Hole 646B stratigraphy.

E. Method for organic and biogenic matter removal for grain-size analysis

Modified from Povea *et al.* (2015)

1.1 Materials

Total of 18 dried core samples typically ~5 g

10% hydrogen peroxide solution

10% acetic acid solution

1.5 M sodium hydroxide solution

Deionised water

1.2 Step one: Organic matter removal

Dry samples were broken up as finely as possible without influencing grain size and put into 50 mL centrifuge tubes. 40 mL of 10% hydrogen peroxide was added to each sample in the tubes, then these were placed in an oven at 60°C with their lids loosely resting on top to allow safe escape of oxygen gas. The samples were left in the oven for ~48 hours, or until the reaction stopped, after which any remaining liquid was poured off. Deionised water was then added to each sample in the tubes, thoroughly shaken by hand and centrifuged, before the liquid was poured off. The resultant clean samples were then dried in an oven at 60°C overnight.

1.3 Step two: Biogenic carbonate removal

35 mL of 10% acetic acid was added to each sample in the centrifuge tubes, which were then placed in an end-over-end turner for a few hours (or until fizzing stopped). Next, the liquid was centrifuged off, deionised water was added and thoroughly shaken, and then the samples were centrifuged again. The liquid was then poured off and then the clean samples were dried in an oven at 60°C overnight.

1.4 Step three: Biogenic silica removal

35 mL of 1.5 M sodium hydroxide added to each sample in the centrifuge tubes, then sonicated in an ultrasonic bath for 5 to 10 minutes. The samples were left in the solution overnight in an oven at 65°C. The next day the tubes were placed in a water bath at 85°C for 2 hours, then the liquid was centrifuged off. Subsequently, 40 mL of 1.5 M sodium hydroxide was added and the samples were placed back in the 85°C water bath for 2 hours. The samples were then left in solution overnight in an oven at 65°C. The next day the liquid was centrifuged off, deionised water added to each sample and shaken thoroughly. They were then centrifuged and the liquid was poured off. This step was repeated two more times, and then the clean samples were dried in an oven at 60°C overnight.

F. Evaluation of alternative age model for IODP Site U1307

Two alternative age models for the late Pliocene and early Pleistocene portion of the shipboard-derived Site U1307 stratigraphy have previously been proposed by Sarnthein *et al.* (2009), which are based on their interpretation of the shipboard palaeomagnetostratigraphy and a discontinuous planktic foraminiferal *Neoglobigerina atlantica* (s) stable oxygen isotope record. The preferred U1307 age model of Sarnthein *et al.* (2009) – ‘Age Model 2’ – assumes that the lowest-most palaeomagnetostratigraphic reversal preserved in the U1307 shipboard INC data is the Gilbert/Gauss reversal ~3.58 Ma, and that the Mammoth reversed subchron (C2An.2r) lies somewhere in the base of A-18H (within a magnetic susceptibility low) and the recovery gap with A-19H. Retuning of our U1307 RPI record to the U1308 stratigraphy guided by Sarnthein *et al.* (2009)’s ‘Age Model 2’ (Fig. S5) highlights the following deficiencies: (1) U1307-U1308 RPI match is poor below the Kaena (bottom); (2) the alignment of U1307 *N. atlantica* (s) $\delta^{18}\text{O}$ and U1308 benthic $\delta^{18}\text{O}$ (and the LR04 stack where U1308 $\delta^{18}\text{O}$ data are not available) below the Kaena (bottom) is unconvincing; and (3) there is no evidence in the inclination record that the reversed Mammoth subchron lies where assumed by Sarnthein *et al.* (2009). Our new RPI correlation to U1308 illustrates that the basal reversal in U1307 is instead the Mammoth top (~3.22 Ma), reducing the proposed maximum age of sediment recovered at U1307 by ~330 kyr.

The last occurrence (LO) of the dinocyst *Operculodinium eirikianum* identified shipboard in the base of U1307A (Expedition 303 Scientists, 2006a) is dated at ~3.3 Ma, but we argue that it should not be used as an initial tie point in constructing our RPI-based age model for this site: (1) its LO is based on a core catcher sample, so its actual last occurrence could be as much as ~10 m higher in the U1307A stratigraphy than reported; and (2) the ~3.3 Ma age assigned to this dinocyst LO is based on the unspliced stratigraphy from Ocean Drilling Program (ODP) Labrador Sea Site 646 that has poor age control (this site lacks a benthic $\delta^{18}\text{O}$ -

Blake-Mizen *et al.* Southern Greenland glaciation and Western Boundary Undercurrent evolution recorded on Eirik Drift during the late Pliocene onset of Northern Hemisphere glaciation

based age model, and is dated only by a reversal-based magnetostratigraphy that does not resolve the Kaena and Mammoth subchrons; Shipboard Scientific Party, 1987).

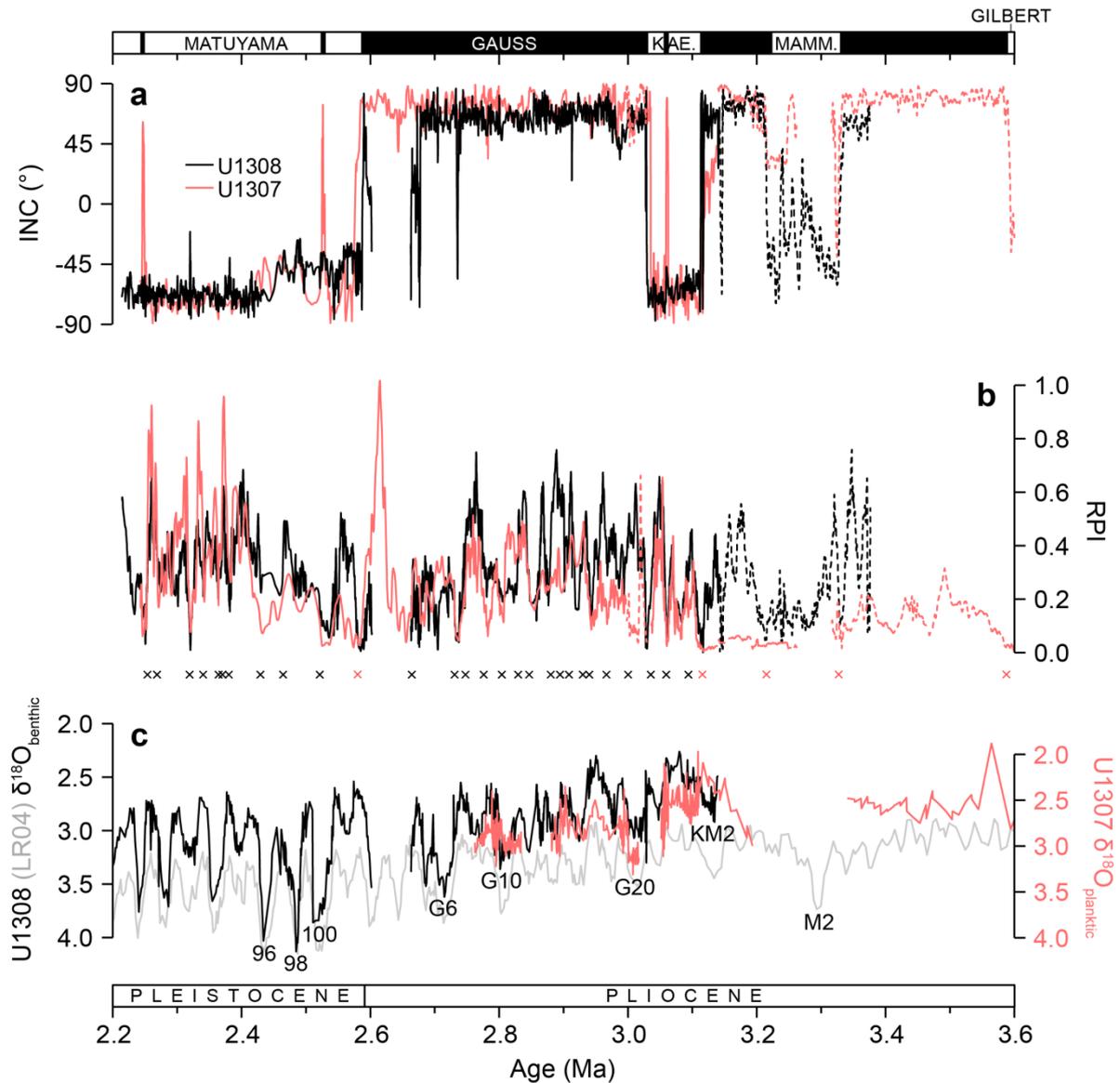


Figure S8. Records of (a) inclination (INC; red - this study; black – Channell *et al.*, 2016), (b) relative paleointensity (RPI; red – this study; black – Channell *et al.*, 2016) and (c) $\delta^{18}\text{O}$ (red – Sarnthein *et al.*, 2009; black – Channell *et al.*, 2016; grey – Lisiecki and Raymo, 2005). The relationship between the U1307 and U1308 $\delta^{18}\text{O}$ datasets shown in (c) is the product of tuning RPIs between U1307 and U1308 based on the ‘Age Model 2’ scenario for the U1307 stratigraphy proposed by Sarnthein *et al.* (2009). Application of this age model results in a poor fit between the two records prior to 3.1 Ma. Solid lines show u-channel-derived data. Dashed

Blake-Mizen *et al.* Southern Greenland glaciation and Western Boundary Undercurrent evolution recorded on Eirik Drift during the late Pliocene onset of Northern Hemisphere glaciation

lines show shipboard-derived split core data. Labels in (c) are Marine Isotope Stages. Red/black crosses indicate reversal-/RPI based tie-points.

G. Evaluation of alternative age model for ODP Site 907

Between 3.5–1.0 Ma, the age model for the Iceland Plateau Site 907 IRD record is based on tuning of the 41-kyr component of IRD abundance to orbital parameters within the constraints of the site's paleomagnetic stratigraphy (Jansen *et al.*, 2000). An alternative age model for this portion of the Site 907 stratigraphy has been proposed, however, by Lacasse and van den Bogaard (2002), based on laser probe $^{40}\text{Ar}/^{39}\text{Ar}$ dating of three single-crystal K-feldspar or biotite grains from discrete tephra layers deposited during the intensification of Northern Hemisphere glaciation (iNHG). Based on this alternative age model, the onset of persistently elevated IRD inputs to Site 907 occurred ~ 2.9 Ma, and not ~ 3 Ma (Lacasse and van den Bogaard, 2002). This alternative age model is unlikely to represent the best estimate of age-depth relationships in the Site 907 stratigraphy because: (1) it invalidates the reversal-based paleomagnetic stratigraphy for Site 907; (2) it places a large peak in IRD abundance in the Site 907 stratigraphy formerly assigned to MIS M2, ~ 3.3 Ma, by Jansen *et al.* (2000) at ~ 3.1 Ma during the mid-Piacenzian warm period, and would represent an ice-rafting event not found in any other IRD record for this time interval from the Nordic Seas and subpolar North Atlantic (e.g., this study; Jansen and Sjöholm, 1991; Kleiven *et al.*, 2002; Kneis *et al.*, 2014). Regardless, based on either age model, if a temporal offset does exist between the onset of consistently-elevated IRD deposition at Site 907 and at Site U1307 it cannot be attributed to age model uncertainty.

Blake-Mizen *et al.* Southern Greenland glaciation and Western Boundary Undercurrent evolution recorded on Eirik Drift during the late Pliocene onset of Northern Hemisphere glaciation

Table S2. Polarity reversals and excursions at Site U1307 determined in this study compared to previously published interpretations.

Hole, core, section, interval (cm)	Depth (rmcd)	Transition interval	Reversal/excursion	Age (ka)	MIS ^b	Shipboard interpretation, age (ka) ^d	Alternative interpretation, age (ka) ^e
U1307B-13H-1 128	117.11	Upper	Unnamed excursion	2236 ^a	86	-	-
U1307B-13H-2 11	117.44	Lower					
U1307A-14H-3 143	124.74	Upper	Gauss/Matuyama C2An.1n (t)	2595 ^b (2581 ^c)	103	Gauss/Matuyama C2An.1n (t), 2581	Gauss/Matuyama C2An.1n (t), 2581
U1307A-14H-4 25	125.06	Lower					
U1307A-17H-2 61	147.81	Upper	Kaena top C2An.1n (b)	3029 ^b (3032 ^c)	G21	Kaena top C2An.1n (b), 3040	Kaena top C2An.1n (b), 3040
U1307A-17H-2 97	148.17	Lower					
U1307A-17H-4 70	150.90	Upper	Unnamed excursion (?)	3060	G22	Kaena bottom C2An.2n (t), 3110 Mammoth top C2An.2n (b), 3220	-
U1307A-17H-4 147	151.67	Lower					
U1307A-18H-2 31	157.06	Upper	Kaena bottom C2An.2n (t)	3116 ^b (3110 ^c)	KM2	Mammoth bottom C2An.3n (t), 3330	Kaena bottom C2An.2n (t), 3110
U1307A-18H-2 82	157.57	Lower					
U1307A-19H-6 60	173.93	Upper	Mammoth top C2An.2n (b)	3207 ^c	KM6	Gilbert/Gauss C2An.3n (b), 3580	Gilbert/Gauss C2An.3n (b), 3580
U1307A-19H-6 120	174.53	Lower					

^aChannell and Guyodo (2004); ^bChannell *et al.* (2016); ^cOgg *et al.* (2012); ^dExpedition 303 Scientists (2006a); ^eSarnthein *et al.* (2009).

Blake-Mizen *et al.* Southern Greenland glaciation and Western Boundary Undercurrent evolution recorded on Eirik Drift during the late Pliocene onset of Northern Hemisphere glaciation

Table S3. List of revised depth-age tie-points for Site U1307 between 117.63–174.48 rmcd based on our new correlation to the Site U1308 RPI.

Depth (rmcd)	Age (ka)	Type	Chronology	Depth (rmcd)	Age (ka)	Type	Chronology
117.63	2253.84	RPI	IODP U1308 ^a	136.86	2846.71	RPI	IODP U1308 ^a
118.51	2268.70	RPI	IODP U1308 ^a	137.47	2880.09	RPI	IODP U1308 ^a
120.52	2319.38	RPI	IODP U1308 ^a	138.13	2894.64	RPI	IODP U1308 ^a
121.30	2340.27	RPI	IODP U1308 ^a	138.56	2908.58	RPI	IODP U1308 ^a
121.86	2364.20	RPI	IODP U1308 ^a	138.80	2930.08	RPI	IODP U1308 ^a
122.05	2370.48	RPI	IODP U1308 ^a	139.31	2940.14	RPI	IODP U1308 ^a
122.40	2380.28	RPI	IODP U1308 ^a	141.12	2966.67	RPI	IODP U1308 ^a
123.11	2429.18	RPI	IODP U1308 ^a	143.76	3000.41	RPI	IODP U1308 ^a
123.53	2464.37	RPI	IODP U1308 ^a	147.97	3029	Reversal	Kaena (t) ^a
123.96	2521.41	RPI	IODP U1308 ^a	151.01	3059.42	RPI	IODP U1308 ^a
124.88	2595	Reversal	Gauss-Matuyama ^a	155.41	3093.99	RPI	IODP U1308 ^a
127.32	2664.45	RPI	IODP U1308 ^a	157.33	3116	Reversal	Kaena (b) ^a
129.10	2730.77	RPI	IODP U1308 ^a	166.48	3149.18	RPI	IODP U1308 ^a
129.73	2747.82	RPI	IODP U1308 ^a	168.96	3167.34	RPI	IODP U1308 ^a
132.00	2776.11	RPI	IODP U1308	170.70	3187.76	RPI	IODP U1308 ^a
135.00	2804.08	RPI	IODP U1308 ^a	174.48	3207	Reversal	Mammoth (t) ^a
136.03	2829.58	RPI	IODP U1308 ^a				

^aChannell *et al.* (2016)

Blake-Mizen *et al.* Southern Greenland glaciation and Western Boundary Undercurrent evolution recorded on Eirik Drift during the late Pliocene onset of Northern Hemisphere glaciation

Table S4: Terrigenous grain-size analysis of Site U1307 sediments: sediment size fraction proportion (in %, ± 2 standard deviations) measured for each discrete sample* from select glacial cycles during iNHG (~ 2.8 – 2.3 Ma).

Sample	Depth (rncd)	Clay <3 μm	Very fine silt 3–10 μm	Fine-medium silt 10–32 μm	Medium-coarse silt 32–63 μm	Sand >63 μm
MIS 88 (IG)	118.30	22.38 ± 7.50	36.82 ± 5.69	25.39 ± 2.69	9.00 ± 0.87	6.40 ± 1.35
MIS 88 (IG)	118.63	17.41 ± 3.06	33.44 ± 2.52	26.57 ± 0.88	13.04 ± 1.30	9.54 ± 2.29
MIS 88 (DG)	119.00	14.74 ± 2.99	33.78 ± 1.33	26.48 ± 0.83	11.57 ± 0.87	13.42 ± 3.04
MIS 88 (G)	119.40	16.14 ± 2.83	35.31 ± 1.56	26.83 ± 2.67	12.03 ± 1.31	9.69 ± 0.96
MIS 88 (IG)	120.20	18.54 ± 4.00	38.60 ± 3.09	28.99 ± 3.08	8.55 ± 2.27	45.32 ± 1.63
MIS 92 (G)	121.24	10.92 ± 2.31	30.64 ± 4.82	32.32 ± 2.36	13.85 ± 3.13	12.26 ± 2.02
MIS 92 (G)	121.44	14.24 ± 3.10	34.37 ± 1.63	28.15 ± 3.08	12.15 ± 1.18	11.09 ± 1.00
MIS 92 (IG)	121.64	15.73 ± 2.46	40.19 ± 0.65	28.16 ± 0.88	9.56 ± 1.20	6.35 ± 0.67
MIS 92 (G)	121.88	18.30 ± 1.04	36.56 ± 0.84	24.20 ± 0.88	8.54 ± 0.78	12.40 ± 1.27
MIS 92 (DG)	122.30	10.15 ± 1.10	27.12 ± 2.55	35.57 ± 1.41	16.86 ± 2.44	10.29 ± 1.60
MIS 104 (IG)	125.82	20.91 ± 1.40	36.53 ± 1.42	30.57 ± 0.57	7.08 ± 1.44	4.89 ± 1.85
MIS 104 (DG)	126.24	14.45 ± 2.17	34.30 ± 1.82	30.07 ± 2.07	11.74 ± 1.38	9.44 ± 0.81
MIS 104 (G)	126.42	15.35 ± 1.58	30.03 ± 1.55	26.36 ± 0.87	11.74 ± 0.68	16.51 ± 1.84
MIS 104 (IG)	126.62	18.47 ± 2.97	33.27 ± 1.74	27.25 ± 0.71	11.46 ± 1.19	9.55 ± 2.13
MIS 104 (DG)	126.82	18.24 ± 4.59	31.75 ± 1.69	29.84 ± 1.20	12.95 ± 0.58	7.23 ± 0.92
MIS G12 (IG)	136.00	16.59 ± 2.19	32.41 ± 0.47	32.27 ± 0.30	13.01 ± 1.11	5.71 ± 1.47
MIS G12 (G)	136.20	13.88 ± 1.41	28.43 ± 0.99	33.82 ± 0.29	17.01 ± 0.59	6.86 ± 0.97
MIS G12 (IG)	136.90	15.18 ± 2.43	33.03 ± 2.26	33.14 ± 0.84	17.03 ± 0.53	7.29 ± 0.95

*Ten repeat grain size measurements were made on a well-mixed aliquot of each sample, and this procedure repeated on a separate subsample. Values reported for each sample are an average of all measurements made. MIS = Marine Isotope Stage. IG = interglacial, G = glacial, DG = deglacial.

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