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Amplitude and timing of the Laschamp geomagnetic dipole low from the global atmospheric ^{10}Be overproduction: Contribution of authigenic $^{10}\text{Be}/^{9}\text{Be}$ ratios in west equatorial Pacific sediments

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1 **Amplitude and timing of the Laschamp geomagnetic dipole low**
2 **from the global atmospheric ^{10}Be overproduction: contribution of**
3 **authigenic $^{10}\text{Be}/^9\text{Be}$ ratios in West Equatorial Pacific sediments.**

4

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19 Abstract

20 Authigenic $^{10}\text{Be}/^9\text{Be}$ ratios were measured along a sediment core collected in the
21 West Equatorial Pacific in order to reconstruct cosmogenic ^{10}Be production variations
22 near the equator, where the geomagnetic modulation is maximum. From 60 to 20 ka, the
23 single significant ^{10}Be production impulse recorded at 41 ka results from the geomagnetic
24 dipole low that triggered the Laschamp excursion. No significant ^{10}Be overproduction
25 signature is recorded at the age of the Mono Lake excursion (~34 ka). A compilation of
26 authigenic $^{10}\text{Be}/^9\text{Be}$ records obtained from sediments was averaged over a 1 ka window
27 and compared with the 1 ka-averaged ^{10}Be flux record of Greenland ice cores. Their
28 remarkable similarity demonstrates that the ^{10}Be production is globally modulated by
29 geomagnetic dipole variations and redistributed by atmosphere dynamics. After
30 calibration using absolute values of the virtual dipole moment drawn from paleomagnetic
31 database, the authigenic $^{10}\text{Be}/^9\text{Be}$ stack allows reconstructing the geomagnetic dipole
32 moment variations over the 20-50 ka time interval. Between 48 ka and 41 ka, the dipole
33 moment collapsed at a rate of $-1.5 \times 10^{22} \text{ A.m}^2.\text{ka}^{-1}$, which will be an interesting criterion
34 for the assessment of the loss rate of the historical field and the comparison of dipole
35 moment loss prior to excursions and reversals. After a 2 ka duration of the minimum
36 dipole moment ($\sim 1 \times 10^{22} \text{ A.m}^2$), a slow increase started at 39 ka, progressively reaching
37 $5 \times 10^{22} \text{ A.m}^2$ at 20 ka. The absence of a significant dipole moment drop at 34 ka, the age
38 of the Mono lake excursion, suggests that the duration and amplitude of the dipole
39 weakening cannot be compared with that of the Laschamp. This study provides a reliable
40 basis to model the production of radiocarbon and *in-situ* cosmogenic nuclides and to
41 improve the calibration of these dating methods.

42

43 1. Introduction

44 The understanding of past and present geomagnetic field behaviour depends on
45 the accuracy and precision of proxy records of geomagnetic field magnitude and
46 direction. Since the last geomagnetic reversal (780 ka ago), the geomagnetic dipole
47 moment appears to have been affected by repeated drops, which generate large
48 directional variations of the geomagnetic vector field [e.g. *Thouveny et al.*, 2008;
49 *Channell et al.*, 2009]. The interpreted global character of these geomagnetic instabilities
50 however strongly depends on the geographic distribution of available paleomagnetic
51 records and on the recording materials (sediments or volcanic rocks), which influence the
52 reliability of the recorded phenomena (amplitude and duration). This is illustrated by the
53 ongoing debate over the occurrence of one or two dipole lows / excursions during the 30-
54 45 ka interval: Laschamp at ~41 ka and possibly Mono Lake at ~33 ka.

55 Quasi-reversed magnetizations recorded in Laschamp and Olby lava flows,
56 (Massif central, France) provided the first evidence of a past transient reversed state of
57 the geomagnetic field [*Bonhommet and Babkine*, 1967]. The first dating by *Bonhommet*
58 *and Zähringer* [1969] using K/Ar methods performed on whole rock led to an age
59 determination between 8 and 20 ka. After almost four decades of geochronologic
60 investigations using K–Ar, thermoluminescence, $^{40}\text{Ar}/^{39}\text{Ar}$ and ^{230}Th – ^{238}U methods [e.g.
61 *Bonhommet and Zähringer*, 1969; *Condomines*, 1980; *Hall and York*, 1978; *Gillot et al.*,
62 1979; *Chauvin et al.*, 1989; *Plenier et al.*, 2007], the most reliable radiometric age
63 datasets finally led to concordant ages of 40.4 ± 2.0 ka [*Guillou et al.*, 2004] and 40.7 ± 1.0

64 ka BP [*Singer et al.*, 2009]. Meanwhile, the Laschamp excursion has been identified
65 worldwide as a large amplitude swing of the magnetization vectors, and/or as a dramatic
66 paleointensity low in lava flows (e.g. *Kristjansson and Gudmundsson*, 1980; *Roperch et*
67 *al.*, 1988; *Levi et al.*, 1990; *Mochizuki et al.*, 2006; *Cassata et al.*, 2008] and sediments
68 [e.g. *Thouveny and Creer*, 1992; *Vlag et al.*, 1996; *Lund et al.*, 2005; *Channell*, 2006].
69 These studies thus established the global extension of the geomagnetic anomaly and mark
70 it as the major geomagnetic crisis to have occurred over the last 50 ka.

71 In contrast with the Laschamp case, the occurrence and/or the age of the Mono
72 Lake (ML) excursion/dipole moment low remain controversial. *Denham and Cox* [1971],
73 while seeking a record of the Laschamp excursion at Mono Lake (California), detected
74 excursions paleomagnetic directions in sedimentary layers of the Wilson Creek
75 formation. These layers were initially radiocarbon dated at ~24 ka, i.e. at an age older
76 than the age attributed at that time to the Laschamp excursion [*Bonhommet and*
77 *Zahringer*, 1969]. The ML excursion was thus considered as distinct from the Laschamp
78 excursion. It was later described in other North-Western American lacustrine sections
79 [e.g. *Negrini et al.*, 1984; *Liddicoat et al.*, 1992; *Coe and Liddicoat*, 1994; *Hanna and*
80 *Verosub*, 1989] as well as North Atlantic sediments [e.g. *Novaczyk and Knies*, 2000;
81 *Channell*, 2006]. The age of the excursion in the Mono Lake sediment sequence has since
82 been revised from ~24 ka [*Denham and Cox*, 1971] to ~32 – 34 ka [*Negrini et al.*, 2000;
83 *Benson et al.*, 2003; *Zic et al.* 2002]. This is, however, still subject to debate. Indeed, ^{14}C
84 ages obtained on lacustrine carbonates and $^{40}\text{Ar}/^{39}\text{Ar}$ ages of sanidine crystals from the
85 Wilson Creek Formation which range from 38 to 41 ka suggest, on the contrary, that the
86 ML excursion should be assigned to the Laschamp excursion [*Kent et al.*, 2002]. The

87 correlation of the Mono Lake RPI stack [Zimmerman *et al.*, 2006] with the GLOPIS
88 record of Laj *et al.* [2004] supports this attribution.

89 Marine sediment relative paleointensity (RPI) records however, often display two
90 successive significant RPI lows attributed to the Laschamp and Mono Lake excursions
91 [Nowaczyk and Knies, 2000; Laj *et al.*, 2000, 2004; Channel *et al.*, 2006; Lund *et al.*,
92 2006]. The Mono Lake RPI low is recorded at ~34.7 ka according to the correlation of
93 the sedimentary paleoclimatic proxies with oxygen isotopes records from Greenland
94 [GICC05 age model; *NGRIP dating group 2006*, and references therein]. Volcanic
95 records of the Mono Lake excursion were also retrieved from New Zealand [Shibuya *et*
96 *al.*, 1992; Mochizuki *et al.*, 2004, 2006, 2007; Cassata *et al.*, 2008 ; Cassidy and Hill,
97 2009], Hawaii [Laj *et al.*, 2002; Teanby *et al.*, 2002] and the Canary Islands [Kissel *et al.*,
98 2011]. The case of the Mono Lake excursion, therefore, remains controversial. Both its
99 age and the amplitude of dipole moment reduction remain uncertain.

100 Deciphering the amplitude and timing of dipole moment lows can be performed
101 using methods independent from rock- and paleomagnetism, such as those based on
102 cosmogenic nuclide production recorded in sediments and ice cores. Meteoric Beryllium-
103 10 (^{10}Be , half-life: 1.387 ± 0.012 Ma [Chmeleff *et al.*, 2010; Korschinek *et al.*, 2010]) is
104 produced through nucleonic cascades in the atmosphere which result from nuclear
105 interactions between the Galactic Cosmic Rays (GCR) and the Oxygen and Nitrogen
106 atmospheric targets. Proportional to the flux of the highly energetic charged particles
107 constituting the GCR, the ^{10}Be production rate is mainly modulated over multi-millennial
108 time scales by the variability of the magnetospheric shielding dominated by the
109 geomagnetic dipole. In appropriate archives, records of ^{10}Be production rates thus

110 provide proxies of the geomagnetic dipole moment variations [e.g. *Lal*, 1988].
111 Independent from paleomagnetic methods, this approach is particularly well suited to
112 confirm or invalidate the worldwide character of reported paleomagnetic features. Early
113 works by *Elsasser et al.* [1956] and *Lal* [1988] allowed establishing an inverse
114 relationship between the globally integrated ^{10}Be atmospheric production and the dipole
115 moment magnitude. During the last decades, this relationship has been broadly confirmed
116 using numerical simulations based on purely physical models [*Masarik and Beer*, 1999,
117 2009; *Wagner et al.*, 2000b], and experimentally supported by changes of ^{10}Be
118 depositional fluxes in ice sheets [e.g. *Muscheler et al.*, 2005] and in marine sediment [e.g.
119 *Robinson et al.*, 1995; *Frank*, 2000; *Carcaillet et al.*, 2003, 2004a; *Christl et al.*, 2003]
120 records.

121 Cosmogenic nuclide production records over the time interval spanning the
122 Laschamp and Mono Lake excursions are provided by ^{10}Be and ^{36}Cl deposition flux in
123 polar ice [e.g. *Raisbeck et al.*, 1992; *Finkel and Nishiizumi*, 1997; *Yiou et al.*, 1997;
124 *Wagner et al.*, 2000 a, b; *Muscheler et al.*, 2005] and by marine sediments using $^{230}\text{Th}_{\text{xs}}$ -
125 normalized ^{10}Be fluxes [e.g. *Frank et al.*, 1997; *Christl et al.*, 2010]. Authigenic $^{10}\text{Be}/^9\text{Be}$
126 records recently obtained from marine sediments [e.g. *Carcaillet et al.*, 2004a, b; *Leduc*
127 *et al.*, 2006; *Ménabréaz et al.*, 2011] encourage further studies in other regions, especially
128 at low latitude, where the modulation of ^{10}Be production rates by the geomagnetic dipole
129 is maximum.

130

131 **2. Environmental setting and sediment description of core MD05-2920.**

132 The studied core MD05-2920 (36.67 m long; 2.51°S, 144.32°E; 1848 m water
133 depth) was retrieved with a giant piston corer during the MD148-PECTEN Cruise aboard
134 the RV/Marion Dufresne in 2005 [*Beaufort et al.*, 2005]. The coring site is located on the
135 north coast of Papua New Guinea, in the Bismarck Sea, at ~100 km off the Sepik and
136 Ramu Rivers estuaries (Figure 1.A). These large rivers and their tributaries drain through
137 erodible volcanic and igneous rock formations distributed along a steep topographic
138 profile (the altitude of these “central mountains” is > 4000 m).

139 The regional climate is dominated by the Asian-Australian monsoon system
140 [*Webster et al.*, 1998; *Wang et al.*, 2003]. The sediment discharge from rivers draining
141 the north slope of the island of New Guinea [*Milliman et al.*, 1999] is thus very
142 considerable. This area is characterized by a very narrow continental shelf (< 5 km),
143 incised principally by the Sepik submarine canyon extending from the river mouth to
144 water-depths greater than 1000m [*Cresswell*, 2000]. After the vertical divergence of the
145 surface plume at the head of the canyon, the Sepik sediments disperse along two distinct
146 routes [*Kineke et al.*, 2000]: much of the sediment is transported down the canyon via
147 near-bottom hyperpycnal flows, while a surface plume, driven by the New Guinean
148 Coastal Current, transports fine sediments eastward (during the NW monsoon) and
149 westward (during the SE monsoon) along the shelf and slope. Evidence of intermediate
150 turbid layers also suggests distal transport along isopycnal surfaces [*Kineke et al.*, 2000].
151 However, the wet conditions throughout the year limit aeolian particle transport to this
152 area, contrarily to contributions from the river inputs [*Kawahata et al.*, 2000]. This
153 depositional setting leads to sedimentation rates on the order of tens of centimeters per ka
154 [*Beaufort et al.*, 2005].

155 The MD05-2920 sequence presents good stratigraphic preservation and is mainly
156 composed of homogeneous (non-laminated) greyish olive clay, with dispersed
157 foraminifers and occasional black lenses of organic matter [*Beaufort et al.*, 2005]. The
158 proportion of the terrigenous fraction in MD05-2920 core top sediments is ~70%, and
159 that of the carbonate fraction is ~25% [*Tachikawa et al.*, 2011].

160 The sedimentary elemental composition as determined by X-ray fluorescence
161 analyses [*Tachikawa et al.*, 2011] does not present a clear glacial-interglacial variability,
162 which implies that glacial conditions have a limited influence on the hydrological cycle,
163 which is rather linked to the intertropical convergence zone.

164

165 **3. Methodology**

166 3.1 Transport correction:

167 ^{10}Be concentration in marine sediments is the result of superimposed
168 contributions. First, it depends on the atmospheric ^{10}Be flux entering the ocean and
169 reflecting its production rate in the atmosphere. ^{10}Be is primarily produced in the
170 stratosphere: this proportion was estimated at 67% by *Lal and Peters* [1967], and at 56%
171 by *Masarik and Beer* [1999]. Once produced, the particle-reactive ^{10}Be set onto aerosols
172 is integrated to the hydrological cycle and removed from the atmospheric reservoir on a
173 yearly time-scale. *Baroni et al.* [2011] calculated a ^{10}Be atmospheric residence time of ~3
174 years which is a combination of a tropospheric residence time of approximately one week
175 and a stratospheric residence time that could be as long as 6 years. This estimation is
176 slightly higher than previous ones [*Beer et al.*, 1990; *Raisbeck et al.*, 1981], and suggests,

177 according to *Heikkilä et al.* [2011], a better ^{10}Be atmospheric mixing than previously
178 assumed. The stratospheric ^{10}Be is transferred to the troposphere at mid-latitudes [e.g.
179 *Bard and Frank*, 2006]. It seems reasonable to suppose that the atmospheric transport
180 may have a negligible effect on ^{10}Be concentration in deep-sea sediments given that ^{10}Be
181 residence time in the ocean is 500 to 1000 times higher.

182 ^{10}Be content in marine sediments is also affected by oceanic reservoir effects
183 (transport, adsorption and deposition processes) that depend on Be scavenging efficiency,
184 residence time and particle composition affinities. Meaningless ^{10}Be concentrations have
185 thus to be normalized and two proxies are currently used to retrieve the cosmogenic
186 production signal: the $^{10}\text{Be}/^{230}\text{Th}_{\text{xs}}$ ratio [e.g. *Frank et al.*, 1997; *Christl et al.*, 2003,
187 2007, 2010; *Knudsen et al.*, 2008] and the authigenic $^{10}\text{Be}/^9\text{Be}$ ratio [*Bourlès et al.*, 1989;
188 *Henken-Mellies et al.*, 1990; *Robinson et al.*, 1995; *Carcaillet et al.*, 2003, 2004a, 2004b;
189 *Leduc et al.*, 2006]. Principles, as well as possible biases resulting from their use, were
190 further summarized in *Ménabréaz et al.* [2011] and references therein. In their study, the
191 authigenic $^{10}\text{Be}/^9\text{Be}$ methodology -confirmed for the first time by a $^{10}\text{Be}/^{230}\text{Th}_{\text{xs}}$ cross-
192 evaluation conducted on the same homogenized samples- has proven to reliably correct
193 for ocean secondary contributions.

194

195 3.2 Beryllium isotopes sampling strategy determined from preliminary paleomagnetic
196 results

197 A detailed paleomagnetic study will be presented in a future paper dedicated to a
198 set of several cores collected in the same area. This paper limits itself to the relative

199 paleointensity curve that has been the basis for Be sampling. Natural and artificial
200 remanent magnetizations were measured on U-Channels using a 2G cryogenic
201 magnetometer 760 SRM model coupled with an in-line Alternating Field (AF)
202 demagnetizer. The Natural Remanent Magnetization (NRM) and Anhyseretic Remanent
203 Magnetization (ARM) were measured after AF steps from 5 to 60 mT. After
204 demagnetization at 30 mT, the NRM intensity /ARM intensity ratios were taken as the
205 best proxy for the Relative PaleoIntensity (RPI) (Figure 2B). The major feature of this
206 preliminary RPI record consists in a deep minimum located at 7.8 m expressing the
207 occurrence of a geomagnetic dipole low (GDL).

208 The ^{10}Be sampling strategy was driven by the position of this GDL. The MD05-
209 2920 sequence was sampled every 30 cm from 4.67 to 9.67 m, and every 10 cm between
210 6.27 and 8.67 m.

211

212 3.3 Sample preparation and Be isotopes extraction:

213 The 34 selected samples were dried and crushed in an agate mortar. Of the
214 resulting homogenized powder, ~1g was leached using a 0.04M hydroxylamine (NH_2OH -
215 HCl) in a 25% acetic acid leaching solution [*Bourlès et al.*, 1989]. This procedure avoid
216 the leaching of detrital Be that would strongly bias the authigenic $^{10}\text{Be}/^9\text{Be}$ ratio through
217 detrital ^9Be contamination.

218 The resulting leaching solution was then split into two aliquots: one 2ml aliquot
219 was separated for natural ^9Be measurements using Flameless Atomic Adsorption
220 Spectrophotometry, and the remaining solution was spiked using 300 μl of a 10^{-3} g/g ^9Be -

221 carrier solution (Sharlau®) before undergoing the chemical extraction procedure
222 summarized below. Beryllium in the leachates was first chelated at pH 7 by
223 acetylacetone. The obtained Be-acetylacetonates were then separated using an organic
224 solvent extraction and decomposed in acid. Beryllium oxy-hydroxydes were finally
225 precipitated at pH 8 before being oxidized to BeO at 800°C to perform AMS (Accelerator
226 Mass Spectrometry) measurements [Bourlès *et al.*, 1989].

227

228 3.4 Measurements:

229 ¹⁰Be concentrations were measured using the new French AMS national facility
230 “ASTER”, operating at 5MV (CEREGE). ¹⁰Be concentrations were calculated from the
231 measured spiked ¹⁰Be/⁹Be ratios (see equation given by Ménabréaz *et al.* [2011])
232 normalized to the NIST 4325 international standard (¹⁰Be/⁹Be = 2.79 x 10⁻¹¹ [Nishiizumi
233 *et al.*, 2007]). Final ¹⁰Be concentrations were all corrected for ¹⁰Be radioactive decay
234 using the half-life determined by Chmeleff *et al.* [2010] and Korschinek *et al.* [2010].
235 Uncertainties in the measured ¹⁰Be/⁹Be ratios and in the calculated ¹⁰Be concentrations
236 may result from counting statistics and instrumental error propagation [Arnold *et al.*,
237 2010], according to the standard propagation of uncertainties equation [e.g. Taylor,
238 1997]. Chemistry blank ratios range from 7.66 10⁻¹⁵ to 1.39 10⁻¹⁴ and are at least 1000
239 times lower than the sample ratios. Measured ratios and their uncertainties are presented
240 in Table 1.

241 Natural ⁹Be concentrations were measured using a graphite-furnace atomic
242 absorption spectrophotometer equipped with a Zeeman effect background correction

243 (Thermo Scientific ICE 3400 installed at the CEREGE). The absorbance of each sample
244 was repeatedly measured at incremental steps of standard addition. The analytical
245 precision of final ${}^9\text{Be}$ concentrations was determined from the reproducibility of standard
246 addition absorptions and the fit of standard addition lines, and ranges from 0.2% to 4.6%
247 (see Table 1).

248

249 **4. The authigenic ${}^{10}\text{Be}/{}^9\text{Be}$ record**

250 Authigenic ${}^{10}\text{Be}$ concentrations vary from 9.56×10^{-15} g/g to 2.44×10^{-14} g/g and
251 authigenic ${}^9\text{Be}$ concentrations vary from 1.79×10^{-7} g/g to 2.60×10^{-7} g/g, resulting in
252 ${}^{10}\text{Be}/{}^9\text{Be}$ ratios that range from 4.57×10^{-8} to 1.20×10^{-7} . Sample concentration and ratio
253 values are listed in Table 1. The ${}^{10}\text{Be}$ concentration profile shows a unique major
254 enhancement reaching a peak at 767 cm which persists along the authigenic ${}^{10}\text{Be}/{}^9\text{Be}$
255 ratio profile (Figure 2A) in the form of ~1.7-fold increase of the ${}^{10}\text{Be}/{}^9\text{Be}$ ratio when
256 compared to the average value calculated over the time interval spanned by this record
257 (7.07×10^{-8}). This principal enhancement is recorded between ~737 cm and ~817 cm.
258 The rest of the profile exhibits reduced variability, with slightly lower values before than
259 after this major increase.

260 The RPI record, proxy of the geomagnetic dipole moment variation, and the
261 authigenic ${}^{10}\text{Be}/{}^9\text{Be}$ ratio, proxy of the cosmogenic nuclide production, are inversely
262 correlated (Figure 2). The main RPI low, located between ~850 cm and ~750 cm, and the
263 main authigenic ${}^{10}\text{Be}/{}^9\text{Be}$ ratio enhancement, located between ~737 and ~817 cm, overlap

264 to a large extent. The RPI minimum leads the authigenic $^{10}\text{Be}/^9\text{Be}$ ratio maximum by ~15
265 cm

266 Despite this slight delay which is coherent with the principle of the post-depositional
267 magnetization lock-in depth, the two paleomagnetic and geochemical signatures must be
268 interpreted as expressions of the same geomagnetic dipole low (GDL).

269

270 **5. Age Model**

271 The age model of core MD05-2920 is based on ten ^{14}C datings performed on the
272 planktonic foraminifera *Globigerinoides ruber* (white), and on the correlation between
273 the benthic $\delta^{18}\text{O}$ record and the reference benthic stack published by *Lisieki and Raymo*
274 [2005] [Figure 3 in *Tachikawa et al.*, 2011]. After correction of the local reservoir age of
275 420 ± 60 yrs [*Mc Gregor et al.*, 2008], conventional ^{14}C ages were calibrated using the
276 IntCal09 calibration set [*Reimer et al.*, 2009] and the Calib 6.1.0 software
277 (<http://intcal.qub.ac.uk/calib/>). The chronological reconstruction is presented in Figure
278 3C. Corresponding data can be found in Table 2. From these data, the sedimentation rates
279 range between 12.3 and 16 cm/ka along the studied part of the MD05-2920 sequence.

280 **6. Discussion**

281 The authigenic $^{10}\text{Be}/^9\text{Be}$ record is plotted on its time scale in Figure 4a. The
282 $^{10}\text{Be}/^9\text{Be}$ ratio peak can be seen recorded at ~40.8 ka, which allows its attribution to the
283 Laschamp-related cosmogenic nuclide production enhancement in the atmosphere (see
284 details on the chronology of the Laschamp event discussed by *Ménabréaz et al.* [2011]).

285

286 6.1 Comparison with the Laschamp ^{10}Be atmospheric overproduction records at mid and
287 high latitudes.

288 The age of the authigenic $^{10}\text{Be}/^9\text{Be}$ maximum value at ~40.8 ka remarkably
289 coincides (within uncertainties) with the age of the authigenic $^{10}\text{Be}/^9\text{Be}$ peak
290 documenting the Laschamp-related GDL recorded at ~41 ka in Portuguese Margin
291 sediment cores ([*Ménabréaz et al.*, 2011]; Figure 4b). Despite very different depositional
292 settings, these authigenic $^{10}\text{Be}/^9\text{Be}$ records are in good overall correspondence over the
293 whole studied time period. During the Laschamp GDL, the presented data indicate that
294 the ^{10}Be production was enhanced by a factor of ~1.7 at that equatorial site (enhancement
295 factors are determined from the average value over the time interval spanned by the
296 records). This agrees, within respective uncertainties, with the ~1.9-fold increase
297 measured in the authigenic $^{10}\text{Be}/^9\text{Be}$ ratio record of the Portuguese Margin [*Ménabréaz et*
298 *al.*, 2011], and also with the ~1.7-fold increase previously reported by *Carcaillet et al.*
299 [2004b] from a neighbouring core (Figure 4b). Such enhancement factors are also
300 compatible with the near-doubling of the global (i.e. latitude integrated) ^{10}Be production
301 during GDL, expected from the physically constrained algorithm [*Elsasser et al.*, 1956;
302 *Lal*, 1988] and from numerical simulations [*Masarik and Beer*, 2009] describing the
303 relation between geomagnetic moment and cosmogenic nuclides global production.

304

305 Furthermore, these sedimentary records of the ^{10}Be atmospheric overproduction
306 linked to the Laschamp GDL are concomitant with the ^{10}Be -flux peak recorded in the

307 Greenland ice cap at Summit (Figure 4) [*Muscheler et al.*, 2005], dated at ~41 ka based
308 on ice laminae counting (GICC05 timescale of the *NGRIP-dating group* [2006] and
309 references therein]. This emphasises the simultaneity of the Laschamp ^{10}Be
310 overproduction records in paleoclimatic archives at very different locations. Over the
311 entire studied time interval, the presented authigenic $^{10}\text{Be}/^9\text{Be}$ records are remarkably
312 similar both in time and in amplitude with the Greenland ^{10}Be -flux. This coherency of the
313 signals observed for different archives located at different latitudes and in such various
314 climatic and environmental conditions most likely allows ruling out the hypothesis that
315 variability in $^{10}\text{Be}/^9\text{Be}$ ratios in core MD05-2920 arises primarily from changes in the
316 continental runoff and oceanic regimes.

317 In addition to validating the pertinence of the authigenic ^9Be normalization to
318 account for secondary effects, these observations more importantly demonstrate that the
319 ^{10}Be production signal in marine sediments is a global signal because of the lack of any
320 significant latitude effects. This remarkable correspondence also confirms that the
321 Greenland ^{10}Be -flux signal over long time-scales is dominantly modulated by the
322 geomagnetic dipole moment [e.g. *Muscheler et al.*, 2005], implying that a significant part
323 of the ^{10}Be deposited in Greenland ice has been homogenized in the atmosphere. This
324 supports a ^{10}Be residence time of several years in the atmosphere [*Baroni et al.*, 2011].

325

326 6.2 A marine stacked record of the ^{10}Be production rates.

327 The authigenic $^{10}\text{Be}/^9\text{Be}$ records obtained from sedimentary sequences from two
328 different regions -West Equatorial Pacific (MD05-2920) and Northeast Atlantic (MD95-

329 2042 and MD04-2811 cores)- are normalized to their own average value and plotted on
330 their own chronologies in Figure 5 (MD05-2920 chronological data are presented in
331 Section 5 and Table 2; MD95-2042 and MD04-2811 chronological data are presented in
332 *Ménabréaz et al.* [2011]).

333 These records are compared to the Greenland ^{10}Be flux (GRIP and GISP2 ice
334 cores), normalized to its own average value and plotted in Figure 5 on its own chronology
335 as established by multiparameter counting of annual layers [*Andersen et al.*, 2006;
336 *NGRIP-dating group*, 2006; *Rasmussen et al.*, 2006]. The glacial part of this time scale
337 has an estimated associated error of 2% back to 40 ka and of 5–10% back to 57 ka. The
338 correspondence is remarkable despite noise and distortion introduced by 1) analytical
339 uncertainties, 2) chronological uncertainties (e.g. linear interpolation between ~37 ka and
340 62 ka in the MD05-2920 core), and 3) recording processes (e.g. changes in sediment
341 accumulation rate and/or sediment properties). A composite record of the normalized
342 authigenic $^{10}\text{Be}/^9\text{Be}$ ratios is thus constructed and arithmetically averaged using a 1000-
343 year sliding window offset by 500 years. Associated uncertainties (1σ) are standard
344 deviations around computed average values.

345 The normalized authigenic $^{10}\text{Be}/^9\text{Be}$ stack (Figure 6; Table 3) indicates that the
346 global 1000-year averaged ^{10}Be production rate increased by a factor of ~1.5 at ~41 ka
347 (age of the Laschamp GDL) compared to the long-term average calculated over the 20-50
348 ka. No other significant enhancement of ^{10}Be production is evidenced over the studied
349 time interval.

350 The comparison of the normalized authigenic $^{10}\text{Be}/^9\text{Be}$ stack with the Greenland
351 ice sheet ^{10}Be -record (smoothed over a 1000-year window and plotted on its own
352 timescale) evidences a remarkable similarity (Figure 6). This is especially noticeable
353 considering that both records are obtained using very different techniques and recording
354 archives. For the first time, the Greenland ^{10}Be deposition millennial-scale flux variations
355 can be compared with an authigenic $^{10}\text{Be}/^9\text{Be}$ reference record composed by low and mid
356 latitudes records. Their similarity confirms that the modulation mechanism is common
357 and that they constitute proxies of the global atmospheric ^{10}Be production. Since there is
358 no geomagnetic modulation of the cosmogenic nuclides production at high latitudes, the
359 similarity of these ^{10}Be production records confirms that the major part of the ^{10}Be
360 deposited in Greenland ice is transported from lower latitudes through atmospheric
361 homogenization [e.g. *Muscheler et al.*, 2005; *Heikkilä et al.*, 2008]. The observed
362 agreement of the Greenland ice record with oceanic sediment records located in high
363 particle flux areas (where ^{10}Be residence time with respect to scavenging is ~ 500 yrs
364 [*Anderson et al.*, 1990; *Ku et al.*, 1990]) also suggests that both signal attenuation and
365 time lags potentially resulting from an oceanic ^{10}Be reservoir effect are minimized.
366 Indeed, such settings reduce the ^{10}Be residence time in the water column and the effects
367 of surface sediment mixing by the burrowing fauna.

368

369 6.3. Construction of a ^{10}Be -derived virtual dipole moment record

370 The normalized authigenic $^{10}\text{Be}/^9\text{Be}$ stack is then used to reconstruct the
371 modulating dipole moment variation. Normalized $^{10}\text{Be}/^9\text{Be}$ values are calibrated using

372 absolute VDM values determined from absolute paleointensities measured on lava flows
373 and drawn from the literature. For coherency, the same absolute VDM values as in
374 *Ménabréaz et al.* [2011] are taken from the GEOMAGIA-50 database, because they are
375 assumed to be representative of the VDM values over the 20-50 ka time period. Since
376 ^{10}Be production rates are inversely proportional to VDM values [*Elsasser et al.*, 1956;
377 *Lal*, 1988], (1) the normalized $^{10}\text{Be}/^9\text{Be}$ maximum value (i.e. $[1.52 \pm 0.26] \times 10^{-8}$) was
378 assigned to the minimum VDM value linked to the LE (i.e. $1.06 \pm 0.05 \times 10^{22} \text{ A.m}^2$ [*Levi*
379 *et al.*, 1990]), (2) the average of the normalized $^{10}\text{Be}/^9\text{Be}$ values lower than 0.78 (mean –
380 1 sigma) was assigned to the average of the VDM values higher than 8.70 A.m^2 (mean +
381 1 sigma), and (3) the average of the normalized $^{10}\text{Be}/^9\text{Be}$ values comprised between 0.78
382 and 1.12 ([mean \pm 1 sigma] interval) was assigned to the average of the VDM values
383 comprised between 3.62 and 8.70 A.m^2 ([mean \pm 1 sigma] interval). This empirical
384 polynomial fit obtained using the absolute VDM values is compared to the curve derived
385 from *Elsasser's* algorithm, obtained using relative VDM values (as noted by *Lal*, [1992])
386 and applied to the normalized ^{10}Be production values derived from this study: their
387 similarity reinforces the validity of the proposed calibration procedure (Figure 7). The
388 polynomial fit between ^{10}Be and absolute VDM data is then applied to the whole ^{10}Be
389 dataset. The ^{10}Be -derived absolute VDM reconstruction is shown in Figure 8 (data are
390 listed in Table 3).

391 The ^{10}Be -derived absolute VDM reconstruction is compared to the reference
392 relative paleointensity stacks, evidencing an overall good agreement with the GLOPIS-75
393 curve [*Laj et al.*, 2004], particularly between 36 and 51 ka (Figure 8). A major difference
394 appears at 30-33 ka, where high RPI values in GLOPIS follow the reduction generally

395 attributed to the Mono Lake excursion. The ^{10}Be -derived VDM record only shows a
396 small amplitude reduction at ~34 ka followed by a progressive increase until ~22 ka,
397 interrupted by another transient reduction at ~26 ka. The ^{10}Be -derived VDM record
398 presents similar averaged values to the SINT 800 stack but points out a larger amplitude
399 and narrower VDM reduction at the Laschamp age, due to the SINT-800 smoothing
400 which results from stacking numerous records.

401 6.4 Geomagnetic implications

402 These cosmogenic nuclide production records and their translation as VDM
403 records contribute to the understanding of the relation between the ^{10}Be production rates
404 and the dipole moment values, and to the timing and quantification of the VDM reduction
405 accompanying the geomagnetic excursions documented in the studied time interval.
406 Indeed, the trends of the ^{10}Be -derived VDM stack result from the construction of robust
407 chronologies along the two studied sedimentary sequences and are further supported by
408 the tight correlation between the sedimentary and ice cores ^{10}Be production records.
409 Estimates of durations and rates of changes are thus enabled.

410 While paleomagnetic sedimentary records rarely provide precise estimations of
411 the duration of excursions due to 1) imperfect remanent magnetization acquisition in low
412 intensity field, 2) remagnetization after recovery of higher intensity field [e.g. *Coe and*
413 *Liddicoat*, 1994; *Roberts and Winklhofer*, 2004], and 3) latitudinal and longitudinal
414 dependency of the geomagnetic vector variation, the ^{10}Be production records provide a
415 global perspective on the timing and amplitude of the VDM variation.

416

417 *The Laschamp Excursion*

418 The global cosmogenic ^{10}Be enhancement at the time of the Laschamp excursion
419 (~41 ka) recorded in the marine reference $^{10}\text{Be}/^9\text{Be}$ record is fully concordant with that
420 reported from ^{10}Be deposition rates in the Greenland ice cores: when the dipole field
421 vanishes (VDM $\sim 1 \times 10^{22} \text{ A.m}^2$), the 1 ka average global ^{10}Be production rate is multiplied
422 by ~ 1.5 , while individual unsmoothed records report a 1.7- to 1.9- fold increase of the
423 ^{10}Be production.

424 The duration of the enhancement phase can be estimated using, either the ^{10}Be
425 value above “mean+1 σ ” which provides a maximum duration of 2500 years, or above
426 “mean + 2 σ ”, which provides a maximum duration of 1500 years. This provides the same
427 estimated duration of VDM loss as that deduced from ice core records [*Wagner et al.*,
428 2000b; *Muscheler et al.*, 2005] and from the GLOPIS-75 stack [*Laj et al.*, 2004].

429 The VDM collapse from $\sim 11 \times 10^{22} \text{ A.m}^2$ to $\sim 1 \times 10^{22} \text{ A.m}^2$ occurred between 48
430 and 41 ka, yielding an average pre-excursion VDM loss rate of $\sim -1.4 \times 10^{22} \text{ A.m}^2.\text{ka}^{-1}$.
431 The main VDM loss occurred after 44 ka with a loss rate of $\sim -1.5 \times 10^{22} \text{ A.m}^2.\text{ka}^{-1}$. It is
432 interesting to note that from archeomagnetic reconstructions [*Gallet et al.*, 2009], the
433 computed VDM loss rate of $\sim -2.7 \times 10^{22} \text{ A.m}^2$ over the last millennium appears
434 significantly higher than the Laschamp pre-excursion rates.

435

436 *The Mono Lake Excursion*

437 The ^{10}Be production reference records do not evidence any global significant
438 increase in the atmospheric production of cosmogenic nuclides at the age of the Mono
439 Lake excursion (~34 ka). The corresponding VDM value ($4 \times 10^{22} \text{ A.m}^2$) obtained from
440 our reconstruction (Figure 8) is consistent with those determined from absolute
441 paleointensities measured on lava flows (though it must be stressed that VDM values
442 computed from low paleointensity data are, by definition, biased by non-dipole field
443 contributions). In Hawaii, 9 of the 11 lavas recording the excursion yield VADM
444 estimates of $\sim 4 \times 10^{22} \text{ A.m}^2$ [Laj *et al.*, 2002; Teanby *et al.*, 2002]. In the Canary Islands,
445 the 3 studied lava flows yield VADM values of (4.3 ± 1.3) , (1.6 ± 0.3) , and $(2.5 \pm 0.8) \times$
446 10^{22} Am^2 [Kissel *et al.* 2011]. Compilation of New Zealand PI data [Mochizuki *et al.*,
447 2006; Cassidy and Hill 2009] provides a VADM value of $\sim 2.5 \times 10^{22} \text{ A.m}^2$ [Kissel *et al.*,
448 2011].

449 The short duration of the GDL associated with the Mono Lake excursion may
450 contribute to the lack of cosmogenic nuclide overproduction signature in marine archives.
451 Taking into account a (standard) bioturbation depth of 15 cm, the related overproduction
452 may not be recorded in these sequences. However, ^{10}Be ice records, despite their high
453 resolution, do not exhibit either any overproduction signature either (see Muscheler *et al.*,
454 2005), contrary to the ^{36}Cl record [Wagner *et al.*, 2000a]. The different response of these
455 two cosmogenic nuclides, similarly produced through nuclear reactions induced in the
456 atmosphere by cosmic ray particles has led to questions concerning the reliability of the
457 ^{36}Cl peak [see Delmas *et al.*, 2004].

458

459 7. Conclusion and perspectives

460 The $^{10}\text{Be}/^9\text{Be}$ record of the MD05-2920 sediment core is the first reliable
461 authigenic $^{10}\text{Be}/^9\text{Be}$ evidence of cosmogenic nuclide ^{10}Be overproduction at low latitude
462 at the age of the Laschamp excursion (41 ka). Together with other records of marine and
463 terrestrial archives it confirms the global synchronicity of the ^{10}Be overproduction in the
464 atmosphere generated by the loss of the geomagnetic dipole. The compilation of
465 authigenic $^{10}\text{Be}/^9\text{Be}$ marine records indicates that the global ^{10}Be production rates at 41
466 ka were enhanced by a ~ 1.5 factor compared to the average over the 20-60 ka interval.
467 The comparison of the authigenic $^{10}\text{Be}/^9\text{Be}$ marine stack with the Greenland ^{10}Be flux
468 record (smoothed by 1000-year averaging) evidences a good coherency of the timing and
469 amplitude of ^{10}Be production at high, mid and low latitudes. This confirms that the ^{10}Be
470 overproduction signal has a global significance, as expected from a geomagnetic dipole
471 moment loss.

472 The calibration of the sedimentary $^{10}\text{Be}/^9\text{Be}$ stack using absolute virtual dipole moment
473 values provides an independent tool to reconstruct geomagnetic dipole moment
474 variations. This allows computing the loss rate leading to the Laschamp dipole minimum
475 ($\sim -1.5 \times 10^{22} \text{ A.m}^2.\text{ka}^{-1}$). This constitutes an interesting criterion to assess the loss rate of
476 the historical field.

477 In contrast with the relevant signatures in the GLOPIS-75 relative paleointensity
478 stack and in absolute paleointensity data sets, the absence of significant cosmogenic
479 response at 34 ka suggests that the Mono Lake dipole low was hardly sufficient to trigger
480 a significant cosmogenic overproduction. This demonstrates that if the Mono lake

481 excursion really occurred at that time, the duration and amplitude of the dipole
482 weakening were very limited compared to that of the Laschamp.

483 The ^{10}Be overproduction quantified in this study constitutes a reliable basis to
484 calibrate radiocarbon production and *in situ* cosmogenic nuclides production. For
485 example, it can help to understand the atmospheric ^{14}C concentration variations recorded
486 near 41 ka and near 34 ka in delta ^{14}C series [e.g. *Hughen et al.*, 2004; *Reimer et al.*,
487 2009], which were probably produced by the Laschamp and Mono Lake geomagnetic
488 dipole lows.

489 The ^{10}Be production peak linked to the Laschamp dipole low can be used as a
490 global tie point for correlation of high-resolution paleoclimatic series obtained from high
491 quality archives.

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822 **Figure Captions :**

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824 Figure 1. A. Location map of the MD05-2920 coring site in the Bismarck Sea, on the
825 Papua New Guinea northern margin. B. Location of the marine sediment cores used in
826 this study: MD05-2020 core located North of Papua New Guinea (NPNG), MD04-2811
827 and MD95-2042 cores located on the Portuguese Margin. Location of the GRIP and
828 GISP2 ice cores at Summit (Greenland).

829 Figure 2. A. The $^{10}\text{Be}/^9\text{Be}$ record along core MD05-2920. B. The NRM and ARM
830 intensities are both demagnetized at 30mT AF step along core MD05-2920 and provide a
831 ratio considered as the best Relative Paleointensity proxy. C. $\delta^{18}\text{O}$ measured on benthic
832 foraminifer *Uvigerina peregrina* (Tachikawa et al., 2011). D. Chronostratigraphic
833 markers in MD05-2920 sediments, and the depth-age relationship. Green dots are
834 radiocarbon dated levels, and the blue dot is the MIS-3 MIS-4 transition tie point.
835 Between these points, the age model is based on linear interpolation.

836 Figure 3. (a) ^{10}Be deposition flux (10^6 atoms/cm²/yr) record at Summit (Greenland) in
837 GIPS2 and GRIP ice cores [Finkel et al., 1997; Yiou et al., 1997; Muscheler et al., 2005].
838 Authigenic $^{10}\text{Be}/^9\text{Be}$ ratios (10^{-8}) records in the Portuguese Margin (b) and Papua New
839 Guinea (c) sediments.

840 Figure 4. Records of ^{10}Be atmospheric production variations between 20 and 60 ka. Dots
841 correspond to the authigenic $^{10}\text{Be}/^9\text{Be}$ ratios from MD04-2811 (blue dots), MD05-2920
842 (red dots) and MD95-2042 (black dots). In grey, the Greenland ^{10}Be deposition flux
843 variations. All data series are normalized to their own mean values, and plotted on their
844 own chronological scales.

845 Figure 5. Variations of the ^{10}Be production in the atmosphere during the 20-50 ka period.
846 In red, the authigenic $^{10}\text{Be}/^9\text{Be}$ composite record, and the associated 1σ uncertainties. It is
847 compared with the Greenland ^{10}Be -Flux record (In grey), and its 1000-year smoothed
848 version (black curve). The Greenland record is normalized to its own mean value, and
849 plotted on its own chronological scale (see text).

850 Figure 6. Calibration of the normalized $^{10}\text{Be}/^9\text{Be}$ (proxy of the normalized ^{10}Be
851 production) values using Virtual Dipole Moments (VDM) values provided by the
852 GEOMAGIA-50 database [Korhonen *et al.*, 2008], and the polynomial fit used to derive
853 Virtual Dipole Moments (VDM): $y = 32.973 - 42.156x + 13.921x^2$. The normalized
854 $^{10}\text{Be}/^9\text{Be}$ intermediate and minimum clusters are calculated using values comprised
855 within the « mean $\pm 1\sigma$ » range, and using values less than « mean - 1σ », respectively.
856 Associated error bars correspond to the standard deviation of the values used for
857 averaging. Grey dots indicate normalized dipole moments obtained after application of
858 Elsasser's algorithm [see Lal, 1992] on the normalized $^{10}\text{Be}/^9\text{Be}$ data.

859 Figure 7. ^{10}Be -based VDM reconstruction (in 10^{22} Am^2) and associated 1-sigma
860 uncertainties, compared to paleomagnetic VADM reconstructions (GLOPIS-75 and SINT
861 800 reference records) over the 20- 50 ka interval. The grey band represents the GLOPIS-
862 75 1 sigma envelop [Laj *et al.*, 2004] plotted on the GICC05 time-scale, and the black
863 curve is the SINT 800 reconstruction [Guyodo and Valet, 1999]. Each data series is
864 plotted on its own time scale. The dotted line shows the present-day VADM value of ~ 8
865 $\times 10^{22} \text{ A.m}^2$.

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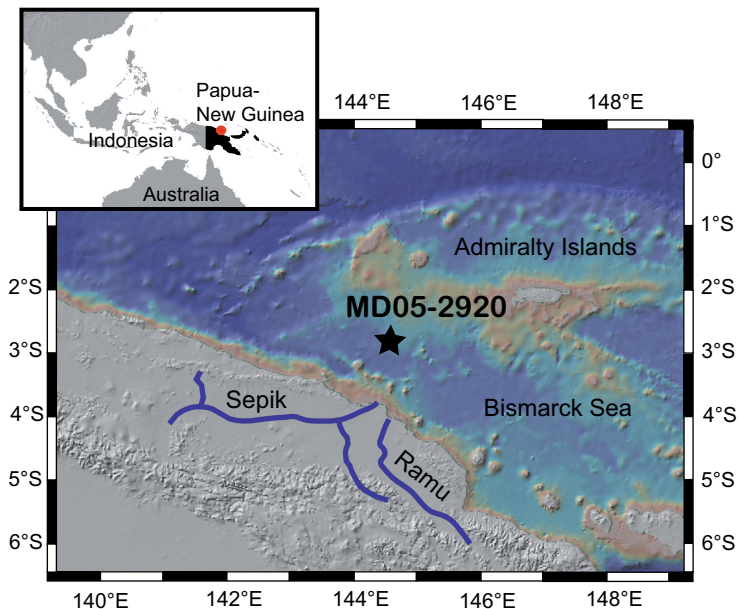
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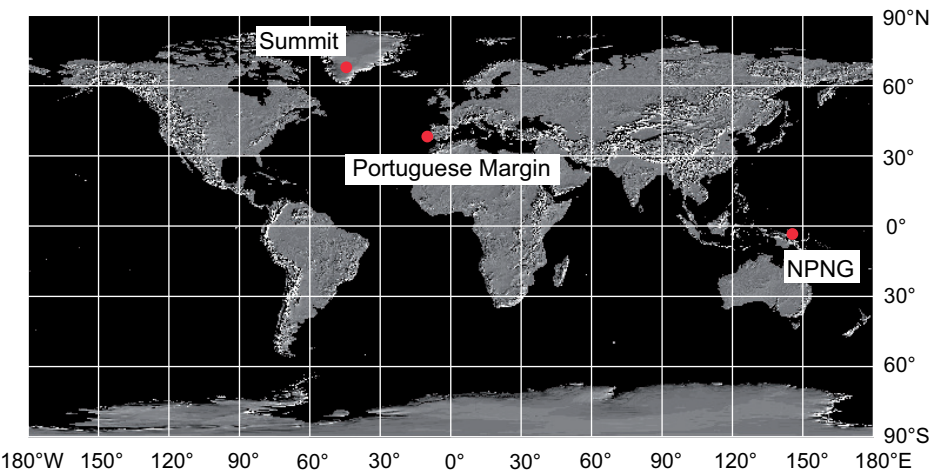
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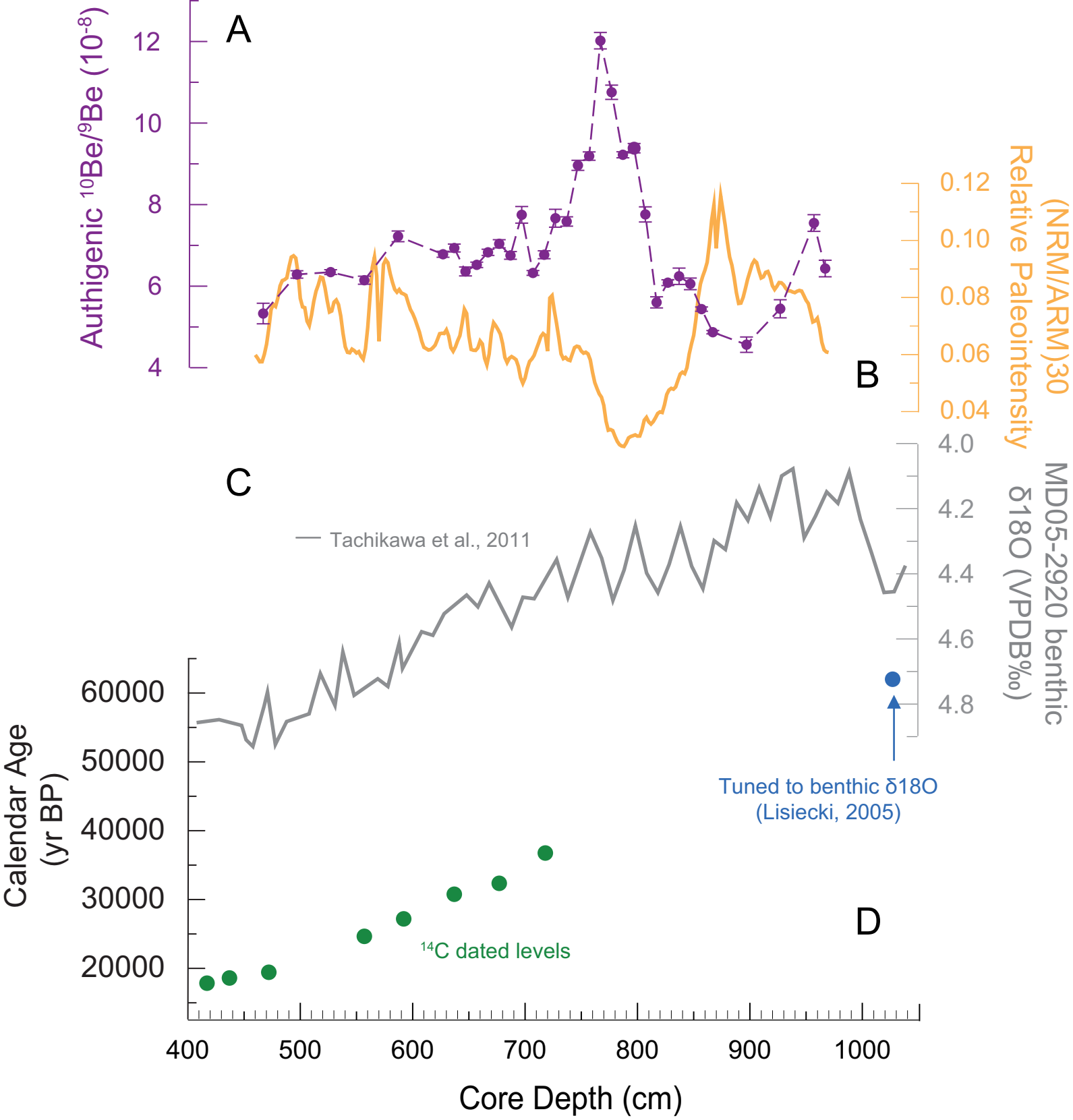
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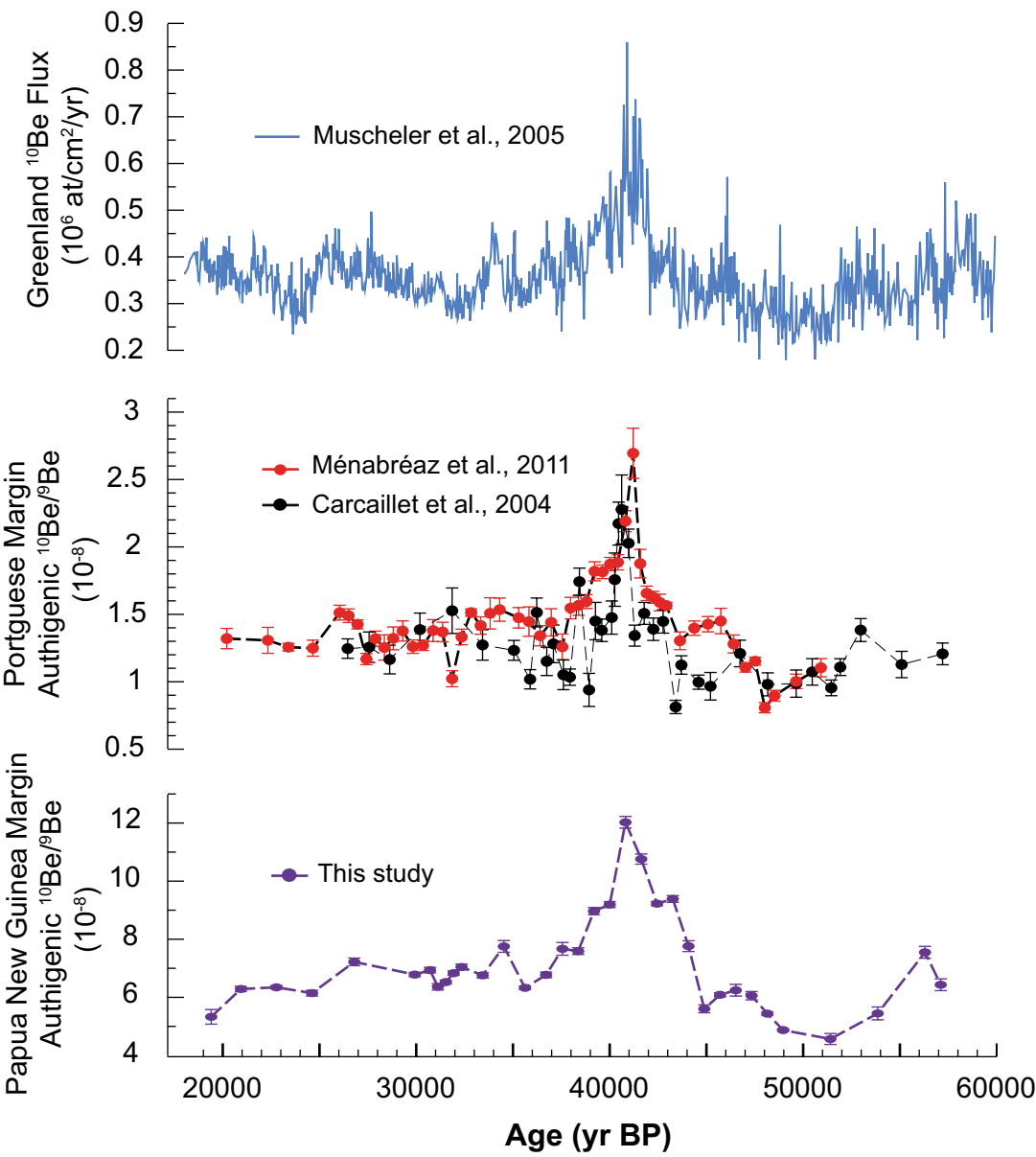
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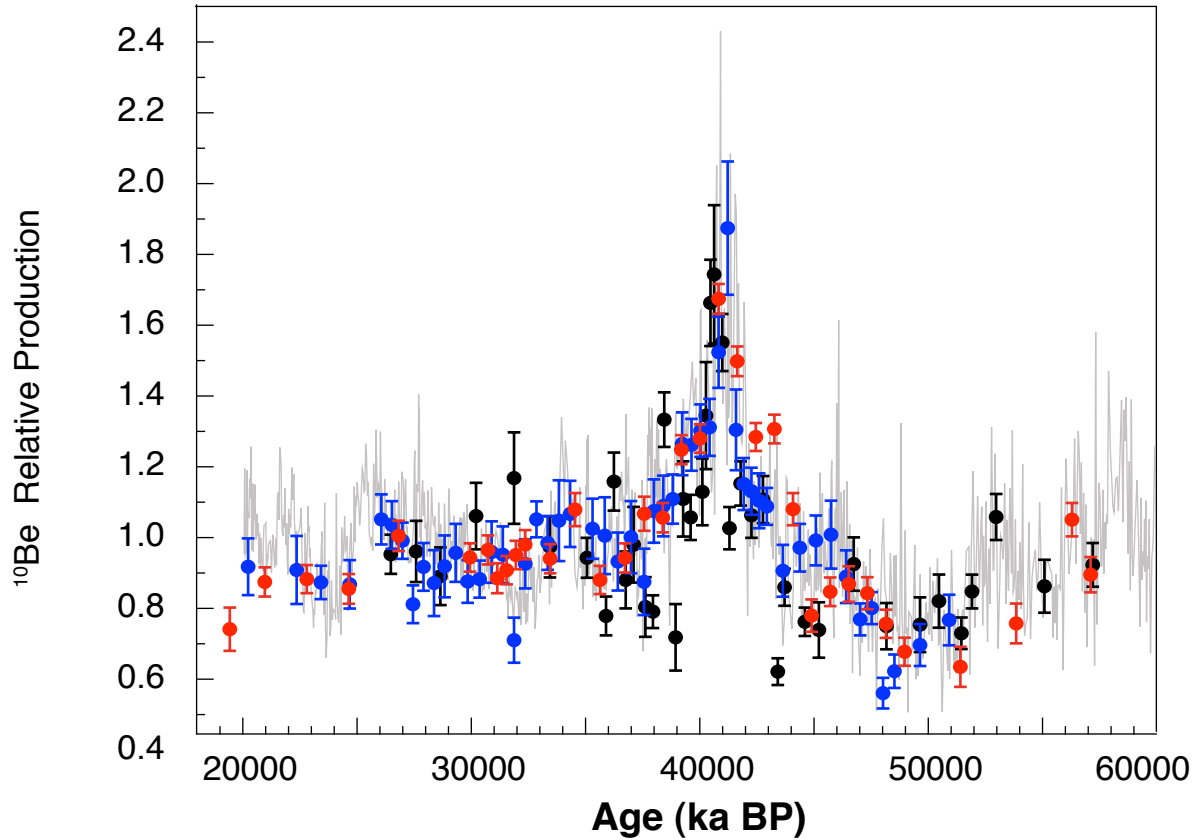


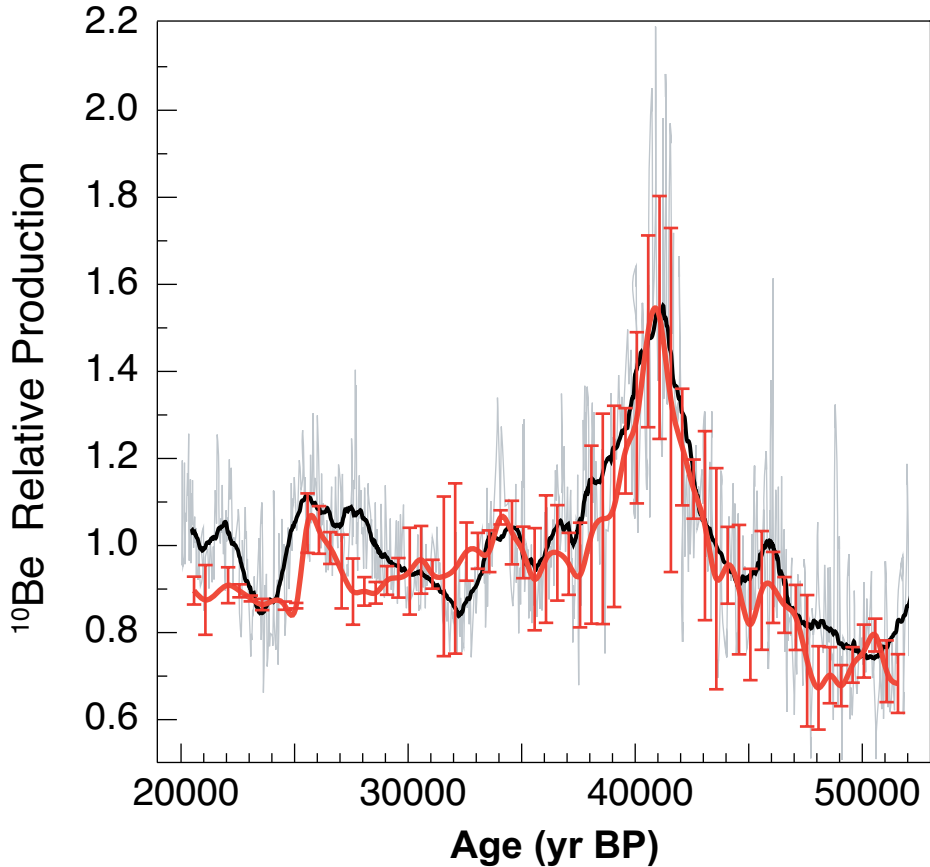
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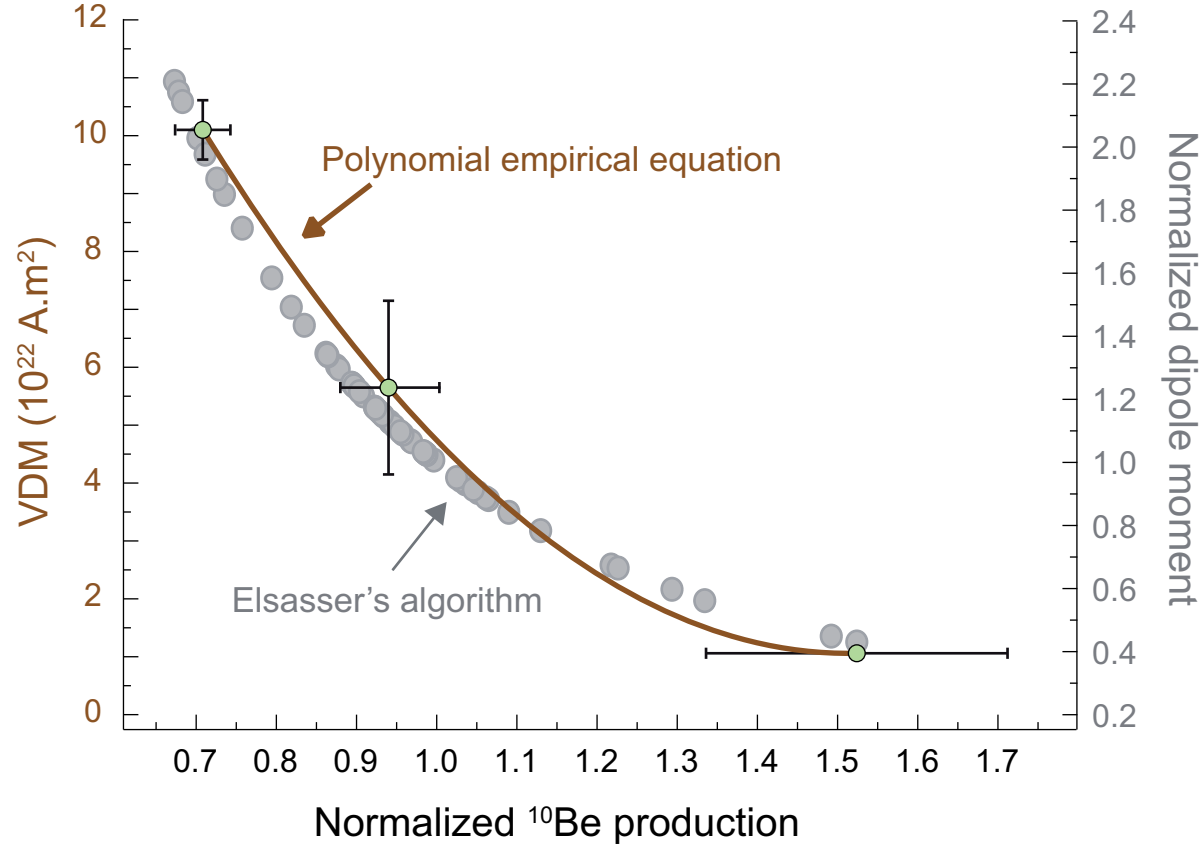












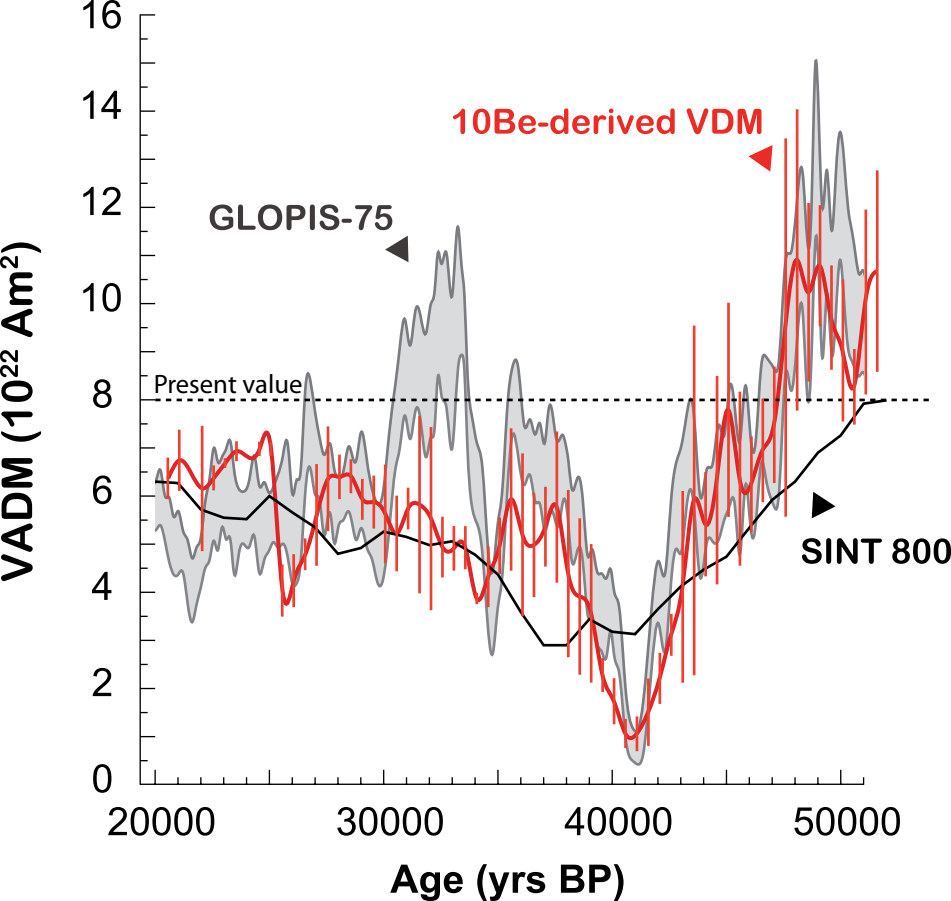


Table1: AMS measurements, Be isotopes concentrations and authigenic $^{10}\text{Be}/^9\text{Be}$ ratios of core MD05-2920 samples.

Sample: Core name- sample depth in core (cm)	Age (BP)	Measured $^{10}\text{Be}/^9\text{Be}$ (10^{-11}) *	Decay-corrected authigenic [^{10}Be] (10^{-14}g/g) *	Authigenic [^9Be] (10^{-7}g/g) *	Authigenic $^{10}\text{Be}/^9\text{Be}$ (10^{-8}) *
MD05 2920-467	19405	2.134±0.021	0.956±0.009	1.793±0.083	5.33±0.26
MD05 2920-797	20937	3.307±0.034	1.535±0.016	2.442±0.026	6.29±0.10
MD05 2920-527	22777	3.148±0.029	1.475±0.014	2.325±0.006	6.34±0.06
MD05 2920-557	24616	3.246±0.031	1.558±0.015	2.536±0.033	6.15±0.10
MD05 2920-587	26798	2.508±0.024	1.701±0.016	2.355±0.037	7.22±0.14
MD05 2920-627	29925	4.097±0.035	1.723±0.015	2.540±0.017	6.78±0.08
MD05 2920-637	30714	4.094±0.035	1.724±0.015	2.486±0.028	6.94±0.10
MD05 2920-647	31121	3.636±0.033	1.541±0.014	2.423±0.036	6.36±0.11
MD05 2920-657	31529	3.546±0.033	1.457±0.014	2.233±0.010	6.52±0.07
MD05 2920-667	31936	3.604±0.032	1.538±0.014	2.252±0.016	6.83±0.08
MD05 2920-677	32343	3.693±0.034	1.477±0.014	2.098±0.023	7.04±0.10
MD05 2920-687	33436	4.419±0.040	1.717±0.016	2.542±0.028	6.75±0.10
MD05 2920-697	34529	4.386±0.038	1.609±0.014	2.076±0.052	7.75±0.21
MD05 2920-707	35622	3.330±0.030	1.546±0.014	2.445±0.007	6.32±0.06
MD05 2920-717	36715	4.787±0.043	1.763±0.016	2.604±0.030	6.77±0.10
MD05 2920-727	37557	2.910±0.060	1.830±0.038	2.387±0.048	7.67±0.22
MD05 2920-737	38372	3.426±0.030	1.865±0.017	2.458±0.031	7.59±0.12
MD05 2920-747	39187	4.987±0.042	2.104±0.018	2.347±0.026	8.96±0.13
MD05 2920-757	40002	5.540±0.047	2.311±0.020	2.514±0.019	9.19±0.11
MD05 2920-767	40816	4.358±0.028	2.439±0.016	2.029±0.032	12.01±0.21
MD05 2920-777	41631	3.994±0.023	2.274±0.013	2.115±0.033	10.75±0.18
MD05 2920-787	42446	3.662±0.025	2.068±0.014	2.242±0.011	9.22±0.08
MD05 2920-797	43261	4.262±0.028	1.892±0.013	2.016±0.021	9.38±0.12
MD05 2920-807	44075	3.829±0.027	1.627±0.012	2.097±0.048	7.76±0.19
MD05 2920-817	44890	3.077±0.023	1.307±0.010	2.333±0.055	5.60±0.14
MD05 2920-827	45705	2.672±0.019	1.294±0.010	2.127±0.020	6.09±0.08
MD05 2920-837	46520	2.850±0.019	1.335±0.009	2.138±0.068	6.24±0.21
MD05 2920-847	47334	3.419±0.022	1.276±0.009	2.107±0.048	6.06±0.15
MD05 2920-857	48149	2.590±0.019	1.227±0.009	2.257±0.015	5.44±0.06
MD05 2920-867	48964	2.531±0.020	1.222±0.010	2.509±0.008	4.87±0.05
MD05 2920-897	51408	1.645±0.015	1.123±0.010	2.459±0.100	4.57±0.19
MD05 2920-927	53852	2.051±0.016	1.400±0.011	2.572±0.103	5.45±0.23
MD05 2920-957	56297	2.834±0.021	1.549±0.012	2.053±0.054	7.55±0.21
MD05 2920-967	57111	3.181±0.023	1.543±0.011	2.398±0.074	6.43±0.21
mean ± std. dev.			1.618±0.397	2.303±0.202	7.07±1.62
mean ± SDOM					7.07±0.28

* errors are 1 sigma, and rounded to the higher value. Sample depths are corrected from the 30 cm surface sediment void.

Table 2: Chronological data for MD05-2920 age model construction.

Sample depth in core MD05-2920 (cm) ^a	¹⁴ C Age ^b (¹⁴ C yr BP)	Calibrated Age (yr BP)	Method
40	2610 ± 35	2151 ± 206	¹⁴ C AMS ages, INTCAL09 *
80	4445 ± 35	4553 ± 271	¹⁴ C AMS ages, INTCAL09 *
119	6040 ± 40	6463 ± 191	¹⁴ C AMS ages, INTCAL09 *
198	8965 ± 45	9544 ± 251	¹⁴ C AMS ages, INTCAL09 *
253	10120 ± 35	11000 ± 252	¹⁴ C AMS ages, INTCAL09 *
302	11835 ± 45	13289 ± 196	¹⁴ C AMS ages, INTCAL09 *
322	12525 ± 40	13973 ± 267	¹⁴ C AMS ages, INTCAL09 *
352	12575 ± 45	14146 ± 396	¹⁴ C AMS ages, INTCAL09 *
417	15130 ± 60	18375 ± 130	¹⁴ C AMS ages, INTCAL09 *
437	15710 ± 60	19405 ± 197	¹⁴ C AMS ages, INTCAL09 *
472	16710 ± 60	19405 ± 197	¹⁴ C AMS ages, INTCAL09 *
557	21030 ± 90	24616 ± 409	¹⁴ C AMS ages, INTCAL09 *
592	22840 ± 90	27162 ± 609	¹⁴ C AMS ages, INTCAL09 *
637	26430 ± 130	30714 ± 557	¹⁴ C AMS ages, INTCAL09 *
677	28560 ± 160	32343 ± 715	¹⁴ C AMS ages, INTCAL09 *
718	32520 ± 240	36824 ± 623	¹⁴ C AMS ages, INTCAL09 *
1027		62000	Tuned to benthic δ ¹⁸ O **

^a All depths in core MD05-2920 are corrected for a top-core 30 cm void.

^b All ¹⁴C ages were determined by *Tachikawa et al.* [2011], and corrected for a regional reservoir age of 420 ± 60 years given by *McGregor et al.* [2008]. Errors are 2-sigma.

* *Reimer et al.* [2009]

** *Lisiecki and Raymo* [2005], *Tachikawa et al.* [2011].

Table 3: Authigenic $^{10}\text{Be}/^9\text{Be}$ marine stacked record, and corresponding ^{10}Be -based VDM record.

Age (BP)	Authigenic $^{10}\text{Be}/^9\text{Be}$ stack (normalized data)	VDM (10^{22} A.m ²)
20572	0.90 ± 0.03	6.4 ± 0.4
21072	0.87 ± 0.04	6.7 ± 0.6
22072	0.91 ± 0.10	6.2 ± 1.3
22572	0.90 ± 0.02	6.4 ± 0.3
23072	0.88 ± 0.01	6.7 ± 0.1
23572	0.86 ± 0.01	6.9 ± 0.2
24572	0.86 ± 0.01	7,0 ± 0.1
25072	0.86 ± 0.01	7,0 ± 0.1
25572	1.05 ± 0.07	4,0 ± 0.5
26072	1.04 ± 0.07	4.2 ± 0.6
26572	1.00 ± 0.04	4.8 ± 0.3
27072	0.94 ± 0.09	5.6 ± 1.1
27572	0.90 ± 0.08	6.4 ± 1.1
28072	0.89 ± 0.04	6.4 ± 0.5
28572	0.89 ± 0.02	6.4 ± 0.4
29072	0.92 ± 0.03	5.9 ± 0.4
29572	0.93 ± 0.04	5.9 ± 0.6
30072	0.94 ± 0.09	5.6 ± 1,0
30572	0.97 ± 0.07	5.2 ± 0.8
31072	0.93 ± 0.04	5.7 ± 0.4
31572	0.93 ± 0.15	5.8 ± 1.8
32072	0.95 ± 0.16	5.5 ± 1.9
32572	0.99 ± 0.06	4.9 ± 0.6
33072	0.99 ± 0.05	4.9 ± 0.5
33572	0.99 ± 0.04	4.9 ± 0.4
34072	1.06 ± 0.02	3.9 ± 0.1
34572	1.03 ± 0.08	4.3 ± 0.6

35072	0.98 ± 0.06	5,0 ± 0.6
35572	0.92 ± 0.12	5.9 ± 1.5
36072	0.97 ± 0.16	5.2 ± 1.7
36572	0.98 ± 0.11	5,0 ± 1.1
37072	0.96 ± 0.07	5.4 ± 0.8
37572	0.93 ± 0.13	5.8 ± 1.6
38072	1.02 ± 0.20	4.4 ± 1.7
38572	1.06 ± 0.22	3.9 ± 1.6
39072	1.09 ± 0.22	3.6 ± 1.4
39572	1.22 ± 0.09	2.3 ± 0.4
40072	1.29 ± 0.18	1.7 ± 0.5
40572	1.49 ± 0.21	1.1 ± 0.3
41072	1.52 ± 0.26	1.1 ± 0.4
41572	1.33 ± 0.31	1.5 ± 0.7
42072	1.23 ± 0.15	2.2 ± 0.5
42572	1.13 ± 0.08	3.1 ± 0.4
43072	1.05 ± 0.25	4.2 ± 2,0
43572	0.92 ± 0.28	5.9 ± 3.6
44072	0.95 ± 0.10	5.4 ± 1.1
44572	0.09 ± 0.15	6.3 ± 2.2
45072	0.82 ± 0.12	7.8 ± 2.2
45572	0.90 ± 0.13	6.4 ± 1.8
46072	0.90 ± 0.07	6.2 ± 1,0
46572	0.86 ± 0.07	7,0 ± 1.1
47072	0.83 ± 0.07	7.5 ± 1.2
47572	0.74 ± 0.15	9.5 ± 3.9
48072	0.67 ± 0.10	10.9 ± 3.1
48572	0.70 ± 0.06	10.2 ± 1.9
49072	0.68 ± 0.04	10.8 ± 1.3
49572	0.73 ± 0.04	9.7 ± 1.1
50072	0.76 ± 0.06	9,0 ± 1.5

50572	0.79 ± 0.04	8.3 ± 0.8
51072	0.71 ± 0.07	10,0 ± 1.9
51572	0.68 ± 0.07	10.7 ± 2.1

errors are 1-sigma