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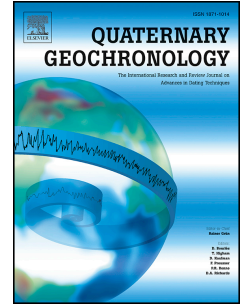
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2 **accretionary wedge sediments: relative paleointensity, oxygen isotopes and reversals**

3

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15

16 **Abstract**

17 Paleomagnetism is a powerful tool for establishing an almost continuous chronostratigraphy for

18 an entire region. When combined with other dating methods, absolute or relative, it can be used

19 to develop a regional reference chronostratigraphic framework. During the summer of 2016,

20 several piston cores were collected along the Atlantic side of the Lesser Antilles onboard the R/V

21 *Pourquoi pas?* as part of the CASEIS Expedition. Core CAS16-24PC was devoted to

22 chronostratigraphic analysis and allowed development of a quasi-continuous Quaternary record.

24 layers, we reconstructed the relative paleointensity (RPI) variations by normalizing the natural
25 remanent magnetization with the laboratory induced anhysteretic remanent magnetization. An
26 age model was developed by comparing our RPI record with the PISO-1500 stack and
27 paleomagnetic axial dipole moment model for the past 2 Myr (PADM-2M) and the planktic
28 oxygen isotopic record ($\delta^{18}\text{O}$) for core CAS16-24PC with the LR04 benthic stack. By combining
29 the $\delta^{18}\text{O}$ stratigraphy with paleomagnetic analyses, we established an age model covering the
30 Brunhes/Matuyama boundary and Jaramillo Subchron back to ~ 1.15 Ma with a mean
31 sedimentation rate of 1.7 cm/kyr. This age model complements the paleomagnetic data from
32 IODP campaigns and volcanic records, and offers almost complete inclination, declination and
33 RPI records as a local reference.

34

35 Keywords: Lesser Antilles, Quaternary, sediment, geomagnetic reversals, relative paleointensity,
36 oxygen isotopes

37

38 **1 Introduction**

39 Earth's magnetic field varies at different timescales and can undergo reversals (normal and
40 reverse polarities) that may be globally recorded quasi-simultaneously in rocks and sediments
41 (Lowrie, 2007). The synchronicity of reversals across different locations makes them important
42 for stratigraphy (Hambach et al., 2008). The last reversal occurred 773 ± 2 ka ago between the
43 normal polarity Brunhes Chron and the reversed polarity Matuyama Chron (Singer, 2014; Simon
44 et al. 2019; Channell et al., 2020). The Matuyama/Brunhes boundary (MBB) delimits the
45 beginning of the Middle Pleistocene (International Commission on Stratigraphy, 2020). In

47 in Earth's magnetic field intensity. Full vector analysis of Quaternary oceanic sediment cores
48 enables quasi-continuous reconstruction of the relative variations in the Earth's magnetic field
49 (e.g., Tauxe, 1993; Weeks et al. 1993; Stoner and St-Onge, 2007; Channell et al., 2012, 2014,
50 2019; Roberts et al., 2013; Liu et al. 2016; Deschamps et al. 2018; Simon et al. 2020). This
51 reconstruction is based on the assumption that magnetic minerals such as magnetite and titan-
52 o-magnetite within the sediment reflect the strength and the direction of the Earth magnetic field at
53 the time they were deposited. This non-destructive approach enables correlation of multiple cores
54 and stratigraphic events for entire regions. While paleomagnetism has been widely used to
55 understand past geomagnetic field fluctuations, increasing global data coverage remains a long-
56 term challenge for understanding geomagnetic field dynamics (Korte and Mandea, 2019), and for
57 chronostratigraphic purposes.

58 There is currently no complete and well-dated Quaternary magnetostratigraphic record for the
59 Lesser Antilles. Most paleomagnetic studies have focused on volcanic rocks (Carlut and
60 Quidelleur, 2000; Genevey et al., 2002; Tanty et al., 2015; Ricci et al., 2018) or Ocean Drilling
61 Program (ODP) cores that cover longer geologic time intervals (DSDP Site 502: Kent and
62 Spariosu, 1983; DSDP Leg 78A: Wilson, 1984; ODP Leg 110: Hounslow et al., 1990), but with a
63 less detailed representation of the Quaternary.

64 The main objective of this study is to establish a complete regional Quaternary
65 chronostratigraphy based on analyses of core CAS16-24PC retrieved during the CASEIS
66 expedition (May 28 to July 4, 2016; Feuillet, 2016) onboard the R/V *Pourquoi Pas?*. The first
67 step was to identify all rapidly deposited layers (RDL) using x-ray fluorescence, and physical and
68 magnetic parameters to create a continuous composite record. The second step was to reconstruct
69 a reliable relative paleointensity (RPI) record and to combine it to reversals and stable oxygen

71 Antilles.

72

Journal Pre-proof

74 The Lesser Antilles are located in the easternmost Caribbean Sea (Fig. 1). They are composed of
75 two volcanic arcs: the inactive early Eocene to Oligocene arc and the current arc that has been
76 active since 38 Ma (Briden et al., 1979; Bouysse and Westercamp, 1990). These arcs resulted
77 from subduction of the south and north Atlantic oceanic crust under Caribbean oceanic crust
78 (McDonald et al., 2000), which led to formation of basaltic volcanic arcs. The current average
79 convergence velocity is 2 cm/yr with a NNE and SSW orientation (DeMets et al., 2000; Symithe
80 et al., 2015). The current system is the product of a western arc migration following subduction
81 of the two aseismic ridges (i.e., Tiburon and St-Luce) beneath the Caribbean plate. The volcanic
82 activity is probably also related to the seismic activity (Davies, 1999; Leclerc et al., 2016).
83 Seismic fault activation or reactivation led to two periods of maximum eruptive activity during
84 the Eocene-Oligocene and the late Miocene-early Pliocene (Bouysse and Westercamp, 1990).

85 The Lesser Antilles arc basement has an asymmetric shape (Picard et al., 2006). Westward of the
86 island arc, the Grenada back-arc basin has a steep and narrow continental shelf. In contrast,
87 eastward of the island arc, the shelf is wider and limited eastwards by the accretionary wedge and
88 the trench. The insular shelf is 250 km wide in the south, and only 80-100 km wide in the north.
89 Similarly, the accretionary prism has a width of 300 km in the south that decreases northward to
90 50 km. The trench is deeper to the north (up to 8 km below sea-level). The sedimentary fill
91 consists of (1) shallow shelf sediment predominantly composed (up to 50 %) of river discharge
92 volcanoclastic sediments (the proportion increases closer to islands shore, up to 50 %) and pelagic
93 and hemipelagic carbonate sediments, (2) deep sea sediments mostly composed of pelagic,
94 hemipelagic and remobilized carbonates, and (3) volcanoclastic turbidite and tephra deposits
95 (Reid et al., 1996; Picard et al., 2006). However, the most important terrigenous sediment sources

97 contribution from aeolian dust from Africa (Reid et al., 1996; Picard et al., 2006).

98 **3 Material and methods**

99 A bathymetric map (Fig. 1) was created with the QGIS software using the General Bathymetric
100 Chart of the Oceans GEBCO_08 grid, version 20100927 (<http://www.gebco.net>) and Digital
101 Elevation Models from the 2016 CASEIS cruise (Feuillet, 2016; Seibert et al., 2020). A 100-m
102 resolution was used with a WGS84/UTM World Mercator (15.5°N) projection. Reference core
103 CAS16-24PC (Fig. 1) was retrieved during the CASEIS expedition at 4023 m depth with a
104 Calypso piston corer on board the R/V *Pourquoi Pas?* (IFREMER fleet). The core is stored in the
105 marine collection of the *Muséum national d'Histoire naturelle* (Paris, France) under the number
106 MNHN-GS-CAS16-24PC and is 19.90 m in length and was divided into fourteen 150 cm
107 sections labeled with Roman numbers I to XIV (Fig. 2-A). Each section was measured on board
108 with a Multi Sensor Core Logger (see below), then split, photographed, described and sampled
109 with u-channels (u-shaped plastic tubes with 2 x 2 cm cross-section and up to 150 cm in length)
110 for continuous paleomagnetic analysis (e.g., Weeks et al., 1993; Stoner and St-Onge, 2007). In
111 2018, the core was sampled with plastic cubes (1.9 x 1.9 x 1.9 cm) for discrete paleomagnetic
112 analyses.

113 **3.1 Multi Sensor Core Logger analysis, CT scanning and x-ray fluorescence**

114 The GEOTEK Multi Sensor Core Logger (MSCL) from ISMER (*Institut des sciences de la mer*
115 *de Rimouski*) was used on board during the expedition to measure the physical and magnetic
116 properties of the sediments (St-Onge et al., 2007). Bulk density (obtained with gamma ray
117 attenuation), P-wave velocities and volumetric magnetic susceptibility (k) were analysed at 1-cm
118 intervals on whole core sections. After splitting the core, archive halves were photographed with

120 Instruments MS2E1 point sensor for magnetic susceptibility. Diffuse spectral reflectance data
121 were also measured at 1-cm intervals using a Minolta CM-2600d spectrophotometer inline with
122 the MSCL. Data were converted into the $L^*a^*b^*$ colour space of the International Commission
123 on Illumination (CIE): L^* ranges from black (0) to white (100), a^* ranges from green (-60) to red
124 (+60), and b^* ranges from blue (-60) to yellow (+60). U-channel samples were scanned at *the*
125 *Institut national de recherche scientifique – Centre eau, terre et environnement* (INRS-ETE,
126 Québec, Canada) with a CT scanner. The resulting digital X-ray images were displayed in
127 greyscale and expressed as CT numbers, which primarily reflect bulk density changes (e.g., St-
128 Onge et al., 2007; Fortin et al., 2013). Major elements from Al to Ba were analysed on archive
129 halves at the UMR CNRS 5805 EPOC laboratory (Bordeaux, France) using an X-Ray
130 fluorescence (XRF) Avaatech core-scanner. XRF-measurements were conducted at 1 cm
131 intervals with two tube voltages (10 and 30 keV).

132 **3.2 Paleomagnetic analysis**

133 **3.2.1 Continuous magnetic measurements**

134 Continuous paleomagnetic data were acquired at 1-cm intervals on u-channel samples using a 2G
135 Enterprises 755SRM-1.65 cryogenic magnetometer at the ISMER. The natural remanent
136 magnetization (NRM) was first measured directly and then stepwise demagnetized and measured
137 after each of 17 alternating field (AF) demagnetization steps (0 to 80 mT, 5 mT increments) until
138 the residual magnetization was less than 15% of the initial magnetization. An anhysteretic
139 remanent magnetization (ARM) was then induced using a 100 mT peak AF and a 50 μ T direct
140 current (DC) bias field. The ARM was measured following the same procedure and increment
141 steps as the NRM. Finally, an isothermal remanent magnetization (IRM) and a saturation IRM
142 (SIRM) were induced in DC fields of 0.3 T and 0.95 T, respectively, using a pulse magnetizer,

144 increments) and 4 steps for SIRM (0, 10, 30, 50 mT). The data were processed with the Excel
145 spreadsheet developed by Mazaud (2005), which provided the paleomagnetic inclination,
146 declination and the associated maximum angular deviation (MAD) values calculated for the
147 characteristic remanent magnetization (ChRM; 10 to 40 mT steps). The pseudo S-ratio was
148 determined using the $IRM_{0\text{ mT}}$ induced at 300 mT divided by $SIRM_{0\text{ mT}}$ induced at 950 mT
149 ($IRM/SIRM$). This pseudoS-ratio, alike the classical S-ratio, is an indicator of the magnetic
150 coercivity and mineralogy (St-Onge et al., 2003). Values close to 1 ($IRM \approx SIRM$) indicate the
151 preponderance of low coercivity minerals, whereas lower values are the expression of higher
152 coercivity minerals ($IRM < SIRM$). The sediments from core CAS16-24PC are characterized by
153 high values of IRM and SIRM (average value of all data $>11\text{ Am}^{-1}$) and the intensity of the SIRM
154 is sometimes slightly lower than the IRM, leading to values of the pseudo-S ratio slightly higher
155 than 1 in specific intervals. This is most likely due to the difficulty of the cryogenic
156 magnetometer to perform precise measurements near the dynamic range of the instrument
157 (Roberts, 2006). As a consequence, it prevents us from using the pseudo S-ratio lonely and
158 continuously downcore, but we still use the average value in hemipelagic facies in conjunction
159 with the other magnetic and imaging data (SEM, see below) to assess the magnetic mineralogy.
160 Also, k_{ARM} is calculated by normalizing the ARM with the bias field value and is used as a
161 magnetic grain size indicator (e.g., Banerjee et al., 1981; King et al., 1982).

162 **3.2.2 Discrete magnetic measurements**

163 To confirm and refine the continuous magnetic measurements results and to improve delimitation
164 of suspected reversals, we measured paleomagnetic data on discrete samples. They were acquired
165 at 40-cm intervals along the whole core except for the uppermost 5 m (10 cm intervals) and
166 sections with suspected geomagnetic reversals (2 cm intervals). We used a 2G Enterprises

168 stepwise demagnetization (10 steps; 5, 10, 15, 20, 30, 40, 50, 60, 70 and 90 mT) of the NRM at
169 the *Institut de Physique du Globe de Paris* (IPGP).

170 Twenty-five bulk samples were selected and measured using a Princeton Measurement
171 Corporation vibrating sample magnetometer (VSM) at the IPGP to obtain hysteresis loops and
172 derived magnetic parameters corrected for high field and mass: coercive force (H_c), remanent
173 coercive force (H_{cr}), saturation magnetisation (M_s) and saturation remanence (M_{rs}). These data
174 were used to estimate the relative magnetic-grain sizes domains (single domain, multidomain and
175 vortex state; Roberts et al.; 2017) based on the Day plot (Day et al., 1977). Attention must be
176 taken when considering these results. As Roberts et al. (2018) pointed out, the Day diagram
177 indicates strict limits for magnetic domains, while these limits depend on multiple parameters
178 (e.g. magnetic grain size, mineralogy, mixing, shape). Three samples with highest magnetic
179 susceptibility values were measured using an AGICO KLY-3 Kappa bridge system to measure
180 the low-field magnetic susceptibility (k_{LF}) at high-temperatures. k_{LF} was measured in a heating-
181 cooling cycle between room temperature and 700°C in a natural atmosphere. These
182 measurements enable determination of Curie temperatures (Hrouda, 1994; Dunlop and Özdemir,
183 2007).

184 **3.2.3 Scanning electron microscopy and energy-dispersive X-ray spectroscopy**

185 The magnetic mineral fraction of five samples was manually separated from bulk powdered
186 sediments using magnets. Imaging and energy-dispersive X-ray spectroscopy (EDS)
187 microanalysis were observed with a EVO MA10 Zeiss scanning electron microscope (SEM).

188 **3.3 Stable isotopes**

190 Nineteen specimens were sampled in RDL and were removed from this study. $\delta^{18}\text{O}$ was
191 determined on 250-315 μm sized *Globigerinoides ruber* planktic foraminifera using a Dual Inlet
192 GV isoprime mass spectrometer at the *Laboratoire des Sciences du Climat et de l'Environnement*
193 (LSCE). The measurements are reported versus the Vienna Pee Dee Belemnite standard (VPDB)
194 with NBS-19 standard at $\delta^{18}\text{O} = -2.20\text{‰}$, with a mean external reproducibility (1σ) of carbonate
195 standards of $\pm 0.05\text{‰}$. Measured NBS-18 $\delta^{18}\text{O}$ values are $-23.27 \pm 0.10\text{‰}$ VPDB.

196 **3.4 Age model**

197 The chronology of the event-free core CAS16-24PC composite record is based on paleomagnetic
198 reversals, RPI variations and the $\delta^{18}\text{O}$ stratigraphy (see below). In this study, the R software
199 package Bacon 2.2 (Blaauw and Christen, 2011) was used to produce a best-fit age model using
200 Bayesian statistics with normal distributions. The parameters used were a 5 cm depth interval
201 (d.by=10) and a 2 cm/ka mean accumulation rate (acc.mean=500).

202 **4 Results**

203 **4.1 Morpho-bathymetry offshore of eastern Martinique**

204 Core CAS16-24PC was retrieved offshore of Martinique Island at the limit between forearc and
205 accretionary wedge sediments (Fig. 1). This limit follows approximately the local minimum
206 negative gravity anomaly (dotted black line in Fig. 1), while the deformation front (white dashed
207 line in Fig. 1) follows the outer prism limit (Westbrook et al., 1984; Bouysse and Westercamp,
208 1990). The core location is surrounded eastward by a deep basin down to 4400 m below sea level
209 (mbsl) and westward by the Martinique basin deeper down to 5100 mbsl. This basin is fed from
210 the south by the St. Lucia, the Caravelle, the Amerique and the Kalanina canyons, together with
211 one canyon stemming from the accretionary prism (Seibert et al., 2020). In a north-south transect,

213 wasting events and frequent turbidity currents. The core site was chosen to collect the best
214 possible chronostratigraphic reference core from the area, with the presence of carbonates for
215 $\delta^{18}\text{O}$ stratigraphy (above the carbonate compensation depth) and the least possible turbidites.
216 Nonetheless, in this volcanic setting, thin volcanoclastic layers are expected (e.g. Reid et al.,
217 1996; Picard et al., 2000; Le Friant et al., 2008).

218 **4.2 Lithology and rapidly deposited layer (RDL) determination**

219 High-definition photographic images of the 19.90 m long piston core are presented in Fig. 2-A
220 and confirm the overall homogeneity of brownish to greyish sediment. The associated CT-scan
221 images have low (dark grey) to medium (light grey) density with traces of bioturbation. This
222 background sediment is interbedded with thinner, darker sediment layers with higher density
223 (white color) that can be related to RDL. In addition to the visual core characterization, down-
224 core variations of physical and magnetic parameters, CT-number, $\ln(\text{Fe}/\text{Ca})$ XRF ratios and
225 sedimentological log established onboard are shown in Fig. 2-B. The same method used by
226 Cassidy et al. (2014) was used to decipher hemipelagic background sediments from RDL.
227 Hemipelagic sediments are identified by traces of bioturbation, fine grains (under silt size), and
228 relatively low and stable physical and magnetic parameter values. The mean ± 2 standard
229 deviation (σ) limits (whitened bands delimited by vertical red lines) of each parameter was
230 calculated, which gives a 95% (2σ) confidence level to delimit hemipelagic background
231 sediments along the core, which is consistent with the photographic and CT-scan images. Outside
232 the red lines, positive spikes in density parameters (gamma density, P-wave velocity and CT-
233 number) and magnetic susceptibility reflect the presence of coarser grains, as well as $\ln(\text{Fe}/\text{Ca})$
234 positive spikes and L^* negative spikes, which reflect darker minerals with higher iron contents.
235 Based on this analysis, magnetic susceptibility and density are the best parameters to establish the

237 low density (density around 1.29 g/cm^3 , CT-numbers between 600 and 1028 HU) and weaker
238 magnetic susceptibility (between 20 to 150×10^{-5} SI). The second facies (e.g. Fig. 3-B) has a sharp
239 contact, coarser basal sediments with high density (up to 1.7 g/cm^3), high CT-number (up to 1977
240 HU), P-wave velocity spikes (up to 1701 m/s) and high magnetic susceptibility (up to 966×10^{-5}
241 SI). The last facies is composed of silt layers with thicknesses of a few cm (e.g. Fig. 3-A)
242 interbedded with hemipelagic background sediments with similar physical and magnetic
243 parameters as facies 2, but without distinctive sharp contacts and normal grading. Facies 2 can be
244 interpreted as classical turbidites because of its sharp contact and normal grading (e.g., Bouma,
245 1964), while facies 3 can be inferred to reflect deposition from suspended particles during a weak
246 turbidity current with a slight normal grading at the base. Along the entire 24PC core, a total of
247 37 distinctive RDL were identified. Their thicknesses vary between 1 cm and 30 cm with a mean
248 thickness of about 6 cm and a 5 cm median value; they related mostly to cryptotephra and
249 reworked volcanoclastic material.

250 **4.3 Creation of an event-free composite record.**

251 One part of the core, from 10.32 m to 12.01 m, presents a specific signal that is mostly not related
252 to turbidite or mass movement deposits. This section has no high values of density indicators or
253 P-wave velocities, but L^* and the photography (section VII_{132cm} to IX_{1cm}) indicate darker
254 sediments. The XRF ratio $\ln(\text{Fe}/\text{Ca})$ has high values, which combined with magnetic
255 susceptibility and negative L^* spikes, could indicate the presence of cryptotephra and reworked
256 volcanoclastic minerals (Cassidy et al., 2014). The average magnetic susceptibility is higher than
257 the background mean + 2 standard deviations, which is the upper limit for the hemipelagic
258 sediments. However, density parameters do not indicate coarsening grain size. This implies
259 magnetic mineral concentration variation, but in the absence of evidence for the deposition of a

261 composite depth) of the piston core is composed of background hemipelagic sediments. The
262 following results are presented for this event-free stratigraphy as a function of meters composite
263 depth (MCD; Table S-1 for depth correspondence) and the magnetic mineralogy was analyzed
264 only on these hemipelagic sediments.

266 SEM imaging and EDS of the magnetic mineral fraction (Fig. 4) enabled the identification of
267 magnetic minerals and reveal grain sizes mostly below 5 μm . The elemental compositions of the
268 analyzed grains are compiled in supplementary material Table S- 2 and reveal a matrix composed
269 mostly of alumino-silicates (Sample 4) with magnetic particles mostly consisting of low-titanium
270 iron oxides grains (TiFeOx, samples 2, 3 and 5) and a few high-titanium iron oxide grains (not
271 shown here). These SEM investigations are in agreement with the VSM measurements where the
272 shapes of hysteresis loops (Fig. 5-A) are characteristic of low-coercivity ferrimagnetic minerals
273 like magnetite or titanomagnetite (Day et al., 1977; Tauxe et al. 1996). In addition, magnetic
274 susceptibility was measured every 18 seconds over a heating-cooling cycle from room
275 temperature to 700°C and back to room temperature (Fig. 5-B). The presence of different
276 magnetic minerals is revealed. High temperature curves suggest the identification of the
277 magnetite Curie temperature (Fig. 5-B) at around 580 °C (Dunlop and Özdemir, 2007). The small
278 peak between 300 °C and 370 °C can be interpreted as Ti-rich titanomagnetite (Gilder and
279 Legoff, 2005). Additionally, the pseudo S-ratio in hemipelagic sediments presents an average
280 value of 0.96. This value is typical of low coercivity minerals such as magnetite (Stoner and St-
281 Onge, 2007). While EDS measurements only reflect mineralogy of minerals bigger than 1 μm
282 and might be different from remanence carriers, the VSM and magnetic susceptibility
283 measurements were made on bulk sediment. These combined results demonstrate an overall
284 magnetic mineralogy dominated by magnetite and titanomagnetite. Such a magnetic composition
285 is typical of volcanic settings and was previously observed for example in the Tobago Basin,
286 Southeast Caribbean (Frank et al., 2016).

287 **4.5 Magnetic grain size and concentration**

289 Day-plot (Day et al., 1977), derived from hysteresis parameters (Fig. 6-A), confirms the relative
290 prevalence of magnetic grains in a vortex state. In the King plot, data for almost all samples from
291 the composite core range around the empirical 0.2 μm line, and except for 3 samples, all are
292 above the 5 μm empirical line (Fig. 6-B). Even though these absolute values should be treated
293 with caution because they were derived from synthetic magnetite (King et al., 1982), they
294 nonetheless suggest a relatively fine magnetic grain size. These results are supported by SEM
295 images (Fig. 4) that reveal particle sizes smaller than or equal to 5 μm . This can explain that the
296 $\text{NRM}_{20\text{ mT}}$, $\text{ARM}_{20\text{ mT}}$ and $\text{IRM}_{20\text{ mT}}$ values vary by less than an order of magnitude along the core
297 (Fig. 7), except from 9.24 to 10.90 MCD (10.32 m to 12.01 m) where the density parameters do
298 not exhibit coarsening grain size, while magnetic concentration increases (Fig. 2-B).
299 Furthermore, the k_{ARM}/k ratio (Fig. 7), which reflects magnetic grain size variations (Banerjee et
300 al., 1981; King et al., 1982; Stoner and St-Onge, 2007), has an overall constant trend.

301 **4.6 Magnetic remanence**

302 Magnetic u-channel measurements are presented in Figs. 8 and 9. Within all the demagnetization
303 steps, the NRM varies between $7.95 \cdot 10^{-4}$ and $2.93 \cdot 10^{-1}$ A/m, while induced magnetization ARM
304 and IRM (SIRM) vary, respectively, between $1.52 \cdot 10^{-4}$ and $8.1 \cdot 10^{-1}$ A/m, and $1.32 \cdot 10^{-1}$ and 5.29
305 A/m. From the top to 12.25 MCD, a strong and stable single-component NRM is isolated from 10
306 to 40 mT (7 demagnetization steps; Fig. 8) with a clear linear trend toward the origin of the
307 orthogonal projection (Fig. 9-A and B). The viscous component is removed easily within the first
308 2 demagnetization steps. The mean MAD is 7° and the median is 5.6° , which are indicative of
309 relatively good directional data. In addition, the inclination fluctuates around the expected
310 inclination for the site latitude for a geocentric axial dipole ($\text{GAD} = 28.28^\circ$) field. From 12.25
311 MCD to the bottom of the core, a stable single-component magnetization is isolated from 10 to

313 interpreted as reversals. While the MAD values are relatively higher below 12.25 MCD (mean
314 and median values of 18° and 15.6°, respectively), the inclinations fluctuate around the GAD
315 values for both polarities (Fig. 8). The declination variations also mirror the inclination changes.
316 In addition, discrete samples are coherent with u-channel data and improve the delimitation of the
317 reversal event at 12.78 MCD (blue dots in Fig. 8). Overall, piston core CAS16-24PC has strong
318 and stable single-component NRM's carried by PSD magnetite or titanomagnetite grains. The
319 event-free composite record thus fulfills the criteria required for relative paleointensity (RPI)
320 determinations (Levi and Banerjee, 1976; Tauxe, 1993; Stoner and St-Onge, 2007).

321 **4.7 Relative paleointensity (RPI)**

322 To reconstruct RPI variations, the NRM needs to be cleaned from the influence of lithological
323 variations by using a normalizer such as ARM, IRM or k (Tauxe, 1993). k can be affected by
324 paramagnetic grains that do not contribute to remanent magnetisation (Forster et al., 1994;
325 Brachfeld and Banerjee, 2000), so we will not use it as a normalizer in this study. Two methods
326 were compared to identify the best normalizer. The most used is the ratio method (e.g. Channell
327 et al., 1997; Channell, 1999; St-Onge et al. 2003; Deschamps et al., 2018), which was calculated
328 here using the average of seven demagnetization steps (10 to 40 mT) of the NRM normalized by
329 ARM or IRM at the same steps (10 to 40 mT). To evaluate the best normalizer, the normalized
330 NRM is plotted against the normalizer and must reveal the lowest coefficient of determination
331 (r^2). Neither normalized remanence proxy correlates with their normalizers, but the r^2 value is
332 lower for ARM (0.028) than for IRM (0.13) (Fig. 10). The second method, based on the pseudo-
333 Thellier method and recognised as the slope method (e.g. Tauxe et al., 1995; Channell, 2002;
334 Snowball and Sandgren, 2004; Xuan and Channell, 2009), uses the slope between the NRM and
335 the normalizer at different demagnetization steps. To evaluate the efficiency of a normalizer

337 normalization gives the same results as for the slope method, discrepancies exist for IRM
338 between both normalization methods. Also, the median r value is 0.99 for ARM, while it is 0.98
339 for IRM. Considering these arguments, ARM is here considered as the best normalizer and we
340 use the ratio method for normalization.

341 **5 Discussion**

342 **5.1 $\delta^{18}\text{O}$ stratigraphy and paleomagnetic dating**

343 Marine sediment cores usually have age models based on stratigraphic alignment of the depth-
344 scale $\delta^{18}\text{O}$ record with an already dated reference $\delta^{18}\text{O}$ record. We establish a first age-model by
345 matching the planktic $\delta^{18}\text{O}$ record of core CAS16-24PC with the benthic reference LR04 stack
346 (Fig. 11), which is an average of 57 globally distributed benthic $\delta^{18}\text{O}$ records (Lisiecki and
347 Raymo, 2005) using a graphic correlation technique with the Analyseries software (Paillard et al.,
348 1996). This correlation assumes a linear sedimentation rate between 2 tie-points. A total of 37 tie
349 points (Table 1) were used and have an excellent correlation ($r^2 = 0.75$). In a second stage, the
350 age-model was refined using paleomagnetic results from core CAS16-24PC (Fig. 8). Inclination
351 and declination variations indicate that core CAS16-24PC covers the Brunhes Chron and the end
352 of the Matuyama Chron, including the Jaramillo Subchron. While the paleomagnetic
353 measurements made on the u-channels were of sufficient quality to identify these reversals,
354 discrete measurements on cubes made it possible to detect the positions of reversals more
355 accurately, especially when sedimentation rates are below 10 cm/kyr (Philippe et al., 2018).
356 Thus, the 3 first polarity transitions were deduced from u-channel and discrete samples. The
357 normal polarity Jaramillo Subchron begins at around 1069 ± 12 ka at 16.32 MCD and ends at
358 around $1001 \text{ ka} \pm 10 \text{ ka}$ at 15.38 MCD (Singer, 2014). The Matuyama/Brunhes boundary (MBB),
359 occurred at 773 ka (Channell et al., 2010; Singer, 2014, Simon et al., 2019), and is recorded at

361 the PISO-1500 paleointensity stack (Channell et al., 2009). The match between the two datasets
362 (Fig. 12) is reasonable for the Brunhes Chron ($r^2 = 0.45$), while for the Matuyama Chron part it is
363 weaker ($r^2 = 0.33$). To complement these results, we also match our RPI curve against the
364 PADM2M model (Ziegler et al., 2011). Within the Brunhes Chron, two intervals are more
365 difficult to match. The first is between 100 and 150 ka, while the second interval is between 300
366 and 500 ka. However, 37 tie points (Fig. 12, green and blue dots) were derived from these two
367 correlations and are listed in Table 2 and error estimations are described in Fig. S-1. The
368 discrepancies between the reference curves and Lesser Antilles RPI before the MBB are likely to
369 be the result of the lower quality data, as illustrated by the demagnetization curves (Fig. 9-D) and
370 high MAD values (Fig. 8). From 900 to 1151 ka, the lack of continuous data points does not
371 permit use of the RPI record as a sole chronostratigraphic marker.

372 **5.2 Final age model and paleomagnetic record**

373 Quaternary sediments can contain numerous time constraints such as geomagnetic polarity
374 reversals, geomagnetic excursions, $\delta^{18}\text{O}$ and RPI stratigraphy (Richards and Andersen, 2013).
375 The combination of RPI and $\delta^{18}\text{O}$ stratigraphy has been used in several studies to establish a
376 stronger age model than with RPI or $\delta^{18}\text{O}$ alone (e.g., Channell et al., 1997, 2009, 2012; Stoner
377 and St-Onge, 2007; Xuan et al., 2016). Our age model (Fig. 13) was generated using 3 reversals,
378 37 tie points derived from the reconstructed RPI record and 37 tie points derived from the $\delta^{18}\text{O}$
379 stratigraphy. This age model has an age of ~ 1.15 Ma at the base of the core (17.33 MCD). The
380 mean sedimentation rate varies around 1.67 cm/kyr, ranging from 1.70 cm/kyr during the
381 Brunhes Chron and 1.58 cm/kyr during the Matuyama Chron. This change between the
382 Matuyama and Brunhes Chrons might reflect change in the astronomical control of paleoclimate
383 across the mid-Pleistocene transition (Schmieder et al., 2000). The sedimentation rate in this age-

385 presence of thicker RDL (> 5 cm), and fewer $\delta^{18}\text{O}$ and paleomagnetic tie points. Nonetheless, the
386 low sedimentation rates are similar to published data from the accretionary prism sediments with
387 Brunhes Chron sedimentation rates between 1 and 3 cm/kyr (Damuth, 1977; Reid et al., 1996)
388 and they are slightly lower than the sedimentation rate reported at 3 km landward of the Barbados
389 subduction deformation front DSDP Site 541 (Wright, 1984), which is about 5 cm/kyr for the
390 Quaternary epoch. This age-depth model is validated by comparison of our data with the stacks
391 mentioned above (Fig. 14).

392 Long sedimentary paleomagnetic records have been published in the equatorial to sub-equatorial
393 Indian Ocean (e.g. Meynadier et al., 1994; Oda et al., 2000) and Pacific Ocean (e.g. Shackleton et
394 al., 1990; Yamazaki and Oda, 2005), and in the North and South Atlantic Ocean (e.g. Haag,
395 2000; Stoner et al., 2003; Hofmann and Fabian, 2007). We establish here a new
396 chronostratigraphy for the sub-equatorial Atlantic Ocean based on reconstruction of RPI,
397 geomagnetic polarity reversals and $\delta^{18}\text{O}$ stratigraphy for the last ~1.15 Ma (Fig. 13). The low
398 sedimentation rates prevent for detailed study of the MBB or the Jaramillo Subchron, but the
399 composite record provides the first late Quaternary magnetostratigraphy (Fig. 14, red curve, and
400 Fig. S-2) in this area and complements younger (Frank et al., 2016) and older (Wilson, 1984;
401 Hounslow et al., 1990) records.

402 Our results make a new contribution to the global coverage of paleomagnetic data by adding a
403 detailed and almost continuous late Quaternary paleomagnetic record from the Lesser Antilles
404 subduction zone. The chronostratigraphy established here will be useful to correlate and date
405 numerous cores from the CASEIS campaign from this area (Seibert et al., 2020), as well as to
406 date the RDL triggered by earthquakes (e.g. seismo-turbidites, Heezen and Ewing, 1952) and
407 tsunamis (e.g. turbidite-homogenite complexes, San Pedro et al., 2017) in these cores.

Journal Pre-proof

410 The combination of strong chronostratigraphic markers based on two magnetic reversals and
411 indirect dating methods based on $\delta^{18}\text{O}$ and RPI correlation enable the construction of an age
412 model for core CAS16-24PC. However, in spite of the quality of the chronology proposed, our
413 study has three limitations. First, we did not use any absolute dating methods (such as
414 radiometric or luminescence dating). Second, reference records used in our study already have
415 their own limitations and uncertainties. For the benthic LR04 stack, the uncertainties are about 6
416 ka from 3 to 1 Ma and 4 ka for the last Ma (Lisiecki and Raymo, 2005). The PISO-1500 stack is
417 comparable to LR04 (Channell et al., 2009), while the PADM2M age model has an estimated
418 uncertainty of 5-10 ka (Ziegler et al., 2011). Third, low sedimentation rate can limit the resolution
419 of the record. However, the mean sedimentation rate in core CAS16-24PC is 1.7 cm/kyr what is
420 similar to cores MD97-2140 and MD97-2143 used in the construction of the PISO-1500 stack
421 (Channell et al. 2009). Considering these three limitations, the uncertainty of the reconstructed
422 RPI is estimated at 10 ka on continuous measurements. Some periods of time were not recorded
423 (for example in core breaks or between the sections) and some tie-points were found at the limit
424 of these hiatuses and imply a greater uncertainty.

425 **6 Conclusions**

426 Magnetic, physical and chemical properties of the sediments enable identification of rapidly
427 deposited layers and makes it able to build an event-free 1.51 Ma composite record for core
428 CAS16-24PC. This sediment acquired a natural remanent magnetization characterized by a
429 strong, well-defined, stable single component magnetization carried by fine-grained magnetite
430 and titanomagnetite. These results allow for reconstructing a reliable relative paleointensity
431 record for the last 900 ka, including the Brunhes/Matuyama boundary. This geomagnetic reversal

433 also record the limits of the Jaramillo Subchron.

434 The $\delta^{18}\text{O}$ stratigraphy combined with the reconstructed relative paleointensity and 3 reversal
435 events recorded in core CAS16-24PC allow establishment of an age-depth model for the last
436 ~1.15 Ma. This age model will be used to investigate earthquakes and tsunami recurrence in
437 Lesser Antilles forearc and accretionary wedge sediments. Finally, even though the reference
438 core was selected to avoid the recording of rapidly deposited layers, we have established a rapid
439 and non-destructive method to identify thin turbidites and tephra layers based on the combination
440 of several magnetic, physical and geochemical properties.

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449 **Data availability:** [https://data.mendeley.com/datasets/wv2bzwc98p/draft?a=3f1a879d-218d-](https://data.mendeley.com/datasets/wv2bzwc98p/draft?a=3f1a879d-218d-4a7a-9d2d-276cdf70d0ce)
450 [4a7a-9d2d-276cdf70d0ce](https://data.mendeley.com/datasets/wv2bzwc98p/draft?a=3f1a879d-218d-4a7a-9d2d-276cdf70d0ce) (Reserved DOI: 10.17632/wv2bzwc98p.1)

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- 708

710 Fig. 1. Bathymetric map of the middle Lesser Antilles Arc, western Atlantic Ocean using GEBCO and
711 CASEIS data. Red and yellow lines highlight, respectively, continental slope canyons and forearc
712 basin canyons. The local minimum gravity anomaly limit (black dotted line), the N 294°E strike
713 of the Tiburon ridge (dashed black lines) and the deformation front (white dashed line with teeth)
714 are derived from Bouysse and Wastercamp (1990). The location of reference core CAS16-24PC
715 is indicated in purple.

716 Fig. 2. A: Full core high-resolution photography with CT-scan images (left: TOP of the core, right:
717 BOTTOM of the core). B: MSCL and XRF data comprising from left to right: P-wave velocity,
718 density, magnetic susceptibility k , L^* , CT-number, and $\ln(\text{Fe}/\text{Ca})$. The shipboard grain size log
719 is also plotted on the right, including red marks that are pure hemipelagic background sediments
720 used to delimit the background signal with $\pm 2\sigma$ (whitened band between red straight lines. RDL
721 are highlighted as grey horizontal intervals.

722 Fig. 3. A: Brownish hemipelagic background sediments defined with low and stable magnetic
723 susceptibility (k), CT-number, density and P-wave velocity, interbedded with a RDL (≈ 5 cm)
724 between 2.70 m and 2.65 m (yellow band). B: Light brown hemipelagic background sediments
725 are defined by low and stable k , CT-number, density and P-wave velocity data. A well-defined
726 RDL between 6.24 m and 6.14 m is darker (dark brown to gray color) with an erosive contact,
727 normal grading (high CT-number, P-wave velocity, density) and high k .

728 Fig. 4. SEM images for representative bulk sample VIII-1137. The left-hand image is a general view
729 of the sample. The right-hand image is focused on small particles ($< 10 \mu\text{m}$) and matrix analysis.
730 Blue bands are $10 \mu\text{m}$ scales. Red objects are elemental analyses, as referenced in Table S1 with
731 the same numbers.

733 loops are shown in red. B: Magnetic susceptibility vs. high temperatures. Red curves for heating,
734 while blue curves are for cooling.

735 Fig. 6. A: Day plot (Day et al., 1977). Red labeled dots are data for the samples presented in figure 5.
736 B: King plot (King et al., 1982).

737 Fig. 7. Magnetic properties of the pelagic sediments (event-free composite record) with concentration,
738 and magnetic grain size-dependent parameters on the left (NRM20 mT, ARM20 mT and IRM20
739 mT). k_{ARM}/k indicates magnetic grain size variations for a magnetic assemblage dominated by
740 magnetite.

741 Fig. 8. Paleomagnetic data including from left to right: ChRM10-40 mT, ARM, IRM and SIRM
742 demagnetization steps. Inclination (red line is the \pm GAD inclination at the site latitude) and
743 declination (blue dots are data for the discrete samples). MAD values $> 15^\circ$ are indicated in dark
744 red and $> 20^\circ$ in red.

745 Fig. 9. Typical demagnetization curves and orthogonal projections of hemipelagic sediments. Filled
746 blue dots represent the range of demagnetization steps used to calculate the characteristic
747 remanent magnetisation (ChRM).

748 Fig. 10. Comparison of the RPI estimated with the average ratios (superposed red and green curves)
749 and slope methods (red and green curves) in the scatter plots on the left, the correlation is reduced
750 for ARM. Demagnetization curves for NRM, ARM and IRM are presented on the right-hand
751 side.

752 Fig. 11. Planktic foraminiferal $\delta^{18}\text{O}$ record for core CAS16-24PC (red curve) correlated to the LR04
753 benthic stack in blue (Lisiecki and Raymo, 2005). Black crosses represents the tie-points
754 determined using the AnalySeries software (Paillard et al. 1996).

756 2009) in black. Upper middle: RPI record for core 24PC ratio-method normalized by ARM in
757 red. Lower middle: paleomagnetic axial dipole moment (PADM2M) model (Ziegler et al. 2011)
758 in grey. Notable tie-points are indicated with green dots (I-1 to I-18) and blue dots (P-1 to P-16).
759 Bottom: compilation of PISO-1500, PADM2M and RPI record (colors are the same as above
760 with the violet curve derived from the developed slope method).

761 Fig. 13. Age-depth model developed using the R package Bacon 2.2 (Blaauw and Christen, 2011) with
762 linear interpolation using 3 paleomagnetic reversals (grey dots) and 37 tie-points (blue and green
763 dots) from RPI (see Fig. 12) and 37 tie-points (red dots) from $\delta^{18}\text{O}$ stratigraphy (see Fig. 11).
764 Age-errors used for each tie point correspond to half the distance samples (see Fig. S-1).
765 Sedimentation rates are derived from the age-depth model, including RDL (gray dotted lines). B
766 = Brunhes Chron, M = Matuyama Chron, J = Jaramillo Subchron.

767 Fig. 14. Planktic foraminiferal $\delta^{18}\text{O}$ record for core CAS16-24PC (blue curve) and RPI record (red
768 curve) using the age-depth model developed in this study (Fig. 13) compared with the LR04
769 benthic (upper black curve), PISO-1500 (lower black curve) and PADM2M (grey curve) stacks.
770 Odd-numbered MIS are highlighted with grey bands. B = Brunhes Chron, M = Matuyama Chron,
771 J = Jaramillo Subchron.

772 Table 1. List of $\delta^{18}\text{O}$ tie points with isotopic composition of the sample and the corresponding LR04
773 point with age, including the error used in the age-depth model.

774 Table 2. List of RPI tie points determined against PISO-1500 (I1-18) and against PADM2M (P1-16),
775 including the error used in the age-depth model.

776

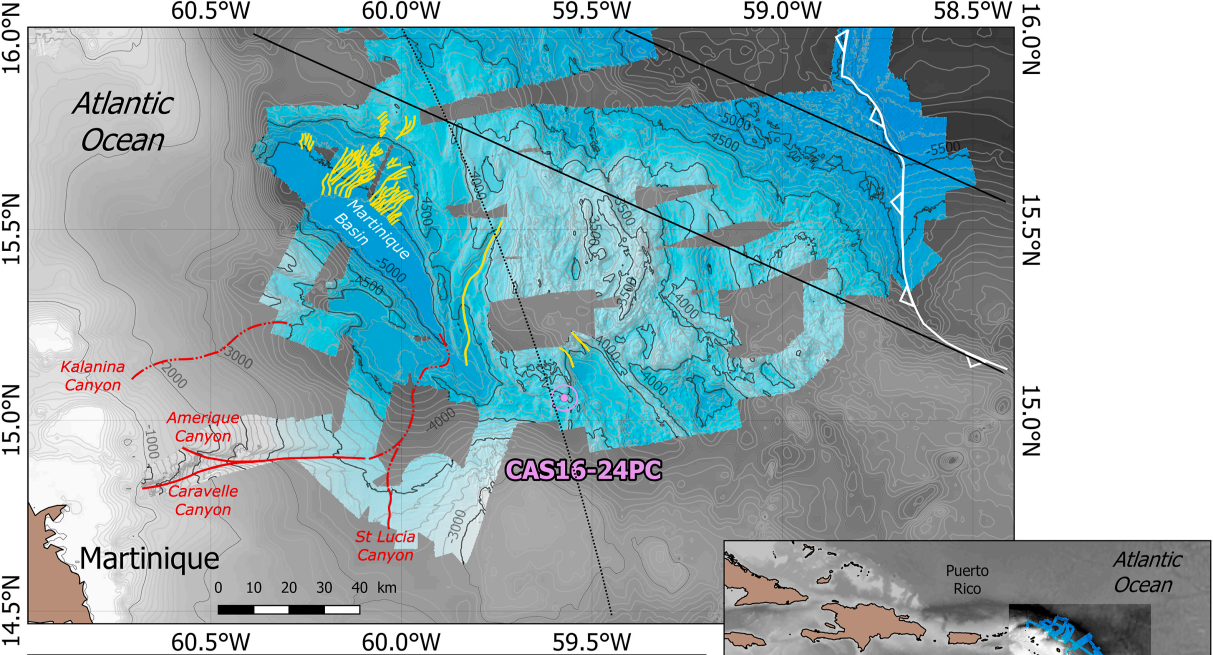
- 778 Figure S-1. Schematic explanation of age-error used for the age-depth model. The half-time gap
779 between a tie-point and its surrounding values is represented in green and purple segments. The
780 longest half-time gap (green segments) is set as this specific tie-point age-error.
- 781 Figure S-2. Inclination, declination and relative paleointensity records for core CAS16-24PC.
- 782 Table S-1: Meter composite depth (MCD) correspondence with original coring depth in meters.
- 783 Table S-2. Major elemental percentages obtained by energy dispersive X-ray spectroscopy (EDS) in
784 samples indicated in Fig. 4.

Composite depth (m)	$\delta^{18}\text{O}_{\text{planktic}} (\text{‰})$	Age (ka)	LR04 $\delta^{18}\text{O}_{\text{benthic}} (\text{‰})$	Error (\pm ka)
0.05	-1.84	4	3.3	3.75
0.9	-0.41	60	4.60	2.5
1.87	-1.72	124	3.27	2.5
2.34	0.25	139	4.87	5
2.95	-0.76	188	4.46	7
3.15	-1.45	203	3.61	4
3.65	-0.95	229	4.02	5
3.85	-1.16	240	3.44	5
4.27	-1.1	285	3.84	4.5
4.67	-1.05	296	4.23	5
4.87	-1.28	317	3.89	5
5.07	-1.74	326	3.23	3
5.37	-0.28	343	4.8	3.5
5.98	-1.18	384	4.08	3.5
6.28	-1.6	405	3.11	4.5
7.13	-0.49	476	4.3	3.5
7.49	-1.31	491	3.47	4.5
7.79	-0.82	510	4.14	3.5
7.94	-0.98	525	3.92	11
8.14	-0.52	547	4.53	11
8.8	-1.66	579	3.57	7
9.42	-1.74	614	3.53	3.5
9.72	-0.2	634	4.93	3.5
10.29	-0.44	676	4.4	3.5
10.99	-1.34	696	3.5	4
11.26	-0.29	720	4.66	4
11.7	-0.79	740	4.11	2.5
11.94	-0.31	756	4.59	4
12.23	-1.31	780	3.48	4
12.81	-1.38	822	3.9	6.5
13.11	-1.37	860	3.45	6.5
14.46	-2.12	954	3.39	3
15.7	-1.17	1022	3.94	3
16.31	-1.41	1068	3.66	8
16.90	-0.68	1126	4.5	4
17.26	-1.51	1148	4.02	9

Table 1. List of $\delta^{18}\text{O}$ tie points with isotopic composition of the sample and the corresponding LR04 point with age, including the error used in the age-depth model.

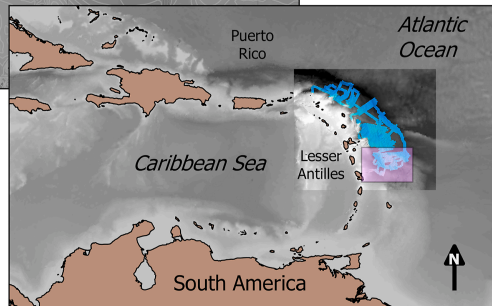
Composite depth (m)	RPI	Age (ka)	Models (10^{22} Am ²)	Error (\pm ka)
0.36 P-1	0.94	20	4.59	2
0.51 P-2	1.52	23	6.01	2
0.68 P-3	1.13	41	3.48	2
1.11 P-4	0.78	65	5.07	2
1.34 I-1	1.26	92	9.62	1
1.41 I-2	0.83	99	3.59	1
2.14 P-5	2.01	139	7.10	2
2.85 P-6	2.23	169	7.10	2
3.07 I-3	0.46	194	2.34	2
3.28 P-7	1.05	203	6.32	1
3.28 I-4	1.05	205	9.24	1
3.63 P-8	1.07	219	5.52	2
3.72 I-5	1.62	231	13.11	1
3.88 P-9	1.10	242	5.16	2
4.43 P-10	0.50	294	4.25	1
5.52 I-6	2.45	348	13.51	2
6.07 I-7	0.68	394	10.81	1
6.44 I-8	1.51	415	11.04	2
7.85 I-9	1.92	510	10.46	2
8.70 I-10	2.16	569	12.34	1
9.12 P-11	0.50	596	4.69	1
9.53 P-12	0.53	617	5.14	2
10.33 P-13	1.59	667	8.21	1
10.55 I-11	0.48	685	3.97	1
11.19 P-14	0.73	712	5.55	1
11.38 I-12	1.83	723	9.48	1
12.24 I-13	0.36	771	1.41	2
13.47 I-14	1.72	871	8.06	1
13.60 I-15	0.58	890	2.62	2
13.87 I-16	2.70	901	7.68	1
14.44 I-17	0.13	936	2.30	2
14.75 I-18	0.88	961	10.32	1
15.48 I-19	0.87	1001	11.27	2
16.01 I-20	0.94	1062	13.21	1
16.33 I-21	0.08	1071	1.05	2
16.57 P-15	1.07	1106	4.76	1
16.74 P-16	0.38	1122	2.92	2

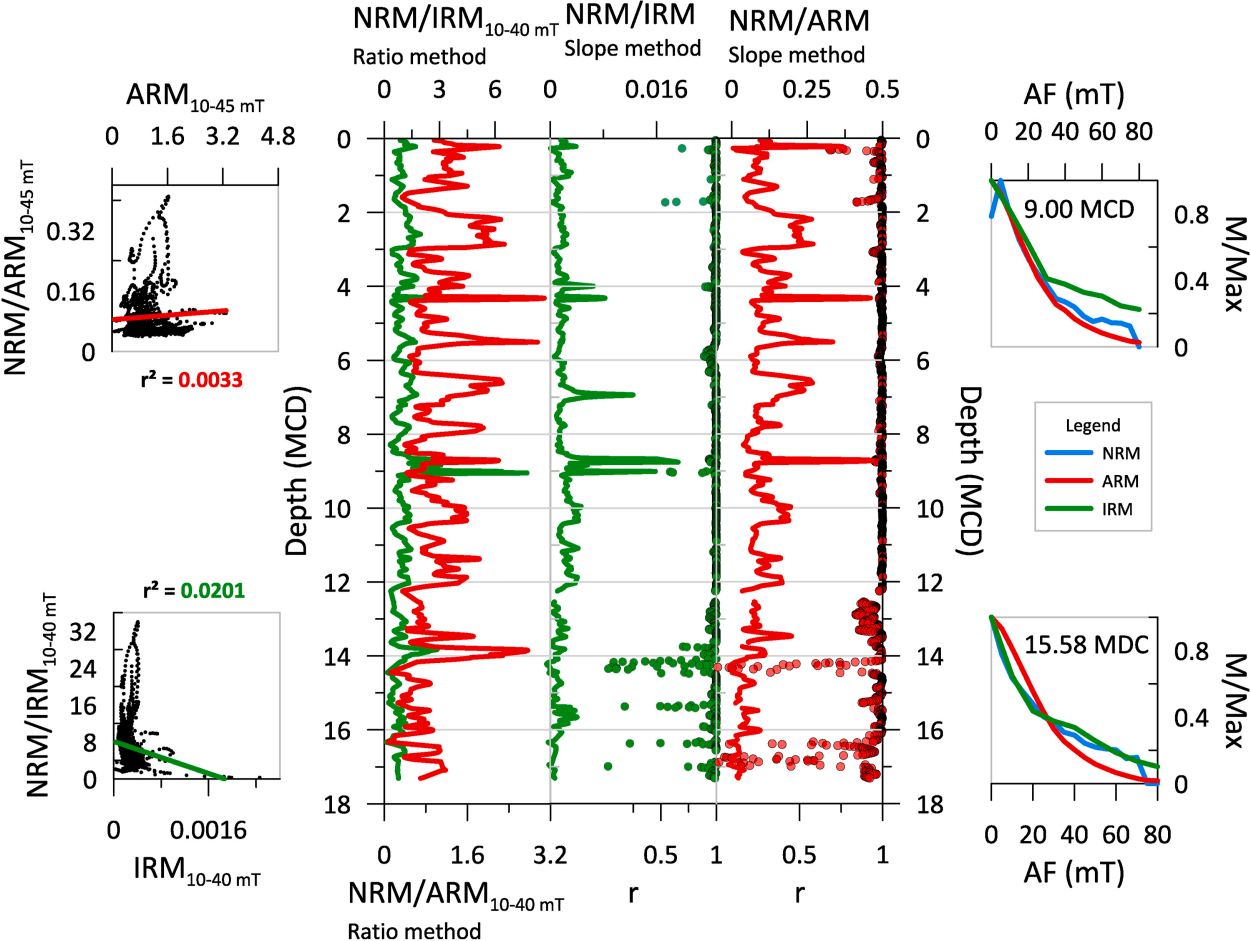
Table 2. List of RPI tie points determined against PISO-1500 (I1-18) and against PADM2M (P1-16), including the error used in the age-depth model.

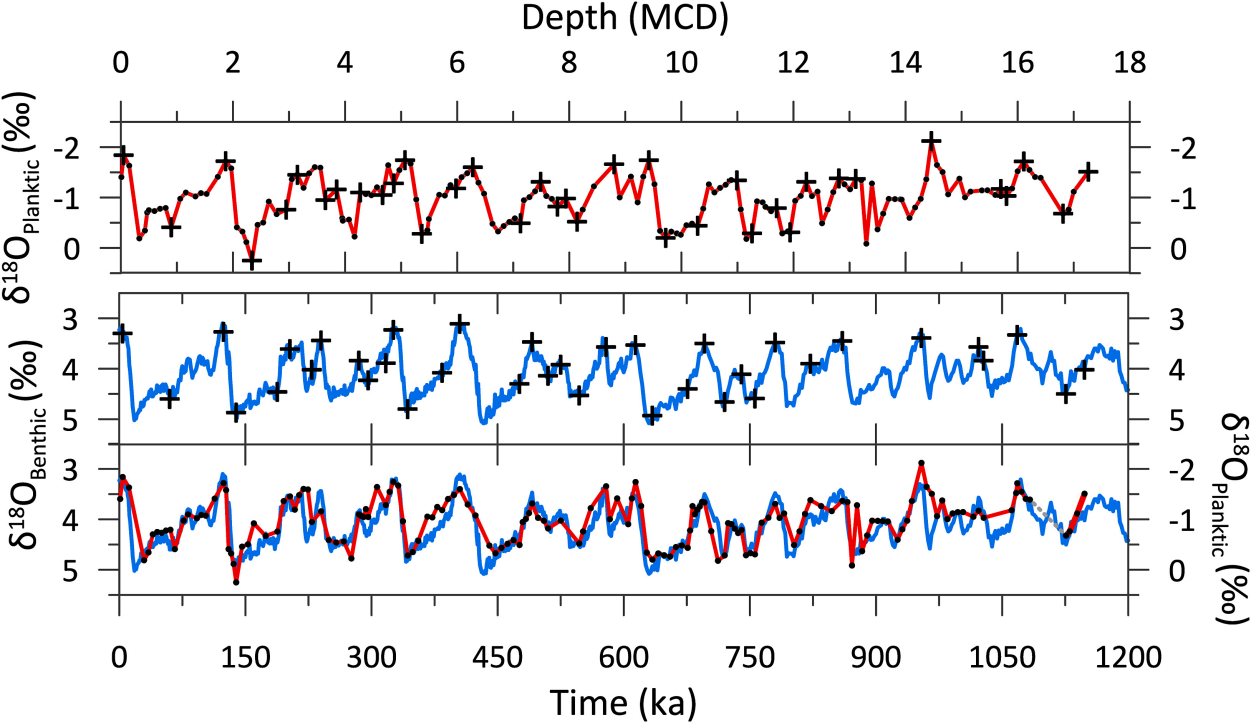


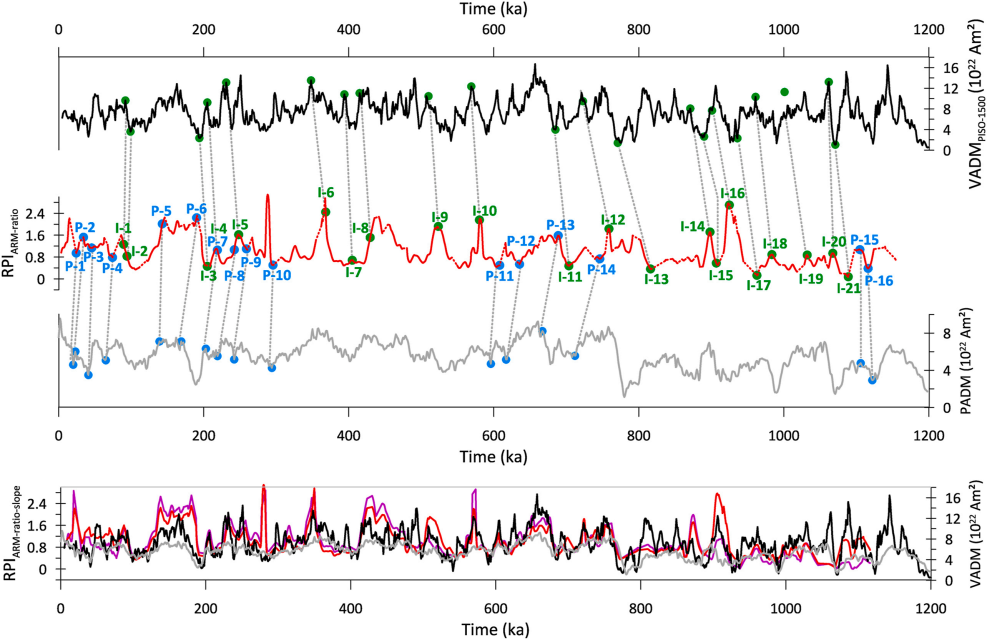
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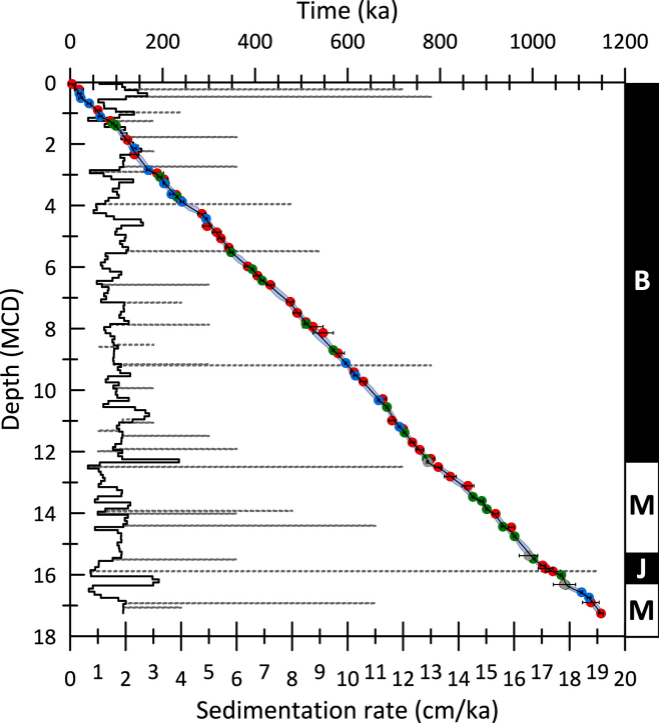
- Piston core 24PC
- Bathymetry CASEIS
- Bathymetry GEBCO
- Land
- Isobath 500 m
- Isobath 100 m
- Continental slope canyon
- Forearc Canyon strike
- N 294°E strike of the Tiburon ridge
- Local minimum gravity anomaly

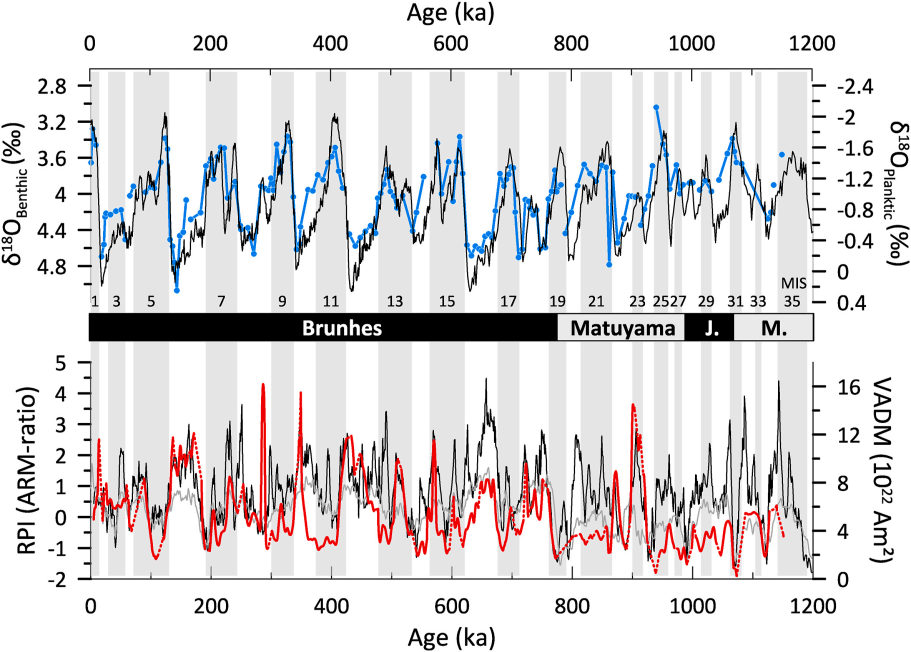


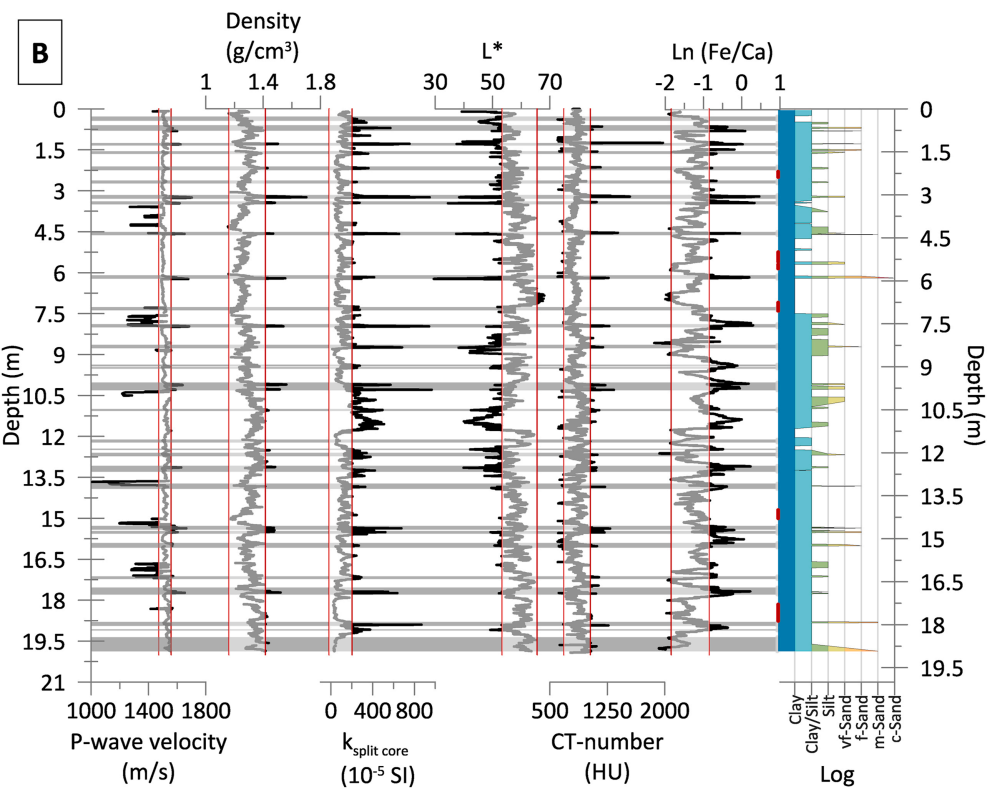
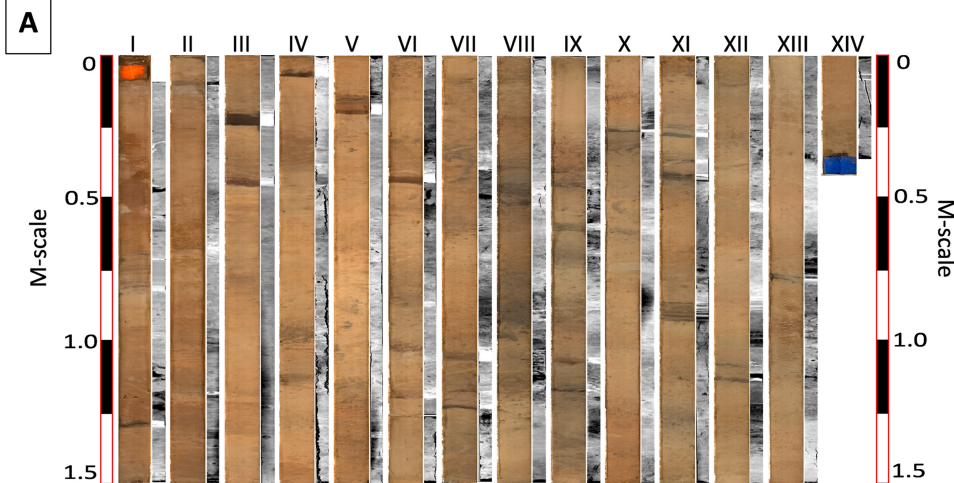


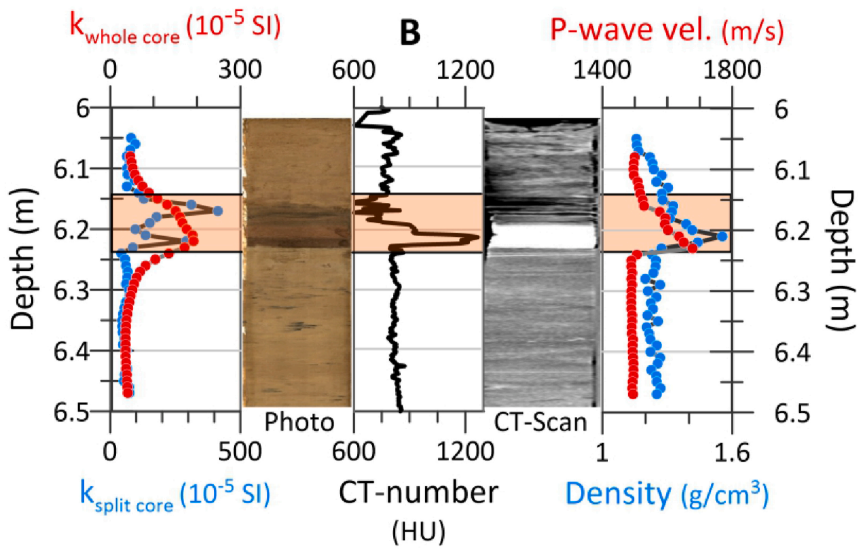
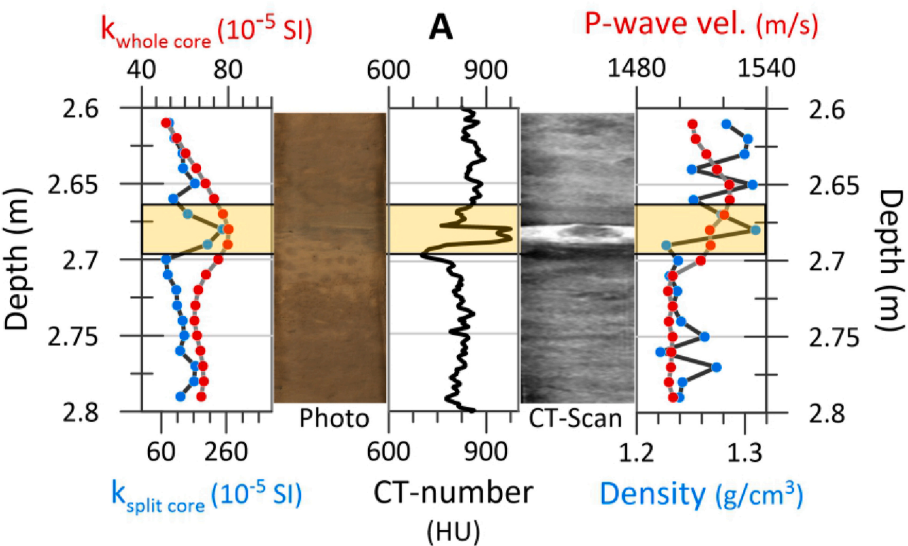


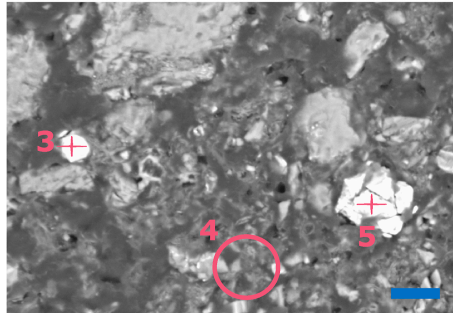
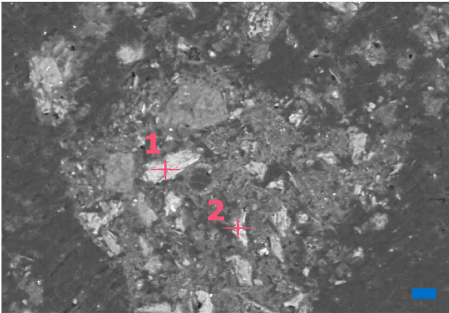




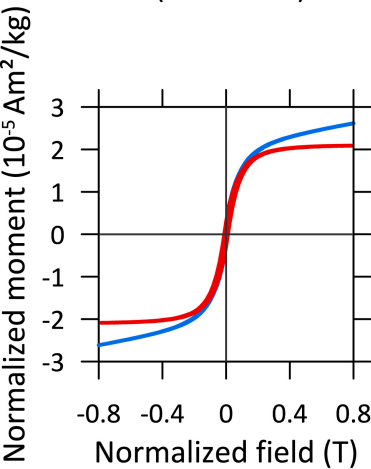




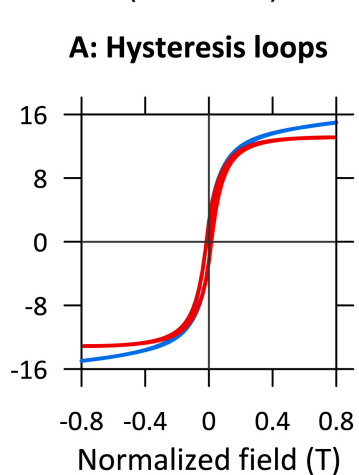




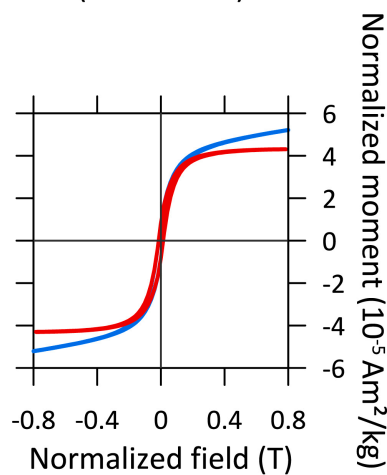
CAS16-24PC-1-95 cm
(0.58 MCD)



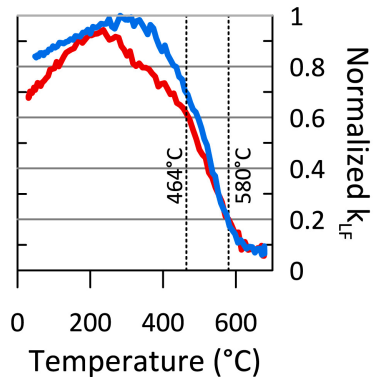
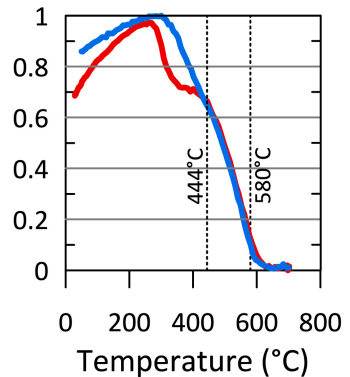
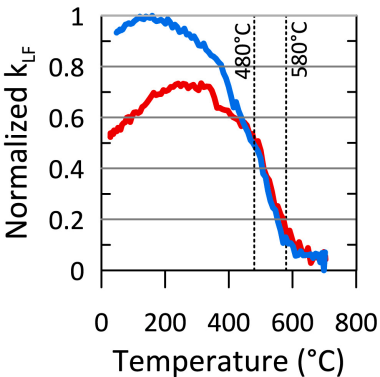
CAS16-24PC-8-1097 cm
(9.82 MCD)

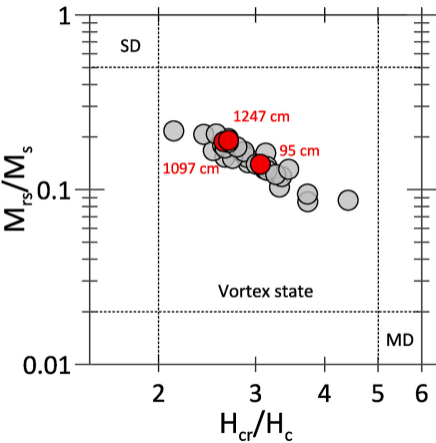
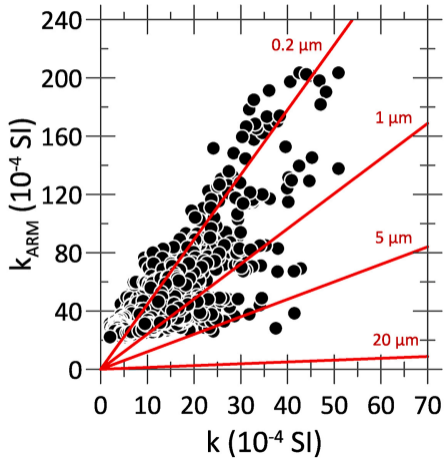


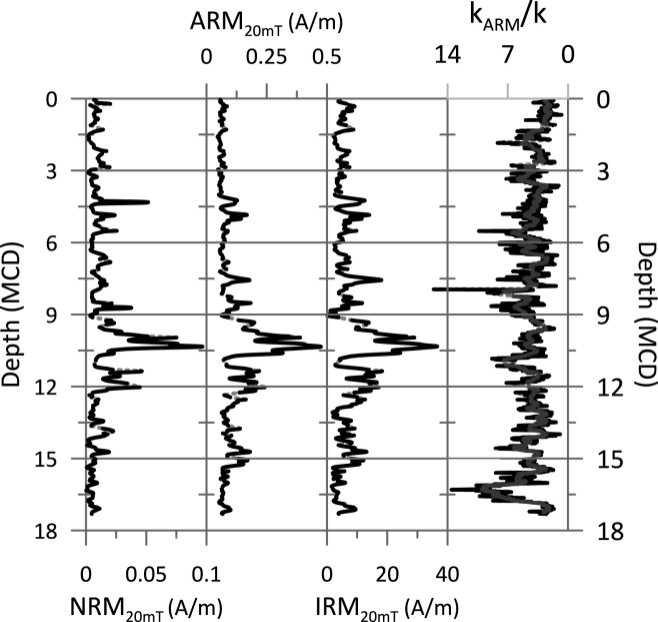
CAS16-24PC-9-1247 cm
(11.33 MCD)

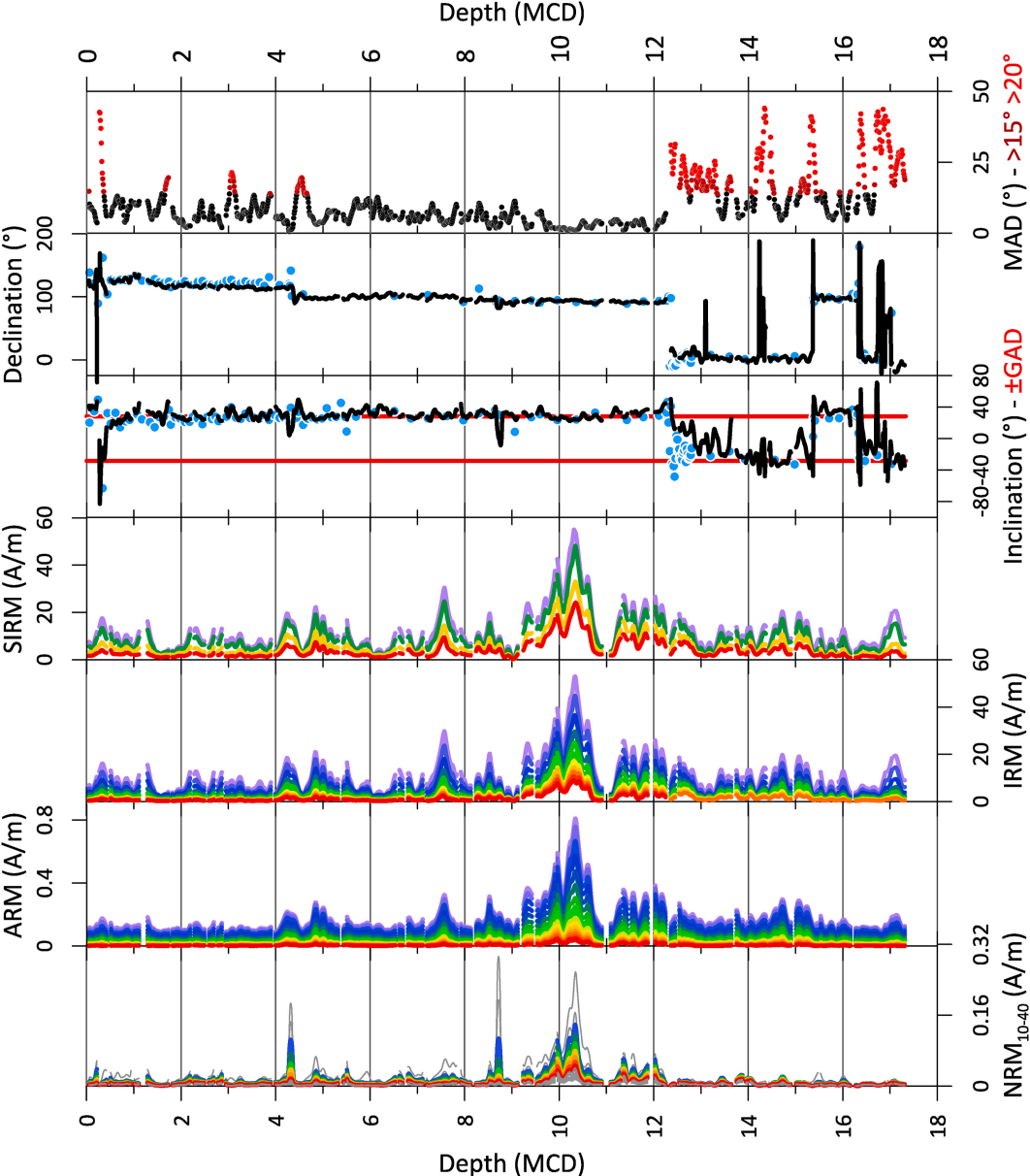


B: Magnetic susceptibility vs. high temperatures



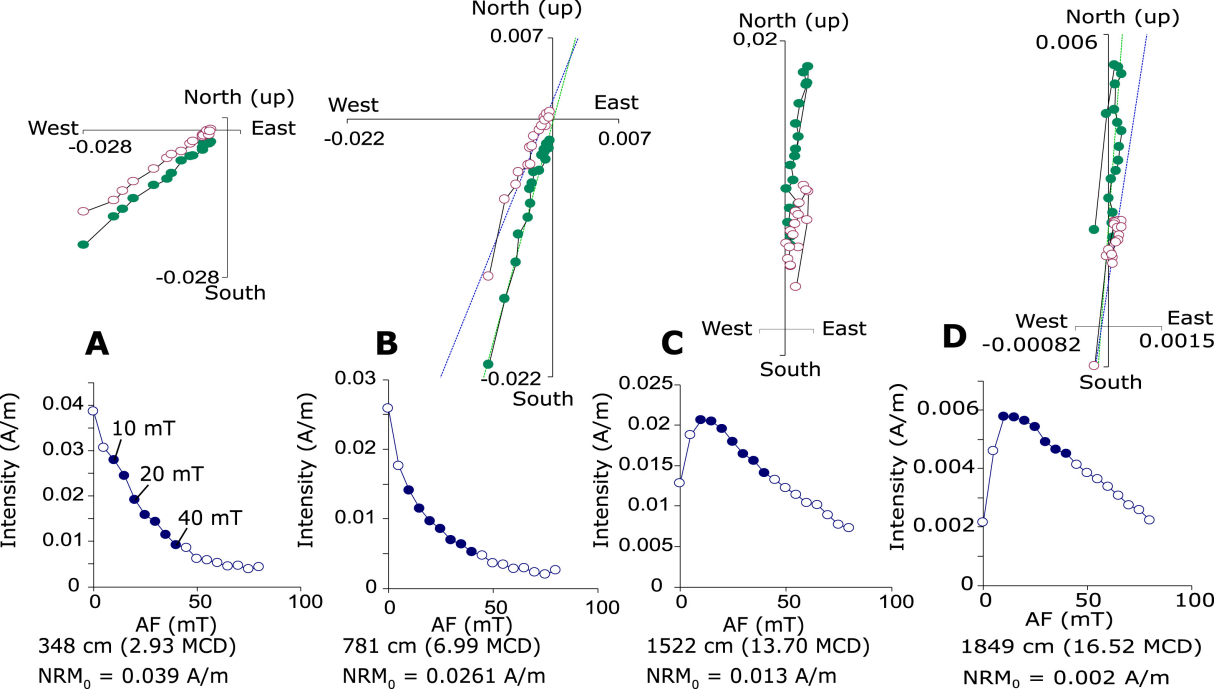
A: Day plot**B: King plot**





Demagnetization steps (mT)





Declaration of interests

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

The authors declare the following financial interests/personal relationships which may be considered as potential competing interests:

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