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# Journal Pre-proof

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1 Chronostratigraphy of a 270-ka sediment record from Lake Selina,  
2 Tasmania: combining radiometric, geomagnetic and climatic dating

3

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21

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23

24 Keywords

25 paleomagnetism, optically stimulated luminescence, radiocarbon, authigenic  $^{10}\text{Be}/^{9}\text{Be}$  ratio,  
26 Pleistocene, lake sediment, Australia

27

28 Highlights

29 ● Chronostratigraphy of a  $266 \pm 8$  ka continuous lake sediment archive from Australia

30 ● We combine radiometric ( $^{14}\text{C}$ , OSL) and relative dating (geomagnetic and climatic)

31 ● Bayesian modeling and dynamic programming integrate dating constraints

32 ● First identification of the Laschamp geomagnetic excursion in Australia

## 33 Abstract

34 Lake sediment archives covering several glacial cycles are scarce in the Southern Hemisphere and  
35 they are challenging to date. Here we present the chronostratigraphy of the oldest continuous lake  
36 sediment archive in Tasmania, Australia; a 5.5 m and 270 ka (Marine Isotope Stage 8) sediment core  
37 from Lake Selina. We employ radiometric dating (radiocarbon and optically stimulated  
38 luminescence) and relative dating (geomagnetic and climate comparisons). Bayesian modeling of the  
39 radiometric ages reaches back to 80 ka (1.7 m) and relative dating using a dynamic programming  
40 algorithm allows dating of the full sequence. Elemental data, magnetic properties and beryllium  
41 isotopes from Lake Selina reveal a close fit to Antarctic ice core climate proxies. Weaker correlation  
42 during the Last Glacial Period (MIS 2–4) is attributed to additional local factors impacting Lake Selina  
43 proxies at a time of climate changes and human arrival into Tasmania. Over that period, full vector  
44 paleomagnetic records and authigenic  $^{10}\text{Be}/^9\text{Be}$  ratios are combined to identify the Laschamp  
45 geomagnetic excursion for the first time in Australia and constrain the chronology. The multi-  
46 method approach provides two preferred age models, indiscernible within their uncertainties, which  
47 allows the use of a geomagnetic dipole-independent (full archive) or a climate-independent (111 ka  
48 to present) age model.

49

## 50 1. Introduction

51 Lake sediments preserve some of the best continuous records of rapid environmental changes, but  
52 those covering multiple glacial-interglacial cycles are notoriously difficult to date. At present, only  
53 one such long continental record is found on the island of Tasmania, south of mainland Australia  
54 (Darwin Crater, Howard and Haines, 2007; Lisé-Pronovost et al., 2019a). Long archives of  
55 environmental change from Tasmania have the potential to elucidate how terrestrial environments  
56 respond to large-scale and rapid climatic change in an under-documented region of the Southern  
57 Hemisphere. Further, developing a precise chronostratigraphy for lake sediments covering the Last  
58 Glacial Period in Australia offers the possibility of investigating the timeline and linkages of  
59 environmental changes across the local and broader region, shedding light on Earth system  
60 dynamics and contributing to debated issues such as megafauna extinction (Roberts et al., 2001;  
61 Wroe and Field, 2006; Rule et al., 2012; Van der Kaars et al., 2017; Saltré et al., 2019) and the impact  
62 and timing of human arrival on the Australian continent (Cosgrove, 1999; Clarkson et al., 2017;  
63 O'Connell et al., 2018). The lacustrine sediments from Lake Selina, a Tasmanian record argued to  
64 span the Last Glacial Cycle (Colhoun et al., 1999), provide an opportunity to explore these questions,  
65 but their tractability depends on the ability to obtain a reliable chronology.

66 Dating lacustrine sediments beyond the range of radiocarbon (>40–50 ka) is challenging. An ability to  
67 combine different dating methods is therefore key to obtaining a reliable chronology (e.g. Colman et  
68 al., 2006; Shanahan et al., 2013; Simon et al., 2020a). Commonly used numerical dating methods for  
69 lacustrine sediments include radiocarbon dating of organic material, optically-stimulated  
70 luminescence dating of quartz and feldspar, U-series dating of carbonates, varve counting, and ash  
71 dating. Other chronological methods include correlation of climate signals (wiggle-matching),  
72 tephrochronology, and paleomagnetic dating. An important consideration in the application of  
73 multiple methods for developing chronologies is to avoid circular reasoning; i.e., ensuring the use of  
74 dating methods that are independent of the variable you are attempting to reconstruct or infer from  
75 analysis of the lake sediments (Blaauw et al., 2012). For example, avoid climate comparison if you  
76 discuss past climate, and avoid magnetostratigraphy if you discuss the Earth's magnetic field.



77 Paleomagnetic dating is useful at the multi-millennial timescale using paleosecular variations of the  
78 geomagnetic field and major instabilities such as polarity reversals and excursions (Channell et al.,  
79 2020). Regional (sub-millennial) paleomagnetic dating is currently infeasible in Australia because  
80 there is a critical lack of empirical data required to build paleomagnetic reference curves, a  
81 limitation that also biases geomagnetic field models (e.g. Panovska et al., 2015; Constable et al.,  
82 2016). The available Pleistocene sedimentary paleomagnetic data from Australia span the past 16 ka  
83 and includes only two full vector lacustrine records from northeast Australia (Lake Eacham and  
84 Barrine, Constable, 1985; Constable and McElhinny, 1985), four directions-only lacustrine records  
85 from southeast Australia (Lake Bullenmerri, Lake Keilambete, Lake Gnotuk, 10 ka, Barton and  
86 McElhinny 1981; Lake Johnston, 9.5 ka, Ankers et al., 2001) and two directions-only lagoon records  
87 from southeast Australia (Big Jibbon Lagoon 5 ka, and Tocal Homestead Lagoon 2 ka, Gale et al.,  
88 2013).

89 Importantly, none of these previous Australian studies meet the modern standards of laboratory  
90 experiments and analytical methods (e.g. Korte et al., 2019), such as magnetic mineral  
91 characterisation to identify the magnetic remanence carriers and assess the reliability of the  
92 paleomagnetic record. Moreover, while it is now routine to perform stepwise demagnetisation to  
93 identify a stable and reliable characteristic remanent magnetisation (ChRM) and calculate the  
94 paleomagnetic directions from the best least-squares line fit by principal component analysis  
95 (Kirschvink, 1980), available Australian studies employ either single step alternating field (AF)  
96 'blanket' demagnetisation or no demagnetisation at all. New state-of-the-art high-quality  
97 sedimentary paleomagnetic and archaeomagnetic data are necessary to confidently isolate the  
98 geomagnetic signal from environmental biases and start building Australian regional reference  
99 curves (sub-millennial behavior). In the meantime, global geomagnetic dipole moment variation can  
100 be used for millennial-timescale paleomagnetic dating by correlating individual records to relative  
101 paleointensity (RPI) stacks (e.g. GLOPIS, Laj et al., 2004) or models (e.g. GGF100k, Panovska et al.,  
102 2018).

103 An alternative means for reconstructing millennial-scale geomagnetic dipole moment variations is  
104 through reconstructing the concentration of cosmogenic radionuclide beryllium-10 ( $^{10}\text{Be}$ ) deposited  
105 in natural archives (Lal and Peters, 1967). Atmospheric  $^{10}\text{Be}$  is produced by spallation reactions when  
106 highly energetic galactic cosmic ray (GCR) particles interact with nitrogen and oxygen atoms. The  
107 incoming rate of these charged GCRs, which directly controls  $^{10}\text{Be}$  production rate, depends on the  
108 strength of geo- and heliomagnetic fields, with a nonlinear inverse relationship: i.e. lower  $^{10}\text{Be}$   
109 production rate at high fields (Beer et al., 2012). In a similar manner to RPI representing  
110 paleomagnetic field intensity by removing depositional and environmental influences (King et al.,  
111 1983; Roberts et al., 2013), normalised  $^{10}\text{Be}$  concentrations from sediment and ice records trace  
112 production variations that are mainly modulated by geomagnetic field strength at the millennial  
113 timescale (e.g. Frank et al., 1997; Muscheler et al., 2005; Horiuchi et al., 2016; Simon et al., 2016a,  
114 2018). Normalised  $^{10}\text{Be}$  can therefore be employed along with RPI proxies to evaluate their  
115 respective limitations (Suganuma et al., 2010, 2011; Valet et al., 2014, 2019; Simon et al., 2019,  
116 2020b) and help recovering reliable records of paleomagnetic dipole moment variations from  
117 lacustrine sedimentary sequences (Nilsson et al., 2011; Du et al., 2018; Tang et al., 2019).

118 In this study, we present the chronostratigraphy of Lake Selina in Tasmania, Australia, combining  
119 radiocarbon, optically stimulated luminescence (OSL) dating, full-vector sedimentary  
120 paleomagnetism, authigenic  $^{10}\text{Be}$  /  $^9\text{Be}$  ratio, and climate comparison.

121

## 122 2. Regional setting

123 Lake Selina (41°52'41"S, 145°36'34"E) is a shallow basin located in the West Coast Range of  
124 Tasmania, southeast Australia, 516 meters above sea level (m asl) and 40 km east of the Southern  
125 Ocean (Fig. 1). The modern lake has a surface area of 190 m<sup>2</sup> and maximum water depth of  
126 approximately 5 m. It has no permanent inflow or outflow; but an ephemeral creek connects Lake  
127 Selina to the larger and higher altitude Lake Westwood (380 m<sup>2</sup>, 605 m asl) ~1 km to the west. The  
128 local geology is dominated by the Ordovician Owen Conglomerate and Cambrian Mount Read  
129 Volcanics. The Owen Conglomerate is a coarse siliciclastic molasse with hematite-stained quartz and  
130 rounded to sub-rounded quartzite pebbles and cobbles in an argillite matrix (Bradley, 1954). While  
131 Mount Selina (~1 km to the east) was prospected for lead and zinc mining, the local mining activities  
132 are associated with the Mount Read Volcanics and include the Rosebery mine for galena at ~12 km  
133 distance and the Henty gold mine ~5 km distance from Lake Selina (Fig. 1C).

134 The climate in western Tasmania is temperate maritime with average temperature ranging from 6 to  
135 16°C and average annual rainfall of 2400 mm (Queenstown; Australian Bureau of Meteorology).  
136 Pleistocene glacial activity has been documented in Tasmania for the Last Glacial Period (MIS 2–3;  
137 Colhoun 1985; Mackintosh et al., 2006) and before, however, with poor age constraints (possibly  
138 MIS 4, 6, 8, and/or 10, 12; Barrows et al., 2002; Augustinus et al., 2017). The West Coast Range has  
139 many cirques and moraines, which provide evidence of paleo-glaciers at Mount Murchison (1279 m  
140 asl) to the north of Lake Selina, Mount Tyndall (1179 m asl) to the south, and Eldon Range (1440 m  
141 asl) to the east. Previous coring of the uppermost 4 m sediments at Lake Selina revealed that the  
142 lake was not ice covered during the Last Glacial Period (Colhoun et al., 1999).

143

## 144 3. Material and methods

### 145 3.1. Coring and sampling

146 Coring of Lake Selina was completed in November 2014 from a floating platform rigged on two  
147 inflatable rafts. Two sites were cored from the approximate lake center (Fig. 1C). Three cores were  
148 recovered including a short gravity core LA (84 cm) and two Nesje (Nesje, 1992) cores N1 (sections A,  
149 B, C; 394 cm) and N2 (sections A, B, C, D; 497 cm) (Fig. 2A). The cores were transported to the  
150 University of Melbourne, stored in a cold room, and split using a GEOTEK core splitter. For each core  
151 section, one half was reserved for micro-XRF (ITRAX) core scanning and archiving, and the other half  
152 for multi-proxy sampling. The sampling strategy includes biogenic proxies (results reported  
153 elsewhere), radiocarbon dating, OSL dating, paleo- and rock-magnetism, and beryllium isotopes.

154

### 155 3.2. Micro-XRF core scanning

156 The cores were scanned using a micro X-ray fluorescence ITRAX core scanner at the Australian  
157 Nuclear Science and Technology Organisation (ANSTO). Optical and X-ray radiograph images of the  
158 split cores were acquired, and elemental composition was measured at 0.02 cm intervals. All  
159 measurements were conducted using a molybdenum (Mo) target tube set at 30 kV and 55 mA with a  
160 dwell time of 10 s. The pass-through radiation (RAD) is a proxy for density (Francus et al., 2015) and  
161 the ratio inc/coh is a proxy for sediment organic content (Woodward and Gadd, 2018). Principal  
162 component analysis (PCA) of the XRF core scanning data was performed to extract the main  
163 components of variance. The selected variables include elements with significant counts (>30 kcps),

164 RAD, inc/coh, and color RGB. Prior to PCA, each variable was scaled by subtracting the mean and  
165 normalising by the standard deviation.

166

### 167 3.3 Radiocarbon dating

168 Samples for radiocarbon dating were taken from core sections LA, N2-A and N2-B (Fig. 2A).  
169 Supplementary Material 1 includes core photographs and radiocarbon samples position. A total of  
170 29 bulk sediment samples of 0.5 cm thickness were analysed at DirectAMS Radiocarbon Dating  
171 Service laboratory. The radiocarbon ages are calibrated using the Southern Hemisphere calibration  
172 curve ShCal20 (Hogg et al., 2020).

173

### 174 3.4 Optically stimulated luminescence (OSL) dating

175 Twenty-three samples for OSL dating (OSL samples) and an additional 21 from above, between and  
176 below each OSL sample for field moisture content and environmental dose rate measurements (DOS  
177 samples), were collected from core sections N1-A, N1-B, N1-C, and N2-D (Fig. 2 and Sup. Mat. 1).  
178 Samples were 3-7 cm thick and collected from clastic sediment facies. Quartz grains (90-125  $\mu\text{m}$   
179 diameter) were isolated for OSL dating, purified using standard procedures (e.g. Wintle, 1997) and  
180 measured on an automated Risø TL-DA-20 luminescence reader (Bøtter-Jensen et al., 2003).

181 Equivalent dose ( $D_e$ ) values were estimated from 6–10 multi-grain aliquots (~300 grains per aliquot)  
182 for each sample using the single aliquot regenerative-dose (SAR) procedure (Wintle and Murray,  
183 2000). Aliquots were preheated at 240°C for 10 s prior to optical stimulation with blue LEDs for 40 s  
184 at 125°C prior to measurement of the natural ( $L_n$ ) and regenerative doses ( $L_x$ ). A fixed test dose  
185 (~5.4 Gy, also preheated at 240°C for 5 s) was applied after each  $L_n$  and  $L_x$  and the induced OSL  
186 signals ( $T_n$  and  $T_x$ ) measured. Also included were duplicate and zero regenerative dose cycles and an  
187 OSL-IR depletion ratio test. Sensitivity-corrected ( $L_x/T_x$ ) dose response curves (DRCs) were then  
188 constructed from the  $L_x$  and  $T_x$  OSL signals, using a saturating exponential function, and the  
189 sensitivity-corrected natural OSL signal ( $L_n/T_n$ ) projected onto the fitted DRC to estimate  $D_e$  by  
190 interpolation.

191 Choice of measurement parameters were tested using a series of dose recovery tests (Galbraith et  
192 al., 1999). Aliquots of sample OSL5 ( $N = 21$ ) were measured using seven different preheat  
193 combinations (180/180, 260/160, 180/160, 200/200, 220/220, 240/220 and 240/240) and a given  
194 dose of ~80 Gy. Accurate measured/given dose ratios were only obtained for two combinations:  
195 240°C for 10 s (PH-1) and 240°C for 5 s (PH-2) (mean =  $1.01 \pm 0.07$ ;  $N = 3$ ) and 220°C for 10 s (PH-1)  
196 and 220°C for 5 s (PH-2) (mean =  $1.00 \pm 0.06$ ,  $N = 3$ ).

197 Environmental dose rates were calculated as the sum of the beta, gamma and cosmic-ray dose rates  
198 external to the grains, plus a small alpha dose rate from inside the quartz grains ( $0.03 \pm 0.01$  Gy/ka).  
199 We assumed that present-day radionuclide activities and dose rates have prevailed throughout the  
200 period of sample burial. Gamma dose rates were estimated from uranium, thorium and potassium  
201 concentrations obtained from a combination of inductively coupled plasma mass spectrometry (ICP-  
202 MS) and optical emission spectroscopy (ICP-OES) and converted to gamma dose rates using the  
203 conversion factors of Guérin et al. (2011). Beta dose rates were estimated from GM-25-5 beta  
204 counting using the procedures described and tested previously (Jacobs and Roberts, 2015), including  
205 a correction for grain size (Brennan, 2003). Cosmic-ray dose rates were calculated following Prescott  
206 and Hutton (1994), adjusting for site altitude, geomagnetic latitude, and the density ( $1.8 \text{ g/cm}^3$  for  
207 sediment and  $1.025 \text{ g/cm}^3$  for water) and thickness of sediment overburden and an average

208 permanent water depth of 5 m. Beta, gamma and cosmic-ray dose rates were corrected for a  
209 moisture content of  $40 \pm 10\%$  for all samples (Nathan and Mauz, 2008). The De divided by the  
210 environmental dose rate gives the burial time of the grains in calendar years ago.

211

### 212 3.5 Magnetic analysis

213 Paleomagnetic  $8 \text{ cm}^3$  cube samples were taken continuously from cores LA, N1-A, N1-B, N1-C, N2-C,  
214 and N2-D (Fig. 2A). Core N2-A (24-81.6 cm section depth) was sampled for duplicate measurements.  
215 A total of 287 cube samples and 24 discrete samples (1 g) were analysed at the Australian National  
216 University. Magnetic analyses of the cube samples include low field volumetric magnetic  
217 susceptibility (klf) measured with an AGICO Kappabridge, and the natural, anhysteretic, and  
218 isothermal remanent magnetizations (NRM, ARM, and IRM, respectively) measured with a 2G  
219 cryogenic magnetometer. The NRM was demagnetized in 13 steps to maximum alternating field (AF)  
220 of 90 mT. The results were analysed in Puffinplot (Lurcock and Wilson, 2012) using unanchored  
221 principal component analysis (PCA). The cores were not azimuthally oriented; the declination  
222 records are aligned by subtracting the average declination of the non-anomalous values from each  
223 section. The ARM was induced using a direct current field of 0.05 mT and peak AF of 100 mT and  
224 demagnetized in 7 AF steps. A first IRM was induced in 1 T field using a 2G pulse magnetizer and  
225 demagnetized in 6 AF steps, and a second IRM was induced in 0.3 T field in the opposite direction  
226 and measured. Useful coercivity indicators from the remanence measurements include the median  
227 destructive field (MDF; the field required to demagnetize half the initial remanence; Dankers, 1981),  
228 the ARM ratio ( $k_{\text{ARM}}/\text{IRM}$ ; Peters and Dekkers, 2003), the S-ratio ( $\text{IRM}_{0.3\text{T}}/\text{IRM}_{1\text{T}}$ ; a proxy for the  
229 relative contributions of low and high coercivity minerals to the total IRM; Liu et al., 2012) and the  
230 HIRM ( $(\text{IRM}_{1\text{T}} + \text{IRM}_{0.3\text{T}})/2$ ; the absolute contribution of high coercivity minerals to the total IRM; Liu  
231 et al., 2012).

232 Magnetic analyses of the 24 discrete samples were used to further characterise the magnetic  
233 mineral assemblage. The analyses include hysteresis loops, IRM acquisition and back field curves  
234 using a Princeton Measurements Corporation Vibrating Sample Magnetometer (VSM). The  
235 remanence ratio (saturation remanence ( $M_{\text{rs}}$ )/saturation magnetisation( $M_{\text{s}}$ )) and the coercivity  
236 ratio (the remanent coercive force ( $H_{\text{cr}}$ )/bulk coercive force ( $H_{\text{c}}$ )) calculated from hysteresis loop  
237 and back field curve data are useful magnetic grain size and mineralogy indicators (Peters and  
238 Thompson, 1998; Roberts et al., 2011). IRM acquisition curves were analysed using MAX UnMix  
239 (Maxbauer et al., 2016). First-order reversal curves were acquired on 4 samples and processed using  
240 FORCinel (Harrison and Feinberg, 2008) with VARIFORC smoothing (Egli, 2013). Thermomagnetic  
241 curves were acquired for the representative samples N1-A18 and N1-C37. High-temperature cycling  
242 of magnetic susceptibility was measured from room temperature to 700 °C using a AGICO MFK2, and  
243 low-temperature cycling of room-temperature saturation isothermal remanent magnetisation  
244 (RTSIRM 5T) was measured down to 10 K using a Quantum Design Magnetic Property Measurement  
245 System (MPMS) at the Australian National University.

246

### 247 3.6 Beryllium isotopes analysis

248 Samples for beryllium isotope analysis were taken from core sections N1-A and N1-B (Fig. 2A) on one  
249 side of the trench left by paleomagnetic cube sampling at corresponding depth intervals. A total of  
250 45 samples were analysed at CEREGE National Cosmogenic Nuclides Laboratory following the  
251 method established by Bournès et al. (1989) and revised by Simon et al. (2016b). Authigenic  $^{10}\text{Be}$  and  
252 its stable isotope  $^9\text{Be}$  were extracted from  $\sim 1$  g dry sample by soaking the samples in 20 ml of

253 leaching solution (0.04 M hydroxylamine (NH<sub>2</sub>OH–HCl) and 25% acetic acid) at 95 ± 5°C for 7 h. A 2-  
254 ml aliquot of the resulting leaching solution was sampled for measurement of the natural <sup>9</sup>Be  
255 concentration using a graphite-furnace atomic absorption spectrophotometer (AAS) with a double  
256 beam correction (Thermo Scientific ICE 3400). The remaining solution was spiked with 300 µl of a  
257 9.8039 × 10<sup>-4</sup> g<sup>-1</sup> <sup>9</sup>Be carrier before Be purification by chromatography to determine accurate <sup>10</sup>Be  
258 sample concentrations from accelerator mass spectrometer (AMS) measurements of <sup>10</sup>Be /<sup>9</sup>Be  
259 ratios. <sup>10</sup>Be sample concentrations were calculated from the measured spiked <sup>10</sup>Be /<sup>9</sup>Be ratios  
260 normalised to the BeO STD-11 in-house standard (1.191 ± 0.013 × 10<sup>-11</sup>) (Braucher et al., 2015).  
261 Authigenic <sup>10</sup>Be concentrations were decay-corrected using the <sup>10</sup>Be half-life (T<sub>1/2</sub>) of 1.387 ± 0.012  
262 Ma (Chmeleff et al., 2010; Korschinek et al., 2010) and the dipole-independent age model (see  
263 below).

264

### 265 3.7 Core correlation

266 Core correlation is based on 24 visible stratigraphic tie-points and one inc/coh data tie-point (Fig.  
267 2A). The correlation of the gravity core (LA) to the Nesje cores (N1 and N2) is set using a large  
268 amplitude change in inc/coh ratio at 75 cm in LA, 29 cm in N1-A, and 25 cm in N2-A (Fig. 2B). The  
269 inc/coh correlation depths correspond to the maximum inc/coh value after the inflection. The  
270 composite depth LAN2 is the reference record (Fig. 2C) used in this work. Data measured on samples  
271 from core N1 (OSL, beryllium, and magnetism) is transferred to the reference depth LAN2 using a  
272 transfer function based on linear interpolation and the 24 stratigraphic tie-points (Fig. 2A; cf. Result  
273 section 4.1 Lithology; Sup. Mat. 2). This assumes constant sedimentation rates between tie-points  
274 within each facies. We note that the stratigraphy of core section N2-A is potentially disturbed; a  
275 sediment gap formed (at 114 cm depth LAN2) during core splitting and two sub-sections were  
276 labelled (N2-A-a and -b; Fig. 2A). Thus, the duplicate magnetic data from core section N2-A is  
277 represented in the figures by points only.

278

## 279 4. Results

### 280 4.1. Lithology

281 Sediments from Lake Selina are characterised by two main facies, I and II (Fig. 3). These sedimentary  
282 facies have been associated with interglacial and glacial time intervals based on previous coring and  
283 radiocarbon dating (Colhoun et al., 1999). Interglacial facies I is homogeneous dark brown to black  
284 mud with lower density (RAD), higher organic content (inc/coh), lower ferrimagnetic content (klf,  
285 NRM, ARM, IRM) and lower bulk coercivity magnetic minerals assemblage (HIRM, S-ratio) than  
286 glacial facies II. Facies II is lighter in color, more clastic (higher Pb, Zn, Rb, Zr, Si, K) and composed of  
287 two main sub-facies (IIa and IIb) that alternate in distinct or mixed layers, sometimes including  
288 laminations. The sub-facies IIa is grey-beige mud and has the lowest organic content, the highest  
289 ferrimagnetic content and the higher magnetic coercivity assemblage. Sub-facies IIb is grey-brown  
290 mud with occasional lighter mud clasts and intermediate values between facies I and facies IIa. A few  
291 pebbles were found during sub-sampling (Fig. 2). Figure 4 illustrates the XRF PCA variables  
292 contribution to the principal components PC1 and PC2. The organic matter proxies (inc/coh, RAD, Br)  
293 and the elements associated with the clastic fraction of the sediment and the local geology (Rb, Zr,  
294 Si, Pb, Zn, K) strongly influence PC1 and are negatively correlated.

295 Cores N1 and N2 have the same stratigraphic sequence and are readily correlated visually (Fig. 2).  
296 However, some of the facies in N1 and N2 have different thickness (discussed in section 5.1). The



297 average facies difference is 1.7 cm thicker in N1 and the most outstanding example is the uppermost  
 298 glacial clastic facies IIa that has a thickness of 1 cm in N2 and 7 cm in N1. Yet, the facies alternation  
 299 provides 24 unambiguous stratigraphic tie-points. The cumulative depth difference is an additional  
 300 35 cm in composite depth LAN1 compared to LAN2 at the deepest tie-point (431 cm depth LAN1 and  
 301 396 cm depth LAN2; Fig. 2; Sup. Mat. 2). There is a series of oblique contacts and a cm-sized dark  
 302 vertical stain (organic debris) near the base of the core in the lowermost facies II (516-544 cm depth  
 303 LAN2).

304

#### 305 4.2. Radiometric dating ( $^{14}\text{C}$ and OSL)

306 A total of 29  $^{14}\text{C}$  (Table 1) and 12 OSL ages (Table 2) were obtained from the TAS1402 cores. The  
 307 oldest age is  $84 \pm 5$  ka at  $174.5 \pm 2.5$  cm depth LAN2 (OSL13; Table 2). OSL signals for samples  
 308 collected below this depth (OSL14 to 23) were in dose saturation, so finite  $D_e$  values and ages could  
 309 not be obtained. Most samples have overdispersion values  $<20\%$  (Table 2), consistent with samples  
 310 that were well-bleached prior to deposition and remained largely undisturbed. The radial plots show  
 311 tight distributions of  $D_e$  values, in which the majority lie within  $\pm 2$  standardised estimates of a  
 312 common value (Sup. Mat. 3). The weighted mean  $D_e$  values for age determination were calculated  
 313 using the central age model (CAM) of Galbraith et al. (1999) (Table 2). The dose rate data are  
 314 provided in Table 2 and shown for all OSL and DOS samples between depths of 81 and 190 cm depth  
 315 LAN2 in Fig. 5. The total dose rates are similar for different samples, and there are no significant  
 316 trends with depth or between different sediment facies. A comparison of beta dose rates  
 317 determined using two independent methods and laboratories verify the reliability of the beta dose  
 318 rate measurements (Fig. 5D). The final optical ages are presented in Table 2. Uncertainties on the  
 319 ages are given at  $1\sigma$  (the standard error on the mean) and were estimated by combining, in  
 320 quadrature, all known and estimated sources of random and systematic error.

321 Figure 6 presents the radiometric dating results and Bayesian age-depth model (rbacon; Blaauw and  
 322 Christen, 2011). The OSL ages (ka before 2018) were corrected by 68 years to make it equivalent to  
 323 the calibrated radiocarbon ages (ka before 1950 – ka cal BP).  $^{14}\text{C}$  and OSL samples overlap over the  
 324 interval 83–177 cm LAN2 depth. Below the first OSL-dated clastic sediment sample at 83–84 cm  
 325 LAN2 depth, the calibrated  $^{14}\text{C}$  ages return age inversions with apparently too ‘old’ and too ‘young’  
 326 ages relative to OSL. Calibrated  $^{14}\text{C}$  ages reach a plateau at ca. 40 ka cal BP (grey symbols in figure 6).  
 327 The two calibrated  $^{14}\text{C}$  ages (D-AMS 027826 and D-AMS 015350; Table 1) that appear too ‘old’ may  
 328 indicate older organic material input to the coring site. The calibrated  $^{14}\text{C}$  ages that appear too  
 329 ‘young’ can be explained by two processes. First, underestimation of  $^{14}\text{C}$  ages by modern carbon  
 330 contamination (Fig. 6B; Pigati et al., 2007). Contamination has greater impact on older materials ( $>35$   
 331 ka BP) that contain proportionally less radiocarbon than younger materials. Figure 6B shows that  
 332 true  $^{14}\text{C}$  ages  $>40$  ka BP with 1-2% modern carbon contamination always returns  $^{14}\text{C}$  ages of about  
 333 30–35 ka (32–38 ka cal BP). This likely explains the vertical distribution of 7 calibrated  $^{14}\text{C}$  ages below  
 334  $\sim 110$  cm, where 4 other samples returned results not distinguishable from background (NDFB; Table  
 335 1, Fig. 6). Second, percolation of carbon down the core through the porous glacial lithologies (e.g.  
 336 Briant and Bateman, 2009; Long and Shen, 2014). This process may explain  $^{14}\text{C}$  ages younger than  
 337 OSL and  $<35$  ka (D-AMS 027827, D-AMS 027828, D-AMS 009084; Table 1). Unlike modern carbon  
 338 contamination, the migration of young carbon in sediments is site-specific and difficult to quantify. It  
 339 is, however, limited by the stratigraphy because percolation takes place through porous material at  
 340 a lithological boundary. Thus, we can reasonably assume that young carbon migration only affects  
 341  $^{14}\text{C}$  ages at depths greater than the uppermost porous detrital layer ( $>83$  cm LA-N2 composite  
 342 depth).

343 To investigate further the young carbon effect, we compare two Bayesian age modeling runs  
 344 (rbacon; Blaauw and Christen, 2011): the first using all calibrated  $^{14}\text{C}$  and OSL ages (run 1) and the  
 345 second using only the 12 calibrated  $^{14}\text{C}$  ages from 0–83 cm and all OSL ages (run 2). With run 2, we  
 346 assume that the  $^{14}\text{C}$  ages at depths >83 cm are affected by old or young carbon contamination,  
 347 and/or downcore percolation, all of which relate to lithological changes during the Last Glacial  
 348 Period. Run 1 and 2 return quite similar age-depth models, however run 1 shows sharper changes in  
 349 sedimentation rate constrained by the extra  $^{14}\text{C}$  ages at depths >83 cm. Key to this exercise is that  
 350 rbacon independently rejected the too ‘old’ and too ‘young’  $^{14}\text{C}$  ages as hypothesized from the  
 351 lithology, providing support to the run 2 sample selection (Sup. Mat. 4). We select the more  
 352 conservative age-depth model run 2, based on 12 calibrated  $^{14}\text{C}$  and 12 OSL ages, to avoid any bias  
 353 induced by unknown young carbon migration and old/young carbon contamination. The inclusion of  
 354 stratigraphic boundaries (Sup. Mat. 2) in the rbacon model to allow for sedimentation changes  
 355 between facies returns the same result within uncertainty as model runs that do not include  
 356 stratigraphic boundaries (Sup. Mat. 4). The age-depth model run 2 based on radiometric dating is  
 357 hereafter referred to as “Radiometric”.

358

### 359 4.3. Magnetic mineral assemblage

360 Ferrimagnetic mineral content varies by up to one order of magnitude between sediment facies I  
 361 and II (Fig. 3; klf, NRM, IRM). The organic-rich facies I has weak average NRMs of  $2 \times 10^{-3} \text{ Am}^{-1}$  and klf  
 362 values of  $7 \times 10^{-6} \text{ SI}$ . The more clastic facies II has average NRM and klf values five to ten times higher,  
 363 of  $21 \times 10^{-3} \text{ Am}^{-1}$  and  $35 \times 10^{-6} \text{ SI}$ , respectively. Slightly coarser magnetic grains (kARM/IRM,  $\text{MDF}_{\text{ARM}}$ ,  
 364 Mr/Ms) and more high coercivity minerals (HIRM) characterize facies II compared to facies I (Fig. 3).  
 365 The magnetic mineralogy is a mixed magnetic assemblage comprising at least two contrasting  
 366 coercivity populations as indicated by wasp-waisted hysteresis loops (Fig. 7) and S-ratio varying  
 367 between 0.4 and 0.8 (Fig. 3). The mineralogy discriminant remanence ratio ARM40mT/SARM (Peters  
 368 and Thompson, 1998) (Fig. 8) and the average Hcr/Hc value of  $4.5 \pm 1.5$  indicate that magnetite is  
 369 the primary remanence carrier (pyrrhotite and greigite have Hcr/Hc <1.5; Peters and Thompson,  
 370 1998; Roberts et al., 2011). Further magnetic mineralogical evidence comes from the temperature-  
 371 dependant analysis. The low temperature analyses do not reveal a magnetite Verwey transition (a  
 372 change in remanence at 120 K) and the RTSIRM curve is almost reversible (Fig. 7B and F), which is  
 373 typical of surficially oxidized magnetite (maghemite), titanomagnetite, and/or goethite as  
 374 remanence carriers (France and Oldfield, 2000; Lacroix and Guyodo, 2017). The goethite Curie  
 375 temperature (120 °C) is occasionally seen in the thermomagnetic curves of magnetic susceptibility  
 376 for facies II samples. None of the (not shown). No thermomagnetic curves are reversible and  
 377 ferrimagnetic minerals are formed and/or grain size changed from about 400 °C (Fig. 7B and F). Yet, t  
 378 his transformation is not diagnostic and may result from heating of other iron oxides, iron sulphides,  
 379 hydroxides, carbonates or silicates originally present in the sediments (Henry, 2007). Finally, five  
 380 samples from facies II (<2% of the samples) acquired a gyroremanent magnetisation (GRM) during  
 381 stepwise AF demagnetisation of the NRM in fields >40 mT, indicating presence of some greigite in  
 382 the lake sediments (e.g., Roberts et al., 2011; Lise-Pronovost et al., 2014).

383 Unmixing of the 24 IRM acquisition curves using MAX UnMix (Maxbauer et al., 2016) indicates a  
 384 minimum of three subpopulations of magnetic mineral particles (P1, P2, P3) with average coercivity  
 385 ranges <20 mT (P1), of  $63 \pm 9 \text{ mT}$  (P2), and >270 mT (P3) (Fig. 7). The lower coercivity population P1  
 386 may correspond to multidomain particles with an average estimated contribution to the IRM of 10 %  
 387 (Maxbauer et al., 2016). The high coercivity population P3 has an average estimated contribution to  
 388 the IRM of 40 % and may correspond to hematite and/or goethite (Egli, 2004). The dominant

389 population is P2 with an estimated contribution of 50 % to the IRM and a medium coercivity range  
390 that corresponds to fine iron oxide particles (Egli, 2004). While the P2 range of IRM unmixed  
391 coercivities better match biogenic magnetite, the  $MDF_{ARM}$  values (20–30 mT), ARM ratio ( $0.38 \text{ mmA}^{-1}$ ),  
392 and dispersion parameter value ( $\log DP_{P2} = 0.4$ ) closely match the typical values for detrital,  
393 pedogenic, extracellular, or aeolian sources (Egli, 2004; Maxbauer, 2016). Further support for a non-  
394 biogenic source is provided by the FORC diagrams (Fig. 7F) which indicate vortex and interacting  
395 single domain particles (Harrison et al., 2018) based on the absence of a central ridge feature  
396 associated with the non-interacting single domain particles, typical of biogenic magnetite (Egli et al.,  
397 2010; Heslop et al., 2014). Yet, it is unclear if P2 has a detrital and/or biogenic origin, as biogenic  
398 magnetite may have oxidized to maghemite. We note that a dual biogenic and detrital origin for the  
399 iron minerals would be consistent with Fe being neutral in the PC1 opposing organic and clastic  
400 variables (Fig. 4). The proportion of the primary magnetic population P2 is maximum in facies II and  
401 accounts for the higher ferrimagnetic concentration (k<sub>lf</sub>, NRM, ARM, IRM; Fig. 3). In summary, the  
402 magnetic analysis results indicate a principal magnetic population of oxidized detrital (titano-)  
403 magnetite (P2) present throughout all sedimentary facies (Fig. 7) and isolated by alternating field  
404 demagnetisation of the ARM and IRM (Fig. 8). Up to half the estimated contributions are attributed  
405 to combined secondary populations of low coercivity ferrimagnetic material (P1) and high coercivity  
406 hematite and/or goethite (P3).

407

#### 408 4.4. Paleomagnetic record

##### 409 4.4.1. Paleomagnetic directions

410 The paleomagnetic directions calculated for all demagnetisation steps (13 steps from 0 to 90 mT)  
411 show inclinations varying near the expected value ( $-61^\circ$ ) for the site latitude and a geocentric axial  
412 dipole (GAD). However, the inclination and declination demagnetisation data are noisy with  
413 maximum angular deviation (MAD) reaching values up to  $40^\circ$  (Fig. 9A). The MAD is approximately  
414 inversely proportional to the NRM intensity, with highest MAD values in weakly magnetic interglacial  
415 sediments (facies I) and the most stable directions in glacial sediments (facies II). There is no  
416 evidence of inclination shallowing or paleomagnetic directional (in)stability tied to specific facies.  
417 Overlapping core sections LA, N1-A and N2-A give reproducible results (Fig. 9A). A two-step selection  
418 process was applied to select a paleomagnetic direction record (Fig. 9B). First, the optimum  
419 characteristic remanent magnetisation (ChRM) interval is selected for each sample. The  
420 demagnetisation steps associated with any viscous overprint ( $<20 \text{ mT}$ ), weak noisy values at higher  
421 fields and GRM ( $>40 \text{ mT}$ ) are left out. Most selected ChRM intervals lie between 10 and 60 mT and  
422 include about 9 demagnetisation steps. Second, we select samples having a minimum of 5  
423 consecutive steps and MAD values  $<20^\circ$ . 76 % of the samples pass these selection criteria and the  
424 selected paleomagnetic directions data are shown in figure 9B. The selected paleomagnetic  
425 directions demonstrate overall small-scale oscillations, in agreement with paleosecular variation,  
426 except for few intervals showing large-scale deviations (Fig. 9B). The distinction between secular  
427 variation and geomagnetic instabilities (e.g. geomagnetic excursions) is often defined empirically by  
428 a deviation of the virtual geomagnetic pole (VGP) by more than  $45^\circ$  accompanied by paleointensity  
429 minima (e.g. Roberts, 2008; Laj and Channell, 2015). Here, we consider that paleomagnetic  
430 inclinations departing  $>45^\circ$  away from the GAD and associated with large amplitude changes in  
431 declination correspond to “excursion” events features (Fig. 9B). These “excursion” events  
432 features are found in the interval 91-116 cm depth LAN2 (facies IIa and IIb), at 302 cm depth LAN2  
433 (facies I), and at 558 cm and 565 cm depth LAN2 (lowermost facies IIa).

434



## 435 4.4.2. Relative paleointensity (RPI)

436 RPI is estimated by normalising the NRM with a rock-magnetic parameter such as the ARM or IRM to  
 437 isolate the geomagnetic signal, considered to be proportional to the ancient field intensity, from the  
 438 concentration of the remanence carrying ferrimagnetic mineral. A set of criteria for rock magnetic  
 439 homogeneity is traditionally used to assess the quality of sediments for RPI reconstruction (King et  
 440 al., 1983; Roberts et al., 2013). 1) The ferrimagnetic concentration should vary by less than one  
 441 order of magnitude (kIf, ARM, IRM), 2) the ideal magnetic mineral is magnetite, and 3) the ideal  
 442 magnetic particle size is single domain (SD). The magnetic mineralogy results indicate that the  
 443 sediments from Lake Selina barely fulfill these criteria (Figures 3, 7, and 8). The ferrimagnetic  
 444 content of facies I and II differs by one order of magnitude and although fine magnetite particles (P2)  
 445 dominate the magnetic mineral assemblage, magnetite is not the only ferrimagnetic mineral.  
 446 Downcore homogeneity of the magnetic mineral assemblage is crucial for RPI reconstruction  
 447 because different magnetic particles can have different recording efficiencies that are difficult to  
 448 reconcile into a continuous RPI record (e.g. Chen et al., 2017). Figures 9C and D compares four RPI  
 449 estimates. The raw data in Figure 9C illustrates efficiency differences in the sediments of Lake Selina,  
 450 with RPI values varying over one order of magnitude. Data deemed unsuitable for RPI include  
 451 isolated peak RPI values up to 5 reflecting NRM overprint (Fig. 9C) and high  $\text{NRM}_{20\text{ mT}}/\text{ARM}_{20\text{ mT}}$   
 452 values  $>0.4$  corresponding to high  $\text{MDF}_{\text{NRM}}$  values  $>30\text{ mT}$  (not shown). These intervals have a  
 453 distinct remanence efficiency apparently associated with some of the facies IIa; they are highlighted  
 454 in grey in figures 3 and 9. Figure 10D presents the selected RPI estimates (without data from the grey  
 455 intervals). The RPI estimates include NRM normalisation by ARM and IRM at 20 mT demagnetisation  
 456 ( $\text{NRM}_{20\text{ mT}}/\text{ARM}_{20\text{ mT}}$  and  $\text{NRM}_{20\text{ mT}}/\text{IRM}_{20\text{ mT}}$ ) and by the portion of remanence removed between 10  
 457 mT and 60 mT demagnetisation ( $\text{NRM}_{10-60\text{ mT}}/\text{ARM}_{10-60\text{ mT}}$ ,  $\text{NRM}_{10-60\text{ mT}}/\text{IRM}_{10-60\text{ mT}}$ ) to specifically target  
 458 the P2 magnetic population. The four RPI proxies are similar (Fig. 9C-D), which supports P2 as the  
 459 main remanence carrier isolated in laboratory AF fields. The interval 10-60 mT has the best  
 460 reproducibility between cores (LA, N1 and N2 overlap; Fig. 9D) and therefore  $\text{NRM}_{10-60\text{ mT}}/\text{ARM}_{10-60\text{ mT}}$   
 461 is selected as the preferred RPI estimate. In the interval displaying excursions directions, eight RPI  
 462 data (3 cubes from core N1 and 5 cubes from N2) did not pass the selection criteria because of their  
 463 different remanence efficiency. A total of six samples (3 cubes from core N1 and 3 cubes from N2)  
 464 have both reliable RPI and excursions directions, including the low RPI interval from 110 to 116 cm  
 465 depth LAN2 (Fig. 10).

466

## 467 4.4.3. Beryllium isotopes

468 The authigenic  $^9\text{Be}$  concentrations range from 0.1 to  $2.2 \times 10^{16}\text{ at.g}^{-1}$ , with an average value of  $0.6 \pm$   
 469  $0.4 \times 10^{16}\text{ at.g}^{-1}$  (Fig. 3). The authigenic  $^{10}\text{Be}$  (decay-corrected) concentrations range from 3.1 to  $36.7$   
 470  $\times 10^8\text{ at.g}^{-1}$ , with an average value of  $12.6 \pm 7.1 \times 10^8\text{ at.g}^{-1}$ . These  $^{10}\text{Be}$  and  $^9\text{Be}$  concentrations and  
 471 variation ranges are larger than those found in most marine records (e.g. Simon et al., 2016a and  
 472 references therein) or existing lacustrine sediments (e.g. Horiuchi et al., 2001; Czymzik et al., 2015;  
 473 Du et al., 2018; Tang et al., 2019), except from the closest lacustrine record (i.e. Lake Pupuke in New  
 474 Zealand; Nilsson et al., 2011). Such large-scale concentration variations indicate large changes of Be  
 475 inputs into the lake controlled by transport and deposition processes, but not associated with  $^{10}\text{Be}$   
 476 production rates (e.g. Simon et al., 2016b). The fact that both  $^9\text{Be}$  and  $^{10}\text{Be}$  concentrations vary  
 477 similarly (correlation coefficient  $r=0.97$ ; Sup. Mat. 5) and display two peak intervals also found in  
 478 elemental ratios (e.g. Pb/Al) and magnetic properties (e.g. magnetic susceptibility, HIRM) associated  
 479 with local runoff (Fig. 3), suggests that  $^9\text{Be}$  and  $^{10}\text{Be}$  have the same transport pathway to the lake as  
 480 clastic detrital particles.

481 The  $^{10}\text{Be}$  is used for reconstructing geomagnetic dipole moment variations by normalising  $^{10}\text{Be}$   
482 concentration by its stable isotope  $^9\text{Be}$  (hereinafter Be ratio) to try minimizing environmental  
483 influences (Bourlès et al., 1989). The Be ratio varies from  $14.7 \times 10^{-8}$  to  $23.7 \times 10^{-8}$ , with an average of  
484  $20.7 \pm 2.3 \times 10^{-8}$ . One single Be ratio peak reaching  $28 \pm 1.0 \times 10^{-8}$  at 95 cm depth LAN2 was identified  
485 in a first batch, but not reproduced by a replicate measurement (Table 3). Excluding this  
486 measurement outlier, the highest Be ratio values ( $>22 \times 10^{-8}$ ) are found in the intervals 91–95 cm and  
487 110–117 cm depth LAN2. The latter interval is coeval with the RPI low and excursions  
488 paleomagnetic inclinations (Fig. 10). In the top 90 cm of the core (Holocene facies I), the Be ratio  
489 decreases significantly and reaches its lowest values ( $<18 \times 10^{-8}$ ). These low Be ratio values are  
490 explained by a dilution of  $^{10}\text{Be}$  inputs between the Holocene interval and the Last Glacial Period  
491 (Willenbring and von Blanckenburg, 2010; Sup. Mat. 5). This is likely associated with large-scale  
492 atmospheric reorganization and environmental change within the lake watershed since the last  
493 glacial to interglacial climate transition (Fig. 3) and higher sedimentation rates. Excluding these  
494 Holocene results, Be ratio and RPI have a correlation coefficient of  $r=-0.36$ , and  $r=-0.5$  for the grey  
495 shading interval of figure 10. Considering the large lithological changes observed during the Last  
496 Glacial Period, these moderately good correlations may relate to unaccounted for environmental  
497 biases on both geomagnetic field intensity proxies and/or to unknown lock-in-depth affecting the  
498 RPI (Roberts and Winklhofer, 2004; Nilsson et al., 2018; Simon et al., 2018, 2020b). The phasing  
499 between RPI low and Be ratio peak between 110–117 cm depth LAN2 suggests a very limited lock-in  
500 depth, but data resolution and lithological changes prevent any strong inference.

501

## 502 5. Discussion

### 503 5.1 Depositional processes in Lake Selina and instability of the Last Glacial Period

504 The difference in facies thickness between the two cores could point to event deposits as suggested  
505 by the occasional mm-sized pebbles in facies IIa (black lozenge on Fig. 2 stratigraphic logs). However,  
506 a pebble is also found in facies I and there are no clear and systematic geochemical, magnetic, and  
507 physical evidence for event deposits as reported elsewhere in lakes, fjords and harbour sediments  
508 (e.g. St-Onge et al., 2004; Guyard et al., 2007; Lisé-Pronovost et al., 2014; 2019b). No basal erosive  
509 contacts are observed and the base of facies units contains no vegetal debris, nor present any  
510 systematically higher density (RAD; Fig. 3), higher concentration of ferrimagnetic minerals (ARM,  
511 IRM; Fig. 3), or abnormally shallow paleomagnetic inclinations as would be expected for  
512 instantaneous event deposits such as floods, extreme precipitation or melt runoff. The facies IIa  
513 from lowermost part of core N2 (516–544 cm depth LAN2) is the only one standing out as a likely  
514 event or disturbed deposit because it has oblique structures (visible in the radiograph and photo,  
515 Sup. Mat. 1), a 3-cm vertical black stain, pebbles, flattened inclinations and sharp large amplitude  
516 changes in magnetic grain size proxies ( $k\text{ARM}/\text{IRM}$  and  $\text{MDF}_{\text{ARM}}$ ; Fig. 3). All other facies I, IIa and IIb  
517 are interpreted as distinct types of normal lake sediment deposits and the pebbles are assumed to  
518 have reached the coring sites through a process independent of facies type (e.g. isolated  
519 meteorological event). The difference in facies thickness between cores can alternatively be  
520 attributed to the coring site N1 being closer than N2 to gullies from Lake Westwood to Lake Selina  
521 (Fig. 1C). This is supported by glacial facies (IIa and IIb) being on average 1.6 times thicker in N1 than  
522 N2 (N1/N2) and coarser magnetic particles deposited at coring site N1 than N2 ( $k\text{ARM}/\text{IRM}$  and  
523  $\text{MDF}_{\text{ARM}}$  data of overlapping sections N2-A; Fig. 3). Detrital input from the gullies likely also accounts  
524 for the southwest of Lake Selina being filled in the modern wetland (Fig. 1C). Future geophysical  
525 surveys will be useful to find out the total sediment thickness accumulated in Lake Selina.

526 The Last Glacial Period (MIS 2, 3 and 4) also stands out as more unstable than the previous glacial  
527 period (MIS 6), with greater and more variable facies thickness differences between cores (N1/N2  
528 ratio of facies thicknesses are  $1.82 \pm 1.68$  for the Last Glacial Period and  $1.17 \pm 0.16$  for MIS 6). The  
529 relatively unstable stratigraphy of the Last Glacial Period may be linked to climate and/or human  
530 influence on the transport of clastic sediments to the lake. The distinct climatic signal of the two last  
531 glacial periods in the Lake Selina data may be attributed to their different orbital configurations (e.g.  
532 Ruddiman, 2006) and the global distribution of continental ice (Rohling et al., 2017). Distinct regional  
533 climate signals for the two last glacials are similarly reported elsewhere (e.g. de Abreu, 2003, Fritz et  
534 al., 2004; Roucoux et al., 2011). While MIS 6 is poorly documented in the Australian region, dramatic  
535 climate changes are documented for the Last Glacial Period in the marine and continental temperate  
536 regions of Australia, from cold MIS 4 (de Deckker et al., 2019) to wet MIS 3, and cold and arid MIS 2  
537 (Fletcher and Thomas, 2010; Bostock et al., 2013; Falster et al., 2018; Kemp et al., 2019; Sniderman  
538 et al., 2019). Lake Mungo, in the Willandra Lakes system on mainland Australia (900 km north from  
539 Lake Selina), was permanent during MIS 4 and 2 and underwent large-scale oscillating lake levels  
540 during MIS 3 (Fitzsimmons et al., 2015). At Lake Selina, the frequent alternation of the more  
541 clastic facies IIa and the more organic facies IIb may suggest similar lake level fluctuations with  
542 changing conditions controlled by an on/off type of control. Examples of on/off switches include  
543 temperatures near freezing point and the hydrological connection from Lake Westwood to Lake  
544 Selina. Gullies may have only been active at temperatures above the freezing point at the altitude of  
545 Lake Westwood, less active during arid periods (e.g. MIS 2), and much more active in wet climate  
546 (e.g. wet MIS 3; Falster et al., 2018; Kemp et al., 2019). While there is no evidence for ice cover on  
547 lakes Selina and Westwood, a sedimentary hiatus in the highest altitude Lake Julia (616 m asl) (Fig.  
548 1C) suggests it was permanently ice-covered during MIS 2 (M.-S. Fletcher unpublished data). In  
549 addition to climate, the second non-exclusive hypothesis for the unstable signal of the Last Glacial  
550 Period is Indigenous land use practices such as vegetation burning in the lake catchment that  
551 would have impacted the availability of clastic material for erosion, since humans have occupied  
552 Tasmania for the last 43,000 years (Kiernan et al., 1983; Cosgrove, 1999). These hypotheses on the  
553 conditions controlling the depositional types will be addressed elsewhere using multi-proxy biotic  
554 and abiotic investigations from Lake Selina and Lake Westwood.

555

## 556 5.2 Radiometric dating

557 The heterogeneous lithology of the Last Glacial Period in Lake Selina is problematic for radiocarbon  
558 dating (Fig. 6). Downcore movement of humic acid and contamination has previously been reported  
559 at Lake Selina by Colhoun et al. (1999), precluding the determination of reliable sedimentation rates.  
560 The Radiometric age model obtained here by combining radiocarbon and OSL ages indicates an  
561 average sedimentation rate of 1.5 cm/ka for the Last Glacial Period (MIS 2, 3, 4) and 5.3 cm/ka for  
562 the uppermost facies I (MIS 1) (Fig. 6; Table 4). The average uncertainty of the radiocarbon and OSL-  
563 dated intervals is  $0.8 \pm 0.4$  ka and is  $8.2 \pm 1.9$  ka at  $2\sigma$ , respectively. The OSL dating uncertainty and  
564 low sedimentation rates at Lake Selina prevent resolving any difference in accumulation rates  
565 between the glacial facies IIa and IIb.

566 MIS 2 is arguably the best dated interval of the Last Glacial Period because it is bracketed by the  
567 precisely radiocarbon-dated Holocene and a stratigraphic change from facies IIa to IIb dated at 30 ka  
568 BP (27,361–33,758 cal BP; Radiometric age model) in agreement with the accepted MIS 3/2 global  
569 boundary (29 ka; Lisiecki and Raymo, 2005). The Radiometric age model indicates that the  
570 uppermost facies IIb accumulated during MIS 2, the coldest period of the Last Glacial Cycle and a  
571 presumably arid period in the southeast of mainland Australia (Falster et al., 2018). In contrast, MIS

572 3 includes a series of clastic facies IIa interbedded with facies IIb. This stratigraphy may be linked to  
573 wetter conditions influenced by Southern Westerly Winds during MIS 3 (Kemp et al., 2019), with  
574 increased runoff episodes capable of transporting more detrital particles to the lake. A local source  
575 for the detrital material is supported by the magnetic, beryllium, and elemental data (Fig. 3). The  
576 magnetic assemblage has a detrital signature, with the same magnetic subpopulations present  
577 throughout sediment facies (Fig. 7), and the elemental composition is closely tied to the local  
578 geology (e.g. Pb/Al maxima in facies IIa; Fig. 3). The interpretation of these depositional processes  
579 inferred from the lithology and environmental magnetism results therefore concur to known climate  
580 variability in southeast Australia since MIS 4 using the radiometric age model. Radiometric dating in  
581 Lake Selina sediments is however limited to the uppermost 1.74 m (ca. 80 ka; Fig. 6).

582

### 583 5.3 Dating the full record with climate comparison

584 The environmental sensitivity reflected by contrasting sediment facies in Lake Selina is a challenge  
585 for radiocarbon dating but can be an advantage for climate-based relative dating. To explore this  
586 idea, we use Match dynamic programming (Lisiecki and Lisiecki, 2002) to synchronize climate proxies  
587 from Lake Selina with well-dated Antarctic ice core climate records. This is based on the assumption  
588 of linked millennial-scale climate in Antarctica and Tasmania, as documented in paleo-records from  
589 the Southern Hemisphere mid-latitudes (e.g. Pahnke, 2003; Barrows et al., 2007; Pugh et al., 2009;  
590 Fletcher and Moreno, 2012; Petherick et al., 2013; Abram et al., 2014) and the modern Southern  
591 Annular Mode (Garreaud, 2007; Hendon et al., 2007). Fletcher and Thomas (2010) further use a  
592 pollen transfer function to estimate temperature change from the Colhoun et al. (1999) pollen  
593 record from Lake Selina, identifying a tight coherence between Antarctic and Tasmanian  
594 temperature changes over the Last Glacial Cycle. This coherence supports our assumption of strong  
595 correlation between Tasmanian and Antarctic climatic variations.

596 Dynamic programming allows time series compression and expansion to account for varying  
597 sedimentation rates between compared natural archives, e.g. ice accumulates faster in cold glacial  
598 climates in Antarctica, while lake sediment accumulates faster during warm interglacial climates in  
599 Tasmania. The method is most efficient for large amplitude changes such as glacial/interglacial  
600 climate transitions. Taking this into account, the proxy data are selected based on their superficial  
601 similarity and six tie-points located at large amplitude changes are chosen (Fig. 11A). We use a  
602 double approach matching concurrently a) the inc/coh organic matter proxy from Lake Selina (Fig. 1)  
603 to the dD temperature proxy from EPICA Dome C (EDC; Jouzel et al., 2007; Fig. 11A) and b) the Pb/Al  
604 local runoff proxy from Lake Selina (Fig. 1) to the EDC dust flux (Lambert et al., 2012; Fig. 11B). The  
605 Southern Hemisphere atmospheric circulation links these proxies, with increased winds over the  
606 continental masses generating more atmospheric dust and more rainfall in western Tasmania (Fig.  
607 1A and B). The Match-Climate age model is presented in Figure 12 and gives a basal sediment age of  
608 266 ka. The average estimated sedimentation rates of Match-Climate are 5.2 cm/ka for interglacial  
609 facies I (MIS 5 and 7) and 2.6 cm/ka for glacial facies II (MIS 2, 3, 4, 6, and 8) (Table 4). These  
610 sedimentation rate estimates are similar to those using the Radiometric age model; however, the  
611 Match-Climate age model returns higher estimates for the Last Glacial Period.

612 Despite recognisable patterns of multi-millennial variability through glacial and interglacial climates,  
613 the low variance of climatic proxies over the interval 20–60 ka prevents any reliable visual matching  
614 in this interval (Fig. 11). The inc/coh comparison is poor over that period, with an R value of 0.25  
615 compared to 0.84 for the entire record (Fig. 11). This compromises the applicability of Match-  
616 Climate age model over MIS 2-3. Providing there are no unidentified event deposits obscuring the

617 climate signal in the Lake Selina sequence, this mismatch may be linked to differences in regional  
618 climates in western Tasmania and in Antarctica over the Last Glacial Period, however, a close  
619 alignment between the climate of these regions is reported in the literature (Williams et al., 2009;  
620 Petherick et al., 2013; Kemp et al., 2019; Sniderman et al., 2019). Another possible explanation for  
621 the mismatch is a misinterpretation of the climate proxies. Additional local factors may influence  
622 one of the proxies in addition to the common climate forcing component. For instance, the ratio  
623 inc/coh in Lake Selina may not be solely controlled by atmospheric temperature. Therefore, the  
624 assumption that more organic sediments reflecting warmer temperatures does not hold for  
625 moderate variations, resulting in low R value during MIS 2–3. As discussed in section 5.1,  
626 precipitation and/or Aboriginal landscape management may have played a crucial role in the  
627 transport of clastic material to Lake Selina over the Last Glacial Period, thus decoupling the  
628 sedimentary record from climate through this interval of relatively small-scale climate fluctuations in  
629 the Antarctic record (when compared to the large orbital-scale glacial-interglacial oscillations).

630

## 631 5.4 Dating with paleomagnetism

### 632 5.4.1 Geomagnetic excursions

633 Up to 10 geomagnetic excursions have been recognised in the past 270 ka (Channell et al., 2020 and  
634 references therein). Of these, the Blake (120 ka), Laschamp (41 ka), and Mono Lake (32 ka)  
635 excursions are amongst the most globally documented from sedimentary and volcanic archives, but  
636 previously not from Australia. At Lake Selina, the interval 110–117 cm depth LAN2 is characterised  
637 by Be ratio maxima, RPI low, and excursions paleodirections associated with the Laschamp  
638 geomagnetic excursion (Fig. 10). While the dipole moment modulates the amount of beryllium  
639 isotopes  $^{10}\text{Be}$  produced in the atmosphere, the total field intensity (dipolar and non-dipolar fields)  
640 can be recorded in sediments. Yet due to low sedimentation rates at Lake Selina, the RPI signal is  
641 likely to be dominated by dipolar variations only. Thus, the Be ratio and RPI records from Lake Selina  
642 are expected to vary similarly and are here referred to as geomagnetic field intensity proxies. The  
643 geomagnetic field intensity low attributed to the Laschamp has an estimated duration of about 5 ka  
644 based on the radiometric age model and the excursions interval extends over a longer  
645 period of about 10 ka. These durations are higher than the generally accepted < 2 ka for the  
646 Laschamp excursion (Channell et al., 2020). While this result may indicate undetected higher  
647 sedimentation rates in Lake Selina, a longer overall process duration for the Laschamp excursion was  
648 also reported from the closest available record, derived from volcanic rocks in New Zealand (Ingham  
649 et al., 2017). Korte et al. (2019) show that duration of the Laschamp can vary regionally depending  
650 on its definition and the model used. Such records of the Laschamp excursion in the Australasian  
651 region are therefore useful to document the non-uniform expression at the Earth's surface of a  
652 global phenomenon.

653 A second Be ratio maxima (91–95 cm depth LAN2, yellow highlight Fig. 10) does not correlate with  
654 an RPI low, but corresponds to a slight inclination shift above the  $\text{GAD}+45^\circ$  limit and large amplitude  
655 declination change seen in N2 (however, not in N1). Thus, attributing this second event to a  
656 geomagnetic excursions feature seems speculative. We note however that its occurrence (from  
657 26,632–32,747 cal BP to 23,812–29,041 cal BP; Radiometric age model) corresponds to a time period  
658 where geomagnetic intensity drops have been noted in different records, with the suggestion of two  
659 excursions at ca. 32 ka (e.g. Liddicoat and Coe, 1979; Lund et al., 2017) and ca. 26 ka (Channell et al.,  
660 2016). This timing also corresponds to a moderate Be ratio increase in core MD05-2930 from the  
661 Gulf of Papua, ca. 3500 km north of the Selina Lake (Simon et al., 2016a). Finally, one data point (302  
662 cm depth LAN2) returning a paleomagnetic inclination departure  $>45^\circ$  from the GAD together with



663 large declination variations and low RPI values during MIS 5e (Fig. 9D) correspond to the age of the  
664 Blake geomagnetic excursion (ca. 125 ka) according to the Match-Climate age model (Fig. 12).  
665 However, due to the high organic content and low ferrimagnetic content of the interglacial  
666 sediments (facies I; Fig. 3), the paleomagnetic directions are too poorly defined (Fig. 9) to confidently  
667 attribute these results to a geomagnetic excursion. Future beryllium isotopes analysis over this  
668 interval could be useful to test this interpretation, providing that Be ratio does not suffer from  $^{10}\text{Be}$   
669 dilution as during the Holocene facies I (see section 4.4.3).

670

#### 671 5.4.2 Geomagnetic field intensity proxies

672 The best agreement of the two geomagnetic field intensity proxies (RPI and Be ratio) is found over  
673 the period 20–60 ka (80–140 cm depth LAN2; Fig. 10), the same interval where Match-Climate  
674 applicability is compromised (c.f. section 5.3). We use the Be ratio as our most reliable geomagnetic  
675 field intensity proxy for Match dynamic programming (Lisiecki and Lisiecki, 2002) because it is not  
676 affected by unknown lock-in delay and can be compared to the marine Be ratio stack of similar  
677 resolution (Simon et al., 2016a). The Match-Dipole age model covers the period 20–111 ka and falls  
678 within the Radiometric age model uncertainty for the period of overlap (Fig. 12). The average  
679 estimated sedimentation rate of Match-Dipole is 1.5 cm/ka for the Last Glacial Period facies II (MIS  
680 2, 3, and 4), which is the same as the Radiometric age model estimate (Table 4). We note that  
681 running Match with the Lake Selina RPI data and other independent paleointensity references  
682 (GGF100k, Panovska et al., 2018; GLOPIS, Laj et al., 2014) returns similar results. Comparison of the  
683 Lake Selina geomagnetic field intensity proxies with the reference stack GLOPIS and the geomagnetic  
684 field model GGF100k (Fig. 13) illustrates that Match-Climate overestimates the age of the  
685 geomagnetic field intensity low associated with the Laschamp geomagnetic excursion (41 ka) by  
686 about 6 ka. Prior to 60 ka and back to 80 ka, the reference curves are not readily correlatable (Fig.  
687 13; Laj et al., 2004; Simon et al., 2016a; Panovska et al., 2018), which may reflect dating difficulties in  
688 those records and limits the performance of Match dynamic programming.

689

#### 690 5.5 Age models and uncertainties

691 There is an overall agreement within age uncertainties between all age models developed in this  
692 study, with an overlap of two to three age models (Radiometric, Match-dipole and Match-climate)  
693 for the interval between 81–223 cm depth LAN2 (Fig. 12). An average age model can be calculated  
694 for that interval, or more practically, the appropriate chronology can be selected for a given  
695 purpose, e.g. a climate-independent chronology for paleoclimate investigations (up to 111 ka), and a  
696 geomagnetic-independent chronology for paleomagnetic investigations (full record) (GEOMAGIA50,  
697 Brown et al., 2015; MagIc; <https://www2.earthref.org/MagIc>). Our preferred age model is the  
698 combination of Radiometric (0–70 ka) and Match-Climate (70–265 ka), independently supported by  
699 the Laschamp geomagnetic excursion. This choice is based on i) the Radiometric age model having  
700 age uncertainties encompassing Match-Dipole and most of Match-Climate age models (Fig. 12), and  
701 ii) Match-Climate being considered more reliable than Match-Dipole before 60 ka, where Be ratio  
702 and RPI are not synchronized (Fig. 10) and do not agree with other dipole reconstructions (Fig. 13).  
703 We select 70 ka as the transition point between our optimal age models (Fig. 12). For paleoclimate  
704 discussions, a second preferred age model is the combination of Radiometric (20–0 ka), Match-  
705 Dipole (20–70 or 111 ka) and Match-Climate (70 or 111–265 ka). This second option is aligned to  
706 global dipole millennial-scale variability and the Laschamp geomagnetic excursion (Fig. 13), and it is  
707 independent of climate over the longest possible period.

708 The overlap in independent age models during MIS 2–3–4 gives a means to estimate the uncertainty  
709 on ages obtained from the Match dynamic programming algorithm at Lake Selina. Match dynamic  
710 programming is deterministic; it gives one best solution and does not provide age uncertainties  
711 (Lisiecki and Lisiecki, 2002). Yet estimate of age uncertainty is desirable for the lowermost interval  
712 between 223–44 cm (111–265 ka) that only has one age model based on climate comparison (Fig.  
713 12). The average age difference between the three overlapping chronologies is  $4.4 \pm 1.6$  ka (Fig. 12  
714 inset), which is smaller than the average uncertainty of the OSL-dated interval ( $8.2 \pm 1.9$  ka). We use  
715 the average deviation of the Match-Climate age to the other age models (Radiometric and Match-  
716 Dipole) over the interval where the three age models overlap to estimate the uncertainty on age of  
717 Match-Climate within the lowermost interval exclusively based on climate tuning. An estimated age  
718 uncertainty of  $5.3 \pm 2.3$  ka is obtained. This age uncertainty estimate is comparable to the average  
719 age differences of  $4.1 \pm 2.8$  ka between the Lake Selina stratigraphic transitions and the globally  
720 recognized MIS boundary ages (Lisiecki and Raymo, 2005) (Figure 12).

721

## 722 6. Conclusions

723 This work presents a chronostratigraphic model of Lake Selina core TAS1402, which is the oldest  
724 continuous continental archive in Tasmania. Radiometric dating (radiocarbon and OSL) and relative  
725 dating (paleomagnetism, authigenic  $^{10}\text{Be}/^9\text{Be}$  ratio and climate comparisons) were combined to  
726 construct a chronology extending to  $266 \pm 8$  ka (MIS 8). The 5.5 m lake sediment archive is longer  
727 than the previously recovered 4 m core (Colhoun et al., 1999), and the sedimentation rate estimates  
728 are 5 cm/ka for interglacial periods and 1.5 cm/ka for glacial periods.

729 High-resolution XRF core scanning, beryllium isotopes and magnetic properties enable a matching of  
730 the sediment archive from Tasmania to ice core records of climate proxies in Antarctica. This  
731 agreement was used as a basis to date material beyond the range of radiometric methods (>80 ka)  
732 via a dynamic programming algorithm (Match; Lisiecki and Lisiecki, 2002). The interval 20–60 ka (MIS  
733 2–3), however, returns poor correlations of climate proxies from Lake Selina to Antarctic records, in  
734 contrast to the rest of the record, and limits the application of climate comparison dating over that  
735 interval. We interpret the misfit to be caused by local controls, other than temperature, on the Lake  
736 Selina proxy  $\text{inc}/\text{coh}$  during a period of rapid climate oscillations and the arrival of humans in  
737 Tasmania (at ca. 43 ka). Paleomagnetic comparison over the interval 20–60 ka effectively constrains  
738 the chronology with geomagnetic field intensity proxies; authigenic  $^{10}\text{Be}/^9\text{Be}$  ratio and relative  
739 paleointensity (RPI). The multi-method age modeling approach used here provided two preferred  
740 age models, nearly indiscernible within their uncertainties, which allows the use of a dipole-  
741 independent (270 ka to present) or a climate-independent age model (111 ka to present). The age  
742 uncertainties are lowest in the radiocarbon-dated uppermost 83 cm of the core ( $0.8 \pm 0.4$  ka) and  
743 are estimated to be within 8 ka for the entire sequence.

744 The period 45–35 ka (defined by the Radiometric age model) is a 14 cm interval that spans human  
745 arrival in Tasmania and the demise of the last remaining megafauna in this region. It has an average  
746 Radiometric age model uncertainty of  $7.2 \pm 0.4$  ka and the average age difference for the three  
747 overlapping age models (Radiometric, Match-Climate, and Match-Dipole) is  $2.5 \pm 1.1$  ka.

748 This work presents the first full-vector paleomagnetic record and environmental magnetism study in  
749 Tasmania. Despite a complex magnetic mineral assemblage with magnetic concentration changes up  
750 to one order of magnitude between glacial and interglacial sediment facies, the main paleomagnetic  
751 remanence carrier is detrital oxidised (titano)magnetite throughout all sediment facies. A set of

752 quality criteria were used to select the most reliable data. The paleomagnetic record from Lake  
 753 Selina returns transitional paleomagnetic directions and low geomagnetic field intensity proxies (RPI  
 754 and authigenic  $^{10}\text{Be}/^9\text{Be}$  ratio) at the time of the Laschamp geomagnetic excursion. Comparison of  
 755 the RPI and Be ratio with published virtual axial dipole moment reconstructions over the Last Glacial  
 756 Period indicates negligible lock-in-depth and supports a genuine geomagnetic signal. Thus, providing  
 757 a practical example for combining authigenic  $^{10}\text{Be}/^9\text{Be}$  ratio and paleomagnetism to constrain the  
 758 chronology of lake sediments.

759

## 760 7. Data availability

761 The data presented here will be available on PANGAEA: <https://www.pangaea.de/>. The  
 762 paleomagnetic data will also be archived with the Magnetics Information Consortium (MagIC):  
 763 <http://earthref.org/MagIC/16529>.

764

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778

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## FIGURE CAPTIONS

Figure 1. Site location A) in the Southern Hemisphere mid-latitudes and B) in the West Coast Range of Tasmania where rainfall is strongly correlated to the westerly winds. A and B are modified from Lise-Pronovost et al (2019). C) Two sites were cored in the approximate center of Lake Selina. The 10 m contours and hydrographic data are from Land Information System Tasmania (LIST) and the satellite imagery from ESRI.

Figure 2. A) Cores LA, N1 and N2 simplified lithology, correlation and sampling. Note the different thickness of the Last Glacial Period sediment facies in cores N1 and N2. Individual core sections are indicated to the left of each lithology log. Core section N2-A was divided in sub-sections a and b after a sediment gap formed during splitting. Black zones indicate sampling locations for radiocarbon ( $^{14}\text{C}$ ), magnetic analysis (mag), optically stimulated luminescence (OSL), and beryllium isotopes (Be). B) Large amplitude change in the XRF ratio Inc/coh is used to correlate the gravity core LA with the Nesje cores N1 and N2. C) The composite depth LAN2 is the reference depth scale used in this work. Data from core N1 is transferred to LAN2 depth using a series of 24 stratigraphic tie-points (they are the horizontal lines of the stratigraphic log) and linear interpolation.

Figure 3. Physical, elemental and magnetic properties of core TAS1402. Upper panel: XRF and beryllium data including pass-through radiation (RAD; density proxy), incoherence/coherence (Inc/coh; organic content proxy), authigenic  $^{10}\text{Be}$  and  $^9\text{Be}$  isotopes and elemental ratio Pb/Al (local runoff proxies), and Ti/K (detrital grain size proxy). Middle panel: magnetic properties including the remanence ratio kARM/IRM, median destructive field of the anhysteretic remanent magnetisation ( $\text{MDF}_{\text{ARM}}$ ), and magnetisation ratio (Mr/Ms) (magnetic grain size proxies), S-ratio (proportion of low coercivity minerals), HIRM (content of high coercivity minerals). Lower panel: ferrimagnetic content proxies including the volumetric magnetic susceptibility ( $k_{\text{LF}}$ ), natural remanent magnetisation (NRM) on log and linear scales, anhysteretic and isothermal remanent magnetizations (ARM and IRM). The simple lithology log with sediment facies is shown for the reference depth LAN2. The overlapping section N2-A for the magnetic grain size proxies is shown to illustrate finer particles are deposited at site N2 compared to site N1. Grey shading indicates facies IIa intervals with approximately one order of magnitude greater remanence acquisition efficiency.

Figure 4. Principal component analysis biplot of the micro-XRF core scanning data. The length and color of the arrow reflect the contribution per variable.

Figure 5. A) Measured moisture contents for OSL and DOS samples between 81 and 190 cm depth LAN2, where OSL ages were obtained. B) Gamma dose rates for the same OSL and DOS samples derived from ICP-MS(OES) measurements of U, Th and K and used for age determination. C) Beta dose rates for the same OSL and DOS samples obtained from GM-25-5 beta counting and used for age determination. D) The ratio of beta dose rates derived from GM-25-5 beta counting and ICP-MS(OES) measurements for the same OSL and DOS samples. The stippled line denotes unity. Error bars are all one standard error. The simple lithology log with sediment facies is shown.

Figure 6. A) Age-depth model from Bayesian modeling (rbacon; Blaauw and Christen, 2011) of calibrated radiocarbon ( $^{14}\text{C}$ ) and optically stimulated luminescence (OSL) ages (Tables 1 and 2). The OSL ages are

obtained from core N1 and transferred to the reference depth LAN2 based on stratigraphic correlations (Supplementary Material 2). B) Contamination effect on radiocarbon ages (Pigati et al., 2007) explains the excluded “too young”  $^{14}\text{C}$  ages that plateau at ca. 40 ka BP.

**Figure 7.** Magnetic mineralogy of Lake Selina interglacial sediments (facies I) and glacial sediment (facies II). Representative sample N1-A18 (81.8 cm depth LAN2) for facies I includes A) hysteresis loop, B) thermomagnetic curves of (top panel) magnetic remanence measured from room temperature to 10 Kelvin saturation isothermal remanent magnetisation (RTSIRM), and (base panel) magnetic susceptibility up to 700 °C, C) unmixed magnetic populations (P1, P2, P3) of the isothermal remanent magnetisation (IRM) acquisition curve (MaxUnmix; Maxbauer et al., 2016), and D) first order reversal curve (FORC) diagram (FORCinel; Harrison and Feinberg, 2008). Representative sample N1-C37 (365.3 cm depth LAN2) for facies II includes E) hysteresis loop, F) thermomagnetic curves, G) unmixed magnetic populations (MaxUnmix; Maxbauer et al., 2016), and H) FORC diagram (FORCinel; Harrison and Feinberg, 2008).

**Figure 8.** The biplot of the magnetic remanence ratios  $\text{IRM}(-300\text{mT})/\text{SIRM}$  and  $\text{ARM}(40\text{mT})/\text{SARM}$  informs on the mineralogy of the remanence carriers (Peters and Thompson, 1998). Greater ARM hardness ( $\text{ARM}(40\text{mT})/\text{SARM}$ ) in facies I also reflects finer magnetic grain size than in facies II (visible in Fig. 2).

**Figure 9.** Paleomagnetic directions results for A) raw data and B) data after selection criteria, and relative paleointensity estimates for C) raw data and D) selected data. A line is not used to connect the data points from duplicate core N2-A (blue) because of possibly disturbed stratigraphy. Grey shading indicates the non-selected intervals with approximately one order of magnitude greater remanence acquisition efficiency (cf. Figure 3). A few individual samples display peak values (out of scale) interpreted as having NRM overprint.

**Figure 10.** The paleomagnetic record from Lake Selina uppermost 2.25 m combines sediment magnetic remanence and authigenic beryllium isotopes. From left to right; the paleomagnetic inclination, relative declination and geomagnetic field intensity proxies (Be ratio and RPI). A line is not used to connect the data points from duplicate core N2-A (blue) because of possibly disturbed stratigraphy. The yellow rectangle at 100-117 cm depth LAN2 indicates excursions directions (duration about 10 kyr) and intensity (duration about 5 kyr) at the time of the Laschamp geomagnetic excursion. The yellow rectangle at 91-95 cm depth LAN2 indicates a younger Be ratio maxima, however with inconclusive magnetic remanence results. The geomagnetic field intensity proxies are scaled and the Be ratio axis is reversed. The grey highlight indicates the best fit among geomagnetic field intensity proxies where lock-in depth in RPI appears to be absent.

**Figure 11.** Proxy data from Lake Selina (this study) and Antarctica EDC ice core (Jouzel et al, 2007; Lambert et al, 2012; Veres et al, 2013) used to build the Match-Climate age model using Match dynamic programming (Liesicki and Liesicki, 2002). A) Temperature proxies incoherence/coherence (inc/coh; this study) and dD (Jouzel et al, 2007), and B) atmospheric circulation proxies Pb/Al (this study) and dust flux (Lambert et al, 2012). The top two curves from Lake Selina are versus depth and Match-Climate age, and the lower curves are the reference proxy data from Antarctica on the Antarctic Ice Core Chronology 2012 (Veres et al, 2013). Vertical dashed lines indicate the six tie-points used for Match programming. Correlation coefficients are indicated for the Match-Climate dated interval 18-266 ka and for the highlighted interval in grey representing the period 20-60 ka (MIS2 and 3).

**Figure 12.** Lake Selina age-depth models. Age of selected stratigraphic changes are indicated on the right according to the age models Radiometric, Match-Climate and Match-Dipole, and compared to the MIS boundary ages (Liesicki and Raymo, 2005). The interpreted MIS for Lake Selina are indicated next to the simplified stratigraphic log.

**Figure 13.** Lake Selina geomagnetic field intensity proxies (Be ratio and RPI) on the age-depth models A) Radiometric, B) Match-Climate, and C) Match-Dipole, compared to D) dipole references (Be stack, Simon et al., 2016; GLOPIS, Laj et al, 2014; GGF100k, Panovska et al, 2018). The highlighted interval in grey represents the period 20-60 ka (MIS2 and 3) and the best fit interval between RPI and Be ratio defined in Fig. 10. Yellow shading indicates the geomagnetic intensity low associated to the Laschamp excursion (41 ka). The comparison illustrates the age overestimation of Match-Climate over the period 20-60 ka BP because of poor correlation in climate proxies (Fig. 9) and possibly disturbed stratigraphy of core N2-A. Note the different trends in dipole references for the period 60-80 ka BP.

Table 1. Radiocarbon ages from TAS1402. The selected samples for age-depth modelling are bold.

Lab ID	Core	Section depth (cm)	Depth LAN2 (cm)	+/- error depth	Radiocarbon age	error (1 sigma)	Min	Max	Mean Age (cal BP)
							Calibrated age (cal BP)	Calibrated age (cal BP)	
surface			<b>0.0</b>		<b>-64</b>				<b>-64</b>
<b>D-AMS 009187</b>	LA	<b>11-11.5</b>	<b>11.3</b>	<b>0.3</b>	<b>2614</b>	<b>27</b>	<b>2548</b>	<b>2756</b>	<b>2652</b>
<b>D-AMS 012498</b>	LA	<b>13.75-14</b>	<b>13.9</b>	<b>0.1</b>	<b>3649</b>	<b>28</b>	<b>3876</b>	<b>3978</b>	<b>3927</b>
<b>D-AMS 017905</b>	LA	<b>17.25-18</b>	<b>17.6</b>	<b>0.4</b>	<b>4589</b>	<b>26</b>	<b>5063</b>	<b>5317</b>	<b>5190</b>
<b>D-AMS 017906</b>	LA	<b>30.5-31</b>	<b>30.8</b>	<b>0.3</b>	<b>7194</b>	<b>33</b>	<b>7945</b>	<b>8010</b>	<b>7978</b>
<b>D-AMS 009086</b>	LA	<b>37-37.5</b>	<b>37.3</b>	<b>0.3</b>	<b>8515</b>	<b>39</b>	<b>9478</b>	<b>9531</b>	<b>9505</b>
<b>D-AMS 014240</b>	LA	<b>42.25-42.5</b>	<b>42.4</b>	<b>0.1</b>	<b>8723</b>	<b>36</b>	<b>9553</b>	<b>9672</b>	<b>9613</b>
<b>D-AMS 014241</b>	LA	<b>52.25-52.5</b>	<b>52.4</b>	<b>0.1</b>	<b>10414</b>	<b>35</b>	<b>12147</b>	<b>12462</b>	<b>12305</b>
<b>D-AMS 014242</b>	LA	<b>62.25-62.5</b>	<b>62.4</b>	<b>0.1</b>	<b>11726</b>	<b>41</b>	<b>13496</b>	<b>13594</b>	<b>13545</b>
<b>D-AMS 014243</b>	LA	<b>67.75-68</b>	<b>67.9</b>	<b>0.1</b>	<b>12123</b>	<b>40</b>	<b>13860</b>	<b>14060</b>	<b>13960</b>
<b>D-AMS 014244</b>	LA	<b>72.25-72.5</b>	<b>72.4</b>	<b>0.1</b>	<b>12457</b>	<b>53</b>	<b>14327</b>	<b>14834</b>	<b>14581</b>
<b>D-AMS 015348</b>	<b>N2-A</b>	<b>23-23.25</b>	<b>73.1</b>	<b>0.1</b>	<b>12965</b>	<b>44</b>	<b>15356</b>	<b>15556</b>	<b>15456</b>
<b>D-AMS 015349</b>	<b>N2-A</b>	<b>30-30.25</b>	<b>80.1</b>	<b>0.1</b>	<b>14736</b>	<b>50</b>	<b>17948</b>	<b>18144</b>	<b>18046</b>
D-AMS 015350	N2-A	34-34.25	84.1	0.1	22857	86	27135	27259	27197
D-AMS 009079	N2-A	37-37.5	87.3	0.3	19048	71	22909	23016	22963
D-AMS 027826	N2-A	38-38.25	88.1	0.1	25130	106	29158	29503	29331
D-AMS 027827	N2-A	46.5-46.75	96.6	0.1	16581	61	19893	20087	19990
D-AMS 027828	N2-A	56.75-57	106.9	0.1	18510	68	22333	22449	22391
D-AMS 027829	N2-A	62.5-62.75	112.6	0.1	32971	180	37067	37557	37312
D-AMS 009080	N2-A	64.75-65.25	115.0	0.3	32778	181	36881	37380	37131
D-AMS 027830	N2-A	70-70.25	120.1	0.1	37330	276	41859	42160	42010
D-AMS 027831	N2-A	77-77.25	127.1	0.1	NDFB				
D-AMS 015351	N2-A	84-84.25	134.1	0.1	34529	219	39396	39829	39613
D-AMS 009082	N2-A	94.25-94.5	144.4	0.1	37542	249	41985	42214	42100
D-AMS 015352	N2-A	102-102.25	152.1	0.1	35920	246	40794	41205	41000
D-AMS 009084	N2-A	110-110.5	160.3	0.3	14883	51	18146	18244	18195
D-AMS 015335	N2-A	118-118.25	168.1	0.1	NDFB				
D-AMS 009088	N2-B	12.5-13	183.8	0.3	33494	182	37750	38718	38234
D-AMS 015354	N2-B	24.5-24.75	195.6	0.1	NDFB				
D-AMS 015355	N2-B	48.25-48.5	219.4	0.1	NDFB				

Table 2. Dose rate data, equivalent dose ( $D_e$ ) and overdispersion (OD) values, and ages from core TAS1402-N1. The samples core section depths were transferred to LAN2 equivalent depth (Sup. Mat. Table 1).

Sample	Core	Section depth (cm)	Average depth LAN2 (cm)	Water content (%)	Environmental dose rate (Gy/ka)					No of aliquots	CAM $D_e$ (Gy)	Overdispersion (%)	OSL age (ka)
					Gamma	Beta	Cosmic	Alpha	Total				
OSL1	N1-A	44-49	83.5	40 (30)	$0.75 \pm 0.05$	$1.26 \pm 0.07$	$0.13 \pm 0.02$	0.03	$2.17 \pm 0.09$	6/6	$55.7 \pm 2.7$	$4.8 \pm 7.6$	$25.7 \pm 1.7$
OSL2/3	N1-A	55-61	90.0	40 (32)	$0.73 \pm 0.03$	$1.15 \pm 0.07$	$0.13 \pm 0.02$	0.03	$2.04 \pm 0.08$	10/10	$49.2 \pm 1.9$	0	$24.2 \pm 1.4$
OSL4	N1-A	67-74	102.2	40 (22)	$0.64 \pm 0.04$	$0.99 \pm 0.05$	$0.12 \pm 0.02$	0.03	$1.78 \pm 0.06$	10/10	$62.2 \pm 2.8$	$14.5 \pm 5.8$	$35 \pm 2.1$
OSL5	N1-A	74-80	111.2	40 (20)	$0.64 \pm 0.03$	$1.04 \pm 0.05$	$0.12 \pm 0.02$	0.03	$1.83 \pm 0.06$	9/10	$75.6 \pm 4.8$	$11.6 \pm 6.9$	$41.3 \pm 3.1$
OSL6	N1-A	80-86	117.2	40 (25)	$0.67 \pm 0.04$	$1.05 \pm 0.05$	$0.12 \pm 0.02$	0.03	$1.87 \pm 0.07$	10/10	$102.1 \pm 11.5$	$32.2 \pm 8.7$	$54.6 \pm 6.6$
OSL7	N1-A	89-92	123.5	40 (31)	$0.67 \pm 0.04$	$1.11 \pm 0.07$	$0.12 \pm 0.02$	0.03	$1.93 \pm 0.08$	9/10	$113.6 \pm 7.0$	$14.6 \pm 5.4$	$58.9 \pm 4.6$
OSL8	N1-A	99-101	132.5	40 (32)	$0.73 \pm 0.05$	$1.12 \pm 0.07$	$0.12 \pm 0.02$	0.03	$2.01 \pm 0.09$	10/10	$113.7 \pm 8.0$	$18.8 \pm 5.7$	$56.7 \pm 4.8$
OSL9	N1-A	105-108	137.3	40 (30)	$0.71 \pm 0.05$	$1.27 \pm 0.07$	$0.12 \pm 0.02$	0.03	$2.13 \pm 0.09$	9/10	$144.7 \pm 7.9$	$11.0 \pm 5.3$	$67.9 \pm 4.9$
OSL10	N1-A	108-111	139.8	40 (29)	$0.71 \pm 0.05$	$1.12 \pm 0.06$	$0.12 \pm 0.02$	0.03	$1.99 \pm 0.08$	9/10	$126.4 \pm 7.5$	$15.6 \pm 4.8$	$63.6 \pm 4.8$
OSL11	N1-B	0-5	158.1	40 (26)	$0.68 \pm 0.04$	$1.00 \pm 0.05$	$0.12 \pm 0.02$	0.03	$1.83 \pm 0.07$	10/10	$138.9 \pm 5.5$	$5.8 \pm 5.6$	$75.8 \pm 4.4$
OSL12	N1-B	5-10	163.8	40 (25)	$0.67 \pm 0.04$	$0.99 \pm 0.05$	$0.12 \pm 0.02$	0.03	$1.81 \pm 0.07$	9/10	$144.5 \pm 16.5$	$30.0 \pm 8.9$	$79.8 \pm 9.7$
OSL13	N1-B	18-23	175.0	40 (24)	$0.64 \pm 0.04$	$0.93 \pm 0.05$	$0.12 \pm 0.02$	0.03	$1.72 \pm 0.06$	9/10	$143.9 \pm 6.1$	$21.7 \pm 6.3$	$83.6 \pm 5$

Table 3. Beryllium isotopes results. Replicate measurements are italic.

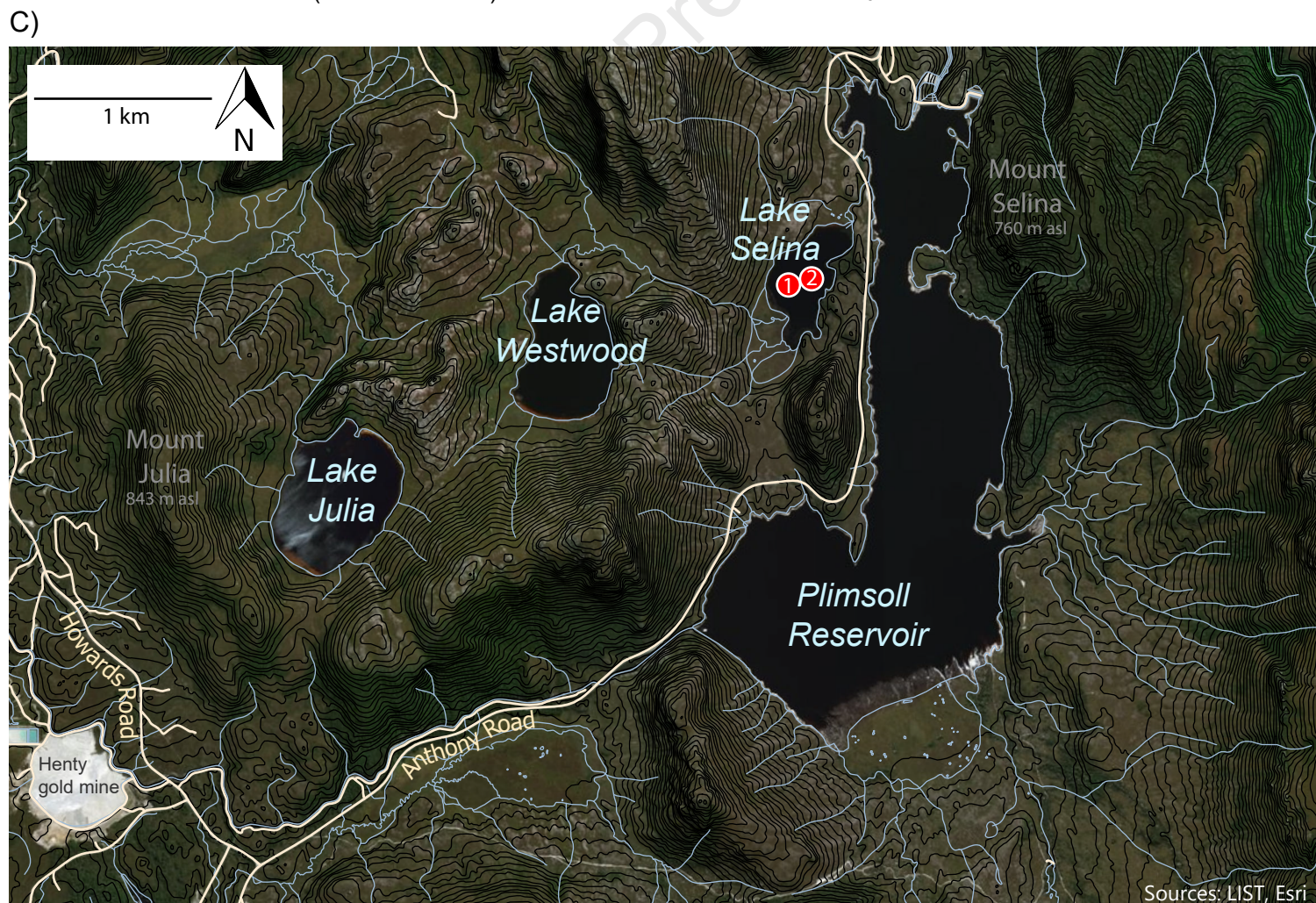
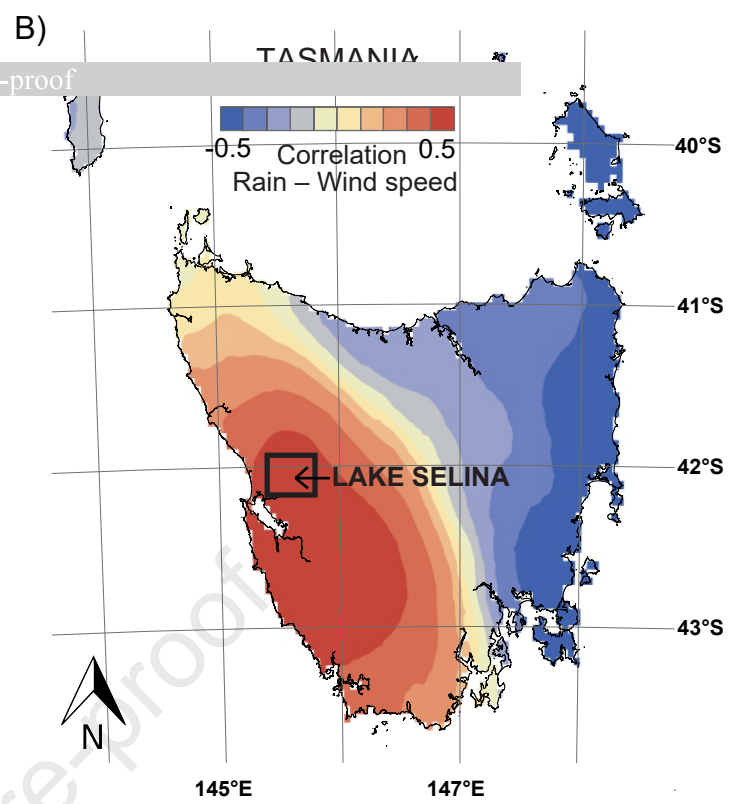
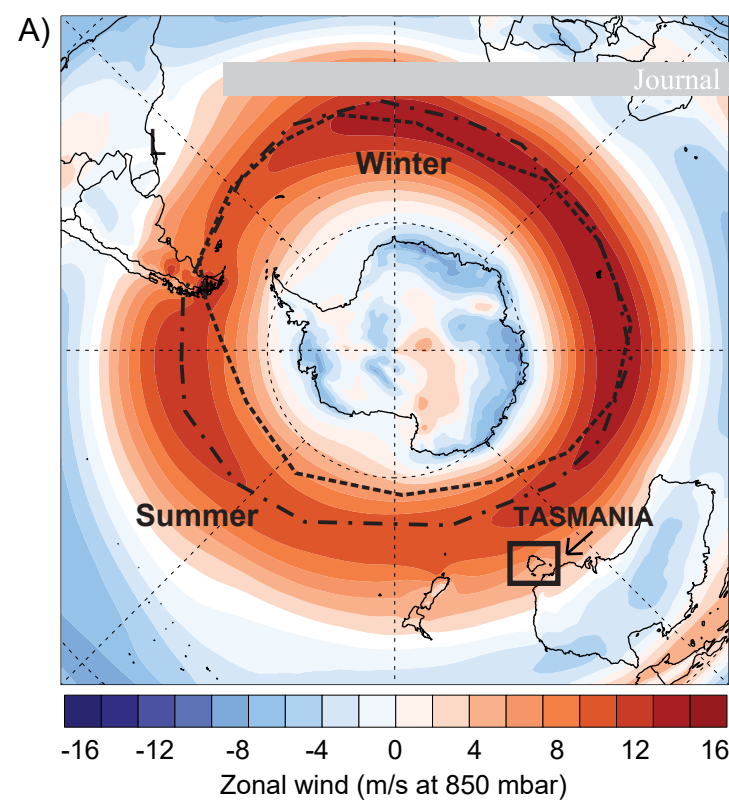
Sample	Core	Section depth (cm)	Depth LAN2 (cm)	Measured $^{10}\text{Be}$ / $^9\text{Be}$ ( $10^{-11}$ )	Authigenic decay corrected $^{10}\text{Be}$ ( $10^8$ at.g $^{-1}$ )	Authigenic $^9\text{Be}$ ( $10^{16}$ at.g $^{-1}$ )	Authigenic decay corrected $^{10}\text{Be}$ / $^9\text{Be}$ ( $10^{-8}$ )
1	N1-A	1.2	43.7	3.146 ± 0.046	8.382 ± 0.124	0.508 ± 0.014	16.513 ± 0.514
4	N1-A	8.4	50.6	3.366 ± 0.048	9.422 ± 0.136	0.630 ± 0.013	14.949 ± 0.375
6	N1-A	13.2	55.3	4.951 ± 0.087	13.121 ± 0.231	0.790 ± 0.006	16.603 ± 0.318
9	N1-A	20.4	62.2	7.052 ± 0.122	15.227 ± 0.264	0.833 ± 0.012	18.281 ± 0.407
12	N1-A	27.6	69.1	7.927 ± 0.130	26.103 ± 0.427	1.771 ± 0.021	14.738 ± 0.298
15	N1-A	34.8	76.1	16.770 ± 0.280	36.718 ± 0.614	2.173 ± 0.065	16.895 ± 0.576
18	N1-A	42	81.8	9.803 ± 0.168	19.496 ± 0.335	1.073 ± 0.021	18.173 ± 0.473
20	N1-A	46.8	83.3	4.504 ± 0.078	9.787 ± 0.169	0.484 ± 0.010	20.230 ± 0.549
22	N1-A	51.6	84.0	6.345 ± 0.105	14.398 ± 0.238	0.726 ± 0.004	19.838 ± 0.346
24	N1-A	56.4	87.4	7.110 ± 0.179	15.505 ± 0.389	0.758 ± 0.010	20.449 ± 0.578
25	N1-A	58.8	89.2	5.744 ± 0.073	13.883 ± 0.176	0.735 ± 0.021	18.887 ± 0.600
26	N1-A	61.2	91.0	6.761 ± 0.170	14.991 ± 0.376	0.632 ± 0.013	23.735 ± 0.760
27	N1-A	63.6	92.8	7.432 ± 0.092	20.083 ± 0.249	0.869 ± 0.024	23.118 ± 0.702
27	<i>N1-A</i>	<i>63.6</i>	<i>92.8</i>	<i>6.084 ± 0.076</i>	<i>21.142 ± 0.263</i>	<i>0.908 ± 0.026</i>	<i>23.292 ± 0.739</i>
28	N1-A	66	94.6	6.182 ± 0.157	12.316 ± 0.312	0.440 ± 0.012	28.004 ± 1.030
28	<i>N1-A</i>	<i>66</i>	<i>94.6</i>	<i>3.716 ± 0.047</i>	<i>12.041 ± 0.151</i>	<i>0.521 ± 0.013</i>	<i>23.126 ± 0.645</i>
29	N1-A	68.4	96.7	3.368 ± 0.050	7.741 ± 0.116	0.393 ± 0.010	19.684 ± 0.597
30	N1-A	70.8	100.0	1.499 ± 0.041	3.504 ± 0.097	0.158 ± 0.004	22.144 ± 0.871
31	N1-A	73.2	103.3	2.613 ± 0.039	5.794 ± 0.087	0.262 ± 0.008	22.090 ± 0.738
32	N1-A	75.6	106.7	2.494 ± 0.065	5.591 ± 0.146	0.257 ± 0.006	21.793 ± 0.742
33	N1-A	78	110.0	3.484 ± 0.089	7.672 ± 0.196	0.332 ± 0.009	23.084 ± 0.869
34b	N1-A	80.4	113.3	2.795 ± 0.071	5.996 ± 0.153	0.252 ± 0.002	23.760 ± 0.648
34a	N1-A	82.8	115.9	4.299 ± 0.109	9.739 ± 0.247	0.421 ± 0.007	23.121 ± 0.705
35	N1-A	85.2	117.4	3.487 ± 0.089	8.117 ± 0.207	0.347 ± 0.004	23.390 ± 0.667
36	N1-A	87.6	118.8	6.764 ± 0.171	15.075 ± 0.380	0.710 ± 0.013	21.226 ± 0.660
37	N1-A	90	121.1	3.572 ± 0.051	7.872 ± 0.113	0.411 ± 0.009	19.135 ± 0.498
38	N1-A	92.4	123.5	5.875 ± 0.148	13.196 ± 0.332	0.626 ± 0.007	21.087 ± 0.579



39	N1-A	94.8	125.9	13.422	± 0.178	29.021	± 0.385	1.481	± 0.043	19.601	± 0.629
40	N1-A	97.2	128.3	9.994	± 0.250	23.112	± 0.577	1.087	± 0.006	21.257	± 0.544
41	N1-A	99.6	130.7	6.179	± 0.078	15.911	± 0.200	0.817	± 0.017	19.479	± 0.478
42	N1-A	102	132.6	10.647	± 0.268	23.735	± 0.597	1.092	± 0.006	21.736	± 0.560
44	N1-A	106.8	135.9	7.579	± 0.190	16.865	± 0.423	0.761	± 0.008	22.160	± 0.599
46	N1-A	111.6	139.9	6.864	± 0.114	15.245	± 0.254	0.761	± 0.011	20.030	± 0.449
48	N1-A	116.4	143.5	6.020	± 0.099	13.467	± 0.221	0.694	± 0.012	19.419	± 0.469
50	N1-A	121.2	146.9	5.447	± 0.091	10.664	± 0.179	0.499	± 0.008	21.375	± 0.499
53	N1-A	128.4	151.9	5.124	± 0.091	11.807	± 0.209	0.528	± 0.014	22.370	± 0.704
1	N1-B	1.2	156.8	5.452	± 0.090	11.620	± 0.193	0.561	± 0.004	20.700	± 0.367
4	N1-B	8.4	164.9	2.020	± 0.035	4.663	± 0.080	0.221	± 0.005	21.108	± 0.632
7	N1-B	15.6	172.0	3.862	± 0.065	8.867	± 0.150	0.392	± 0.004	22.630	± 0.445
9	N1-B	20.4	175.0	1.265	± 0.021	3.099	± 0.051	0.135	± 0.004	23.007	± 0.749
12	N1-B	27.6	182.7	3.137	± 0.046	6.992	± 0.103	0.361	± 0.011	19.352	± 0.642
15	N1-B	34.8	189.4	3.298	± 0.047	7.783	± 0.111	0.390	± 0.005	19.940	± 0.368
18	N1-B	42	196.2	4.689	± 0.069	10.606	± 0.156	0.500	± 0.010	21.197	± 0.532
21	N1-B	49.2	202.9	2.719	± 0.041	6.123	± 0.091	0.279	± 0.006	21.972	± 0.571
24	N1-B	56.4	209.7	1.891	± 0.030	4.484	± 0.072	0.198	± 0.005	22.694	± 0.655
27	N1-B	63.6	217.6	2.246	± 0.038	5.231	± 0.088	0.226	± 0.005	23.136	± 0.630
29	N1-B	68.4	222.3	3.661	± 0.053	8.328	± 0.120	0.400	± 0.008	20.825	± 0.518

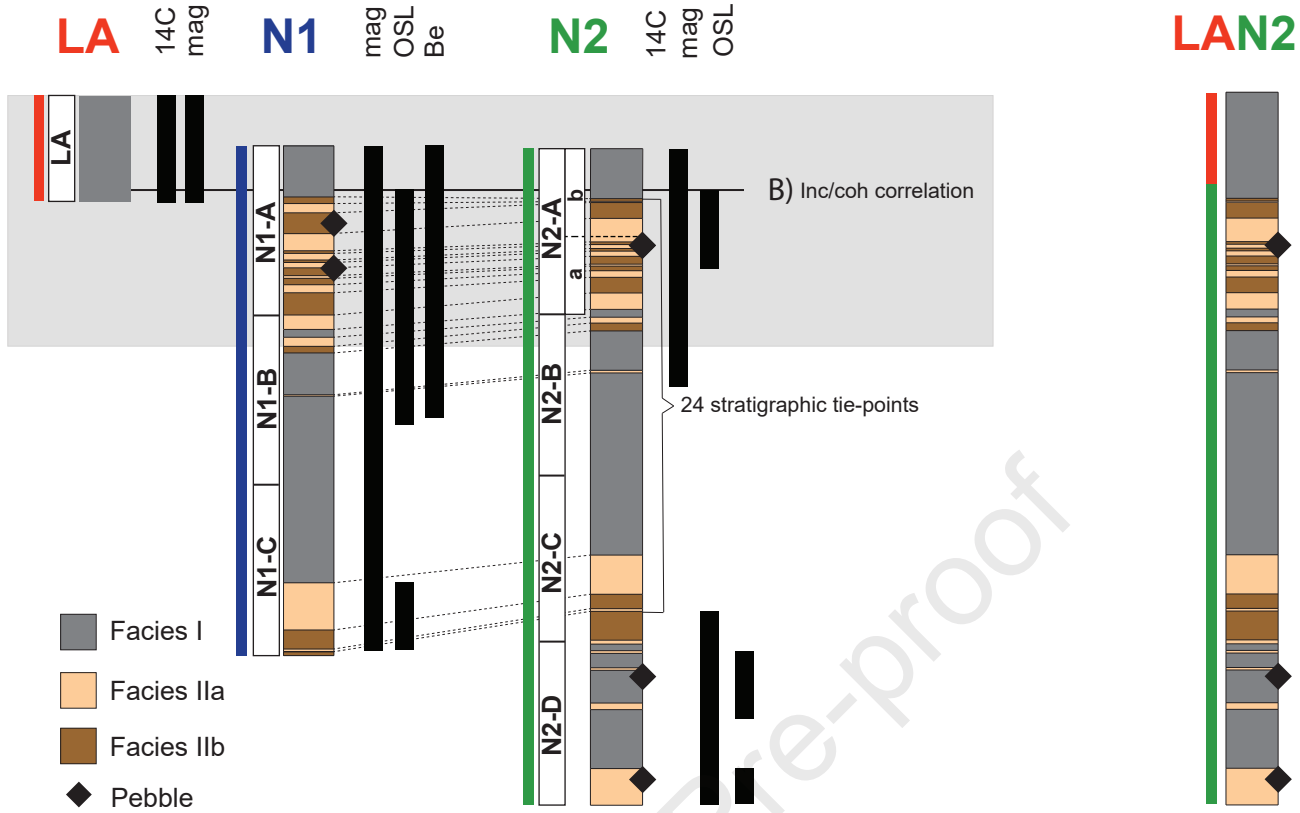
Table 4. Average estimated sedimentation rates for each age model. The assigned marine isotope stages (MIS) for Lake Selina are shown in Figure 11. In cases where an age model does not cover an entire MIS, the covered sub-stages are indicated in parentheses.

	<b>Average sedimentation rate (cm/ka)</b>		
	<b>Radiometric</b>	<b>Match-Climate</b>	<b>Match-Dipole</b>
<i>Facies</i>			
I	5.1	5.2	1.6
II	1.5	2.6	1.5
<i>MIS</i>			
MIS 1	5.3		
MIS 2, 3, 4	1.5	2.5	1.5
MIS 5	1.9 (MIS 5a)	6.2	1.6 (MIS 5a, 5c)
MIS 6		2.4	
MIS 7		3.0	
MIS 8		1.9	

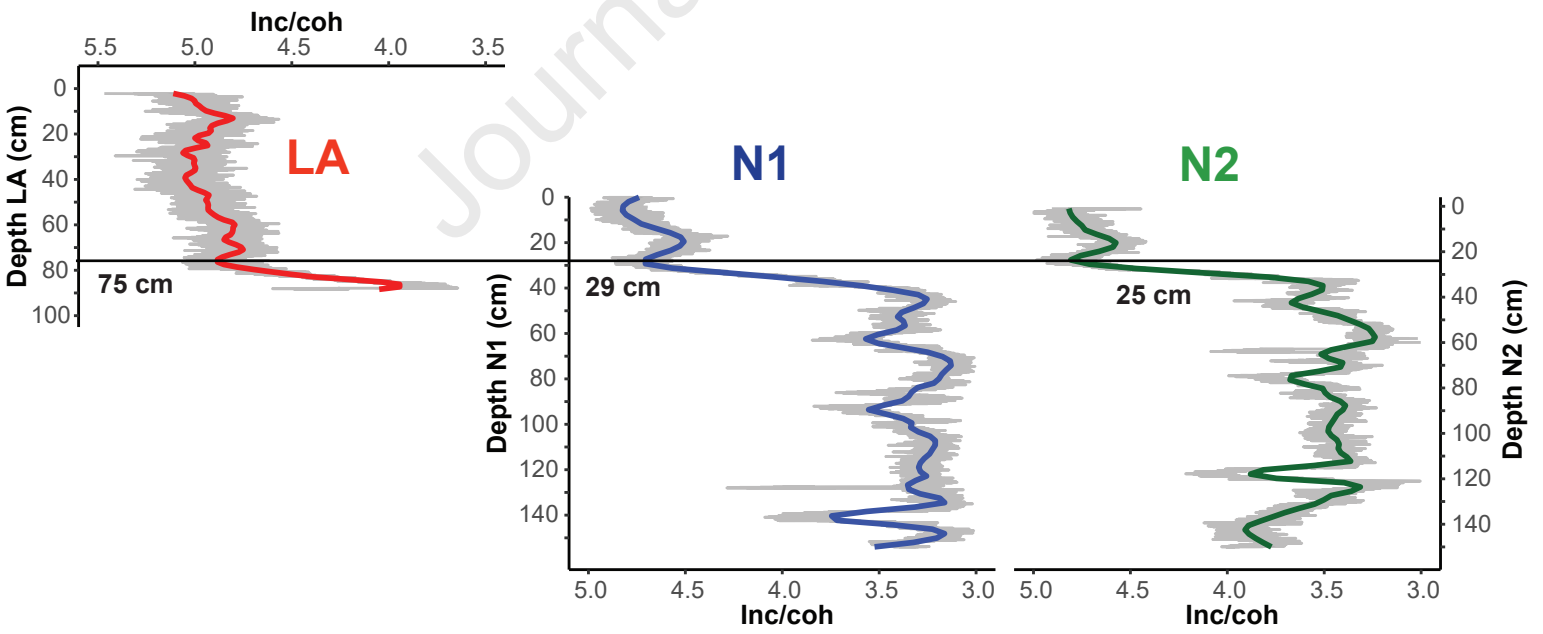


A) Cores correlation and sampling

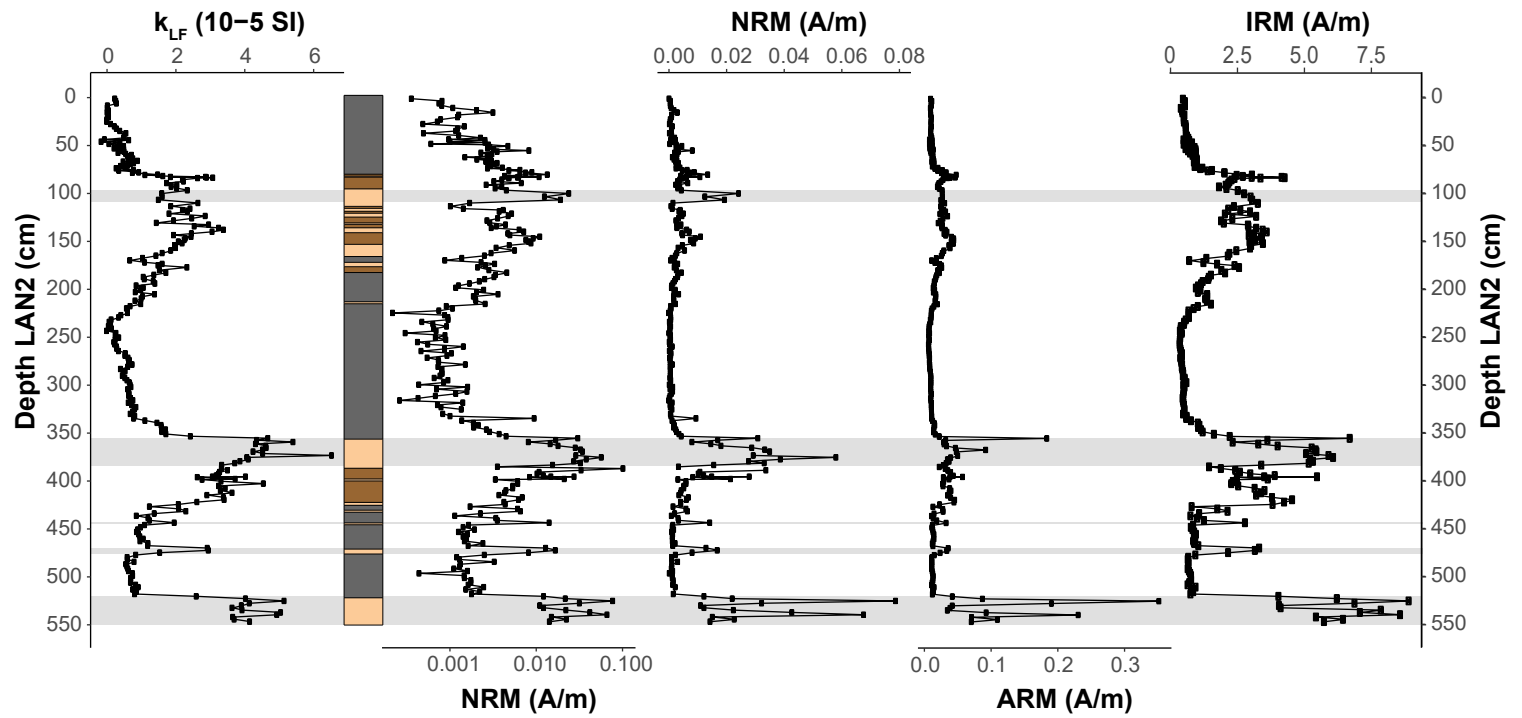
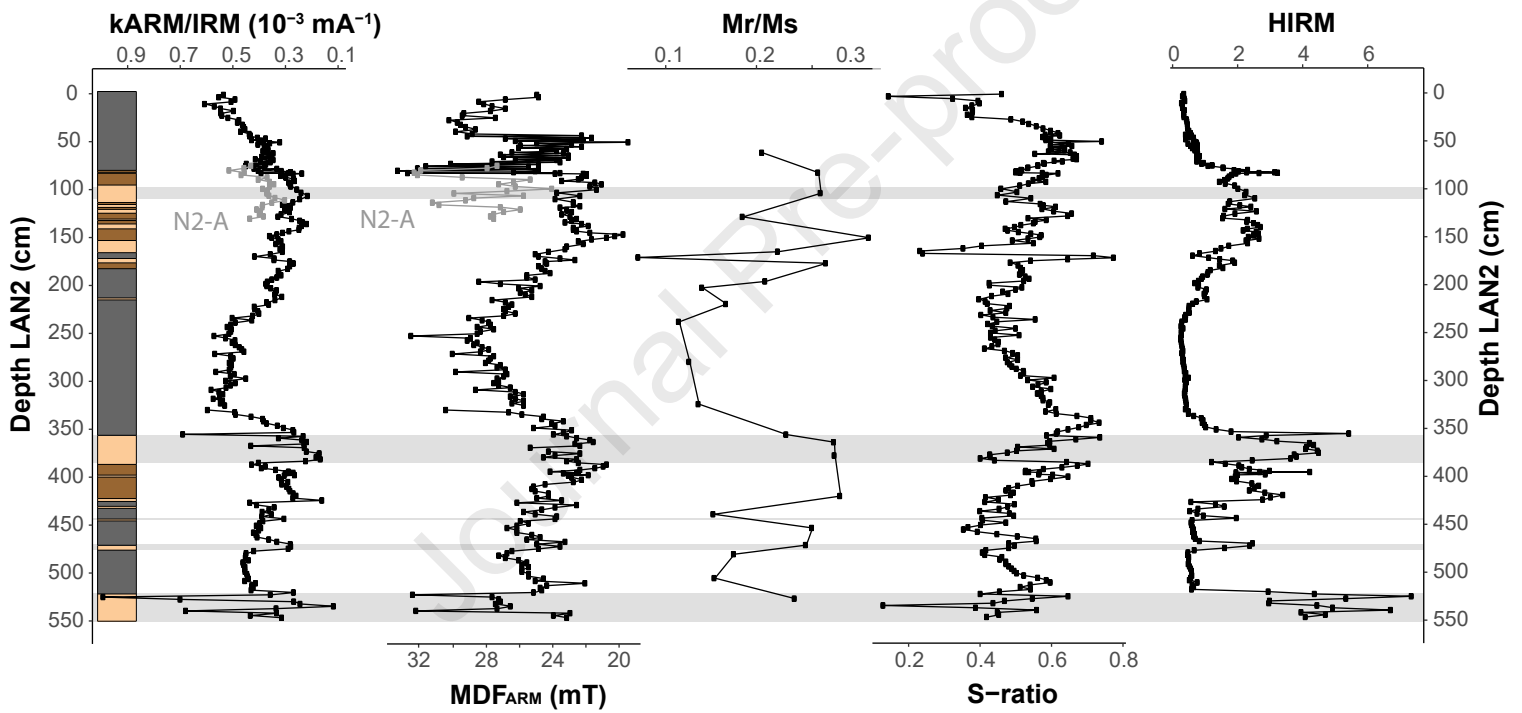
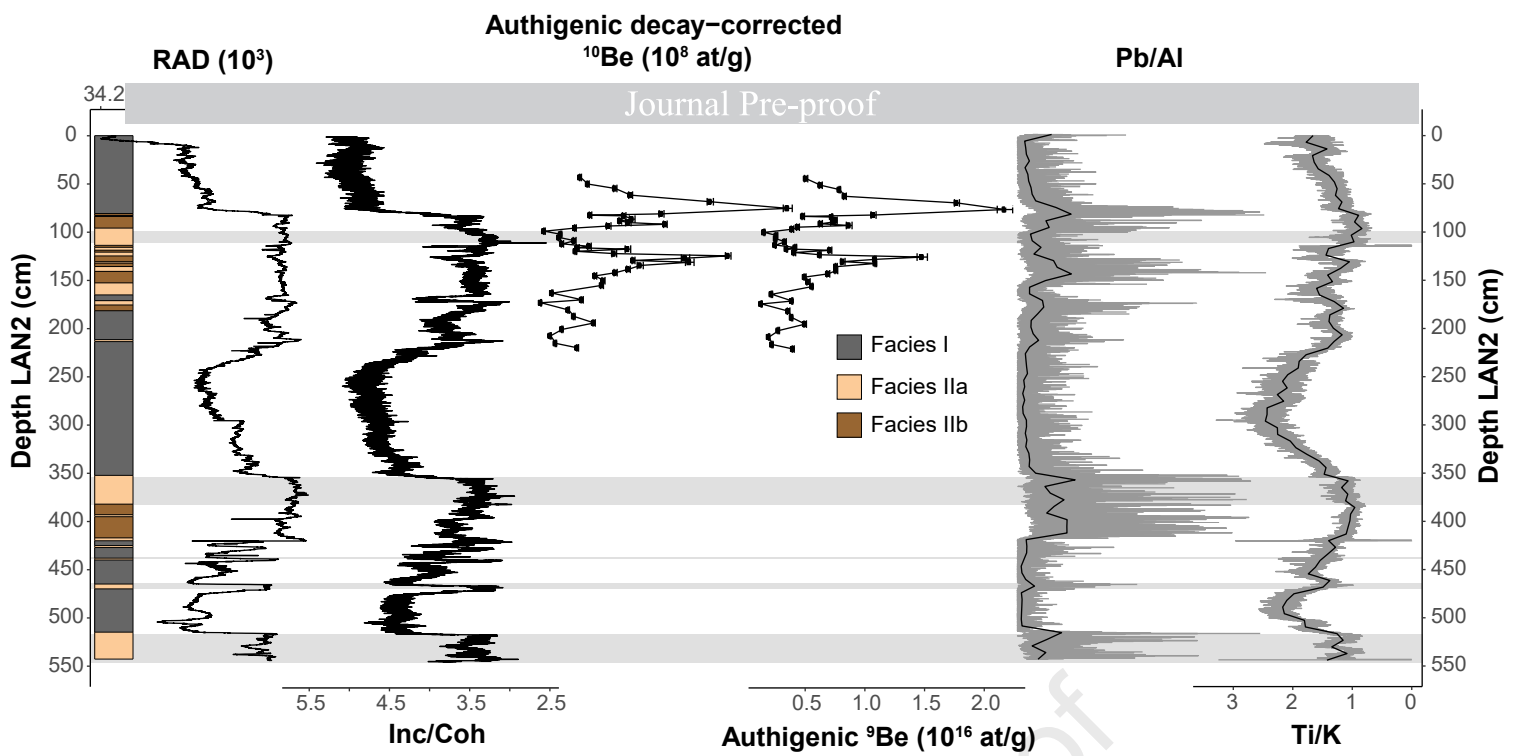
C) Reference depth scale

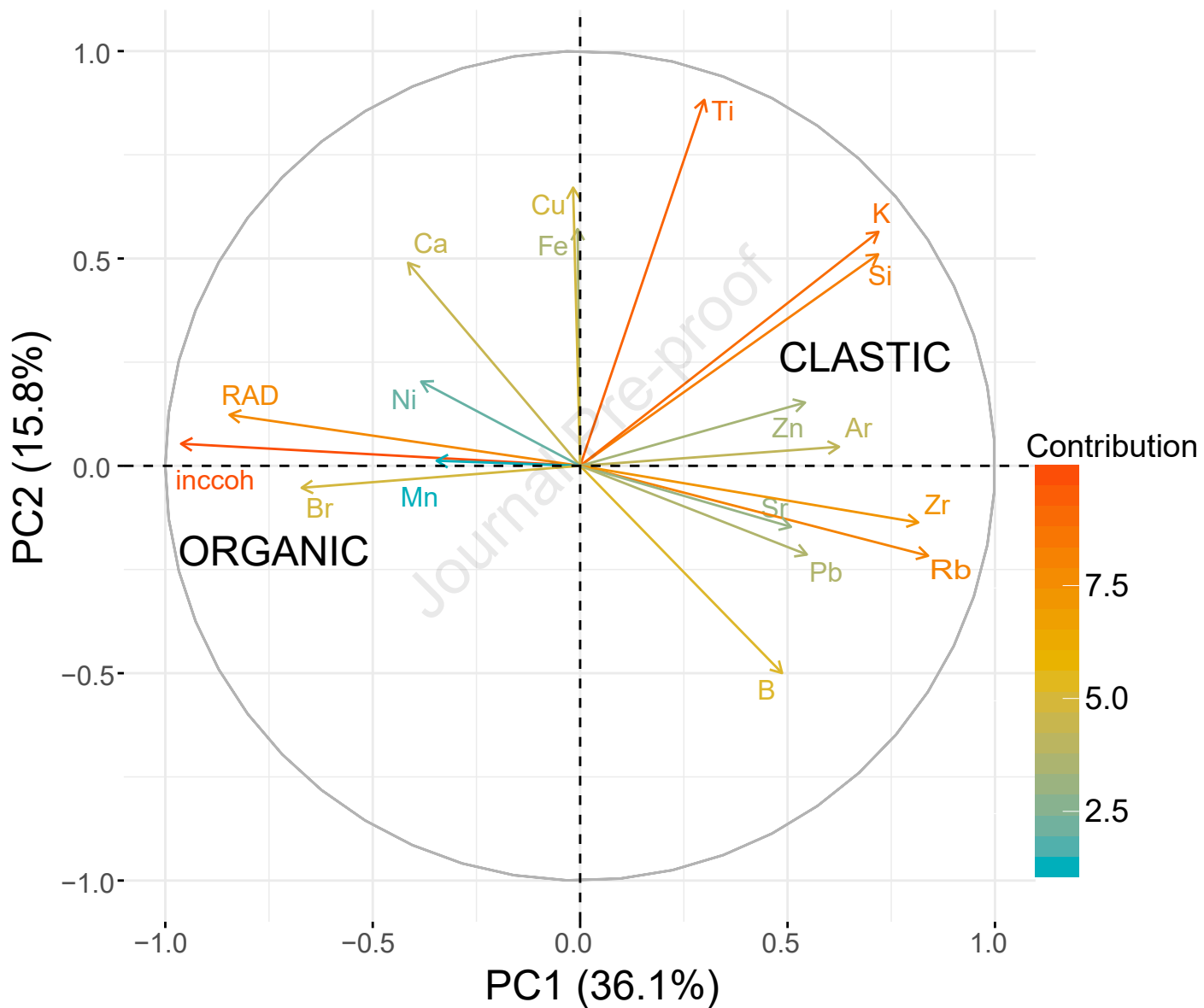


B) Inc/coh correlation



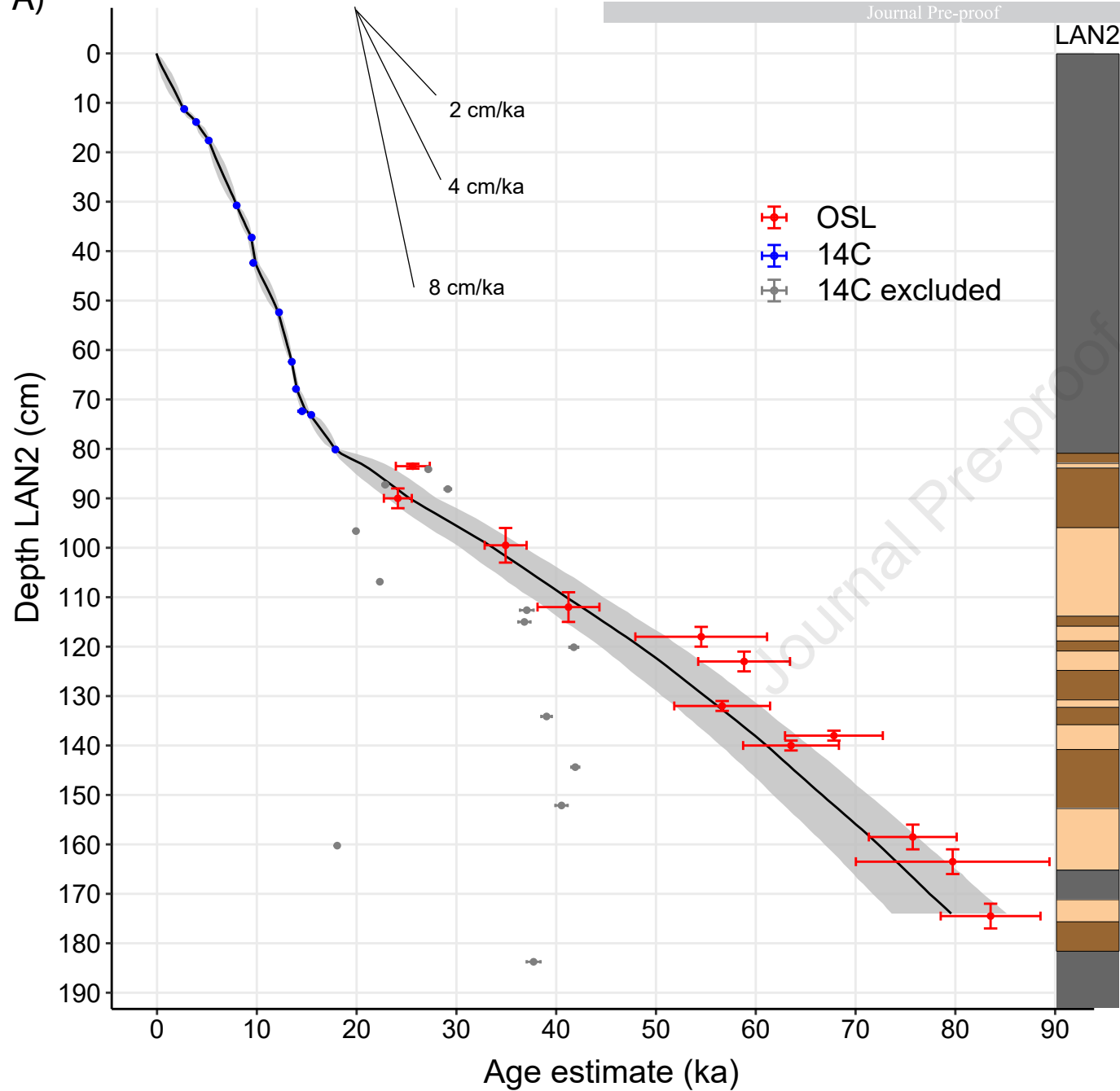




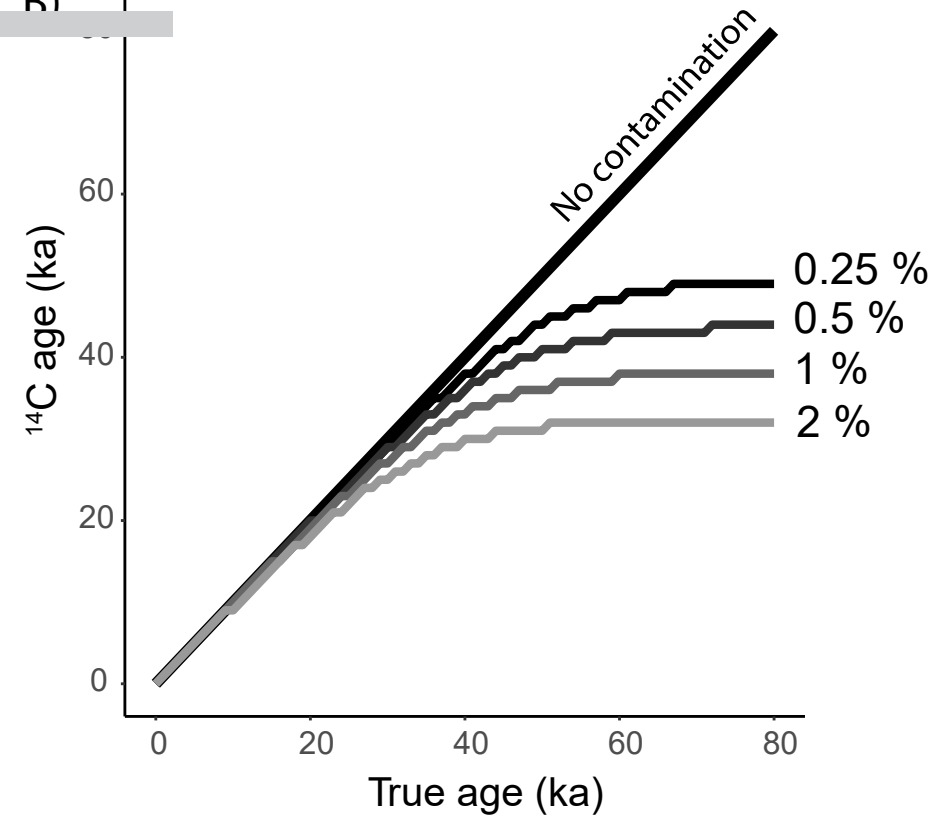




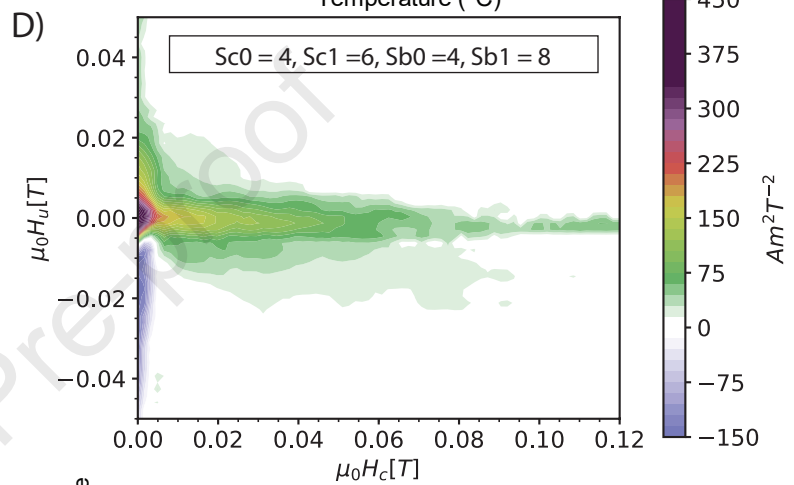
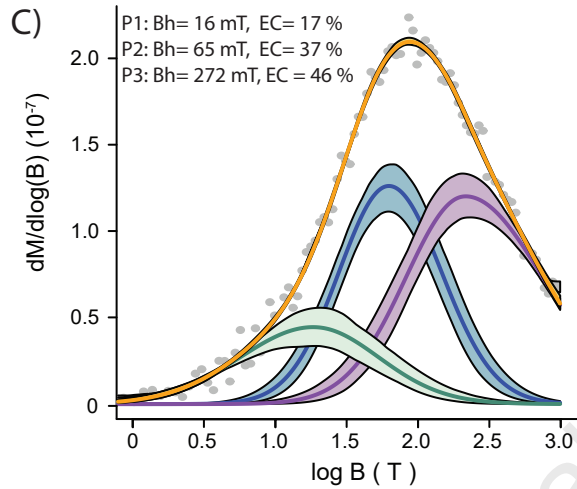
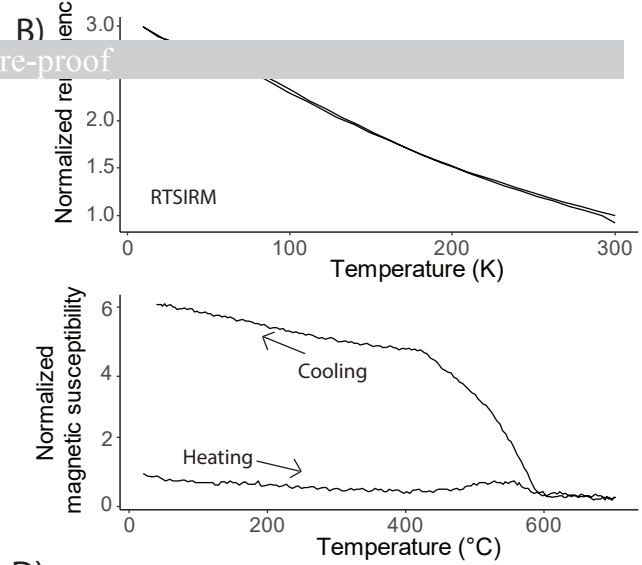
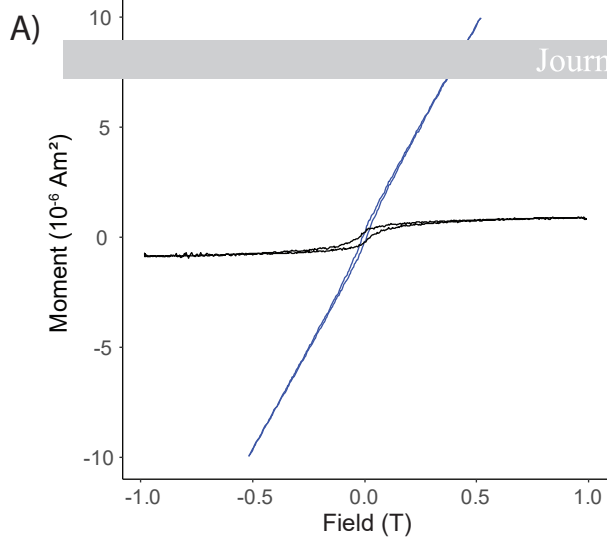
A)



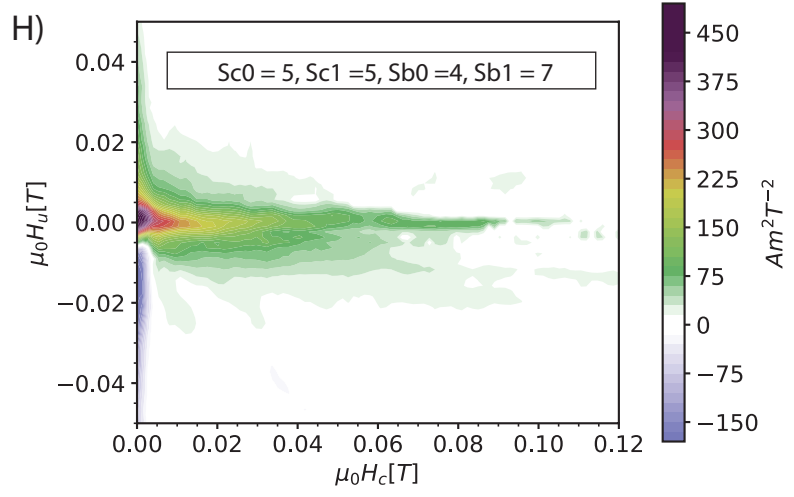
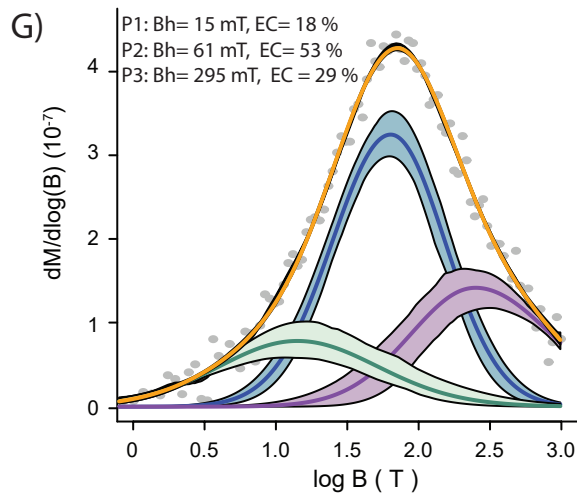
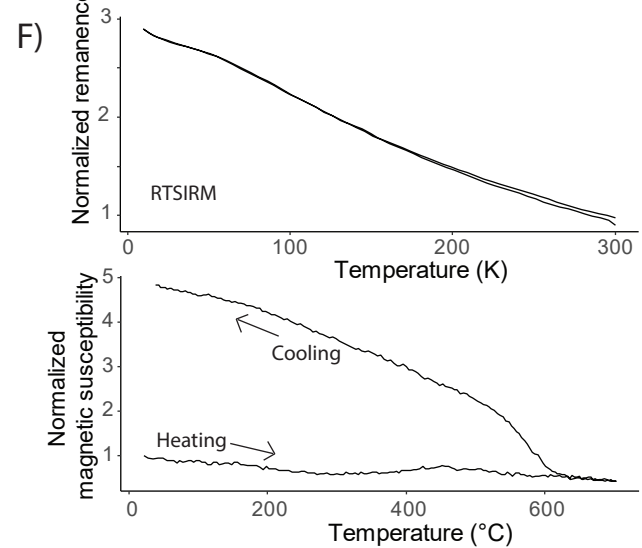
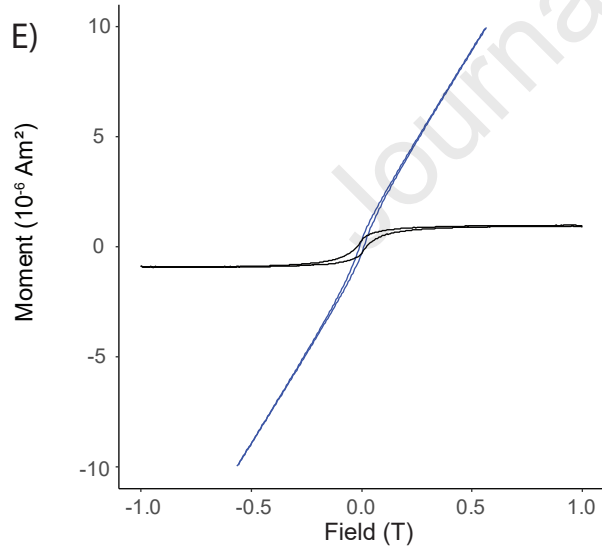
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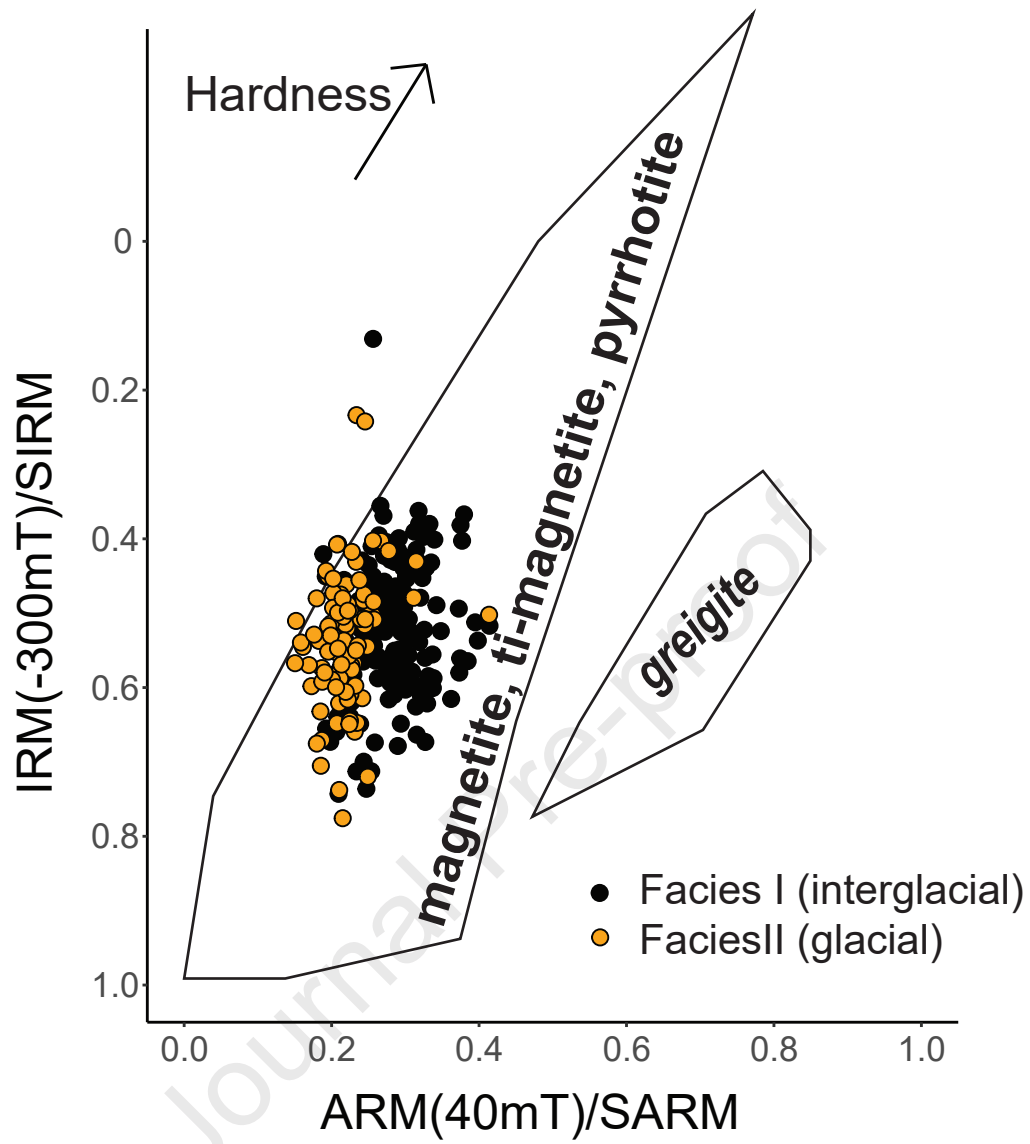


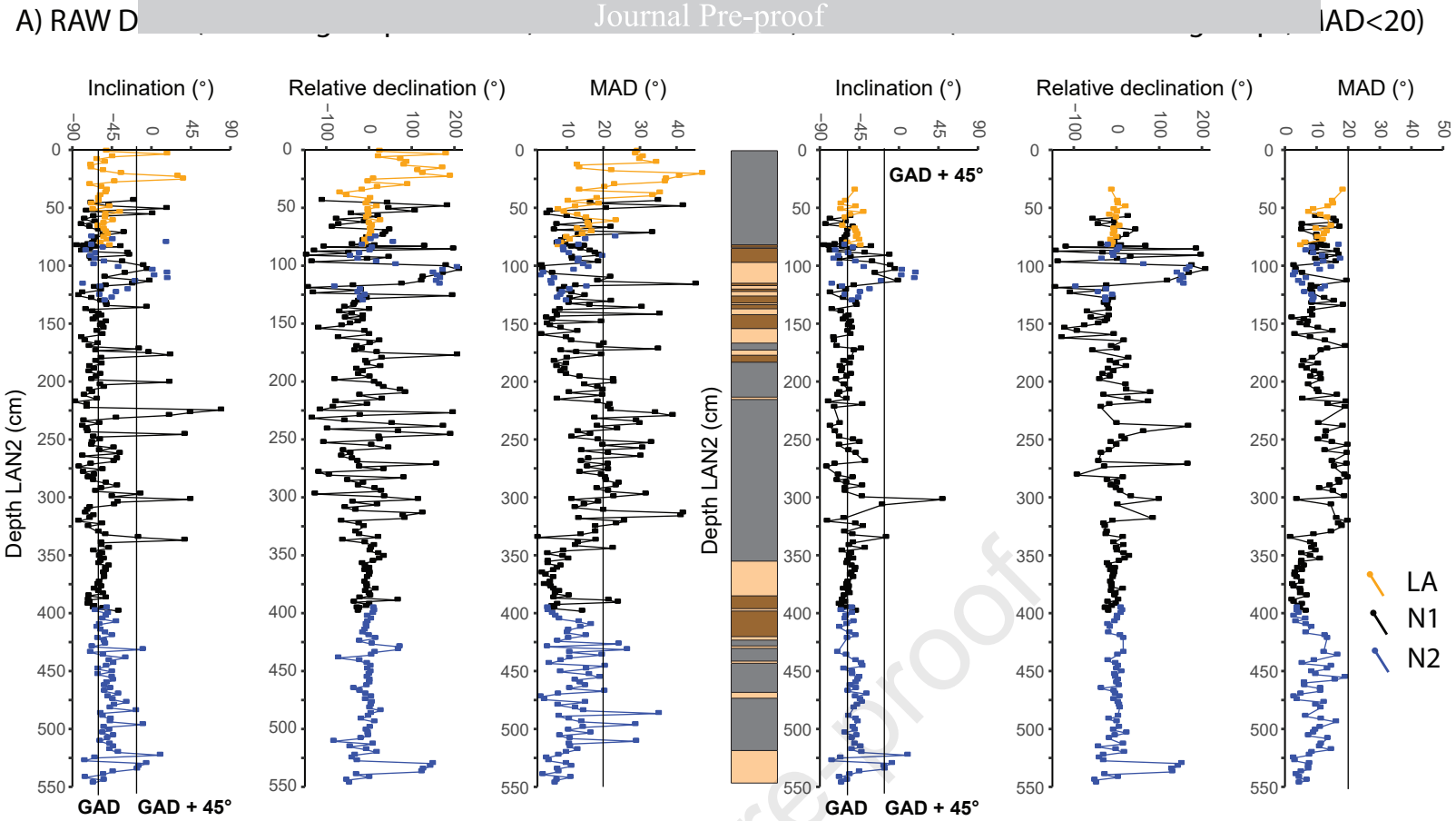
Facies I - Interglacial sediments



Facies II - Glacial sediments

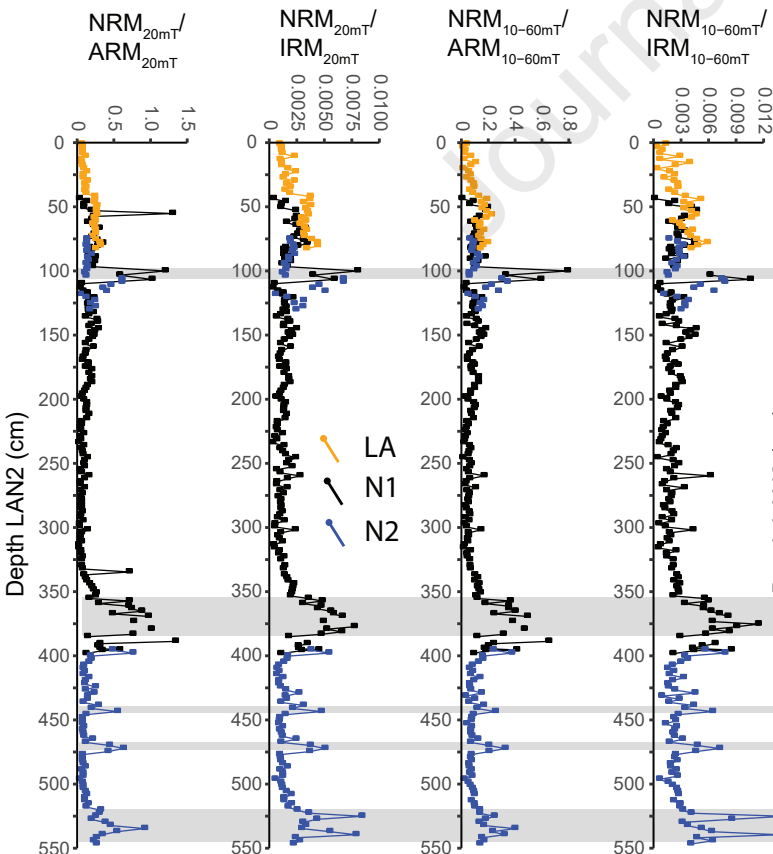




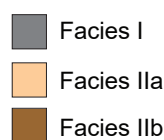
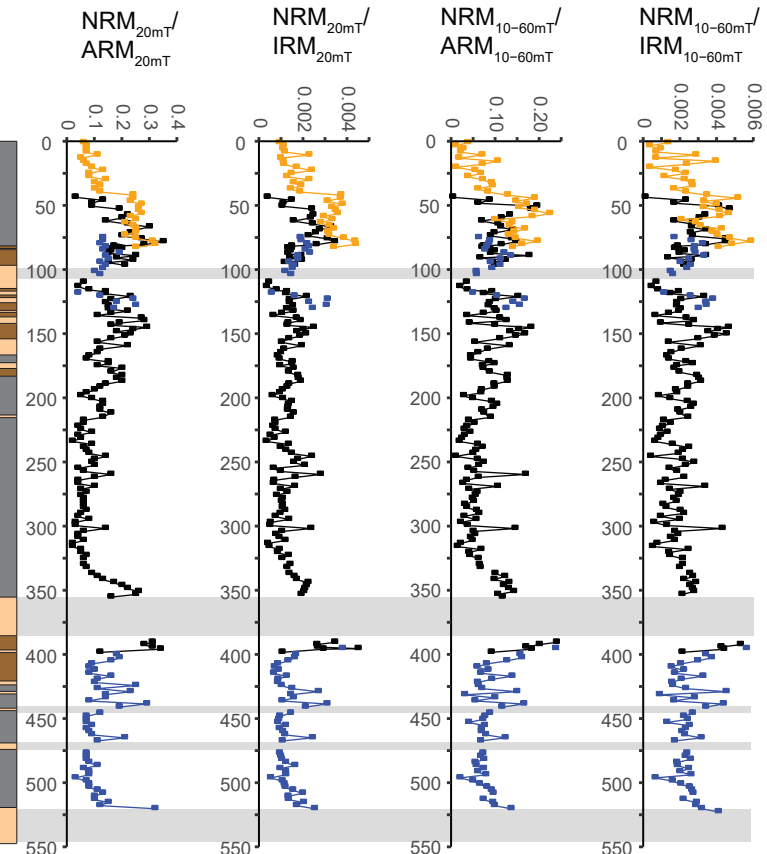


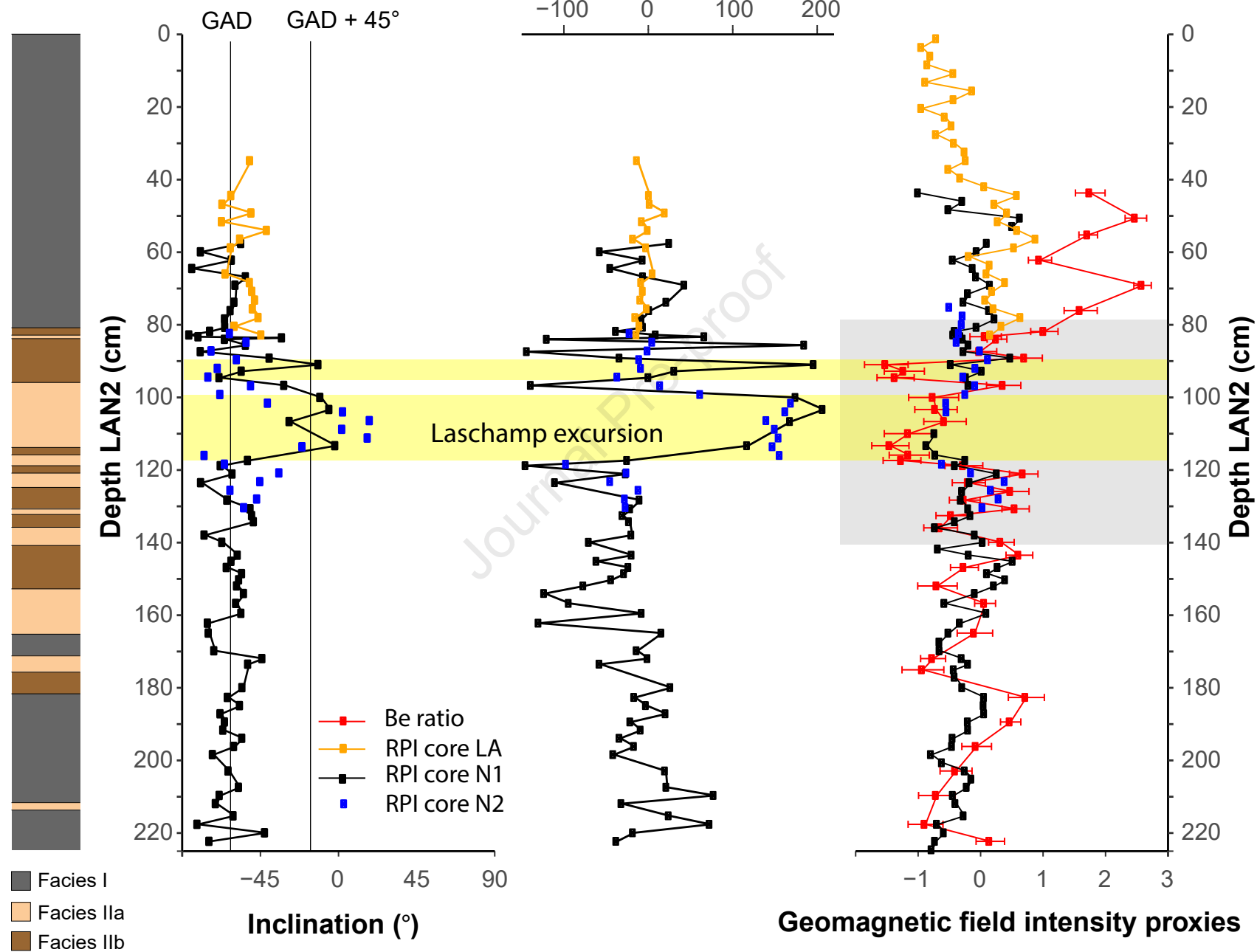
RELATIVE PALEOINTENSITY ESTIMATES

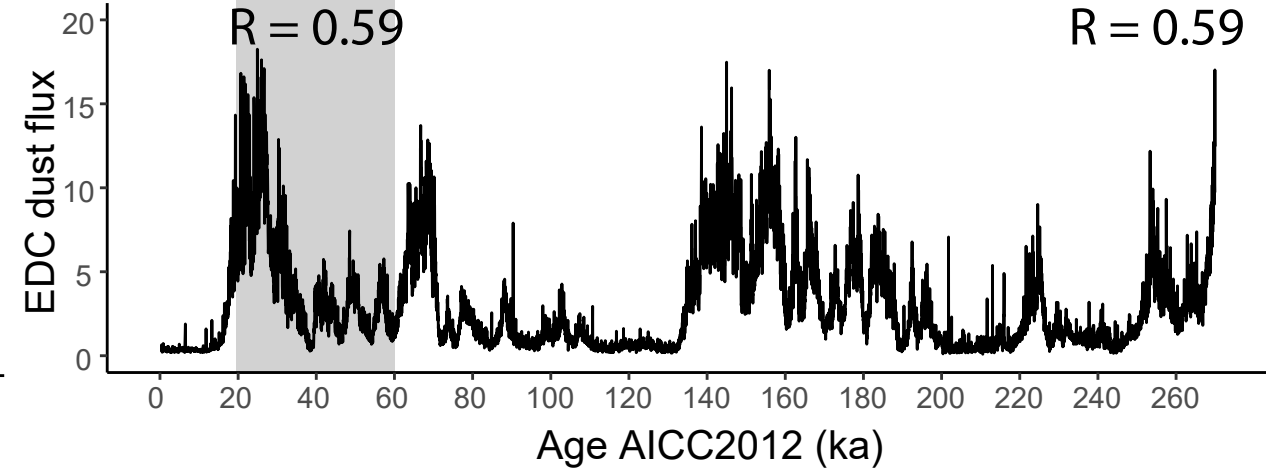
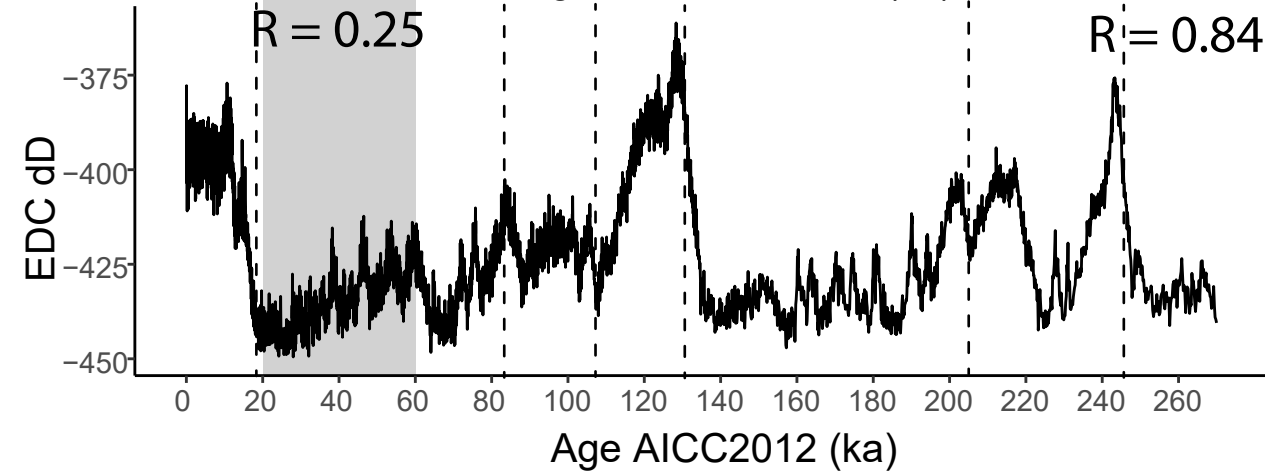
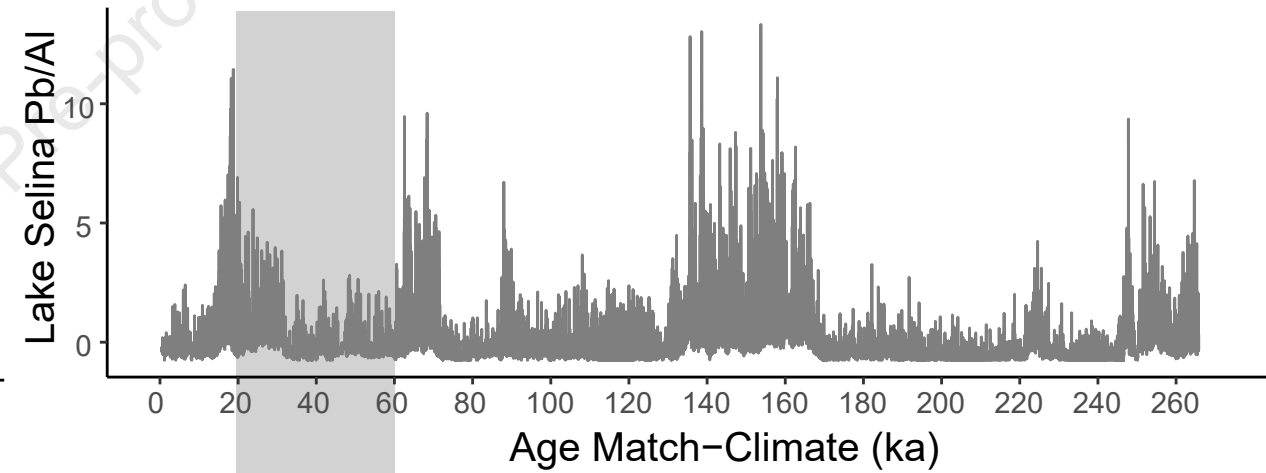
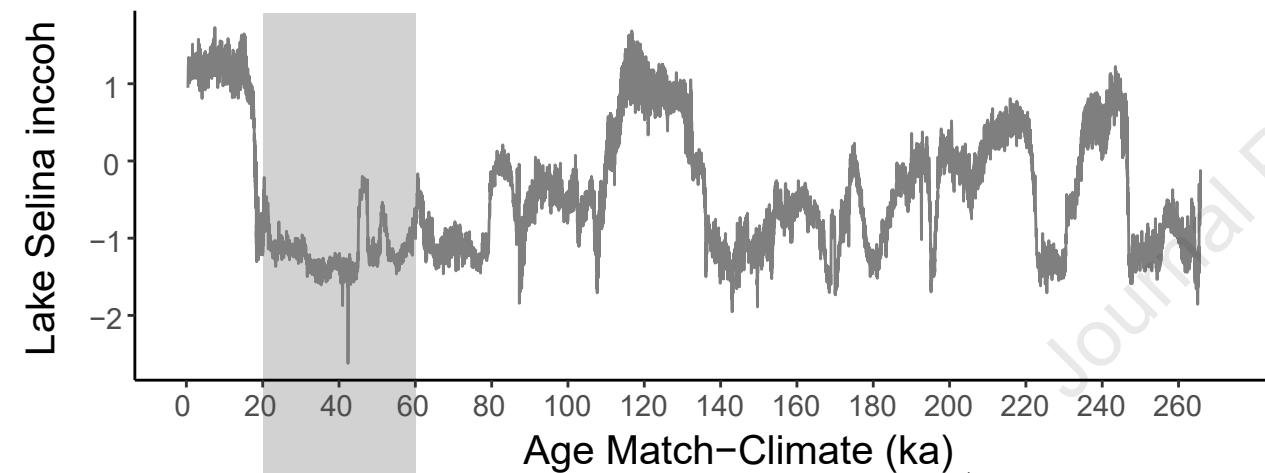
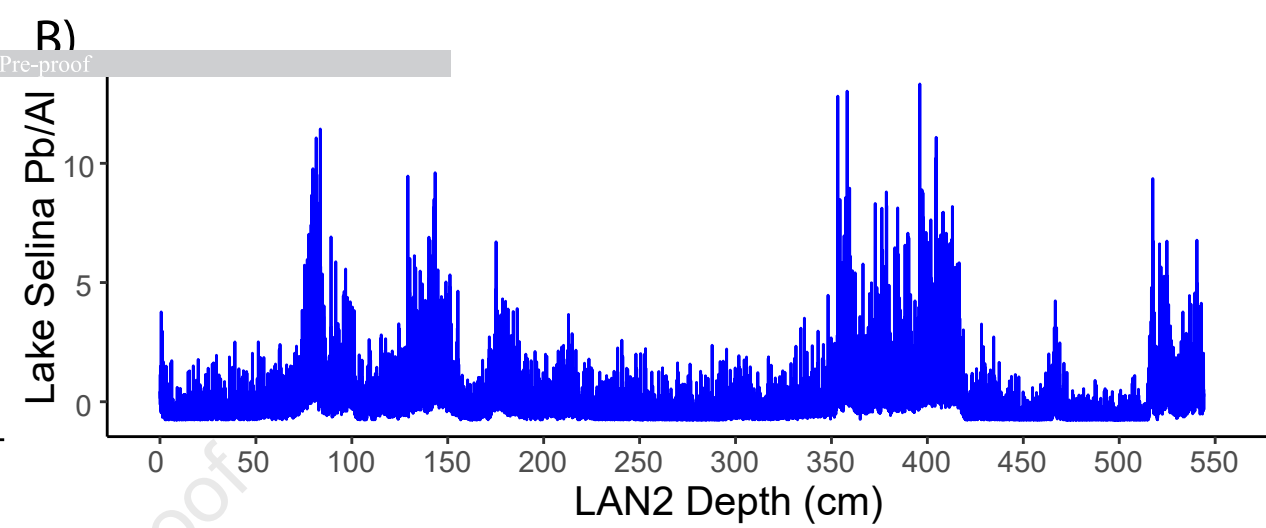
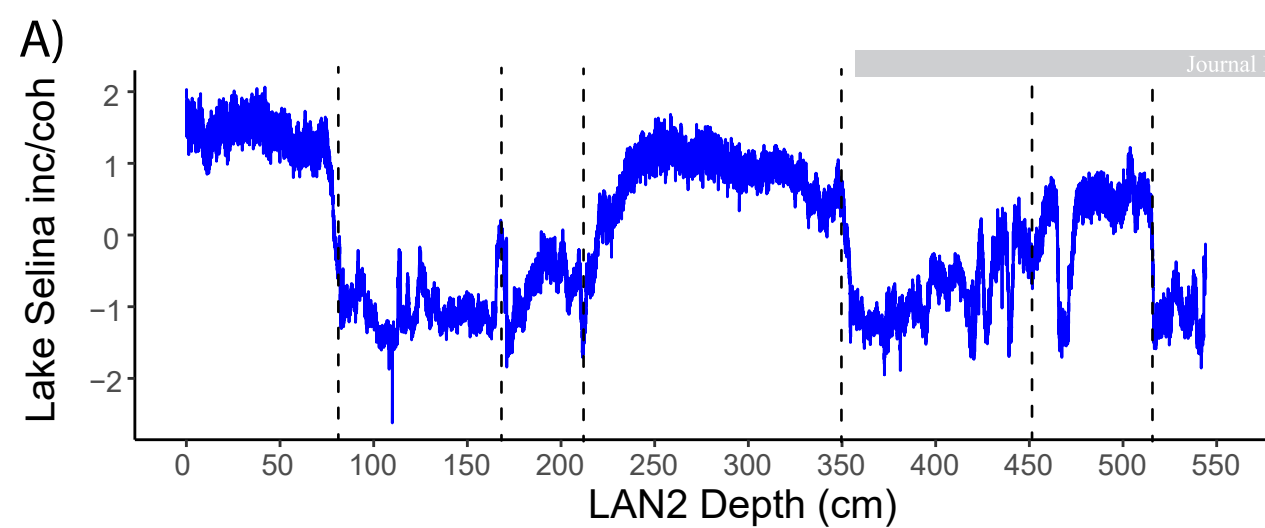
C) RAW DATA



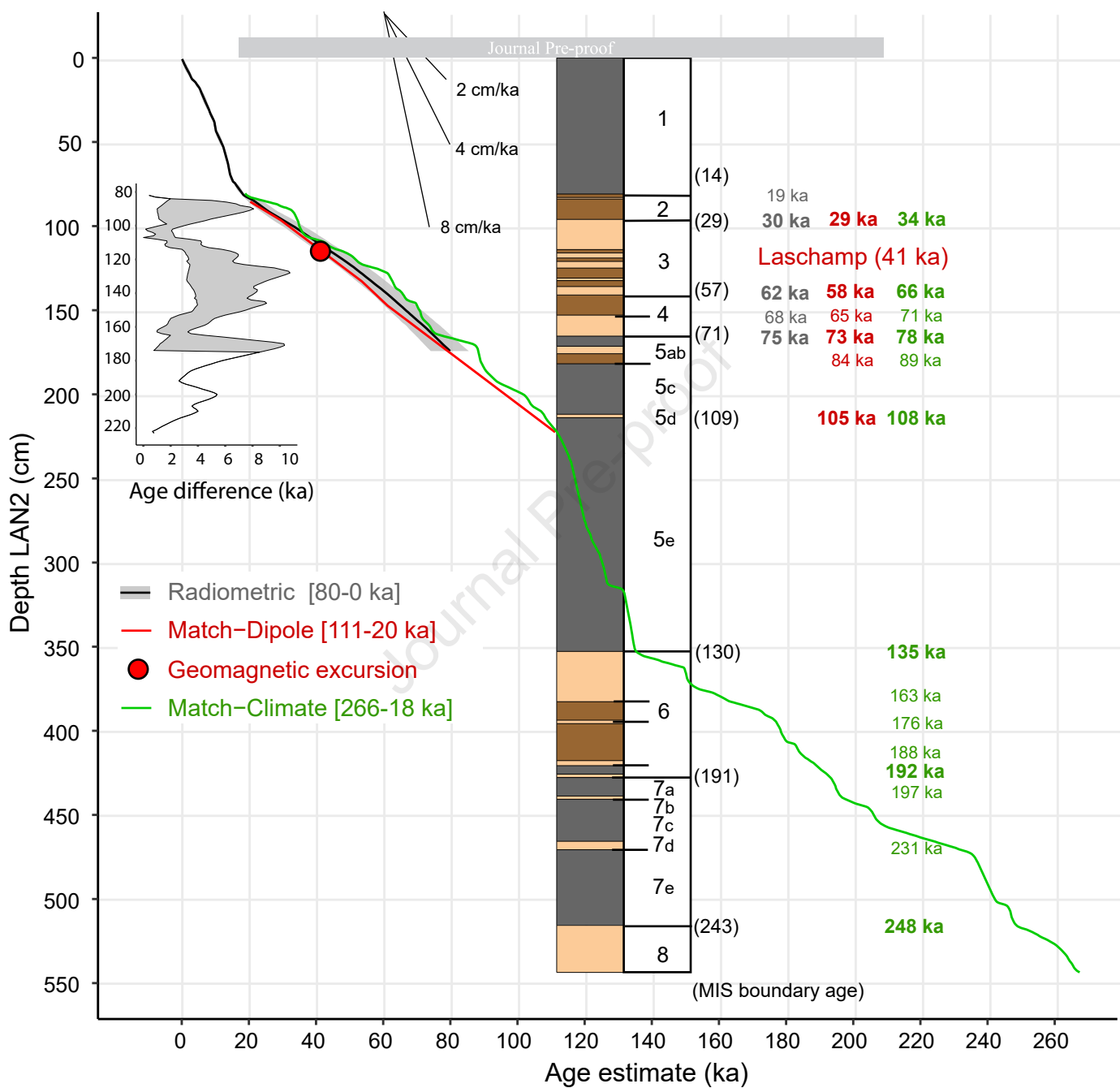
D) SELECTED

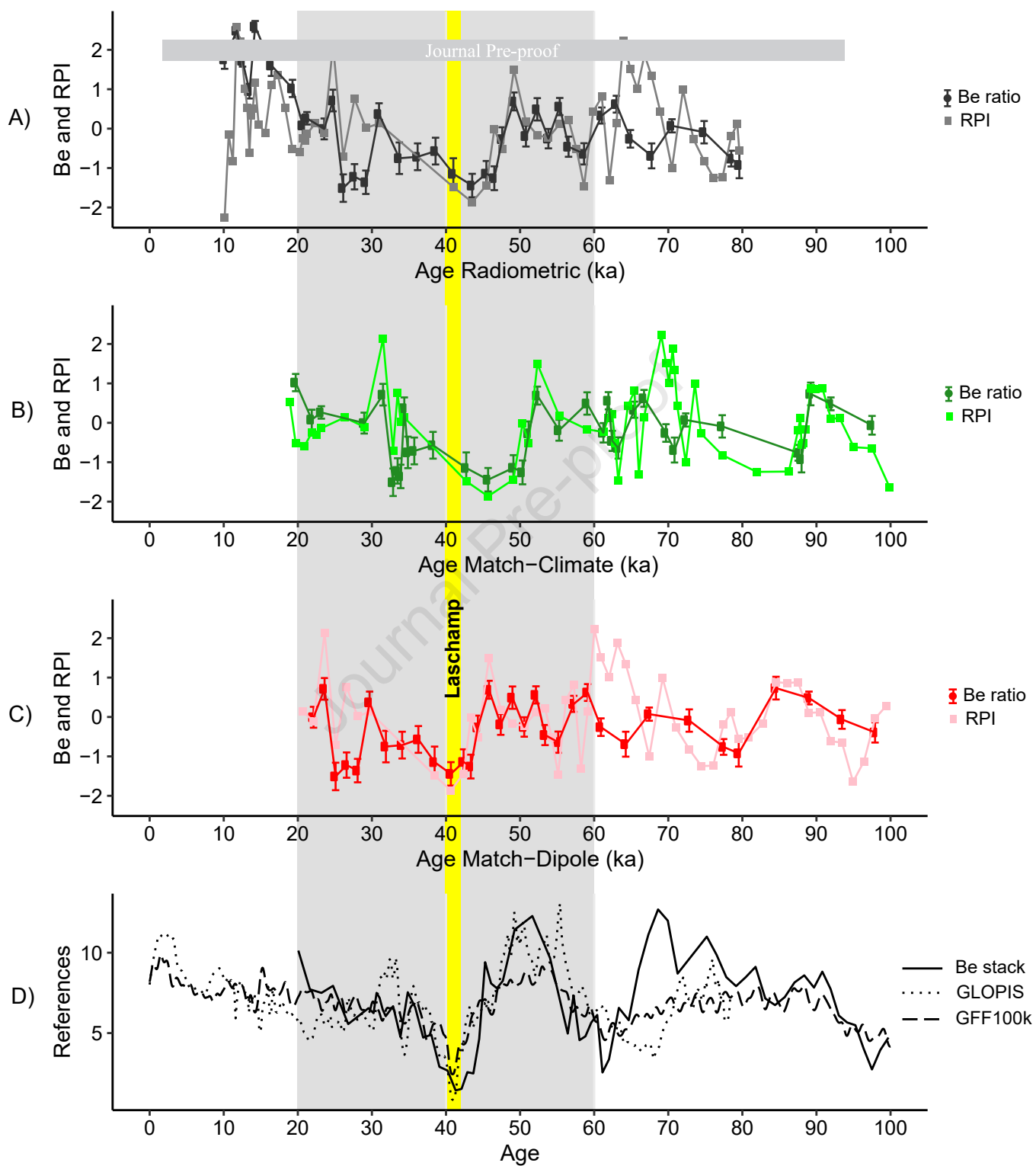












**Declaration of interests**

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:

Journal Pre-proof