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1 **Modeling the decomposition signal and correcting bulk organic data from a**
2 **peat deposit, a case study at low latitudes (Cameroon)**

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34 **Abstract**

35 Organic compounds are widely used for paleoclimatic and paleoenvironmental reconstructions. Bulk
36 organic proxies, however, are more complicated to interpret due to the multiple causes of variation
37 in climatic and environmental conditions and the degree of diagenetic alteration. As labile
38 compounds, rich in easily degradable function and generally richer in heteroatoms such as oxygen and
39 nitrogen, decompose, the remaining organic matter becomes progressively richer in refractory carbon
40 and its carbon content increases. Thus, in a peat deposit composed almost entirely of organic matter,
41 total organic carbon (TOC) is expected to increase with time and depth, which could mask the
42 paleoclimatic signal. We propose a simple model for peat sediments to remove the decomposition
43 signal based on a logarithmic function fitted with a partial dataset where decomposition is the main
44 parameter. The subtraction of the obtained logarithmic function to the raw data (i.e., measured data)
45 leads to "residual" data. We discuss the influence of different parameters (water table depth,
46 vegetation, microbial community) on the "residual" data and their possible link to paleoclimatic and
47 paleoenvironmental variations. This method is tested on bulk elemental and isotopic data obtained
48 from a new peatcore from the Adamawa Plateau (North-East Cameroon) covering nearly 10 ka cal BP.
49 Comparison with Rock-Eval® parameters highlights similar variations between the Hydrogen Index and
50 residual TOC variations, supporting the interpretation based on residual TOC. Our approach allows to
51 extract paleoenvironmental information from decomposition-prone bulk organic proxies and can be
52 generalized to peat deposits where decomposition plays a major role in controlling bulk data.

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55 **Key words: Peat deposit, Decomposition model, Bulk organic data**

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71 Introduction

72 Paleoclimatic proxies based on organic compounds rely on the assumption that these compounds are
73 resistant to diagenesis. Many organic compounds can record pre-depositional and syn-depositional
74 information such as climatic conditions (e.g., temperature, hydrology) or environmental conditions
75 (e.g., major vegetation types, pH, methanogenesis) (Sachse et al., 2012; Diefendorf and Freimuth,
76 2017; Inglis et al., 2019; Naafs et al., 2019). This signal can, however, be overprinted by post-
77 depositional processes such as decomposition (Gupta and Kawahata, 2006) or contamination by other
78 reservoirs (Makou et al., 2018).

79 Peat deposits have been successfully used to reconstruct past climatic variations at all latitudes using
80 diverse proxies such as pollen, plant macrofossils or biomarkers (e.g., Pancost et al., 2003; Booth et
81 al., 2004; Baker et al., 2014). Bulk analysis of organic matter such as Rock-Eval® or elemental analysis
82 are routine techniques in paleoenvironmental studies. These non-specific analyses give first insights
83 into the sources, nature and decomposition stage of organic matter (e.g., Talbot and Johannessen,
84 1992; Kuhry and Vitt, 1996; Baker et al., 2014). However, in bulk organic data, the paleoclimatic or
85 paleoenvironmental signal is often hidden by the signal of decomposition (Kuhry and Vitt, 1996;
86 Drollinger et al., 2019). Some proxies were specifically developed to reconstruct the peat
87 decomposition stage such as colorimetric determination (Chambers et al., 2011; Biester et al., 2014)
88 or thermal analyses (Upton et al., 2018; Cooper et al., 2019), while others such as the C/N ratio are
89 sometimes used as an indicator of organic matter decomposition (Kuhry and Vitt, 1996) or of organic
90 matter sources (Baker et al., 2014). In intertropical regions, peat deposits have a large paleoclimatic
91 potential especially when applying organic geochemistry techniques (e.g., Sukumar et al., 1993;
92 Bonnefille et al., 1995; Aucour et al., 1996, 1999; Behling and Safford, 2010). Until now, no study has
93 specifically focus on correcting bulk organic data in tropical peat.

94 Peatlands are highly impacted by organic matter decay (Clymo, 1984). Peats are usually divided in two
95 distinct layers: the upper layer (the acrotelm), above the mean water table depth and exposed to oxic
96 conditions, and the lower layer (the catotelm) situated under the mean water table depth and
97 dominated by anoxic conditions (Ingram, 1978; Clymo, 1984). Clymo (1984) and Clymo et al. (1998)
98 focused on models of peat mass accumulation and decay rules and indicated that about 90% of the
99 original biomass is lost in the upper oxic part of the peat within an indicative period of 10 to 100 years
100 but selective decay continues in the anoxic layer.

101 Peat organic matter is often mainly composed of different pools of organic compounds with different
102 sensitivities to decomposition (Dodla et al., 2012) and different $\delta^{13}\text{C}$ composition (Benner et al., 1987).
103 The labile organic compounds such as carbohydrates and proteins composed of easily hydrolyzable
104 functions rich in oxygen and nitrogen rapidly decompose after deposition while recalcitrant organic
105 compounds such as lignin and lipids, having long carbon chains and aromatic structures rich in carbon,
106 can be selectively preserved (Dodla et al., 2012; Wang et al., 2015). Thus, despite global mass loss and
107 carbon loss during organic matter decomposition (Clymo, 1984), the remaining organic matter, mainly
108 composed of recalcitrant compounds, becomes more and more concentrated in carbon with
109 increasing time and depth (Fig. 1). In the oxic layer, aerobic decomposers preferentially use ^{12}C ,
110 resulting in an increase in ^{13}C content (Ågren et al., 1996; Drollinger et al., 2019). The mineral fraction
111 in the peat is a key point to understand the variations in TOC. If the mineral component is very low,
112 the TOC will follow the trend of the organic matter and increase with depth and time (e.g., Upton et
113 al., 2018). In mineral-rich peat, or when mineral fraction supply varies, the increasing trend will no
114 longer be visible because the increasing proportion of the mineral fraction due to the decomposition
115 of the organic matter will mask the evolution of the carbon in the organic matter (e.g., Baker et al.,
116 2014; Njagi et al., 2021).

117 While the effects of early diagenesis on $\delta^{13}\text{C}$, $\delta^{18}\text{O}$ or Mg/Ca of foraminifera and carbonate sediments
118 have been studied for decades (Bernier, 2020), and several corrections of these proxies from
119 diagenetic overprint have been proposed (e.g., Schrag et al., 1995; Rosenthal and Lohmann, 2002),
120 only a few studies exist on the diagenetic effects of organic matter decomposition on bulk parameters
121 used for paleoenvironmental studies. The importance of TOC corrections for petroleum studies is well
122 known and several corrections specific to source rocks are available but they have not been applied
123 to paleoenvironmental reconstructions until recently (Hart and Hofmann, 2022). In a marine core from
124 the Timor Sea, Gupta and Kawahata (2006) observed a high impact of diagenesis on labile components
125 (amino acids) and proposed a correction of the measured total organic carbon (TOC) using a linear
126 equation. They suggest that corrected organic carbon more accurately reflects the original organic
127 carbon deposited on the sea floor and thus is more appropriate for primary productivity estimates.
128 In this paper, we focus on bulk organic data from a peatcore situated on the Adamawa Plateau in
129 Cameroon. Several maar lakes from the Adamawa Plateau have been studied for paleoclimatic and
130 paleoenvironmental reconstruction (e.g., Ngos III et al., 2003; Vincens et al., 2010; Nguetsop et al.,
131 2011; Lebamba et al., 2016; N’anga et al., 2021) but until now no peat deposit has been investigated.
132 We propose a method to deconvolute the diagenetic signal from the climatic signal in peat. We
133 develop a model that allows subtraction of a mean decomposition signal from bulk parameters such
134 as TOC, $\delta^{13}\text{C}$ and C/N. Residual data provide a robust means of interpreting palaeoclimatic and
135 palaeoenvironmental variations in parameters for which the decomposition signal is the main mode
136 of variability (in particular TOC).

137

138 **1. Study site and materials**

139 **1.1. Study site**

140 The NGaoundaba peat deposit (N7.13542° E13.69009°, 1175m above sea level) is located in North
141 Cameroon near the city of NGaoundéré on the Adamawa Plateau, occupying a volcanic crater situated
142 next to Lake Tabéré. Its current surface area covers ~18 000 m², and the maximum depth is still
143 unknown. According to Letouzey (1958), the peat deposit is situated in the soudano-guinean shrubs
144 savannas from the Adamawa Plateau. Nowadays, the peat is covered by sedges (Cyperaceae), that are
145 annually burned. Tropical peatlands of Africa from similar settings are largely sedge-dominated and
146 peat composition is a mix of sedges and woody remains from the surrounding forests (Dargie, 2015).

147 **1.2. Sampling**

148 A 6-meter-long core was collected on February 2019 using a Russian peat corer. The core is composed
149 of twelve consecutive 50-cm sections and consists of dark brown peat containing plant debris. The
150 color is consistently dark brown over the whole core. The NGaoundaba peat deposit is continuous and
151 essentially composed of peat without sandy or clayey zones. The core was subsampled at 2.5-cm
152 intervals corresponding to a resolution of around 40yr. Samples were freeze-dried prior to analysis.

153

154 **2. Analytical methods**

155 **2.1. Dating**

156 The chronological framework of NGaoundaba sequence is based on 30 radiocarbon dates carried out
157 at the LMC14 Laboratory (Saclay, France) following the procedure from Dumoulin et al. (2017) and
158 Moreau et al. (2013). Measurements were performed on bulk peat sediment. Data were compiled

159 using the R-software Bacon v2.5 (Blaauw and Christen, 2011) and calibrated using the SHCal20
160 calibration curve (Hogg et al., 2020) and postbomb calibration curve NH3 (Hua et al., 2013).

161

162 **2.2. Rock-Eval® thermal analysis**

163 Rock-Eval® analyses were performed on 69 samples using 10 mg of ground peat and a Rock-Eval® 6
164 (Vinci Technologies, France). The Rock-Eval® thermal analysis is based on continuous measurement of
165 effluents, i.e., hydrocarbons (HC), carbon monoxide (CO), and carbon dioxide (CO₂), released during
166 the thermal cracking of organic compounds under pyrolytic conditions (from 200 to 650°C at 25°C min⁻¹
167 in an inert atmosphere), followed by the combustion of residual organic compounds (from 200 to
168 850°C in an oxidative atmosphere). The resulting thermograms were used to calculate the standard
169 parameters by integrating the amounts of HC, CO, and CO₂ within the defined temperature limits
170 (Lafargue et al., 1998; Behar et al., 2001). The Total Organic Carbon (TOC parameter) is calculated as
171 the sum of all C effluents (i.e., HC, CO, and CO₂) under a temperature limit (Lafargue et al., 1998) and
172 usually represents the amount of organic C. When inorganic carbon forms are negligible, the percent
173 C is calculated as the sum of organic carbon (TOC) and MinC parameter. The Hydrogen Index (HI)
174 corresponds to the amount of HC released during pyrolysis relative to percent C. In organic samples
175 with the same source of organic inputs, a decrease in HI is usually related to the decomposition of H-
176 rich organic compounds (dehydrogenation), leading to an enrichment of aromatic compounds during
177 diagenesis in lake deposits (Meyers and Lallier-Vergès, 1999) or pedogenesis in soils (Disnar et al.,
178 2003; Barré et al., 2016). In addition to the above standard parameters, Disnar et al. (2003) proposed
179 using the shape of thermograms to obtain additional information about OM quality. In the present
180 study, we used the I index to quantify the preservation of the thermolabile C pool (Sebag et al., 2016).
181 This index is derived from the integrated S2 areas in specific temperature ranges (200--400 °C, 400--
182 460 °C, and > 460 °C), and is usually interpreted in terms of specific thresholds of the thermal stability
183 of organic compounds, separating the thermolabile, thermoresistant, and thermorefractory C pools
184 (Disnar et al., 2003; Sebag et al., 2006; Saenger et al., 2013). The I-index has recently been used to
185 measure the degree of decomposition of a tropical peat to identify an event of increased
186 decomposition of organic matter after deposit (Garcin et al., 2022).

187 While it is well calibrated for sedimentary organic matter (i.e., kerogens), Rock-Eval® percent C needs
188 to be corrected in soils and litters (Disnar et al., 2003). The correction factor calculation depends on
189 an estimation of the decomposition state of the sediment. As decomposition state is a focus of this
190 study, we chose to use elemental and stable isotope analyses rather than Rock-Eval® analyses for data
191 processing to avoid the use of any pre-processing assumption of decomposition state.

192

193 **2.3. Elemental and stable isotope analyses**

194 A total of 66 samples was ground and homogenized, then measured for TOC, total nitrogen, and δ¹³C
195 composition of TOC. Between 2 and 4 mg of non-acidified sediment were combusted in tin capsules
196 using an elemental analyser (Vario Micro Cube Elementar) coupled to a thermal conductivity detector
197 and an isotope ratio mass spectrometer IRMS (Elementar Vision). The carrier gas was helium.
198 Samples were first passed through a combustion furnace at 950°C under an excess of oxygen with
199 copper oxide as an oxidation catalyst, producing CO₂, N_xO_x, and H₂O. Then, a reduction furnace at
200 550°C reduced N_xO_x to N₂ and removed excess O₂ using reduced copper. H₂O was trapped and
201 removed using a phosphorous pentoxide chemical trap. N₂ and CO₂ were separated using a purge and
202 trap desorption column. The C/N ratio was calculated as the ratio between TOC and wt%N.

203 A working standard (IVA sediment) was analysed every 10 samples and used to normalize the mass
 204 spectrometer signals. Triplicates were analysed every 3 samples and standard deviations were
 205 calculated. The absence of carbonates in the samples was confirmed by acid-treating 6 samples (1
 206 every meter) using vapour of 12N hydrochloric acid for 72 hours and then performing a comparative
 207 analysis.

208

209 3. Data processing

210 3.1. General information

211 The method proposed in this study is divided into 4 steps (Fig. 2) and described in detail below.
 212 Kinetics of organic compound degradation follow a simple first order equation (Westrich and Berner,
 213 1984; Grossi et al., 2003) with different reactivity depending on the compounds. The increase in the
 214 proportion of recalcitrant compounds leads to a progressive increase in carbon concentration despite
 215 the total mass loss (Fig. 1). Using a different approach, Clymo (1984) and Clymo et al. (1998) developed
 216 models of peat accumulation. Based on equation 1, α denotes the manner of decay and constant,
 217 linear and quadratic rules were tested by Clymo et al. (1998). Using a linear decay rule α_L^* , the plot of
 218 peat growth (cumulative carbon M_t) follows the logarithmic equation (2) with p^* the rate of addition
 219 of dry mass at the top of the peat deposit (Clymo et al., 1998).

$$220 \quad \frac{d\mu}{dt} = -\alpha \times \mu \text{ with } \mu = \frac{m_t}{m_0} \quad (1)$$

$$227 \quad M_t = \frac{p^*}{\alpha_L^*} \times \ln(1 + \alpha_L^* \times t) \quad (2)$$

221 Based on these observations and on the logarithmic trend followed by TOC with age and depth in our
 222 dataset and in several others (e.g., Upton et al., 2018; Wei et al., 2012; Garcin et al., 2022), we chose
 223 to use a natural logarithmic function (noted "ln") in two possible forms presented in equations 3 and
 224 4 using age as the variable. The parameter c can be linked to the initial C content in freshly deposited
 225 organic matter, while a and b are related to the kinetic constants and the proportion of labile versus
 226 recalcitrant organic matter.

$$228 \quad f(x) = a \times \ln(b \times t) + c \quad \forall t \in \mathfrak{R}^{+*} \quad (3)$$

$$236 \quad f(x) = a \times \ln(b \times t + d) + c \quad \forall t \in]-d/b, +\infty[. \quad (4)$$

229 Data processing was performed using the Python software. The different parameters a, b, c and d of
 230 each function are optimized using the curve.fit() function of the SciPy Python library. The difference
 231 between the two forms is slight; equation 4 allows an offset on the t axis of the data and the domains
 232 of the function defined as the set of t-values accepted by the function are slightly different for the two
 233 functions. By definition, the ln function is only defined for strictly positive values, equation 3 is defined
 234 for strictly positive values \mathfrak{R}^{+*} , while equation 4 is defined for t above the ratio of d and b parameters
 235 $-d/b$ (i.e. for $t \in]-d/b; +\infty[$).

237 The choice of the function for each parameter was determined by comparing the determination
 238 coefficient r^2 between the raw data and the curve-fit for each form. We called "raw data" the initial
 239 data prior to any data treatment.

240 3.2. Data normalization

241 As logarithmic functions can only be used for positive and non-zero values, raw data were normalized
 242 (scaling of all values in the range of 0 and 1) prior to any analysis (Fig. 2 - Step1). Practically, all values

243 were normalized using the MinMaxScaler transforms from Python library sklearn.preprocessing based
244 on the following equation 5:

$$247 \quad x_{scaled} = \frac{x - x_{min}}{x_{max} - x_{min}} \quad (5)$$

245 We add 0.001 to all normalized ages to avoid having a zero t -value that cannot be used in a
246 logarithmic function.

248

249 **3.3. Total curve fitting and derivative calculation**

250 The method is based on two successive curve fittings, which are presented in the next two subsections.
251 The objective is to find a curve fit based only on a part of the initial dataset where the decomposition
252 is predominant.

253 A first option would use the upper oxic layer thickness (acrotelm) as most of the decomposition occurs
254 in this layer (Clymo, 1984). However, no measurement of this parameter is available for the
255 NGaoundaba peat deposit. Based on the mean water table depth, upper oxic layer thicknesses
256 between 5 and 35 cm were estimated for several temperate peat deposits (Kuhry and Vitt, 1996; Evans
257 et al., 1999; Weijers et al., 2004, 2006; Liu et al., 2010; Drollinger et al., 2019). In a tropical peat swamp
258 forest, Jauhiainen et al. (2005) reported a water table varying between 80 cm under and 20 cm above
259 the peat sediment surface throughout the year with maximum occurrences of values between 10 and
260 30 cm under the peat sediment surface. The upper oxic layer thickness is usually on the order of tens
261 of centimeters. In the case of the NGaoundaba dataset, neither measurement of the upper layer
262 thickness nor estimation based on literature can be used for the modeling as it would involve only a
263 few points and would lead to poorly constrained results.

264 To determine the age where the decomposition is the main parameter controlling bulk organic data,
265 we first fit the logarithmic function f using all x -values, then we calculate the derivative of this function
266 f' and determine the range of x -values where $f'(x) > 1$, thus finding the inflection point of the curve
267 (Fig.2 - Step 2 and Fig.3 - Light colored curves). The value $f'(x) > 1$ was fixed arbitrarily. This choice avoids
268 using the whole dataset for the modeling and avoids purely circular reasoning.

269

270 **3.4. Partial curve fitting**

271 A new function g is then fitted based only on the reduced range of x -values (Fig. 2 - Step 3 and Fig. 3).
272 In other words, only a part of the original data is selected for the final curve fitting g . For each
273 parameter, the selected x -values are indicated in dark color on Figure 3. A total of 13, 17 and 11 x -
274 values were selected for TOC, $\delta^{13}\text{C}_{bulk}$ and C/N, respectively.

275

276 **3.5. Curve fit subtraction**

277 The last step is the subtraction of the function g (i.e., the modeled data) to the raw data. The resulting
278 data are named the residual data. Thus, positive residual values indicate that the raw data values are
279 superior to those of the modeled data, indicating an excess of this parameter compared to the
280 modeled data. Residual data are smoothed (Fig. 2 - Step 4) using a moving average method.

281

282 4. Results

283 4.1. Age model

284 Calibrated ages range from -39 a cal BP at 10 cm depth to 9642 a cal BP at the bottom (590 cm) (Fig.
285 4, Tab. 1). The age-depth model presents a quite uniform sediment accumulation rate except around
286 150 cm-depth (2.5 ka cal BP) and 200 cm-depth (4.0 ka cal BP). These two periods are not situated at
287 the limit between two core sections, as indicated on figure 4.

288

289 4.2. Elemental raw data

290 TOC increases with increasing depth and age ranging from 26.76 to 53.5 % for elemental and stable
291 isotope analyses (Fig. 5(A)).

292 The C/N ratio roughly increases with depth and age. Until 4.5 ka cal BP, values increase nearly linearly
293 (Fig. 5(C)). From 4.5 to around 5 ka cal BP, the C/N ratio decreases and then stabilizes until 7.1 ka cal
294 BP. The bottom of the core is marked by a general increase of the C/N ratio with small-scale variations.
295 The range of values of elemental raw data is consistent with data reported for other African peat
296 deposits (e.g., Aucour et al., 1999; Baker et al., 2014; Strobel et al., 2019).

297

298 4.3. Rock-Eval® I-index and HI

299 The I-index ranges from 0.06 to 0.37. The I-index variations can be divided into two periods (Fig. 5(D)).
300 The first period from the present to 6.4 ka cal BP is characterized by a regular decrease in I-index from
301 0.37 to 0.08. The second period from 6.4 ka cal BP to the bottom of the core presents more variations,
302 but with a narrower amplitude between 0.2 and 0.06 (Fig. 5(D)).

303 The HI ranges from 195 to 427 (Fig. 6(G)). The HI is between 250 and 300 from 5.8 ka cal BP until
304 present except between 4 and 3 ka cal BP where values are around 225. The highest HI values are
305 found from 7.9 to 5.8 ka cal BP and at the bottom of the core. From 9.4 to 7.9 ka cal BP, HI values are
306 under 300 (Fig. 6(G)).

307

308 4.4. Isotopic raw data

309 The $\delta^{13}\text{C}_{\text{bulk}}$ presents a general decrease with increasing depth and age ranging from -15.6 to
310 -26.3‰ (Fig. 5(B)). The $\delta^{13}\text{C}_{\text{bulk}}$ decreases from the top of the core to 3 ka cal BP from values around -
311 16‰ to values around -22‰. Between 3 and 4 ka cal BP, $\delta^{13}\text{C}_{\text{bulk}}$ is close to -25‰. The $\delta^{13}\text{C}_{\text{bulk}}$ increases
312 between 4 ka cal BP and 5.8 ka with most values between -23 and -20.4‰ compared to values around
313 -25 between 5.8 and 9.4 ka cal BP. The $\delta^{13}\text{C}_{\text{bulk}}$ also increases from 9.4 ka to the bottom of the core
314 ranging from -21.4 to -20.4 ‰ during this period.

315

316 4.5. Processed residual data

317 Processed residual data range from -0.126 to 0.181, -0.189 to 0.815 and -0.090 to 0.782 for TOC,
318 $\delta^{13}\text{C}_{\text{bulk}}$ and C/N respectively. Both residual TOC, residual $\delta^{13}\text{C}_{\text{bulk}}$ and residual C/N are mostly positive
319 (Fig. 6(C), (D) and (E) respectively).

320 Residual TOC values are decreasing from the bottom of the core until 8 ka cal BP with values close
321 zero between 9 and 8 ka cal BP (Fig. 6(D)). From 8 to 4 ka cal BP, residual TOC are between 0.1 and 0.2
322 except around 5.5 ka cal BP where values are slightly lower. From 2.5 ka cal BP until present, TOC
323 values present rapid variations between -0.13 and 0.12.

324 The residual $\delta^{13}\text{C}_{\text{bulk}}$ presents a general decreasing trend and positive values from the bottom of the
325 core to around 4 ka cal BP (Fig. 6(E)). The period of less depleted values from 5.8 to around 4 ka cal BP
326 observed in the raw $\delta^{13}\text{C}_{\text{bulk}}$ data (Fig. 5(B)) is still visible in the residual $\delta^{13}\text{C}_{\text{bulk}}$ with positive residual
327 data ranging from 0.2 to 0.5.

328 The residual C/N presents similar variations as the raw C/N data with a globally increasing trend with
329 depth and age (Fig. 6(F)). Values are close to zero until 2.5 ka cal BP and between 5 and 7.5 ka cal BP.
330 The highest values occur between 8.5 and 9 ka cal BP.

331

332 **5. Discussion**

333 **5.1. Interpreting the residual data**

334 The log-generated data represent the part of the signal due to decomposition. Residual values close
335 to 0 indicate a good correspondence between the original data and the log-generated data (Fig. 6).
336 Higher values (both positively and negatively) indicate a greater difference between raw data and the
337 model and suggest either a change in the decomposition compared to the model or the influence of
338 other parameters that will be discussed in the next paragraphs.

339 In a simple approach, the carbon accumulation in peatlands can be described as a balance between
340 vegetation inputs and decomposition (Frolking et al., 2001; Kurnianto et al., 2015). Even if microbial
341 organic matter is generally not considered in the mass balance, microbial activity is a key point of
342 decomposition processes (Moore, 1989). The TOC excess is thus linked to either reduced
343 decomposition or an increase in deposition compared to the mean decomposition signal (and vice-
344 versa). Changes in paleo -climate and -environment exert a strong influence on these parameters.

345

346 **5.2. Influence of water table depth**

347 An increase in wetness leads to changes in peatland surface structure and an increase in carbon
348 sequestration by increasing litter production and decreasing litter decay losses (Belyea and Malmer,
349 2004). These changes are driven by an increase in acrotelm thickness due to a rise of the water table
350 (Belyea and Malmer, 2004). Moisture content, redox potential and pH control the microbial
351 community structure and can limit or enhance the presence of key decomposers (Girkin et al., 2020),
352 even in deep peat deposits (Kluber et al., 2020). Furthermore, Ise et al. (2008) demonstrate the
353 existence of a positive feedback between accumulation of soil organic carbon, water table depth and
354 decomposition rates, highlighting the high sensitivity of peatland to climate variations. Hodgkins et al.
355 (2018) demonstrated that the recalcitrance of tropical peat deposits is higher than that of boreal peat
356 deposits which allows them to persist despite warmer and sometimes drier conditions. Tropical peat
357 also accumulates because of the low decomposition of roots and woods under anoxic conditions when
358 the water table is high (Chimner and Ewel, 2005).

359 The positive values of residual TOC before ca 4 ka cal BP (Fig. 6(D)) could indicate a period of increased
360 wetness leading to a rise of the water table favouring preservation of the organic matter. This period
361 is consistent with the timing of the African Humid Period (AHP), beginning around 14.5 ka cal BP and

362 ending between 6 and 4 ka cal BP (among others, Demenocal et al., 2000; Kröpelin et al., 2008; Amaral
363 et al., 2013). The period between 9 and 8 ka cal BP that presents residual TOC close to zero could
364 correspond to a relatively drier intermission during the AHP.

365

366 **5.3. Influence of changes in vegetation**

367 Major shifts in vegetation also have a strong influence on both decomposition and deposition of
368 organic matter (Belyea and Malmer, 2004). Tropical peat deposits may be visually uniform despite
369 considerable changes in vegetation inputs (Page et al., 1999). The visual homogeneity of the
370 NGAoundaba peat deposit does not exclude changes in the vegetation. A reciprocal relation between
371 carbon storage and plant species diversity has been identified in a tropical peat deposit from Sumatra
372 (Shimamura and Momose, 2007). Changes in the sources of organic matter have an influence on the
373 initial proportion of thermo-labile vs thermo-stable organic matter (Wang et al., 2015) and, thus, on
374 the evolution of this organic matter during decomposition. Major changes in vegetation are reported
375 at the end of the African humid period (AHP) (Maley and Brenac, 1998; Vincens et al., 2010) and could
376 explain part of the observed variations in residual TOC by changing the type and amount of vegetation
377 inputs to the peatland.

378 To identify the period where the signal is most influenced by vegetation changes at the NGAoundaba
379 peat deposit, we use $\delta^{13}\text{C}_{\text{bulk}}$, both raw and residual (Fig. 5(B) and fig. 6(E)). Skrzypek et al. (2010)
380 demonstrated that the primary $\delta^{13}\text{C}$ signal of peat-forming plants is preserved and that decomposition
381 has only a secondary influence on the $\delta^{13}\text{C}_{\text{bulk}}$ signal. The latter has been directly used to describe
382 changes in C3/C4 vegetation (Baker et al., 2014).

383 The two main periods observed in the residual $\delta^{13}\text{C}_{\text{bulk}}$ data before and after around 4 ka cal BP are
384 roughly consistent with the timing of the AHP (Fig. 6(E)). The raw $\delta^{13}\text{C}_{\text{bulk}}$ values close to -25‰ until
385 5.8 ka cal BP are consistent with a C3-dominated environment suggesting predominance of forest taxa
386 (Fig. 5(B)). The period from 5.8 to around 4 ka cal BP highlighted in both residual and raw $\delta^{13}\text{C}_{\text{bulk}}$
387 presents less depleted values. It may suggest a short transition to C4-dominated savanna taxa
388 followed by a short return of the C3-dominated environment between 4 and 3 ka cal BP, with raw
389 $\delta^{13}\text{C}_{\text{bulk}}$ around -25‰. After 3 ka cal BP, the signal is consistent with an increasing influence of C4-type
390 plants.

391 Human activities such as turning peatland into cultured field also can have an impact on the
392 accumulation of carbon (Wang et al., 2017) and, consequently, on the $\delta^{13}\text{C}_{\text{bulk}}$ signature (Lane et al.,
393 2008). The influence of human occupation on sedimentary archives is reported in several studies from
394 West Africa during late Holocene such as around 2.6 ka cal BP at lake Barombi MBo (Cameroon)
395 (Garcin et al., 2018) or after 2.2 ka cal BP in the Congo River Basin (Bayon et al., 2019). It cannot be
396 excluded that the recent history of the NGAoundaba peat deposit was influenced by human
397 occupation.

398

399 **5.4. Influence of C-cycling and microbial organic matter**

400 Peat deposits can shift from a sink to a source of CO₂ and CH₄ gases reflecting changes in the microbial
401 communities and associated biogeochemical processes such as methanotrophy and methanogenesis
402 (Jauhiainen et al., 2005; Roehm, 2005). Even if the proportion, in terms of the mass of the microbial
403 organic matter, is not large, the specific $\delta^{13}\text{C}$ signature of some microbial compounds may not be
404 negligible. Indeed, some carbon sources such as methane can present important variations of their

405 carbon isotopic composition and values as depleted as -110‰ for direct reduction of CO₂ to CH₄
406 (Whiticar, 1999). The δ¹³C of methanotrophs' biomass is generally close to the carbon source signature
407 even if depletion or enrichment of cellular components can be observed depending on metabolic
408 pathway or environmental parameters (Jahnke et al., 1999). Thus, even a small proportion of highly
409 ¹³C-depleted microbial biomass could significantly modify the δ¹³C_{bulk}.

410 Mosses such as Sphagnum spp. can use recycled carbon in wet environments (Price et al., 1997;
411 Nichols et al., 2009). By extension, δ¹³C_{bulk} has also been proposed to reflect variations in the uptake
412 of C recycled from methanogenesis with enriched values corresponding to periods of high
413 methanogenesis (Jones et al., 2010). However, Sphagnum are not present at the NGaoundaba
414 peatland, which is dominated by sedges, so δ¹³C_{bulk} cannot be interpreted this way.
415 As an example, if we consider CH₄ with a δ¹³C signature close to -60‰, taking into account both direct
416 reduction of CO₂ to CH₄ and acetotrophic methanogenesis (Whiticar, 1999). In a first approximation,
417 we can consider that bulk organic matter produced by consumption of this methane will have a similar
418 δ¹³C signature as the source CH₄. Even a 1% input of this microbial organic matter in the bulk will lead
419 to a decrease of more than 0.5‰ of the δ¹³C_{bulk}. Present bulk data don't give us much information
420 about microbial activity. However, it is important to keep in mind that a microbial signal can
421 superimpose on the primary δ¹³C_{bulk} signal and either increase or mask pre-existing variations. Kuhry
422 and Vitt (1996) compared the evolution of bulk C/N in both the acrotelm and catotelm in a Sphagnum
423 peat, suggesting a predominant impact of peat decomposition on C/N values. The increase in acrotelm
424 is interpreted as a preferential N loss in oxic conditions while the decrease in catotelm is consistent
425 with a predominance of methanogenesis affecting C content while N is no longer affected by
426 decomposition and is much less mobile in the sediment under anoxic conditions. These observations
427 are not consistent with the general increase observed in C/N in the NGaoundaba peat deposit (Fig.
428 6(G)). This increase follows the general increase observed in TOC and is interrupted by a short interval
429 from 9 to 8 ka cal BP when the C/N increases and might suggest a change in the origin of the organic
430 matter (Talbot and Johannessen, 1992).

431

432 **5.5. Validation and limits of the approach**

433 The decreasing I-index observed in the NGaoundaba peat deposit from 6.4 ka cal BP to the surface
434 reflects the decrease of thermolabile compounds and the relative increase in thermostable
435 compounds, reflecting the progressive decomposition of labile organic matter and the increase in
436 thermal stability of organic matter from 6.4 ka cal BP until present (Sebag et al., 2016) (Fig. 5(D)).
437 During this period, raw TOC increases, indicating an enrichment in thermostable C-rich compounds
438 relatively to thermolabile compounds more rich in heteroatoms. Between the bottom of the core and
439 6.4 ka cal BP, the I-index remains rather stable; it is likely that most thermolabile compounds are
440 decomposed and it is consistent with a raw TOC around 50% with limited variations during this period.

441 Our method is based on the observation that for this particular peat deposit, the TOC increases
442 progressively with depth and follows nearly exactly a logarithmic curve. The function used for the
443 modelling will depend on the dynamics of the system (Clymo et al., 1998). Thus, this method is
444 undoubtedly not applicable to all peat deposits. This increasing trend is expected for peats composed
445 almost entirely of organic matter. The presence of mineral-rich intervals such as clayey or sandy
446 intervals or a steady input of a large amount of mineral matter will affect this trend (Baker et al., 2014;
447 Njagi et al., 2021). Such records are therefore not ideal for the modeling presented in this study. The
448 method gives a lot of weight to the top values of the core and slight analytical errors in these values
449 could have a strong impact on the model.

450 As vegetation may be the primary factor controlling $\delta^{13}\text{C}_{\text{bulk}}$ (Skrzypek et al., 2010), the curve fitting
451 might be distorted by the primary impact of vegetation changes. It may explain why the curve fitting
452 for $\delta^{13}\text{C}$ is not as good as for TOC with the modeling function rapidly diverging from the data (Fig. 3).
453 The residual $\delta^{13}\text{C}_{\text{bulk}}$ (Fig. 6(E)) erases two periods of relatively enriched $\delta^{13}\text{C}_{\text{bulk}}$ from 5.8 to 4 ka cal BP
454 and from the bottom of the core to 9 ka cal BP visible in the raw $\delta^{13}\text{C}_{\text{bulk}}$ record (Fig. 5(B)).

455 The comparison between residual TOC and HI from the same peatcore shows striking similarities (Fig.
456 6(D) & (G)). The HI can be controlled by the degree of decomposition but can also reflect changes in
457 the origin of the organic matter (Sebag et al., 2016). The higher values observed between 8 and 6 ka
458 cal BP and at the bottom of the core tend to indicate well preserved terrestrial organic matter that is
459 consistent with the TOC data and the wettest period of the African Humid Period. The decrease in HI
460 between 9 and 8 ka cal BP suggests a change in the origin of the organic matter or in its decomposition
461 and is consistent with the increase in the C/N ratio observed at the same period.

462

463 **Conclusions**

464 The present case study based on a West-African peat deposit presents several interesting
465 characteristics: it is nearly continuous, macroscopically homogeneous without clayey passages, and
466 some bulk parameters such as TOC present a strong logarithmic trend with age. Thus, it is possible to
467 propose a method to isolate and subtract the part of the variation only due to the decomposition of
468 organic matter and interpret the residual variations as modifications of the environmental conditions.
469 The residual TOC obtained with our method highlights well-known climatic transitions and variations
470 such as the African Humid Period and presents similar variations as other parameters such as HI.
471 Results obtained from residual C/N and $\delta^{13}\text{C}_{\text{bulk}}$ are more questionable, suggesting a greater impact of
472 parameters other than decomposition on the variations, such as vegetation shifts (Skrzypek et al.,
473 2010; Baker et al., 2014), changes in microbial processes (Moore, 1989; Girkin et al., 2020) or detrital
474 input (Amorim et al., 2022) associated with differences in the way the decomposition processes affect
475 these parameters (Ågren et al., 1996). Based only on bulk organic parameters, it is difficult to precisely
476 constrain the origin and timing of the proposed environmental and climatic changes, thus the use of
477 other proxies such as pollen data and biomarkers will help to confirm or infirm the climatic and
478 environmental interpretation proposed here.

479 This method is certainly not applicable to all peat deposits and we recommend a careful analysis of
480 the data and testing before applying it. We recommend applying this method to records composed
481 almost entirely of organic matter and without significant variation in the mineral fraction. This method
482 is a quick and straightforward way to obtain a first overview of the paleoclimatic and
483 paleoenvironmental variations from a new sediment core based on bulk analytical methods
484 (Elemental Analysis and Rock-Eval®). It can help to identify periods of great interest for more intensive
485 investigations (e.g., pollen, biomarkers), by highlighting the main transitions visible in bulk organic
486 data.

487

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494

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Figure captions

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773 **Fig. 1: Evolution of carbon mass and carbon concentration with time in OM-dominated sediment**
774 **reflecting the progressive mineralization of organic matter. Labile compounds are richer in**
775 **heteroatoms than resistant compounds, leading to an increase in C concentration as labile**
776 **compounds decompose while the global mass of carbon decreases.**

777

778 **Fig. 2: Description of the different steps of the procedure from raw to residual data.**

779

780 **Fig. 3: Normalized data of the NGaoundaba peat deposit plotted against age (ka cal BP &**
781 **normalized): (A) TOC, (B) $\delta^{13}\text{C}_{\text{bulk}}$ and (C) C/N. The light colored curves represent the curve fitting**
782 **based on the whole dataset and the dark colored curves represent the curve fitting based on the**
783 **partial dataset using only depth with a positive derivative. The function selected for each parameter**
784 **is indicated on the left of each panel, the constants a, b, c and d are fitted for each parameter and**
785 **x-range. The selected x-range for each parameter is indicated in dark color (encircled points). The x-**
786 **range limit is indicated in both age and depth for each parameter.**

787

788 **Fig. 4: Age-depth model of NGaoundaba peatland core. The corresponding section number is**
789 **indicated for each ^{14}C sample.**

790

791 **Fig. 5: Raw data (i.e., before data processing) from the NGaoundaba peat deposit: (A) TOC (%), (B)**
792 **$\delta^{13}\text{C}_{\text{bulk}}$ (‰), (C) C/N from Elemental and stable isotope analysis and the I-index from Rock-Eval[®]**
793 **analysis. Visible transitions are indicated with vertical red dashed lines.**

794

795 **Fig. 6: Retreated NGaoundaba peat deposit data and regional climatic and environmental settings:**
796 **(A) Insolation 0°N ($\text{W}\cdot\text{m}^{-2}$) from Berger et al. (1981), (B) $\delta\text{D}_{\text{n-alkC}_{31}}$ corrected for vegetation changes in**
797 **Lake Barombi (Southwest Cameroon) (Garcin et al., 2018), (C) Simplified pollen data from Lake**
798 **MBalang (about 25 km from the NGaoundaba peat deposit) (Vincens et al., 2010) (D) TOC (%), (E)**
799 **$\delta^{13}\text{C}_{\text{bulk}}$ (‰) and (F) C/N as well as (G) Hydrogen Index (HI) from Rock-Eval[®] analysis. The African**
800 **Humid Period is highlighted in green, the termination of this period has been represented as a color**
801 **gradient to illustrate the uncertainties and different possible timings of the termination. The mean**

802 value of HI is indicated as a dashed line. Visible transitions are indicated with vertical red dashed
803 lines.

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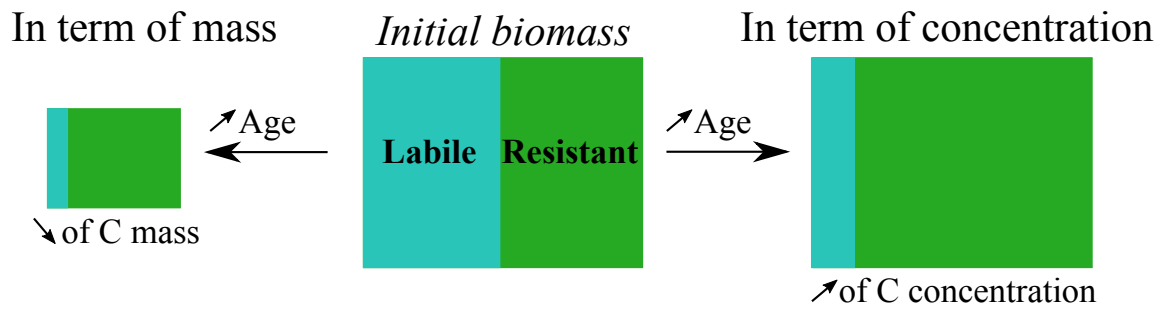
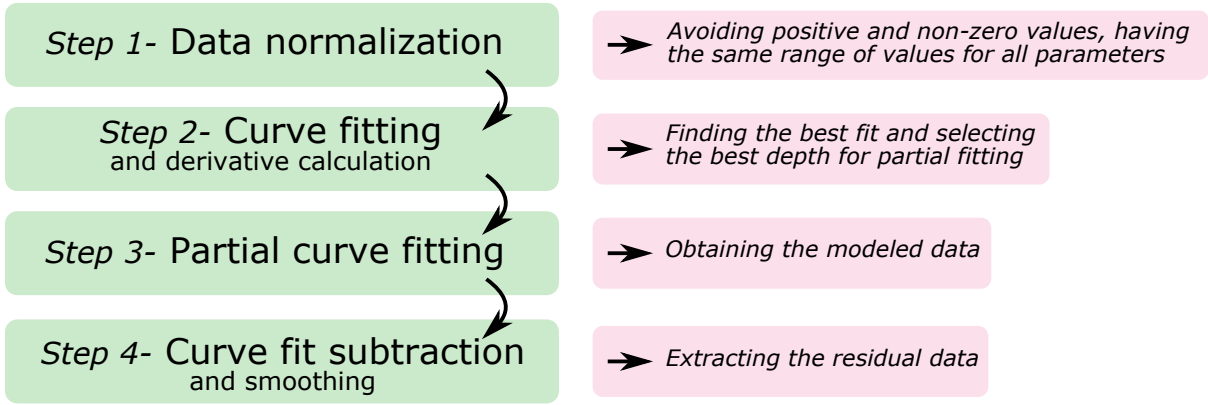


Fig. 1:

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Fig. 2:

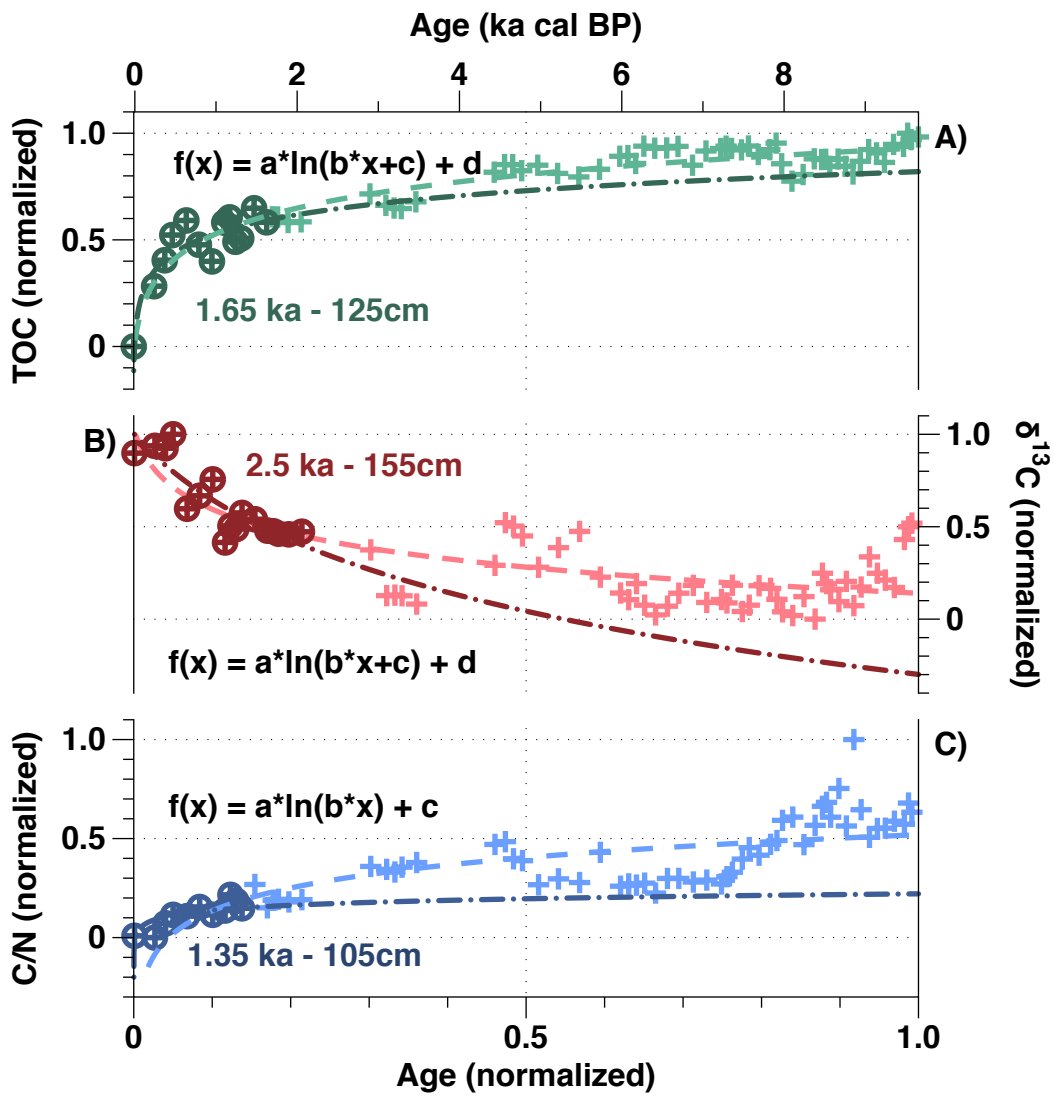


Fig. 3:

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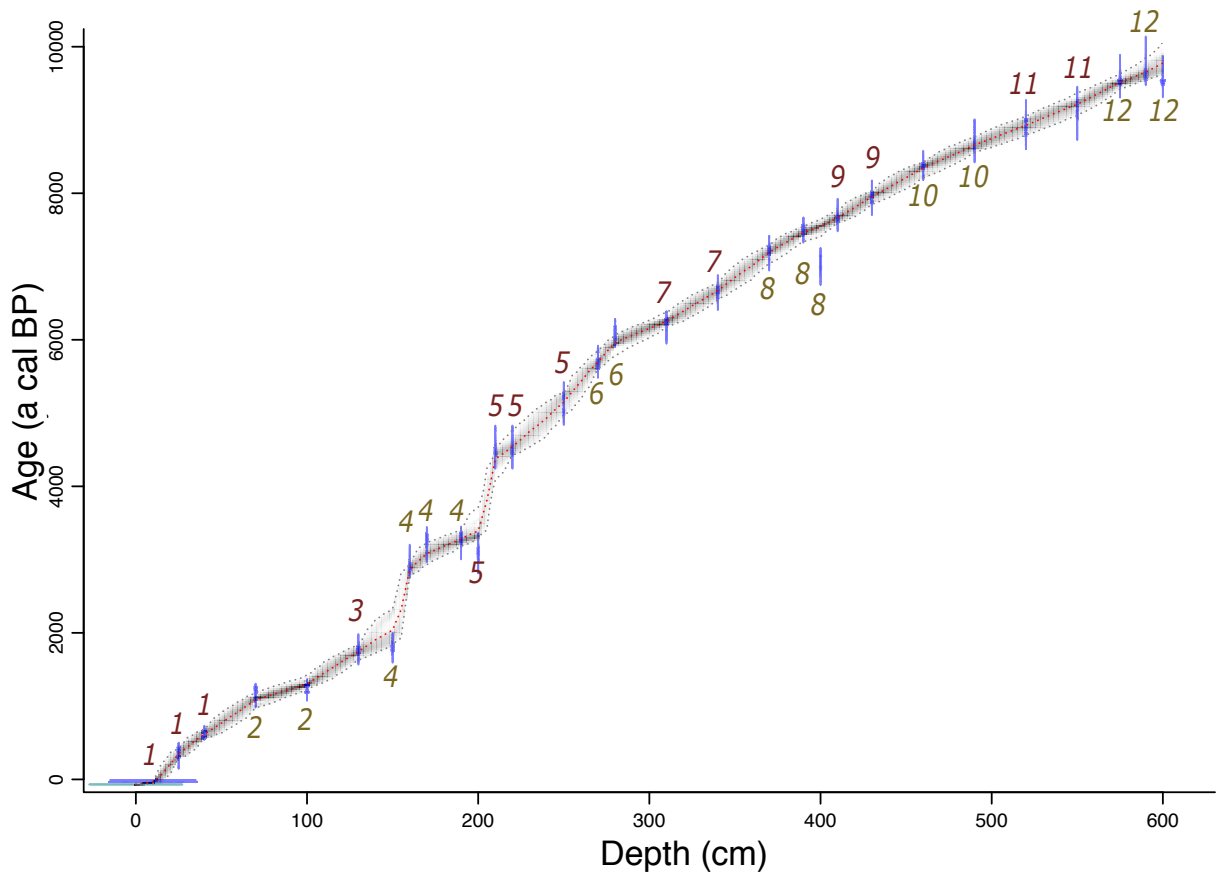


Fig. 4:

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Sample code	Depth (cm)	Uncalibrated Age	Calibrated age (a cal BP)
58976	10	120 pMC	-39
62587	25	330 ± 30 BP	343
58977	40	725 ± 30 BP	614
58978	70	1310 ± 30 BP	1093
58979	100	1390 ± 30 BP	1290
58980	130	1870 ± 30 BP	1741
62588	150	1910 ± 30 BP	1981
64844	160	2840 ± 30 BP	2871
58981	170	3090 ± 30 BP	3082
58982	190	3130 ± 30 BP	3274
62589	200	2975 ± 30 BP	3345
64845	210	4070 ± 30 BP	4417
58983	220	4070 ± 30 BP	4531
58984	250	4475 ± 30 BP	5180
62590	270	5000 ± 35 BP	5690
58985	280	5335 ± 30 BP	5960
58986	310	5460 ± 30 BP	6255
58987	340	5875 ± 30 BP	6668
58988	370	6335 ± 35 BP	7201
64846	390	6660 ± 30 BP	7465
58989	400	6170 ± 30 BP	7547
62591	410	6870 ± 40 BP	7665
58990	430	7180 ± 40 BP	7955
58991	460	7600 ± 35 BP	8361
58992	490	7925 ± 40 BP	8653
58993	520	8080 ± 35 BP	8926
58994	550	8210 ± 40 BP	9223
62592	575	8605 ± 45 BP	9511
64847	590	8700 ± 40 BP	9642
58995	600	8590 ± 40 BP	9759

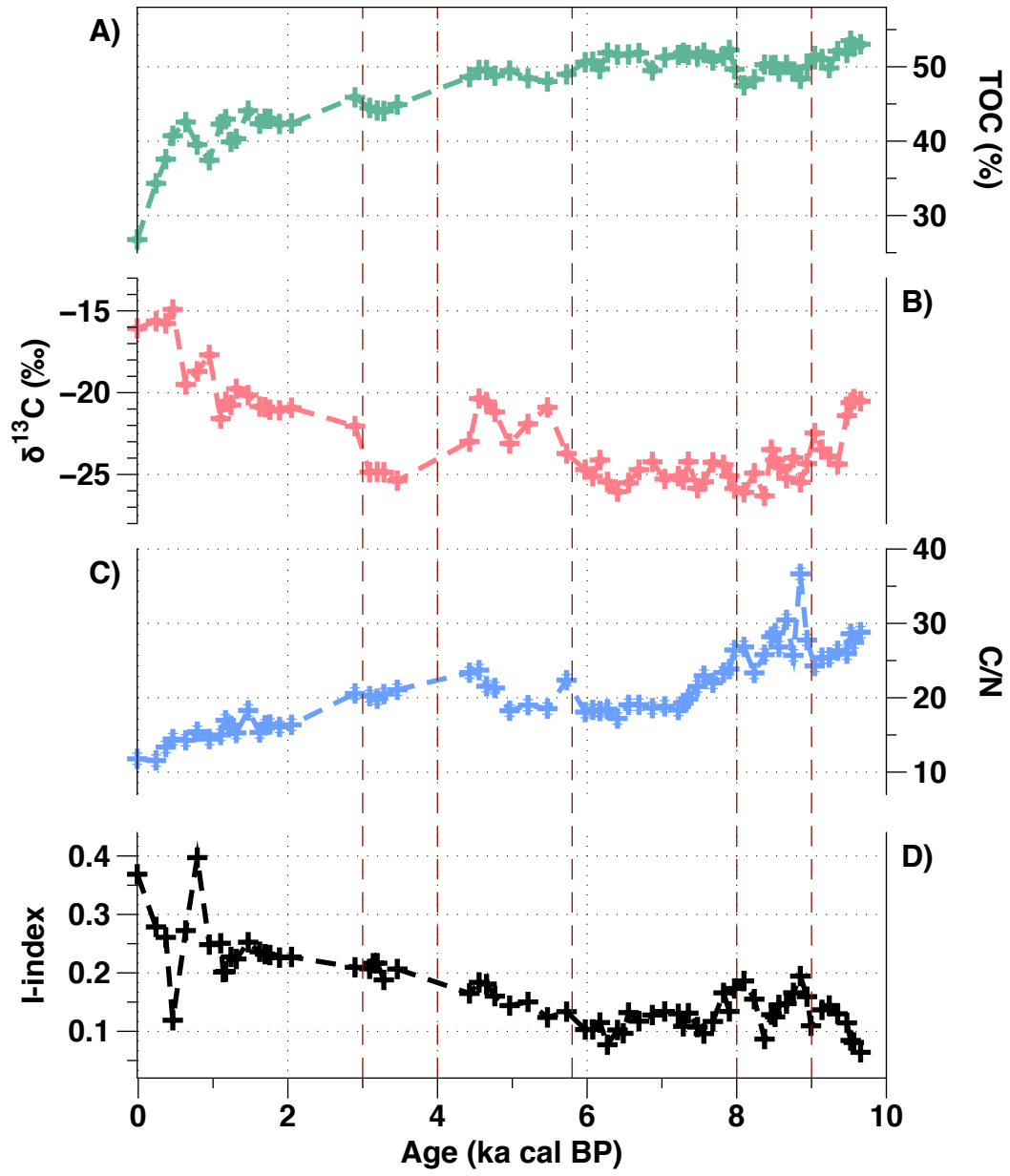
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888 **Table 1: Chronology of NGaoundaba peatland with uncalibrated AMS ¹⁴C ages (BP : before present)**
889 **and calibrated median ¹⁴C age (a cal BP).**

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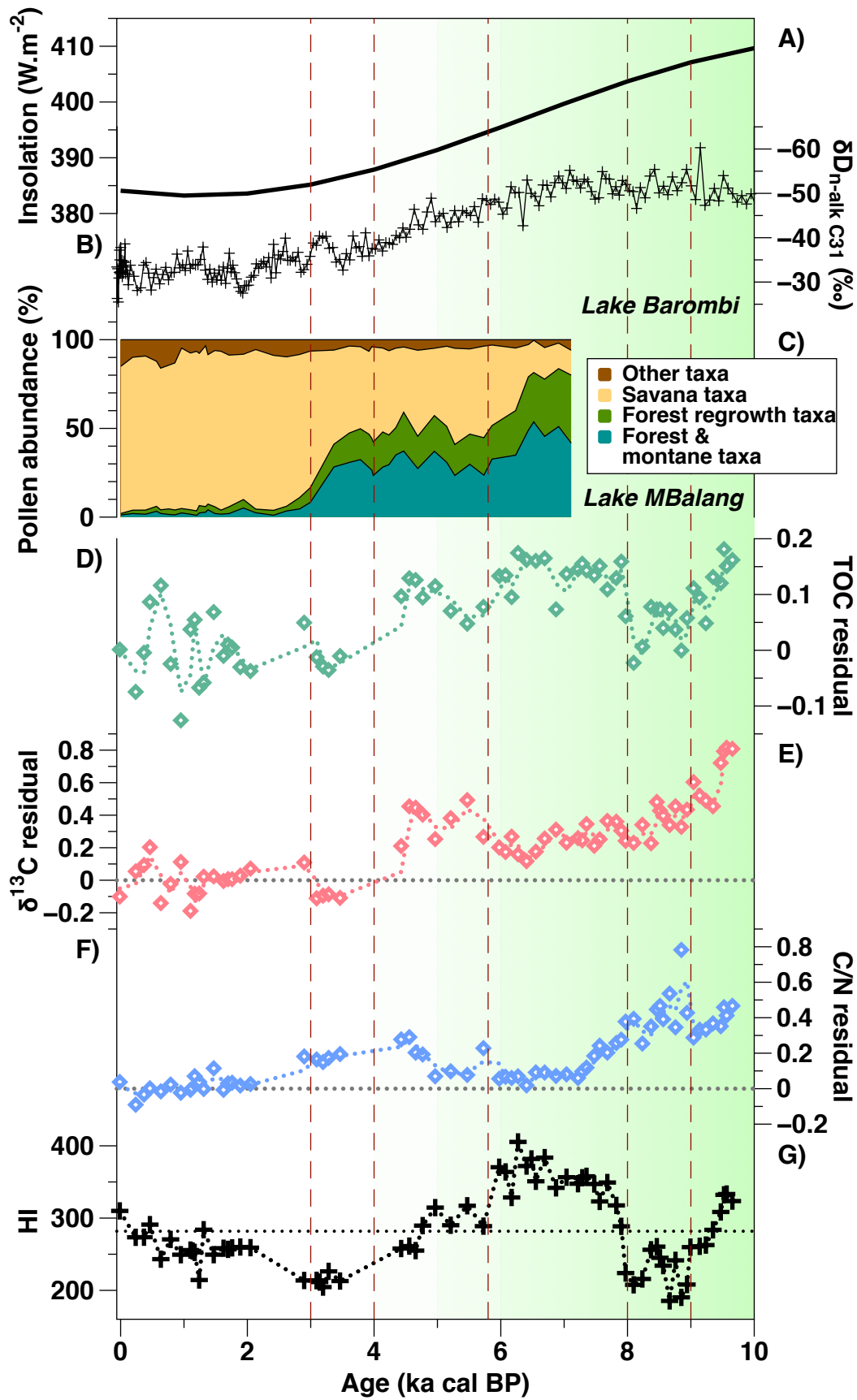
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Fig. 5:



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Fig. 6: