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## **Modeling the Decomposition Signal and Correcting Bulk Organic Data From a Peat Deposit, a Case Study at Low Latitudes (Cameroon)**

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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)**

3  
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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  $^{12}\text{C}$ ,  
110 resulting in an increase in  $^{13}\text{C}$  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'nanga 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 m<sup>2</sup>, 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 (CO<sub>2</sub>), released during  
166 the thermal cracking of organic compounds under pyrolytic conditions (from 200 to 650°C at 25°C min<sup>-1</sup>  
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 CO<sub>2</sub> 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 CO<sub>2</sub>) 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 δ<sup>13</sup>C  
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 CO<sub>2</sub>, N<sub>x</sub>O<sub>x</sub>, and H<sub>2</sub>O. Then, a reduction furnace at  
200 550°C reduced N<sub>x</sub>O<sub>x</sub> to N<sub>2</sub> and removed excess O<sub>2</sub> using reduced copper. H<sub>2</sub>O was trapped and  
201 removed using a phosphorous pentoxide chemical trap. N<sub>2</sub> and CO<sub>2</sub> 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,  $\alpha$  denotes the manner of decay and constant,  
 217 linear and quadratic rules were tested by Clymo et al. (1998). Using a linear decay rule  $\alpha_L^*$ , 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 CO<sub>2</sub> and CH<sub>4</sub> 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 CO<sub>2</sub> to CH<sub>4</sub>  
406 (Whiticar, 1999). The δ<sup>13</sup>C 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 <sup>13</sup>C-depleted microbial biomass could significantly modify the δ<sup>13</sup>C<sub>bulk</sub>.

410 Mosses such as Sphagnum spp. can use recycled carbon in wet environments (Price et al., 1997;  
411 Nichols et al., 2009). By extension, δ<sup>13</sup>C<sub>bulk</sub> 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 δ<sup>13</sup>C<sub>bulk</sub> cannot be interpreted this way.  
415 As an example, if we consider CH<sub>4</sub> with a δ<sup>13</sup>C signature close to -60‰, taking into account both direct  
416 reduction of CO<sub>2</sub> to CH<sub>4</sub> 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 δ<sup>13</sup>C signature as the source CH<sub>4</sub>. 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 δ<sup>13</sup>C<sub>bulk</sub>. 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 δ<sup>13</sup>C<sub>bulk</sub> 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.**

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788 **Fig. 4: Age-depth model of NGaoundaba peatland core. The corresponding section number is**  
789 **indicated for each  $^{14}\text{C}$  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<sup>®</sup>**  
793 **analysis. Visible transitions are indicated with vertical red dashed lines.**

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795 **Fig. 6: Retreated NGaoundaba peat deposit data and regional climatic and environmental settings:**  
796 **(A) Insolation  $0^\circ\text{N}$  ( $\text{W}\cdot\text{m}^{-2}$ ) 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<sup>®</sup> 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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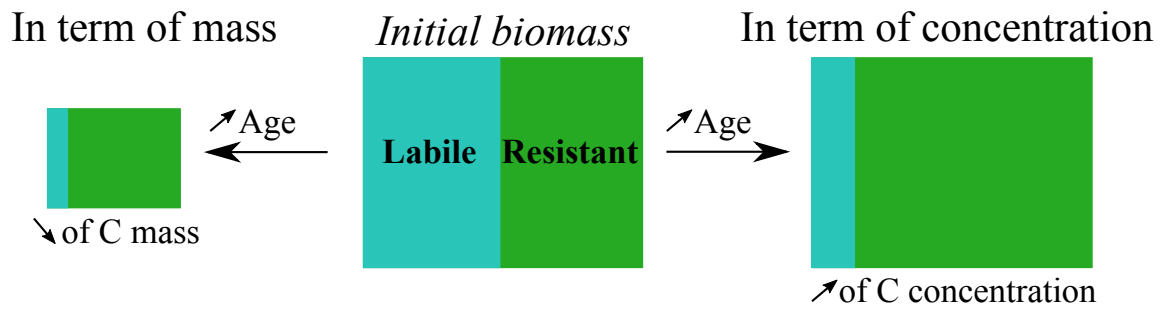
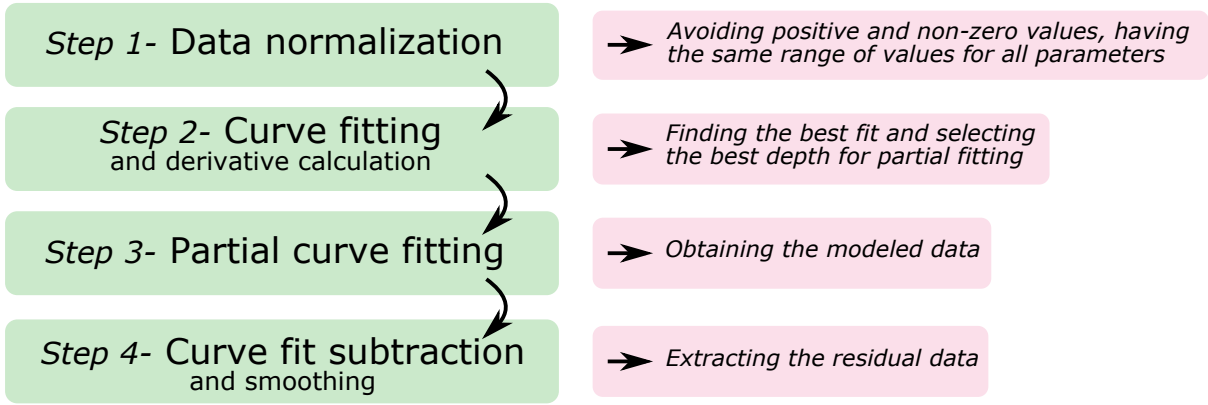


Fig. 1:

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**Fig. 2:**

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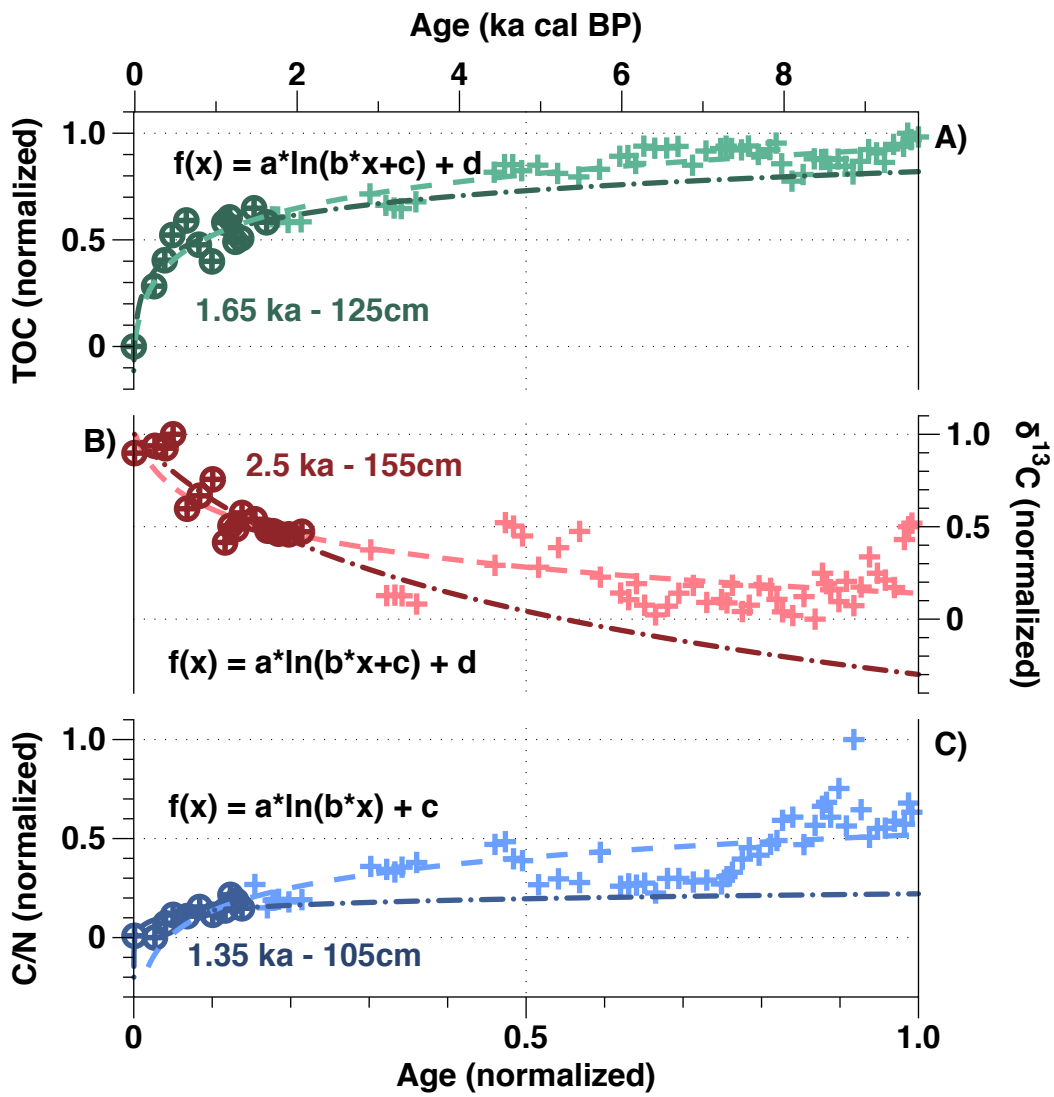


Fig. 3:

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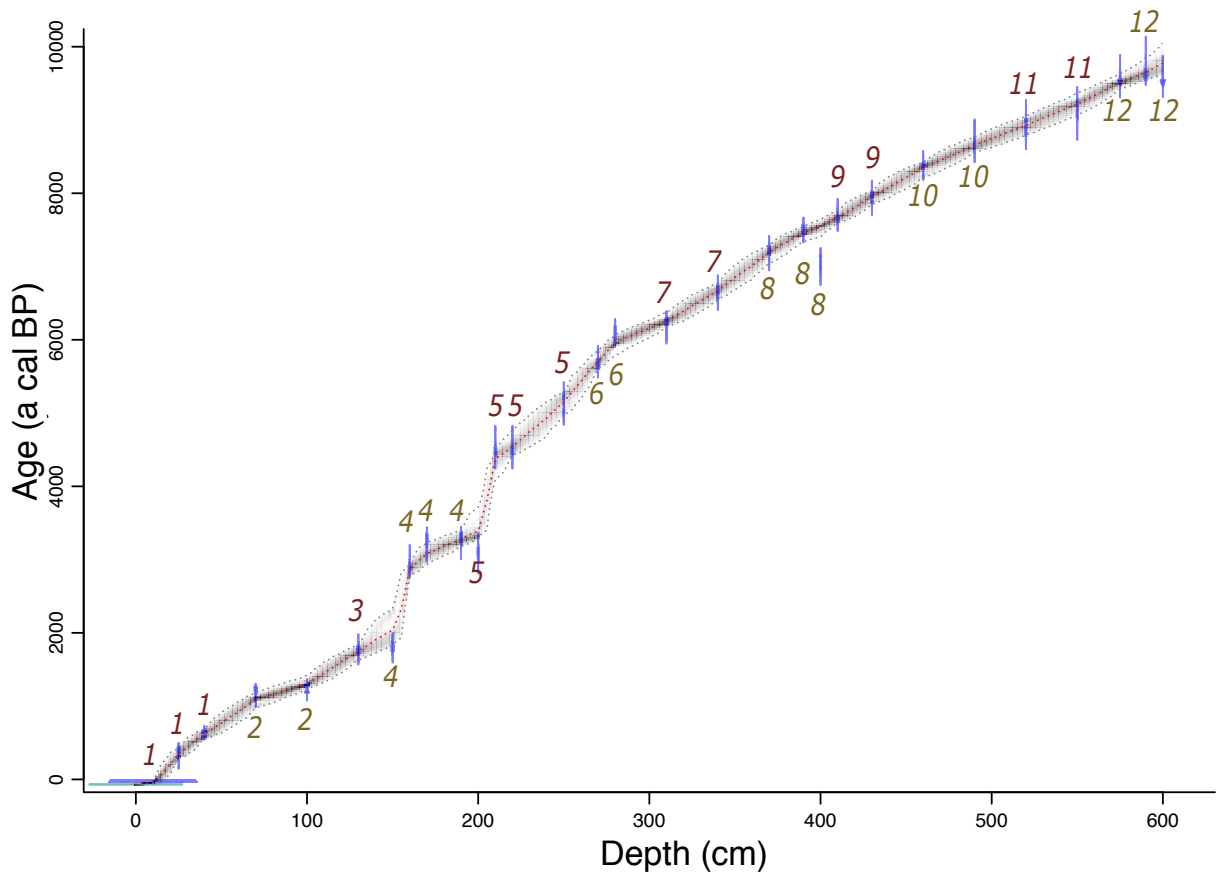


Fig. 4:

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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

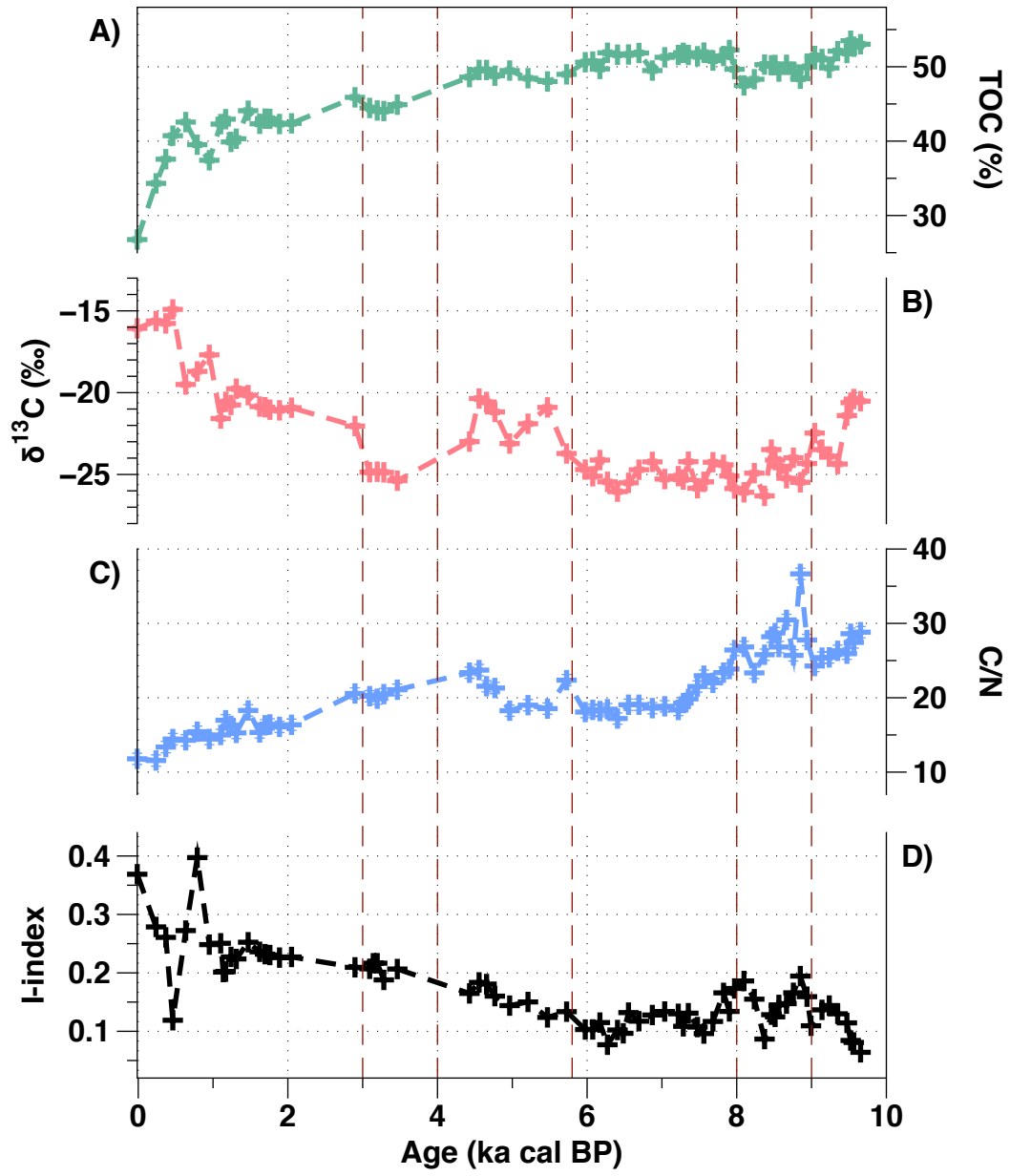
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888 **Table 1: Chronology of NGaoundaba peatland with uncalibrated AMS <sup>14</sup>C ages (BP : before present)**  
889 **and calibrated median <sup>14</sup>C age (a cal BP).**

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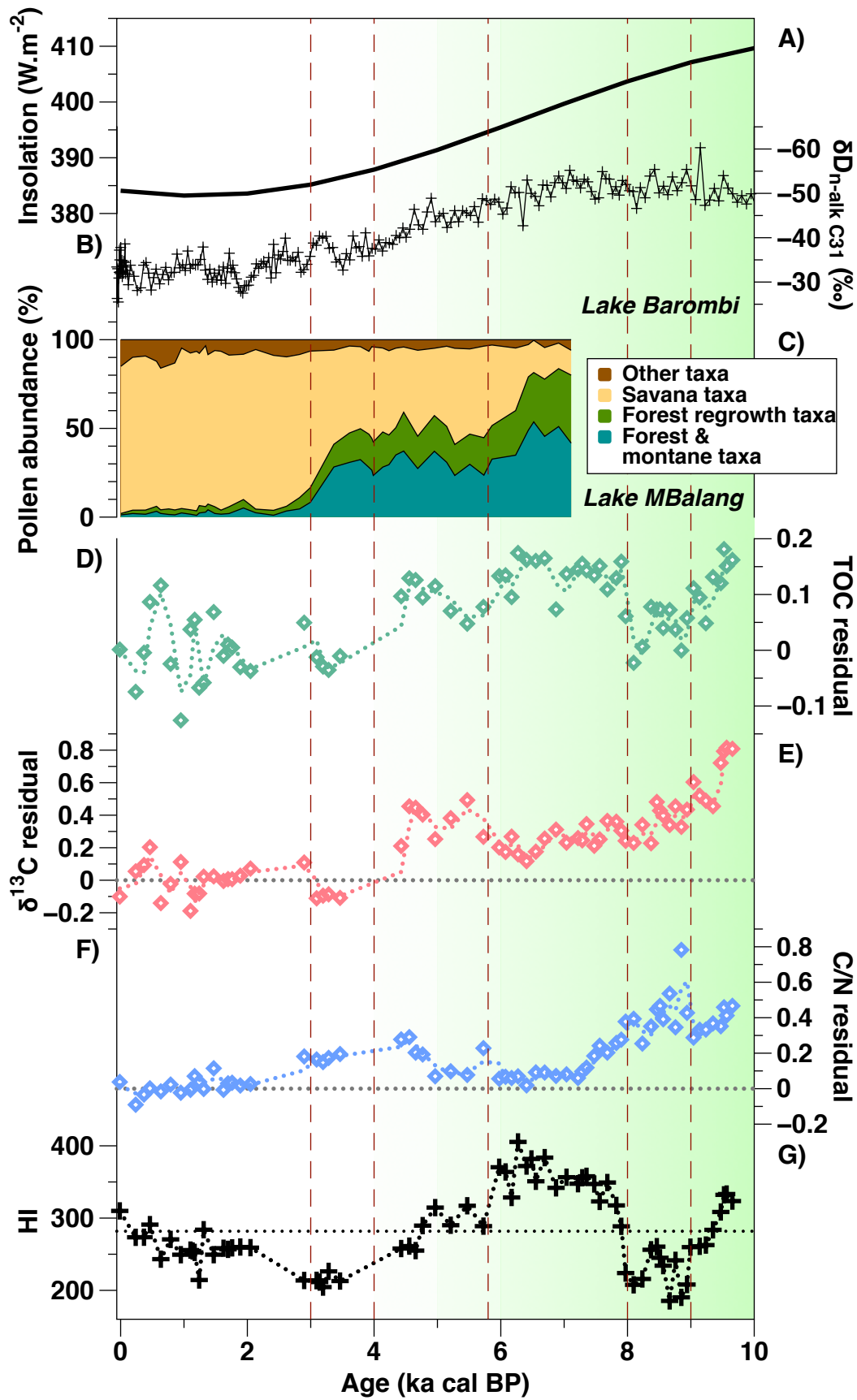
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Fig. 5:



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Fig. 6: