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1 Impact of climate change on the ecology of the Kyambangunguru Crater Marsh
2 in southwestern Tanzania during the Late Holocene

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32 **Abstract**

33 Instrumental records of temperature and hydrological regimes in East Africa evidence frequent droughts
34 with dramatic effects on population and ecosystems. Sources of these climatic variations remain largely
35 unconstrained, partly because of a paucity of Late Holocene records. Here, we present a multi-proxy
36 analysis of a 4-m continuous sediment core collected in the Kyambangunguru crater marsh, in southwest
37 Tanzania, covering the last 4000 yrs (cal. BP). We used microscopic (macro-remains, microfossils,
38 palynofacies, pollen), elemental (carbon, nitrogen contents), molecular (br GDGTs, *n*-alkanes) and
39 compound-specific isotopic ($\delta^2\text{H}$ *n*-alkanes) investigations to reconstruct the environmental history of
40 the marsh. The multi proxy record reveals that, 2500 years ago, the marsh underwent a major ecological
41 transition from a lake to a peatland. Temperature and hydrological reconstructions evidence warmer and
42 drier conditions between 2200 and 860 cal. BP, which probably triggered the establishment of a
43 perennial peatland. This study is one of the first combined temperature and precipitation record of Late
44 Holocene in the region and highlights changes in the spatial distribution of the East African climate
45 regimes. Several cold periods are observed, between 3300 and 2000 cal. BP and since 630 cal. BP, the
46 latter corresponding to the Little Ice Age. Moreover, wetter conditions are reported during the Medieval
47 Climate Anomaly in contrast to other north-eastern African records suggesting that Tanzania is located
48 at the transition between two hydro-climatic zones (north-eastern versus southern Africa) and has
49 experienced variable contributions of these two zones over the last millennium.

50

51 **Keywords**

52 Holocene, Paleoclimatology, Paleolimnology, East Africa, Continental biomarkers, Organic
53 geochemistry, Stable isotopes, Palynology.

54

55 **1. Introduction**

56 Tropical highlands are major sources of food and freshwater for more than 35 tropical countries
57 (Williamson, 2014). The climate dynamics and variability of these topographically complex
58 environments, however, remain poorly studied. While it has been shown that the Quaternary climatic
59 trends in East Africa were primarily controlled by orbital forcing (e.g. Garcin et al., 2006; Tierney et al.,
60 2008), the shorter scale climate dynamics of this region is largely unconstrained. Notably, mid- to late
61 Holocene records of many East African lakes (e.g. Gasse, 2000; Wanner et al., 2011) suggest rapid and
62 frequent, high amplitude, climatic fluctuations at the centennial scale. These fluctuations and their
63 consequences are not well understood due to a general lack of highly resolved records (Nicholson et al.,
64 2013). Furthermore, the timing and intensity of these events are not always synchronous from site to
65 site (Tierney et al., 2011, 2013). Here, we present detailed records of climate and ecosystem changes
66 from a sequence of sediments covering the late Holocene (the last 4000 years) in the Kyambangunguru
67 marsh. This marsh is located in the Rungwe Volcanic Province (RVP; southwest Tanzania), a highland
68 representing one of the four major food crop producing regions in the country (Majule, 2010).

69 Marshes and peatlands have a great potential for quantitative high-resolution palaeoclimatic records
70 (Amesbury et al., 2012; Blackford, 2000) notably in the tropics (e.g. Bonnefille et al., 1990; Bourdon et
71 al., 2000; Page et al., 2011; Rucina et al., 2010; Swindles et al., 2018). However, they are highly dynamic
72 ecosystems where the vegetation cover and the hydrology functioning can be totally modified at a
73 centennial scale (Loisel and Yu, 2013). This may complicate the interpretation of climatic proxies,
74 notably those based on biological markers as their fluctuations may be related to ecological, local change
75 rather than regional climatic change. A major challenge in using marsh/peat records as climatic archives
76 is thus to disentangle biological signals linked to dynamic changes of the peatland ecosystem itself from
77 those that are driven by local to regional environmental change (Chambers et al., 2012; Morris et al.,
78 2015). The focus of this study is to investigate the internal, ecological changes within the marsh in the
79 context of regional climatic variations. We aim to retrieve detailed (quantitative) air temperature and
80 (qualitative) hydrological condition records of the late Holocene from the southernmost part of East
81 Africa to test whether rapid and high amplitude climatic events (e.g. Russell and Johnson, 2005; Wanner

82 et al., 2011) were recorded in this area in comparison to other East African records. Additionally, the
83 multi-proxy approach, combining microscopic observations and geochemical characterization, intends
84 to determine potential feedbacks of these rapid climatic events in the tropical highland wetlands as well
85 as potential human impact in the region.

86 Analysis of pollen, non-pollen palynomorphs (NPPs), macro-remains, palynofacies and bulk elemental
87 (C and N content) determination was conducted to characterize the ecological states of the wetland,
88 complemented by biomarker-based proxies to determine past variations in air temperature and
89 hydrology. Branched glycerol dialkyl glycerol tetraethers (br GDGTs) and compound specific long
90 chain *n*-alkane hydrogen isotopic composition ($\delta^2\text{H}_{\text{wax}}$) were used for mean annual air temperature and
91 hydrological conditions reconstruction, respectively. Br GDGTs are membrane lipids produced by
92 unknown bacteria (Sinninghe Damsté et al., 2000) whose relative abundances in environmental samples
93 have been shown to correlate with temperature and pH (Weijers et al., 2006, 2009). This enabled the
94 reconstruction of past pH and air temperatures from the br GDGT distribution in sediments, peats and
95 soils (Nichols et al., 2014; Peterse et al., 2011; Weijers et al., 2007a). Long chain *n*-alkanes are
96 constituents of the epicuticular wax layer of leaves (Eglinton and Hamilton, 1967). It has been shown
97 that their hydrogen isotopic composition ($\delta^2\text{H}_{\text{wax}}$) reflects the hydrogen isotopic composition of the water
98 taken up by the plants (e.g. Estep and Hoering, 1980; Sauer et al., 2001; Sessions et al., 1999).
99 Accordingly, they can be used to reconstruct variations in local palaeohydrology as shown in several
100 lacustrine sedimentary archives from the Quaternary and the Holocene in East Africa (e.g. Loomis et
101 al., 2015; Powers et al., 2005; Tierney et al., 2008; Verschuren et al., 2000). The combined use of these
102 two proxies allows distinguishing the temperature from the hydrological signal which has been a major
103 limitation in lake-based East African climatic reconstructions (Verschuren, 2003). Moreover, in settings
104 with high sedimentation rates like marshes, they can offer highly detailed and independent
105 reconstruction of the temperature and the hydrological conditions.

106

107

108

109 **2. Regional setting: the Rungwe Volcanic Province and the Kyambangunguru marsh**

110 The Rungwe Volcanic Province (RVP; SW Tanzania; Fig. 1A), is a large volcanic mountain region
111 (1500 km²) located at the triple junction of the Malawi Rift, Rukwa/Tanganyika Rift and the Usanga
112 Basin in the southern part of the East African Rift System (Fontijn et al., 2010, 2012). The RVP is
113 delimited by the Poroto Mountains in the north, Lake Malawi in the south and the Livingstone
114 escarpment in the west (Fig. 1B). The area is known to be seismically active with volcanic eruptions
115 occurring from the late Miocene (9.2 Ma) to the 19th century, with hot spring activity still found today
116 (Branchu et al., 2005). The region contains three major stratovolcanoes: the Ngozi, Kyejo and Rungwe
117 (Fontijn et al., 2010, 2012). South of these high-altitude sites and north of Lake Malawi lies the Karonga
118 plain. Several monogenic maar-type craters were created during late Pleistocene phreatomagmatic
119 explosions along the Mbaka fault system and are now filled by closed lake hydro-systems (Fontijn et
120 al., 2012; Fig. 1B). The region belongs to the humid equatorial zone of Africa, mainly determined by
121 the migration of the Intertropical Convergence Zone (ITCZ), a key atmospheric feature of tropical
122 atmospheric circulation with low-pressure air masses accompanied by high precipitation. The ITCZ
123 reaches its southernmost position (centred at ca. 15°S, Fig. 1A) in January, resulting in seasonal
124 fluctuations between hot humid conditions from November to May and relatively colder and dry
125 conditions from June to October (Fig. 1C). The RVP is among the most humid regions of Tanzania
126 along with the coastal zone (Basalirwa et al., 1999). It is characterized by a different rainfall distribution
127 with persisting rainfall in April-May. Nivet et al. (2018) showed that the Indian tropical Ocean and the
128 Austral Ocean are the main sources of moisture in the area, with only a minor influence of the Congo
129 Air Mass. Thus, the currently observed variability of the regional rainfall is likely highly impacted by
130 the Indian Ocean Dipole, through Sea Surface Temperature anomalies. Over the last century, climatic
131 trends from the RVP point towards drier conditions associated with a shorter rain season (Williamson
132 et al., 2014) and a continuous increase in temperature ($\approx 1^\circ\text{C}$ for the last 100 yr.; Branchu et al., 2005).
133 Typical vegetation of the region includes Zambezian Miombo-type woodland at low altitude and
134 Afromontane vegetation at higher altitude (Garcin et al., 2006, Williamson et al., 2014). In many
135 locations, the woodland has been replaced by diverse crops (banana, rice, cocoa, tea, coffee, maize;

136 Coffinet et al., 2017; Williamson et al., 2014). The RVP is today one of the main agricultural resources
137 of Tanzania (Majule, 2010).

138 The Kyambangunguru marsh (9°22' S - 33°47' E, 660 m a.s.l.) is located in one of the numerous maar
139 craters of the RVP, between the Mbaka River and the Mbaka fault. These maar craters are essential
140 water and biodiversity resources for the region. At Kyambangunguru, no human activity has been
141 recorded nor is known within the crater (no land or water use). Human settlement expands in the plains
142 surrounding the volcano (mainly family-scale farming) but not on its slopes. The inner marsh covers
143 about 0.04 km² and its catchment area – limited to the crater slopes elevated ca. 100 m above the water
144 table – is relatively small (0.20 km²; Delalande et al., 2008a, Fig. 1D). According to the Lwifwa Masoko
145 station of the University of Dar es Salaam located at Lake Masoko, 7.5 km to the south east, mean
146 precipitation (P) is up to 2099 mm.yr⁻¹, with April being the most humid month (470 mm in average)
147 and September the driest one (8 mm in average; Nivet et al., 2018). Air temperature fluctuates around
148 22 °C throughout the year; July is the coldest month (19 °C on average) and November the warmest (25
149 °C on average). The crater depression is filled with peat like deposits overgrown by marsh-type
150 vegetation and the water level (H) varies around 70 cm of amplitude over the year (Fig. 1C). At the end
151 of the rainy season, the marsh resembles a shallow lake with floating vegetation mats and patches of
152 free water surface (Fig. 1E) while the water table considerably decreases during the dry season (Fig.
153 1C). On a monthly scale, water level fluctuations (ΔH) correlate with the rainfall (P): $\Delta H = 1.03 P - 18$
154 (in cm; n = 12; r = 0.97). During the humid season, the marsh water is warm (around 25 °C), low
155 mineralized and slightly acidic (pH around 5.8). On the contrary, during the dry periods, it is
156 characterized by higher mineral concentration as well as pH increase up to around 6.4, because of water
157 evaporation at the surface (Delalande, 2008; Delalande et al., 2008a). The water residence time in the
158 marsh is short (a few months), based on the isotopic water budget of the marsh in- and outputs ($\delta^{18}\text{O}$
159 and $\delta^2\text{H}$ of H₂O; Delalande et al., 2008a), suggesting a prominent influence of climate in the marsh
160 water budget. The isotopic signature of the marsh water during the humid season ($\delta^2\text{H-H}_2\text{O} = -7\text{‰}$) is
161 close to the mean annual isotopic signature of precipitation recorded at Lwifwa Masoko station (Nivet
162 et al., 2018). At the end of the dry season, the marsh water isotopic signature ($\delta^2\text{H-H}_2\text{O}$) becomes more
163 ²H-enriched (between 7 and 16‰, Fig. 1C) demonstrating significant evaporation resulting from the

164 drier climatic conditions. The slopes of the crater are steep and covered by Zambebian-type (Miombo)
165 forest, dominated by *Brachystegia*, *Uapaca* and *Acalypha* tree species, all common to the region (White,
166 1983). Shrubs of Rubiaceae and Myriaceae families are present at the edge of the marsh. The vegetation
167 of the marsh is dominated by sedges (*Carex*, *Cyperus*) while floating (*Nymphaea*) and submerged
168 macrophytes are abundant in the depressions filled with water during the rainy season.

169

170 **3. Materials and methods**

171 3.1. Core retrieval and sampling

172 A 4-m long core was collected with a Wright corer in the Kyambangunguru marsh in December 2012.
173 The coring process was stopped at 4 m because of the thickness of the Rungwe Pumice tephra (Fontijn
174 et al., 2012) and thus covers the most recent history of the Kyambangunguru wetland. The core was
175 sampled in 1 cm thick slices at the Lwifwa-Masoko station of University of Dar es Salaam and kept at
176 -20 °C until further treatment. 21 samples were chosen for dating while 35 samples were selected every
177 12 cm to perform the elemental (carbon and nitrogen content), molecular (br GDGTs, *n*-alkanes) and
178 isotopic ($\delta^2\text{H}_{\text{wax}}$) analyses. Total organic carbon (C_{org}) and nitrogen (N) contents were determined after
179 decarbonation by elemental analysis at the Service Central d'Analyse du CNRS, Villeurbanne, France.
180 105 1 cm-thick samples (ca. every 3 cm) were selected for plant macrofossil analysis. Among these, 44
181 samples (distributed evenly along the core) were additionally analysed for pollen and non-pollen
182 palynomorphs (NPPs). Within these 44 samples, 12 were used for palynofacies determination.
183 Additionally, 5 surface soil samples (0-5 cm) from the catchment area were collected between the marsh
184 and the top of the crater.

185

186 3.2. Absolute chronology and sediment accumulation rate

187

188 The chronology of the core is based on 21 Accelerator Mass Spectrometry (AMS) dates performed on
189 bulk Total Organic Matter (TOM; 17 samples) and wood fragments (4 samples; Table 1) all along the
190 core. Samples were subjected to acid-alkali-acid treatment to remove the mineral phase. AMS- ^{14}C and

191 associated $\delta^{13}\text{C}$ analyses were conducted on aliquots prepared according to the following AMS protocol:
192 burning at 860 °C for 30 min under vacuum, in the presence of a Cu (II)-oxide/Cu (III)-oxide mix and
193 of Ag string. The obtained CO_2 was graphitized on powdered Fe with H_2 at 650 °C for 100 min, and
194 graphite was compressed in analytical pellets. Residual CO_2 gases were used for associated ^{13}C
195 measurements on a SIRA 10 and are expressed in delta notation per mil versus V-PDB (Vienna Pee Dee
196 Belemnite). Graphite preparation and $\delta^{13}\text{C}$ measurements were made at the GEOPS Laboratory
197 (University of Paris-Saclay, France). ^{14}C counting was performed using the AMS facility at the LMC14
198 laboratory (Laboratoire de Mesure du Carbone 14; Artemis, Saclay; Cottureau et al., 2007). Analytical
199 uncertainties, are $\pm 0.1\%$ for $\delta^{13}\text{C}$ and between 0.2 and 0.5 pMC (percentage of Modern Carbon) for ^{14}C
200 activity.

201 Calibrated radiocarbon ranges were obtained using OxCal 4.3 software (Bronk Ramsey, 2009) with the
202 ShCal13 (Hogg et al., 2013) and Bomb13SH3 (Hua et al., 2013) atmospheric curves as the calibration
203 set (Table 1). A Bayesian age-depth model was used to establish an absolute chronology based on these
204 calibrated ^{14}C date ranges. The age-depth model was constructed applying a *P_Sequence* function, with
205 the parameters $k_0=1$ and $\log_{10}(k/k_0)=1$, in the OxCal v. 4.3 software (Bronk Ramsey, 1995, 2008).
206 Additionally, boundaries reflecting potential changes in the rate of deposit accumulation were
207 introduced to the model, based on observations of micro- and macrofossils (see section 4.5 and 4.6).
208 These boundaries were defined as follows: (i) 417.5 cm: top of the tephra layer and bottom of the model,
209 (ii) 191 cm: abrupt change from lacustrine to marshland conditions, (iii) 115.5 cm: distinct increase in
210 water table and (iv) 0 cm: top of the core. Computing of the age-depth model led to the exclusion of two
211 samples (SacA40028 and SacA38523). Dates with the lowest individual agreement between the
212 modelled and the calibrated date, i.e. SacA40077 and SacA40076, were also excluded from the
213 calculations until the lowest critical value of the agreement index (A_{model}) suggested by Bronk Ramsey
214 (2008) for model reliability ($A_{\text{model}} = 60\%$) was achieved. Final A_{model} value of the chronology was 63%.
215 The age is presented as a μ (mean) value of the modelled age expressed as calibrated year before present
216 i.e. AD 1950 (cal. BP; Fig. 2), rounded to tens.

217 A mean value for the sedimentation accumulation rate (SAR), expressed in cm year^{-1} , was determined
218 as the median value of the probability distribution of the modelled age (μ) for each depth (in cm) at
219 which a date was modelled. The applied formula was the following:

$$220 \quad \text{SAR} = \frac{1}{\mu_{\text{depth}-0.5} - \mu_{\text{depth}+0.5}} \quad (1)$$

221

222 3.3. Plant macrofossil analysis

223 About 5 cm^3 of sediment were rinsed with warm water and sieved at 0.25 mm. Macrofossils were studied
224 in transmitted light with a Nikon SMZ800 stereoscopic microscope at a magnification of 10 to 200.
225 Species determination of individual plant macrofossils was performed based on the data from
226 Velichkevich and Zastawniak (2006, 2009). The data were presented as numbers of detected
227 macrofossils and were presented as diagram drawn in the POLPAL software (Nalepka and Walanus,
228 2003). Analysis was performed at the Adam Mickiewicz University in Poznań (Poland).

229

230 3.4. Pollen, non-pollen palynomorphs and microscopic charcoal

231 Samples for pollen, non-pollen palynomorph and charcoal analysis were prepared using standard
232 laboratory procedures: adding 10% HCl to dissolve carbonates, heating in 10% KOH to remove the
233 humic fraction and at least 24-hour treatment with HF to remove the mineral fraction followed by
234 acetolysis (Berglund and Ralska-Jasiewiczowa, 1986). *Lycopodium* tablet of known number of spores
235 ($n=20848$, produced by Lund University) was added to each sample for calculation of microfossil
236 concentration (Stockmarr, 1971). Pollen and spores were counted with a biological microscope under
237 $400\times$ and $1000\times$ magnification until the number of at least 500 pollen grains was obtained (Vincens et
238 al., 2003, 2007). Pollen grains were identified using atlases (Gosling et al., 2013), internet-based
239 databases such as African Pollen Database (<http://apd.sedoo.fr/>), and the Universal Pollen Collection
240 (Institut des Sciences de l'Évolution Montpellier; <http://www.palyno.org/>). Non-pollen palynomorphs
241 (NPPs) were identified using available literature (Gelorini et al., 2011; Miola, 2012; van Geel et al.,

242 2011). The NPP type numbers follow the convention of ‘HdV-number’ and ‘UG-number’, in which
243 acronym ‘HdV’ means Hugo de Vries Laboratory of the University of Amsterdam (The Netherlands),
244 whereas ‘UG’ is Universiteit Gent (Belgium) (Miola, 2012). Percentages of pollen grains originating
245 from forest and savannah communities were calculated as the ratio of an individual taxon and the TPS
246 (total pollen sum); the TPS consists of the sum of AP (arboreal pollen) and NAP (non-arboreal pollen
247 but excludes any taxa originating from aquatic and wetland plants as well as spores and NPPs).
248 Percentages of aquatic and wetland pollen taxa, as well as spores and NPPs, were calculated as the ratio
249 of an individual taxon or NPP type and the TPS enlarged by this taxon or NPP type.

250 Microscopic charcoal particles (size range 0.02-0.5 mm) were counted on the same microscopic slides
251 as the ones used for pollen counting until the total number of charcoal particles and *Lycopodium* spore
252 standard was at least 200 in each sample (Finsinger and Tinner, 2005). Values are expressed as the
253 charcoal accumulation rate (CHAR_{micro}) in grains cm⁻² year⁻¹ and were calculated based on the
254 following formula proposed by Davis and Deevey Jr.(1964):

$$255 \quad \text{CHAR}_{\text{micro}} = \text{CHAC}_{\text{micro}} \times \text{SAR} \quad (2)$$

256 where CHAC_{micro} is the concentration of microscopic charcoal particles (in grains or particles cm⁻³) and
257 SAR is the sediment accumulation rate (in cm year⁻¹). The diagrams were prepared using the POLPAL
258 software (Nalepka and Walanus, 2003) and the analyses were performed at the Adam Mickiewicz
259 University in Poznań (Poland).

260

261 3.5. Quantitative palynofacies analysis

262 Sample preparation included treatment with 50 ml of HF overnight at 40°C followed by 50 ml of HCl
263 for 30 min to eliminate siliciclastic and carbonate minerals. Samples were then rinsed until neutral pH.
264 Quantification of the different types of organic matter (OM) is based on the methodology developed by
265 Graz et al. (2010) using incorporation of a standard solution of *Cupressus* pollen at 10 mg.ml⁻¹ in each
266 sample. Samples were prepared in thin sections before optical investigations using a transmitted light
267 microscope with a 50× magnification. Particle identification was performed using the methodology

268 described by Boussafir et al. (2012) and Graz et al. (2010). Particles were classified into 3 types: (i)
269 Ligno-Cellulosic tissues (LC) at different stages of degradation comprising fresh tissues detected as
270 translucent LC (tLC), slightly degraded/amorphised LC (saLC) with cell structures that are still partially
271 recognizable and totally degraded/amorphous LC, characterised by red aggregates of amorphous OM
272 (rAOM) as originally described by Graz et al. (2010), (ii) mycelium fragments (myc; Graz et al., 2010)
273 and (iii) planktonic remains (algal organic matter, algOM; Boussafir et al., 2012). Whenever necessary,
274 particle identification was aided using UV excitation. The total mass of each particle type was
275 determined after 40 counting according to Eq. 3:

$$276 \quad m_{part} = \frac{m_{std} \times A_{part}}{A_{std}} \times \frac{d_{part}}{d_{std}} \quad (3)$$

277 with m_{part} : mass of the particle; m_{std} : mass of the standard; A_{part} : counted surface of the particle; A_{std} :
278 counted surface of the standard; d_{part} : density of the particle; d_{std} : density of the standard. Densities used
279 for the calculation were determined by Graz et al. (2010). Results are expressed as relative abundance
280 (in terms of mass) of each particle to the total. Sample preparation and analysis were performed at the
281 University of Orléans (France).

282

283 3.6. Biomarker analyses

284 3.6.1. Lipid extraction

285 After freeze-drying, samples were submitted to a modified Bligh-Dyer extraction as described in
286 Coffinet et al. (2015). The total lipid extract was then separated into three fractions on a silica column
287 with a succession of (i) DCM, (ii) DCM:acetone (2:1, v:v) and (iii) DCM:MeOH (1:1, v:v) followed by
288 pure MeOH as solvents (Coffinet et al., 2015). The apolar fraction containing the *n*-alkanes and the
289 intermediate polarity one (fraction 2) containing the GDGTs were analysed. Prior to analysis, the apolar
290 fraction was further separated on silver nitrate impregnated silica columns (10%, w:w) in Pasteur
291 pipettes (with heptane and then DCM as eluents) in order to purify the linear *n*-alkanes for compound-
292 specific $\delta^2\text{H}$ analysis.

293 3.6.2. *n*-Alkane analyses

294 *n*-Alkanes were analysed at Sorbonne University (Paris, France) by gas chromatography coupled to a
 295 mass spectrometer (GC-MS) using an Agilent Network 6890 GC System coupled with a 5973 Mass
 296 Selective Detector, with electron impact at 70 eV. 1 µl was injected and separation was achieved using
 297 a Restek RXI-5 Sil MS silica capillary column (30 m × 0.25 mm i.d., 0.50 µm film thickness) with He
 298 as the carrier gas at 1 ml min⁻¹ flow rate. Initial temperature was set at 50 °C and increased to 320 °C at
 299 4 °C min⁻¹. Samples were injected in splitless mode and the injector temperature was 280 °C.

300 The average chain length (ACL; Eq. 4) describes the *n*-alkane distribution profile of a sample and is
 301 used to determine the predominant origin of the *n*-alkanes. Typically, *n*-alkanes with chain with less
 302 than 21 carbon atoms are suggested to be produced by algae and cyanobacteria (e.g. Han et al., 1968)
 303 while *n*-alkanes with more than 25 carbon atoms more likely originate from terrestrial higher plants
 304 (Eglinton and Hamilton, 1967). *n*-Alkanes with chain length between 21 and 25 carbon atoms are
 305 predominantly found in aquatic macrophytes (Ficken et al., 2000).

$$306 \quad ACL = \frac{\sum C_i \times i}{\sum C_i} \quad \text{where } i \text{ spans from 21 to 35} \quad (4)$$

307 The carbon preference index (CPI; Eq. 5) is a ratio assessing the relative importance of odd over even
 308 homologues and reflects the degree of maturity (organic matter degradation) of a sample. Immature
 309 samples have very high CPI (>> 1; Killups and Killups, 2005). Unusually low CPI (below 3) in recent
 310 sediments are generally considered as polluted by a source of mature organic matter (petroleum, wood
 311 burning; Bray and Evans, 1961).

$$312 \quad CPI = 0.5 \times \left(\frac{\sum C_{odd\ 25-33}}{\sum C_{even\ 24-32}} + \frac{\sum C_{odd\ 25-33}}{\sum C_{even\ 26-34}} \right) \quad (5)$$

313 The P_{aq} index (Eq. 6) was developed by Ficken et al. (2000) and is a proxy for the relative contribution
 314 of emergent/terrestrial macrophytes compared to submerged/floating ones. Contribution of submerged
 315 and floating macrophytes is considered as high when P_{aq} values are higher than 0.4 and insignificant
 316 when P_{aq} values are lower than 0.1.

$$317 \quad P_{aq} = \frac{C_{23} + C_{25}}{C_{23} + C_{25} + C_{29} + C_{31}} \quad (6)$$

318 *n*-Alkane hydrogen isotopic composition ($\delta^2\text{H}$ *n*-alkane) was measured at Newcastle University (United-
319 Kingdom) using a Delta V+ isotope-ratio mass spectrometer (IRMS, Thermo Fisher) connected to a GC
320 Ultra Trace (Thermo Fisher), a Finnigan GC Combustion III (Thermo Fisher) and a high temperature
321 conversion (HTC) system set up at 1400 °C. The GC temperature was set to start at 50 °C and then to
322 raise to 250 °C at 15 °C min⁻¹ and from 250 °C to 320 °C at 5 °C min⁻¹. Temperature was then held at
323 320 °C for 15 min. Every sample was analysed in duplicate and the ²H/¹H ratio was reported on the V-
324 SMOW (Vienna standard mean ocean water) scale and expressed in delta per mil (‰). A mixture of
325 *n*-C₁₆ to *n*-C₃₀ alkane standard and 5 α androstane standard (A. Schimmelmann, Indiana University) was
326 run at the beginning and at the end of each sequence. Standard error of the measurements of the
327 individual long chain *n*-alkanes (C₂₃-C₃₁) from this standard mix ranged between 0.3‰ and 1.1‰.

328 3.6.3. GDGT analysis

329 GDGTs were analysed at Sorbonne University (Paris, France) with a high-pressure liquid
330 chromatography coupled to a mass spectrometer with an atmospheric pressure chemical ionization
331 source (HPLC-APCI-MS, Shimadzu LCMS-2020). Separation was achieved with a Prevail Cyano
332 column (2.1 mm x 150 mm, 3 μm ; Alltech, Deerfield, IL, USA) at 30 °C, using a mixture of hexane and
333 isopropanol at 0.2 ml min⁻¹ according to Coffinet et al. (2015). Elution began at 99% A/1% B for 5 min
334 followed by a linear gradient to 98% A/2% B in 45 min. A second linear gradient led to a mixture of
335 90% A/10% B in 10 min, maintained for 10 min and returned to the initial conditions (99% A/1% B) in
336 14 min, maintained for 10 min. Injection volume was 10 μl . Single ion monitoring (SIM) of the [M+H]⁺
337 ions was used to detect the GDGTs.

338 Mean annual air temperatures (MAAT) were estimated using the calibration specifically developed for
339 East African lakes by Loomis et al. (2012):

$$340 \text{MAAT} = 22.77 - 33.58 \times f(\text{III}) - 12.88 \times f(\text{II}) - 418.53 \times f(\text{IIc}) + 86.43 \times f(\text{Ib}) \quad (7)$$

341 where $f(x)$ is the fractional abundance of the compound *x* relative to the total br GDGTs and the roman
342 numerals correspond to the different br GDGT compounds according to Weijers et al. (2007b)
343 numbering.

344 pH was calculated using the soil calibrations developed by Tierney et al. (2010b) and based on the
345 cyclisation ratio of br GDGTs (CBT; Weijers et al., 2007b):

$$346 \quad \text{pH} = 10.32 - 3.03 \times \text{CBT} \quad (8)$$

$$347 \quad \text{CBT} = -\log\left(\frac{[\text{Ib}] + [\text{IIb}]}{[\text{I}] + [\text{II}]}\right) \quad (9)$$

348 where the roman numerals correspond to the different br GDGT compounds according to Weijers et al.
349 (2007b) numbering.

350

351 **4. Results**

352 4.1. Lithology

353 The bottom of the core corresponds to a pumice-rich tephra. The overlaying deposits (ca. 418 to 200
354 cm) consist of an alternating layer of peat, organic gyttja and silty clay material, with charcoal-rich
355 layers (Fig. 2). Above 200 cm, the sediment mostly consists of peat at diverse stages of decay. Large
356 amounts of higher plant macrofossils are observed at depth ranging between 175 and 75 cm, while the
357 decomposition of plant fragments is more pronounced from 75 cm to the top of the sequence. The upper
358 clayey part of the sequence contains two additional tephra layers at 65.5-63 cm and at 57-51 cm (Fig.2).

359

360 4.2. Chronostratigraphy, age-depth model and sedimentation accumulation rate

361 As shown in Table 1 and Figure 2, dates were obtained all along the core and generally fit with the
362 stratigraphy. Nevertheless, two age reversals are observed at the base of the core, based on Total Organic
363 Matter (TOM) measurements (Fig. 2). The first one occurs between 371.5 and 402 cm depth
364 (SacA44028) and the second corresponds to the bottom tephra layer of the core between 417.5 and 420.5
365 cm depth (SacA38523; Table 1). This could be attributed to stratigraphic misplacement or cross-
366 contamination during coring. Therefore, these dates were not included in the age model. All the wood
367 remain-based ages are very close to those from bulk TOM or slightly older (mean standard deviation

368 between the bulk TOM and the wood remain age is +2.3%). This minor mismatch could be due to the
369 additional transport time of the wood remains from the catchment into the marsh sediment while the
370 bulk TOM signature is expected to be predominantly autochthonous (see sections 4.4 to 4.6).
371 Alternatively, “younger” TOM ages could result from microbial activity or organic matter derived from
372 root development occurring after sedimentation (Trumbore, 2009). The $\delta^{13}\text{C}$ values obtained after
373 sample preparation on residual CO_2 vary from -20.0‰ to -29.5‰ along the core (average value -25.3‰;
374 Tab. 1). This range is compatible with organic carbon originating from the vegetation, likely C3 type
375 plants, and consistent with $\delta^{13}\text{C}$ values measured at the nearby Lake Masoko (Gibert et al., 2002).
376 Compared to the mean TOM $\delta^{13}\text{C}$ value (-24.8‰), the mean wood $\delta^{13}\text{C}$ value is lower (-27.2‰) also in
377 agreement with a C3 type plant signature.

378 The investigated sediment sequence from Kyambangunguru shows continuous sedimentation of ca.
379 4080 calibrated years, spanning from ca. 4020 and -60 cal. BP (Fig. 2). The “apparent” sediment
380 accumulation rate (SAR) ranges from 0.03 to 0.18 cm yr^{-1} (i.e., each cm thick sample of the profile
381 records 6.3 to 33 years). The highest SAR was recorded in the bottom section of the profile (between
382 0.10 and 0.18 cm yr^{-1}), in the 417-191 cm section (from 4015 to 2280 cal. BP). Intermediate SAR values,
383 ranging between 0.07 and 0.11 cm yr^{-1} , were recorded between 191 and 8 cm (from 2280 to 130 cal.
384 BP), whereas the lowest SAR (0.03-0.06 cm yr^{-1}) was recorded in the top section of the profile between
385 8 and 0 cm (from 130 to -60 cal. BP). The σ error of the modelled dates ranged between 0.2 (top of the
386 profile) and 53 years.

387

388 4.3. Palynofacies

389 The Kyambangunguru core is largely dominated (ca. 80% of the OM throughout the core) by ligno-
390 cellulosic tissues (LC) from vascular plants (Fig. 3). Microscopic investigation of the LC reveals 3 stages
391 of degradation. On average, fresh, well-preserved, tLC tissues represent 19% of the total OM (Fig. 3),
392 while slightly degraded, saLC particles are the most abundant type of OM (ca. 39% on average) in the
393 core. Amorphous rAOM represents ca. 11% of the total LC OM with an increase to 20% in the section
394 from ca. 3140 to 830 cal. BP (300-62.5 cm). In addition to LC, fungal mycelia are observed in the layer

395 spanning the period between 239 and 60 cm (ca. 2670 to 800 cal. BP). Grey cell fragments and granular
396 amorphous OM, related to planktonic-derived material, are found throughout the core (Fig. 3), but are
397 particularly abundant at the base (below 330 cm, i.e. ca. 3360 cal. BP) and at the top of the profile (from
398 89.5 to 0 cm, ca. 1150 cal. BP to modern).

399

400 4.4. Plant macrofossil analysis

401 Plant macrofossils are dominated by macrophytes (sedges, submerged/floating plants and algae) and
402 wood remains. *Eleocharis* sp. is present all along the core, while *Nymphaea* sp. (floating macrophyte),
403 disappeared between 160 and 100 cm (ca. 1960 and 1270 cal. BP; Fig. 4). In addition, in the layer
404 between 415.5 and 281.5 cm (ca. 4000 to 3000 cal. BP), macrofossils are composed of remains of algae
405 (*Nitella* sp., Fig. 4) and submerged/floating plants (*Potamogeton* sp. and *Caldesia parnassiflora*; Fig.
406 4). These species are replaced by *Juncus* sp. from 239.5 to 139.5 cm (ca. 2670 to 1730 cal. BP). During
407 this interval, there is a relative increase in abundance of wood remains in comparison to the other
408 sections of the core. *Nitella* sp. together with *Carex* sp. and *Chara* sp. are the dominant remains found
409 in the sequence from 139.5 to the surface (1730 cal. BP to modern; Fig. 4). At 105 cm (ca. 1330 cal.
410 BP), *Aldrovanda vesiculosa* seed was found.

411

412 4.5. Pollen, NPPs and microcharcoal

413 265 taxa of pollen and spores as well as non-pollen palynomorphs (NPPs) – organic walled microfossils
414 which are not pollen – were identified by palynological analysis. The results of this analysis were
415 separated into two groups: indicators of terrestrial vegetation, i.e. vegetation surrounding the lake/marsh
416 and indicators of aquatic vegetation reflecting mainly lake-marsh vegetation. Three pollen zones
417 reflecting compositional changes in terrestrial (woodland and open land) communities (labelled as
418 KY/tp-1 to -3 with tp: terrestrial pollen; Fig. 5) were defined. Additionally, the KY/tp-1 zone was
419 subdivided into three subzones (KY/tp-1a-c). Pollen and NPPs reflecting lake/marsh vegetation changes
420 enabled the establishment of 6 zones (KY/lpn-1-6 with lpn standing for local pollen and NPPs; Fig. 6).

421 4.5.1. Terrestrial vegetation

422 The KY/tp-1 zone (416–303 cm; ca. 4000–3160 cal. BP) is characteristic of the highest percentages of
423 *Acalypha*, Moraceae, *Macaranga* and *Piptadenia/Piptadeniastrum/Entada* (Fig. 5). Moraceae and
424 *Macaranga* pollen reached their maxima (27–28% and 11%, respectively) at 353–333 cm (ca. 3520–
425 3380 cal. BP). Poaceae pollen increased until 397 cm (ca. 3850 cal. BP) when they distinctly decreased
426 while Proteaceae pollen revealed a first maximum at 378 cm (ca. 3700 cal. BP). During the KY/tp-1b
427 subzone (353–333 cm; ca. 3520–3380 cal. BP), Poaceae percentages fell to their minimum values (8%).
428 During the KY/tp-1c subzone (333–303 cm; ca. 3380–3160 cal. BP) pollen values of Moraceae and
429 *Macaranga* dropped simultaneously with the rise in Poaceae percentages and the appearance of palm
430 pollen (*Elaeis guinensis* and *Raphia* type). During the entire KY/tp-1 zone charcoal (0.02–0.5 mm),
431 accumulation rate (CHAR) was high and relatively stable without any distinct maximum (8450–37940
432 particles cm⁻² yr⁻¹; Fig. 5).

433 The KY/tp-2 zone (303–63 cm; ca. 3160–830 cal. BP) was characteristic of the highest values of *Uapaca*
434 (15–24%; 162.5–133.5 cm; ca. 1990–1660 cal. BP) in the profile. However, at 123–113.5 cm (ca. 1540–
435 1430 cal. BP) they substantially declined, simultaneously to a rapid increase in Poaceae values. At the
436 beginning of the zone, *Raphia* pollen increased rapidly (303–283 cm; ca. 3160–3010 cal. BP; up to
437 15.5%). Values of Proteaceae increased from 234.5 cm (ca. 2630 cal. BP) and reached the maxima after
438 113.5 cm (ca. 1430 cal. BP). Simultaneously, or slightly prior to the Poaceae maximum in this zone
439 (123–113.5 cm ca. 1540–1430 cal. BP), values of *Ricinus communis* and *Macaranga* increased (142.5
440 cm; ca. 1760 cal. BP), whereas percentages of *Nauclea* type (123 cm; ca. 1540 cal. BP), *Elaeis guinensis*
441 (143 cm; ca. 1770 cal. BP), *Raphia* type (132.5 cm; ca. 1650 cal. BP) and later Rutaceae (113.5 cm; ca.
442 1430 cal. BP) decreased substantially or disappeared. Among pollen taxa related to arboreal plants
443 occupying montane forest zones, *Apodytes* cf. *dimidiata*, *Olea*, *Podocarpus* and *Prunus africana* type
444 were the most common. During the KY/tp-2 zone, three distinct phases of increase in CHAR values,
445 potentially related to an increase in fire activity, were identified. These occurred at 297.5, 248.5 and
446 117.5 cm (ca. 3120, 2740, and 1480 cal. BP respectively) and were intersected by periods with CHAR

447 values lower than during the KY/tp-1 zone. From 83 cm (ca. 1070 cal. BP), CHAR values started to
448 increase gradually.

449 The KY/tp-3 zone (63–5 cm; 830 to 60 cal. BP) was characterized by the highest percentages of Poaceae
450 (67–90%), the continuous presence of *Ricinus communis* and distinct drops of Proteaceae and *Uapaca*.
451 At 50 cm (ca. 690 cal. BP), *Syzygium* pollen value revealed a substantial increase (up to 16%). In general,
452 CHAR values were higher than in the previous zone and two distinct maxima of their values were
453 recorded at 50 and 10 cm (ca. 690 and 160 cal. BP).

454 4.5.2. Aquatic/marsh vegetation

455 During the KY/lpn-1 zone (416–383 cm; ca. 4000–3740 cal. BP) the algae *Tetraedron trigonum* type,
456 *Tetraedron incus/caudatum* and *Coelastrum reticulatum* reached their maxima (Fig. 6). These algae
457 taxa rapidly declined between 412 and 393 cm (ca 3970 and 3820 cal. BP). Along the KY/lpn-2 zone
458 (383–303 cm; 3740–3160 cal. BP), Alismataceae (cf. *Caldesia*) was regularly observed (0.8–6.5%).
459 Pollen of Nymphaeaceae (*Nymphaea* type) and their epidermis (UG-1241), *Potamogeton* and algae such
460 as *Scenedesmus*, *Tetraedron minimum*, *Pediastrum* undiff., *Pediastrum angulosum* and *Botryococcus*
461 were frequent during these two zones. At 323 cm (ca. 3310 cal. BP), *Nymphaea* type percentages
462 increased simultaneously with a decline in *Potamogeton*. The KY/lpn-3 zone (303–193 cm; ca. 3160–
463 2300 cal. BP) was characterized by an increase in Cyperaceae pollen and tissue fragment percentages
464 (rise in fungal NPP UG-1176 and UG-1197) and a simultaneous drop in *Scenedesmus* and *Tetraedron*
465 *minimum*.

466 The KY/lpn-4 zone (193–113 cm; ca. 2300–1430 cal. BP) was characterized by a prominent increase in
467 Cyperaceae percentages (12–77%). The spores of *Lycopodiella caroliniana* were regularly present (0.2–
468 18%) together with amoeba *Assulina muscorum*. Between 143 and 123 cm (ca. 1770 and 1540 cal. BP),
469 monolet spores reached their maximum values in the core (64–69%). Values of *Nymphaea* type, UG-
470 1241 (*Nymphaea* tissues), *Potamogeton*, Alismataceae (cf. *Caldesia*), *Pediastrum angulosum*,
471 *Botryococcus* and *Scenedesmus* dropped markedly. NPPs of fungal origin increased distinctly, notably
472 UG-1197 and *Entorrhiza* in the deeper part of the zone and HdV-172, UG-1077, UG-1176 and UG-
473 1107 in the upper part.

474 The KY/lpn-5 (113–46.5 cm; ca. 1430–650 cal. BP) and KY/lpn-6 (46.5–5 cm; 650 to 60 cal. BP) zones
475 were characterized by the increase in frequency of algae, mainly *Botryococcus*, *Scenedesmus* and
476 *Pediastrum angulosum* and, among submersed macrophytes, of *Nymphaea* type pollen and their related
477 tissues (UG-1241). In the KY/lpn-5 zone, Hallorrhagaceae appeared (1.5–24%). *Entorrhiza* and fungal
478 type UG-1107 reached maxima in the profile (59–76%). In the KY/lpn-6 zone, Alismataceae (cf.
479 *Caldesia*) and *Potamogeton* became more frequent while Cyperaceae pollen percentages distinctly
480 dropped.

481

482 4.6. Elemental analysis

483 C_{org} content is high throughout the core (39.0 - 57.8%; mean $51.0\% \pm 6.2$; Suppl. Table 1), except in the
484 tephra layer at the base of the core (5%). Total nitrogen (TN) varies from 1.5 to 3.6% except for the
485 tephra layer (0.2%). Slightly lower TN values are observed between 199.5 and 100 cm (ca. 2350 and
486 1270 cal. BP; ca. 2%, Suppl. Table 1) leading to higher C/N ratios that range from 16 to 38 (Fig. 7 and
487 Suppl. Table 1). In the other sections of the core, C/N ratios are relatively invariant at ca. 15.

488

489 4.7. Br GDGT abundance and distribution

490 Br GDGTs are abundant throughout the core (mean $121.5 \pm 81.3 \mu\text{g g}^{-1}$ of dry wt. peat), with maximal
491 concentration observed at 150.5 cm (ca. 1850 cal. BP; Suppl. Table 1). CBT varies between 0.53 and
492 2.01 with a mean value of 1.31 and is higher above 180.5 cm (ca. 2180 cal. BP; 1.41-2.01; Suppl. Table
493 1). MBT is comprised between 0.65 and 0.89, with a mean value of 0.77 (Suppl. Table 1). The highest
494 MBT values (>0.80) are found between 194 and 45 cm (ca. 2310 and 630 cal. BP).

495

496 4.8. *n*-Alkane distribution and their $\delta^2\text{H}$ composition

497 Mid- to long-chain *n*-alkanes ($>C_{21}$) dominate most of the samples, with C_{23} and C_{25} being the most
498 abundant (on average 23% and 21%, respectively; Suppl. Fig. 1). Two *n*-alkane distribution patterns are

499 observed in the core (Suppl. Fig. 1), defined as patterns A and B. Pattern A was identified in the sections
500 from 417 to 180 cm and from 30 cm to the surface (ca. 3670 -2300 cal. BP and ca. 430 cal. BP-modern
501 respectively). It is dominated by odd numbered *n*-alkanes, the most abundant one being C₂₃, with a
502 decreasing trend in relative abundance from C₂₃ to C₃₁. Pattern B (Suppl. Fig. 1) was observed in the
503 sections from 193 to 29.5 cm (ca. 2300 to 430 cal. BP). This distribution is characterized by a flattening
504 of the *n*-alkane profile. The C₂₉⁺ *n*-alkanes (up to C₃₅ in some samples) as well as the C₂₃⁻ *n*-alkanes,
505 especially C₁₉, increase while the C₂₃, C₂₅ and C₂₇ decrease, in comparison to pattern A. In pattern B,
506 even numbered *n*-alkanes are also found in larger amount than in pattern A.

507 The CPI index confirms a strong odd-over-even predominance throughout the core (mean 8.1 ± 3 ; Suppl.
508 Table 2) while the ACL index varies between ca. 25 and 27 (Fig. 7; Suppl. Table 2). The P_{aq} index is
509 systematically higher than 0.4, ranging between 0.50 and 0.85 (Fig. 7; Suppl. Table 2).

510 The δ²H values of odd mid to long chain *n*-alkanes (C₂₃-C₃₁) varies between -96‰ and -172‰ (Suppl.
511 Table 2) and are higher in the deepest part of the core, between 417 and 391 cm (ca. 4010 and 3800 cal.
512 BP; between -96‰ and -122‰; Suppl. Table 2). Long chain *n*-alkanes (C₂₉ and C₃₁) are more ²H-
513 enriched (-130‰ on average; Suppl. Table 2) than mid chain (C₂₃ and C₂₅; -150‰ on average; Suppl.
514 Table 2) compounds.

515

516 **5. Discussion**

517 The combination of palaeobotanical, elemental and molecular analyses enabled the reconstruction of the
518 Holocene ecological history of the Kyambangunguru wetland together with regional climate. The proxy
519 analyses revealed the presence of three major phases in the Kyambangunguru wetland ecosystem
520 development over the last 4.0 ka cal. BP for which the interaction with regional and global climatic
521 changes are discussed. The chronology of the sediment record was compared to published records of
522 the region (Filippi and Talbot, 2005; Fontijn et al., 2012; Garcin et al., 2007) showing overall good
523 consistency. The tephra layer at the base of the core, dated at ca. 4.1 ka cal. BP, corresponds to the
524 Rungwe Pumice deposits identified at Lakes Masoko and Malawi and dated at 4.3 ka cal. BP and 4.3 to

525 3.6 ka cal. BP, respectively. The second tephra observed at ca. 0.9 ka cal. BP in the Kyambangunguru
526 core can be related to the Aphyric Pumice deposits observed at 1.2 ka cal. BP in the Lake Masoko
527 sediment record and between 1.1 and 0.6 ka cal. BP in the Lake Malawi one. The last tephra layer
528 identified at Kyambangunguru is harder to define. It is dated at ca. 0.8 ka cal. BP and could either
529 correspond to the Aphyric Pumice described earlier or to the Ngozi Tuff, even though this event was
530 dated between 0.5 and 0.3 ka cal. BP at Lakes Masoko and Malawi. This regional comparison, even if
531 it stresses its inherent time uncertainty, further validates the age-model used in this study.

532

533 5.1. Ecosystem, air temperature and precipitation variability at the Kyambangunguru crater marsh

534 Three units (labelled as Units I-III; Fig. 7), representing the major developmental phases of marshland
535 ecosystem over the last 4.0 ka cal. BP, were distinguished based on the local plant community changes
536 inferred from pollen, non-pollen palynomorph and plant macro-fossil analyses. In addition, the origin
537 of the OM and its degree of preservation were determined using the palynofacies and C/N records
538 (Bourdon et al., 2000) while the distribution of br GDGTs allowed reconstructing the pH of the site,
539 following the approach described by Weijers et al. (2007b) and Tierney et al. (2010b). For the climatic
540 reconstruction, the mean annual air temperature (MAAT) and the hydrological conditions over the last
541 4.0 ka cal. BP years were reconstructed using the br GDGT distribution and the hydrogen isotopic values
542 of the *n*-alkanes ($\delta^2\text{H}_{\text{wax}}$), respectively. Br GDGT-derived MAAT were calculated with the East African
543 lacustrine calibration developed by Loomis et al. (2012), i.e. Eq. 7, which takes into account *in situ*
544 production of br GDGTs in lakes, as discussed in the Suppl. Information. The $\delta^2\text{H}_{\text{wax}}$ was determined as
545 the weighted average of the $\delta^2\text{H}$ values of the C₂₃ and the C₂₅ *n*-alkanes which are expected to have
546 recorded mainly the wetland water isotopic composition ($\delta^2\text{H}\text{-H}_2\text{O}$) variations during the accumulation
547 of the sedimentary sequence (as discussed in the Suppl. Information). Because the study site has no
548 outflows, the $\delta^2\text{H}\text{-H}_2\text{O}$ variations depend essentially on the rates of precipitation and evaporation and
549 the $\delta^2\text{H}_{\text{wax}}$ can be interpreted as changes in the precipitation to evaporation ratio (P/E ratio; Gonfiantini,
550 1986; Sachse et al., 2004).

551

552 5.1.1. Unit I: ca. 4.0 – 2.3 ka cal. BP (417–193 cm) – shallow lake, cold and wet conditions

553 The palaeoenvironmental proxies suggest a relatively stable and persistent lake environment during this
554 period. Planktonic remains are observed in the palynofacies (Fig. 3) and in the non-pollen palynomorph
555 profile (NPP; Fig. 6) and the C/N values (ca. 15; Fig. 7) indicate a mixed aquatic and terrestrial source
556 of the OM (Meyers, 1997). The GDGT derived-pH of ca. 7 (Fig. 7) is consistent with current pH values
557 of the RVP crater lakes (ranging between 5.8 and 8.7; Delalande, 2008), further supporting the
558 occurrence of a lake at this time. The high abundance of macro- and microremains (Figs. 4 and 6) from
559 algae (*Nitella* sp.) and submerged/floating macrophytes (*Potamogeton* sp. and *Nymphaea* sp.) and the
560 relatively high P_{aq} values (0.7-0.8; Ficken et al., 2000) suggest that the lake was relatively shallow and
561 its water column likely harboured abundant macrophyte vegetation. The growth of *Caldesia*
562 *parnassifolia* for instance may indicate a water depth lower than one meter (Gupta and Beentje, 2017;
563 Sinkevičienė, 2016). In addition, the NPP record reveals a characteristic pattern of successive and rapid
564 decline of *Tetraedron trigonum* type, *Coelastrum reticulatum* and *Tetraedron incus/caudatus*, all before
565 3.8 ka cal. BP (Fig. 6). This reduction is possibly related to the spread of *Nymphaea* whose floating
566 leaves may have limited the light availability, although it did not seem to impact *Tetraedron minimum*
567 and *Scenedesmus*, which might be more resistant species. The second stage of microalgal community
568 retreat, which affected *Scenedesmus* and *Tetraedron minimum*, occurred at ca. 3.2 ka cal. BP and seem
569 to have also been stimulated by *Nymphaea* expansion. Simultaneously, an increased input of fungal
570 remains, UG-1176 and UG-1197, may indicate spreading of emerged marsh plants around the lake. This
571 is further supported by the high C_{org} values (> 50%; Suppl. Table 1) and the predominance of
572 lignocellulosic-derived OM in the palynofacies record partly originating from *Cyperus* (Laggoun-
573 Défarge et al., 2008a) according to microscopic observation (Fig. 3).

574 Unit I is marked by an abrupt decrease in mean annual air temperature (MAAT) of about three degrees
575 (from 26 °C to 23 °C, Fig. 8) at ca. 3.9 ka cal. BP. The MAAT continues to decrease by another three
576 degrees to 20 °C at 2.7 ka cal. BP (Fig. 8). This trend is synchronous with the “3.3–2.5 ka. BP” cold
577 Holocene event introduced by Wanner et al. (2011) which was also observed – to a lower extent – at ca.

578 4.0 – 3.0 ka cal. BP in Lakes Tanganyika and Malawi (Powers et al., 2005; Tierney et al., 2008; Fig. 8)
579 and in Lakes Turkana and Challa (Berke et al., 2012;
580 Sinninghe Damsté et al., 2012). At the same time, the $\delta^2\text{H}_{\text{wax}}$ record shows more negative values until
581 ca. 3.0 ka cal. BP (up to -168 ‰; Fig. 9), which can be interpreted as an increase in the P/E ratio either
582 due to higher precipitation or/and lower evaporation rates. A wetter environment is also inferred from
583 the terrestrial pollens, with a decrease in pollen supply from grassland communities from ca. 3.7 to 3.4
584 ka cal. BP and an expansion of mountain forest and Zambezian Miombo woodland communities
585 (Moraceae and *Macaranga* at ca. 3.5 – 3.4 ka cal. BP, *Uapaca* optimum, presence of *Apodytes cf.*
586 *dymidata*, *Olea*, *Podocarpus* and *Prunus africana*). Wetter conditions at the same period were also
587 inferred from a pollen record at Lake Masoko (Vincens et al., 2003). Notably, a synchronous *Uapaca*
588 pollen optimum occurred both at Masoko and Kyambangunguru (after ca. 3.1 ka cal. BP, Vincens et al.,
589 2003). $\delta^2\text{H}_{\text{wax}}$ values remain negative during unit I, although two positive peaks in $\delta^2\text{H}_{\text{wax}}$ are noticeable
590 at ca. 3.6 and 2.8 ka cal. BP possibly indicating brief dry events interrupting the overall wetter period.
591 These dry events are in agreement with observations by Russell et al. (2003) and Russell and Johnson
592 (2005) at Lake Edward (Uganda; Fig. 9). Peaks in Mg content in calcite at 3.6 and 2.8 ka cal. BP were
593 indeed linked to evaporative concentration of the lake (increase in the $[\text{Mg}^{2+}]/[\text{Ca}^{2+}]$ ratio in lake water)
594 in response to short but pronounced drought events.

595 *5.1.2. Unit II: 2.3 – 1.4 ka cal. BP (193–113 cm) – marsh/peatland formation under sustained warm*
596 *and drier conditions*

597 An abrupt, environmental change occurred at ca. 2.3 ka cal. BP leading to the establishment of a peatland
598 with variable hydrological conditions (Fig. 7). The onset of unit II is marked by an apparent decrease in
599 the sedimentation rate from 0.13 cm yr⁻¹ to 0.08 cm yr⁻¹ (Fig. 2), an increase in the Cyperaceae tissue
600 and ferns (monolete spores) accumulation and a strong increase of the C/N ratio (up to 40), all
601 evidencing peat soil formation. High C/N ratios (usually > 30 and sometimes up to 100) are indeed
602 typical for peatlands, due to the high preservation of the OM in relation to the prevailing acidic and
603 anoxic conditions (Laggoun-Défarge et al., 2008b; Meyers, 1997). Simultaneously, a substantial
604 decrease or almost total disappearance of algae (*Scenedesmus*, *Pediastrum*, *Botryococcus* and

605 *Tetraedron minimum*) and aquatic plants such as *Nymphaea* spp. (Fig. 6) reflect a substantial fall in the
606 water table. However, irregular appearances of these taxa, as well as of the macroalgae *Nitella* sp.,
607 indicate episodic inundation of the wetland (Fig. 4). The main constituents of the peatland community
608 were *Juncus* sp., *Eleocharis* sp., ferns and perhaps other Cyperaceae species (Figs. 4 and 6). Permanent
609 waterlogged conditions, and low pH (down to 5.2, Fig. 7) contributed to the establishment of plant taxa
610 restricted to humid acidic environments, such as *Drosera*, *Lycopodiella caroliniana* and *Aldrovanda*
611 *vesiculosa* (Figs. 4 and 6; (Gałka et al., 2015), and the appearance of protists frequently found in
612 peatlands, such as the testate amoeba *Assulina muscorum* (van Geel, 1978). The increase in reddish
613 amorphous OM (rAOM) and the appearance of fungal mycelia at that time (Fig. 3) also suggest the start
614 of a terrestrialization process and the development of vascular plants within the marsh (Bourdon et al.,
615 2000). The increase in wood remains in the macrofossil profile, related to the development of trees and
616 shrubs at the edges of the marsh or even in the marsh itself, supports this interpretation.

617 Unit II displays the highest $\delta^2\text{H}_{\text{wax}}$ values of the core (up to -136 ‰; Fig. 9), suggesting low P/E ratio
618 and thus the driest period of the last 4.0 ka cal. BP, consistent with several East-African records – Lake
619 Edward (Uganda), Lake Turkana (Kenya), Lake Tanganyika (Tanzania) (Nash et al., 2016 and
620 references therein). A major drought event is widely described at ca. 2.0 ka cal. BP, followed by a second
621 period between ca. 1.7 and 1.0 ka cal. BP, during which successive minor drought events occurred (Alin
622 and Cohen, 2003; Russell et al., 2007; Russell and Johnson, 2005; Verschuren and Charman, 2008). In
623 the Kyambangunguru $\delta^2\text{H}_{\text{wax}}$ record, a positive excursion that could be interpreted as a drought event is
624 observed at ca. 2.2 ka cal. BP. However, the major dry period seems to have occurred later, between ca.
625 1.7 and 1.4 ka cal. BP (centred at 1.5 ka cal. BP, Fig. 9). Around the same time (ca. 1.7 ka cal. BP),
626 *Uapaca* pollen suddenly declined and fire activity increased, which contributed to the opening of the
627 woodland canopy and a spread of the grassland communities (Fig. 5). The same abrupt changes are
628 noticeable in charcoal and pollen data at Lake Masoko (Thevenon et al., 2003; Vincens et al., 2003) and
629 in the Amboseli basin in Kenya (Rucina et al., 2010). Alternatively, these vegetation changes could have
630 been induced by enhanced human activity. Indeed, a first spread of human settlement is considered to
631 have occurred during the Late Iron Age, around 1.5 ka cal. BP (Marchant and Taylor, 1998; Vincens et
632 al., 2003). However, several sites in East Africa (Lakes Turkana, Tanganyika, Naivasha, Challa, Edward

633 and Sacred Lake; see references in Marchant et al., 2018) support a widespread increase in aridity as the
634 main cause explaining the sedimentological evidence of droughts, as discussed by Marchant et al.
635 (2018). Notably, we promote that $\delta^2\text{H}_{\text{wax}}$ records, such as this study and Konecky et al. (2014), could
636 help disentangling the human impact from the climatic one as they should only be marginally impacted
637 by human activities. According to the br GDGT data, MAAT remained high during unit II, oscillating
638 around 21.5-22 °C (Fig. 8), suggesting a dry and warm period during unit II. During the same period,
639 higher temperatures are also reported in the records from Lake Challa (Sinninghe Damsté et al., 2012),
640 Lake Turkana (Berke et al., 2012) and Lake Tanganyika (Tierney et al., 2008) (Fig. 8).

641 *5.1.3. Unit III: ca. 1.4 ka cal. BP – modern (113–0 cm) – periodically flooded marsh and*
642 *transition to colder conditions*

643 At the onset of Unit III, an increase in the water table led to the reappearance of microalgae assemblages
644 (mainly composed of *Botryococcus*, *Scenedesmus* and *Pediastrum angulosum*), and macrophytes such
645 as *Nymphaea*, Alismataceae and Halorrhagaceae at ca. 1.4 ka cal. BP. However, until ca. 0.7 ka cal.
646 BP, high percentages of Cyperaceae and of *Entorrhiza* spores are observed, which indicate that the
647 peatland was subjected to pronounced water table fluctuations but without open water stages. Indeed,
648 *Entorrhiza* is a genus of parasites that infect the roots of the Juncaceae (rush) and Cyperaceae (sedge)
649 families when they are no longer in water (Riess et al., 2015; Vánky, 1998). At ca. 0.7 ka cal. BP, the
650 water table likely increased substantially, which supported a spread of *Nymphaea* and contributed to the
651 sharp decrease in Cyperaceae. Since ca. 0.4 ka cal. BP, *Tetraedron minimum* and *Scenedesmus* blooms
652 became more frequent, and the structure of the microalgal communities resembled the one from the base
653 of the core (start of Unit I). This time, however, large amounts of *Carex* remains and large fluctuations
654 in pH (from 5.2 to 7.8; Fig. 7) suggest the coexistence of peat patches, likely acidic, and water
655 depressions where the pH may have been higher, supporting the development of microalgal
656 communities. This corresponds to the current status of the marsh. Continuous monitoring of the site
657 during the last decade indicates the presence of a seasonally evolving ecosystem, i.e. (i) a shallow lake
658 colonized by large sedge mats during the rainy season and (ii) a waterlogged marsh-like regime at the
659 end of the dry season due to high evaporation rates (Delalande et al., 2008a, 2008b).

660 The period that corresponds to unit III at Kyambangunguru starts with relatively warm mean annual air
661 temperatures, at ca. 22 °C, followed by an abrupt cooling to 20 °C at ca. 0.5 ka cal. BP (Fig. 8). A similar
662 abrupt event was also identified at Lake Tanganyika (Tierney et al., 2010a; Fig. 8), in the Ethiopian
663 Highlands and in southern Africa (Nicholson et al., 2013 and references therein) and could coincide
664 with the transition between a “Medieval Climate Anomaly” (MCA; based on Jones et al. (2001) time
665 boundaries: ca. 1.0 – 0.8 ka cal. BP) and a Little Ice Age (LIA; 0.7 – 0.1 ka cal. BP; Matthews and
666 Briffa, 2005) in the African continent. The occurrence of climatic shifts in Africa that could be related
667 to the European MCA and LIA events is currently under debate but a growing body of research seems
668 to support such a cross-latitude connection (e.g. Lüning et al., 2018; Russell and Johnson, 2007; Tierney
669 et al., 2013). At Kyambangunguru, the last 500 yrs exhibit the coldest temperatures of the record which
670 would support the existence of a “Little Ice Age” equivalent in East Africa, in agreement with records
671 from nearby Lake Malawi (Branchu et al., 2010; Powers et al., 2011; Fig. 8). At the transition between
672 units II and III (ca. 1.4 – 0.8 ka cal. BP) the $\delta^2\text{H}_{\text{wax}}$ values decrease, consistent with the reestablishment
673 of higher P/E ratio under wetter conditions and the spreading of Proteaceae and the recovery of the
674 *Uapaca* woodlands, as observed in the pollen record. At 0.8 ka cal. BP, the canopy density started to
675 decrease, the fire activity increased, which led to a spread in the grassland communities, maintained
676 until nowadays. Similar conditions were recorded at Lake Masoko (Vincens et al., 2003). Intensification
677 of human activity in the region could be responsible for such vegetation changes, especially as no shift
678 is observed in the $\delta^2\text{H}_{\text{wax}}$ values during this period. Additional high-resolution reconstructions in the
679 RVP and the neighbouring provinces are necessary to better assess the relative contribution of human
680 versus climatic nature of these recent environmental shifts.

681

682 5.2. Implications for climate dynamics in East Africa during the late Holocene and its impact on 683 highland wetland ecology

684 The temperature and P/E ratio records at Kyambangunguru show a high variability over the last 4.0 ka
685 cal. BP. The main two warm periods, before ca. 3.7 ka cal. BP and between ca. 2.2 and 1.0 ka cal. BP,
686 were generally accompanied by drier conditions (Figs. 8 and 9). A major dry event at 4.2 ka cal. BP is

687 a common feature in many tropical and temperate records and was interpreted as a southward migration
688 of the ITCZ (e.g. Gasse, 2000; Mayewski et al., 2004). A severe drought has also been observed in many
689 other East African sites around 2.0 ka cal. BP (Marchant et al., 2018) and could have been related to the
690 dry and warm period centred at ca. 1.5 ka cal. BP at Kyambangunguru. The discrepancy in timing
691 between the RVP and the other tropical East African records may be related to the unique location of
692 the RVP at the southern end of the tropical climatic belt ($\approx 10^\circ\text{S}$), i.e. at the transition between two
693 hydro-climatic regimes (the north-eastern and the southern ones). Two cold and wet periods were
694 identified, between ca. 3.3 and 2.0 ka cal. BP and since ca. 0.6 ka cal. BP. The latter period is concurrent
695 with the European Little Ice Age and suggests that this cold event may have occurred globally (e.g.
696 Brown and Johnson, 2005). The warm and dry period at ca. 1.5 ka cal. BP and the two cold and wet
697 events all correspond to cold periods identified by Wanner et al. (2011) supporting the idea that the
698 Holocene climatic variability is at least partly driven by global scale events (e.g. solar and volcanic
699 activities, changes in the thermohaline circulation). However, it emphasizes that these processes seemed
700 to have contrasting effects at different latitudes (cooling vs. warming).

701 A major ecological shift that led to the transition from a shallow lake to a peatland at Kyambangunguru
702 started at ca. 2.5 ka cal. BP, in a rather wet environment, 300 years before the $\delta^2\text{H}_{\text{wax}}$ exhibit an abrupt
703 shift towards drier conditions (Fig 9). This time offset suggests that the ecological change recorded at
704 Kyambangunguru was primarily due to a hydroseral succession, i.e. the natural progressive colonisation
705 and infilling of a freshwater lake by different types of macrophytes leading to its transition to a swamp
706 which could eventually turn into a forest (Charman, 2002). A drier peatland was established at ca. 2.2
707 ka cal. BP and sustained until ca. 1.4 ka cal. BP likely as a result of combined warm and dry conditions,
708 which significantly lowered the P/E ratio of the wetland (Fig. 8 and 9). This is consistent with the abrupt
709 change at ca. 2.2 ka cal. BP observed in the biotic communities: planktonic and *Nymphaea* communities
710 disappeared, while *Cyperus* spp. took over in only a century (Fig. 6). This behaviour suggests a threshold
711 effect, which may reflect an autogenic hydroseral development triggered by gradual warming
712 conditions. Previous studies have shown the ability of *Nymphaea* spp. to form mats floating at the
713 surface of the water (Charman, 2002; Ellery et al., 1990) and pointed it out as a common step in the
714 hydroseral succession leading to the conversion of a freshwater body to a swamp (Kratz and DeWitt,

1986; Swan and Gill, 1970). The development of floating mats could be accelerated in the tropics, as the warm air temperatures hold up high vegetation productivity rates (Talling et al., 1998). The slight warming recorded at ca. 2.6 ka cal. BP (by the br GDGTs; Fig. 8) could then have acted as a positive feedback in promoting high rates of vegetation productivity. The consequences of this fast development of floating mats are (i) large inputs of OM in the sediment, which progressively infilled the lake, (ii) a vertical expansion of the mats by peat accumulation, and (iii) a horizontal expansion and partial covering of the surface of the lake. From ca. 3.3 to 2.3 ka cal. BP (Fig. 6), opposite trends are observed in the distribution of plankton and *Nymphaea* suggesting that the more *Nymphaea* mats would cover the lake surface, the less light would penetrate the water column leading to a decrease in microalgae abundance, supporting the proposed mechanism of lake infilling.

This terrestrialization process was interrupted at ca. 1.4 ka cal. BP, when pronounced fluctuations in the water table are revealed by the presence of *Enthoriza* in the microfossil record (ca. 1.4 – 0.7 ka cal. BP; Fig. 6). The water balance, recorded by the $\delta^2\text{H}_{\text{wax}}$, increased over this period, suggesting a relative increase in precipitation (Fig. 9) in agreement with the spread of Proteaceae and *Uapaca* in the terrestrial pollen. This likely re-flooded the marsh and created patches of open water, where phytoplankton, and then *Nymphaea*, could re-colonise the marsh (Fig. 6). A similar ecosystem change was encountered at the same period in a mountainous marsh in Madagascar (Bourdon et al., 2000). This period of water fluctuations, synchronous with the Medieval Climate Anomaly (MCA; 1.0 – 0.8 ka cal. BP), is the only period of the investigating 4000 yr record when conditions were warm and wet at Kyambangunguru, suggesting that its origin may be different from the other recorded climatic changes. The MCA is described as warm and dry in most north-eastern Africa (Lüning et al., 2017; Nash et al., 2016; Nicholson et al., 2013) and humid in southern Africa (e.g. Nash et al., 2016; Tyson and Lindsay, 1992; Woodborne et al., 2015). A recent literature review by Lüning et al. (2018) suggests a transition zone across Tanzania with increased signs of humidity along a NE-SW transect. Notably, Buckles et al. (2016) and Finch et al. (2017) recorded wet conditions at Lake Challa until ca. 800 BP and at the Kwasebuge peat bog until ca. 675 BP, respectively. Hence the Kyambangunguru record presented here extends this transitional zone to the southeast, in agreement with the hydroclimatic interpretation of a biogenic silica record by Johnson et al. (2004) at Lake Malawi.

743

744 **6. Conclusions**

745 The detailed multi-proxy analysis of a 4-m peat core covering the late Holocene (4000 years) reveals
746 rapid and profound ecological changes of the Kyambangunguru wetland in the Rungwe Volcanic
747 Province (RVP), southwestern Tanzania. Around 2.2 ka cal. BP, a shallow crater lake turned into a
748 peatland. Starting at ca. 0.9 ka cal. BP the water level in the marsh increased, creating a shallow lake
749 during the rainy season and a peaty marsh during the dry season. These significant ecological
750 fluctuations correlated with major changes of the Late Holocene East African climate. Notably, the air
751 temperatures remained high and the reconstructed precipitation low between ca. 2.2 and 0.9 ka cal. BP,
752 which allowed the peatland terrestrialization to sustain. This study represents the first detailed late
753 Holocene quantitative air temperature reconstruction from the RVP region. We identified a succession
754 of cold/warm/cold events, largely in phase with the other regional East African climate records and with
755 the cold periods identified worldwide by Wanner et al. (2011). This further supports that global scale
756 processes may be the main drivers of the Holocene climatic variability. Moreover, warm conditions
757 during the MCA followed by abrupt cooling during the LIA were observed at Kyambangunguru and
758 elsewhere in East Africa suggesting that these two recent events occurred globally. The precipitation
759 pattern at Kyambangunguru during these two events is opposite to most of the more north-eastern
760 African records and rather resembles the southern African climatic records. Recent additional
761 precipitation records in Tanzania also show such a pattern. Tanzania seemed thus to be located at the
762 transition zone between two hydro-climatic poles (north-eastern and southern Africa) and to have
763 experienced a variable relative contribution of these two poles over the last millennium. This study
764 further demonstrates that peatlands and marshes provide valuable, high resolution climatic archives in
765 the tropics that offer novel avenues of research for understanding linkages between the Holocene climate
766 variability and ecosystem change in the tropics.

767

768

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780

781 **Figure and Table captions:**

782 Figure 1: Regional setting and climatic conditions of the studied site. A: geographical location of the
783 Rungwe Volcanic Province (RVP, adapted from Delalande, 2008), the green lines outline the maximal
784 positions of the intertropical convergence zone (ITCZ) over the year, the blue lines delimitate the
785 equatorial rain region and the histograms show the monthly mean precipitation values (mm) at the main
786 East African weather stations; B: topographical map of the RVP and location of the Kyambangunguru
787 marsh; C: lake level (black dots), precipitation (blue bars) and marsh water $\delta^2\text{H}$ composition (red dots)
788 variations over one calendar year; D-E: pictures of the Kyambangunguru crater marsh taken in July
789 2014.

790 Figure 2: Lithology (A) and Bayesian age-depth model (B). The A values indicate the agreement
791 between the modelled and the calibrated age and the A_{model} the agreement of the model itself. TOM
792 samples are in black, wood samples in red and tephra ones in grey. Horizontal dotted lines represent the
793 model boundaries (see text for details) and excluded dates are identified as outliers.

794 Figure 3: Quantitative palynofacies: relative abundance of the main organic aggregates analysed by
795 photonic microscopy in transmitted light. In green: ligno-cellulosic tissues (LC): (i) fresh LC detected
796 as translucent LC (tLC) in light green, (ii) slightly degraded/amorphised LC (saLC) in yellow-green and
797 (iii) totally degraded/amorphous LC, characterised by red aggregates of amorphous OM (rAOM) in dark
798 green. In orange: mycelium (myc) fragments. In blue: planktonic remains (algOM).

799 Figure 4: Plant macro-fossil diagram. Results are given in absolute numbers. For the legend of the
800 lithology column see Fig. 2. Roman numbers indicate the three ecological units as defined in the
801 discussion part.

802 Figure 5: Woodland pollen diagram for the Kyambangunguru marsh showing relative percentages of
803 the selected taxa. The grey pattern shows a 10x magnification. For the legend of the lithology column
804 see Fig. 2. Roman numbers indicate the three ecological units as defined in the discussion part.

805 Figure 6: Ferns, aquatic pollen and NPP diagram for the Kyambangunguru marsh showing relative
806 percentages of the selected taxa. The grey pattern shows a 10x magnification. For the legend of the
807 lithology column see Fig. 2. Roman numbers indicate the three ecological units as defined in the
808 discussion part.

809 Figure 7: Lithology (A; see details in Fig. 2), total organic carbon over total nitrogen atomic ratio (B),
810 br GDGT-derived pH (C; based on Tierney et al. (2010b) calibration) and *n*-alkane distribution indices
811 (D-E; ACL and Paq, respectively). The three units discussed in the text are also represented along with
812 their ecological interpretation.

813 Figure 8: (a) Lake Tanganyika TEX₈₆-derived lake surface temperature (LST) of the last 4000 (black
814 contour red dots) and 1500 (red plain dots) years (Tierney et al., 2010a and 2008 respectively) and (b)
815 Lake Malawi TEX₈₆-derived LST of the last 4000 (black contour yellow dots) and 700 (yellow plain
816 dots) years (Powers et al., 2005, 2011) compared to (c) Kyambangunguru br GDGT-derived MAAT
817 (present study). The three units described in the text are also represented (dashed lines).

818 Figure 9: (a) Lake Edward Mg (mol %; Russell and Johnson, 2005) and (b) Kyambangunguru $\delta^2\text{H}_{\text{wax}}$
819 records for the last 4000 years (present study). The three units described in the text are also represented
820 (dashed lines).

821 Table 1: AMS Radiocarbon chronology of the core KYAM12 (2012). Calibrated median age and range
822 (2σ) were obtained using OxCal 4.3 software (Bronk Ramsey and Lee, 2013) with the ShCal 13
823 atmospheric curve (Hogg et al., 2013) and Bomb13SH3 (Hua et al., 2013).

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Highlights

- 4000-year continuous peat record of the Late Holocene in East Africa (SW Tanzania).
- Multi proxy approach evidences major ecological changes at ca. 2500 yr cal. BP.
- Detailed temperature and hydrology records highlight several cold events.
- More recent climate conditions seem connected to South African climate variability.

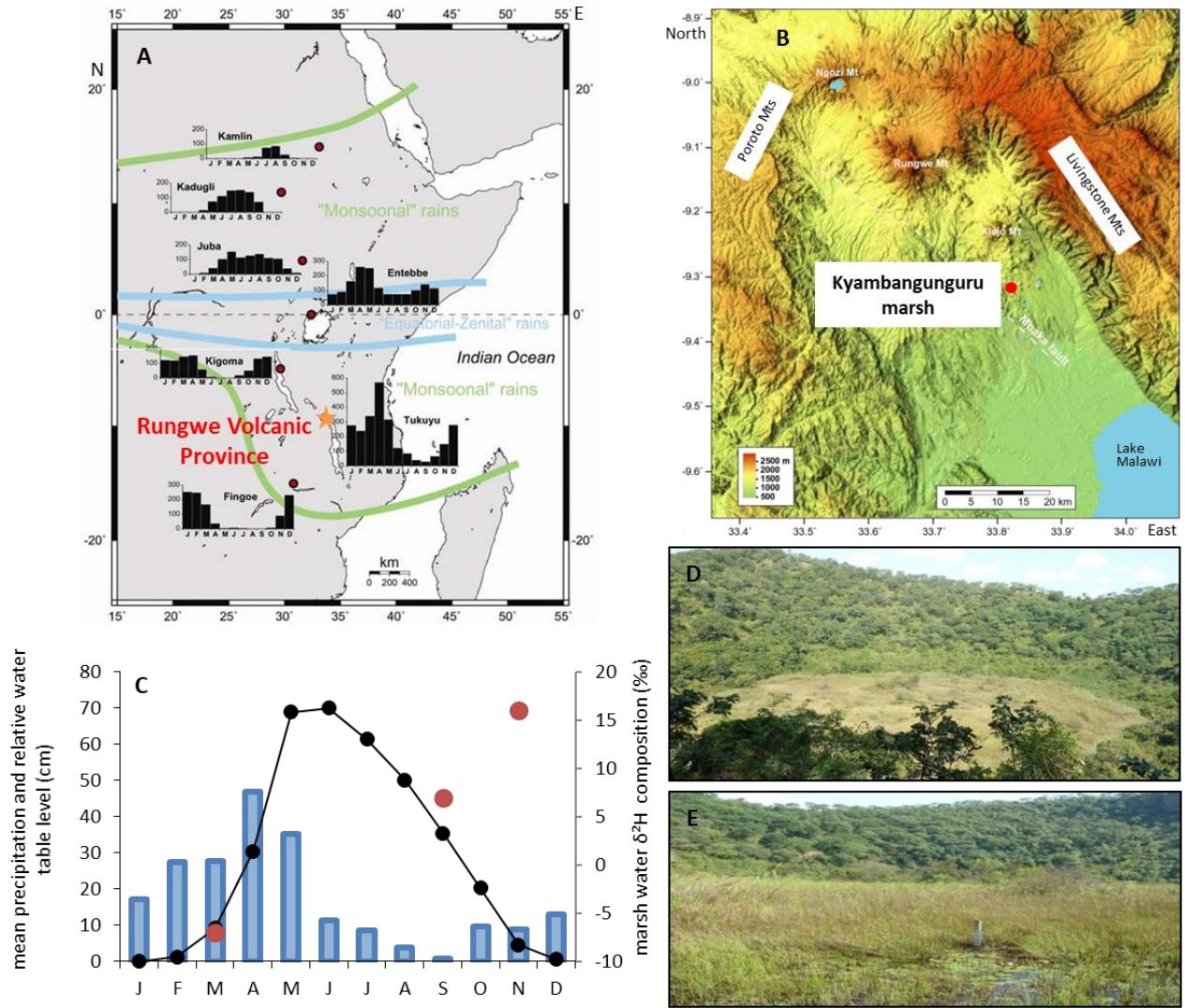


Figure 1.

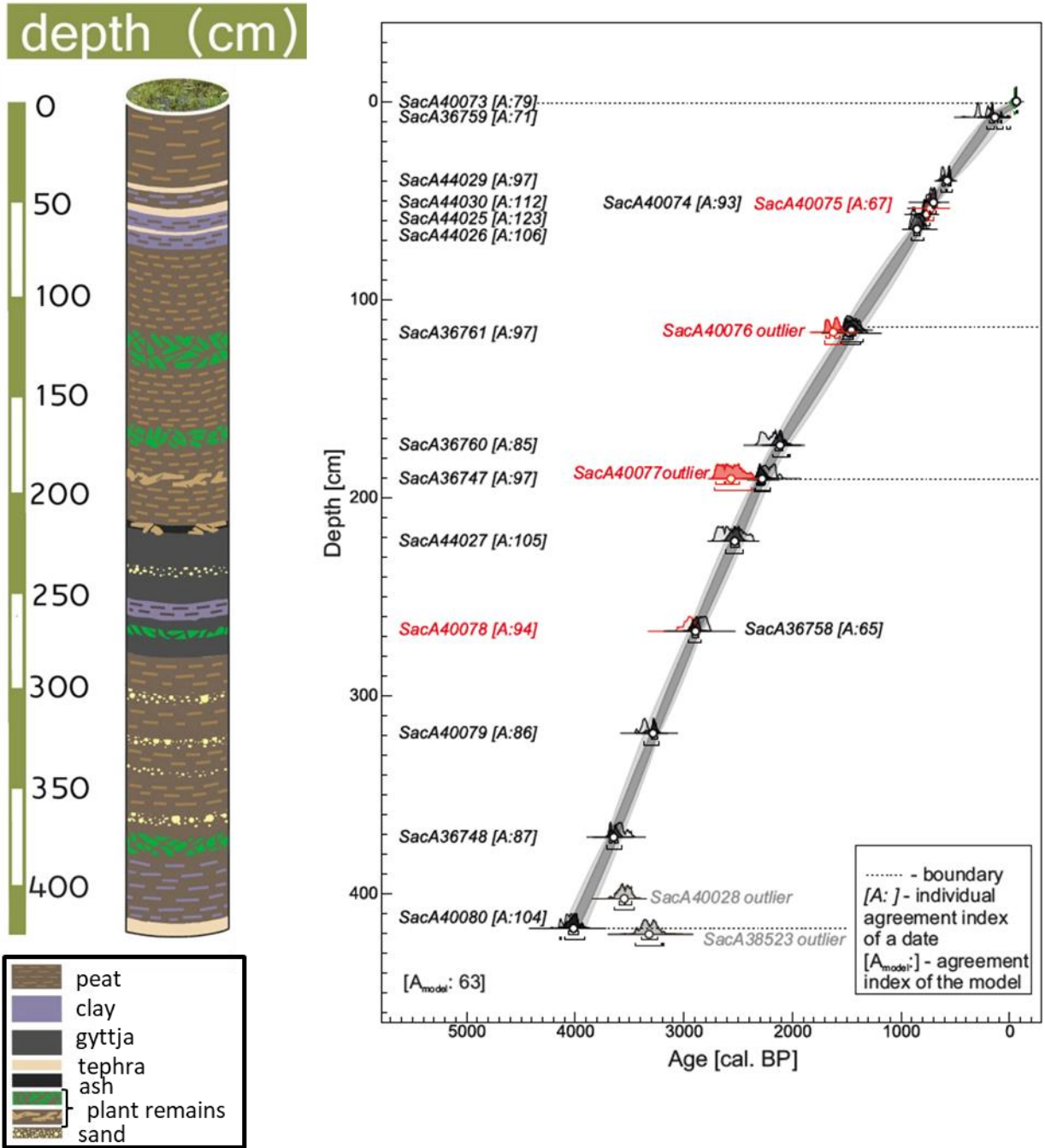


Figure 2.

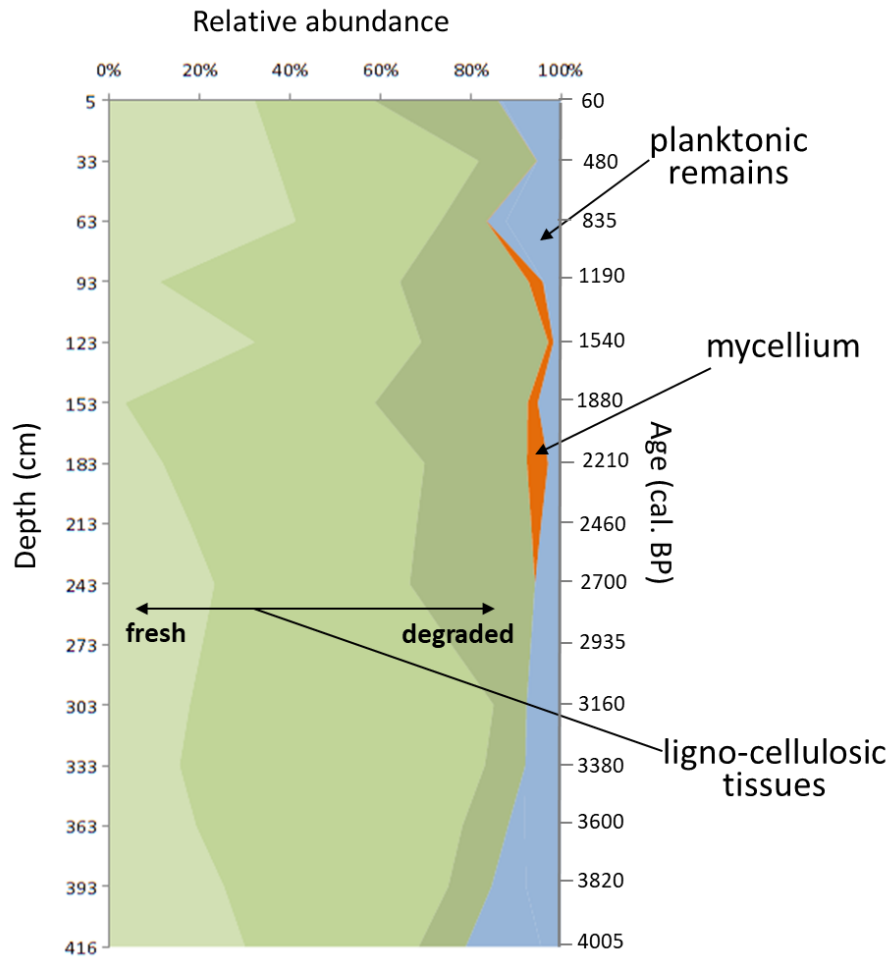


Figure 3.

Figure 4.

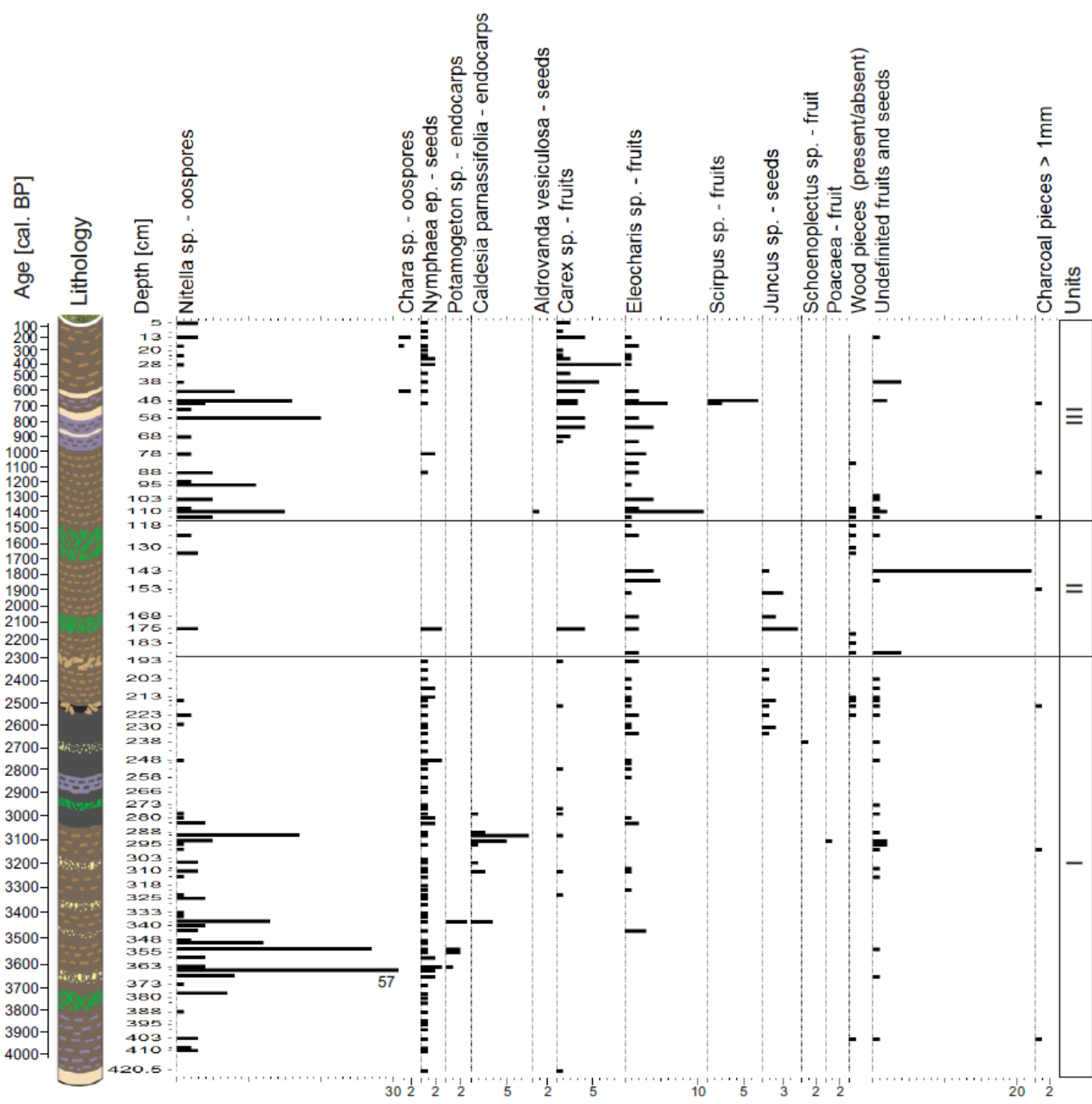


Figure 5.

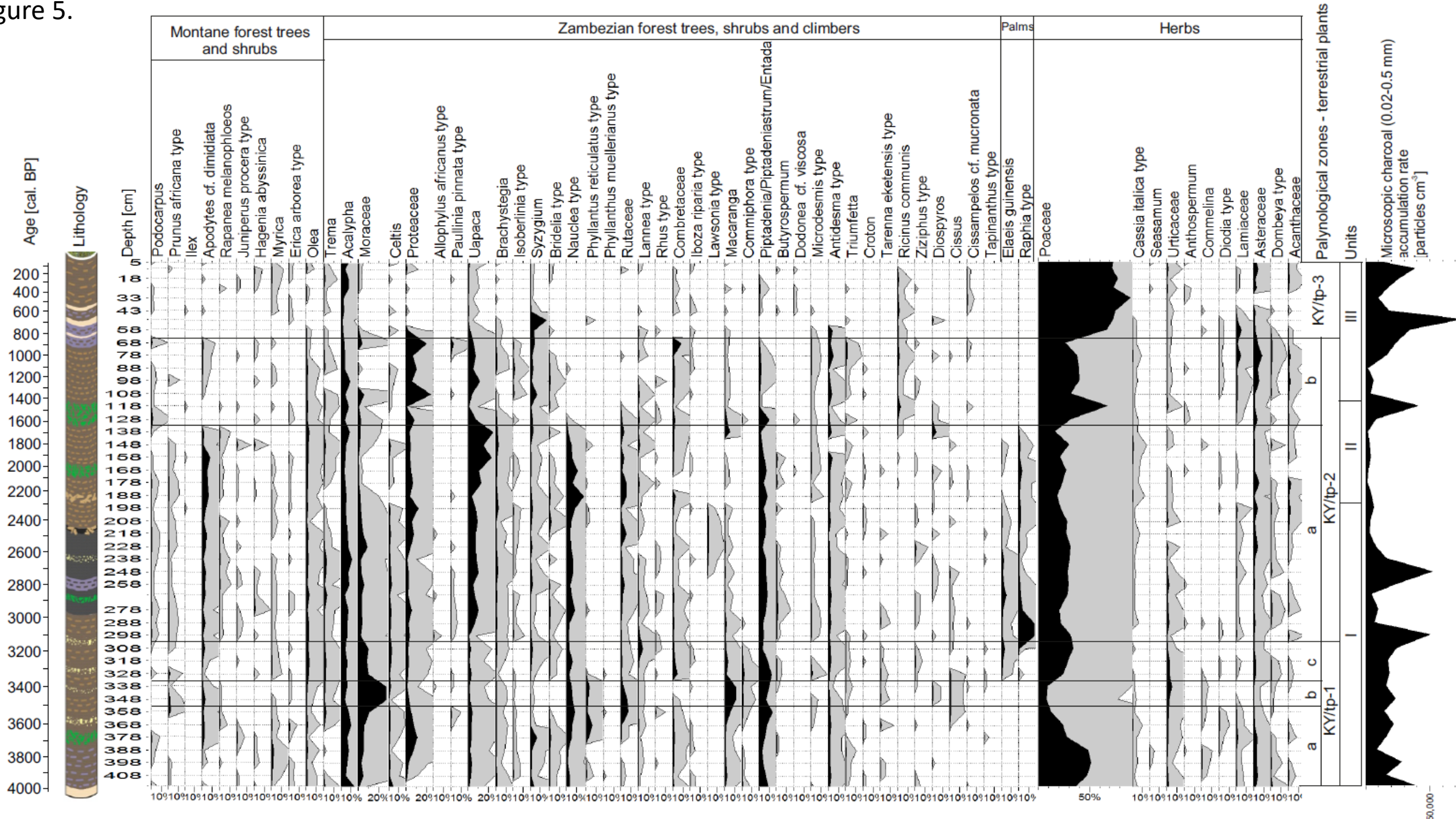
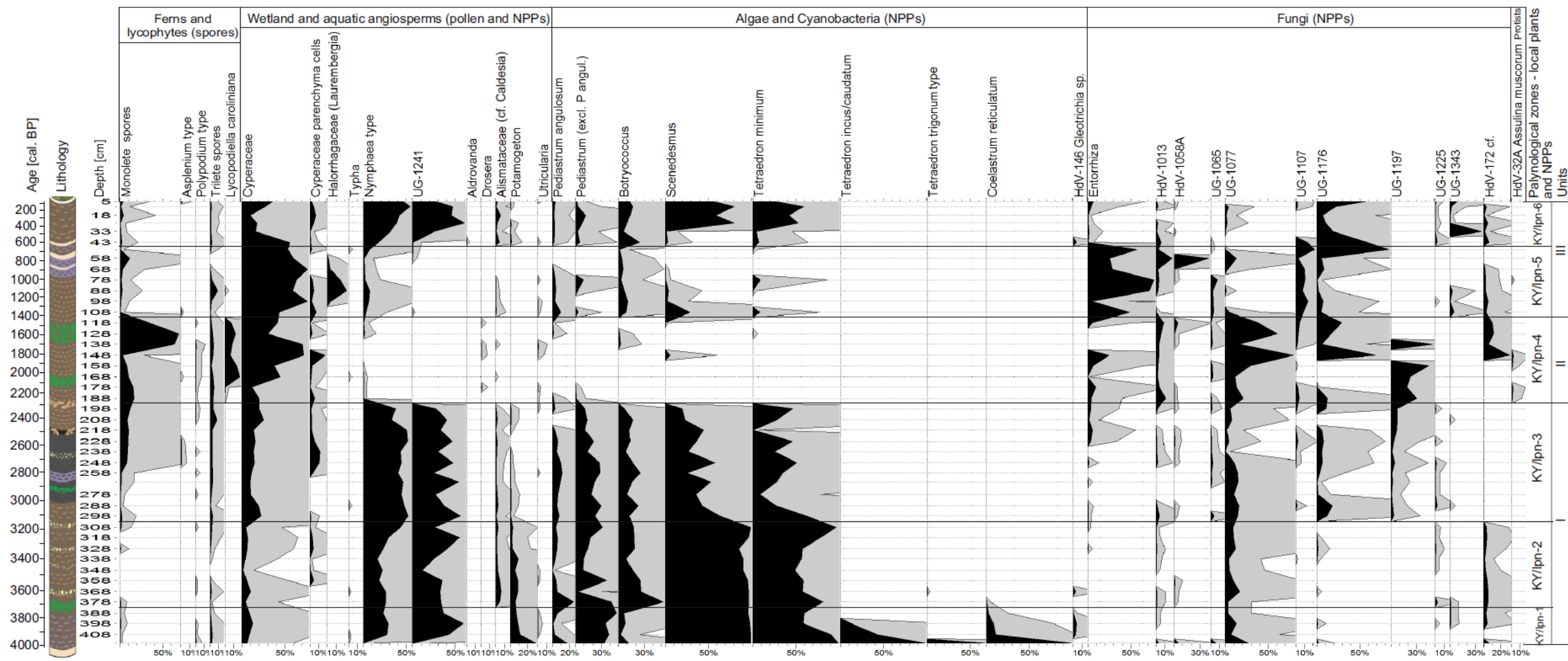


Figure 6.



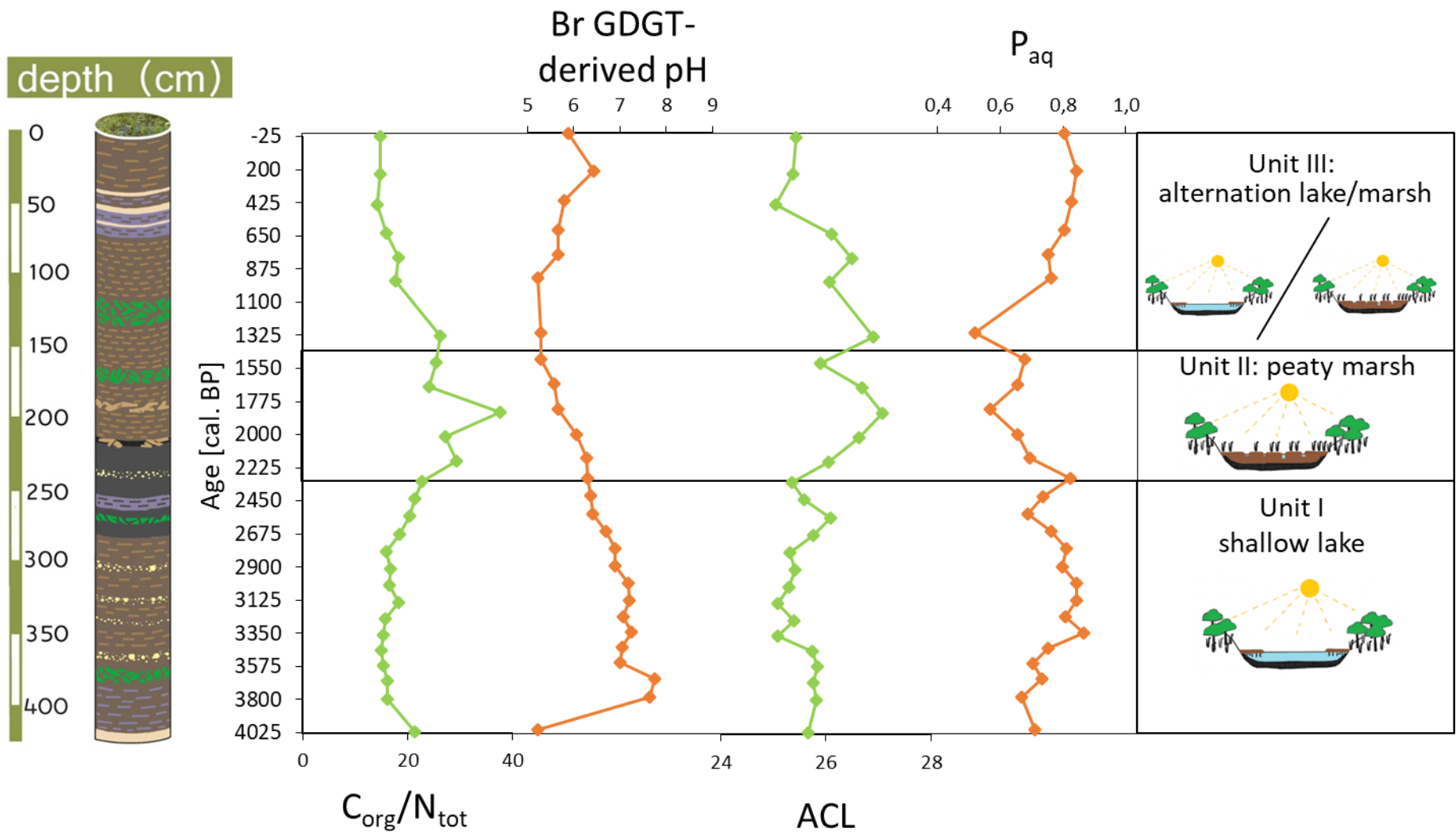


Figure 7.

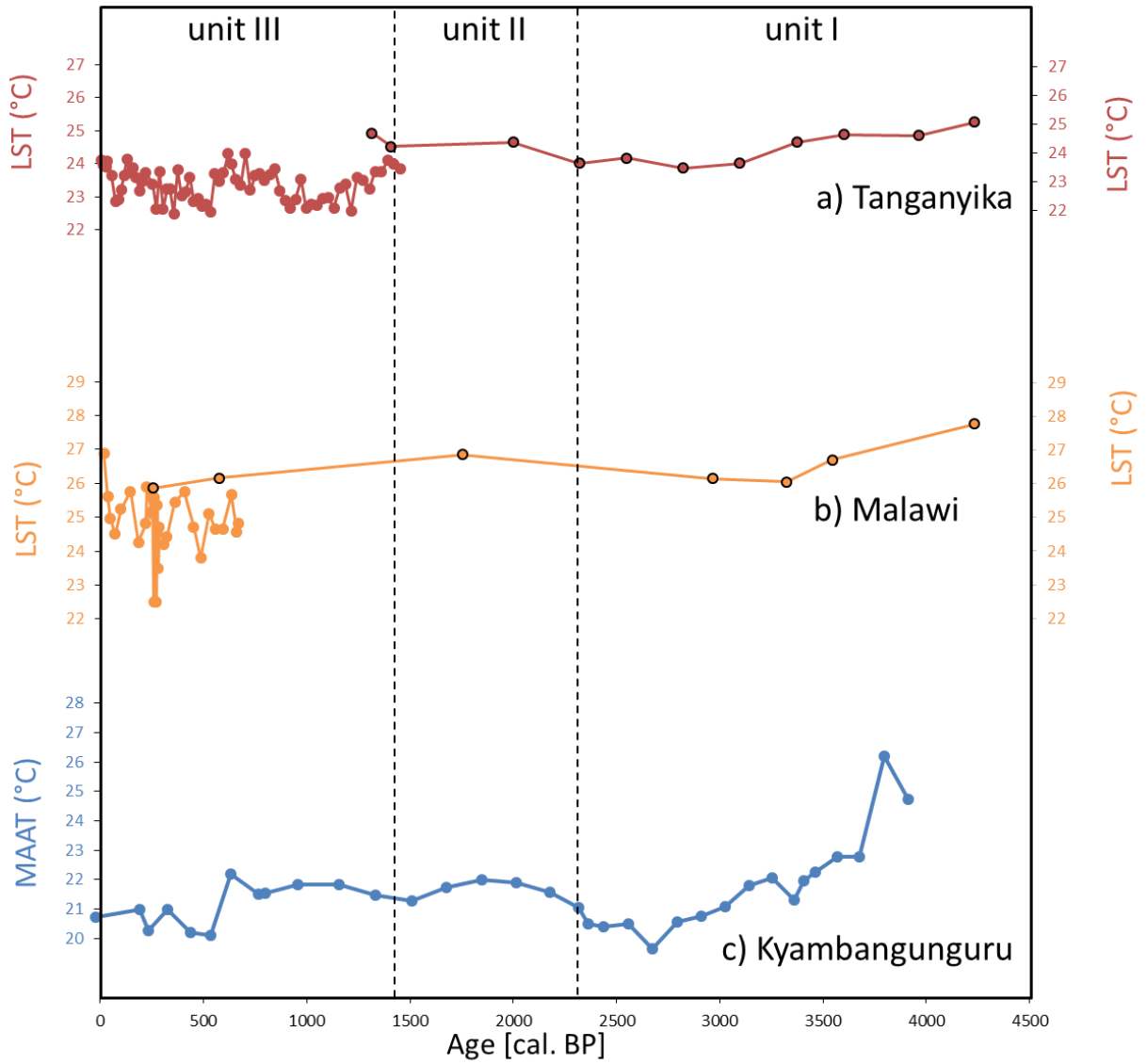


Figure 8.

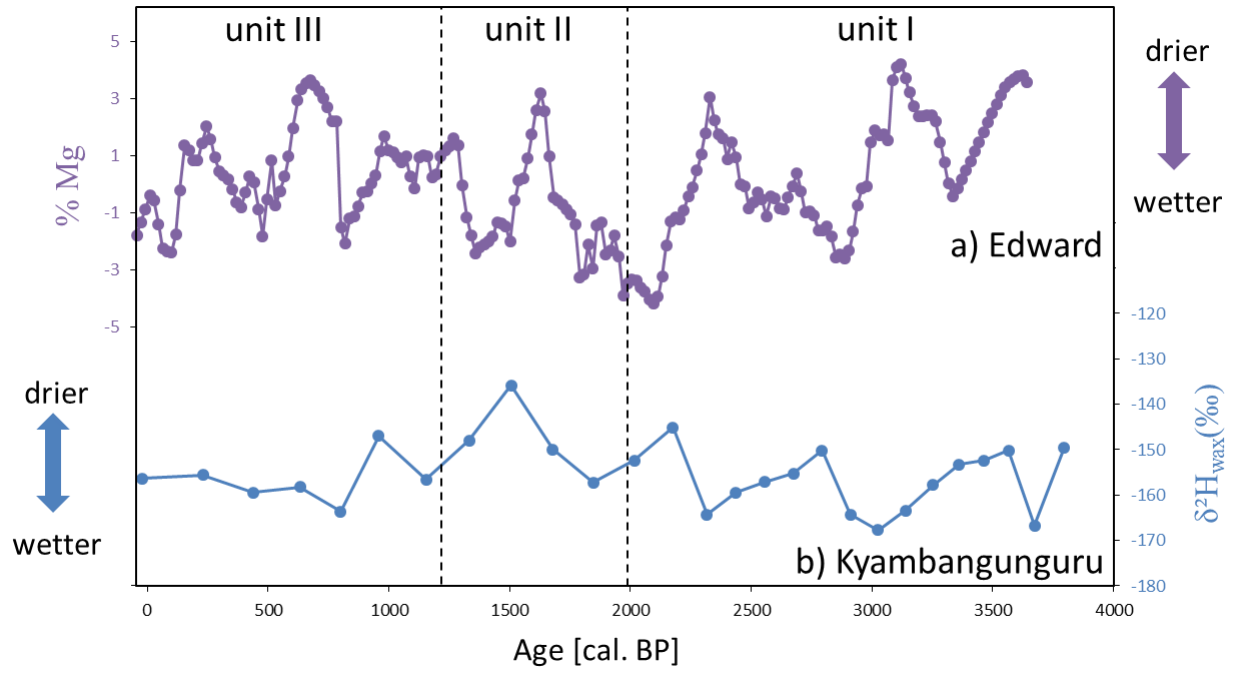


Figure 9.

Table 1.

Depth (cm)	AMS ¹⁴ C measure ref	Type of material	Carbon mass (mg)	¹⁴ C activity (pMC)	Measured ¹⁴ C age (cal. BP)	Calibrated age range [cal. BP; 95.4% (2 σ) range]	Modelled age ± σ error (cal. BP)	δ ¹³ C (‰)
0.5	SacA40073	TOM	1.82	112.21 ± 0.33	Post 1950	-8--8 (3.3 %) -43--46 (88.4 %) -47--47 (3.7 %)	-45±5	-29.5
8.0	SacA36759	TOM	0.75	96.86 ± 0.43	255 ± 35	435--408 (1.9 %) 325--254 (39.4 %) 225--143 (54.2 %)	135±50	-24.5
40.0	SacA44029	TOM	0.75	92.48 ± 0.24	630 ± 30	646--588 (58.7 %) 574--535 (36.7 %)	574±28	-23.8
51.0	SacA44030	TOM	0.75	89.99 ± 0.27	845 ± 30	763--673	701±15	-22.9
54.0	SacA40074	TOM	0.55	89.53 ± 0.29	890 ± 30	800--682	730±14	-28.0
54.0	SacA40075	Wood	0.91	90.27 ± 0.36	825 ± 30	742--666	729±14	-25.2
57.0	SacA44025	TOM	0.75	89.20 ± 0.23	920 ± 30	905--861 (14.1 %) 842--829 (1.3 %) 820--724 (80 %)	765±17	-21.4
64.0	SacA44026	TOM	0.75	88.55 ± 0.24	975 ± 30	921--774	852±30	-20.2
116.5	SacA36761	TOM	0.50	82.03 ± 0.48	1590 ± 45	1538--1348 (94.6 %) 1331--1324 (0.8 %)	1472±47	-25.5
116.5	SacA40076	Wood	1.37	80.47 ± 0.28	1745 ± 30	1702--1639 (39.3 %) 1633--1542 (56.1 %)	Not applied	-28.2
173.5	SacA36760	TOM	0.75	75.90 ± 0.38	2215 ± 40	2318--2059	2113±37	-20.1
190.5	SacA36747	TOM	0.75	75.68 ± 0.31	2240 ± 35	2328--2146 (90.1 %) 2131--2103 (5.3 %)	2278±34	-25.0
190.5	SacA40077	Wood	1.30	73.12 ± 0.26	2515 ± 30	2719--2379	Not applied	-27.5
222.0	SacA44027	TOM	0.75	73.18 ± 0.22	2510 ± 30	2715--2379	2531±42	-26.8
267.5	SacA36758	TOM	0.75	70.72 ± 0.37	2785 ± 40	2945--2759	2892±30	-27.9

267.5	SacA40078	Wood	1.40	69.87 ± 0.25	2880 ± 30	3062–2857	2893±30	-27.9
319.0	SacA40079	TOM	1.11	67.41 ± 0.25	3170 ± 30	3445–3423 (4.8%) 3411–3236 (90.6 %)	3283±34	-28.6
371.5	SacA36748	TOM	0.75	65.50 ± 0.29	3395 ± 35	3692–3657 (10.4 %) 3651–3478 (85 %)	3646±37	-28.2
402.0	SacA44028	TOM	0.75	65.76 ± 0.21	3365 ± 30	3639–3456	Not applied	-26.4
417.5	SacA40080	TOM Tephra	1.04	62.75 ± 0.24	3745 ± 30	4151–3960 (91.1 %) 3950–3926 (4.3 %)	4015±47	-20.0
420.5	SacA38523	TOM Tephra	0.30	67.47 ± 0.42	3160 ± 50	3450–3206 (93.8 %) 3197–3182 (1.6 %)	Not applied	-23.6