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1 The Gauls experienced the Roman Warm Period: oxygen isotope study of
2 the Gallic site of Thézy-Glimont, Picardie, France

3 Thibault Clauzel^a, Pascale Richardin^{b,b'}, Jannick Ricard^c, Yves Le Béchenec^c, Romain
4 Amiot^a, François Fourel^d, Brian Phouybanhdyt^{b,e}, Arnauld Vinçon-Laugier^a, Jean-Pierre
5 Flandrois^f and Christophe Lécuyer^{a,*, §}

6
7 a: Univ Lyon, Univ Lyon 1, ENSL, CNRS, LGL-TPE, F-69622, Villeurbanne, France

8 b: Centre de Recherche et de Restauration des Musées de France C2RMF, Palais du Louvre, Porte des Lions, 14
9 quai François Mitterrand, 75001 Paris, France

10 b': PRETECH – Préhistoire et Technologie, CNRS UMR7055, Université Paris Nanterre, Nanterre, France

11 c: TRAME (Université de Picardie Jules Verne) & Unité Mixte de Recherche n° 8546, AOrOc, CNRS, ENS
12 Ulm, Paris, France

13 d: Laboratoire d'Ecologie des Hydrosystèmes Naturels et Anthropisés, CNRS UMR 5023, Université Claude
14 Bernard, Lyon 1, France

15 e: GEOTRAC - Géochronologie Traceurs Archéométrie LSCE- Laboratoire des Sciences du Climat et de
16 l'Environnement [Gif-sur-Yvette] : DRF/LSCE

17 f: Université de Lyon, CNRS, UMR 5558, Laboratoire de Biométrie et Biologie Evolutive, Villeurbanne, France

18
19 *: also at Institut Universitaire de France

20 §: corresponding author, email address: christophe.lecuyer@univ-lyon1.fr

21
22 **Abstract (190 words)**

23 The Roman Warm Period (from ≈300 BCE to ≈300 AD) is a climatic optimum, which had a
24 key role in the development of the Roman civilization. This study provides new Mean Air
25 Temperatures (MATs) inferred from the oxygen isotope composition of 80 bones and teeth
26 apatite from 8 humans and 8 animals of the Gallic site of Thézy-Glimont, Picardie, France,
27 dated between the 3rd and 2nd century BCE. Various bones from the cephalic, axial and

28 appendicular skeleton of three human individuals were sampled. All bones have similar
29 phosphate $\delta^{18}\text{O}$ values for each individual, showing that the oxygen isotope signal of
30 phosphate groups may be homogeneously recorded in bones. The sampled individuals were
31 assumed to drink water from the nearby Avre river collecting local meteoric waters. The
32 MATs were reconstructed combining (1) oxygen isotope fractionation equations between
33 phosphatic tissues ($\delta^{18}\text{O}_p$) and drinking water ($\delta^{18}\text{O}_w$), and (2) relationships between MATs
34 and $\delta^{18}\text{O}_w$. Depending on the used MAT- $\delta^{18}\text{O}_w$ relationship, the reconstructed MATs are
35 $10.2^\circ\text{C}\pm 1.4^\circ\text{C}$ and $10.5^\circ\text{C}\pm 1.7^\circ\text{C}$, which are comparable to present days (10.9°C). This study
36 shows that the Gallic civilization also experienced relatively high temperatures since the
37 initiation of the Roman Warm Period.

38

39 Keywords: oxygen isotopes; human skeleton; climate; Gauls, Roman Warm Period

40

41 **1. Introduction**

42

43 Although the Holocene is considered to be a very stable interglacial period, climatic
44 variations from the centennial- to millennial-scale have been identified (Holzhauser et al.,
45 2005; Caroli and Caldara, 2007; Lirer et al., 2014; Luterbacher et al., 2016; Cisneros et al.,
46 2016). These variations are considered to be key elements in the rise and fall of civilizations.
47 For instance, the Medieval Warm Period witnessed the development of rural populations in
48 Europe linked to very productive harvests. Cities developed and major cultural innovations
49 took place during warm periods due to land shortages and increasing food requirements
50 (Fagan, 2008). On the other hand, at the end of the Late Bronze Age, most eastern
51 Mediterranean urban centers were either destroyed or abandoned throughout the Middle East
52 and the Aegean area. Some authors have recently suggested that a centuries-long

53 megadrought caused the widespread collapse of Bronze Age Palatial civilization (Kaniewski
54 et al., 2010; Drake, 2012). The Roman Warm Period (RWP), also called the Roman Climatic
55 Optimum or the Roman Humid Period, is a climatic variation of interest clearly identified in
56 the European record. The RWP is identified as a key period for the development of the
57 Roman culture all around the Mediterranean basin.

58 The concept of the RWP in Europe was first evoked by Lamb (1995) who provided a
59 synthesis of archaeological, glaciological and palynological studies. The temporal definition
60 of the RWP varies from one study to another, which may be due to the diversity of climate
61 proxies and records in Europe. On average, its period extends between ≈ 300 BCE and ≈ 300
62 AD. The most direct proxies of the RWP are archaeological testimonies. Pliny the Elder (23-
63 79 AD) testifies that, contrary to what had been the custom in earlier centuries, vines and
64 olive trees were cultivated farther north in Italy. Ptolemy of Alexandria (100-168 AD) noticed
65 that, around 140 AD, it rained every month of the year except August. McCormick et al.
66 (2012) developed an online database referencing archaeological evidence of climate
67 conditions during the Roman Period.

68 Although the RWP has been intensively studied in the European record, its climate
69 characteristics remain unclear. Most of the studies of continental western Europe (essentially
70 Spain, France and Italy) suggest a warm (e.g. McDermott et al., 2001; Roberts et al., 2001;
71 Desprat et al., 2003; Holzhauser et al., 2005; Frisia et al., 2005; Berger and Bravard, 2012;
72 Esper et al., 2012) and humid (e.g. Nesje et al., 2000; Wick et al., 2003; Ferrio et al., 2006;
73 Magny et al., 2007; Migowski et al., 2006; Zanchetta et al., 2007) climatic episode.

74 All the studies reconstructing the level of the Dead Sea also suggest a humid episode
75 (Frumkin et al., 1991; Heim et al., 1997; Bookman et al., 2004; Migowski et al., 2006). The
76 data coming from coastal and marine sediments from all over the Mediterranean basin
77 highlight a dry climate prevailing during the RWP (Schilman et al., 2001; Marquer et al.,

78 2008; Piva et al., 2008; Combourieu Nebout et al., 2009). This conflicts with the data coming
79 from continental Europe presented above, and suggests that the Mediterranean basin was
80 under the influence of a distinct climate regime during this episode.

81 High latitude records show a rather cold episode as documented in Iceland (Dahl-Jensen
82 et al., 1998; Seidenkrantz et al., 2007). There are some records of a cold climate period in
83 mid-latitudes, either in continental (Issar and Yakir, 1997; Baier et al., 2004; Mangini et al.,
84 2005; Büntgen et al., 2011) or oceanic (Bond et al., 2001; Taricco et al., 2009; Wang et al.,
85 2012; Lirer et al., 2014) environments. These temperature discrepancies may be explained by
86 the existence of three cold and dry intervals during the RWP, which are named Roman I to III
87 (Lirer et al., 2014; Margaritelli et al., 2016).

88 Variations of solar irradiance are currently invoked to explain the RWP (Bond et al.,
89 2001; Jiang et al., 2005; Mangini et al., 2005; Wang et al., 2012). Some studies highlight
90 correlations between solar activity variations and oceanic (Taricco et al., 2009; Lirer et al.,
91 2014) or continental (Mangini et al., 2005; Margaritelli et al., 2016) records, suggesting that
92 variations in solar activity may be the main mechanism responsible for the RWP. Dermody et
93 al. (2012) have presented a model of a climatic “see-saw” between western/coastal Europe
94 and Mediterranean/Eastern Europe, where solar activity variations trigger a change in the
95 position and intensity of jet streams through thermal variations of the North Atlantic Ocean.
96 However, these interpretations must be taken with caution, as the role of solar activity as a
97 driver for climate variations is questionable and has been challenged in the last decades (e.g.
98 CSIRO, 1983; Haigh, 2007; Lockwood, 2012; Oldenborgh et al., 2013).

99 All the presented records suffer from significant dating uncertainties and, for most of
100 them, are not focused on the RWP. No climatic reconstruction coming from human or animal
101 bones is available so far for this period. This study aims to present stable isotope proxy
102 records of Mean Air Temperature (MAT) inferred from vertebrate remains. Samples have

103 been collected from an archaeological site, well constrained in space and time, which is
104 located in the North of France where an especially important lack of data has been noticed.
105 The studied population belongs to the Gallic civilization. The Gauls, who came from Celtic
106 populations, occupied most of Western Europe during the Iron Age (800 BCE – 50 BCE).
107 Between the 2nd and 1st century BCE, the Roman Empire was involved in territorial conquests
108 from minor Asia through Northern Africa and Greece to Western Europe, which includes
109 Gallic territories. Therefore, the study of Gallic archaeological sites may bring information
110 dealing with the transition from the Iron Age Cold Epoch and the RWP or the first part of the
111 RWP, until ≈ 0 AD when all populations are expected to have been under the control of the
112 Roman Empire.

113 The climate characteristics of the RWP can be studied at mid-to-low latitudes using the
114 oxygen isotope composition of meteoric waters ($\delta^{18}\text{O}_{\text{mw}}$), which strongly depends on air
115 temperature, humidity, the trajectories of humid air masses and the amount of precipitation
116 (Dansgaard, 1964). Reconstructing the past $\delta^{18}\text{O}_{\text{mw}}$ variations of meteoric waters permits an
117 estimation of the past variations of those climate parameters.

118 It is known that the oxygen isotope composition of phosphate group ($\delta^{18}\text{O}_{\text{p}}$) from
119 mineralized tissues of mammalian individuals (bones, teeth) is related to the oxygen isotope
120 composition of drinking water ($\delta^{18}\text{O}_{\text{w}}$) (Longinelli, 1984; Luz et al., 1984; Luz and Kolodny,
121 1985; Kohn, 1996). During pre-industrial times, most of the consumed water came from local
122 sources (wells, springs, nearby watercourses), which are derived from local meteoric waters
123 (Daux et al., 2005; Darling, 2006). The contribution of water from food varies from one
124 species to another, according to the water-turnover of the animal (Kohn, 1996). Water
125 turnover scales to body-mass for most terrestrial vertebrates (Eberhardt, 1969). Humans are
126 medium omnivores with moderate water turnover, around 5 to 10% of total body water per
127 day (Shimamoto and Komiya, 2000), which means that the $\delta^{18}\text{O}$ of water ingested is strongly

128 and mainly imprinted by the $\delta^{18}\text{O}$ of drinking water. The cooking operation of brewing or
129 stewing may lead to an isotopic enrichment of the water contained in food compared to
130 meteoric water (Brettell et al., 2012), but its effect is not unanimous (e.g. Daux et al., 2008).

131 Therefore, measuring the oxygen isotope composition of phosphatic tissues from
132 terrestrial mammals allows the $\delta^{18}\text{O}_{\text{mw}}$ of contemporary meteoric waters to be calculated,
133 hence climate parameters such as MAT to be quantified. We demonstrate in this study that the
134 reconstructed MATs at Thézy-Glimont were similar to those of today.

135

136 **2. Material and methods**

137

138 *2.1. The Gallic site of Thézy-Glimont*

139

140 The Gallic site of Thézy-Glimont is located in the Valley of the Somme, in the region of
141 Picardie, France, 15 km southeast from the city of Amiens. (Figure 1). The site was excavated
142 in the context of rescue archaeology by the Service d'Archéologie Préventive Amiens
143 Métropole (SAAM) between November 2012 and May 2013 (Le Béchenec et al., 2016).

144 Picardie is located at mid-latitude (49°N-50°N) in the northwest of Europe, climatically
145 under the prevailing fresh and humid westerlies from the North Atlantic Ocean. The climate is
146 rated as Cfb (warm temperate, fully humid with warm summers) according to the Köppen-
147 Geiger classification (Kottek et al., 2006) with a yearly average temperature of 10.9°C and a
148 small seasonality amplitude of $\pm 5.4^\circ\text{C}$ throughout the years 1980 to 2010 (data from Météo
149 FranceTM). The yearly precipitation rate is 631 mm, with a small seasonality amplitude of
150 ± 7 mm, July and December being the two most humid months.

151 Thézy-Glimont (49°48'N – 2°25'E) is located at the bottom of the Avre Valley, which
152 is a tributary of the Somme river. The Avre river is assumed to have been the drinking water

153 source of the inhabitants of Thézy-Glimont as it is located on the edge of the archaeological
154 site and no water wells have been observed.

155 The SAAM highlighted 17 rectangular graves during the excavation. The graves contain
156 horses (*Equus caballus*), bovids (*Bos taurus*) and a human (*Homo sapiens*) for most of them.
157 Sometimes a pig (*Sus scrofa*) has been found. All deposits seem to be intentional and
158 organized. They were filled with the same sediment (white chalk) coming from the digging,
159 whereas the tradition of Gallic funeral practices is usually cremation. The excavation report
160 chooses to present the graves as “offering graves”, even if their true nature is not fully
161 understood (Le Béchenec et al., 2016). A total of 8 human, 12 horse, 39 bovid and 3 pig
162 skeletons were excavated from the 17 graves.

163 All 8 humans, 2 horses, 5 bovids and 1 pig were sampled from this corpus (see Table 1).
164 A great number of sampled individuals come from TGV 1135 as it was the first grave to be
165 excavated, the biggest of the archaeological site and the one with the highest number of
166 individuals. The humans are all young males without any noticeable disease. The human
167 identified as sample TG-02 is the youngest, around 15 years old, while TG-05 is the oldest
168 (between 30 and 59 years old). The forensic expertise suggested a quick burying after death
169 before the rigor mortis happened. Although the

170 Eighty samples (bones and teeth) were collected (Table 1). Bone samples were collected
171 using a toothless diamond saw blade fitted to a DremelTM drill. Various elements coming
172 from the axial, cephalic and appendicular skeletons of TG-01, TG-02 and TG-03 were
173 sampled along with 19 bones for TG-01, 14 for TG-02 and 19 for TG-03, respectively (Figure
174 2). The bone samples were first abraded using a surgical knife in order to remove the altered
175 layer at the surface and grossly crushed using a hammer. All samples were washed with
176 acetone to degrease them, rinsed three times with UltrapureTM water and dried in an oven
177 under medium vacuum, at 60°C, for 24 hours. Clean samples were then crushed in fine

178 powder using a Pulverisette 7TM planetary ball mill composed of zirconium oxide. Teeth were
179 extracted from the skulls by gently pulling them from their sockets. They were rinsed with
180 UltrapureTM water and cleaned using a toothbrush. The powder was obtained by drilling the
181 enamel of the teeth with a DremelTM diamond-head drill bit.

182

183 2.2. *Dating of the archaeological site*

184

185 During the excavation of the archaeological site, Beta AnalyticsTM realized ¹⁴C datings
186 on 13 samples, giving dates with 68.3% probability (1 σ) between 200 BCE and 5 AD. New
187 ¹⁴C dates were realized on collagen extracted from 7 human, 2 horse and 1 bovid bones. The
188 protocol of the extraction of soluble collagen was based on the method described by Longin
189 (1971) and has been previously described in detail by Richardin et al. (2017).

190 All measurements were performed with Accelerator Mass Spectrometry techniques
191 from the Artemis AMS facility at Saclay (France) (Moreau et al., 2013). Calendar dates were
192 determined using the OxCal v.4.2 procedure and the most recent calibration curve data for the
193 Northern Hemisphere, IntCal13 (Reimer et al., 2013). Calibration of radiocarbon ages was
194 performed considering that the contribution of marine products to the diet of the Gauls
195 remained marginal. Calibrated date ranges (Table 1) correspond to 95.4% probability (2 σ).
196 The archaeological site is dated with 95.4% certainty between 350 BCE and 60 BCE with the
197 highest probability being between 350 BCE and 200 BCE, one century earlier than previously
198 proposed. There are no obvious differences among and between the humans and the animals,
199 therefore all the graves seem contemporary, their digging possibly spread over a century.

200

201 2.3. *Bone turnover and isotope homogeneity*

202

203 The bone tissue is vascularized, composed at 60 to 70% of hydroxyapatite crystals
204 tightly inter-grown with an organic matrix predominantly composed of type I collagen for 20
205 to 30%. Several ion substitutions can be observed in the crystalline matrix (LeGeros, 1981;
206 Wopenka and Pasteris, 2005). Enamel from teeth has much larger bioapatite crystals and a
207 very low organic content, which makes it less sensitive to diagenesis (Tütken and
208 Vennemann, 2011). Bone and tooth formation starts in the embryo, but human teeth complete
209 accretion long before bone growth has finished. The human permanent dentition is expected
210 to be fully formed at about 12 years, while a complete bone structure can be achieved at the
211 age of 20. Tooth enamel and dentine grow by accretion and do not renew afterwards. On the
212 contrary, there is a set-up of re-absorption of old bone (both organic and inorganic matrix)
213 and deposition of new bone tissue. At maturity, this mechanism happens to be continuous and
214 the bone is said to “turnover”. For a young adult, the total bone tissue is expected to turnover
215 every five to six years, although it may depend on the bone region or the individual health
216 status (Glimcher, 2006). Therefore, the stable isotope composition of the bone tissue only
217 represents the last 5 to 10 years of the individual life, while that of a tooth will represent the
218 first 12 years of life. Some studies use the discrepancy of bone turnover to highlight a change
219 of the diet (e.g. Lamb et al., 2014).

220 Because of poor conservation of the material in the field, the lack of articulations
221 between bones or the preciousness of samples, most isotope studies on bone material are realized
222 on one bone (generally the femur or the rib) and assume the value is representative of the
223 whole individual characteristics (e.g. Prowse et al., 2004; Jay and Richards, 2006). However,
224 if bone turnover is highly variable at the intra- and inter-individual scale, we can wonder to
225 what extent the isotope study of one bone is representative of the whole skeleton? To
226 investigate this issue, several parts (bones and teeth) of the same skeleton have been sampled
227 from the three best conserved and most complete skeletons of human individuals (TG-01, TG-

228 02 and TG-03) excavated in Thézy-Glimont (see Figure 2). Intra-comparisons of the oxygen
229 isotope composition of mineralized tissues were realized for each individual.

230

231 2.4. *Oxygen isotope ratios of apatite phosphate*

232

233 Oxygen isotopic analyses were performed with stable isotope mass spectrometry
234 techniques. Phosphate ions were extracted from the apatite using a wet chemistry procedure.
235 After the dissolution of about 30 mg of each sample in 2 mL of 2M HF at ambient
236 temperature for 24 hours, the CaF₂ precipitate was separated from the sample solution by
237 centrifugation. The phosphate ions were neutralized with KOH and isolated using a strong
238 anionic resin (Amberlite IRN 78) before being precipitated as silver phosphate following the
239 method first described by Crowson et al. (1991) and slightly modified by Lécuyer et al.
240 (1993). The oxygen isotope data were obtained using a high-temperature pyrolysis continuous
241 flow technique developed by Lécuyer et al. (2007) and Fourel et al. (2011). For each sample,
242 five aliquots of 400–500 µg of silver phosphate were mixed with 500 µg of carbon loaded into
243 3.5x5 mm silver foil isotope grade capsules (EurovectorTM). Pyrolysis was performed at
244 1450°C using a PYROcubeTM elemental analyzer (Elementar GmbH-Germany) interfaced in
245 continuous flow mode with an isotopic ratio mass spectrometer IsoprimeTM (Elementar UK
246 Ltd UK). Measurements were performed with CO gas separated by purge and trap technology
247 within the PYROcubeTM system at the Laboratoire de Géologie de Lyon and at the
248 Laboratoire d'Ecologie des Hydrosystèmes Naturels et Anthropisés, both hosted by the
249 Université Claude Bernard Lyon 1. The official unit for isotopic ratios is the Urey (Ur),
250 although the unit commonly used is the permil (‰, 1‰ = 1mUr). For practical reasons, data
251 are reported as δ¹⁸O values with respect to V-SMOW in ‰ units.

252 Oxygen isotope compositions were drift corrected to account for instrumental
 253 variations. Results were calibrated against silver phosphate samples of known $\delta^{18}\text{O}$ values
 254 precipitated from the NBS120c standard for which the $\delta^{18}\text{O}$ value was fixed at 21.7‰ (V-
 255 SMOW) (Lécuyer et al., 1993; Chenery et al., 2010; Halas et al., 2011), as well as with
 256 NBS127 (barium sulfate) having a certified value of 9.3‰ (V-SMOW) (Hut, 1987). NBS120c
 257 standards were converted into silver phosphate along with the samples from each chemistry
 258 batch and considered as “internal standards”. They have $\delta^{18}\text{O}$ values that range from 21.4‰
 259 to 22.2‰ with a mean value of 21.76 ± 0.25 ‰ ($n=12$). The analytical uncertainty for samples
 260 is always lower than 0.3‰ (2σ).

261

262 **3. Results**

263

264 *3.1. Intra-skeletal variability of the $\delta^{18}\text{O}_p$ from human bones*

265

266 The intra-skeletal variability of the $\delta^{18}\text{O}_p$ values from the three human individuals
 267 TG-01, TG-02, TG-03 range from 16.1‰ to 16.9‰ with a mean value of 16.6‰ for TG-01,
 268 from 17.4‰ to 17.9‰ with a mean value of 17.6‰ for TG-02 and from 16.4‰ to 17.3‰
 269 with a mean value of 16.8‰ for TG-03 (Figure 2). The standard deviation, representing the
 270 intra-skeletal variability of oxygen isotope composition, is respectively 0.2‰, 0.2‰ and
 271 0.3‰ for TG-01, TG-02 and TG-03. Local deviations of bones $\delta^{18}\text{O}_p$ from the average
 272 skeleton $\delta^{18}\text{O}_p$ value has been calculated for each individual as follows:

273

$$\delta^{18}O_{skeleton}^{bone}(\text{‰}) = \delta^{18}O_{V-SMOW}^{bone} + \delta^{18}O_{skeleton}^{V-SMOW} + \delta^{18}O_{V-SMOW}^{bone} \times \delta^{18}O_{skeleton}^{V-SMOW} \times 10^{-3}$$

274

275 No localized and repeatable enrichment or depletion in ^{18}O relative to the average $\delta^{18}\text{O}_p$
276 value is observed among the three individuals.

277

278 3.2. *Oxygen isotope values of phosphatic tissues*

279

280 The oxygen isotope compositions of apatite phosphate from the 80 samples were
281 analyzed (Figure 3, Supplementary Table 1). The $\delta^{18}\text{O}_p$ values of human bones and teeth
282 range from 15.1‰ to 18.0‰. Teeth are systematically enriched in ^{18}O relative to bones by
283 respectively 0.9‰, 0.3‰ and 0.9‰ for TG-01, TG-02 and TG-03. Human bones have a mean
284 $\delta^{18}\text{O}_p$ value of 16.2‰ with a standard deviation of 0.8‰. TG-02, who is the youngest
285 analyzed individual, is significantly enriched by 1.3‰ compared to the other humans. The
286 analyzed pig has a mean $\delta^{18}\text{O}_p$ value of 15.6‰ with a standard deviation of 0.5‰. The bones
287 from 5 bovids have a mean $\delta^{18}\text{O}_p$ value of 17.3‰ with a standard deviation of 0.2‰, while
288 the two teeth from individuals TG-B4 and TG-B5 have a mean value of 16.1‰ and a standard
289 deviation of 0.1‰. Teeth are significantly depleted in ^{18}O by 1.3‰ relative to bones. No
290 significant isotopic difference is observed relative to sex (TG-B1 versus the others) or to the
291 grave the individuals were buried in (TG-B5 versus the others). Molars from the two horses
292 have a mean value of 15.8‰ with a standard deviation of 0.6‰. In the case of the horse TG-
293 CH1, the bone $\delta^{18}\text{O}_p$ value is 1.4‰ higher than that of the tooth. The bones of herbivores
294 (horse, bovid) have a mean $\delta^{18}\text{O}_p$ of 17.5‰, while the average $\delta^{18}\text{O}_p$ for omnivores (human,
295 pig) is 15.9‰.

296

297 4. Discussion

298

299 4.1. *Sample preservation and diagenesis*

300

301 The bones from Thézy-Glimont showed a visible brownish layer of alteration, which
302 was removed using a surgical knife. Some of the human teeth showed attrition, which is a
303 common feature for the gallic civilization. It may be due to the habit of eating gritter bread
304 containing sediment particles. Attrition may have altered mechanically the teeth without
305 scrambling the isotopic signal recorded in enamel.

306 In apatite, P-O bond is known to be very resistant to inorganic isotope exchange with
307 aqueous fluids at surface temperatures (e.g. Blake et al., 1997; Lécuyer et al., 1999). No trace
308 of microbially induced diagenesis has been detected. An expected offset of the $\delta^{18}\text{O}_p$ signal
309 between teeth and bones from the same individual has been noticed (Touzeau et al., 2013;
310 Martin et al., 2017). All three individuals TG-01, TG-02 and TG-03 have very low $\delta^{18}\text{O}_p$
311 intra-skeletal variability (Figure 2). When considering all the measurements from each
312 skeleton the three standard deviations of 0.2‰, 0.2‰ and 0.3‰, respectively, are in the same
313 order of magnitude as the analytical uncertainty ($\pm 0.3\%$). No noticeable and localized
314 isotopic enrichment or depletion relative to the average $\delta^{18}\text{O}_p$ value has been noticed. All
315 these observations support the hypothesis of the preservation of the pristine oxygen isotope
316 compositions of phosphatic tissues.

317

318 4.2. *Homogeneity of oxygen isotope values in human skeleton apatite phosphate*

319

320 The mapping of the oxygen isotope composition of human bone apatite at the skeletal
321 level of observation (Figure 2) revealed that, in the case of $\delta^{18}\text{O}_p$ values, the study of one
322 selected bone is representative of the composition of the whole individual. We observe an
323 isotopic homogeneity at the skeleton scale, although the bones have various turnover rates.
324 Two complementary explanations are proposed: TG-01, TG-02 and TG-03 being young, they

325 might have only realized one cycle of bone turnover at most and their diet did not change
326 during this interval of time. As migrations can be recorded by apatite $\delta^{18}\text{O}_p$ values (e.g.
327 Lightfoot and O'Connell, 2016), we also propose that they were sedentary and lived all their
328 life in Picardie, France, which could be confirmed with supplementary analysis such as the
329 use of strontium isotopes. The oxygen isotope variability inside a bone has not been
330 investigated, but the epiphysis of some bones (e.g. the humerus) and the diaphysis of others
331 (e.g. the tibia) were analyzed and no noticeable offset was detected. With caution, following
332 studies of the oxygen isotope composition of bones in archaeological sedentary populations
333 may assume that one bone is representative of the whole individual. The $\delta^{18}\text{O}_p$ value of
334 vertebrates is also dependent on body temperature (Kohn, 1996) and has been used to study
335 the thermophysiology of terrestrial (Barrick and Showers, 1994, 1995; Amiot et al., 2006) and
336 marine vertebrates (Bernard et al., 2010; Séon et al., 2020). Thermal variations (or regional
337 heterothermy), which exist at a human body scale, are not documented at the light of our
338 study of the oxygen isotope ratios of phosphatic tissues.

339

340 4.3. *The $\delta^{18}\text{O}_p$ record in teeth and bones from individuals of Thézy-Glimont*

341

342 There is a very low $\delta^{18}\text{O}_p$ variability observed between humans and animals from
343 Thézy-Glimont. The $\delta^{18}\text{O}_p$ values from herbivore bones are on average 1.7‰ higher than
344 those from omnivorous individuals. This can be explained by the diet of the herbivores
345 mainly eating grass and leaves containing water which is ^{18}O -enriched in relation to the plant
346 evaporative system (Barbour, 2007). Comparing the oxygen isotope value from herbivores
347 and omnivores (mainly humans) enables to address the question of the influence of food
348 processing on $\delta^{18}\text{O}_p$. Although no information is available about the dietary practices on this
349 archaeological site, stewing or brewing may be detected through an isotopic enrichment of

350 body water later translated in a higher $\delta^{18}\text{O}_p$ of human bones compared to herbivores (Brettell
351 et al., 2012). This effect is not detected in this study, which leads to two possible conclusions:
352 either the inhabitants of Thézy-Glimont had dietary practices which did not include brewing
353 or stewing, or these cooking processes are not detected in the oxygen isotope composition of
354 bone apatite. This latter conclusion agrees the works of Daux et al., (2008) on modern human
355 teeth samples and Chesson et al. (2010) on the similarity of oxygen isotope values of different
356 modern beverages and meteoric waters in a same location.

357 The $\delta^{18}\text{O}_p$ value of human teeth enamel is higher than the $\delta^{18}\text{O}_p$ value of bones,
358 especially for TG-01 (+0.9‰) and TG-03 (+0.9‰). These values are comparable to the offset
359 between infant and adult bones of $\approx+1.2\text{‰}$ found by Britton et al. (2015) which can be
360 attributed to a specific diet during childhood. A large part of the child's diet consists of milk,
361 which is ^{18}O -enriched from the mother's body water. The second and third molars,
362 respectively M2 and M3, are considered post-weaning teeth and do not record any breast-
363 feeding. Although the M3 is not discernable from the oxygen isotope value of bones from
364 TG-02 and is ^{18}O -depleted compared to the two other teeth for TG-03, the M2 has similar
365 values than pre-weaning teeth (M1 for TG-01 and PM1 for TG-03). On the contrary,
366 herbivore teeth are depleted in ^{18}O compared to bones: -1.3‰ for bovids and -1.4‰ for the
367 horse TG-CH1. Herbivore hypsodont teeth, which grow continuously and are often used to
368 study climate seasonality (e.g. Bernard et al., 2009), record a signal which is more limited in
369 time than bones. For the last decade, the average interannual change in yearly MAT of
370 Picardie, France, is $\pm 1^\circ\text{C}$ (data from historique-meteo.net) and can go as high as 2 – 2.5°C
371 (for instance between 2013 and 2014). Translated into $\delta^{18}\text{O}$ of local meteoric waters, this
372 interannual change is $\pm 0.6\text{‰}$ on average and can go as far as a $\pm 1.6\text{‰}$, which is in the range
373 of the teeth-bones offset noticed for the herbivores. We propose that herbivore teeth recorded
374 more time-limited interval and colder climatic conditions, hence presenting an offset from the

375 bones $\delta^{18}\text{O}$ record, which represents a pluriannual time scale. Bones are a better material to
376 reconstruct MAT on the pluriannual time scale than teeth which record either a drinking water
377 environmental signal mixed with breastfeeding in the case of humans or a time-limited signal
378 in the case of herbivores. Therefore, only the oxygen isotope values coming from bones are
379 considered for the calculation of $\delta^{18}\text{O}_w$ of drinking water and MATs reconstruction.

380 TG-02 is ^{18}O -enriched compared to all the other humans and there is no significant
381 difference between the $\delta^{18}\text{O}_p$ of his bones and his teeth (see Figure 3). This isotopic offset can
382 be attributed to his young age (estimated at 15 years) keeping in mind that the set-up of bone
383 turnover is only achieved around 20 years-old. The bones from TG-02 might still record a
384 residual signal from the breast-feeding period, mixed with a signal coming from drinking
385 water. Therefore, this individual was removed from the reconstruction of the $\delta^{18}\text{O}$ of
386 environmental drinking water.

387 The oxygen isotope composition of phosphatic tissues from individuals with a medium
388 to low water turnover is mostly representative of the $\delta^{18}\text{O}$ of the drinking water. Thézy-
389 Glimont is located in a valley bottom, a few meters away from the Avre river, which is
390 assumed to be the main drinking water source of local inhabitants. The $\delta^{18}\text{O}_p$ record is,
391 therefore, representative of the climate parameters linked to the Avre river. The relationship
392 between the $\delta^{18}\text{O}_p$ of phosphatic tissues and the $\delta^{18}\text{O}_w$ of drinking water has been determined
393 empirically in numerous groups of vertebrates. These oxygen isotope fractionation equations
394 mainly depend on the basal metabolism, body mass and diet (herbivorous vs. carnivorous vs.
395 omnivorous) of species. Several equations have been used in this study and are represented in
396 Figure 4 as follows:

397

398 Daux et al. (2008) for humans:

399

$$\delta^{18}O_w(\pm 1.51\text{‰}) = 1.54 \times \delta^{18}O_p - 33.72$$

400

401 Longinelli (1984) for pigs:

402

$$\delta^{18}O_w = \frac{\delta^{18}O_p - 22.71}{0.86}$$

403

404 D'Angela and Longinelli (1990) for bovids:

405

$$\delta^{18}O_w = \frac{\delta^{18}O_p - 24.90}{1.01}$$

406

407 Bryant et al. (1994) for equids:

408

$$\delta^{18}O_w(\pm 2.46\text{‰}) = \frac{\delta^{18}O_p - 22.90}{0.6863}$$

409

410 The errors associated with the linear regression of empirical data ($y = mx + b$) have
411 been computed as the mean standard error for the “y” value when not available in the original
412 publication and are $\pm 0.2\text{‰}$ for the equation of Longinelli (1984) and $\pm 0.1\text{‰}$ for the equation
413 of D'Angela and Longinelli (1990). We highlight that these uncertainties may be
414 underestimated as both of the isotope fractionation equations have been established with a
415 very low number of data, respectively 7 average $\delta^{18}O$ values for Longinelli (1984) and 5
416 average $\delta^{18}O$ values for D'Angela and Longinelli (1990).

417 Using these equations, $\delta^{18}O_p$ were converted into $\delta^{18}O_w$ values (Figure 4,
418 Supplementary Table 2), which were interpreted as reflecting the oxygen isotope composition
419 of the Avre river during the 2nd to 3rd century BCE. The mean $\delta^{18}O_w$ inferred from humans

420 and animals is -8.4‰ , with a standard deviation of $\pm 1\text{‰}$. The propagation of uncertainty,
421 represented as error bars in Figure 4, have been calculated as the quadratic sum of the
422 analytical uncertainty, errors associated with the equations and variability in the case of
423 several measurements obtained for a single individual (TG-01, TG-02, TG-03 and TG-C1).
424 Taking into account these uncertainties, all the $\delta^{18}\text{O}_w$ values are homogenous and we can
425 affirm that they represent one main water source.

426 The present-day isotopic composition of the Avre river water was measured by Négrel
427 and Petelet-Giraud (2005) twice in a year (Avril and October 2001), with respective $\delta^{18}\text{O}_w$
428 values of -6.86‰ and -7.00‰ . The $\delta^2\text{H}$ and $\delta^{18}\text{O}$ values of the Avre river match the Global
429 Meteoric Water Line (Craig, 1961) as well as the Brest-Orléans isotopic rainwater domain,
430 confirming that the signal from the Avre river reflects the composition of local meteoric
431 waters. The oxygen isotope composition of meteoric waters is dependent on temperature, rate
432 of precipitation and the trajectory of humid air masses (Craig and Gordon, 1965; Rozanski et
433 al., 1993; Gat, 1996). During the last two millennia, the climatic moisture delivered through
434 Europe has been driven by the North Atlantic Oscillation (NAO) (Smith et al., 2016) and it is
435 reasonable to consider that there has not been any major change in the balance between the
436 water cycle reservoirs for this interval of time (Darling, 2006). For mid- to high-latitudes,
437 rough linear correlations are observed between the $\delta^{18}\text{O}$ of meteoric water and air temperature
438 (e.g. Dansgaard, 1964; Rozanski et al., 1992; Fricke and O'Neil, 1999). Such correlations
439 result from isotopic compositions that obey a Rayleigh distillation law in both evolving
440 reservoirs of vapor water and condensed water. Condensation in clouds globally proceeds
441 from warm low latitudes toward cold high latitudes. Air temperature in the residual clouds
442 tends to decrease resulting in an increasing isotopic difference between precipitation and
443 water vapor.

444 For Europe, the slope between $\delta^{18}\text{O}_{\text{mw}}$ and Mean Air Temperature (MAT) is considered
445 to be 0.58‰ per °C (Rozanski et al., 1992). We combine the $\delta^{18}\text{O}_w$ (now assumed to be local
446 $\delta^{18}\text{O}_{\text{mw}}$) obtained from measurements performed on individuals from Thézy-Glimont to
447 relationships between $\delta^{18}\text{O}_{\text{mw}}$ and MATs. Several relationships (Dansgaard, 1964; Rozanski
448 et al., 1992; Fricke and O’Neil, 1999; von Grafenstein et al., 1996; Lécuyer, 2013) are
449 available in the literature. These different equations may provide sizeable differences in the
450 calculated air temperature (Daux et al., 2005). We chose two equations we considered the
451 most appropriate in the framework of this study. First, the equation provided by von
452 Grafenstein et al. (1996), which has been established at Lake Ammersee, Bavaria, Germany
453 (48°00’N - 11°07’E), only located $\approx 2^\circ$ of latitude south of Thézy-Glimont. The equation is
454 expressed as follows:

$$MAT (^{\circ}C) = \frac{\delta^{18}O_{mw} + 14.48}{0.58 (\pm 0.11)} \quad (1)$$

456
457 Secondly, we also considered the isotope fractionation equation established by Lécuyer
458 (2013) as it has been constructed from European meteorological stations excluding those from
459 the Mediterranean basin, which is under the influence of a specific climatic mode (Lécuyer et
460 al., 2018). The equation is expressed as follows:

$$MAT (^{\circ}C) = 1.41 (\pm 0.09) \times \delta^{18}O_{mw} + 22.02 (\pm 0.83) \quad (2)$$

461
462
463
464 A reconstruction of mean annual $\delta^{18}\text{O}_{\text{mw}}$ value at the location of Thézy-Glimont (49°N,
465 2°E) using The Online Isotopes in Precipitation Calculator (Bowen and Revenaugh, 2003;
466 IAEA/WMO, 2015; Bowen, 2020) gives a value of $-6.7 \pm 0.2\text{‰}$, which is very close to the

467 April (-6.86‰) and October (-7.00‰) $\delta^{18}\text{O}_w$ values measured on the Avre river by Négrel
468 and Petelet-Giraud, (2005). We emphasize that at mid latitudes, sinusoidal-like seasonality in
469 meteoric water $\delta^{18}\text{O}_{\text{mw}}$ values mathematically leads to spring and autumn $\delta^{18}\text{O}_{\text{mw}}$ values
470 matching the mean annual one. Therefore, we use these inter-season oxygen isotope values to
471 test both $\delta^{18}\text{O}_{\text{mw}}$ -MAT equations for the Picardie region. The calculated temperatures are
472 $13.1\pm 1.6^\circ\text{C}$ and $12.3\pm 1.6^\circ\text{C}$ in April and $12.9\pm 1.6^\circ\text{C}$ and $12.1\pm 1.6^\circ\text{C}$ in October for equations
473 (1) and (2), respectively. The isotope data from April 2001 may be biased as floods happened
474 in the Somme department during this month, which led to an ^{18}O enrichment of the Avre river
475 water sampled for the study. As far as October data are concerned, the calculated
476 temperatures are near the average temperature of 11.6°C for this month (data from Météo
477 FranceTM). Those results indicate that both equations are reliable to infer MAT from $\delta^{18}\text{O}_{\text{mw}}$
478 and could safely be applied to the late Iron Age.

479 When applying equations (1) and (2) to the calculated $\delta^{18}\text{O}_{\text{mw}}$, the reconstituted air
480 temperatures for the late Iron Age are on average $10.5^\circ\text{C}\pm 1.7^\circ\text{C}$ for equation (1) and
481 $10.2^\circ\text{C}\pm 1.4^\circ\text{C}$ for equation (2) (Figure 5). Therefore, the results for both of these equations
482 being very close, those calculated MATs are ascribed to Thézy-Glimont during the 3rd to 2nd
483 century BCE. These calculated MAT well compare the present-day average temperature of
484 10.9°C recorded in Amiens, Picardie. Our result implies that mid-latitudes also experienced a
485 climate optimum that can be attributed to the RWP, with continental temperatures close to
486 present-days. This MAT is consistent with results obtained from other studies realized at
487 mid-latitudes, such as palynological data from Tinner et al. (2003), sedimentological data
488 from Holzhauser et al. (2005) and geochemical data obtained from speleothems by
489 McDermott et al. (2001).

490 We assign such mild air temperatures to what is considered to be the initiation of the
491 RWP. The timing of the RWP is still unsure as well as its climatic characteristics. There may

492 always remain some discrepancies in the European temperature record for the RWP because
493 no global coherent warm period has occurred on Earth (Neukom et al., 2019) excluding the
494 current climate change. Nonetheless, this study provides a new climatic anchor point which is
495 well constrained in space and time, in a region lacking climatic data for the RWP.

496

497 **5. Conclusion**

498

499 The Gallic site of Thézy-Glimont provides an insight into the climatic conditions
500 occurring during the 3rd to 2nd century BCE in the North Gaul. An oxygen isotope mapping of
501 three human skeletons revealed a great homogeneity of $\delta^{18}\text{O}_p$ values within the range of
502 analytical uncertainty. This result is of paramount importance in archaeological studies as it
503 means that, in the case of $\delta^{18}\text{O}_p$ values of sedentary populations, the use of one bone as
504 representative of the whole skeleton is justified.

505 The $\delta^{18}\text{O}_p$ of bones from Gauls and associated animals record the $\delta^{18}\text{O}$ value of drinking
506 water, which illustrates the climatic conditions prevailing in Picardie, France. The
507 reconstructed air temperatures are $10.2^\circ\text{C}\pm 1.4^\circ\text{C}$ and $10.5^\circ\text{C}\pm 1.7^\circ\text{C}$ depending on the selected
508 $\delta^{18}\text{O}_{\text{mw}}$ -MAT equation. Here, we emphasize that those estimated mean air temperatures are
509 similar to present-day ones. Mild air temperatures recorded in Picardie, France, during the
510 beginning of the RWP is a new result that needs to be taken into consideration for
511 understanding the climatic regime of the RWP in Europe. This study also shows that
512 archaeological artefacts may provide quantified climatic anchor points with a high spatial and
513 temporal resolution, playing a key role in our knowledge of the Late Holocene short-term
514 climatic changes.

515

516 **Declaration of competing interest**

517

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

520

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522

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526

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874

875 **Tables**

876

877 Table 1: Grave, nature, identification and corresponding dates (at 95.4% certainty) inferred
878 from ¹⁴C dating of the samples from Thézy-Glimont. The archaeological data come from Le
879 Béchenneq et al. (2016).

880 * corresponds to the three individuals for which various bones were sampled, see Figure 2.

881 §: If there are two or three dates from the same sample, the radiocarbon dates have been
882 combined before calibration. Such a combination is checked for internal consistency by a
883 chi-squared test which is performed automatically by the OxCal™ program.

884

885 **Figures**

886

887 Figure 1: (a) Geographic map of France with the location of the Gallic site of Thézy-Glimont,
888 Picardie, France. PM = Prime Meridian. The datum used for latitude and longitude is WGS84.
889 Picture was taken from Google Earth™. (b) General plan of the archaeological site. 1: Area
890 where the excavated graves are located. 2: Intermediate area. 3: Enclosure with iron furniture
891 and weapons. In gray is represented a trench from World War 1 (1918). The background
892 image was taken from Google Maps™.

893

894 Figure 2: Sampling points and intra-skeletal $\delta^{18}\text{O}_p$ (‰, V-SMOW) variations represented as
895 local deviation from the average $\delta^{18}\text{O}_p$ value of bones coming from TG-01, TG-02 and TG-03
896 individuals.

897

898 Figure 3: $\delta^{18}\text{O}_p$ (‰, V-SMOW) values of bones (blue squares) and teeth (red triangles) from
899 humans and animals of Thézy-Glimont. Bones and teeth which are aligned vertically
900 correspond to a same individual. Humans, bovids and horses are represented by increasing
901 sample number from left to right.

902

903 Figure 4: Comparison of oxygen isotope fractionation equations between apatite phosphate
904 ($\delta^{18}\text{O}_p$, ‰, V-SMOW) and drinking water ($\delta^{18}\text{O}_w$, ‰, V-SMOW) proposed by Daux et al.

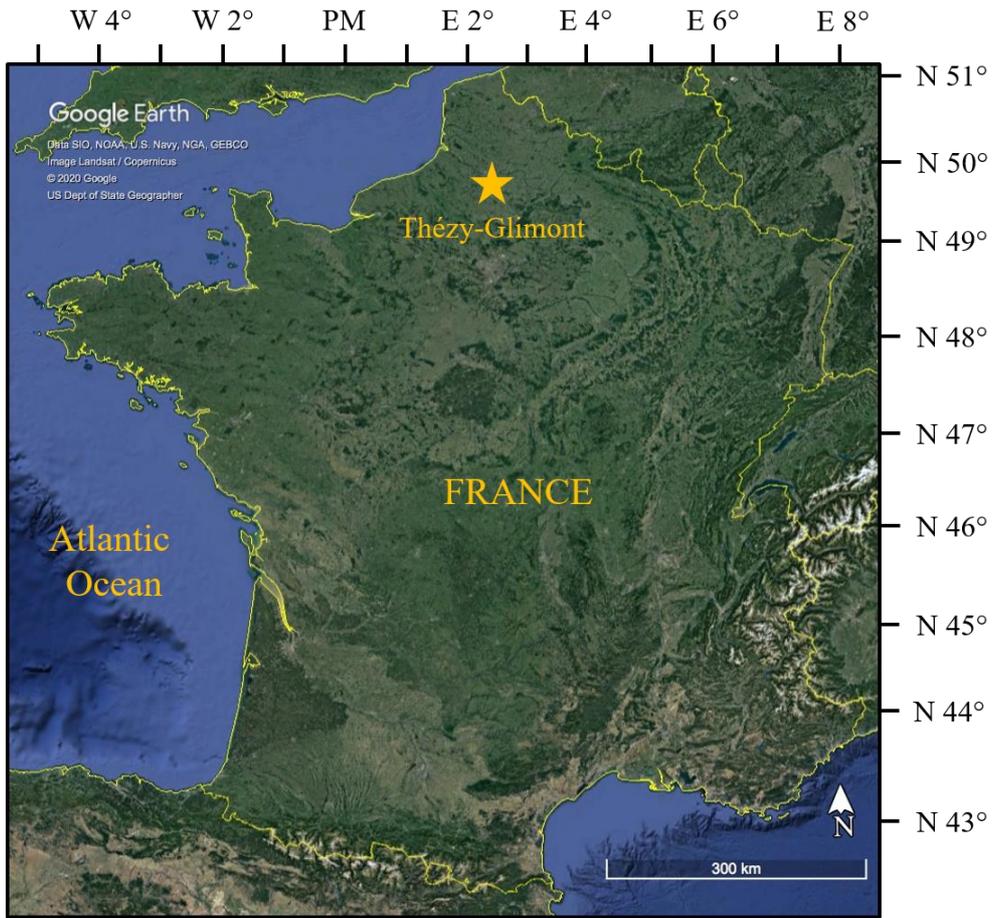
905 (2008) for humans, Longinelli (1984) for pigs, D'Angela and Longinelli (1990) for bovids
906 and Bryant et al. (1994) for horses.

907

908 Figure 5: $\delta^{18}\text{O}_w$ values (‰, V-SMOW) of drinking water inferred from the bones of humans
909 and animals from Thézy-Glimont. Humans and bovids are represented by increasing sample
910 number from left to right. Error bars represent the total uncertainties. The red dashed lines
911 represent the modern $\delta^{18}\text{O}_w$ values measured in the Avre river by Négrel and Petelet-Giraud
912 (2005) in April (A) and October (O) 2001.

913

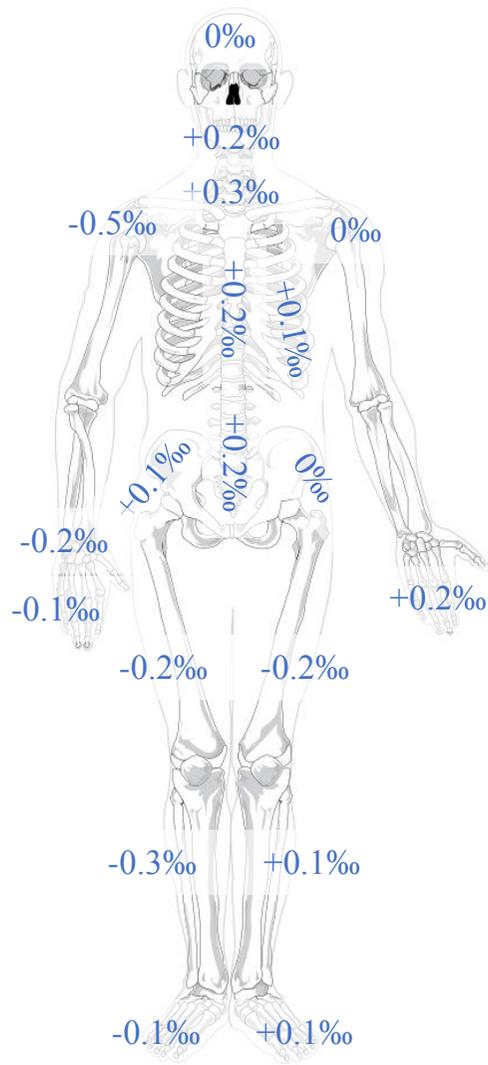
914 Figure 6: Reconstructed Mean Annual Temperatures (MAT) inferred from the oxygen isotope
915 composition of drinking water, using the oxygen isotope fractionation equation of von
916 Grafenstein et al. (1996) (blue squares) and Lécuyer (2013) (red triangles). Humans and
917 bovids are represented by increasing sample number from left to right. The horizontal dashed
918 line corresponds to the modern yearly MAT in Picardie, France, averaged throughout the
919 years 1980 to 2010 (data from Météo FranceTM).



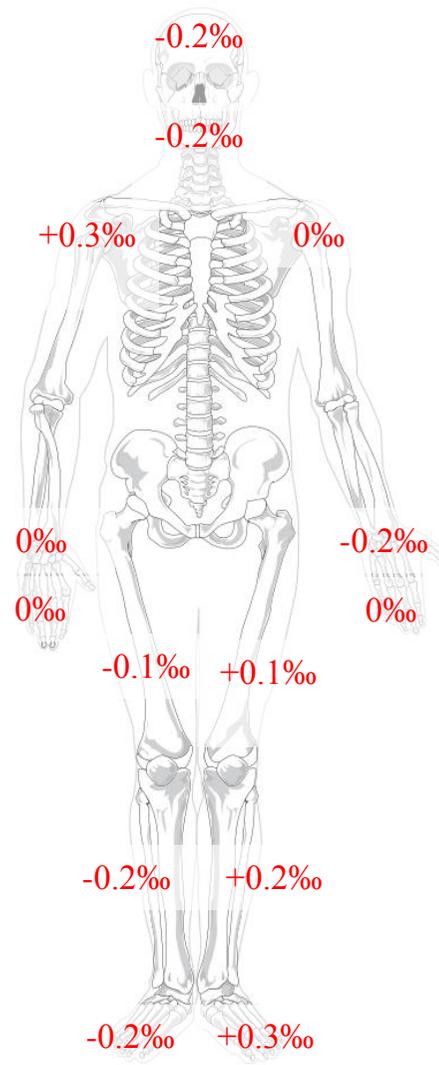
(a)



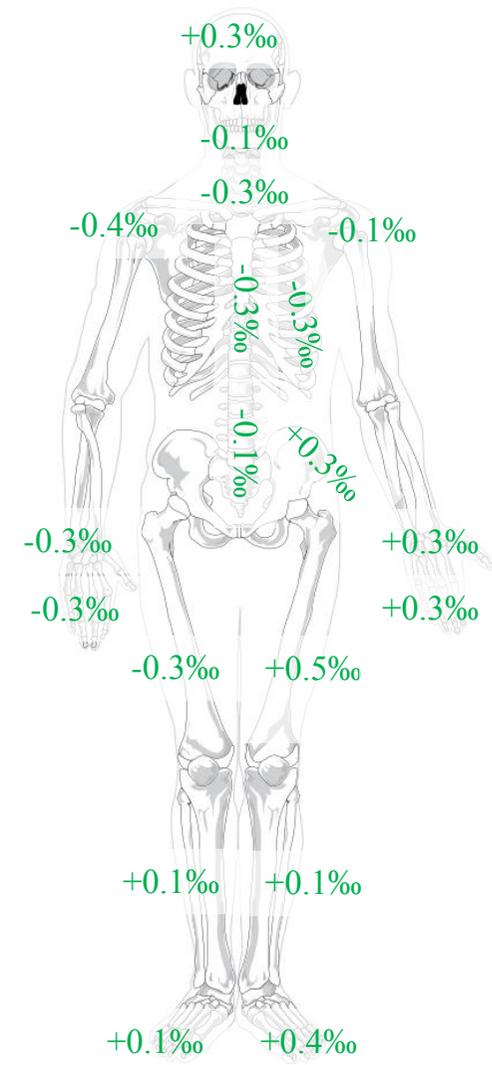
(b)



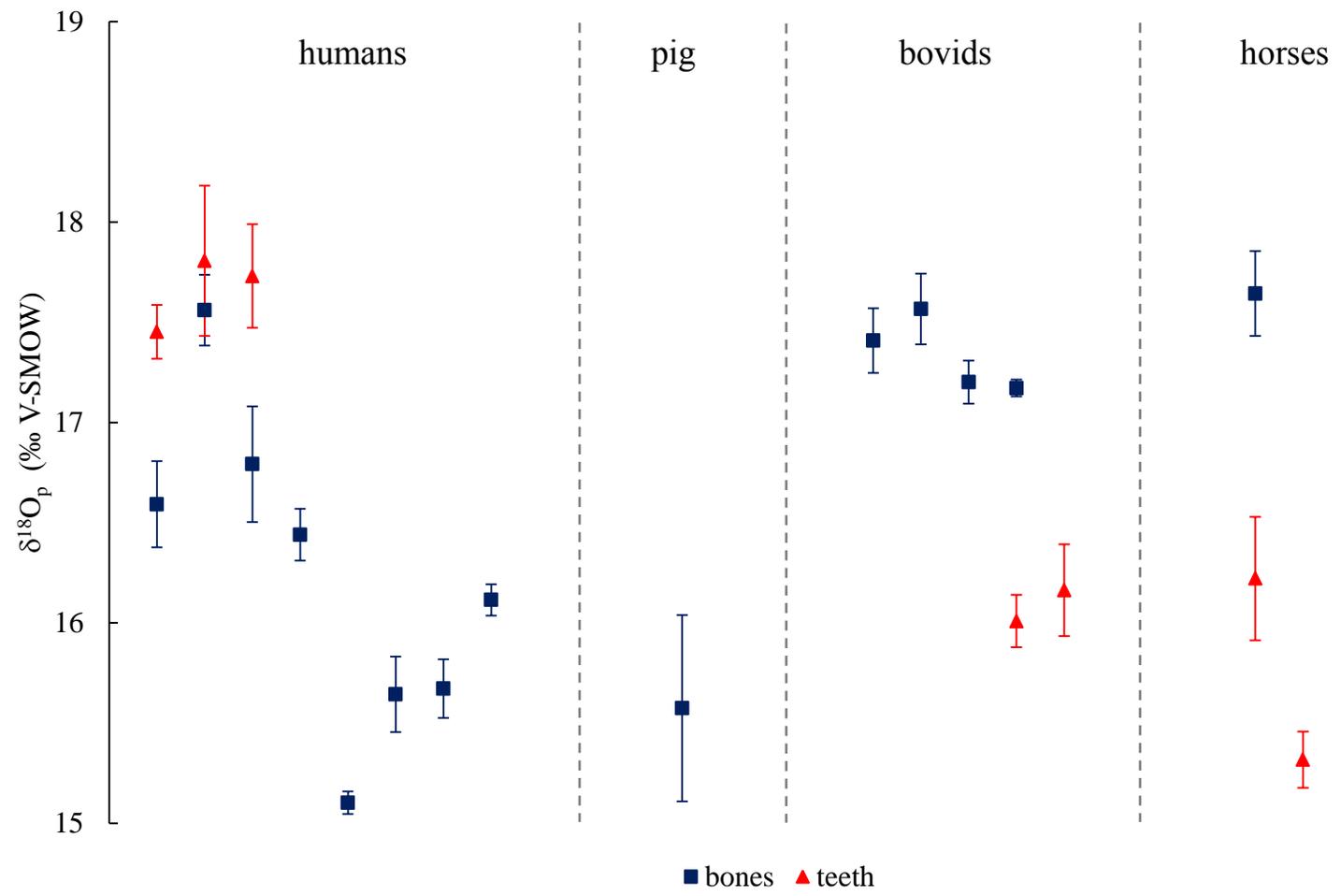
TG-01
 Mean $\delta^{18}\text{O}_p$ (V-SMOW): 16.6‰
 Standard Deviation: 0.2‰

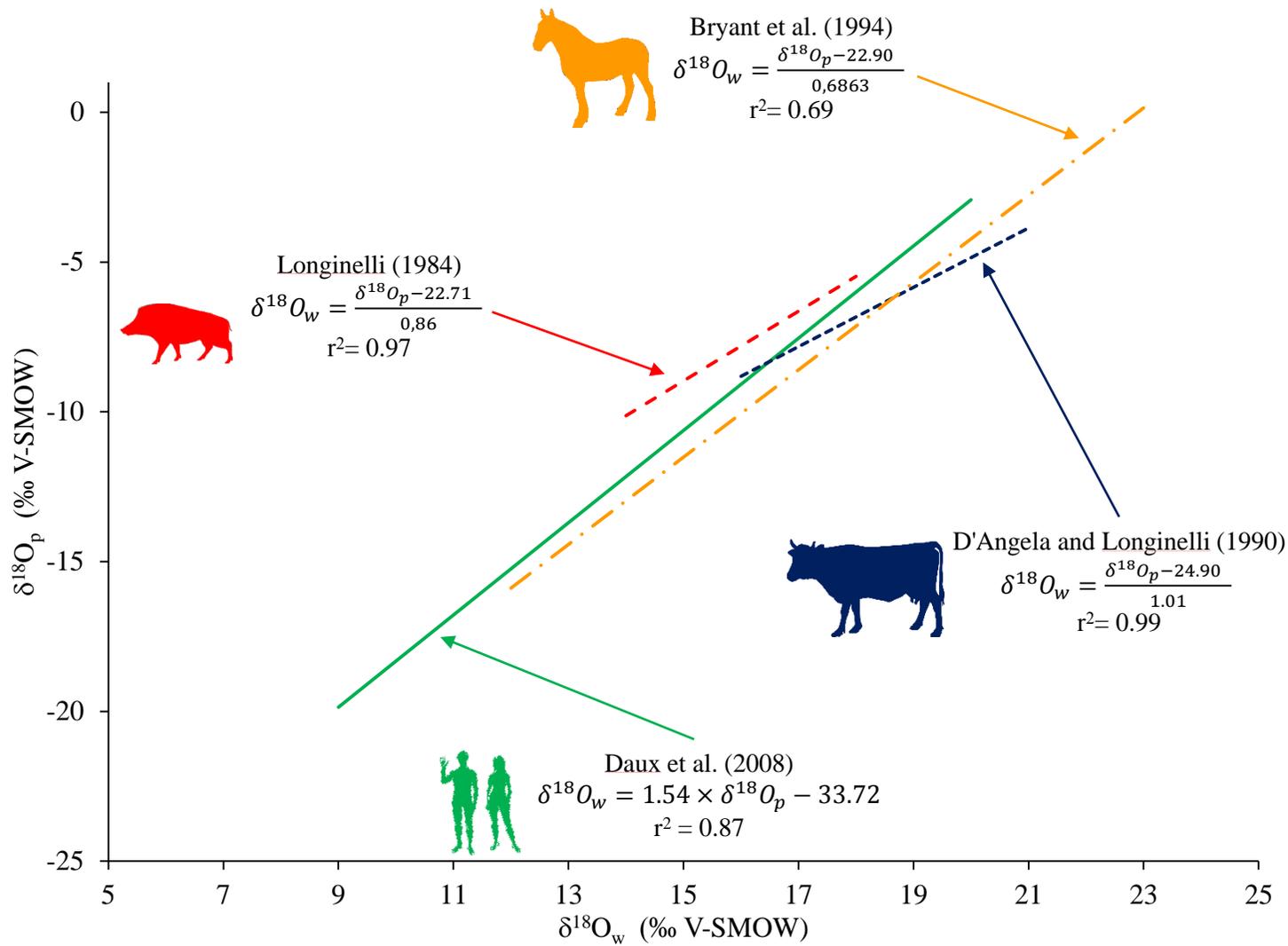


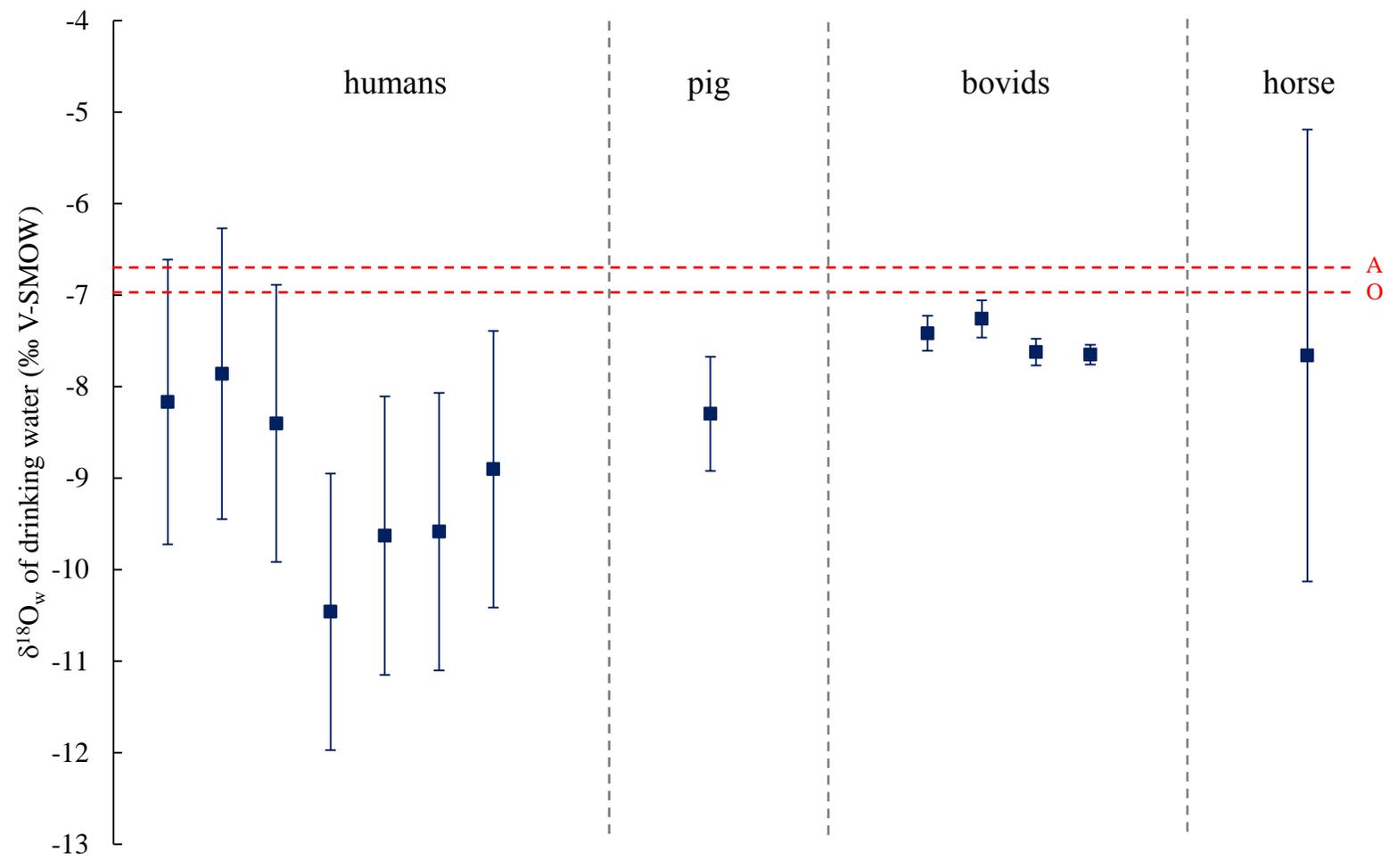
TG-02
 Mean $\delta^{18}\text{O}_p$ (V-SMOW): 17.6‰
 Standard Deviation: 0.2‰

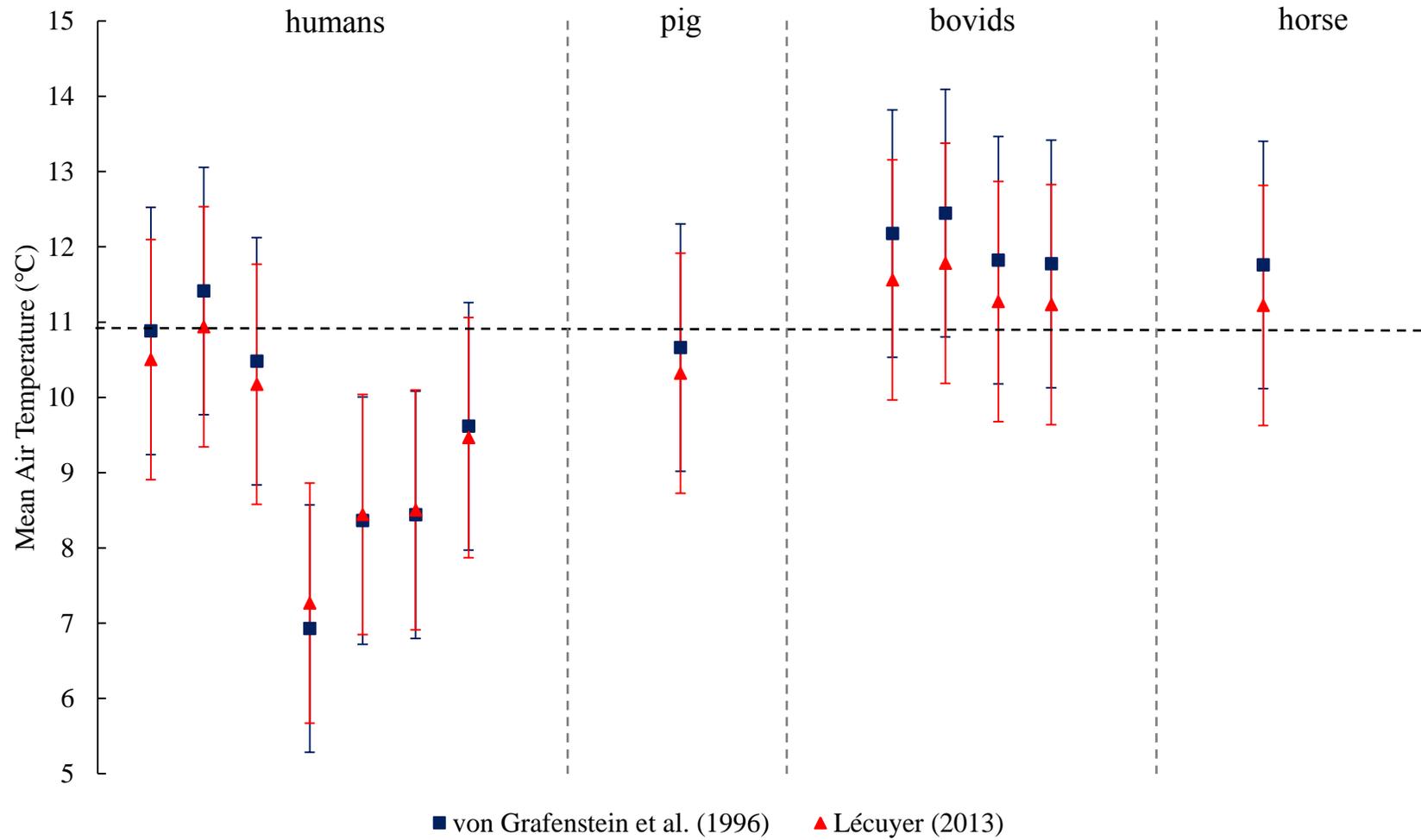


TG-03
 Mean $\delta^{18}\text{O}_p$ (V-SMOW): 16.8‰
 Standard Deviation: 0.3‰









Grave	Individual	Type	Sex	Age (years)	Sample type	¹⁴ C dates (BP)	Calibrated dates - 2σ (95.4%) [§]
TGV 1135	TG-01	<i>Homo sapiens</i>	Male	< 30	*	2160 ± 30 2190 ± 30	358 cal BCE (56.7%) 280 cal BCE 258 cal BCE (2.6%) 243 cal BCE 236 cal BCE (36.1%) 170 cal BCE
					1st premolar (PM1)	-	-
					1st molar (M1)	-	-
					2 nd molar (M2)	-	-
TG-CH1	<i>Equus caballus</i>	Male	3	Left phalanx	2130 ± 30	350 cal BCE (10.5%) 308 cal BCE 210 cal BCE (84.9%) 52 cal BCE	
				3rd molar (M3)	2150 ± 30 2180 ± 30 2190 ± 30	356 cal BCE (59.0%) 286 cal BCE 234 cal BCE (36.4%) 170 cal BCE	
TG-B1	<i>Bos taurus</i>	Female	4	Right phalanx	-	-	
TG-B2	<i>Bos taurus</i>	Male	young	Left talus	-	-	
TG-B3	<i>Bos taurus</i>	Male	2	Left phalanx	-	-	
TG-B4	<i>Bos taurus</i>	Male	4	Left phalanx	-	-	
TG-CO1	<i>Sus scrofa</i>	Male	-	3rd molar (M3)	2170 ± 30 2205 ± 30	359 cal BCE (59.2%) 274 cal BCE 261 cal BCE (36.2%) 184 cal BCE	
				Right Radius	-	-	
				Rib	-	-	
				Vertebra	-	-	

TGV 1249	TG-02	<i>Homo sapiens</i>	Male	15	*	2200 ± 30 2170 ± 30	358 cal BCE (59.2%) 276 cal BCE 260 cal BCE (36.2%) 180 cal BCE
					1st premolar (PM1)	-	-
					1st molar (M1)	-	-
					3rd molar (M3)	-	-
TGV 1450	TG-03	<i>Homo sapiens</i>	Male	Adult	*	2185 ± 30	361 cal BCE (95.4%) 172 cal BCE
					1st premolar (PM1)	-	-
					2nd molar (M2)	-	-
					3rd molar (M3)	-	-
TGV 1219	TG-04	<i>Homo sapiens</i>	Male	Adult	Right femur	2195 ± 30 2175 ± 30	358 cal BCE (59.0%) 276 cal BCE 260 cal BCE (36.4%) 180 cal BCE
TGV 1138	TG-05	<i>Homo sapiens</i>	Male	30 to 59	Right femur	2180 ± 30 2145 ± 30	356 cal BCE (46.7%) 284 cal BCE 256 cal BCE (0.5%) 250 cal BCE 235 cal BCE (46.0%) 156 cal BCE 134 cal BCE (2.1%) 116 cal BCE
	TG-CH2	<i>Equus caballus</i>	Male	5	Molar	2120 ± 30 2135 ± 30	344 cal BCE (4.3%) 322 cal BCE 206 cal BCE (87.9%) 88 cal BCE 76 cal BCE (3.2%) 58 cal BCE
TGV 1140	TG-06	<i>Homo sapiens</i>	Male	Adult	Right femur	2170 ± 30 2145 ± 30	355 cal BCE (40.5%) 289 cal BCE 232 cal BCE (51.0%) 151 cal BCE 136 cal BCE (3.9%) 114 cal BCE
TGV1206	TG-07	<i>Homo sapiens</i>	Male	Adult	Right femur	2200 ± 30 2225 ± 30	364 cal BCE (95.4%) 203 cal BCE
TGV 1246	TG-08	<i>Homo sapiens</i>	Male	Adult	Tibia	-	-

TGV 1222	TG-B5	<i>Bos taurus</i>	Male	2 to 3	3rd molar	2135 ± 30 2160 ± 30 2115 ± 30	346 cal BCE (8.2%) 320 cal BCE 206 cal BCE (87.2%) 102 cal BCE
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