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

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Abstract (190 words)

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

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

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

1. Introduction

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

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

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

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

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

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

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

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

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

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

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

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

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

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

2. Material and methods

2.1. The Gallic site of Thézy-Glimont

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

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

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

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

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

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

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

178 powder using a Pulverisette 7™ 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 Ultrapure™ water and cleaned using a toothbrush. The powder was obtained by drilling the
181 enamel of the teeth with a Dremel™ diamond-head drill bit.
182
183 *2.2. Dating of the archaeological site*
184
185 During the excavation of the archaeological site, Beta Analytics™ realized ^{14}C datings
186 on 13 samples, giving dates with 68.3% probability (1σ) between 200 BCE and 5 AD. New
187 ^{14}C 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 (2σ).
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

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

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

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

2.4. Oxygen isotope ratios of apatite phosphate

Oxygen isotopic analyses were performed with stable isotope mass spectrometry techniques. Phosphate ions were extracted from the apatite using a wet chemistry procedure. After the dissolution of about 30 mg of each sample in 2 mL of 2M HF at ambient temperature for 24 hours, the CaF_2 precipitate was separated from the sample solution by centrifugation. The phosphate ions were neutralized with KOH and isolated using a strong anionic resin (Amberlite IRN 78) before being precipitated as silver phosphate following the method first described by Crowson et al. (1991) and slightly modified by Lécuyer et al. (1993). The oxygen isotope data were obtained using a high-temperature pyrolysis continuous flow technique developed by Lécuyer et al. (2007) and Fourel et al. (2011). For each sample, five aliquots of 400–500 µg of silver phosphate were mixed with 500 µg of carbon loaded into 3.5x5 mm silver foil isotope grade capsules (Eurovector™). Pyrolysis was performed at 1450°C using a PYROcube™ elemental analyzer (Elementar GmbH-Germany) interfaced in continuous flow mode with an isotopic ratio mass spectrometer Isoprime™ (Elementar UK Ltd UK). Measurements were performed with CO gas separated by purge and trap technology within the PYROcube™ system at the Laboratoire de Géologie de Lyon and at the Laboratoire d'Ecologie des Hydrosystèmes Naturels et Anthropisés, both hosted by the Université Claude Bernard Lyon 1. The official unit for isotopic ratios is the Urey (Ur), although the unit commonly used is the permil (‰, 1‰ = 1mUr). For practical reasons, data are reported as $\delta^{18}O$ values with respect to V-SMOW in ‰ units.

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

3. Results

3.1. Intra-skeletal variability of the $\delta^{18}O_p$ from human bones

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

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

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

3.2. Oxygen isotope values of phosphatic tissues

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

4. Discussion

4.1. Sample preservation and diagenesis

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

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

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

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

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

4.3. The $\delta^{18}O_p$ record in teeth and bones from individuals of Thézy-Glimont

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

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

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

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

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

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

Daux et al. (2008) for humans:

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

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

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

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 ±0.2‰ for the equation of Longinelli (1984) and ±0.1‰ 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

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

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

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

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

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

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

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

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

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

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

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

5. Conclusion

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

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

Declaration of competing interest

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

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Tables

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

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

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

Figures

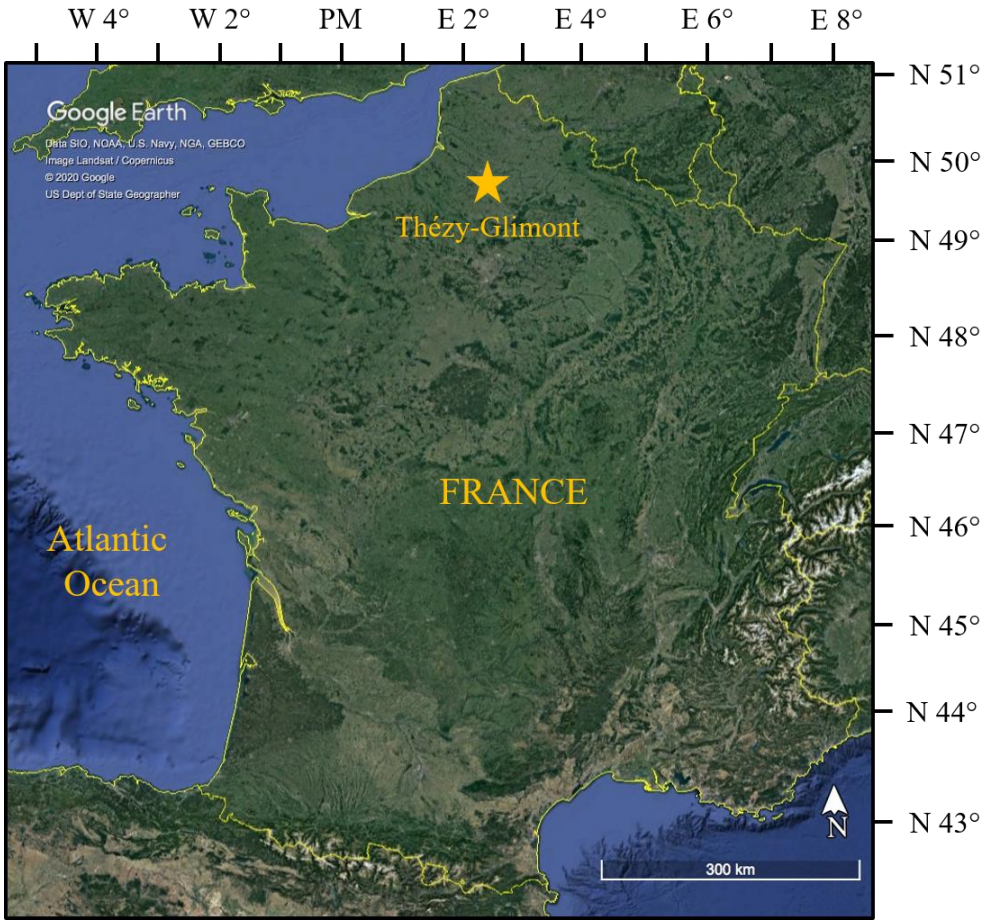
Figure 1: (a) Geographic map of France with the location of the Gallic site of Thézy-Glimont, Picardie, France. PM = Prime Meridian. The datum used for latitude and longitude is WGS84. Picture was taken from Google Earth™. (b) General plan of the archaeological site. 1: Area where the excavated graves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

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

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

Figure 4: Comparison of oxygen isotope fractionation equations between apatite phosphate ($\delta^{18}O_p$, ‰, V-SMOW) and drinking water ($\delta^{18}O_w$, ‰, 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: $\delta^{18}O_w$ 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 $\delta^{18}O_w$ 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 France™).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Google Earth
Data SIO, NOAA, U.S. Navy, NGA, GEBCO
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© 2020 Google
US Dept of State Geographer
Thézy-Glimont
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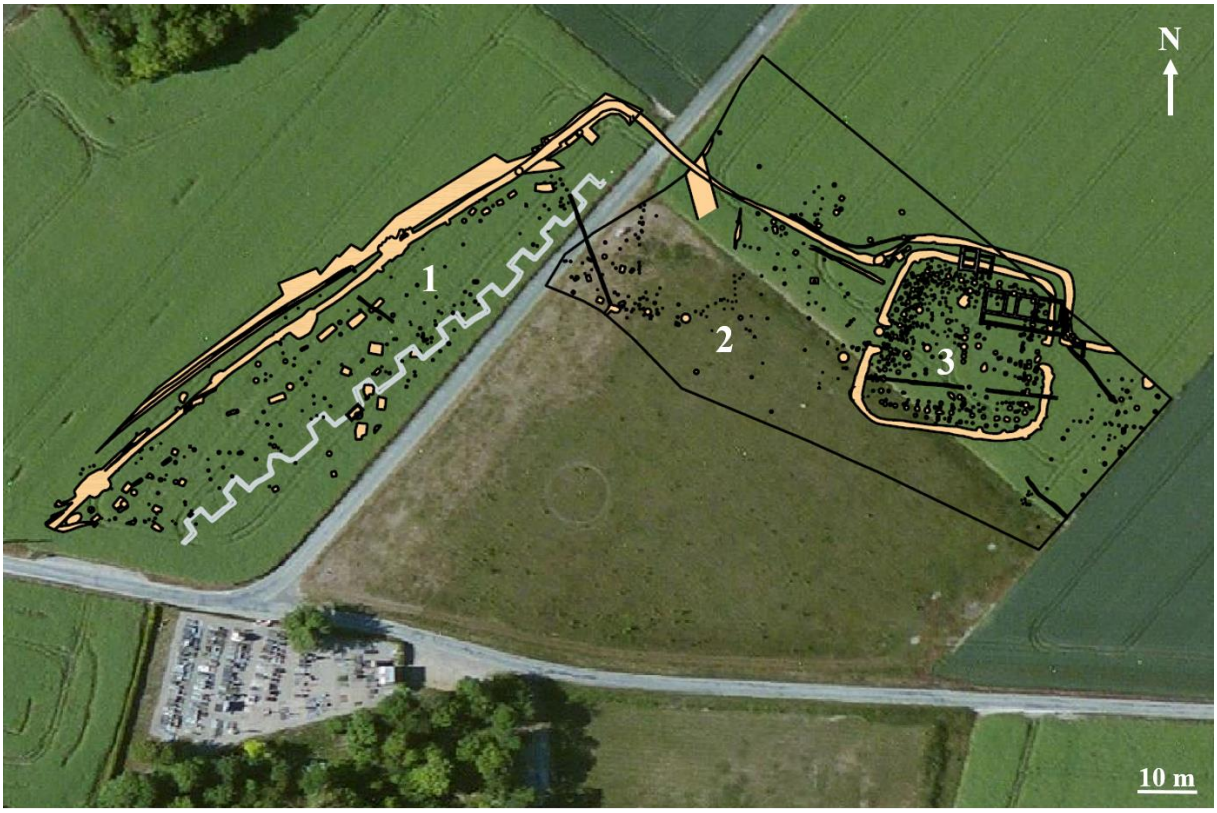
(a)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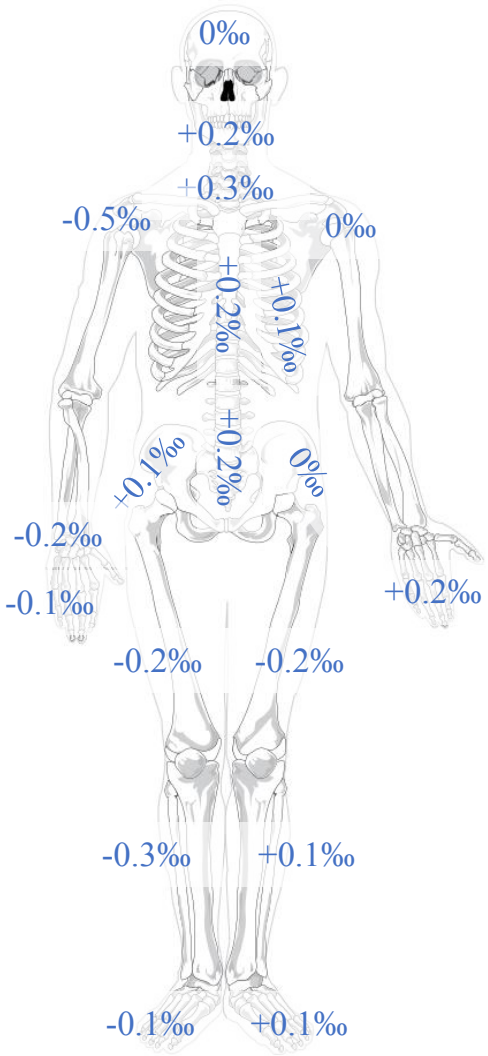
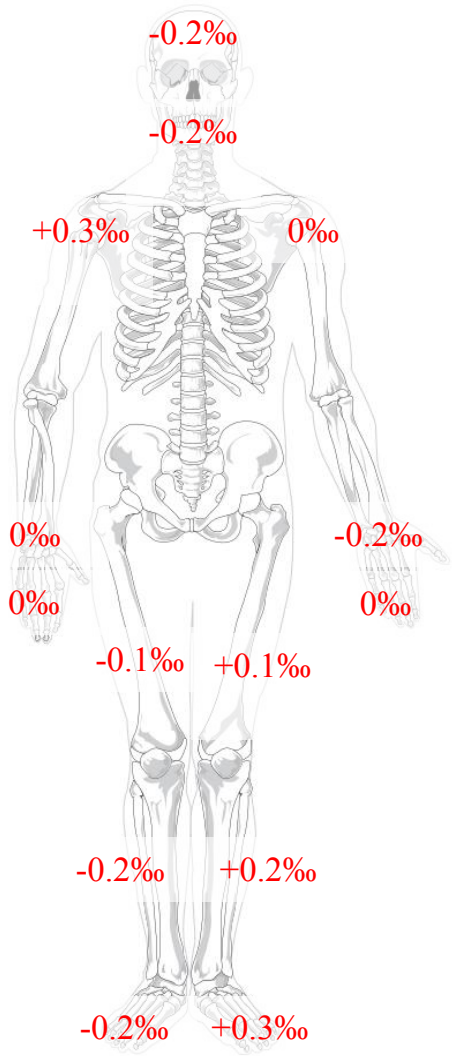


TG-01

Mean $\delta^{18}O_p$ (V-SMOW): 16.6‰

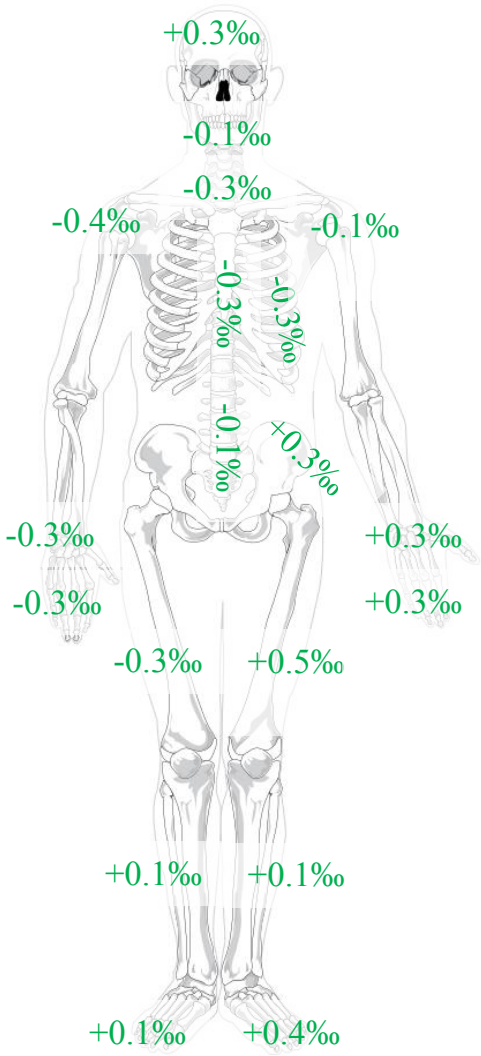
Standard Deviation: 0.2‰



TG-02

Mean $\delta^{18}O_p$ (V-SMOW): 17.6‰

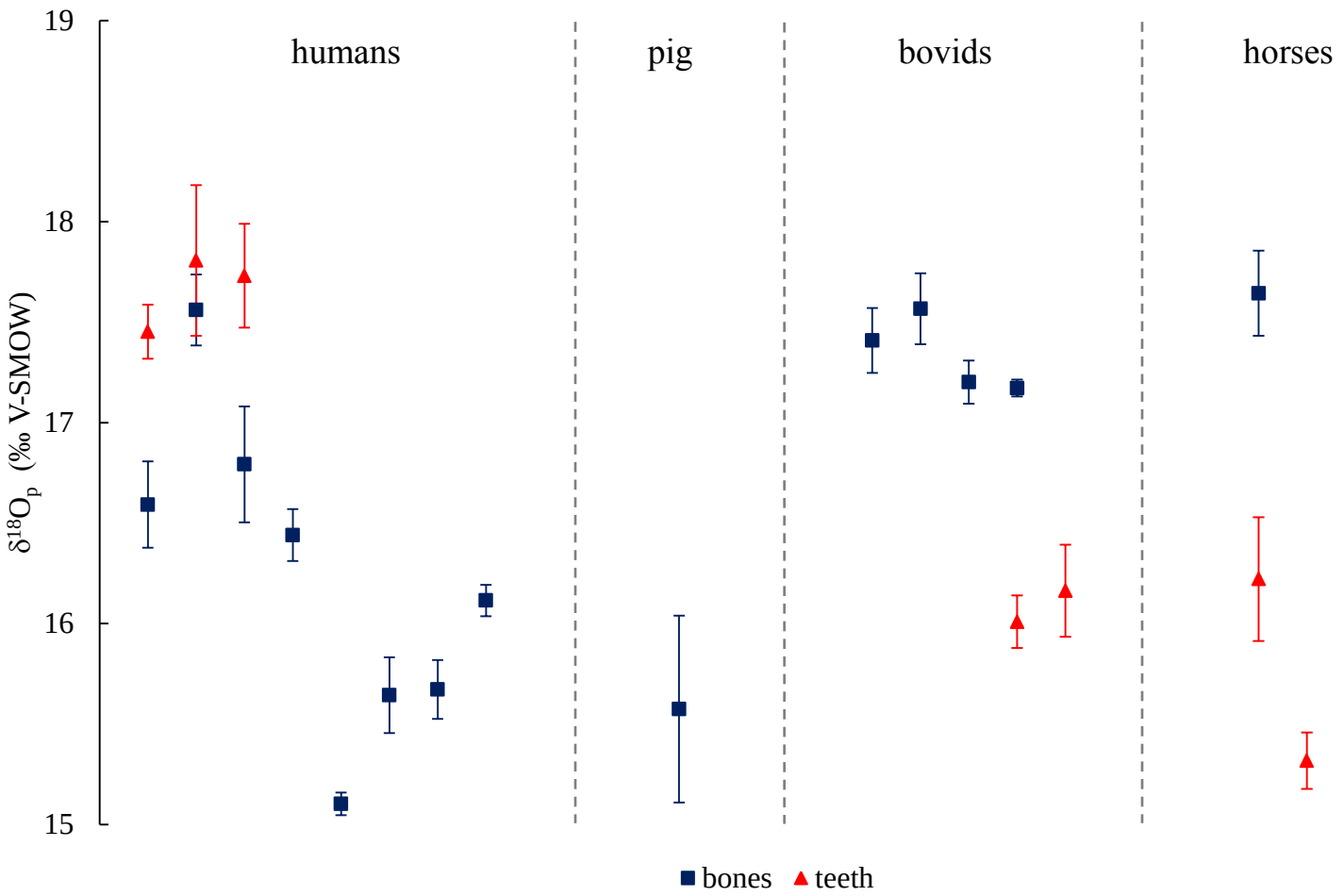
Standard Deviation: 0.2‰



TG-03

Mean $\delta^{18}O_p$ (V-SMOW): 16.8‰

Standard Deviation: 0.3‰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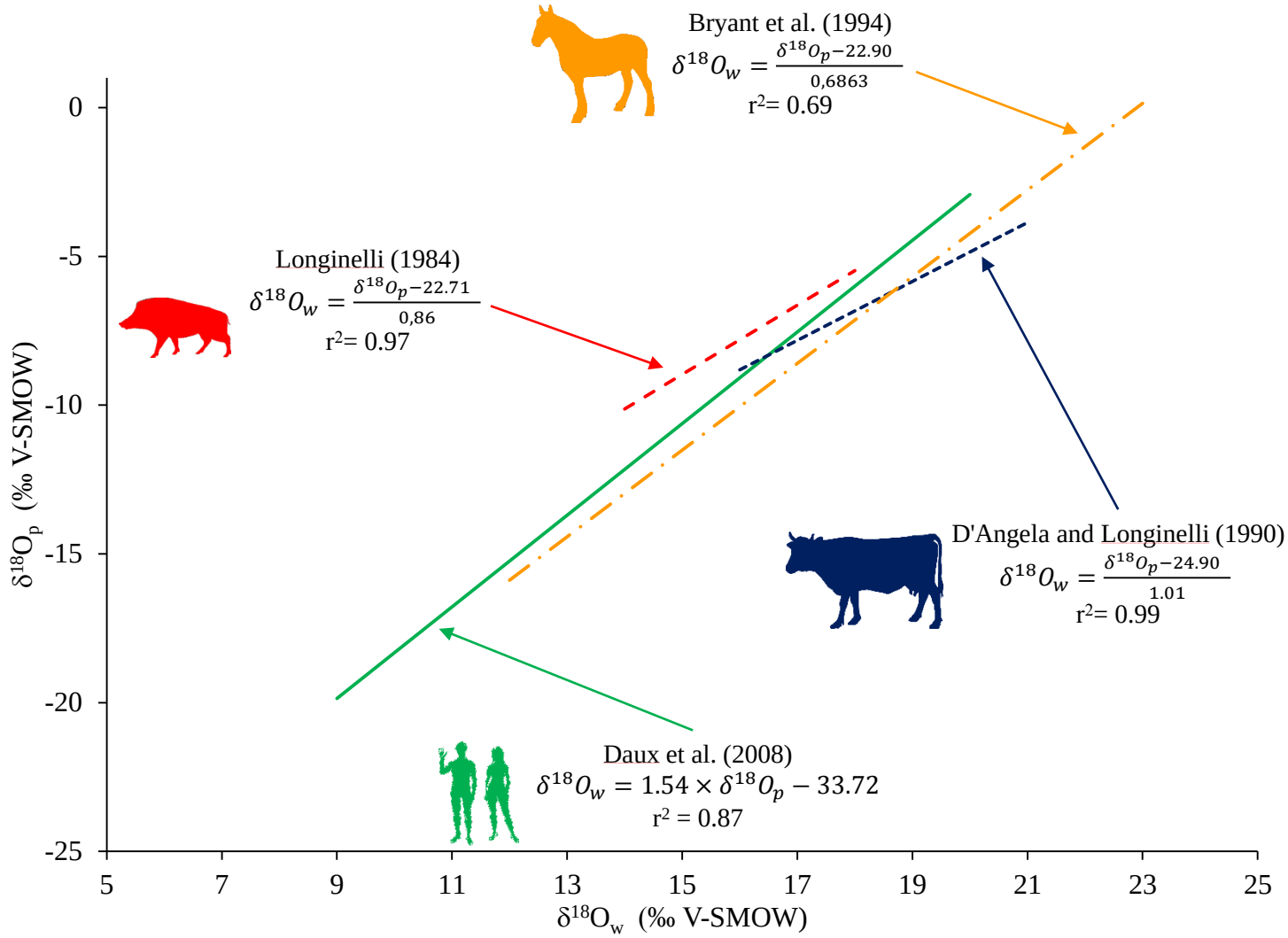
humans
pig
bovids
horses
δ18Op (‰ V-SMOW)
bones
teeth


Bryant et al. (1994)
$\delta^{18}O_w = \frac{\delta^{18}O_p - 22.90}{0,6863}$
r²= 0.69
Longinelli (1984)
$\delta^{18}O_w = \frac{\delta^{18}O_p - 22.71}{0,86}$
r²= 0.97
D'Angela and Longinelli (1990)
$\delta^{18}O_w = \frac{\delta^{18}O_p - 24.90}{1.01}$
r²= 0.99
Daux et al. (2008)
$\delta^{18}O_w = 1.54 \times \delta^{18}O_p - 33.72$
r² = 0.87
0
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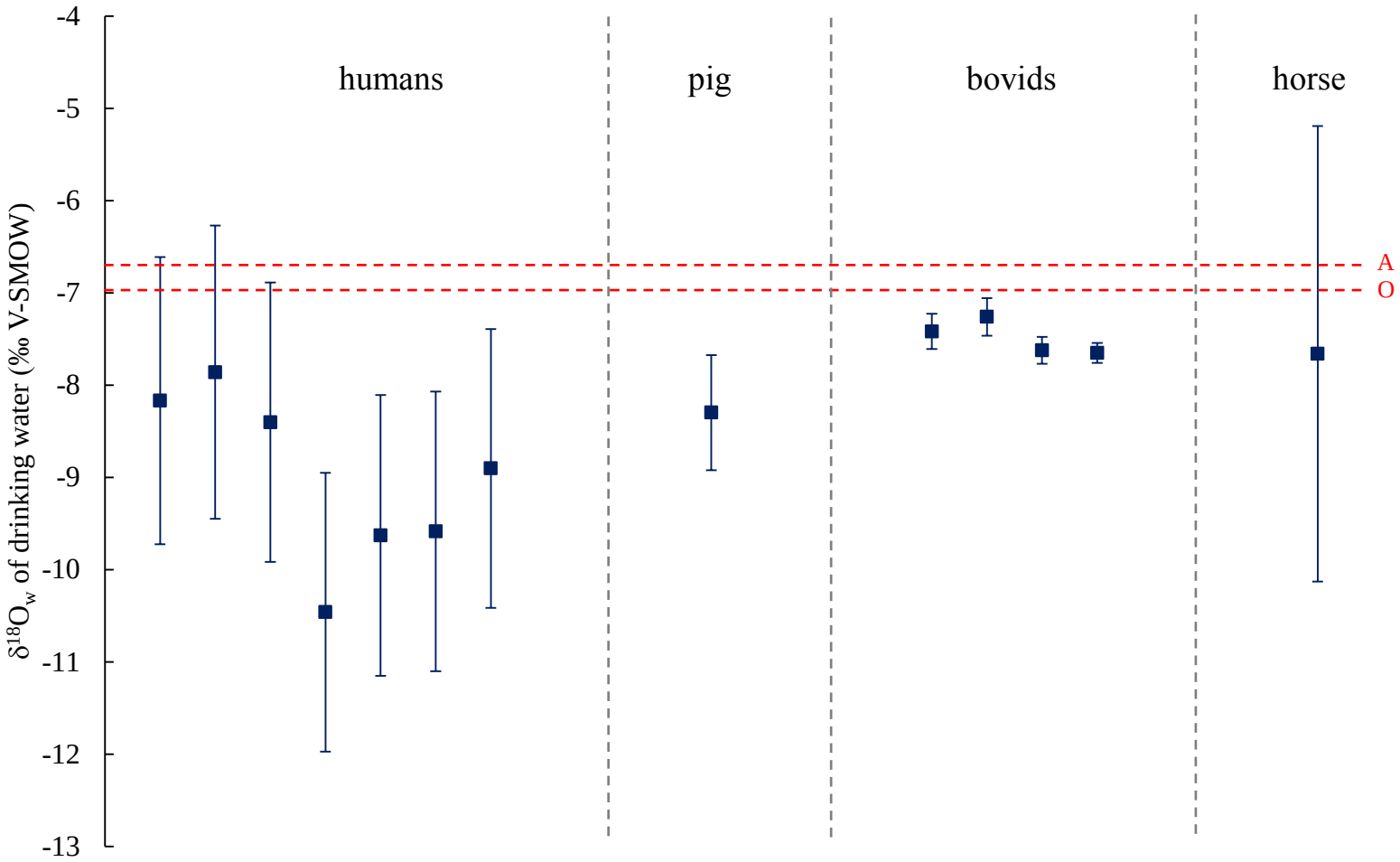
δ¹⁸O_p (‰ V-SMOW)
δ¹⁸O_w (‰ V-SMOW)


humans
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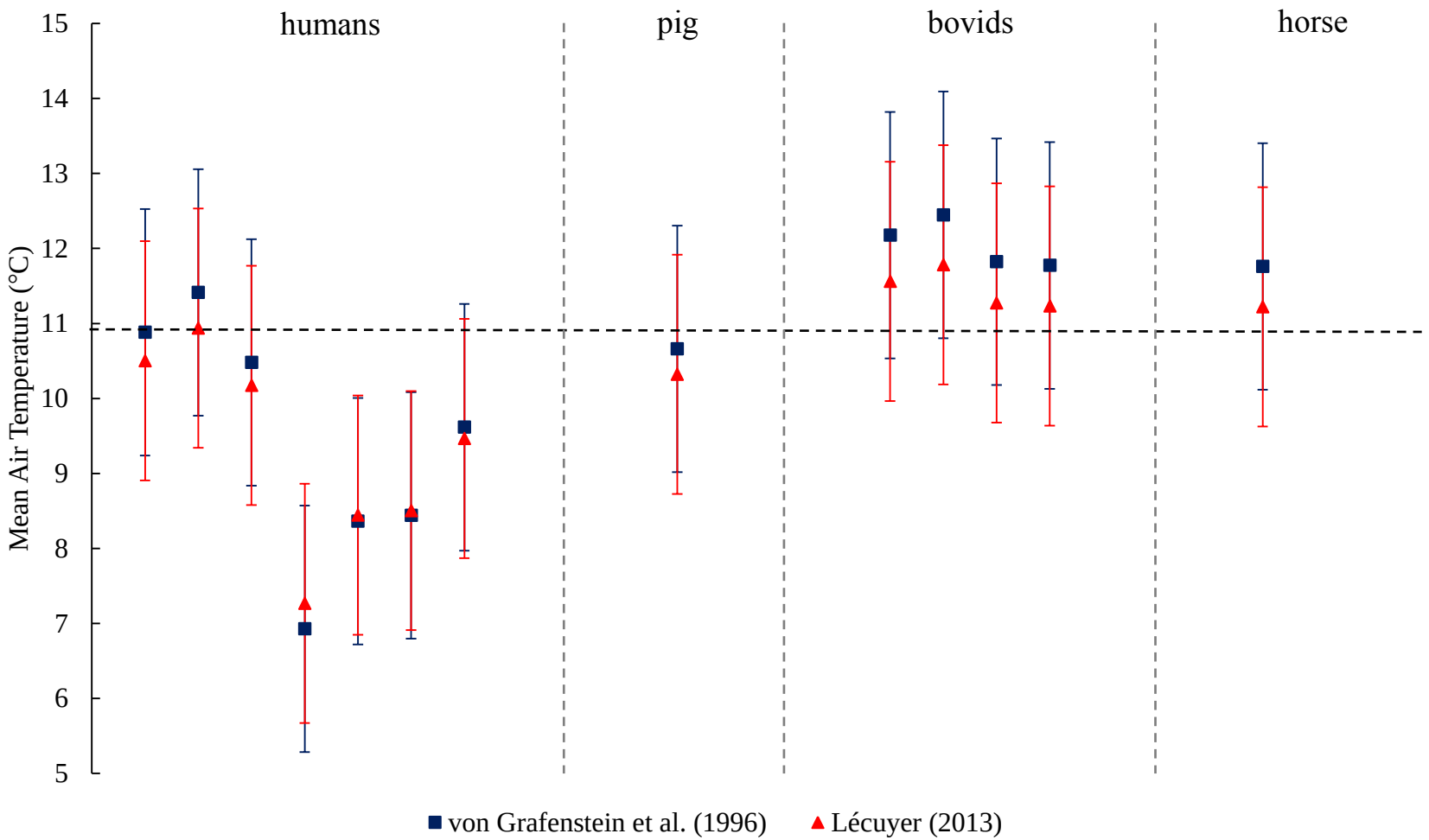
δ18Ow of drinking water (‰ V-SMOW)


humans
pig
bovids
horse
Mean Air Temperature (°C)
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von Grafenstein et al. (1996)
Lécuyer (2013)

Grave	Individual	Type	Sex	Age (years)	Sample type	^{14}C dates (BP)	Calibrated dates - 2σ (95.4%)[§]
TGV 1135	TG-01	*Homo sapiens*	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)	-	-
					2nd molar (M2)	-	-
	TG-CH1	*Equus caballus*	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	*Bos taurus*	Female	4	Right phalanx	-	-
	TG-B2	*Bos taurus*	Male	young	Left talus	-	-
	TG-B3	*Bos taurus*	Male	2	Left phalanx	-	-
	TG-B4	*Bos taurus*	Male	4	Left phalanx	-	-
					3rd molar (M3)	2170 ± 30 2205 ± 30	359 cal BCE (59.2%) 274 cal BCE 261 cal BCE (36.2%) 184 cal BCE
	TG-CO1	*Sus scrofa*	Male	-	Right Radius	-	-
					Rib	-	-
					Vertebra	-	-

TGV 1249	TG-02	*Homo sapiens*	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	*Homo sapiens*	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	*Homo sapiens*	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	*Homo sapiens*	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	*Equus caballus*	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	*Homo sapiens*	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	*Homo sapiens*	Male	Adult	Right femur	2200 ± 30 2225 ± 30	364 cal BCE (95.4%) 203 cal BCE
TGV 1246	TG-08	*Homo sapiens*	Male	Adult	Tibia	-	-

TGV 1222	TG-B5	*Bos taurus*	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