1   **A Middle Pleistocene intense monsoonal episode from the Kapthurin Formation, Kenya: Stable**
2   **isotopic evidence from bovid teeth and pedogenic carbonates**

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25   **Abstract**

26   Carbon and oxygen isotope values of bovid tooth enamel and pedogenic carbonates were used to
27   reconstruct the paleoenvironments associated with Middle Pleistocene archaeological and
28   paleontological sites from the Kapthurin Formation, Kenya. Fossil teeth and pedogenic carbonates
29   were collected from lacustrine and fluvial facies (K3 and K3') of the Kapthurin Formation (~543 – 509
30   ka). Twenty teeth from six bovid tribes and twenty four pedogenic carbonates were sequentially
31   sampled for stable oxygen and carbon analysis. The primary aims of this work are to reconstruct the
32   vegetative habitats of bovids and environmental landscapes from the Kapthurin Formation, assessing
33   the presence of C₃, C₄, or mixed vegetation and foragers. We also assess general paleoclimate factors,
34   such as overall aridity, seasonal shifts in water availability, and broad habitat reconstructions. Carbon
35   isotope values suggest a wide range of foraging strategies available to bovids, and are characterized by
36   both C₃ dominated and C₄ dominated diets, with little evidence for mixed feeders, while pedogenic
37   carbonates suggest the landscape combined a mix of C₃ and C₄ vegetation sources, structurally similar
38   to wooded grasslands. Paleoclimatic conditions, based on the reconstructed mean annual water deficit
39   from $\delta^{18}\text{O}_{\text{enamel}}$ values suggest a hyper mesic environment, most comparable to the modern Ituri
40   rainforest, indicating a much wetter climate than the modern Lake Baringo Basin. This environment
41   coincides with an intensified Middle Pleistocene African monsoonal system at onset of Marine Isotope
42   Stage 13 (~533 ka). The Middle Pleistocene Lake Baringo regional habitat appears to have been one
43   dominated by woodland areas, perennially active streams, marsh and sumpland environments, and a
44   surrounding grassland ecosystem. This lush basin would have provided hominins with a diverse and
45   productive ecosystem. Perhaps not surprisingly, archaeological sites are found in all habitats, including
46   wet, forested, and grassland environments.

47   **Key Words**

48   Middle Pleistocene, Aridity, Paleoenvironment, Marine Isotope Stage 13

49   **1. Introduction**

50   Changes in environments have long been associated with the evolution and technological
51 choices of hominins (Ambrose 1998; Bräuer 2008; Darwin 1871; deMenocal 1995; Potts 1996; Reed
52 1997; Vrba 1993). Stable carbon and oxygen ($\delta^{13}\text{C}$; $\delta^{18}\text{O}$) values derived from herbivorous mammal
53 biogenic carbonated apatite (tooth enamel) and pedogenic carbonates provide valuable data for
54 environmental reconstructions (Cerling 1984; DeNiro and Epstein 1978; Kingston, et al. 1994; Koch, et
55 al. 1989; Levin, et al. 2011; Quade, et al. 1992). The stable isotope values of tooth enamel are
56 independent of taxonomic identification of fossil specimens, and provide a useful way to confirm or
57 falsify environmental reconstructions based on the habitat preferences of modern analogs. Despite
58 potential taphonomic biases, Western and Behrensmeyer (2009) suggest that modern bone assemblages
59 accurately represent the relative proportions of different taxonomic groups of living animals, implying
60 that fossils of animals are good indicators for past habitats. Unfortunately, not all habitats are equally
61 represented in the fossil record (Behrensmeyer and Hill 1988; Hill 1979) and habitats differentially
62 “fossilize”. Isotope analysis of tooth enamel can be used to verify reconstructions based on habitat
63 preferences of modern analogs and relative taxonomic abundance, because the preferences of modern
64 taxonomic groups may have differed in the past.

65   Pedogenic carbonates form at depth (>40cm) in equatorial soils at negative water balance
66 conditions through the combined contribution of plant decay and diffusion (CO_2) at equilibrium with
67 soil water (Cerling and Quade 1993). Pedogenic carbonate carbon isotope values reflect soil CO_2
68 carbon isotope values and = oxygen isotope values of pedogenic carbonates reflect soil water oxygen
69 isotope values (Cerling 1984). Thus, the isotopic composition of pedogenic carbonates can serve as a
70 useful proxy for local vegetation and hydrological habitats (Kingston 2007; Levin, et al. 2011; Quinn, et
71 al. 2013).

72   Here, we use stable isotopes of carbon and oxygen from tooth enamel to reconstruct the dietary

73 habits of fossil bovids collected at paleontological and archaeological sites from the Middle Pleistocene
74 sediments of the Kapthurin Formation, Kenya. These dietary data, associated with taxonomic
75 uniformitarian behavior of bovids (Bibi, et al. 2009), are used to reconstruct a landscape exploited by
76 Acheulian hominins and chimpanzees between 543 ± 3 ka and 509 ± 9 ka (Leakey, et al. 1969;
77 McBrearty, et al. 1996; McBrearty and Jablonski 2005). We also use stable isotopes of carbon and
78 oxygen derived from pedogenic carbonates to augment these paleoecological reconstructions. Finally,
79 we use stable isotopes from both pedogenic carbonates and tooth enamel to test the regional
80 paleoclimatic predictions (Rossignol-Strick, et al. 1998; Schefuß, et al. 2003) of an intense monsoonal
81 system affecting East Africa at ~528 ka. This research is directly relevant to comprehending the
82 paleoenvironmental context of the terminal Acheulian and the beginnings of the Middle Stone Age
83 technocomplexes, and the foraging behavior of hominins.

84   The Middle Pleistocene, 781 – 126 ka, formerly the Ionian (International Commission on
85 Stratigraphy 2014), of Africa records a significant time period in human evolution; it marks the
86 appearance of *Homo sapiens*, the development of new technologies often referred to as “modern”, and
87 the Middle Stone Age shift to the occupation of diverse habitats (Clark 1988; Henshilwood and Marean
88 2003; McBrearty and Brooks 2000). Acheulian technology has long been assumed to be relatively
89 uniform from its inception during the early Pleistocene, 2.58 Ma – 781 ka, formally the Calabrian and
90 Gelasian (International Commission on Stratigraphy, 2014) through its eventual disuse during the later
91 Middle Pleistocene (Clark 1977; Isaac 1972; Leakey 1971). Researchers have documented the
92 variability in lithic manufacture, including the production of flakes via the Levallois method and the
93 systematic production of blades during the terminal Acheulian (Adler, et al. 2014; Johnson and
94 McBrearty 2012; McBrearty 2001; Tryon, et al. 2005). Some of these technological innovations occur
95 prior to the appearance of the Middle Stone Age technocomplex, and could indicate a period of
96 technological experimentation and innovation, focused around perennial freshwater springs and

97 biomass rich habitats (Johnson, et al. 2009; Johnson and McBrearty 2012), similar to foraging behavior
98 associated with some later Middle Stone Age hominins (Curnoe, et al. 2006; McCarthy, et al. 2010).
99 The Kapthurin Formation preserves evidence of late Acheulian behavior and the only documented
100 fossil *Pan* remains (McBrearty and Jablonski 2005). Detailed environmental reconstructions will help
101 elucidate this period of innovation that precedes the Middle Stone Age, the habitat range of hominins,
102 and the habitat range of chimpanzees during the Middle Pleistocene.

103

104 **2. Background**

105 *2.1 Kapthurin Formation*

106 Lying to the west of Lake Baringo, the Kapthurin Formation is part of the Middle Pleistocene
107 sedimentary sequence of the Tugen Hills succession in the Kenyan Rift Valley (Figures 1 and 2). The
108 formation consists of interstratified alluvial, lacustrine, and volcanic sediments, is exposed over an area
109 of 150 km², and has an observed thickness of approximately 125m (Martyn 1969; Tallon 1978; Tryon
110 and McBrearty 2002). All ages quoted are direct dates generated with the ⁴⁰Ar/³⁹Ar method (Deino and
111 McBrearty 2002) (Figure 1). Martyn (1969) subdivided the formation into 5 members; K1, K3, and K5
112 are informally named the Lower, Middle, and Upper Silts and Gravels and are largely composed of
113 terrigenous sediments from the Tugen Hills; K2, the Pumice Tuff Member, dated to 543 ± 3 ka, and K4,
114 the Bedded Tuff Member, with successive eruptive facies dated to 284 ± 12 ka and 235 ± 2 ka separate
115 the Silts and Gravel Members; K3 is also subdivided by the Grey Tuff, dated to 509 ± 9 ka, and by
116 fluviolacustrine facies to the west (K3) and lacustrine facies to the east (K3'); K3' sediments consist of
117 red, black, and green claystones and siltstones (Tallon 1978), and the sedimentary and geochemical
118 structures of the clays and silts indicate deposition in a shallow lake that alternated between freshwater
119 and saline-alkaline (Johnson and McBrearty 2012; McBrearty, et al. 1996; Renaut, et al. 1999).

120 The Kapthurin Formation lies unconformably above the Chemeron Formation, and is underlain

121 by the Ndaou Trachymugearite, which is dated to 1.57 Ma (Deino, et al. 2002; Hill, et al. 1986) using
 122 the K/Ar technique, and provides a maximum age for the Kapthurin Formation. All of the Kapthurin
 123 Formation's sediments are normally magnetized (Dagley, et al. 1978), and thus postdate the Matuyama-
 124 Brunhes Boundary, which is currently estimated at 775 ka using the $^{40}\text{Ar}/^{39}\text{Ar}$ method (Coe, et al.
 125 2004), providing a maximum age for the Kapthurin Formation.
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Figure 1:
 Generalized stratigraphic section of the Kapthurin Formation. Section is after Deino and McBrearty (2002). All dates are derived from the $^{40}\text{Ar}/^{39}\text{Ar}$ method, except the Ndaou Trachymugearite, reported by Hill *et al.* (1986) using the K-Ar method. Shaded areas represent volcanic layers, light areas represent fluviolacustrine sediments.

Figure 2: Satellite image (IKONOS 1.8 m resolution) with relevant archaeological and paleontological sites containing fossil bovid teeth and pedogenic carbonates sampled for stable isotope values ($\delta^{13}\text{C}$ and $\delta^{18}\text{O}$). Inset map is not to scale.

The focus of this research is the time interval represented by fossils recovered from deposits that lie above K2, the Pumice Tuff Member (543 ± 3 ka), and below the Grey Tuff (509 ± 9 ka). McBrearty (1999) noted the excellent faunal preservation below the Grey Tuff, and reconstructed the ancient environments from taxonomic identification of fauna, suggesting dense vegetation and marsh depositional conditions, possibly near a braided river system, with grasslands farther from the rivers. Tallon (1978) also interpreted paleochannels found above and below the Grey Tuff as a product of a braided river system. Johnson, et al. (2009) documented a fossiliferous tufa in the southern exposure

141 of the lacustrine facies (K3'), stratigraphically below the Grey Tuff. The presence of this tufa
142 demonstrates a rich wetland environment in an otherwise semi-arid environment, confirming earlier
143 predictions. McBrearty and Jablonski (2005) document the presence of *Pan* fossils at Loc. 99, which
144 together with other fauna at the site (hippopotamus, crocodile, catfish, colobine monkey, and bushpig),
145 indicates a locally wooded habitat near the shore of an alternating fresh and alkaline lake, with marshy
146 patches. Hominin fossils identified as *Homo sp.* and *Homo erectus* (Leakey, et al. 1969; Wood and Van
147 Noten 1986) have also been found in this interval, in the K3 fluvial facies, close in time and space
148 (<1km) to the fossil *Pan* locality, indicating that chimpanzees and hominins occupied habitats in near
149 proximity to one another *ca* 500 ka (McBrearty and Jablonski 2005).

150 The archaeology of the deposits from K3 and K3' in the Kapthurin Formation has been
151 described by several researchers (Cornelissen 1992; Johnson and McBrearty 2010; Johnson and
152 McBrearty 2012; McBrearty 1999; McBrearty, et al. 1996; Tryon 2003). This previous work has
153 documented that Middle Pleistocene toolkits from the Kapthurin Formation were highly diverse, and
154 include handaxe assemblages (LHA, GnJh-17), assemblages lacking *fossiles directeurs* that are
155 dominated by flakes and cores (GnJh-42, 50), the early presence of prepared core technology (LHA,
156 GnJh-17), and systematic blade production (GnJh-42, 50). Tryon, et al. (2005) suggested that regional
157 variability in Acheulian methods of flake production, including the Levallois method, present in the
158 Kapthurin Formation, demonstrate variability in artifact manufacturing techniques before the Middle
159 Stone Age transition, not regionalization of Acheulean lithic manufacture (McBrearty 2001; McBrearty,
160 et al. 1996). These technological innovations could be related to methods of adaptive niche
161 construction and exploitation of patchy, resource rich environments (Johnson and McBrearty 2012),
162 although higher resolution environmental reconstructions are necessary to test these assumptions, and
163 are currently in progress (Leslie, et al. 2014).

164

165 2.2 Stable Carbon Isotopes

166 Throughout this paper, results are reported and discussed in the standard per mil (‰) notation:

167 Eq. 1: $\delta^X = (R_{\text{sample}}/R_{\text{standard}} - 1) \times 1000$; $X = {}^{13}\text{C}, {}^{18}\text{O}$; $R = {}^{13}\text{C}/{}^{12}\text{C}, {}^{18}\text{O}/{}^{16}\text{O}$ The isotope standard is

168 Vienna Pee Dee Belemnite (VPDB). The stable carbon isotopic composition of herbivorous tooth

169 enamel ($\delta^{13}\text{C}_{\text{enamel}}$) directly reflects the isotopic composition of plants consumed by the animal during

170 periods of enamel formation and mineralization (Ambrose and DeNiro 1986; Cerling and Harris 1999;

171 DeNiro and Epstein 1978). There are three main photosynthetic pathways available to plants

172 prevailing in East Africa, C_3 , C_4 , and CAM (Crassulacean Acid Metabolism), of which only the first

173 two serve as primary dietary sources for mammalian herbivores. C_3 plants use a 3-carbon compound

174 and C_4 plants utilize a 4-carbon compound intermediate phase to fix CO_2 from the atmosphere during

175 photosynthesis (Ehleringer, et al. 1997; Farquhar, et al. 1989; Osmond, et al. 1982). C_3 plants, which

176 represent the most common photosynthetic adaptation, are generally trees and closed habitat

177 vegetation, as well as high latitude grasses (Osmond, et al. 1982). The C_4 pathway is most commonly

178 represented by mid-latitude grasses, wetland sages, and arid landscape chenopods (Ehleringer, et al.

179 1997). $\delta^{13}\text{C}$ values of C_3 plants in closed canopy environments in sub-Saharan Africa are highly

180 negative relative to the standard (VPDB), averaging -31.4‰, while open canopy C_3 plants average -

181 27.8‰, and savanna and bushland C_3 plants average -27.0‰; Conversely, the $\delta^{13}\text{C}$ values of sub-

182 Saharan African C_4 plants are elevated relative to VPDB, mesic C_4 plants average -11.8‰ and xeric C_4

183 plants average -13.1‰ (Cerling, et al. 2003).. Studies of extant ruminant ungulate mammals from

184 differing habitats display an average enrichment factor ($\epsilon_{\text{enamel-diet}}$) associated with carbon incorporation

185 into tooth enamel during formation of $14.1 \pm 0.5\%$ (Cerling and Harris 1999). More recent work by

186 Passey, et al. (2005b) has shown that the enrichment factor varied over 4‰ when comparing the $\delta^{13}\text{C}$

187 values of both diet and enamel of voles, rabbits, pigs, and cattle. However, given the conclusions by

188 Passey, et al. (2005b) that these differences in enrichment are mainly a result of diet, and the

189 assumption that the fossil bovids under study are ruminant herbivores, we apply the enrichment factor
190 from Cerling and Harris (1999). With this known enrichment factor, $\delta^{13}\text{C}_{\text{enamel}}$ values can be used to
191 reconstruct diets, following Cerling, et al. (2015b), as C_4 hypergrazers ($\delta^{13}\text{C}_{\text{enamel}} \geq -2\text{‰}$), C_4 grazers
192 ($\delta^{13}\text{C}_{\text{enamel}} \geq -1\text{‰}$), C_3 closed canopy browsers ($\delta^{13}\text{C}_{\text{enamel}} \leq -14\text{‰}$), C_3 hyper-browsers ($\delta^{13}\text{C}_{\text{enamel}} \leq -$
193 12‰), C_3 browsers ($\delta^{13}\text{C}_{\text{enamel}} \leq -8\text{‰}$), and C_3/C_4 mixed feeders $-(1\text{‰} \leq \delta^{13}\text{C}_{\text{enamel}} \leq -8\text{‰})$. Following
194 Cerling, et al. (2003), the diets of bovids must be considered as part of a grazing to browsing spectrum,
195 where hypergrazers consume over 95% C_4 grasses, grazers consume between 70-95% C_4 grasses,
196 hyperbrowsers or frugivores consume over 95% C_3 browse, browsers consume between 70-95% C_3
197 browse, and mixed feeders consume more than 30% C_4 grasses and more than 30% C_3 browse.

198 Pedogenic carbonates are calcite nodules, clast coverings, and cemented horizons that are
199 formed in soils found in sub-humid to arid regions of the world. Birkeland (1999) suggests that the
200 dissolution of Calcium bearing minerals from dust and soil, release Calcium ions, which are
201 transported downward with soil water and are eventually re-precipitated as pedogenic carbonates at
202 depth in soils ($>40\text{cm}$). Pedogenic carbonates formation can be inhibited by a few factors, including
203 seasonal bias in formation, mainly forming during warm, dry seasons, not during the mean growing
204 season (Breecker, et al. 2009). The formation of pedogenic carbonates is also inhibited by densely
205 forested environments, because high precipitation and acid leaching repress carbonate precipitation,
206 indicating a possible C_4 bias when using pedogenic carbonates for environmental reconstructions
207 (Tucker and Wright 1990). The formation time of pedogenic carbonates, as individual nodules or
208 horizons, varies between 10^2 and 10^5 years (Jenny 1980). Equatorial pedogenic carbonates at depths
209 greater than 30 cm below surface vegetation in soils with high respiration rates incorporate soil CO_2
210 associated with decaying organic matter derived directly from surface vegetation during soil
211 development (Cerling 1984; Cerling and Quade 1993). Assuming minimal fluctuations in atmospheric
212 CO_2 during the Pleistocene (Cerling 1991), the main source of variation in carbon isotope values of

213 pedogenic carbonates ($\delta^{13}\text{C}_{\text{pc}}$) from the Kapthurin Formation should be derived from local decay
 214 products of surface vegetation (Levin, et al. 2011; Wynn 2004). Positive fractionation of carbon
 215 isotopes between soil CO_2 and pedogenic carbonates is determined by both kinetic (Cerling, et al.
 216 1991) and temperature dependent equilibrium (Romanek, et al. 1992) processes, resulting in a
 217 combined ^{13}C enrichment between plants and pedogenic carbonates of 12-14‰ under high respiration
 218 and 14-17‰ under low respiration (Levin, et al. 2011). More recently, Cerling, et al. (2011) developed
 219 a regression equation to calculate the fraction of woody canopy cover from $\delta^{13}\text{C}_{\text{pc}}$ values, using the
 220 following isotope values to discriminate differences in vegetation cover: Forests, a landscape with
 221 continuous stands of trees at least 10 m tall ($\delta^{13}\text{C}_{\text{pc}} < -11.5\text{‰}$); Woodland/Bushland/Shrubland, where
 222 woodlands are an open stand of trees at least 8 m tall and a field layer dominated by grass and woody
 223 cover over 40%, bushlands are an open stand of bushes between 3 and 8 m tall and woody cover over
 224 40%, and shrublands are an open or closed stand of bushes up to 2 m tall ($\delta^{13}\text{C}_{\text{pc}}$ between -11.5 and -
 225 6.5‰); Wooded Grasslands, land covered with grassland and between 10 and 40% tree and shrub
 226 coverage ($\delta^{13}\text{C}_{\text{pc}}$ between -6.5 and -2.3‰); Grasslands, land covered with grasses and less than 10% is
 227 covered with trees and shrubs ($\delta^{13}\text{C}_{\text{pc}} > -2.3\text{‰}$).
 228

229 2.3 Stable Oxygen Isotopes

230 The $\delta^{18}\text{O}$ values measured in medium and large-sized herbivore tooth enamel carbonate
 231 ($\delta^{18}\text{O}_{\text{enamel}}$) is derived from drinking water and diet, mediated by physiology (Bryant and Froelich
 232 1995; Kohn 1996). Tooth enamel can reflect meteoric water composition and thus the degree of
 233 environmental aridity, depending upon the species studied (Levin, et al. 2006). In obligate drinkers the
 234 $\delta^{18}\text{O}_{\text{enamel}}$ values directly track the $\delta^{18}\text{O}$ value of drinking water ($\delta^{18}\text{O}_{\text{drinking water}}$), which is associated
 235 with local temperature, precipitation, latitude, altitude, aridity, and evaporative processes, and are a
 236 good proxy for changes in drinking water composition, local rainfall, or moisture source (Dansgaard

1964). Other animals obtain most of their ingested water from the food they consume and are termed non-obligate drinkers. Non-obligate drinkers' water sources range from roots and stems to fruit and leaves. The $\delta^{18}\text{O}_{\text{enamel}}$ values of leaf eating non-obligate drinkers are often elevated when compared with meteoric water due to preferential transpiration of isotopically light water (H_2^{16}O) from the leaf surface, while the water content of roots, stems, and fruit usually tracks the isotopic composition of the drinking water source (Ehleringer and Dawson 1992; Sternberg 1989). Leaf water $\delta^{18}\text{O}$ enrichment (relative to stem water $\delta^{18}\text{O}$ values) is most expressed under hot and arid conditions, thus further elevating the $\delta^{18}\text{O}_{\text{enamel}}$ measured in arid-adapted animals that obtain their water from leaves (Ayliffe and Chivas 1990; Dongmann, et al. 1974). Levin, et al. (2006) suggest tracking fluctuations in meteoric water availability using an aridity index by separating mammals as Evaporative Insensitive (EI), obligate drinkers, and Evaporative Sensitive (ES), or non-obligate drinkers whose $\delta^{18}\text{O}$ values track the isotopic enrichment associated with leaf water. As environmental aridity increases, the disparity between EI and ES $\delta^{18}\text{O}_{\text{enamel}}$ values should increase, providing an independent measure of aridity from meteoric water isotopic composition. Thus, $\delta^{18}\text{O}_{\text{enamel}}$ values from both obligate and non-obligate drinkers can be useful in reconstructing the isotopic composition of local meteoric water, paleoenvironmental conditions, and paleoaridity.

The $\delta^{18}\text{O}$ values of pedogenic carbonates ($\delta^{18}\text{O}_{\text{pc}}$) are controlled by the oxygen isotopic composition of soil water and temperature. The $\delta^{18}\text{O}$ value of soil water is determined by the $\delta^{18}\text{O}$ value of rainfall water and the degree of soil water evaporation. Soil water is recharged through infiltration and surface run-off, and accumulations of rainfall over the course of a season (Breecker, et al. 2009). The $\delta^{18}\text{O}$ values of rainfall vary with climate, elevation, and distance from humidity source (Dansgaard 1964). In East Africa, the main determinants of rainfall $\delta^{18}\text{O}$ values are rainfall amounts, elevation, and distance from humidity sources (Levin, et al. 2009; Rozanski, et al. 1996). Because surface water is subject to increased evaporation, due to increased time available for evaporation,

261 $\delta^{18}\text{O}_{\text{pc}}$ values are often elevated in near surface samples, where there is increased potential for soil
262 water evaporation (Breecker, et al. 2009; Quade, et al. 1989). Finally, $\delta^{18}\text{O}_{\text{pc}}$ values are a function of
263 soil temperature, because the fractionation between soil water and pedogenic carbonates is temperature
264 dependent ($-0.2\text{‰}/^{\circ}\text{C}$) (Kim and O'Neil 1997). Passey, et al. (2010) demonstrated that soil
265 temperatures during the last 4 million years of the Turkana Basin exceed those of Mean Annual
266 Temperatures (MAT) today. More recently, Cerling, et al. (2015a) have shown that soil temperatures,
267 taken over 25 cm below surface in open environments in East Africa, are elevated by as much as 7°C
268 when compared with MAT, while soil temperature in forest soils are similar to MAT estimates.

269

270 *2.4 Paleoclimate Predictions*

271 Deep sea formation of organic-rich layers (sapropel) in the eastern Mediterranean Sea indicate
272 periods of increased fresh water discharge from the Nile River and have been explained by intensified
273 monsoonal activity in East Africa during periods of Northern Hemisphere insolation maxima
274 (Rossignol-Strick and Paterne 1999; Rossignol-Strick, et al. 1982). The formation of sapropels in the
275 Mediterranean has also been previously linked with hominin activity in East Africa (McDougall, et al.
276 2008). Others (Deino, et al. 2006; Kingston, et al. 2007) have demonstrated orbital forcing of cooling
277 and heating events are well grounded in East African field data. Sapropel A has been identified by
278 Rossignol-Strick, et al. (1998) and correlated to an insolation maximum at $\sim 525\text{--}528$ ka using
279 astronomic calibration (Hilgen 1991). Sapropel A formation exhibits a relatively low monsoon index
280 (30) and exceptionally negative $\delta^{18}\text{O}$ values, measured in foraminifera (*Globigerinoides ruber*) from
281 the sapropel, which indicate marked decrease in salinity and/or an increase in the water input from the
282 Nile. The intensity of African monsoonal activity and sea surface temperature in the tropical Atlantic
283 Ocean has been linked with turnover events of equatorial African C_3 and C_4 vegetation that are largely
284 controlled by aridity (Schefuß, et al. 2003). Schmieder, et al. (2000) document the presence of thick

285 interstratified layers of the giant diatom *Ethmodiscus rex* in deep ocean cores from the South Atlantic
286 Ocean at the onset of Marine Isotope Stage (MIS) 13, approximately 533 ka (Lisiecki and Raymo
287 2005). These diatoms are commonly found today in warm equatorial oceans (Mikkelsen 1977) and
288 Schmieder, et al. (2000) suggest that the presence of these laminated layers indicate a period of extreme
289 paleoceanographic variability in temperature and circulation in the South Atlantic Ocean throughout
290 MIS 13 (~533–478 ka) (Lisiecki and Raymo 2005). Kukla and Cílek (1996) report a peak in the
291 magnetic susceptibility stack from the Chinese Loess plateau that suggests soil formation in a warm
292 and humid climate during MIS 13. Clark, et al. (2006) interpret this peak in the magnetic susceptibility
293 stack as evidence for increased monsoonal activity over Asia and Africa. Hillenbrand, et al. (2009) also
294 note the probable collapse of the Western Antarctic Ice Sheet during MIS 15–13 (~621–478 ka),
295 suggesting it is most likely caused by a prolonged interglacial period that is marked by increased
296 biological productivity and lithogenic sediment supply. These proxies provide good evidence for a
297 period of intense precipitation in equatorial Africa during MIS 13, which is largely controlled by an
298 intensified African monsoonal system, and indicate an increase in C₃ vegetation in East Africa (Figure
299 3). We will test these climatic predictions with African paleoecological data relevant to MIS 13.

Figure 3: Composite paleoclimate data indicating an intensified African monsoonal system during MIS 13 (highlighted). CO₂ atmospheric composition measurements are from Allan Hills BIA (Site 27), Antarctica, from data in Higgins, et al. (2015a). LR04 benthic δ¹⁸O stack from 57 globally distributed records reported in Lisiecki and Raymo (2005). Core MD 900961 from Rossignol-Strick, et al. (1998) displays periods of sapropel formation in the Mediterranean Sea, generally correlated with intensified monsoonal systems. Monsoon index is controlled by insolation variations, and display variations in the intensity of the African monsoonal system over time, after Rossignol-Strick (1983). During MIS 13, CO₂ atmospheric pressure is high, indicating warmer global temperatures, and precipitation in equatorial Africa increases dramatically, indicated in the monsoon index.

3. Methods

Fossil bovid teeth (n=20) were collected from paleontological sites stratigraphically below the Grey Tuff (~509 ka). Sample sizes are small due to the nature of fossil preservation in the Kapthurin Formation, paleontological specimens are relatively rare, and generally only preserved during the time interval under study. Other paleontological specimens were discarded if they were not identified to tribe, or had incisions or breaks in surface enamel, which could be indicative of diagenetic alteration, or

308 were currently under study by other researchers. The specimens are housed in the Department of
309 Paleontology, National Museums of Kenya, Nairobi, not all specimens are accessioned; field
310 identification numbers are used in place of accession numbers. Sites from the Kapthurin Formation
311 sampled include GnJh-23, GnJh-38, GnJh-41, GnJh-47, GnJh-52, GnJh-53, GnJh-56, GnJh-58, and
312 Locality 4 (Figure 2). Sampled teeth were identified to bovid tribe according to Gentry (1978) criteria.
313 Bovid tribes sampled include: Neotragini, Reduncini, Tragelaphini, Alcelaphini, Cephalophini, and
314 Bovini. Bovids were chosen for analysis because extant bovids inhabit diverse environments and
315 should provide evidence for variation in past habitats, if it exists (Bibi, et al. 2009).

316 Coupled samples of $\delta^{13}\text{C}_{\text{enamel}}$ and $\delta^{18}\text{O}_{\text{enamel}}$ derived from serial samples of complete teeth,
317 display dietary choices and changes in aridity during enamel formation, and thus offer a longer span of
318 time to reconstruct habitats (Fricke and O'Neil 1996; Koch, et al. 1989; Passey, et al. 2005a). Tooth
319 formation can occur over periods of months or years depending upon the taxon (Hoppe, et al. 2004;
320 Stevens, et al. 2011; Zazzo, et al. 2005; Zazzo, et al. 2002). Drilling samples perpendicular to the
321 growth axis, samples derived near the occlusal surface of the tooth (apex) represent early periods of
322 tooth formation, while samples near the tooth cervix represent late periods of tooth formation (Passey,
323 et al. 2005a). Balasse (2003) reports that even though some isotopic mixing may occur given a simple
324 sampling strategy perpendicular to the growth of the tooth, starting at the cervix and ending at the apex,
325 the chronological order of enamel formation is preserved. Thus, sampling intra-tooth isotopic variation
326 has become an important tool for reconstructing seasonal variability in climate, ecology, and behavior
327 (Balasse, et al. 2003; Fricke, et al. 1998; Kohn, et al. 1998; Zazzo, et al. 2002).

328 Kenyan cultural heritage laws do not permit the export of fossil teeth, so all sampling was
329 performed at the National Museums of Kenya. A dental drill (Vogue 6000) with a 1 mm diamond drill
330 bit was used to sample the teeth serially, following the protocols of Balasse (2003). Fossil teeth were
331 sampled incrementally, with individual samples drilled perpendicular to the growth axis of the tooth.

Each tooth was drilled at an interval of approximately 1.5 mm, with initial samples taken at the cervix (late formation) and final samples taken at the apex (early formation) to reflect the animal's diet and water availability during enamel formation. Fossil teeth were chosen on the basis of site, taxon, tooth type, and preservation quality. We primarily sampled molar teeth to avoid early juvenile dietary signatures (Bedaso, et al. 2013). EI and ES fauna were identified following the criteria of Levin, et al. (2006) as well as modern physiological and behavioral adaptations (Kingdon 1988). Diagenesis of fossil enamel is a possibility, but Wang and Cerling (1994) have demonstrated that tooth enamel is consistently resistant to diagenetic alteration, and generally preserves the original isotopic signature. The $\delta^{13}\text{C}$ signatures of the fossil bovids sampled in this study are not outside the known ranges of extant faunal isotopic signals (Cerling, et al. 2003). Browser representatives, such as individual tragelaphini teeth, average -10.4‰ (Field ID 80), -7.9‰ (Field ID 5, not an average, only one sample is represented), -11.2‰ (Field ID 232), and -8.8‰ (Field ID 25), similar to modern ranges of -8.5‰ and -10.0‰ (Cerling, et al. 2015b). Hyper-grazer representatives, such as individual alcelaphini teeth, average 1.2‰ (Field ID 99), 1.8‰ (Field ID 245), and 1.0‰ (Field ID 304) and are very close to the modern ranges of > 2‰ (Cerling, et al. 2015b). These results suggest that these fossils have not undergone significant diagenesis, and are within modern ranges once the 1.5‰ shift in modern carbon values is considered (Leuenberger, et al. 1992) (Table 1).

Each serial sample was composed of approximately 10 mg of powdered tooth enamel and was prepared following Balasse (2003). Samples were treated with 3% sodium hypochlorite (0.1 M NaOCl solution per mg of sample) for 24 hours to remove organic matter, and were neutralized with deionized water (dH₂O). Dried enamel samples were then reacted with acetic acid (0.1 M C₂H₄O₂) for four hours to remove any diagenetic carbonates. Samples were submitted to the *Centre National de la Recherche Scientifique, Museum National d'Histoire Naturelle* (CNRS/MNHN) for stable isotope ratio analysis. The isotope analyses were performed using a Thermo-Finnigan Kiel IV carbonate analyzer coupled to

356 a DeltaV mass spectrometer.

357 Analytical precision was obtained using internal standards (MarbreLM) calibrated against NBS-19.

358 These internal standards were run intermittently with the tooth enamel samples, with standard

359 deviations for $\delta^{13}\text{C}$ of 0.016‰ and for $\delta^{18}\text{O}$ of 0.05‰. To estimate $\delta^{18}\text{O}_{\text{drinking water}}$ values, we converted

360 the $\delta^{18}\text{O}_{\text{enamel}}$ from the VPDB to the VSMOW scale using the following approximation from Sharp

361 (2007):

362 Eq. 2: $\delta^{18}\text{O}_{\text{VSMOW}} = 1.03091 \times (\delta^{18}\text{O}_{\text{enamel}}) + 30.91$

363 Next, we calculate $\delta^{18}\text{O}_{\text{drinking water}}$ values of obligate drinkers (EI) using the conversion equation

364 provided by Chenery, et al. (2012), who converted human carbonate (VSMOW) into phosphate oxygen

365 ($\delta^{18}\text{O}_{\text{P}}$) values (VSMOW), combining their regression equations with the drinking water equation from

366 Daux, et al. (2008), creating the following equation:

367 Eq. 3: $\delta^{18}\text{O}_{\text{drinking water (phosphate)}} = 1.590 \times \delta^{18}\text{O}_{\text{VSMOW(carbonate)}} - 48.634$

368 We chose Equation 3, due to the low uncertainty for calculating $\delta^{18}\text{O}_{\text{drinking water}}$ values ($\pm 1\text{‰}$, 2σ), when

369 compared with larger errors (between $\pm 1\text{‰}$ and $\pm 3\text{‰}$) associated with drinking water conversion

370 (Pollard, et al. 2011). Following Bedaso, et al. (2013), we calculate $\delta^{18}\text{O}_{\text{drinking water}}$ values using the

371 most negative value of each obligate drinker's tooth sampled, to minimize evapotranspiration effects

372 that would contribute to higher $\delta^{18}\text{O}$ values. These equations were developed for human phosphate and

373 carbonate values, so we calibrate these equations by testing them on data provided in Levin, et al.

374 (2006), reconstructing $\delta^{18}\text{O}_{\text{drinking water}}$ values from modern fauna and measured water values from the

375 following sites: Arabuko-Sokoke (-3.9‰ vs. reported -2.3‰), Athi (-2.1‰ vs. reported -4.0‰), Mpala

376 (+0.25‰, vs. reported -3.3‰), and Tsavo (-2.1‰, vs. reported -4.0‰). These estimates are within the

377 error ranges reported by Pollard, et al. (2011) when reconstructing drinking water values. To

378 reconstruct aridity, we calculated the enrichment between ES and EI taxa (after Levin, et al. 2006):

379 Eq. 4: $\mathcal{E}_{\text{ES-EI}} = ((R_{\text{ES}}/R_{\text{EI}}) - 1) \times 1000$

380 The Anderson-Darling test for normality was used to determine if $\delta^{13}\text{C}_{\text{enamel}}$ and $\delta^{18}\text{O}_{\text{enamel}}$
 381 values were normally distributed. *P*-values demonstrate that neither distribution is normal, and thus
 382 non parametric tests for significance are appropriate. Pairwise Mann-Whitney U tests (two-tailed) were
 383 applied to sample localities, as long as three teeth were sampled from each location. Mann-Whitney U
 384 tests were also applied to bovid tribes and to ES and EI identified specimens. Isoscapes of predicted
 385 carbon and oxygen isotope ratios were produced with ArcView 10.2. These data were interpolated
 386 between data points using inverse distance weighting, an interpolation technique appropriate for
 387 ecological data (Johnston 1998). Statistics were calculated using Microsoft Excel and *R* software.

388 Pedogenic carbonates nodules (n=24) were collected during 2011 and 2012 field seasons, from
 389 20 different paleontological, archaeological, and geological settings. All samples were collected from
 390 well-developed paleosols, stratigraphically sampled below the Grey Tuff or individual paleontological
 391 and archaeological sites within the time interval under study (543-509 ka). Given these constraints,
 392 some time-averaging must be expected. Pedogenic carbonates were examined microscopically for
 393 cracks, infilling, and other diagenetic reworking by detrital carbonate; any with such alterations were
 394 discarded from analyses. Nodules were embedded in an epoxy-resin (EpoFix) and then cross-sectioned
 395 using a precision saw (Buehler Isomet 1000). Cross-sections of nodules were sampled using a high
 396 precision micromill (New Wave) with a 0.5 mm diamond drill bit. Nodules were serially sampled at 1
 397 mm intervals to determine any internal nodular variability during formation (n=254), the serial
 398 samples have been averaged and are provided with individual standard deviations (Table 2). Powdered
 399 samples were submitted to the *Centre National de la Recherche Scientifique, Museum National*
 400 *d'Histoire Naturelle* (CNRS/MNHN) for stable isotope ratio analysis. The isotope analyses were
 401 performed using a Thermo-Finnigan Kiel IV carbonate analyzer coupled to a DeltaV mass
 402 spectrometer.

403 Analytical precision was obtained using internal standards (MarbreLM) calibrated against NBS-19.

404 These internal standards were run intermittently with the carbonate samples, with standard deviations
405 for $\delta^{13}\text{C}$ of 0.03‰ and for $\delta^{18}\text{O}$ of 0.05‰. Fraction woody cover (f_{wc}) was estimated following
406 Cerling, et al. (2011), by first converting the values of $\delta^{13}\text{C}_{\text{pc}}$ to the isotopic equivalent of organic
407 carbon ($\delta^{13}\text{C}_{\text{om}}$), subtracting 14‰. We also applied the regression equation below (Eq 5) to estimate
408 fraction woody cover during this time interval.

409 Eq. 5: $f_{\text{wc}} = \{\sin[-1.06688 - 0.08538(\delta^{13}\text{C}_{\text{om}})]\}^2$

410

411 **4. Results and Discussion**

412 *4.1 Carbon isotopic composition of tooth enamel by tribe*

413 The results of stable isotope inquiry are displayed in Table 1. The range of $\delta^{13}\text{C}_{\text{enamel}}$ values
414 measured for fossils from the 543 – 509 ka time interval of the Kapthurin Formation (-11.5 to +2.3 ‰)
415 spans extant bovid adaptations to C_3 and C_4 diets (Cerling, et al. 2015b). The $\delta^{13}\text{C}_{\text{enamel}}$ values indicate
416 consistent consumption of both C_3 and C_4 vegetation, with a few specimens falling within the expected
417 range of mixed feeders. Generally the dietary preferences of individuals sampled did not show
418 variation through serial sampling, with the notable exception of Field ID 25, a lower third molar of a
419 traglephine whose diet shifted seasonally to a mixed C_3/C_4 diet from a strict C_3 browsing diet during
420 tooth formation (Figure 4). When teeth data are compared by tribe using pairwise Mann-Whitney U
421 tests, alcelaphini, bovini, reduncini, and tragelaphini are all significantly different from one another (P
422 < 0.05) in $\delta^{13}\text{C}_{\text{enamel}}$ values. Neotragini and cephalophini were not included in statistical comparisons
423 due to sample sizes of $n=1$ for each tribe.

424

Figure 4: Average stable isotope values $\pm 1\sigma$ of serial bovid tooth enamel samples, displayed against x-axis ($\delta^{13}\text{C}$) and y-axis ($\delta^{18}\text{O}$). Samples are shown with respect to assumed diet type. Shaded and light areas reference C_3 , Mixed C_3/C_4 , and C_4 diets.

Alcelaphini- The diet of alcelaphines is dominated by C_4 vegetation, and does not appear to show any dietary shifts during tooth formation, although sample sizes are too small to permit statistical analyses (Table 1). The $\delta^{13}\text{C}_{\text{enamel}}$ values of sampled molars average 1.2‰ (Field ID 99), 1.8‰ (Field ID 245), and 1.0‰ (Field ID 304), indicating that the alcelaphines were grazers, feeding on C_4 grasses for the majority of their diet in open savanna environments (Table 1). These fossil data represent the second highest $\delta^{13}\text{C}_{\text{enamel}}$ values of the bovid taxa sampled, averaging between 1.0 and 1.8‰, and are very close to the modern ranges associated with alcelaphines ($>2\%$) indicating a similarly selective diet of C_4 grasses (Cerling, et al. 2015b). As in other studies, these data suggest that fossil alcelaphines had a slightly more plastic diet that was at least 1‰ lower than modern analogs, indicating a slight percentage of C_3 dietary sources and must be considered grazers, not hyper-grazers (Cerling, et al. 2015b; Kingston and Harrison 2007). The presence of alcelaphines during this time interval indicates the existence of the grassland ecosystems that they commonly inhabit today.

Bovini- The diet of two of the bovines appear to be dominated by C_4 vegetation with average $\delta^{13}\text{C}_{\text{enamel}}$ values of 0.4‰ (Field ID 266) and -0.2‰ (Field ID 254), while Field ID 228 has an average value of -2.9‰, indicating a C_3 dietary component (Table 1). Both Field ID 228 and 254 have sequential

442 $\delta^{13}\text{C}_{\text{enamel}}$ values that trend more positive during tooth formation (-3.7 to -0.8‰, and -1.5 to 1.0‰
443 respectively), possibly indicating a shift in diet during tooth formation. The other tooth (Field ID 266)
444 has relatively stable $\delta^{13}\text{C}$ values throughout tooth formation (between 0.7 and -0.5‰). Field IDs 254
445 and 266 average $\delta^{13}\text{C}_{\text{enamel}}$ values are indicative of a grazing adaptation, similar to extant bovines
446 whose $\delta^{13}\text{C}_{\text{enamel}}$ values average $0.9 \pm 3.3\text{‰}$ (Cerling, et al. 2015b). Field ID 228 $\delta^{13}\text{C}_{\text{enamel}}$ values
447 average -2.9‰, and while indicating a C_3 - C_4 mixed feeder dietary adaptation, are still within the range
448 of extant bovine carbon enamel values (Cerling, et al. 2015b). Taken together, these enamel values
449 indicate that during this time interval bovines in the Baringo Basin incorporated a large, yet not
450 exclusive, portion of C_4 vegetation into their diet. While extant bovines diet can range from pure
451 grazing to pure browsing depending on the habitat, the contribution of C_3 vegetation to the diet of Field
452 ID 228 could presumably be from C_3 browse, grass, shrub, or aquatic sources (Cerling, et al. 2015b).
453 Here, we interpret the presence of bovines to indicate the existence of open C_4 grasslands and C_3
454 browse, shrub, or aquatic plants given the low $\delta^{18}\text{O}_{\text{enamel}}$ average values associated with Field IDs 266
455 (-1.9‰), 228 (-1.7‰), and 254 (-2.2‰) (Table 1) (See section 4.2).

456

457 *Cephalophini*- The $\delta^{13}\text{C}_{\text{enamel}}$ values of the single *Cephalophini* specimen average -8.0‰, suggesting a
458 browser dietary adaptation, although one that is quite different from modern analogs whose $\delta^{13}\text{C}_{\text{enamel}}$
459 values range between -10.6 and -14.3‰ (Cerling, et al. 2015b). Due to the small size of the molar,
460 only two serial samples were manually drilled for $\delta^{13}\text{C}_{\text{enamel}}$ analysis resulting in values of -7.7 and -
461 8.2‰. These $\delta^{13}\text{C}$ values are between 2 and 6‰ more positive than extant values for *cephalophini* and
462 would suggest an open woodland habitat, with contributions from C_4 grasses, or a mixture of C_3
463 browse, shrub, herb, or grass plants whose $\delta^{13}\text{C}$ values range between -20 and -37‰ (Kohn 2010). This
464 data is important despite the small sample size, because specimens of *Cephalophini* generally do not
465 preserve well in the fossil record and due to the divergence of fossil $\delta^{13}\text{C}_{\text{enamel}}$ values from extant

466 values (Bibi 2013; Cerling, et al. 2015b; Gentry, et al. 2010; Kingdon 1988).

467

Field ID	Serial #	Tribe	Genus	Tooth	Site	$\delta^{13}\text{C}$	$\delta^{18}\text{O}$	$\bar{x} \delta^{13}\text{C}$	$\bar{x} \delta^{18}\text{O}$
58	1	Reduncini +	<i>Redunca</i>	LM3	GnJh-23	1.2	-1.6	1.5	-0.5
	2					1.3	0.0		
	3					1.4	-0.3		
	4					2.2	0.0		
70	1	Reduncini +	<i>Redunca</i>	LM3	GnJh-23	1.9	-0.1	1.9	-0.1
	2					1.5	0.0		
	3					2.1	-0.3		
80	1	Tragelaphini ++	<i>Tragelephus</i>	LM	GnJh-23	-10.7	-2.8	-10.4	-2.0
	2					-10.2	-1.2		
99	1	Alcelaphini +	<i>Damaliscus</i>	UM3	GnJh-23	0.8	1.7	1.2	-0.4
	2					0.9	0.7		
	3					1.6	-0.6		
	4					1.3	-2.0		
	5					1.2	-1.9		
48	1	Neotragini ++	<i>Ourebia</i>	UM3	GnJh-23	-0.6	-0.9	-0.2	-0.1
	2					0.3	-0.5		
	3					-0.2	1.0		
57	1	Reduncini +	<i>Redunca</i>	UM3	GnJh-23	0.2	0.3	0.1	-0.1
	2					0.6	-0.2		
	3					-0.6	-0.5		
35	1	Reduncini +	<i>Reduncini</i>	UM	GnJh-38	0.6	1.1	1.8	2.5
	2					2.2	2.6		
	3					2.2	3.3		
	4					2.0	3.1		
5*	2	Tragelaphini ++	<i>Tragelephus</i>	UP3	GnJh-38	-7.9	-1.1	-7.9	-1.1
31	1	Reduncini +		UM	GnJh-38	2.2	0.3	2.5	-0.4
	2					2.2	-0.5		
	3					3.1	-1.1		
237	1	Cephalophini ++	<i>Sylvicapra</i>	LM3	GnJh-41	-8.2	-2.1	-8.0	-2.4
	2					-7.7	-2.8		
92	1	Reduncini +	<i>Redunca</i>	LM3	GnJh-41	1.8	-0.6	1.9	0.4
	2					2.1	0.4		
	3					1.7	1.3		
244	1	Reduncini +	<i>Kobus</i>	UM	GnJh47	0.6	-0.8	1.4	-0.6
	2					2.0	-0.1		
	3					1.6	-0.8		

468

469

Table 1: Serial and average stable isotope values of fossil bovid teeth sampled. Samples were collected serially and have been averaged from serial samples. $\bar{x}$ =averages, *=averages are not possible, n=1. + Evaporative Insensitive, ++ Evaporative Sensitive, LM=Lower Molar, UM=Upper Molar, UP=Upper Premolar.

Field ID	Serial #	Tribe	Genus	Tooth	Site	$\delta^{13}\text{C}$	$\delta^{18}\text{O}$	$\bar{x} \delta^{13}\text{C}$	$\bar{x} \delta^{18}\text{O}$
266	1	Bovini ++	<i>Syncerus</i>	LM3	GnJh-47	-0.5	-1.8	0.4	-1.9
	2					0.5	-1.6		
	3					0.5	-1.8		
	4					0.8	-1.3		
	5					0.1	-2.1		
	6					0.3	-2.3		
	7					0.6	-2.1		
	8					0.7	-1.7		
245	2	Alcelaphini +	<i>Damaliscus</i>	UM	GnJh-47	1.7	-0.1	1.8	-0.4
	3					2.0	-0.4		
	4					1.5	-0.8		
228	1	Bovini ++	<i>Syncerus</i>	UM	GnJh-52	-0.8	-0.3	-2.9	-1.7
	2					-3.1	-1.7		
	3					-3.1	-1.6		
	4					-3.5	-2.3		
	5					-3.7	-2.4		
232	1	Tragelaphini ++	<i>Scriptus</i>	UM	GnJh-53	-10.9	-2.3	-11.2	-2.7
	2					-11.5	-3.1		
254	1	Bovini ++	<i>Syncerus</i>	UM	GnJh-56	1.0	-2.5	-0.2	-2.2
	2					0.7	-2.3		
	3					-0.4	-1.9		
	4					-1.0	-2.1		
	5					-1.5	-2.3		
255	1	Reduncini +	<i>Redunca</i>	LM	GnJh-56	1.7	-0.6	2.1	-1.5
	2					2.2	-1.6		
	3					2.3	-2.3		
304	1	Alcelaphini +	<i>Damaliscus</i>	UM	GnJh-58	1.1	0.8	1.0	0.3
	2					0.8	1.4		
	3					0.8	0.7		
	4					1.4	-0.7		
	5					1.1	-0.8		
25	1	Tragelaphini ++	<i>Tragelaphus</i>	LM3	Loc. 4	-7.6	-3.0	-8.8	-1.8
	2					-7.7	-2.8		
	3					-8.2	-3.0		
	4					-8.7	-1.8		
	5					-8.1	-2.2		
	6					-8.1	-2.2		
	7					-10.5	-0.1		
	8					-11.3	0.4		

Table 1 (continued): Serial and average stable isotope values of fossil bovid teeth sampled. Samples were collected serially and have been averaged from serial samples. $\bar{x}$ = averages, * = averages are not possible, + Evaporative Insensitive, ++ Evaporative Sensitive, LM=Lower Molar, UM= Upper Molar, UP=Upper Premolar.

475 *Neotragini*- The average $\delta^{13}\text{C}_{\text{enamel}}$ values (-0.2‰) of the single Neotragini specimen are consistent
476 with a C₄ grazing adaptation (Table 1). These values mirror the extant counterpart, *Ourebia ourebi*,
477 the only true grazer in the Neotragini family (extant $\delta^{13}\text{C}_{\text{enamel}}$ values average -1.2‰) when compared
478 with other neotragines that rely on C₃ vegetation as the major component of their diet (extant $\delta^{13}\text{C}_{\text{enamel}}$
479 values average between -10.4 and -24.1‰) (Cerling, et al. 2015b; Sponheimer, et al. 2003). Here, we
480 interpret the Neotragini specimen to indicate of an open, C₄ grassland habitat.

481
482 *Reduncini*- The $\delta^{13}\text{C}_{\text{enamel}}$ values for reduncines are consistent with hyper grazing and grazing dietary
483 adaptations, averaging 1.5‰ (Field ID 58), 1.9‰ (Field ID 70), 0.1‰ (Field ID 57), 1.8‰ (Field ID
484 35), 2.5‰ (Field ID 31), 1.9‰ (Field ID 92), 1.4‰ (Field ID 244), and 2.1‰ (Field Id 255) (Table 1).
485 These results suggest that the fossil reduncines' diets were focused exclusively on C₄ vegetation and
486 thus would have no seasonal $\delta^{13}\text{C}_{\text{enamel}}$ variability. These results are consistent with extant dietary
487 choices of reduncines, whose $\delta^{13}\text{C}_{\text{enamel}}$ values average 0.7 to 2.3‰ (Cerling, et al. 2015b). We
488 interpret these fossil reduncine $\delta^{13}\text{C}_{\text{enamel}}$ values to indicate the presence of grasslands, but also mesic,
489 marshy environments dominated by C₄ grasses, perhaps in a sumpland environment fed by seasonal
490 rains or a perennially active river (See section 4.2).

491
492 *Tragelaphini*- The $\delta^{13}\text{C}_{\text{enamel}}$ values of tragelaphines are largely consistent with browser dietary
493 adaptations although there is evidence for one C₃-C₄ mixed feeder, averaging -10.4‰ (Field ID 80), -
494 7.9‰ (Field ID 5, not an average, only one value sampled), -11.2‰ (Field ID 232), and -8.8‰ (Field
495 ID 25) (Table 1). These results suggest that the fossil tragelaphines relied on C₃ vegetation for the
496 majority of their diets, and on C₄ vegetation for smaller components of their diets. There are many
497 possible explanations for the C₃ components of the tragelaphines diets, including browse, shrubs, herbs,
498 or fruits, although given the associated negative $\delta^{18}\text{O}_{\text{enamel}}$ values (averaging between -1.1 and -2.7‰)

shrubs or browse associated with perennial water sources, such as riparian woodlands, are the most likely explanations. One tooth (Field ID 25) shows marked variation in $\delta^{13}\text{C}_{\text{enamel}}$ values, and these values trend more positive during tooth formation (-7.6 to -11.3‰), indicating a possible seasonal shift away from a C_4 diet component to a greater reliance on C_3 browse material. These fossil $\delta^{13}\text{C}_{\text{enamel}}$ values are similar to those of extant tragelaphines values that average between -8.5 and -14.9‰ and usually indicate closed forested environments (Cerling, et al. 2015b; Kingdon 1988; Sponheimer and Lee-Thorp 2003; Sponheimer, et al. 2003). Here, we interpret the $\delta^{13}\text{C}_{\text{enamel}}$ values of tragelaphines to generally indicate open and closed woodland environments, possibly in riparian habitats given the associated negative $\delta^{18}\text{O}_{\text{enamel}}$ values (See section 4.2).

4.2 Oxygen isotopic composition of tooth enamel and aridity

The total range of $\delta^{18}\text{O}_{\text{enamel}}$ values sampled spans -3.1 and +3.3‰, providing the seasonal contribution of meteoric and leaf water sources, food, thermos-physiological adaptations, and possibly migration (Figure 4 and Table 1) The largest intra-tooth range in $\delta^{18}\text{O}_{\text{enamel}}$ values occurs in samples from an alcelaphine tooth (Field ID 99) with 3.8‰, and in a tragelaphine tooth (Field ID 25) with 3.5‰. The smallest intra-tooth range in $\delta^{18}\text{O}$ values come from a reduncine tooth (Field ID 70) at 0.3‰ and a bovine tooth (Field ID 254) at 0.6‰. Generally, these specimens do not show directional shifts in the $\delta^{18}\text{O}_{\text{enamel}}$ values sampled. However, a single tragelaphine tooth (Field ID 25), does show a seasonal trend towards more negative oxygen isotope values (0.4 to -3.0‰) during enamel formation that coincide with a shift away from a C_3 dominated diet to one with increased C_4 components (-11.3 to -7.6‰). , This trend is consistent with early formation of C_3 dominated browse associated with ^{18}O enriched water due to leaf water transpiration. As C_4 vegetation contribution in the diet increases during late tooth formation the oxygen isotope values begin to track meteoric water values more faithfully (Table 1) (Kohn, et al. 1996; Kohn, et al. 1998). Conversely, an alcelaphine tooth (Field ID

99), shows a directional trend toward positive $\delta^{18}\text{O}_{\text{enamel}}$ values, but without an alteration in a C_4 grazing diet (Table 1). This pattern could indicate the contribution of ephemeral water sources as a supplement to the alcelaphines water intake during the dry season. When tribes are compared using Mann-Whitney U pairwise tests, both reduncini and alcelaphini are significantly different from bovini and tragelephini, but not from each other in $\delta^{18}\text{O}_{\text{enamel}}$ values ($P = 0.737$). Likewise, bovini and tragelephini are not significantly different in their $\delta^{18}\text{O}_{\text{enamel}}$ values ($P = 0.441$). Finally, when all teeth are compared on the basis of ES or EI status, their $\delta^{18}\text{O}_{\text{enamel}}$ values are significantly different ($P < 0.01$).

The stable oxygen and carbon values of fossil fauna, in conjunction with dietary habits of their modern counterparts, were useful when classifying each bovid tribe as obligate drinkers (EI), who derive the majority of their $\delta^{18}\text{O}_{\text{enamel}}$ values from meteoric water, or non-obligate drinkers (ES), who derive large components of their $\delta^{18}\text{O}_{\text{enamel}}$ values from dietary water sources (after Levin, et al. 2006). Within the Kapthurin fauna sampled here, we classify tragelaphines, neotragines, bovines (*Syncerus* sp.), and cephalophines as ES specimens, and reduncines and alcelaphines (*Damaliscus* sp.) as EI specimens. We chose to classify *Syncerus* sp. as ES contra to Levin, et al. (2006), who suggest that the stable oxygen isotope values of bovines are similar to ES specimens but not at the $P < 0.05$ level, but had a P value 0.074 (Levin, et al. 2006). We also classify alcelaphines (*Damaliscus* sp.) as EI species based on the criteria set forth in Levin, et al. (2006), their modern counterparts are hyper-grazers (Cerling, et al. 2015b; Cerling, et al. 2003), have medium to large body sizes, are obligate drinkers, and as a tribe are sometimes found in proximity to water sources (Kingdon 1988). The classification of alcelaphines follows data from Levin, et al. (2006) which showed no relationship between alcelaphines (impala) and water deficit (P value 0.311), and is contra Bedaso, et al. (2010) and Bedaso, et al. (2013), who classify alcelaphines as ES, but do not provide a justification for this classification. Surprisingly, EI classified taxa have $\delta^{18}\text{O}_{\text{enamel}}$ values that are overall more positive on average (-0.02%) than ES

specimens (1.8‰), contra to expectations that EI specimens should have lower $\delta^{18}\text{O}_{\text{enamel}}$ values than ES specimens. We attribute this to the overall wet conditions during this time interval, which suggest that the preferential evapotranspiration of $\delta^{18}\text{O}_{\text{enamel}}$ values is not exacerbated by arid conditions (Figure 3). This is similar to data reported by Levin, et al. (2006), where in the Ituri Rainforest, EI specimens have more positive $\delta^{18}\text{O}_{\text{enamel}}$ values than ES specimens.

Figure 5 shows the calculated paleoaridity using $\delta^{18}\text{O}_{\text{enamel}}$ values, and the relationship of $\epsilon_{\text{ES-EI}}$ to the associated water deficit (a measure of mean annual precipitation and potential evapotranspiration), calibrated and modified from Levin, et al. (2006). Due to the small sample sizes from the Kapthurin Formation, we rely on the robust sample sizes from many taxa in Levin, et al. (2006) not included in our analysis (giraffidae, rhinocerotidae, proboscidea, hippopotamidae, equidae, and suidae). Again, due to the small sample sizes, we did not calculate the aridity index for any individual paleontological sites, but instead calculate it for the time interval under study, assuming that the fauna was deposited before the deposition of the Grey Tuff (509 ± 9 ka), although some time averaging must be expected. This method of correlation, following widespread, mappable temporal markers of paleo-landscapes is common, and at best provides a paleo-average of a landscape (Bedaso, et al. 2010). Our sample sizes for EI specimens is $n=11$ individual teeth, and for ES specimens is $n=9$ individual teeth. In Levin, et al. (2006), the authors note that for the aridity index to provide robust significance, 10 samples of both ES and EI taxa are necessary from individual sites. Following Levin, et al. (2006), we have calculated a power analysis (significance (α) = 0.05, power level ($1-\beta$) = 0.2, two-tailed test), indicating that 10 samples are necessary from each population to determine a difference of 1.5‰ between populations of ES and EI $\delta^{18}\text{O}_{\text{enamel}}$ values with a standard deviation of 1.4‰. When a significance level of (α) = 0.10 is used, sample sizes of 8 are acceptable, although again, our samples represent pooled data from several sites. Although the sample sizes of specimens are not large enough to satisfy the statistical demands of Levin, et al. (2006), they are comparable to

571 sample sizes of some datasets used in calculating their original aridity index.

572 We calculate the water deficit for the Kapthurin Formation during this time interval as 143 mm
573 by rewriting the equation in Figure 5 as:

574 Eq 6: $WD = (\epsilon_{ES-EI} + 2.348)/0.004$

575 This indicates a water deficit most similar to the modern Ituri Rainforest (-90 mm), based on the
576 oxygen isotope enrichment among Kapthurin ES and EI taxa, $\epsilon_{ES-EI} = -1.77$. When questionable taxa,
577 such as alcelaphines and bovines (Bedaso, et al. 2013; Levin, et al. 2006) are removed from the EI
578 calculations, a similar oxygen isotope enrichment between ES and EI taxa is found, $\epsilon_{ES-EI} = -1.73$,
579 suggesting water surplus of 154 mm. The similar ϵ_{ES-EI} values of questionable and strict ES and EI
580 taxa indicate that this was a very mesic period, quite different from the conditions predicted by overall
581 drying trends inferred from other Pleistocene fossil localities (deMenocal 2004), and more similar to a
582 variable period of increased rainfall (Potts 1996). The resemblance of deficit values for this time
583 interval to the modern Ituri Rainforest's water deficit, does not necessarily indicate a dense closed
584 rainforest, but may instead suggest a period of increased precipitation from monsoonal activity, as
585 suggested by Rossignol-Strick, et al. (1998) and Clark, et al. (2006), or a period of greatly reduced
586 potential evapotranspiration. Given the elevated greenhouse gasses during this time period (Higgins, et
587 al. 2015a), we expect that temperature conditions were warm and were coupled with increased
588 potential evapotranspiration in Equatorial Africa (Lüthi, et al. 2008), however these warm conditions
589 would also be offset by the increased precipitation from the intensified African monsoon hypothesis for
590 MIS 13 (Rossignol-Strick, et al. 1998),. If this interpretation is correct, we may infer a more precise
591 age of ~533 ka (Lisiecki and Raymo 2005) for the fauna under consideration here.

592 The average $\delta^{18}O_{\text{drinking water}}$ values, as estimated by applying Equations 2 and 3 to obligate
593 drinker taxa, is projected at $-0.9 \pm 1.5\%$. This estimate is more positive than the average $\delta^{18}O_{\text{drinking}}$

594 $\delta^{18}\text{O}_{\text{water}}$ value ($-2.5 \pm 2.4\text{‰}$) of several modern sources in Kenya provided in Levin, et al. (2009), but
595 within the standard deviation. This estimate is somewhat higher than other Pleistocene $\delta^{18}\text{O}_{\text{drinking water}}$
596 values. Bedaso, et al. (2010) estimate the average drinking water values at -1.9‰ for the Middle
597 Pleistocene of the Afar region of Ethiopia. The higher estimates from the Kapthurin Formation may be
598 due to evaporation during dry seasons, and possibly the effects of Lake Baringo as a source for water;
599 modern Lake Baringo has a $\delta^{18}\text{O}_{\text{drinking water}}$ value of 6.6‰ (Levin, et al. 2009), and thus represent the
600 most positive estimate for Kapthurin $\delta^{18}\text{O}_{\text{drinking water}}$ values during this time interval. This discordant
601 relationship between $\delta^{18}\text{O}_{\text{drinking water}}$ values and the aridity index merits further exploration. Our
602 sample sizes may not be robust enough for calculation of the aridity index and could be indicating a
603 much wetter environment than is merited. To determine the effects of small sample sizes, we have
604 contrasted the aridity index results with oxygen values from pedogenic carbonates below (see section
605 4.4). Taxonomic diversity is another possible issue, although the family Bovidae should be consistent
606 enough to provide a relative indicator of different habitats (Bibi, et al. 2009). However, the lack of very
607 positive $\delta^{18}\text{O}_{\text{enamel}}$ values in C_3 browsing taxa is a faithful indicator of mesic environments, due to the
608 exacerbated ^{18}O enrichment associated with leaf water in more xeric environments (Dongmann, et al.
609 1974). The negative $\delta^{18}\text{O}_{\text{enamel}}$ values of browsing taxa are accurate indicators of these mesic
610 conditions. The lack of C_3 dominated environments and the presence of C_4 habitats during this time
611 interval indicates that a monsoonal system that fed riparian woodland environments and significant
612 grassland ecosystem components is the most likely explanation for the low water deficit based on the
613 Kapthurin Formation aridity index.

Figure 5: Relationship between the enrichment of oxygen isotopes in tooth enamel of modern and fossil ES and EI taxa, ϵ_{ES-EI} , and aridity (Water Deficit), based on calibration data from Levin, et al. (2006). All modern environments provided in Levin, et al. (2006) with both ES and EI taxa are displayed. Site location and sample sizes are listed next to plotted values. Only taxa that were positively identified as ES or EI in Levin, et al. (2006) are included. Kapthurin data are shaded. Error bars represent standard deviations of oxygen isotope values of both ES and EI

4.3 Paleo-environmental interpretations from tooth enamel

Sample sizes are generally not large enough at individual paleontological sites for statistical comparisons among sites. In light of this, we present environmental indications based on pooled stable isotope values for the time interval under study. With this procedure, some time averaging among sites that are overlain by the Grey Tuff (509 ± 9 ka) must be assumed. The variety of habitats and paleosurfaces may well represent individual short time intervals within the larger time interval (543 ± 3 ka to 509 ± 9 ka) under study. This is likely, due to the sedimentology of the deposit, tectonic and erosional history of the Lake Baringo Basin, and modern geomorphology found in the area today (Martyn 1969; Tallon 1978). However, the preservation of the teeth within this large time interval appears to coincide with increased monsoonal activity during MIS 13 (~ 533 – 478 ka) (Lisiecki and Raymo 2005), and may have caused an increased sedimentation rate, creating ideal conditions for the

631 rapid burial of fossils. We also assume that fossilized animal remains are faithful indicators of past
632 habitats, based on the work of Western and Behrensmeyer (2009) and Behrensmeyer, et al. (2007).
633 Within these limitations, we assume that the average carbon and oxygen isotopic values of bovid tooth
634 enamel represent the general paleoenvironmental setting of the Lake Baringo Basin, allowing for the
635 selective feeding preferences of some species, during the time interval under study (543 ± 3 ka to $509 \pm$
636 9 ka).

637   Average stable isotope values from serial tooth enamel data indicate a diverse environment
638 between 543 and 509 ka, although oxygen isotope values suggest that the fauna was deposited during a
639 monsoonal episode ~ 533 ka (Lisiecki and Raymo 2005; Rossignol-Strick, et al. 1998). Individual
640 serial enamel values from Table 1 show that both wooded and open habitats were available to hominins
641 during this time interval, but overall the stable isotope values from Figures 5, 6, and 7 indicate that the
642 habitats were very mesic. There is little evidence for mixed C_3 and C_4 foragers, possibly due to
643 sampling bias. The aridity indices calculated for this time interval (after Levin, et al. 2006), suggest a
644 small water deficit and are most comparable to values for environments in the modern Ituri Rainforest
645 (Figure 5). While these results suffer from small sample sizes of EI and ES for individual taxa, there
646 are not any apparent elevated $\delta^{18}O_{\text{enamel}}$ values associated with C_3 browsing taxa, a faithful indicator of
647 mesic conditions (Dongmann, et al. 1974). These results are somewhat consistent with other
648 researchers' paleoenvironmental interpretations based on faunal taxonomic identification and geologic
649 observations, although these hyper-mesic conditions suggest a wetter interval than previously predicted
650 (Johnson, et al. 2009; McBrearty 1999; Renaut, et al. 2000; Tallon 1978). Our data indicate that the
651 average paleoenvironmental conditions during this time period were characterized by wooded and
652 marsh conditions near perennially active rivers and lake-shores, with semi-arid grassland components
653 of the landscape at some distance from these water sources. These results are consistent with a
654 strengthened monsoonal system affecting East Africa during MIS 13, as has been suggested elsewhere

655 (Clark, et al. 2006; Rossignol-Strick, et al. 1998). Substantial precipitation is suggested by the
656 paleochannels observed in the fluvial facies (K3) of the formation (Figure 1) (Tallon 1978), and
657 suggest a more precise date for the fossil fauna at the onset of MIS 13 (~533 ka) (Lisiecki and Raymo
658 2005).

659 Archaeological remains of hominins from this interval seem to indicate exploitation of several
660 different habitats (Figures 4 and 6), although higher resolution environmental reconstructions are
661 necessary to test this hypothesis. Hominin fossil remains have been found at the site of EKG (Leakey,
662 et al. 1969; Wood and Van Noten 1986), near the sites of GnJh-38 and Loc. 4. The combined isotope
663 data from these locations indicates a riparian woodlands, with a drier grassland environment at some
664 distance from the water source (closest to the site of EKG). We suggest that based on the average
665 paleoenvironmental reconstructions for this time period, hominins would have had the opportunity to
666 exploit many rich local habitats, ranging from wet, densely vegetated marsh and forests, to drier
667 grassland landscapes. Higher resolution environmental reconstructions are necessary before we can
668 begin to test the relationship of hominin technological choices to explicit and site microecology.

669

670 *4.4 Carbon and oxygen composition of pedogenic carbonates*

671 Pedogenic carbonates $\delta^{13}\text{C}_{\text{pc}}$ values average $-3.5 \pm 2.3\text{‰}$ and range from -9.6 to 0.2‰ ,
672 indicating that wooded grassland ($n=12$) was the dominant form of vegetation cover recorded by
673 pedogenic carbonates, indicated by the average $f_{\text{wc}} = 0.19$ (Table 2 and Figure 6). Notably, the
674 grassland category is also prevalent ($n=9$). Finally, there is some evidence for the
675 woodland/bushland/shrubland category ($n=3$). There is no evidence for densely forested environments
676 preserved in the pedogenic carbonate record. These reconstructions augment the dietary
677 reconstructions of fossil bovids, and provide evidence for all of the habitats reconstructed in section
678 4.1. Interestingly, the area with the most wooded cover (66%) is Loc. 99, the fossil chimpanzee locality

Sample	$\bar{x} \delta^{13}\text{C}$	$\sigma_{\delta^{13}\text{C}}$	$\bar{x} \delta^{18}\text{O}$	$\sigma_{\delta^{18}\text{O}}$	n	$\bar{x} f_{\text{wc}}$	Site
DL-1-2011	-2.1	0.57	-2.9	0.28	8	0.09	Below GT
DL-2-2011	-2.2	1.37	-4.8	0.98	20	0.11	Below GT
DL-3-2011	-2.1	0.27	-5.7	0.52	16	0.09	Below GT
DL-4-2011	-2.7	0.79	-5.6	0.59	12	0.13	Below GT
DL-5-2011	-2.7	0.32	-6.8	0.60	10	0.12	Below GT
DL-6-2011	-1.1	0.46	-4.5	0.45	10	0.05	Below GT
DL-7-2011	-0.3	0.94	-4.6	0.65	10	0.03	Below GT
DL-7-1-2011	-1.7	1.28	-5.6	2.08	10	0.08	Below GT
DL-8-2011	-3.7	0.89	-3.8	0.40	10	0.19	Below GT
DL-9-2011	-3.1	0.90	-4.5	1.65	10	0.15	Below GT
DL-9-1-2011	-2.6	0.48	-5.2	1.07	10	0.12	Below GT
DL-10-2011	-1.5	0.48	-7.8	0.84	10	0.06	Below GT
DL-10-1-2011	-1.6	0.12	-8.6	0.38	10	0.07	Below GT
DL-1-2012	-6.5	0.29	-3.7	0.41	10	0.40	GnJh-39
DL-2-2012	0.2	0.41	-0.4	0.93	6	0.01	GnJh-39
DL-3-2012	-4.4	0.89	-3.4	0.39	15	0.24	Below GT
DL-4-2012	-4.8	0.99	-3.5	0.54	15	0.27	GnJh-39
DL-5-2012	-4.8	0.50	-2.5	0.57	10	0.26	Below GT
DL-7-2012	-5.6	0.21	-2.9	1.49	9	0.32	EKG
DL-9-2012	-6.8	0.24	-3.5	0.54	10	0.42	BKS
DL-10-2012	-6.0	0.27	-4.1	0.27	10	0.35	JKS
DL-11-2012	-5.3	0.22	-5.4	0.28	10	0.30	GnJh-23
DL-15-2012	-9.6	0.15	-3.8	0.29	3	0.66	Loc. 99
DL-20-2012	-2.3	1.39	-4.1	0.42	10	0.11	GnJh-27

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681 Table 2: Average stable isotope values and standard deviations of serially sampled fossil
682 pedogenic carbonates. Fraction Woody cover estimates and standard deviations are also
683 displayed. Below GT refers to pedogenic carbonates collected from paleosols
stratigraphically below the Grey Tuff (543 ka). N=number of samples, $\bar{x}$ =average,
 σ =standard deviation, and f_{wc} =fraction woody cover estimate.

Figure 6: Average stable isotope values $+1\sigma$ of serial pedogenic carbonate samples, displayed against x-axis ($\delta^{13}\text{C}$) and y-axis ($\delta^{18}\text{O}$). Samples are shown with respect to assumed vegetation type. Shaded and light areas reference C_3 , Mixed C_3/C_4 , and C_4 environments.

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Pedogenic carbonate $\delta^{18}\text{O}_{\text{pc}}$ average $-4.5 \pm 1.7\text{‰}$ and range from -8.6 to -0.4‰ , indicating a

diverse range of oxygen values (Figure 6 and Table 2). The lowest $\delta^{18}\text{O}_{\text{pc}}$ values are the best estimates

for meteoric water $\delta^{18}\text{O}$ values, because they represent carbonate formation that records the least

amount of evaporation. Samples DL-10-2011, and DL-10-1-2012, represent the lowest oxygen isotope

values (-7.8 and -8.6‰ respectively), and were sampled from a paleosols adjacent to the Chemichemi

Tufa (Johnson, et al. 2009), a record of a perennially active stream that was situated near lake margins

and was fed by increased ground water inputs during this time interval (543-509 ka). These minimum

values are also lower than minimum values of $\delta^{18}\text{O}_{\text{pc}}$ provided in Levin, et al. (2011) from the Turkana

Basin (Nachukui Formation, ~ 750 ka, $\sim -3\text{‰}$). The minimum $\delta^{18}\text{O}_{\text{pc}}$ values from the Kapthurin

Formation can also be juxtaposed by the average $\delta^{18}\text{O}$ (VSMOW) values of bodies of water in the Lake

Baringo Basin today, the lake has an average value of 6.6‰ and the Molo River has a an average value

of -2.9‰ (Levin, et al. 2009). These results suggest a maximum difference of -5.7‰ in oxygen isotope

values of meteoric water during this time interval and the present, which may indicate lower meteoric

water values from rainfall, a shift in aridity, or a difference in soil temperature. The latter possibility

can be fully excluded as greenhouse gas concentration (CO_2 and CH_4) in the atmosphere is elevated

700 during this time interval (Higgins, et al. 2015b). The average $\delta^{18}\text{O}_{\text{pc}}$ value (-4.5‰) from this time
701 interval is also lower than the average $\delta^{18}\text{O}$ (VSMOW) value of meteoric water in Kenya today (-
702 2.9‰). Interestingly, the average $\delta^{18}\text{O}_{\text{pc}}$ value is over 3‰ lower than the projected drinking water
703 value from ES taxa (-0.9‰), indicating that the ES taxa probably incorporated significant components
704 of standing or other evaporated water into their drinking water. Taken together with the evidence from
705 the fossil bovid record, we suggest that these values indicate a period of increased rainfall, that are not
706 obliterated by high evaporative climate conditions associated with insolation maximum and overall
707 warmer global conditions during MIS 13.

708

709 **5. Conclusions**

710 We have described here isotopic analyses of bovid fossil teeth and pedogenic carbonates from
711 the Kapthurin Formation, collected from localities within the Middle Silts and Gravels (K3) member of
712 the Kapthurin Formation, stratigraphically above the Pumice Tuff (K2) (543 ± 3 ka) and below the
713 Grey Tuff (509 ± 9 ka). The teeth represent taxa from six tribes; they were sampled serially to
714 reconstruct dietary vegetation and ancient habitats. These fauna represent the full range of dietary
715 adaptations, ranging from pure C_3 to pure C_4 , although there is little evidence for mixed (C_3/C_4)
716 foraging behavior. The stable oxygen isotope data suggests that most bovid taxa were EI, tracking
717 meteoric water faithfully, while a few were ES, their values tracking the oxygen isotope composition of
718 the C_3 vegetation they consumed. Applying the modified aridity index of Levin, et al. (2006), we
719 estimate the oxygen enrichment between these two groups as $\epsilon_{\text{ES-EI}} = -1.77$, approximating a
720 paleowater deficit of 143 mm. The $\delta^{13}\text{C}_{\text{pc}}$ values suggest a landscape dominated by wooded grassland
721 vegetation, with open grassland and woodland influences, and the $\delta^{18}\text{O}_{\text{pc}}$ values suggest a wetter
722 environment than the modern basin, and confirm the presence of a perennially active stream during this
723 time interval. These values suggest a paleohabitat that was perforated with very mesic and mildly xeric

724 components, with riparian woodlands, sumplands, and fringe grasslands ecosystems.

725   These general paleoenvironmental indicators suggest a habitat that was rich in biodiversity, and
726 accords well with faunal identifications that have portrayed the environment as closed, wet, and marshy
727 habitats near an alternating alkaline and freshwater lake with arid components farther afield
728 (McBrearty 1999). A Middle Pleistocene African monsoonal system that dominated this time interval
729 is the most likely explanation for the rapid sedimentation rates, numerous paleochannels, good fossil
730 preservation and burial, and modified paleoaridity index of -1.77 during this time interval (~543 – 509
731 ka). This research represents the first confirmation of this Middle Pleistocene African monsoonal event
732 with African paleoecological data (Clark, et al. 2003; Rossignol-Strick, et al. 1998). This
733 paleolandscape would have shown exceptional ecological diversity, providing hominins with a number
734 of different local habitats and a broad range of resources. From the archaeological evidence, it seems
735 that hominins exploited both wet and dry locales with wooded, marshy, and grassland vegetation.
736 Hominins during this time period also began to experiment with the systematic production of blades
737 (Johnson and McBrearty 2010), a new technology which may have been developed as a response to the
738 ecological diversity of the Lake Baringo Basin, although higher resolution environmental
739 reconstructions are necessary to verify this suggestion.

740   Shifts in environments have been cited as stimuli for evolutionary and technological change
741 among hominins (Ambrose 1998; Bräuer 2008; Darwin 1871; deMenocal 1995; Potts 1996; Reed
742 1997; Vrba 1993). Researchers have noted a general cooling trend during the Pliocene that continues
743 into the Pleistocene, resulting in more xeric and open grassland conditions in East Africa (Bobe and
744 Behrensmeyer 2004; Coppens 1975; deMenocal 1995; deMenocal 2004; Vrba 1985). Isotopic data
745 from the Kapthurin Formation suggests, contrary to expectations (deMenocal 1995), that during this
746 time interval (543 – 509 ka) the Lake Baringo region provided hominins and chimpanzees with hyper
747 mesic, wooded refugium, in addition to nearby grasslands, both sources of valuable resources.

748 Increased monsoonal activity at the onset of MIS 13 (~533 ka) may have driven the appearance of this
749 hyper mesic environment, which may have stimulated the production and use of new technologies,
750 although further exploration and additional examples are necessary to test this hypothesis. This
751 correlation allows for more precision when dating the Kapthurin Formation sediments deposited during
752 K3 and K3', and suggests a time interval of ~533 – 509 ka. This reconstruction accords with those of
753 Kingston, et al. (1994) who noted that the Tugen Hills succession does not provide evidence for a shift
754 to C₄ dominated environments throughout the Neogene and Pliocene, and that a range of local habitats
755 existed at any given point in time. These results highlight the need to ground global or regional
756 climate reconstructions in local paleoecological data from individual archaeological and
757 paleontological sites.

758

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