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Middle Pleistocene climate and habitat change at Zhoukoudian, China, from the carbon and oxygen isotopic record from herbivore tooth enamel

Mabry Gaboardi[a,*], Tao Deng[b], Yang Wang[a]

[a]Department of Geological Sciences, 108 Carraway Building, Florida State University Tallahassee, Florida 32306-4100, USA
[b]Chinese Academy of Sciences, Institute of Vertebrate Paleontology and Paleoanthropology, Beijing 100044, China

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Abstract

The Pleistocene deposits at Zhoukoudian, often referred to as the "Peking Man" site, contain dental remains from a diverse group of herbivores, including *Equus sanmeniensis*, *Cervus elaphus*, *Cervus nippon*, *Megaloceros pachyosteus*, *Sus lydekkeri*, and *Dicerorhinus choukoutienensis*. The carbon and oxygen isotopic compositions of structural carbonate within the enamel of these teeth are used to reconstruct the paleodiet and paleoenvironment of the mammals. The δ^{13}C values of enamel from Zhoukoudian range from −2.3‰ to −13.0‰, indicating that these mammals consumed between ~25% and 100% C_3 plants. The presence of significant amounts of C_4 plants in the diets of some herbivore species indicates that at the onset of the Middle Pleistocene local habitats included mixed C_3/C_4 vegetation. By approximately 470,000 yr ago, C_3 plants dominated the diets of herbivores studied, suggesting that the abundance of C_4 flora had decreased in the area. For all deer analyzed in this study, the values of δ^{13}C and δ^{18}O decrease substantially from about 720,000 to 470,000 yr ago. This trend may be due to a strengthening of the winter monsoon during the Middle Pleistocene.

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Keywords: Zhoukoudian; Pleistocene; Carbon isotopes; Oxygen isotopes; *Homo erectus*; Paleomonsoon; Paleoecology

Introduction

Zhoukoudian, located about 50 km to the southwest of Beijing (see Fig. 1), is of great significance in human evolution studies. Fossils attributed to *Homo erectus*, originally called *Sinanthropus pekinensis*, and *Homo sapiens* attest to the intermittent occupation of the area by hominids during the Middle and Late Pleistocene. Discovered among the plentiful hominid remains were quartzite flakes and choppers, burnt bones, abundant vertebrate fossils, and possible ash (Black et al., 1933; Jia, 1980).

In an attempt to better understand the local climate and habitat in which early humans lived, many traditional paleoclimatological studies have been performed at the "Peking Man" site, including pollen analysis, faunal assemblage studies, and sedimentological investigations (Goldberg et al., 2001; Jia, 1980; Liu, 1985; Wu et al., 1985; Zhu and Zhou, 1994). An independent method for reconstructing the local paleoclimate has been established involving the analysis of isotope fractionation within the enamel of herbivore molars. Because herbivore teeth are abundant within the sedimentary deposits at Zhoukoudian, this type of analysis can be of use to paleontologists and anthropologists attempting to piece together the paleoenvironment in which hominids at Zhoukoudian lived. In addition, paleoclimate and paleovegetation records from this site can be compared with other records, such as the deep-sea oxygen isotope records and the Chinese loess-paleosol sequences. This can improve our understanding of the natural variability of the Earth's climate system and how terrestrial plants have had to adapt to changes in the climate.

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* Corresponding author.
E-mail address: gaboardi@gly.fsu.edu (M. Gaboardi).

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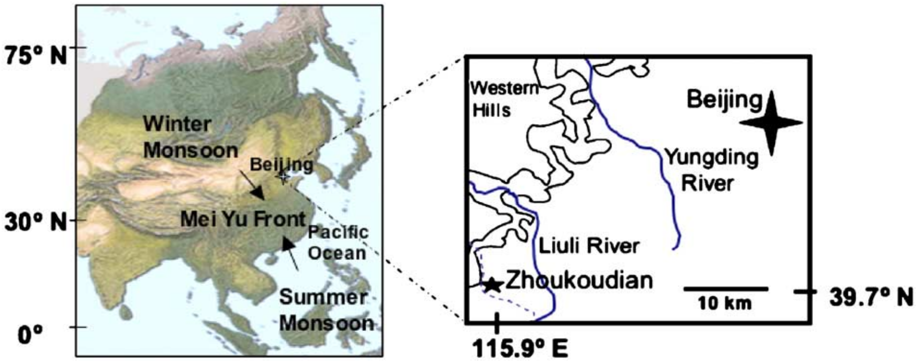

Figure 1. Map showing the monsoonal systems that influence Beijing, China and the location of Zhoukoudian in relation to Beijing.

In this study, the carbon and oxygen isotopic ratios of structural carbonate in fossil tooth enamel from a diverse group of herbivores from Zhoukoudian are analyzed. The objective is to determine the ratio of C_3 to C_4 plants in the diet of these species and reconstruct the paleoenvironment at the site.

Stratigraphy, chronology, and geographic setting of the site

Within Zhoukoudian is a series of caves produced by the intersection of two vertical joints in the surrounding Ordovician limestone. The three caves sampled in this study include Locality 1, Locality 22, and Upper Cave. The main cave, Locality 1, is divided into 17 stratigraphic units. The deposits within this cave are indicative of a "gradually filled, open-air cave" environment (Goldberg et al., 2001). Vertebrate fossils have been found in the 13 uppermost units (Lin, 1994). *Homo erectus* fossils are found in Layer 10 to Layer 3, as are many artifacts. Reviews of Locality 1 can be found in Boaz and Ciochon (2004), Wu and Poirier (1995), and Wu et al. (1985).

Many dates have been published for the layers of Locality 1. Table 1 displays the estimated layer ages used in this paper. It should be noted that most previous paleoclimate reconstructions assign much younger dates to the upper layers of Locality 1 (e.g., Xu et al., 1997). These younger dates are based largely on electron spin resonance (ESR) and uranium-series (U-series) dating of bone, tooth, or antler material. Because the dated material is not a closed system, assumptions about the uptake of uranium must be made before dates can be assigned. The authors accept as most reliable the U-series date of Shen et al. (1996) and the correlation of the Brunhes–Matuyama (B/M) boundary by Qian et al. (1985). U-series dating was done on a speleothem—a closed system, with high-precision thermal-ionization mass spectrometry (TIMS) to produce a date of 414,000 ± 13,000 yr for Layer 2 (Shen et al., 1996). ^{40}Ar/^{39}Ar dating was performed on the B/M boundary to get a date of 780,000 yr (Spell and McDougall, 1992). This reversal is correlated with Layer 14 of Locality 1 by Qian et al. (1985).

Zhou et al. (2000) use these dates and a synthetic climatic index curve to correlate the layers of Locality 1 with the marine oxygen isotope (MOI) stages and the Baoji section of the Chinese loess. This correlation can be seen in Figure 2, along with an alternative correlation by Xu et al. (1997). Using their correlation, Zhou et al. (2000) extrapolate approximate ages for the layers of Locality 1. These estimated ages are used in this paper to construct a time scale for interpreting isotopic data. Because of uncertainty in dating these layers, samples in this study are attributed to a layer rather than to an absolute age.

Additionally sampled are Upper Cave and Locality 22, also within Zoukoudian. Accelerator radiocarbon dating of fossil bone from this layer produces dates of 33,200 ± 820 and 33,460 ± 2000 ^{14}C yr B.P. (Chen et al., 1992).

Locality 22 is a fissure deposit located 40 m above the modern Zhoukou River (Lin, 1994). No absolute dates have been measured from this site. Based on relative dating of fauna, Locality 22 is believed to be younger than Locality 1 and older than Upper Cave, with an estimated age range of 100,000 to 200,000 yr old (Chia et al., 1959).

Geographically, Zhoukoudian is at the western edge of the Great North China Plains, just to the east of the Western Hills (see Fig. 1). Zhoukoudian is named after the village located to the south, but the Chinese name for the site itself is Longgushan, or Dragon Bone Hill (Boaz and Ciochon, 2004). Locality 1, Upper Cave, and Locality 22 are all located on Dragon Bone Hill, between 40 and 160 m above sea level (Liu, 1985). The Zhoukou River winds past Zhoukoudian to the west. Water level varies with precipitation from almost dry in the spring to turbulent in the late summer (Lin, 1994). Currently, the average temperature is 11.8°C and the average annual precipitation is about 650 mm with most of the precipitation falling in the summer. There are strong seasonal variations in both temperature and precipitation in the area due to the influence of the summer and winter monsoonal systems.

Table 1
Dates (10^3) of stratigraphic units at locality 1

Layer	^{40}Ar/^{39}Ar age of the Brunhes–Matuyama magnetic reversal	TL age, fired quartz grains	U-series age, tooth, bone, or antler (except* = speleothem)	Fission-track age, sphene from ash	ESR Age tooth, bone, or antler (except* = speleothem)	Age estimates from Zhou et al. (2000)
1/2			***414 ± 13**[7] 230 +30/−23[11] *421 +110/−54[8]		*221 ± 84[3]	420
3			249 ± 51[10] 256 +60/−40[11]		283 ± 30; 312 ± 36[1] 363 ± 48; 386 ± 54[1] 282 ± 45[3]	470
4		292 ± 26[5] 321 ± 28[5]		306 ± 56[2] 299 ± 55[2]		500
5						510
6					278 ± 22; 376 ± 28[1] 325 ± 21; 404 ± 27[1]	570
7					368 ± 85[4] 385 ± 85[4] 396 ± 46[4]	590
8/9			>350[11]		418 ± 48[3] 423 ± 80[4]	640
10		417[5] >592[5]	340 +100/−60[11]	462 ± 45[2]	326 ± 22; 448 ± 30[1] 382 ± 38; 594 ± 59[1]	670
11					585 ± 105[4] 578 ± 66[3]	680
12			310 ± 70/−50[11] 320 ± 86[11]		669 ± 84[3]	690
13						720
14	**780**[6 and 9]					780

Data are cited from the authors indicated: 1—Grun et al. (1997) (early uptake; linear uptake estimates); 2—Guo et al. (1991); 3—Huang et al. (1991); 4—Huang et al. (1993); 5—Pei (1985); 6—Qian et al. (1985); 7—Shen et al. (1996); 8—Shen and Jin (1991); 9—Spell and McDougall (1992); 10—Yuan and Chen (1980); 11—Zhao et al. (1985).

Dates in bold print are those which the authors accept as the most reliable. Estimated ages (based on bold dates) from Zhou et al. (2000) are used within this paper to construct a time scale against which to report isotopic data. Due to uncertainty with numeric age assignment, the Layer within which the tooth was found is reported with data.

The monsoons are global systems driven by differing ocean-continent heat capabilities. The summer monsoon is powered by warm air from above the Pacific Ocean moving onto the Asian continent. When it reaches East Asia from the south and east, it is laden with moisture and, near 30°N, encounters cold air flowing from the northwest. The contact of these two air masses forms the Mei-Yu front that drops large amounts of rain over Eastern Asia. In warmer months the Mei-Yu front may shift north, dropping precipitation at increasingly higher latitudes. By mid-September, the continent cools sufficiently to form a high-pressure cell at about 100°E, 50°N. Cold air then flows down from the Mongolia–Siberia region. Thus the winter monsoon is characterized by cold, dry, continental air which produces frigid temperatures and very little precipitation as far south as the South China Sea (Liu and Ding, 1998). As these monsoons are interconnected to global circulation patterns, changes in local, monsoonally driven weather should be related to global climate changes.

In the monsoon regions in China, the stable isotopic composition of precipitation varies with monsoon intensity (Johnson and Ingram, 2004). Because Zhoukoudian is located almost 10° north of where the Mei-Yu front originates, the site has strong potential to record changes in the monsoonal intensity over time. These changes at Zhoukoudian should be observable in the oxygen isotopes preserved in tooth enamel.

The distribution of C_3 and C_4 vegetation in the area at present has been strongly modified by human settlement and crop cultivation. However, changes in the relative abundan-

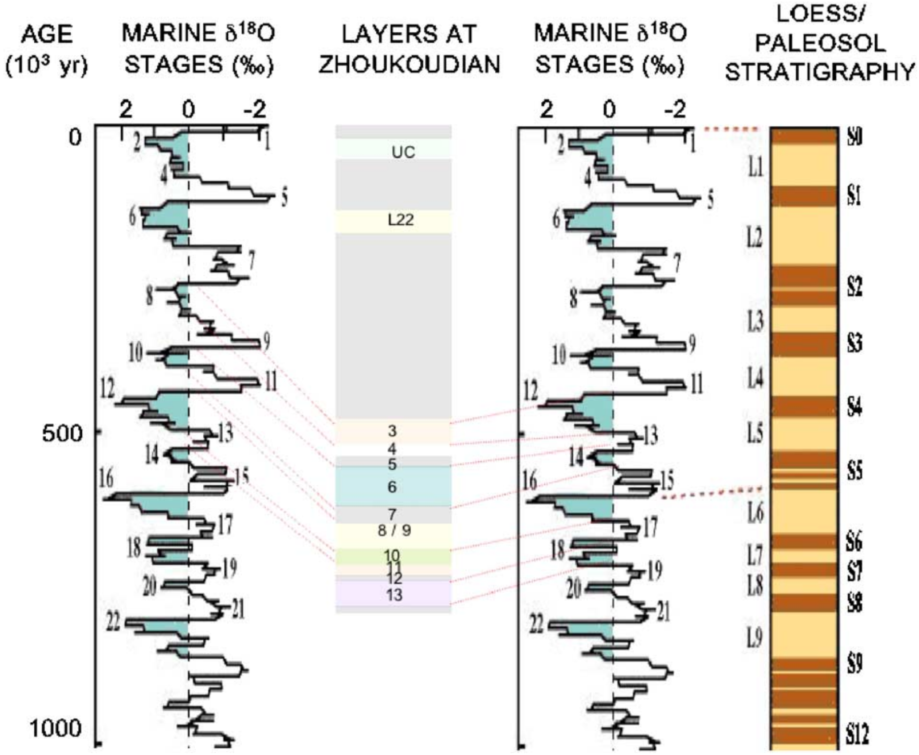

Figure 2. Correlation between layers at Zhoukoudian and marine oxygen isotopes stages based on Xu et al. (1997) on left. Correlation between layers at Zhoukoudian and marine oxygen isotope stages based on Zhou et al. (2000) on right. Even stages correspond to glacial periods; odd stages are interglacial. Loess/paleosol sequence from the Baoji section of the Chinese Loess Plateau on far right. "S" corresponds to paleosol; "L" corresponds to loess. Diagram of marine oxygen isotope stages and loess/paleosol stratigraphy modified from Porter (2001) (after Rutter et al., 1991; Williams et al., 1988).

ces of C_3 and C_4 plants over time are recorded in the loess record (Vidic and Montañez, 2004; Zhang et al., 2003) and can be estimated from the carbon isotopes of tooth enamel.

Rationale

Bioapatite is the mineral that makes up bones, tooth enamel, and dentine. All three materials have been used in previous studies with tooth enamel being the most resistant to diagenesis. For this reason, and because tooth enamel is not reworked throughout the life of the organism, enamel is the preferred medium from which to extract "structural" carbonate for carbon and oxygen isotopic measurements (Lee-Thorp and van der Merwe, 1987, 1991). Carbon isotopic ratios measured from tooth enamel indicate the proportion of C_3 and C_4 plants in the diets of the herbivorous species studied. C_3 plants, which photosynthesize through the Calvin cycle, comprise the majority of plant species including trees, shrubs, and cool season grasses. δ^{13}C values for C_3 plants range from −22‰ to −35‰ with an average of −27‰ (Fogel and Cifuentes, 1993; O'Leary, 1988; Smith and Epstein, 1971). C_4 plants, mostly warm season grasses, photosynthesize through the Hatch–Slack pathway. They are less depleted in ^{13}C, with a δ^{13}C range of −19‰ to −9‰ and an average of −13‰ (O'Leary, 1988; Smith and Epstein, 1971). Although C_3 plants have more variable δ^{13}C signatures due to moisture and/or light stress, they are consistently more negative than C_4 species.

During animal uptake and tooth enamel production, the δ^{13}C values from plants are further fractionated, with a consistent enrichment of 12–14‰ observed (Cerling and Harris, 1999; Kreuger and Sullivan, 1984; Lee-Thorp et al., 1989). A ratio of the C_3 to C_4 plants in the diet of the animals can be calculated using a simple mass balance equation. Changes in that ratio over time may indicate habitat changes. When coupled with δ^{18}O values, these changes may attest to fluctuations in climate at the locality.

The primary factor in determining the δ^{18}O of body water is the δ^{18}O of the meteoric water ingested by the animal (Longinelli, 1984). Other important factors include humidity, diet, and behavior (Luz and Kolodony, 1985; Luz and Kolodny, 1989). Because of the influences of diet and behavior, it is best to analyze a single genus over time and then compare inter-genus trends in order to examine environmental changes (Koch et al., 1994; Kohn et al., 1998). Oxygen isotope fractionation in meteoric water is mass dependent and follows the Raleigh distillation process such that it is affected by the amount of original vapor remaining and the temperature of original condensation (Dansgaard, 1964). With colder temperature and/or more rainout, δ^{18}O values become more negative.

Methods

Following an established procedure (Wang et al., 1994), the enamel was manually cleaned and separated from the dentine using a hand-held Dremel© and approximately 300 mg of tooth was then ground into a fine powder. The powder was then treated with weak (5%) sodium hypochlorite overnight to remove any remaining organics, rinsed with deionized water over glass filter paper and freeze dried. Next, the powder was allowed to react overnight under a weak vacuum with 1 M acetic acid to eliminate any nonstructural carbonates. It was again washed and freeze dried. About 100 mg of the bioapatite powder was placed in a tube, evacuated on a vacuum line, and allowed to react for about 48 h with 100% phosphoric acid at 25°C. The CO_2 produced was extracted using a vacuum line and analyzed on a Mat Delta S Finnigan mass spectrometer. The isotopic results are reported in the standard per mil (‰) notation as $\delta^{13}C$ and $\delta^{18}O$ relative to the international carbonate standard VPDB:

$$\delta = \left[\left(R_{sample}/R_{standard}\right) - 1\right] \times 1000 \qquad (1)$$

where $R = {}^{13}C/{}^{12}C$ or ${}^{18}O/{}^{16}O$. Analytical precision for both $\delta^{13}C$ and $\delta^{18}O$ is better than 0.1‰.

We analyzed 70 samples from six species from Layers 13 through 3 at Locality 1 and from Locality 22 and Upper Cave. The species include *Equus sanmeniensis* (horse), *Megaloceros pachyosteus* (Irish elk), *Dicerorhinus choukoutienensis* (rhino), *Cervus nippon* (sika deer), *Cervus elaphus* (red deer), and *Sus lydekkeri* (pig).

Results and discussion

$\delta^{13}C$ of tooth enamel

The $\delta^{13}C$ values of tooth enamel from Zhoukoudian fall between −2.3‰ and −13.0‰ (Figs. 3 and 4, Table 2), indicating that the diets of these herbivores consisted of ~25% to 100% C_3 plants. Some species are isotopically distinct from others, reflecting differences in dietary preferences and resource utilization. The data show that *E. sanmeniensis* and *C. nippon* had incorporated some C_4 plants into their diets, whereas *D. choukoutienensis* and *M. pachyosteus* were feeding predominantly on C_3 plants. The presence of significant amounts of C_4 plants in the diets of horses (*E. sanmeniensis*) and deer (*C. nippon*),

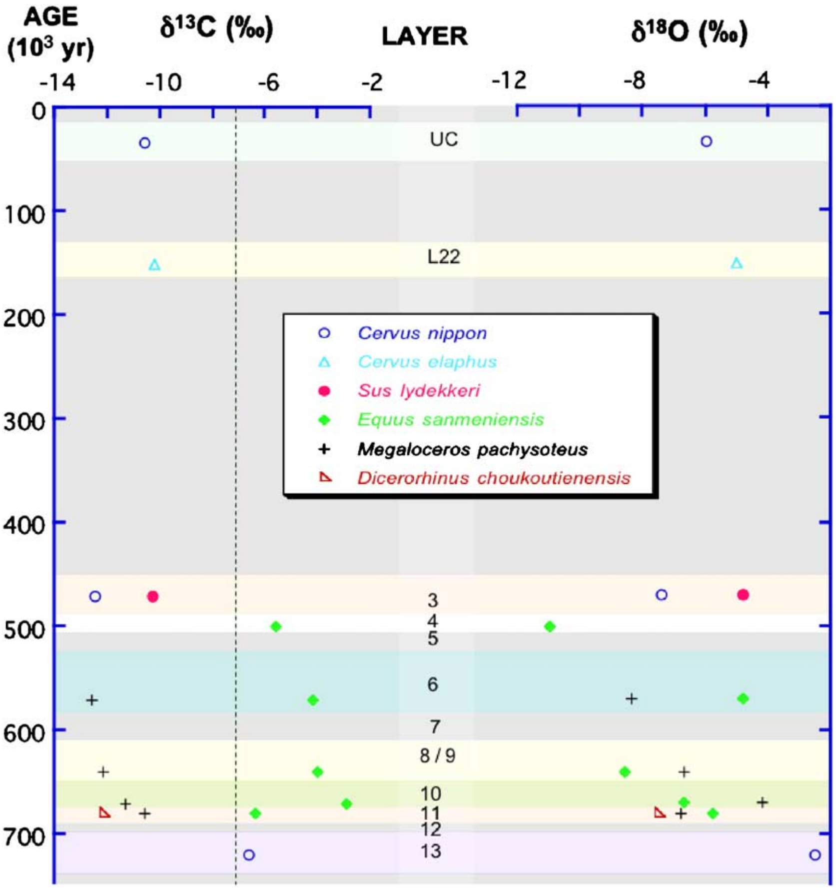

Figure 3. Variations of the means of the $\delta^{13}C$ and $\delta^{18}O$ values of fossil tooth enamel over time in relation to stratigraphic layers at Zhoukoudian. UC = Upper Cave, L22 = Locality 22, and all numbered Layers are from Locality 1. In this study, *Cervus* (*Pseudaxis*) *grayi* and *Cervus hortulorum* are classified as subspecies of the living sika, *Cervus nippon*.

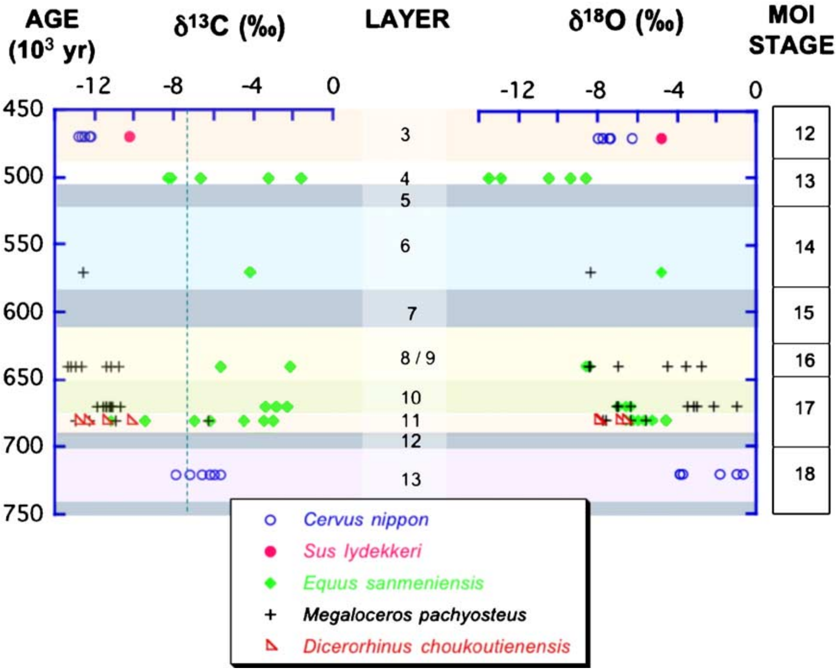

Figure 4. Variations of the δ^{13}C and δ^{18}O values of fossil tooth enamel over time in relation to stratigraphic layers at Locality 1 of Zhoukoudian. In this study, *Cervus* (*Pseudaxis*) *grayi* and *Cervus hortulorum* are classified as subspecies of the living sika, *Cervus nippon.* Marine oxygen isotope stages are as correlated by Zhou et al. (2000).

suggests that C_4 vegetation was an important component of local biomass from the time of deposition of Layer 13 to Layer 4. The isotopic differences among the species suggest that local habitats likely included mixed C_3/C_4 grasslands and woodlands at the onset of the Middle Pleistocene.

Table 2
Carbon and oxygen isotopic compositions of fossil tooth enamel from Zhoukoudian

Museum number	Species	Layer/location	δ^{13}C	δ^{18}O
IVPP TS 269	*Cervus nippon*	Upper Cave	−12.1	−8.4
IVPP TS 270	*Cervus nippon*	Upper Cave	−10.6	−5.2
IVPP TS 271	*Cervus nippon*	Upper Cave	−11.4	−6.1
IVPP TS 272	*Cervus nippon*	Upper Cave	−11.0	−6.3
IVPP TS 273	*Cervus nippon*	Upper Cave	−10.6	−4.9
IVPP TS 274	*Cervus nippon*	Upper Cave	−9.7	−5.3
IVPP TS 275	*Cervus nippon*	Upper Cave	−10.6	−5.8
IVPP TS 276	*Cervus nippon*	Upper Cave	−11.0	−6.2
IVPP TS 277	*Cervus nippon*	Upper Cave	−9.9	−6.5
IVPP TS 278	*Cervus nippon*	Upper Cave	−10.8	−5.8
IVPP TS 279	*Cervus elaphus*	Locality 22	−9.7	−5.8
IVPP TS 280	*Cervus elaphus*	Locality 22	−9.6	−6.5
IVPP TS 281	*Cervus elaphus*	Locality 22	−10.2	−5.7
IVPP TS 282	*Cervus elaphus*	Locality 22	−10.3	−3.5
IVPP TS 283	*Cervus elaphus*	Locality 22	−10.3	−1.8
IVPP TS 284	*Cervus elaphus*	Locality 22	−10.9	−6.8
IVPP TS 251	*Sus lydekkeri*	Layer 3	−10.3	−4.8
IVPP TS 254	*Cervus nippon*	Layer 3	−12.8	−7.5
IVPP TS 255	*Cervus nippon*	Layer 3	−12.5	−6.3
IVPP TS 256	*Cervus nippon*	Layer 3	−12.3	−8.0
IVPP TS 257	*Cervus nippon*	Layer 3	−12.7	−7.8
IVPP TS 258	*Cervus nippon*	Layer 3	−12.2	−7.4
IVPP TS 203	*Equus sanmeniensis*	Layer 4	−6.7	−12.9
IVPP TS 209	*Equus sanmeniensis*	Layer 4	−8.2	−13.5
IVPP TS 210	*Equus sanmeniensis*	Layer 4	−1.6	−8.6
IVPP TS 212	*Equus sanmeniensis*	Layer 4	−3.3	−10.5
IVPP TS 218	*Equus sanmeniensis*	Layer 4	−8.3	−9.4

Table 2 (*continued*)

Museum number	Species	Layer/location	$\delta^{13}C$	$\delta^{18}O$
IVPP TS 217	*Equus sanmeniensis*	Layer 6	−4.2	−4.8
IVPP TS 236	*Megaloceros pachyosteus*	Layer 6	−12.6	−8.4
IVPP TS 202	*Equus sanmeniensis*	Layer 8–9	−5.7	−8.6
IVPP TS 205	*Equus sanmeniensis*	Layer 8–9	−2.2	−8.5
IVPP TS 234	*Megaloceros pachyosteus*	Layer 8–9	−13.4	−8.5
IVPP TS 235	*Megaloceros pachyosteus*	Layer 8–9	−12.7	−8.4
IVPP TS 237	*Megaloceros pachyosteus*	Layer 8–9	−11.2	−3.6
IVPP TS 238	*Megaloceros pachyosteus*	Layer 8–9	−13.2	−4.5
IVPP TS 239	*Megaloceros pachyosteus*	Layer 8–9	−10.8	−7.0
IVPP TS 240	*Megaloceros pachyosteus*	Layer 8–9	−11.4	−8.5
IVPP TS 241	*Megaloceros pachyosteus*	Layer 8–9	−13.0	−2.8
IVPP TS 206	*Equus sanmeniensis*	Layer 10	−2.9	−6.6
IVPP TS 207	*Equus sanmeniensis*	Layer 10	−3.4	−7.0
IVPP TS 211	*Equus sanmeniensis*	Layer 10	−2.3	−6.4
IVPP TS 242	*Megaloceros pachyosteus*	Layer 10	−11.3	−3.0
IVPP TS 243	*Megaloceros pachyosteus*	Layer 10	−11.6	−6.4
IVPP TS 244	*Megaloceros pachyosteus*	Layer 10	−11.4	−3.2
IVPP TS 245	*Megaloceros pachyosteus*	Layer 10	−11.2	−3.5
IVPP TS 246	*Megaloceros pachyosteus*	Layer 10	−11.4	−1.0
IVPP TS 247	*Megaloceros pachyosteus*	Layer 10	−11.9	−2.2
IVPP TS 248	*Megaloceros pachyosteus*	Layer 10	−10.7	−7.1
IVPP TS 249	*Megaloceros pachyosteus*	Layer 10	−11.1	−7.0
IVPP TS 201	*Equus sanmeniensis*	Layer 11	9.5	−5.7
IVPP TS 204	*Equus sanmeniensis*	Layer 11	−6.2	−4.6
IVPP TS 208	*Equus sanmeniensis*	Layer 11	−4.5	−6.3
IVPP TS 213	*Equus sanmeniensis*	Layer 11	−7.0	−6.0
IVPP TS 214	*Equus sanmeniensis*	Layer 11	−3.5	−6.3
IVPP TS 215	*Equus sanmeniensis*	Layer 11	−3.0	−6.6
IVPP TS 216	*Equus sanmeniensis*	Layer 11	−11.2	−5.3
IVPP TS 220	*Dicerorhinus choukoutienensis*	Layer 11	−10.2	−6.9
IVPP TS 221	*Dicerorhinus choukoutienensis*	Layer 11	−11.4	−7.9
IVPP TS 222	*Dicerorhinus choukoutienensis*	Layer 11	−12.4	−6.6
IVPP TS 223	*Dicerorhinus choukoutienensis*	Layer 11	−12.8	−8.0
IVPP TS 230	*Megaloceros pachyosteus*	Layer 11	−11.0	−7.6
IVPP TS 231	*Megaloceros pachyosteus*	Layer 11	−12.3	−6.4
IVPP TS 232	*Megaloceros pachyosteus*	Layer 11	−6.3	−5.6
IVPP TS 233	*Megaloceros pachyosteus*	Layer 11	−13.0	−7.8
IVPP TS 259	*Cervus nippon*	Layer 13	−5.7	−3.7
IVPP TS 260	*Cervus nippon*	Layer 13	−7.9	−1.0
IVPP TS 261	*Cervus nippon*	Layer 13	−6.0	−3.9
IVPP TS 262	*Cervus nippon*	Layer 13	−7.2	−0.7
IVPP TS 263	*Cervus nippon*	Layer 13	−6.2	−3.9

Combined data from all deer species analyzed (*C. nippon, C. elaphus, M. pachyosteus*) show a highly significant decrease in $\delta^{13}C$ over time (Table 3, citing 1). More specifically, $\delta^{13}C$ of *Cervus* teeth decreases significantly (Table 3, citing 2) from an average of -6.6 ± 0.8‰ in Layer 13 to -12.5 ± 0.7‰ in Layer 3. This change indicates that the deer were eating less C_4 vegetation over time, shifting from a mixed C_3/C_4 diet in Layer 13 to a pure C_3 diet by Layer 3.

E. sanmeniensis shows no significant change in $\delta^{13}C$ with time, maintaining a mixed C_3/C_4 diet throughout its sampled interval. *D. choukoutienensis* and *S. lydekkeri* are each analyzed within a single layer, preventing the analysis of trends over time. Both consumed C_3 plants.

The species analyzed indicate the presence of both C_3 and C_4 plants in the area from the time of deposition of Layer 13 to Layer 4. Since the deposition of Layer 4, there is no clear signal of C_4 plants in the area. Analysis of teeth from *E. sanmeniensis* in Layer 3 might indicate a continued presence of C_4 plants. The very strong shift to C_3 plants in the diet of the deer, however, suggests a changing ecosystem. The observed $\delta^{13}C$ values are therefore interpreted as an overall decrease in the abundance of C_4 vegetation during the Middle Pleistocene. It should be noted that the interpreted shift could possibly be due to behavioral changes in the deer sampled.

The $\delta^{13}C$ values of *Cervus* become slightly less negative above Layer 3 but are still within the range of a pure C_3 diet. There may have been many more variations in plant abundances since the deposition of Layer 3, as there is much time not sampled.

$\delta^{18}O$ *of fossil tooth enamel*

Within *C. nippon*, there is a highly significant decrease in $\delta^{18}O$ from Layer 13 to Layer 3 (Table 3, citing 3).

Table 3
Statistics cited

Citation	Tests performed	Values
1	Compared mean δ^{13}C of all deer in and below Layer 11 to mean of all deer above Layer 11	
	Parametric: *t* test (2-tailed) with equal variances not assumed	$P < 0.01$
	Non-parametric: Mann–Whitney	$P < 0.01$
2	Compared mean δ^{13}C of *Cervus* in Layer 13 to mean of *Cervus* in Layer 3	
	Parametric: *t* test (2-tailed) with equal variances not assumed	$P < 0.001$
	Non-parametric: Mann–Whitney	$P < 0.01$
3	Compared mean δ^{18}C of *Cervus* in Layer 13 to mean of *Cervus* in Layer 3	
	Parametric: *t* test (2-tailed) with equal variances not assumed	$P < 0.001$
	Non-parametric: Mann–Whitney	$P < 0.001$
4	Correlated *E. Sanmeniensis* δ^{18}O values with time	
	Parametric: Pearson	$P < 0.001$
	Non-parametric: Spearman	$P < 0.001$
5	Compared mean δ^{18}O of *Cervus* in Layer 3 to mean of *Cervus* above Layer 3	
	Parametric: *t* test (2-tailed) with equal variances not assumed	$P < 0.002$
	Non-parametric: Mann–Whitney	$P < 0.02$
6	Correlated δ^{13}C and δ^{18}O values over time for all deer	
	Parametric: Pearson	$P < 0.001$
	Non-parametric: Spearman	$P < 0.001$

E. sanmeniensis, represented in Layers 11,10, 8/9, 6, and 4, from approximately 680,000 to 500,000 yr ago, also displays a significant decreasing trend with time (Table 3, citing 4). The decrease in δ^{18}O from the time of deposition of Layer 13 to Layer 3 indicates a shift to lower temperatures and/or more winter precipitation-conditions indicative of a strong winter monsoon. Although *M. pachyosteus* does not show this trend, its values are not representative of as wide a time range as *E. sanmeniensis* or *C. nippon*. δ^{18}O values of *M. pachyosteus* from Layer 3 would be helpful in determining if *M. pachyosteus* does indeed not display the pattern, or if we simply have not sampled over a great enough time. If the former is the case, it could be due to behavioral differences in *M. pachyosteus*, for example, migrating during the cold. Because *M. pachyosteus* was much larger than the other deer sampled, behavioral differences would not be surprising. No interpretation of trends from *D. choukoutienensis* or *S. lydekkeri* is possible, as only one layer is sampled in this study.

Following the strong decrease in δ^{18}O, there is a small, but significant (Table 3, citing 5), increase in δ^{18}O of *Cervus* between the time of deposition of Layer 3 and Locality 22. There is no significant change from the time of deposition of Locality 22 to Upper Cave. It is difficult to interpret this shift, as there is much missing time between data points.

The significant correlation of δ^{13}C and δ^{18}O values for all deer sampled (Table 3, citing 6) suggests that the change in the diet of the deer was related to a changing climate at Zhoukoudian. The decrease in the δ^{18}O values over time suggests a strengthening of the winter monsoon and subsequent shortening of the growing season, which is consistent with a decrease in local C_4 biomass as indicated by the decrease in δ^{13}C values from Layer 13 to Layer 3. A denser sample set and serial sampling of individual teeth from different layers for isotopic analysis could help provide a better understanding of the nature of the climate change.

Comparison with other records

Sedimentological studies suggest that the cave deposits of Locality 1 correspond to a global climatic pattern (Liu, 1985; Teilhard de Chardin, 1941) and can be matched to marine oxygen isotope stages of the deep-sea sediments. Figure 2 shows two possible correlations with the astronomically tuned marine oxygen isotope stages, one proposed by Zhou et al. (2000) and the other by Xu et al. (1997). Our tooth enamel isotope data seem to support the correlation by Zhou et al. (2000), with Layer 4 deposited during an interglacial (MOI 13) that is mild, short-lived, and followed by a particularly strong glacial period. The δ^{18}O values of tooth enamel from the same species at Zhoukoudian become more negative by Layer 4, suggesting a strengthening winter monsoon which would be inconsistent with the comparatively warm interglacial (MOI 9) as suggested by Xu et al. (1997).

Many paleoclimatic reconstructions have been attempted at this site, one of the most extensive being the study by Zhou et al. (2000). In order to make the correlation mentioned above, they create a synthetic climate index based on previously published sporopollen data (from Kong et al., 1985) and sediment data (from Xie et al., 1985) from Zhoukoudian. According to Zhou et al. (2000), the curve is closely related to biological and lithological stratigraphy (for a review of biological and lithological stratigraphy at Zhoukoudian, see Wu and Poirier (1995) and Goldberg et al. (2001), respectively). The synthetic climate index created shows glacial and interglacial peaks, which correlate to those in the deep-sea cores and Baoji loess section.

Our data do not show the saw-toothed glacial/interglacial peaks present in the synthetic climate index of Zhou et al. (2000). Instead, the oxygen values become more negative over time, indicating an intensification of the East Asian winter monsoon over several glacial/interglacial cycles. Based on magnetic susceptibility of the stacked Chinese loess, Heslop et al. (2002) also report a strengthening of the winter monsoon beginning after 680,000 yr ago. They interpret this shift as a component of the mid-Pleistocene transition (MPT)—the time at which the global ice volume and the East Asian monsoon shifted from a 41,000- to a 100,000-yr periodicity. In their assessment, the winter monsoon reached its peak intensity around 520,000 yr ago and, following that, has displayed 100,000-yr periodicity, strengthening during glacial periods, weakening in the interglacial periods.

Changes in the winter monsoon intensity with resulting changes in the temperature and precipitation would have affected the plant community. A decline in the relative abundance of C_4 plants on the Chinese Loess Plateau during glacial periods has been documented (Vidic and Montañez, 2004; Zhang et al., 2003). Our data reported here also suggest a decrease in C_4 plant abundance as the winter monsoon intensified.

Finally, it is notable that human occupation of Zhoukoudian represents the last known *H. erectus* fossils from mainland China. Evidence for the presence of this species at Locality 1 is found from Layer 10 through Layer 1, suggesting *H. erectus* weathered the climate deterioration during the mid-Pleistocene. However, the strengthening of the winter monsoon around 520,000 yr ago appears in the loess to be gradual when compared to the subsequent intensification at 410,000 yr ago. Is there a rapid drop in δ^{18}O values in tooth enamel from Layers 1 and 2? Could this have affected the *H. erectus* populations at Zhoukoudian? With analysis of more teeth we can create a stronger data set to better explore the dynamics of the East Asian winter monsoon during the Middle Pleistocene.

Complicating factors for this analysis include the possibility of animal migration and the influence of weaning on the isotopic values of the teeth sampled. If animals were migrating for long distances, their tooth enamel may not provide an accurate record of isotopic values representative of Zhoukoudian. Also, there is some evidence that teeth formed while the animal is nursing can have slightly different isotopic signals than teeth formed after weaning (Hoppe and Amundson, 2001). For this reason, the third molar is the preferred tooth to sample. Because of the destructive nature of sampling and the importance of the site, many of the teeth sampled were already broken. For some teeth, it was not clear if they were first, second, or third molars. Tooth wear analysis was used to ensure that the same individual was not sampled more than once.

Conclusion

Seventy samples of fossil herbivore teeth, found within the Pleistocene deposits at the Peking Man Site, were analyzed for carbon and oxygen isotopic composition. The δ^{13}C values range from $-2.3‰$ to $-13.0‰$, representing ~25% to 100% C_3 plants in the diets of these animals. Such diets suggest a local habitat of mixed C_3/C_4 vegetation at the onset of the Middle Pleistocene. By approximately 470,000 yr ago, C_3 plants dominated the diets of herbivores measured, suggesting that the abundance of C_4 vegetation had decreased in the area. For all deer analyzed in this study, the values of δ^{13}C and δ^{18}O decrease substantially from about 720,000 to 470,000 yr ago, suggesting a strengthening of the winter monsoon in the Middle Pleistocene and a subsequent shortening of the growing season for plants.

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References

Black, D., Teilhard de Chardin, P., Young, C.C., Pei, W.C., 1933. Fossil man in China: the Choukoutien cave deposits with a synopsis of our present knowledge of the Late Cenozoic in China. Geological Society of China, Beijing.

Boaz, N.T., Ciochon, R., 2004. Dragon Bone Hill. Oxford Univ. Press, New York.

Cerling, T.E., Harris, J.M., 1999. Carbon isotope fractionation between bioapatite in ungulate mammals and implications for ecological and paleoecological studies. Oecologia 120, 347–363.

Chen, T., Hedges, R.E.M., Yuan, Z., 1992. The second batch of accelerator radiocarbon dates for the Upper Cave site of Zhoukoudian. Acta Anthropologica Sinica 11, 112–116.

Chia, L.P., Chao, T.K., Li, Y.S., 1959. Some new fossil localities in the Choukoutien region. Vertebrata PalAsiatica 1, 47–51 (in Chinese).

Dansgaard, W., 1964. Stable isotopes in precipitation. Tellus 16, 436–438.

Fogel, M.L., Cifuentes, L.A., 1993. Isotope fractionation during primary production. In: Engel, M.H., Macko, S.A. (Eds.), Organic Geochemistry. Plenum, New York, pp. 73–98.

Goldberg, P., Weiner, S., Bar-Yosef, O., Xu, Q., Liu, J., 2001. Site formation processes at Zhoukoudian. Journal of Human Evolution 41, 483–530.

Grun, R., Huang, P.H., Wu, X., Stringer, C., Thorne, A.G., McCulloch, M., 1997. ESR analysis of teeth from the paleoanthropological site of Zhoukoudian, China. Journal of Human Evolution 32, 83–91.

Guo, S.L., Liu, S.S., Sun, S.F., Zhang, F., Shou, S.H., Hao, X.H., Mang, W., Shang, F., Lui, J.F., 1991. Fission track dating of the 4th layer of the Peking Man Site. Acta Anthropologica Sinica 10, 73–77.

Heslop, D., Dekkers, M.J., Langereis, C.G., 2002. Timing and structure of the mid-Pleistocene transition: records from the loess deposits of northern China. Paleogeography, Paleoclimatology, Paleoecology 185, 133–143.

Hoppe, K., A., Amundson, R., 2001. Interpreting the significance of stable isotopic variations within mammalian teeth: evaluating the influence of biological vs. environmental factors. Geological Society of America, A113–A114.

Huang, P.H., Jin, S.Z., Liang, R.Y., Lu, Z.J., Zheng, L.Z., Yuan, Z.X., Chai, B.X., 1991. Study of ESR dating for the burial age of the first skull of Peking Man and chronological scale of the site. Chinese Science Bulletin 36, 1457–1461 (in Chinese).

Huang, P.H., Jin, S.Z., Peng, Z.C., Liang, R.Y., Lu, Z.J., Wang, Z.R., Chen, J.B., Yuan, Z.X., 1993. ESR dating of tooth enamel: a comparison with U-Series, FT and TL dating at the Peking Man Site. Applied Radiation and Isotopes 44, 239–242.

Jia, L., 1980. Early Man in China. Foreign Languages Press, Beijing.

Johnson, K., Ingram, B., 2004. Spatial and temporal variability in the stable isotope systematics of modern precipitation in China: implications for paleoclimate reconstructions. Earth and Planetary Science Letters 220, 365–377.

Koch, P.L., Fogel, M.L., Tuross, N., 1994. Tracing the diets of fossil animals using stable isotopes. In: Lajtha, K., Michener, B. (Eds.), Stable Isotopes in Ecology and Environmental Science. Blackwell, Oxford, pp. 63–92.

Kohn, M.J., Schoeninger, M.J., Valley, J.W., 1998. Variability in oxygen isotope compositions of herbivore teeth: reflections of seasonality or developmental physiology? Chemical Geology 152, 97–112.

Kong, Z., Du, L., Wu, Y., Yu, Q., Yi, M., Ren, Z., 1985. Vegetational and climatic changes since Paleogene at Zhoukoudian and its adjacent regions. In: Wu, R.K., Ren, M.E., Zhu, X.M., Yang, Z.G., Hu, C.K., Kong, Z.C., Xie, Y.Y., Zhao, S.S. (Eds.), Multi-Disciplinary Study of the Peking Man Site at Zhoukoudian. Science Press, Beijing, pp. 119–154 (in Chinese).

Kreuger, H.W., Sullivan, C.H., 1984. Models for carbon isotope fractionation between diet and bone. In: Turland, J.F., Johnson, P.E. (Eds.), Stable Isotopes in Nutrition, ACS Symposium Series. American Chemical Society, pp. 205–222.

Lee-Thorp, J., van der Merwe, N.J., 1987. Carbon isotope analysis of fossil bone apatite. South African Journal of Science 83, 712–715.

Lee-Thorp, J., van der Merwe, N.J., 1991. Aspects of the chemistry of modern and fossil biological apatites. Journal of Archaeological Science 18, 343–354.

Lee-Thorp, J., Sealy, J.C., van der Merwe, N.J., 1989. Stable carbon isotope ratio differences between bone collagen and bone apatite and their relationship to diet. Journal of Archaeological Science 16, 585–599.

Lin, S., 1994. "Zhoukoudian Peking Man Site." Institute of Vertebrate Paleontology and Paleoanthropology. Chinese Academy of Sciences, Beijing.

Liu, Z.C., 1985. Sequence of sediments at Locality 1 in Zhoukoudian and correlation with loess stratigraphy in Northern China and with the chronology of the deep-sea cores. Quaternary Research 23, 139–153.

Liu, T.S., Ding, Z., 1998. Chinese loess and the paleomonsoon. Annual Review of Earth Planetary Science 26, 111–145.

Longinelli, A., 1984. Oxygen isotopes in mammal bone phosphate: a new tool for paleohydrological and paleoclimatological research? Geochimica et Cosmochimica Acta 48, 385–390.

Luz, B., Kolodony, Y., 1985. Oxygen isotope variation in phosphate of biogenic apatites, IV. Mammal teeth and bones. Earth and Planetary Science Letters 75, 29–36.

Luz, B., Kolodny, Y., 1989. Oxygen isotope variation in bone phosphate. Applied Geochemistry 4, 317–323.

O'Leary, M.H., 1988. Carbon isotopes in photosynthesis. Bioscience 38, 328–336.

Pei, J., 1985. Thermoluminescence of the Peking Man Site and other caves. In: Wu, R.K., Ren, M.E., Zhu, X.M., Yang, Z.G., Hu, C.K., Kong, Z.C., Xie, Y.Y., Zhao, S.S. (Eds.), Multi-Disciplinary Study of the Peking Man Site at Zhoukoudian. Science Press, Beijing, pp. 258–260 (in Chinese).

Porter, S.C., 2001. Chinese loess record of monsoon climate during the last glacial–interglacial cycle. Earth-Science Reviews 54, 115–128.

Qian, F., Zhang, J.X., Yin, W.D., 1985. Magnetic stratigraphy from the sediment of the west wall and test pit of Locality 1 at Zhoukoudian. In: Wu, R.K., Ren, M.E., Zhu, X.M., Yang, Z.G., Hu, C.K., Kong, Z.C., Xie, Y.Y., Zhao, S.S. (Eds.), Multi-Disciplinary Study of the Peking Man Site at Zhoukoudian. Science Press, Beijing, pp. 251–254 (in Chinese).

Rutter, N., Ding, Z.L., Lui, T.S., 1991. Comparison of isotope stages 1–61 with the Baoji-type pedostratigraphic section of north-central China. Canadian Journal of Earth Sciences 28, 985–990.

Shen, G., Jin, L., 1991. Restudy of the upper age limit of Peking Man Site. Acta Anthropopogica Sinica 10, 273–277 (in Chinese).

Shen, G., Gu, D., Gahleb, B., Yuan, Z., Chai, B., 1996. Preliminary results on U-series dating of Peking Man site with high precision TIMS. Acta Anthropopogica Sinica 15, 210–217.

Smith, B., Epstein, S., 1971. Two categories of $^{12}C/^{13}C$ ratios for higher plants. Plant Physiology 47, 380–384.

Spell, T.L., McDougall, I., 1992. Revision to the age of the Brunhes–Matuyama boundary and the Pleistocene geomagnetic polarity time scale. Geophysical Research Letters 19, 1181–1184.

Teilhard de Chardin, P., 1941. Early Man in China. Institute of Geo-Biology, Peking.

Vidic, N.J., Montañez, I.P., 2004. Climatically driven glacial–interglacial variations in the C_3 and C_4 plant proportions on the Chinese Loess Plateau. Geology 32, 337–340.

Wang, Y., Cerling, T.E., MacFadden, B.J., 1994. Fossil horses and carbon isotopes: new evidence for Cenozoic dietary, habitat, and ecosystem changes in North America. Paleogeography, Paleoclimatology, Paleoecology 107, 269–279.

Williams, D.F., Thunell, R.C., Tappa, E., Tio, D., Raffi, I., 1988. Chronology of the Pleistocene oxygen isotope record: 0–1.88 m.y. B.P.. Paleogeography, Paleoclimatology, Paleoecology 64, 221–240.

Wu, X., Poirier, F., 1995. Human Evolution in China: A Metric Description of the Fossils and a Review of the Sites. Oxford Univ. Press, New York.

Wu, R.K., Ren, M.E., Zhu, X.M., Yang, Z.G., Hu, C.K., Kong, Z.C., Xie, Y.Y., Zhao, S.S., 1985. Multi-disciplinary study of Peking Man at Zhoukoudian. Science Press, Beijing (in Chinese).

Xie, Y., Xing, H., Xu, J., Zhou, B., Huang, Y., Liu, Y., Yang, J., Qu, Z., 1985. The sedimentary environment of the Peking Man Period. In: Wu, R.K., Ren, M.E., Zhu, X.M., Yang, Z.G., Hu, C.K., Kong, Z.C., Xie, Y.Y., Zhao, S.S. (Eds.), Multi-disciplinary Study of Peking Man at Zhoukoudian. Science Press, Beijing, pp. 185–215 (in Chinese).

Xu, Q., Jin, C., Tong, H., Dong, W., Liu, J., Cai, B., 1997. Three glacial cycles during Peking Man's time. In: Tong, Y. (Ed.), Essays in Honor of Professor Chungchien Young on the Hundredth Anniversary of his Birth. China Ocean Press, Beijing, pp. 213–226.

Yuan, S.X., Chen, T.N., 1980. Uranium-series dating of bones at Zhoukoudian. Acta Anthropologica Sinica 10, 189–193.

Zhang, Z.H., Zhao, M., Lu, H., Faiia, A.M., 2003. Lower temperature as the main cause of C_4 plant declines during the glacial periods on the Chinese Loess Plateau. Earth and Planetary Science Letters 214, 467–481.

Zhao, S.S., Xia, M., Zhang, Z.H., Liu, M.L., Wang, S.X., Wu, Q.F., Ma, Z.B., 1985. Uranium-series dating of the Peking Man Site. In: Wu, R.K., Ren, M.E., Zhu, X.M., Yang, Z.G., Hu, C.K., Kong, Z.C., Xie, Y.Y., Zhao, S.S. (Eds.), Multi-disciplinary Study of the Peking Man Site at Zhoukoudian. Science Press, Beijing, pp. 246–250 (in Chinese).

Zhou, C., Lui, Z., Wang, Y., 2000. Climatic cycles investigated by sediment analysis in Peking Man's Cave, Zhoukoudian, China. Journal of Archaeological Science 27, 101–109.

Zhu, Y., Zhou, M., 1994. The Asian monsoon variation and the habituation environment of fossil man in Northern China. Catena 22, 121–131.