

Supplementary information for

Arabian hydroclimate and hominin population dynamics over the last 500,000 years

Liang Zhou¹, Zhaoxia Jiang^{1*}, Sanzhong Li^{2,1}, Juan C. Larrasoana^{3,4}, Andrew P. Roberts⁵, Shixia Yang^{6,7}, Qingsong Liu⁸, Liang Chen^{9,10}, Zhengxin Yin^{9,10}, Wanxin Liu¹, Yulong Guan¹, Yuzhen Zhang¹, Michael D. Petraglia^{7,11,12}

¹Frontiers Science Center for Deep Ocean Multispheres and Earth System, Key Lab of Submarine Geosciences and Prospecting Techniques, Ministry of Education, College of Marine Geosciences, Ocean University of China, Qingdao 266100, China

² Laoshan National Laboratory, Qingdao 266237, China

³Departamento de Ciencias, Universidad Pública de Navarra, 31006 Pamplona, Spain

⁴CN IGME, CSIC, 50059 Zaragoza, Spain

⁵Research School of Earth Sciences, Australian National University, Canberra, ACT 2601, Australia

⁶Institute of Vertebrate Paleontology and Paleoanthropology, Chinese Academy of Sciences, Beijing, 100044, China

⁷ Australian Research Centre for Human Evolution, Griffith University, Brisbane, 4111, Australia

⁸Department of Ocean Science and Engineering, Centre for Marine Magnetism (CM²), Southern University of Science and Technology, 518055 Shenzhen, China

⁹South China Sea Marine Survey and Technology Center, State Oceanic Administration, Guangzhou 510300, China

¹⁰Key Laboratory of Marine Environmental Survey Technology and Application, Ministry of Natural Resources, Guangzhou 510300, China

¹¹School of Social Science, University of Queensland, Brisbane, 4072, Australia

¹²Human Origins Program, National Museum of Natural History, Smithsonian Institution, Washington, D.C., 20560, USA

* Corresponding author: Zhaoxia Jiang (jiangzhaoxia@ouc.edu.cn)

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Supplementary Note 1 for location and geological setting.

Modern western equatorial Indian Ocean (WEIO) climate is dictated by seasonal wind direction changes associated with the Indian monsoon system¹. At the end of May and early June, a meridional shift of the subsolar point drives northward propagation of the Intertropical Convergence Zone (ITCZ, **Fig. S1**), which brings moist air from the Southern to the Northern Hemisphere and causes strong convection over the northern Indian Ocean^{2,3}. As the subpolar front and ITCZ propagate further north in mid-June, Northern Hemisphere land warms by absorption of increased solar radiation at a greater rate than the Indian Ocean, forming a strong ocean-land thermal gradient. The gradient drives strong low-level southwesterly ISM winds, which transport water vapour from the tropical Indian Ocean and the Arabian Sea toward continental Afro-Asia^{4,5} (**Fig. S1**). The WEIO is located close to the world's largest dust sources, the Saharo-Arabian desert belt, which provides ~70% of global dust emissions^{6,7}. The main drivers of dust entrainment are southwest summer monsoon winds and the dust-loaded Shamal that blows from the northwest⁸ (**Fig. S1**). Shamal winds develop along the pressure gradient that forms between the low-pressure monsoon system over India and the high-pressure system over the eastern Mediterranean and are enhanced by orographic effects along the Persian Gulf⁹. These winds can override moist near-surface southwest monsoon winds and transport large dust quantities toward the Arabian Sea, where it is scavenged and wet-deposited by summer monsoonal precipitation^{10,11}. This mid-tropospheric transport also results in dust delivery to the more distal equatorial Indian Ocean¹¹⁻¹³.

Supplementary Note 2 for research materials and age model.

Sediment core CJ02-08 (1.64°S, 65.64°E, 3,710 m water depth) was recovered from the western slope of the Central Indian Ridge in the western equatorial Indian Ocean (**Fig. S1**) during the cruises with “Haice 3301” organized by South China Sea Marine Survey and Technology Center (Guangzhou, China) from December 2020 to January 2021. This core includes the studied 542 cm-thick sedimentary sequence, which consists of hemipelagic bioturbated nannofossil oozes and nannofossil clays. Samples for magnetic

measurements were obtained sequentially from a split core profile using $2\text{ cm} \times 2\text{ cm} \times 2\text{ cm}$ plastic cubes. All artifacts referred to in this study are curated in Ocean University of China, Qingdao, China. They are available for further research.

The age model for core CJ02-08 (**Fig. S2**) for the past 500 kyr is based on that of Chen et al.¹, and provides a temporally continuous record. To improve the initial age model for core CJ02-08 presented in ref.¹⁴, a refined chronological framework of core CJ02-08 was established by assigning RPI minima to geomagnetic excursions¹⁵ and to abrupt Ba/Ti increases to glacial terminations¹⁶, respectively (**Fig. S2**).

The new Ba/Ti stratigraphic method relies on the occurrence and abundance patterns of biogenic barium (Ba), which is widespread and well preserved in equatorial Indian Ocean sediments. Biogenic Ba accumulation in sediments is a reliable proxy for marine export production¹⁷. Schmitz¹⁸ hypothesized that a Ba accumulation rate increase in sediments underlying the equatorial Indian Ocean divergence is linked directly to overlying high productivity. Marine sedimentary Ba has both a biogenic (Ba_{bio}) and terrigenous ($\text{Ba}_{\text{detrital}}$) component, so estimates of past export productivity using Ba must distinguish between these sources. Use of bulk Ba/Ti, or Ba/Al, enables evaluation of relative export productivity changes (i.e. via normalization to a predominantly terrigenous element)¹⁹. Relative abundance fluctuations of increased biogenic Ba with export production in interglacial intervals have been correlated to oxygen isotope records in late Quaternary equatorial Indian Ocean sediments^{20,21}. Downcore Ba/Ti variability for core CJ02-08 is comparable to Ba/Al and Ba variations in nearby cores, both of which have a well-constrained chronology^{20,21}. We interpret Ba/Ti to represent a biological/terrestrial input ratio at our study site, which varies with glacial/interglacial productivity changes due to global ice volume variations.

The core CJ02-08 chronology was established by correlating a high-resolution Ba/Ti record with the global $\delta^{18}\text{O}$ LR04 stack¹⁶. This correlation is based on transient biogenic Ba accumulation increases at each deglacial transition across the equatorial Indian Ocean^{20,21}. Seven age tie points were obtained by correlating Ba/Ti to the global $\delta^{18}\text{O}$ LR04 stack. Moreover, eight other age control points for core CJ02-08 are obtained from geomagnetic palaeointensity minima in $\text{NRM}_{30\text{ mT}}/\text{SIRM}_{30\text{ mT}}$ to geomagnetic

excursions in the global PISO-1500 RPI stack¹⁵. The 12 initial RPI age control points of Chen et al.¹⁴ include some uncertain tie points. We eliminated 4 of these tie points for our new age model. The eight age tie points used from RPI dating are consistent with the age constraints from Ba/Ti.

Combining 15 age control points from the two methods (**Fig. S2e**), we developed a refined age model for core CJ02-08 via linear interpolation between age tie points. The resulting age-depth model is almost linear and well constrained because of the number of reliable age tie points (**Fig. S2e**). The mean sedimentation rate is 1.0 cm/kyr for core CJ02-08, which is consistent with that of other western equatorial Indian Ocean sediment cores^{20,22}.

Supplementary Note 3 for Sr-Nd isotopic compositions and dust sources.

Strontium and neodymium isotopes for 12 samples were analyzed using a High-Resolution Multi-Collector Inductively Coupled Plasma Mass Spectrometer (HR-MC-ICPMS). Samples were treated with 20% acetic acid and 15% hydrogen peroxide to remove carbonate and organic matter, respectively. The residue was then transferred to a Teflon bottle and dissolved completely in a HF-HNO₃-HClO₄ mixture. Sr and Nd were extracted from the solution using standard ion-exchange procedures. Sr and Nd isotopic ratios were normalized to the NBS987 standard and the JNdi-1 standard, respectively. During analysis, Sr isotopic compositions of the NBS987 standard yielded a 0.710320 ± 0.000006 value (recommended value 0.71008-0.71060), while Nd isotopic compositions of the JNdi-1 standard yielded a 0.512116 ± 0.000005 value (recommended value 0.512108-0.512122). Nd results are expressed as $\epsilon_{\text{Nd}} = [({}^{143}\text{Nd}/{}^{144}\text{Nd}_{\text{(measured)}}/{}^{143}\text{Nd}/{}^{144}\text{Nd}_{\text{(CHUR)}})-1] \times 10^4$, using a CHUR value of 0.512638²³.

⁸⁷Sr/⁸⁶Sr ratios and ϵ_{Nd} values, measured on the detrital fraction of six samples from core CJ02-08, are summarized in **Table S1**. ⁸⁷Sr/⁸⁶Sr ratios range from 0.710274 to 0.711640 with a mean value of 0.711019; ϵ_{Nd} values vary from -8.66693 to -7.75986 with a mean value of -8.13197. Generally, ⁸⁷Sr/⁸⁶Sr and ϵ_{Nd} remain relatively stable (**Fig.**

S3), which indicates that sediment provenance might have been uniform throughout the studied period. Major dust sources have been delimited in previous studies based on mineralogical and geochemical records along the surrounding desert belt (Fig. S3, e.g., North Africa, Southwest Asia, and India)²⁴⁻³⁶. Dust is generated widely across three geochemically distinct southwest Asian source areas. Our Nd and Sr isotope data for the cleaned bulk terrigenous fraction from core CJ02-08 align closely with Arabian compositions, which have relatively low radiogenic Sr and less negative radiogenic Nd, so they can be explained by dust from this source alone (Fig. S3). Sr and Nd isotopic signatures of the lithogenic fraction of CJ02-08 sediments are remarkably consistent, which imply that terrigenous source regions remained unchanged over the past 500 kyr.

Supplementary Note 4 for dust proxies.

High-coercivity magnetic minerals (i.e., hematite and goethite) are common atmospheric dust constituents emitted from the Saharo-Arabian Desert belt^{37,38}. They have long atmospheric residence times, can be transported over long distances, and their concentrations are reliable proxies for atmospheric dust contents³⁹⁻⁴¹. Magnetic properties point, consistently with what was found for WEIO sediments from nearby core CJ02-01, to the presence of mainly biogenic magnetite and high-coercivity aeolian goethite/hematite¹³.

High-temperature magnetic susceptibility (χ -T curves) of representative samples were measured from room temperature to 700°C to investigate magnetic mineralogy using a MFK1-FA Kappabridge susceptibility meter equipped with an AGICO CS4 oven at a 967 Hz frequency. χ -T heating curves contain a drop at 580°C that attests to the dominance of magnetite, which masks a subdued signal associated with hematite that is only evident as a residual susceptibility above 580°C (Fig. S4e).

To further investigate magnetic mineral concentration and grain size changes, we measured first-order reversal curve (FORC) diagrams and Isothermal remanent magnetization (IRM) acquisition curves using a Lake Shore 8600 vibrating sample magnetometer. FORC diagrams were processed using the FORCinel software v3.06 of Harrison & Feinberg⁴². IRM acquisition curves were decomposed following the

approach of Kruiver et al.⁴³. FORC diagrams can be used to identify the magnetic domain state of fine magnetic particles and to discriminate among different components in magnetic mineral assemblages^{44,45}. FORC diagrams for core CJ02-08 samples have elongated contours along the horizontal axis with small vertical spread (**Fig. S4a-d**), which represents single domain (SD) behaviour with weak magnetostatic interactions⁴⁴, associated typically with noninteracting biogenic SD particles⁴⁶. Diverging outer contours due to vortex state particles are also observed (**Fig. S4a-d**), which are due to detrital particles in the magnetic vortex state^{47,48}. On the B_c axis, contours can extend to 100 mT (**Fig. S4a-d**), which demonstrates the presence of high-coercivity magnetic particles^{49,50}.

IRM acquisition curves can be decomposed into coercivity fractions^{43,51} to assist discrimination among magnetic minerals. Four components are recognized for all samples (**Fig. S4a-d**; **Table S2**). Component 1 ($B_{1/2}$ = 11.8-16.7 mT; dispersion parameter (DP) = 0.3-0.4) likely represents ultrafine magnetite (maghemite) just above the SP/SD threshold size (~20-25 nm)⁵³. Component 2 is the dominant magnetic mineral with a relatively constant mean coercivity ($B_{1/2}$ = 32.1-35.5 mT) and narrow DP (0.22-0.24). Combined with corresponding FORC diagrams (**Fig. S4**), component 2 likely represents biogenic SD magnetite. Component 3 has higher coercivity ($B_{1/2}$ = 78.0-97.4 mT) and has been interpreted as due to aeolian magnetite with surface oxidation^{53,54}. Component 4 ($B_{1/2}$ = 261.7-375.6 mT; DP = 0.24-0.35) contributes <10% to the bulk SIRM and is attributed to high coercivity hematite/goethite. Although goethite might be more abundant than hematite, it hardly contributes to the measured magnetic properties because 10% of its IRM at most is likely to be acquired at applied fields up to 3 T^{55,56}. Hence, only hematite is interpreted to constitute the high coercivity fraction of core CJ02-08 sediments based on magnetic measurements.

$IRM_{1T@AF_{80mT}}$ is often used to evaluate hematite concentrations^{7,40}. However, interpretation of this parameter can be complicated by hematite property changes, such as the coercivity distribution or isomorphous cation substitution⁵⁷. The L-ratio ($L\text{-ratio} = (IRM_{1T} + IRM_{0.3T}) / (IRM_{1T} + IRM_{0.1T})$) enables assessment of $IRM_{1T@AF_{80mT}}$ interpretation in terms of high-coercivity mineral concentrations⁵⁷. In the studied

samples, L-ratio values have a relatively narrow range (**Fig. S5**), which indicates that Al substitution in hematite is relatively stable and that $IRM_{1T}@AF_{80mT}$ is a reliable hematite concentration proxy.

Characteristic peak positions for hematite and goethite in second-derivative DRS curves (535 and 425 nm for hematite and goethite, respectively) can be used to identify these minerals more clearly; they demonstrate that both hematite and goethite occur in the sediments (**Fig. S4f**). Goethite variations can be only assessed from DRS measurements using its corresponding band (425 nm) intensity (I_{Gt}).

Dust proxies can also be derived from X-ray fluorescence (XRF) scans of marine sediment cores. Principal component analysis (PCA) is applied widely to determine the relationship between elements as palaeoclimate indicators in marine sediments^{58,59}. Two principal components explain 93.9% of the total variance (**Fig. S6a**). The first principal component (PC1) explains 81.8% of geochemical data variance, including Al_2O_3 , K_2O , Fe_2O_3 , Na_2O , MgO , and Ti (**Fig. S6a**). Correlation coefficients are 0.91 between MgO and Al_2O_3 , and 0.92 between Ti and MgO , respectively (**Fig. S6b**). CaO has a strong negative correlation with PC1 elements (**Fig. S6c**). PC2 explains 12.2% of the variance (**Fig. S6a**). Al_2O_3 , K_2O , Fe_2O_3 , MgO , and Ti contents correlate negatively with CaO (**Fig. S6c**), which indicates that these elements occur in terrigenous phases (because carbonate production is high in tropical settings which can dilute terrigenous elements). Generally, Si is concentrated in the coarse fraction (e.g., quartz). SiO_2 has lower correlation with other elements, which indicates that it occurs in the terrigenous fine-grained fraction. Ti content also correlates positively with Al_2O_3 , K_2O , and MgO (**Fig. S6c**), which indicates that Ti is also held by the terrigenous fraction.

PCA of major elements indicates that Ti , Mg , and Al are held by terrigenous phases in the distal marine sediments of core CJ02-08. Ca is a measure of biogenic carbonate ($CaCO_3$) abundance in marine sediments^{60,61}. Relative terrigenous and marine constituents can be best deduced from titanium (Ti) and calcium (Ca). The conservative element Ti is restricted to lithogenic sediments and is inert to diagenesis^{60,62}. Ca provides a measure of biogenic carbonate ($CaCO_3$) abundance in marine sediments^{60,61}, and relative terrigenous (Arabian dust) and biogenic variations are deduced from Ti/Ca .

In addition to the general distinction between land-derived and marine sediments following ref.⁶³, we use downcore XRF-scanning to distinguish between aeolian dust and fluvial muds by examining the ratio of magnesium (Mg) over aluminium (Al). Mineralogical and chemical studies of sediment samples from the Arabian Sea indicate that Mg can be used as a potential Arabian Peninsula provenance indicator⁶⁴. The Mg-rich mineral distribution has a persistent pattern with a maximum close to the Arabian Peninsula and decreasing values in more distal parts of the Arabian Sea^{64,65}. This has been interpreted as deposition of dust plumes enriched in Mg-bearing minerals, such as palygorskite, a Mg-rich clay found in sabkhas and Mesozoic rocks of the Arabian Peninsula, or dolomite that forms today in coastal sabkhas around the Persian Gulf^{65,66}. In contrast to Mg, Al has low concentrations in sediments off the Arabian Peninsula⁶⁴, whereas Al clearly indicates aluminosilicates which are mostly present in fluvial sediments⁶³. Accordingly, Prins et al.⁶⁷ proposed that an excess of Mg to Al, i.e., from Mg/Al values of sediment samples, can be used to distinguish an “Arabian aeolian source” from a “Indus fluvial source”.

We compare temporal Mg/Al and Ti/Ca trends with hematite and goethite contents across the studied interval. All four proxies have synchronous variations, with significant positive correlations between Mg/Al, Ti/Ca, and the two aeolian proxies. Although all dust proxies likely respond to aeolian variability, additional processes, sometimes unrelated to atmospheric circulation (e.g. fluvial terrigenous and biogenic alteration), may also influence sediment compositions⁶⁸. To limit such biases, we assess variance in common among all four proxies with palaeoclimatic proxies using simple PCA. PCA results indicate that the first component (PC1: i.e. aeolian dust variability) for all of our proxies ($IRM_{1T@AF80mT}$, I_{Gt} , Mg/Al, and Ti/Ca) explains 60.8% of common variance compared to only 19.0% in PC2, which indicates clearly that all of our proxies are controlled mainly by a common forcing, i.e. Arabian dust dynamics (**Fig. S7**). Visual comparison of the four proxies and PC1 indicates that the timing of variability is similar, which indicates a common link to dust content (**Fig. S8**).

Supplementary Note 5 for spectral analysis of dust records.

The chronology of core CJ02-08 is based essentially on orbital tuning. Therefore, cycles present in target curves (PISO-1500 and LR04 stack)^{15,16} may be introduced into our dust record during age tuning. To address this concern and to verify the authenticity of orbital periodicities in the dust record, we performed spectral analysis directly on the raw CJ02-08 depth-series data (before orbital tuning to the PISO-1500 and LR04 records) using the R package for Astrochron⁶⁹ of the untuned depth-domain CJ02-08 dust record. Results clearly reveal the presence of pervasive ~90-100 cm, ~41-46 cm, and ~20-23 cm cycles (**Fig. S9**). We quantified the spectral power of these periodicities in the depth domain: the pervasive depth cycles retain significant spectral power (exceeding the 90% confidence level of the red noise model) in the untuned data (**Fig. S9**), with no meaningful power reduction relative to the tuned age-series. Considering a sedimentation rate of 1.0 cm/ka, these prominent depth cycles correspond to orbital eccentricity (~100-kyr), obliquity (~41 kyr), and precession (~23 kyr)-related cycles, respectively, as identified in the tuned age-series. This result confirms that orbital (especially eccentricity and precession) cycles in the CJ02-08 dust record are inherent to the original stratigraphic signal and are not an artifact introduced by subsequent tuning to the PISO-1500 and LR04 records.

To further mitigate the risk of false cycle detection⁷⁰, the more rigorous Astrochron-multitaper method (MTM) analysis with 90% confidence level and false positive testing was used to refine spectral peak interpretability. Spectral and wavelet analyses for all four age-series dust proxy records indicate prominent periodicities at 100, 41, 35, and 23 kyr since 500 ka. These cycles exceed the 90% red noise confidence level in the primary spectral analysis, and also pass explicit false positive testing with confidence level above 90% (**Fig. S10**).

Supplementary Note 6 for establishment of a stacked dust index.

Four new dust proxy records from environmental magnetism ($IRM_1 T@AF_{80 mT}$), diffuse reflectance spectroscopy (I_{Gt}), and geochemical elemental ratios (Mg/Al and Ti/Ca) are linked independently to aeolian dust variability via source area and atmospheric changes. We stacked these independent aeolian dust records to derive a

more robust, combined record of dust variability to allow comparison with regional palaeoclimatic, palaeoenvironmental, and archaeological records. The dust index is constructed by simple stacking, which is accomplished by normalizing each record to unit variance, and averaging^{68,71}. The Min-Max normalization method is adopted for data standardization, which is a linear transformation of the original data and scales normalized results to the 0 to 1 range. The corresponding transformation function is expressed as follows: $x^* = (x - \min) / (\max - \min)$, where max denotes the maximum data value and min represents the minimum value. The dust index curve for core CJ02-08 is generated by standardized data processing and calculation steps as follows: first, individual proxy data (IRM_{1T} , AF_{80mT} , I_{Gt} , Mg/Al , and Ti/Ca) are normalized separately using min-max normalization, which transforms linearly the original data to the 0 to 1 range and effectively eliminates dimensional differences among proxies. Subsequently, the arithmetic mean of these normalized proxy datasets is calculated to obtain a continuous dust index series for core CJ02-08. Finally, the dust index curve is plotted by mapping calculated dust index values to their corresponding sedimentary age sequence; error values at each age interval are determined as the standard deviation of the above normalized proxies.

Supplementary Note 7 for estimating hunter-gatherer group population sizes.

Population size is one of the most relevant demographic factors that influence hominin evolution^{72,73}. Aside from effective population size estimates based on present-day hominin genetic data⁷⁴, whose translation into population size is not straightforward⁷³, we are aware of only a handful of studies where Pleistocene hominin population sizes have been estimated^{73,75,76}. We use demographic and ecological data from Kelly⁷⁷ and Binford⁷⁸ for subtropical hunter-gatherer groups (collected and updated by Marlowe⁵¹) to estimate population densities of past hominin groups in different subtropical belts of Arabia during dry (present day-like) and wet (GAP) conditions. Hunter-gatherer groups are considered useful analogues for, at least, Late Pleistocene hominins⁷⁹; recent evidence assembled by Goren-Inbar and Belfer-Cohen⁸⁰ suggests that they might also be valid for providing demographic information about early members of our genus.

Data considered include mean annual rainfall, population density, primary biomass, and local group area; by calculating the total primary biomass (a reliable habitat quality measure, see Marlowe⁷⁹) of the corresponding local area, we can calculate the amount of biomass available for each individual within the home range. Finally, we estimate the surface of different savannah vegetation belts suitable for hominin occupation during the Eemian GAP to estimate the total hominin population in Arabia during representative GAPs, considering population density variations as a function of mean annual precipitation determined from hunter-gatherer analogues. Following Larrasoña et al.⁸¹ and Cerling et al.⁸², the three subtropical belts considered here as prime hominin habitats are grasslands (100-250 mm/year rainfall; <10% woody cover), wooded grasslands (250-1200 mm/year rainfall; 10-40% woody cover), and woodlands (1200-1800 mm/year rainfall; 40-80% woody cover).

There is a positive power-law correlation between rainfall and population density and a similar correlation between primary biomass and rainfall, which suggests that hunter-gatherer population density increases as a function of primary biomass available for exploitation, which is, in turn, favoured by rainfall⁸³. The home range (local group area) of hunter-gatherer groups increases with decreasing rainfall, although its correlation is less robust. Nevertheless, this indicates that hunter-gatherers from drier regions need larger home ranges to access necessary resources. It appears that most subtropical hunter-gatherer groups rely on a similar biomass amount (50-70 tonnes per individual within their range) for survival, regardless of rainfall. These findings suggest that hunter-gatherers thrived in equilibrium with the carrying capacity of their environment across the full range of subtropical climate conditions.

Based on the above inferences, Larrasoña et al.⁸³ estimated population densities for the four prime hominin savannah belts that can be divided according to monsoon rainfall and its impact on structural vegetation classification. Density values are 0.01, 0.056, 0.205, and 0.463 individuals/km² for grassland, dry wooded grassland, wet wooded grassland, and woodland belts, respectively, with an average of 0.15 individuals/km². These values fall within the low range typical for chimpanzees (0.2-1 individuals/km²)⁸⁴ and are lower than expected for a mammal with the body size of an

average hominin (~ 5 individuals/km²)⁸⁵, and are, therefore, considered conservative estimates. Given the lack of hominin adaptation to desert and tropical forest ecosystems before the latest Pleistocene-Holocene, we use a hominin population density of 0 individuals/km² in desert and tropical forest belts. A conservative population density of 0.1 individuals/km² is considered for the Mediterranean area. The total population for each vegetation belt was calculated after subtracting the surface occupied by large lakes.

Supplementary Information figure and table

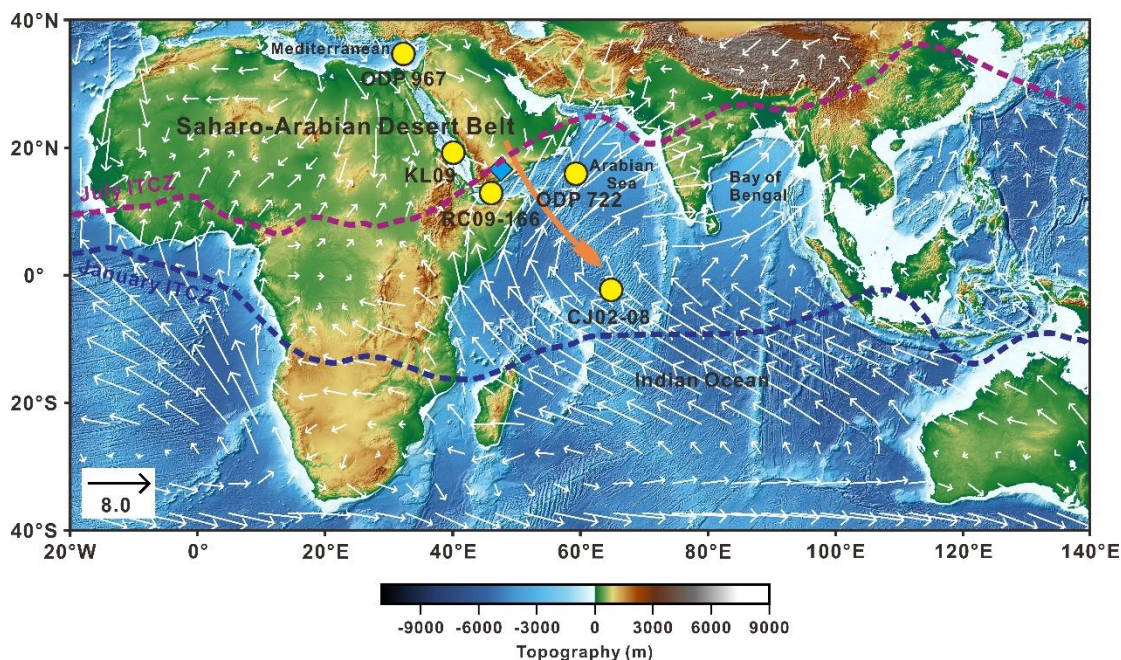


Fig. S1. Location of core CJ02-08 (yellow circle) and geographical setting of the Indian Ocean. Background topography (ETOPO1) from National Oceanic and Atmospheric Administration (NOAA) National Geophysical Data Center (<https://www.ngdc.noaa.gov/ngdc.html>) and Figures were created using Generic Mapping Tools (<https://www.generic-mapping-tools.org/>). ETOPO1 data are produced by NOAA and are publicly available accordance with the NOAA data-use policy (https://www.ngdc.noaa.gov/mgg/global/dem_faq.html). Data for the present-day summer season (June-July-August, JJA) wind vector (white arrows, units: m/s) at 850 hPa between 1979 and 2020 are obtained from NOAA Physical Sciences Laboratory (PSL), available at <https://psl.noaa.gov/cgi-bin/data/composites/printpage.pl>. NOAA PSL data are publicly available and may be used in accordance with the NOAA PSL data-use and citation policy (<https://psl.noaa.gov/data/obs/download/Citation.html>). Modern Intertropical Convergence Zone (ITCZ) locations in July and January are indicated by dashed lines. Marine and terrestrial records include core CJ02-08 (this study), Oceanic Drilling Program (ODP) site 722⁸⁶, International Ocean Discovery Program (IODP) site U1467⁹, ODP site 967⁸⁷, KL 09⁷, RC09-166⁸⁸ (yellow circles), and Mukalla Cave⁸⁹ (blue diamonds).

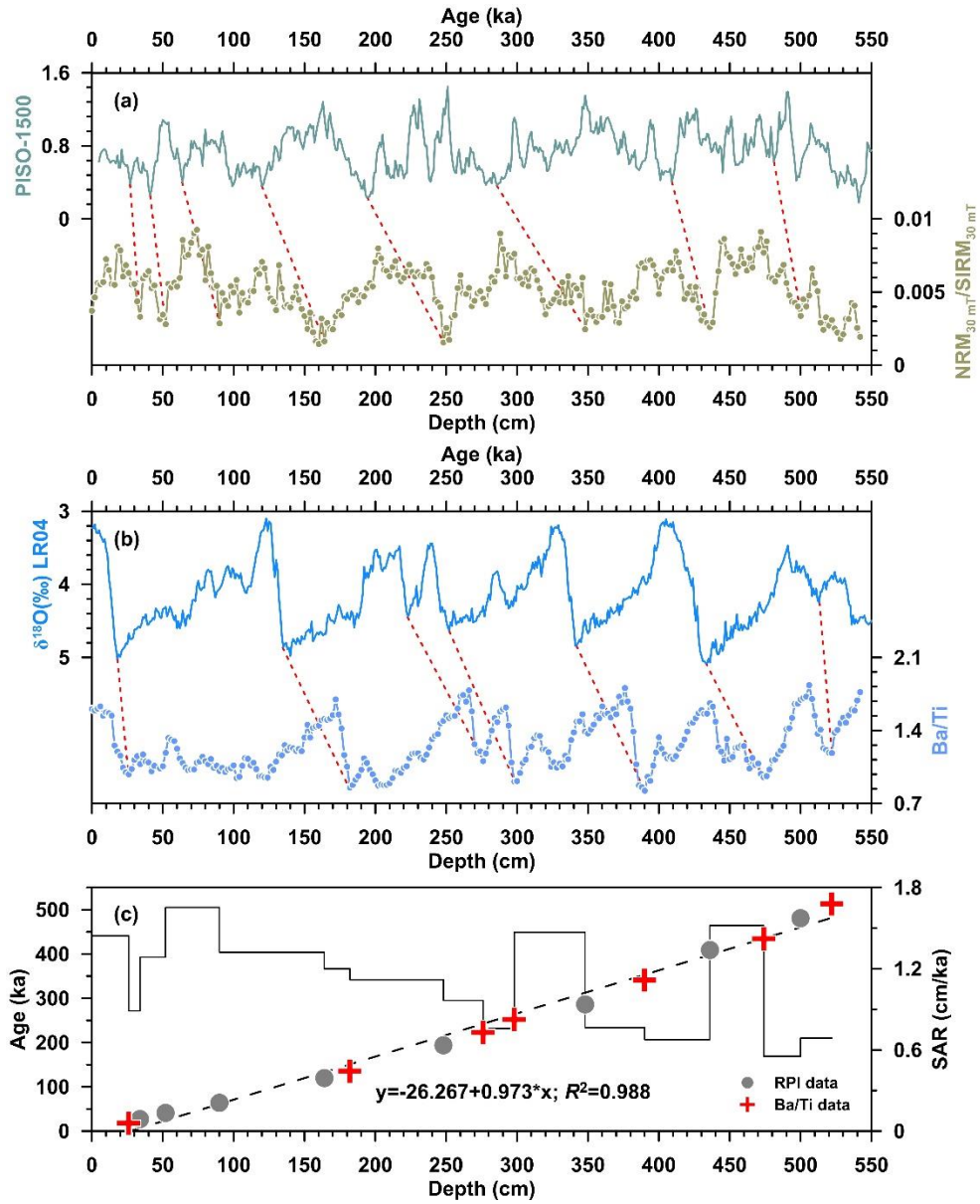


Fig. S2. Age model for core CJ02-08. (a) Relative palaeointensity (RPI) records (RPI = $NRM_{30mT}/SIRM_{30mT}$) for core CJ02-08¹⁴ matched with the PISO-1500 stack¹⁵. (b) Ba/Ti for core CJ02-08 matched with the LR04 marine oxygen isotope stack¹⁶. (c) Depth-age plot and sediment accumulation rates (SARs) for core CJ02-08. Fifteen age tie points were obtained from geomagnetic minima (grey dots) and by correlating abrupt Ba/Ti increases to each major glacial termination (red crosses), indicated by red dashed lines. The black dashed line represents a linear correlation between age and depth.

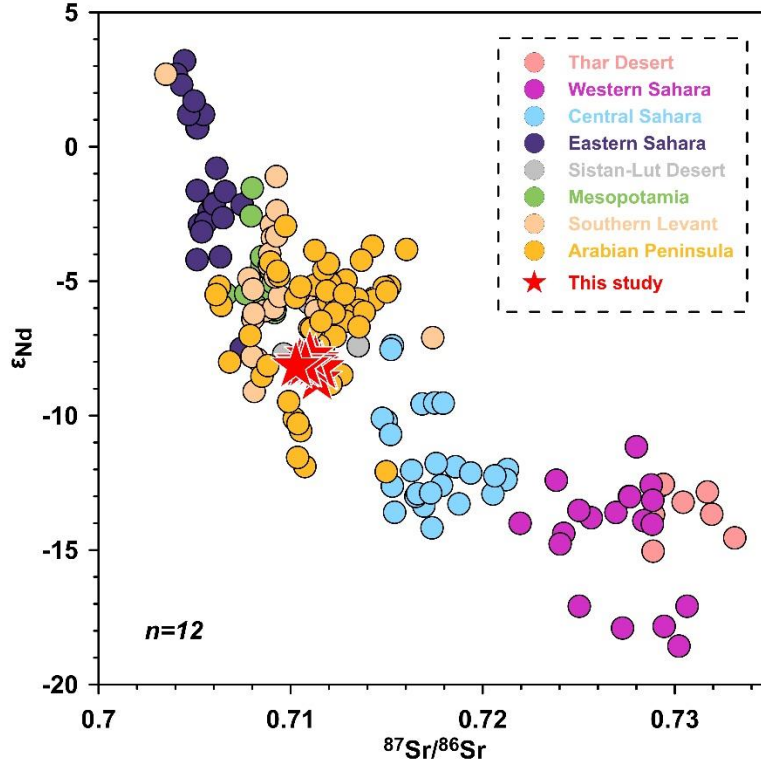


Fig. S3. Comparison of ϵ_{Nd} and $^{87}Sr/^{86}Sr$ data from core CJ02-08 (red stars) compared to potential dust sources (circles). Potential dust sources include the Thar Desert²⁴ (pink), Western Sahara^{25,26} (purple), Central Sahara^{25,27} (cyan), Eastern Sahara^{28,29} (dark blue), Sistan-Lut Desert³⁰ (grey), Mesopotamia³¹⁻³³ (green), Southern Levant^{34,35} (light yellow), and Arabian Peninsula³⁶ (deep yellow). Note the strong correspondence between the radiogenic isotope composition of samples from CJ02-08 and the Arabian Peninsula geochemical fingerprint.

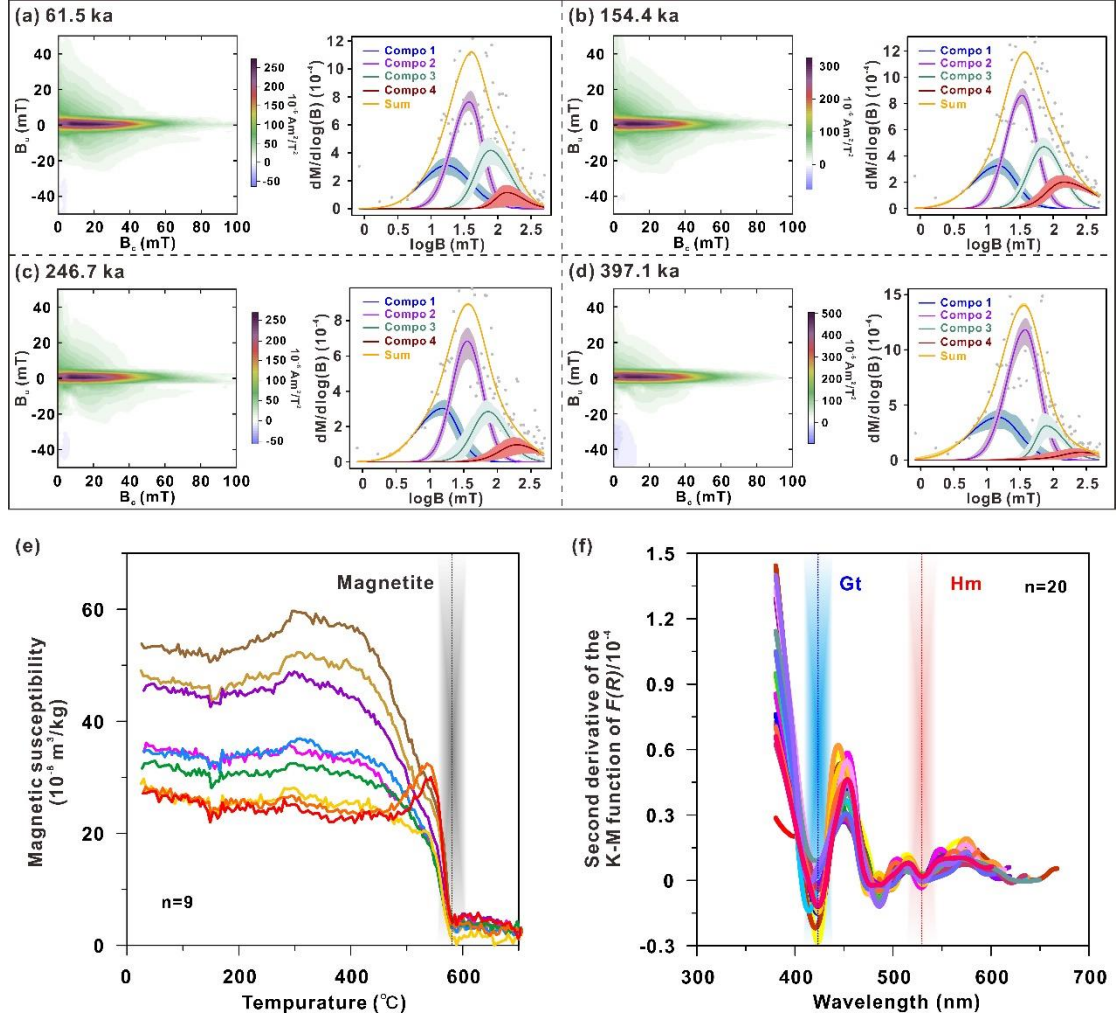


Fig. S4. The results of rock magnetism and diffuse reflectance spectroscopy from representative samples. (a)-(d) First-order reversal curve (FORC) diagrams and Isothermal remanent magnetization (IRM) decomposition curves for representative samples. In IRM decomposition curves, blue, purple, green, red, and orange curves indicate components 1, 2, 3, and 4 and the fitted total spectrum, respectively. (e) Temperature-dependence of magnetic susceptibility curves (heating curves) for selected samples. The black dotted line marks the Curie temperature of magnetite. (f) Second derivative of diffuse reflectance spectroscopy (DRS) for typical samples. Blue and red dotted lines indicate characteristic peak positions for goethite (Gt) and hematite (Hm), respectively.

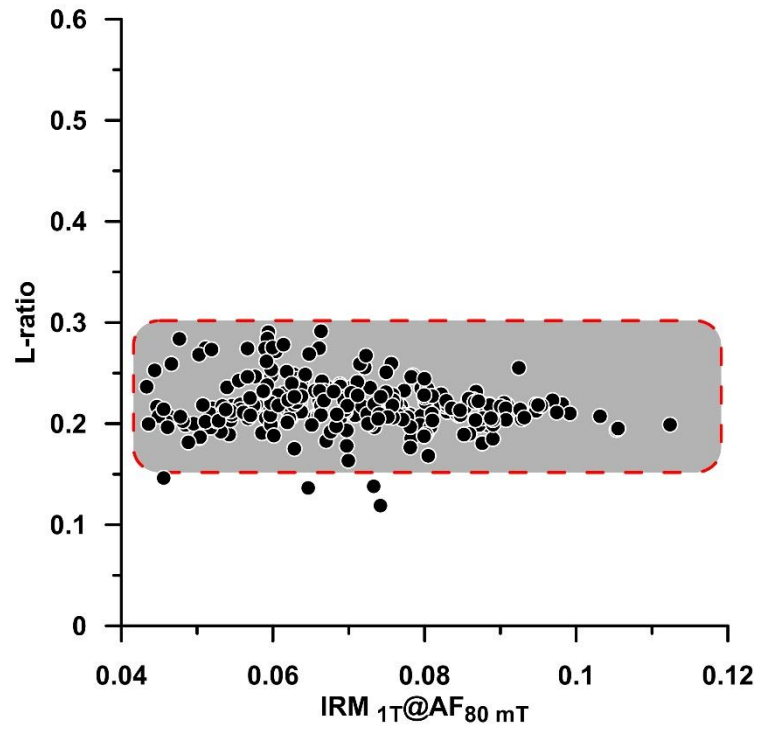


Fig. S5. Correlation between the L-ratio and IRM_{1T@AF80 mT} for samples from core CJ02-08.

The gray shading indicates that the L-ratio values have a relatively narrow range.

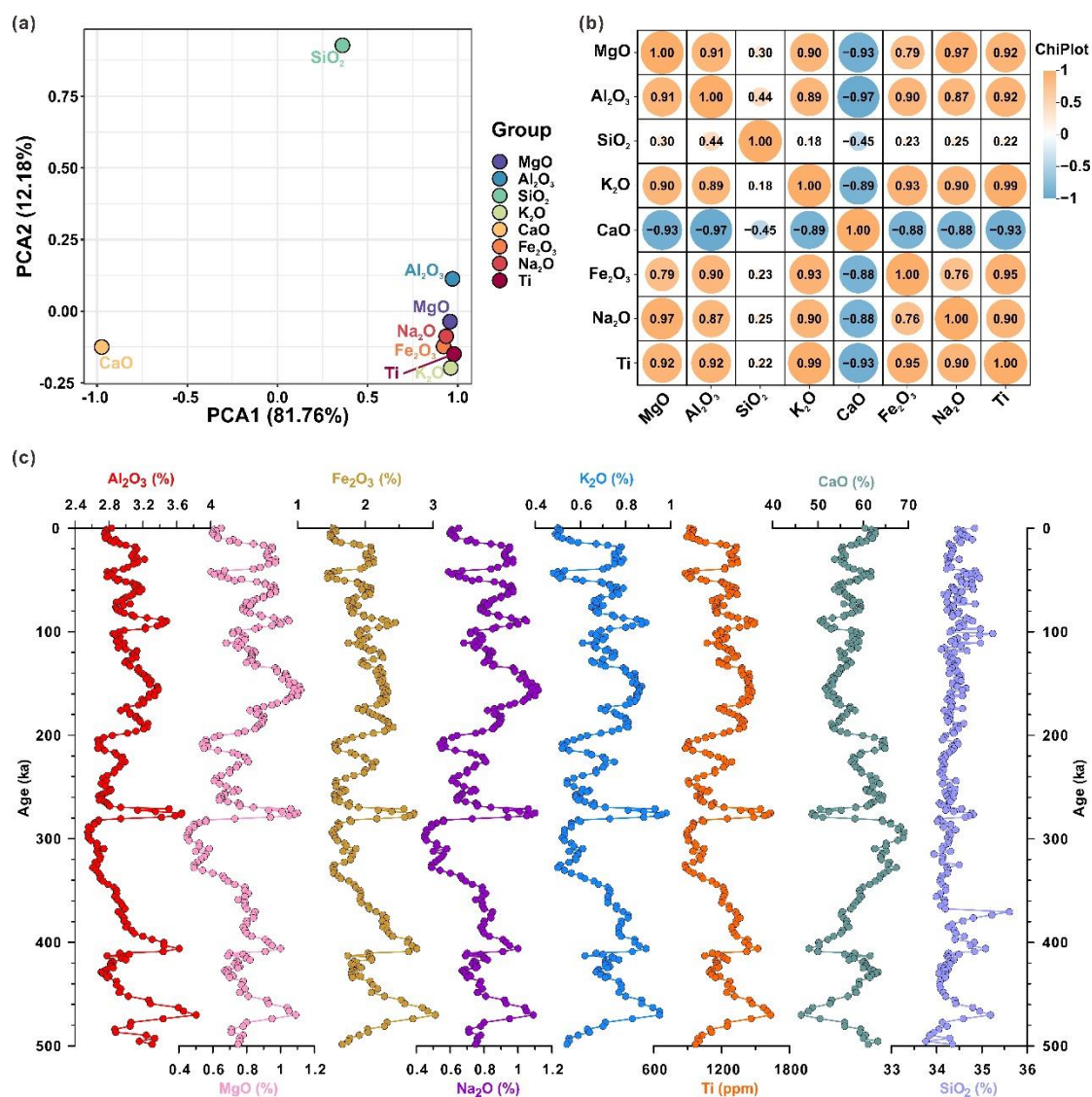


Fig. S6. The results of major element analysis from core CJ02-08 samples. (a) Biplot of the first two principal components (PC1 and PC2) with loadings for the analyzed major elements. (b) Correlation heatmap analysis for the major elements. (c) Major element variations with respect to age.

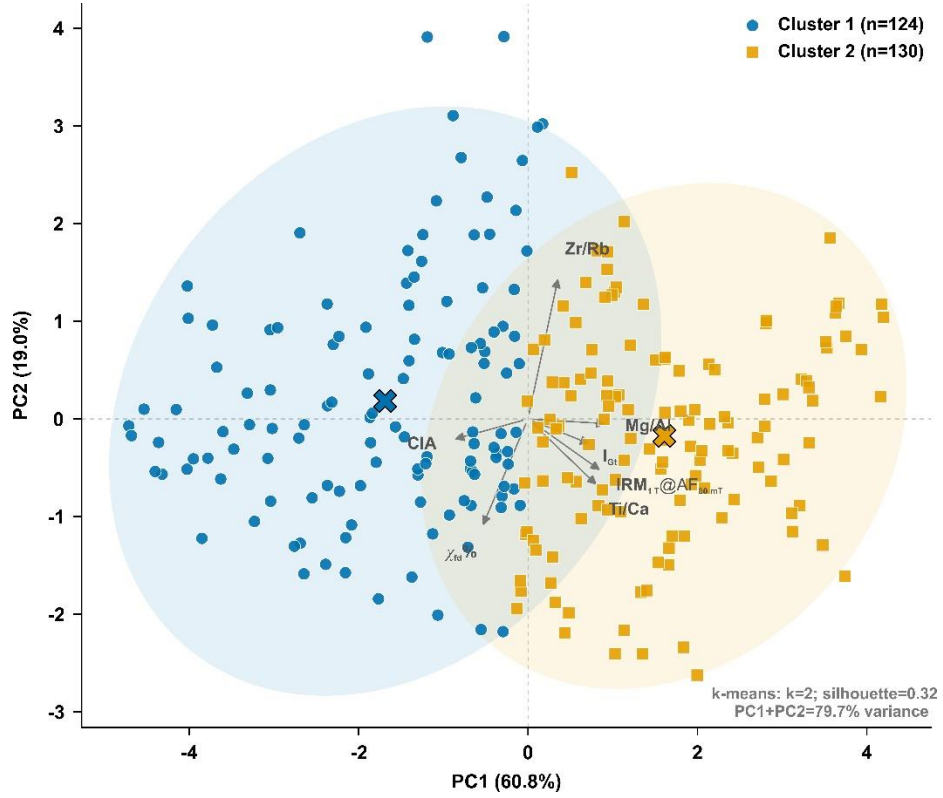


Fig. S7. Principal component analysis results for palaeoclimatic proxies for CJ02-08. Samples are projected onto the first two principal components (PC1 and PC2), which together explain 79.7% of the total variance. K-means clustering ($k = 2$) separates the samples into two groups, shown by blue circles and orange squares, with shaded ellipses indicating the distribution of each cluster. Grey arrows indicate the loadings of chemical index of alteration (CIA), percentage frequency dependent magnetic susceptibility ($\chi_{fd}\%$), Zr/Rb, Mg/Al, goethite band intensity derived from diffuse reflectance spectroscopy (I_{Gt}), isothermal remanent magnetization acquired at 1 T after alternating field demagnetization at 80 mT ($IRM_{1T@AF_{80mT}}$) and Ti/Ca on the first two principal components. Crosses mark the cluster centroids. The silhouette coefficient of the two-cluster solution is 0.32.

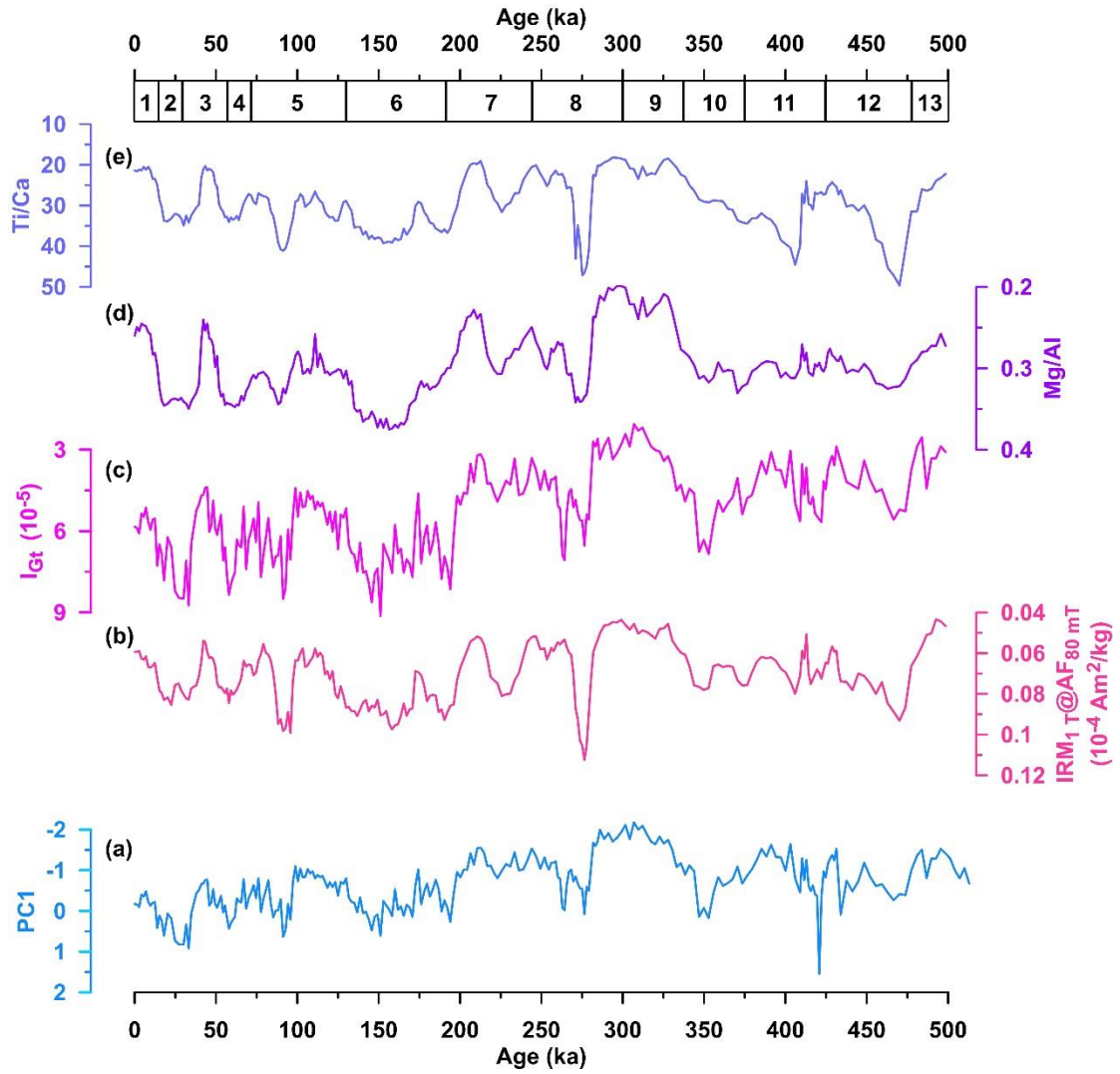


Fig. S8. Four dust proxy records and principal component 1 (PC1) variations with age. (a) principal component 1 (PC1) from principal component analysis. (b) Isothermal remanent magnetization acquired at 1 T after alternating field demagnetization at 80 mT ($IRM_{1\ T@AF\ 80\ mT}$). (c) Goethite band intensity derived from diffuse reflectance spectroscopy (I_{Gt}). (d) Mg/Al. (e) Ti/Ca.

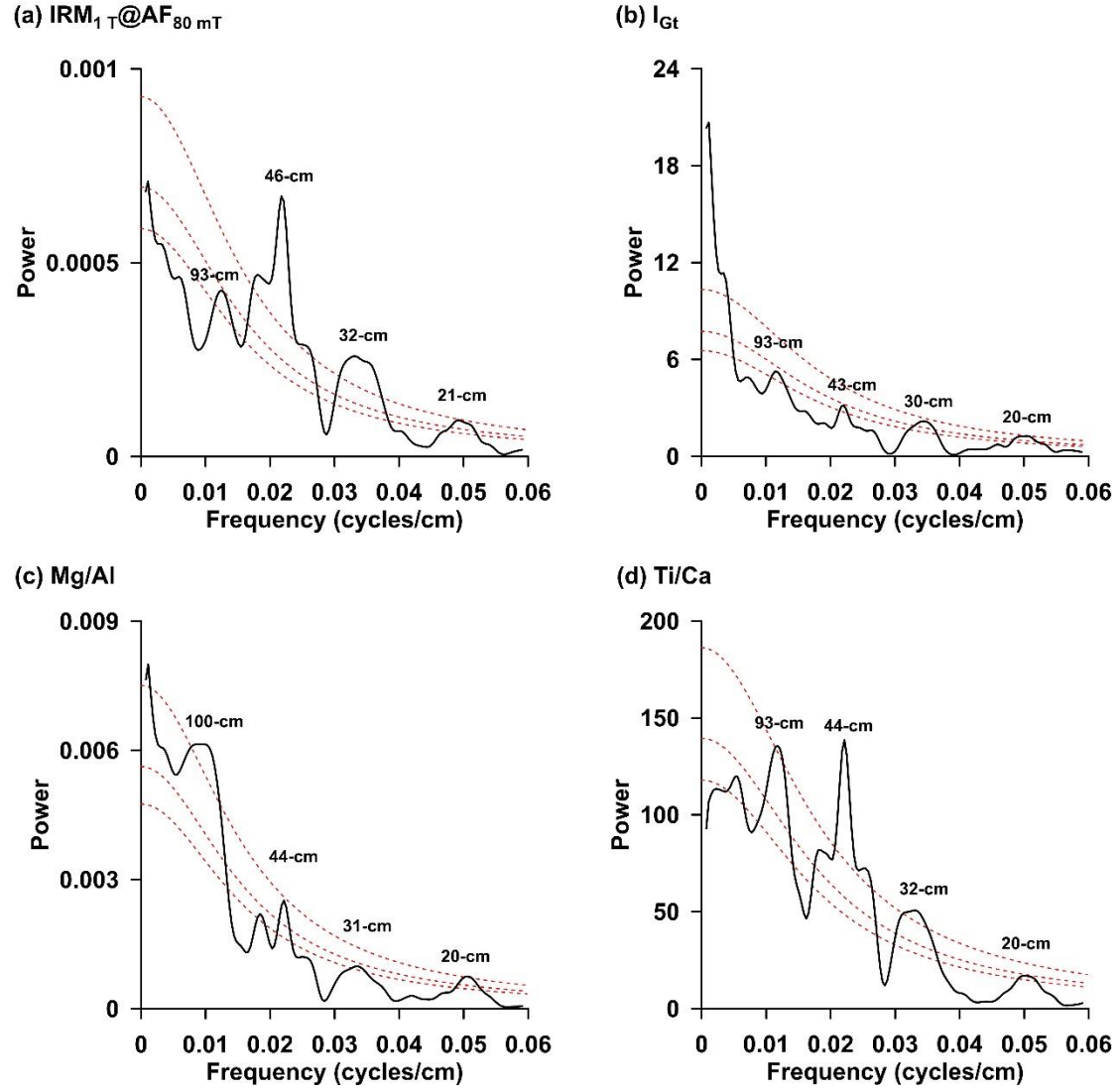


Fig. S9. Power spectral estimates for depth-series dust proxies using the Astrochron R package⁶⁹. (a) Isothermal remanent magnetization acquired at 1 T after alternating field demagnetization at 80 mT ($IRM_{1T}@AF_{80mT}$). (b) Goethite band intensity derived from diffuse reflectance spectroscopy (I_{Gt}). (c) Mg/Al . (d) Ti/Ca . The red dash lines represent 99%, 95%, and 90% confidence level curves.

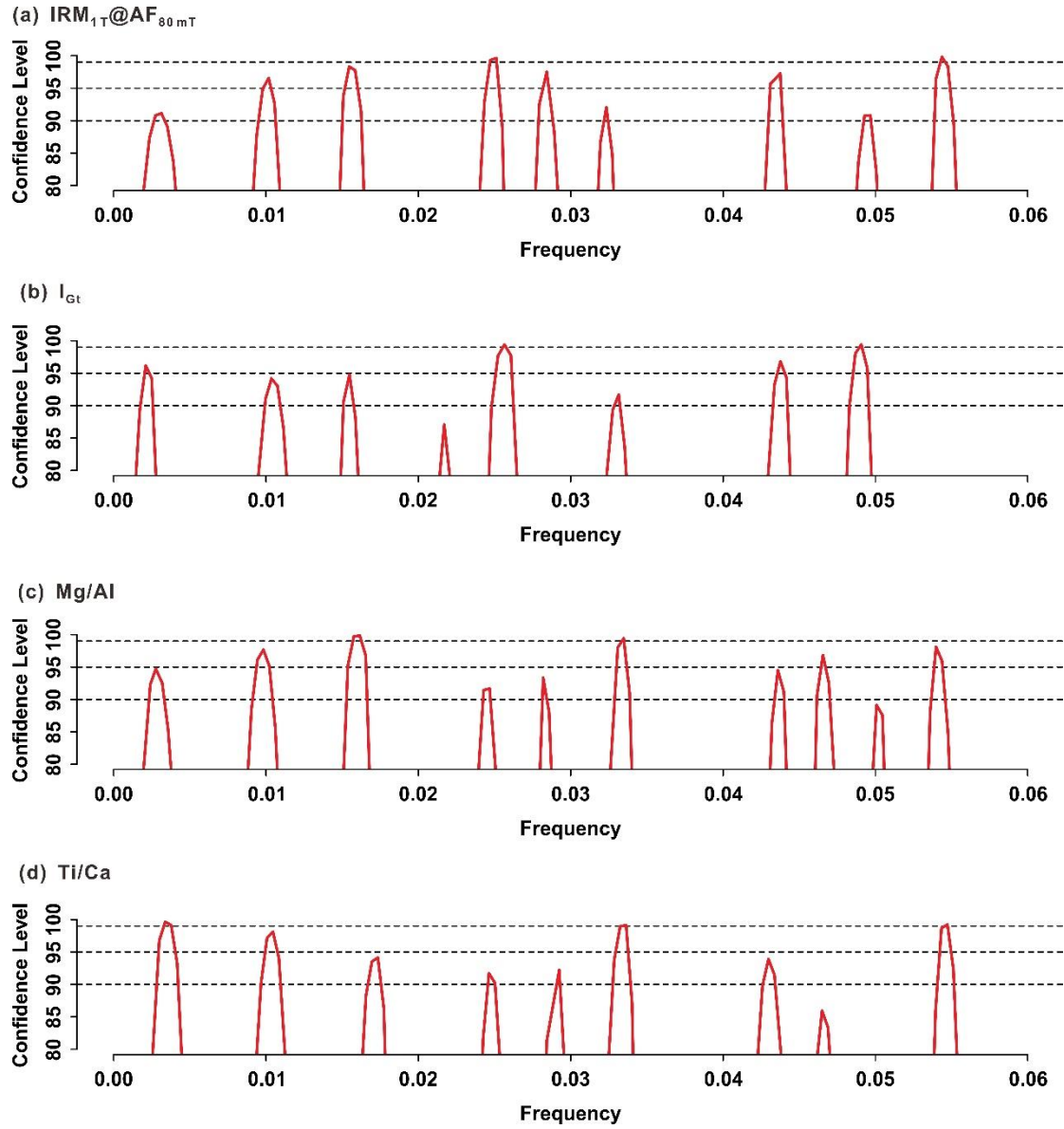


Fig. S10. Results of false testing for age-series dust proxies using the Astrochron R package⁶⁹.

(a) Isothermal remanent magnetization acquired at 1 T after alternating field demagnetization at 80 mT ($IRM_{1T}@AF_{80mT}$). (b) Goethite band intensity derived from diffuse reflectance spectroscopy (I_{Gt}). (c) Mg/Al . (d) Ti/Ca . The red dash lines represent 99%, 95%, and 90% confidence level curves.

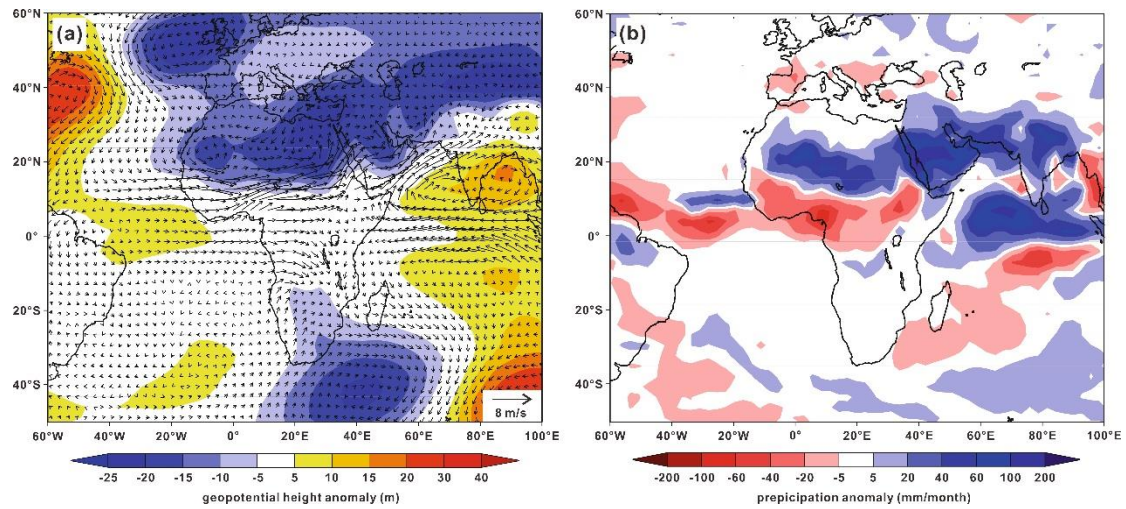


Fig. S11. Climate model simulation of horizontal wind (arrows), geopotential height, and precipitation anomalies for the Eemian, MIS 5.5. (a) Horizontal wind in m/s (arrows), and geopotential height in m (colours) at the 850 hpa level. (b) Summer precipitation (in mm/month) with vegetation feedbacks are included in the simulation. Model results are expressed as Eemian minus pre-industrial for the parameters shown. The horizontal wind, geopotential height, and precipitation anomalies reconstructions were newly generated by the authors based on the model results of Herold & Lohmann⁹⁰. Figures were created using Panoply software v5.3.3 (<https://www.giss.nasa.gov/tools/panoply/>).

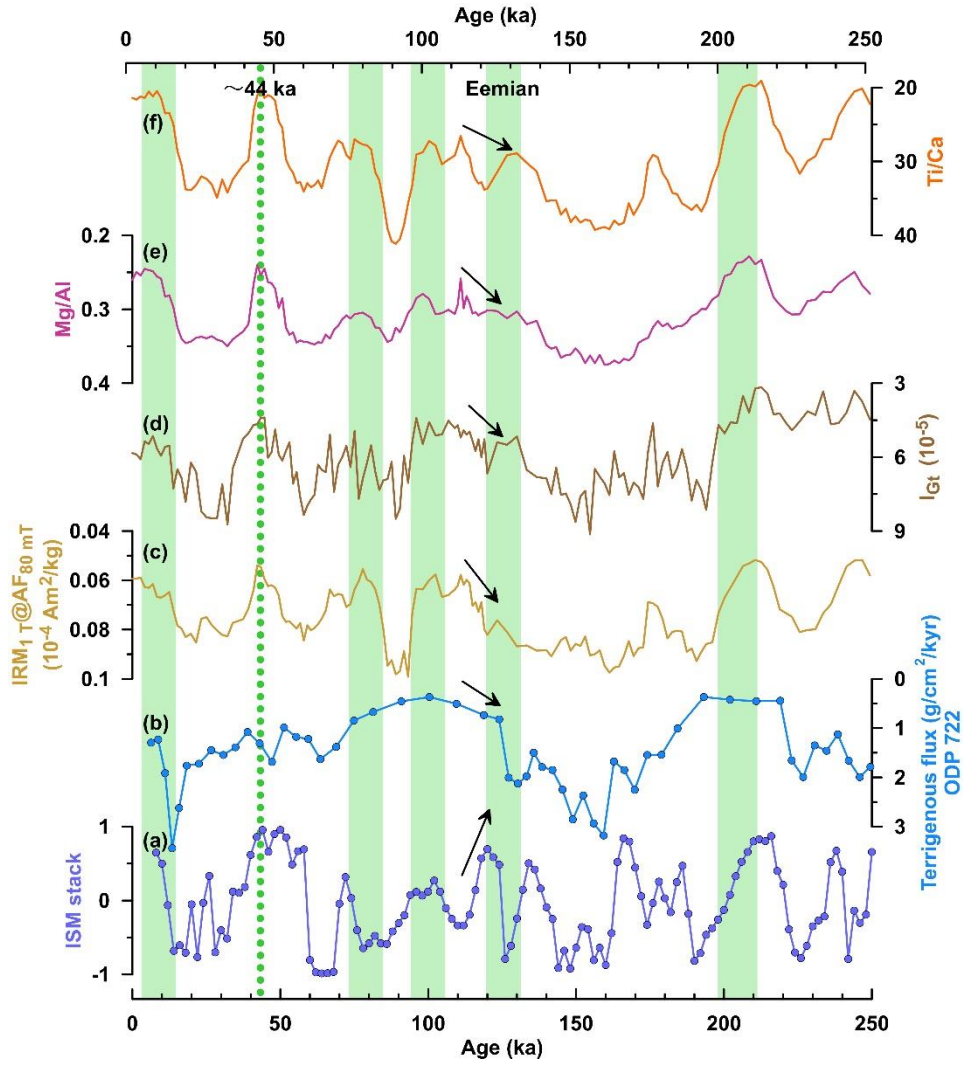


Fig. S12. Arabian dust and Indian summer monsoon (ISM) intensity variations. (a) Arabian Sea Indian Summer Monsoon (ISM) Stack⁶⁸. (b) Terrigenous input record from Oceanic Drilling Program (ODP) site 722, Arabian Sea⁸⁶. (c) Isothermal remanent magnetization acquired at 1 T after alternating field demagnetization at 80 mT ($IRM_1 T@AF_{80\text{ mT}}$). (d) Goethite band intensity derived from diffuse reflectance spectroscopy (I_{Gt}). (e) Mg/Al. (f) Ti/Ca. The green bars represent Green Arabia periods (GAPs). Green dashed lines indicate rapidly decreased dust contents at ~41 ka.

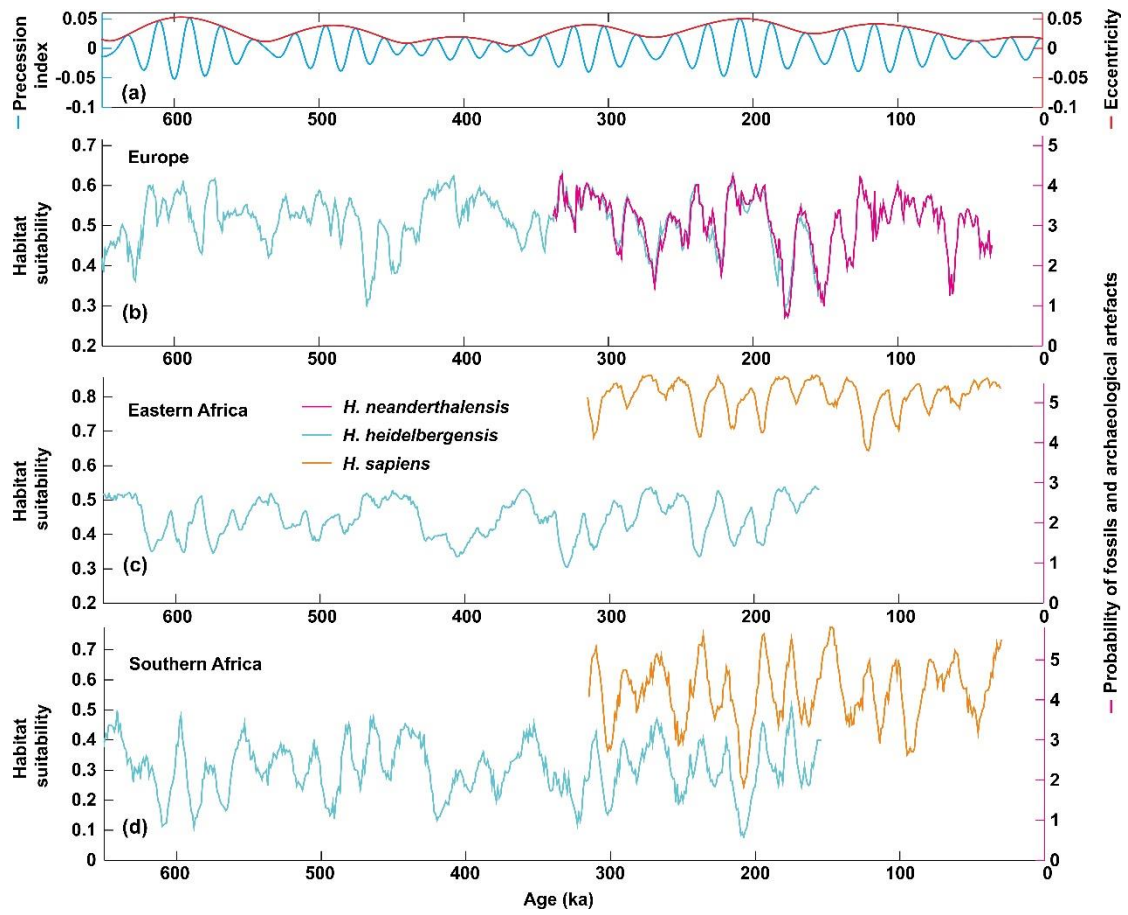


Fig. S13. Hominin species succession in Europe and Africa. (a) Orbital eccentricity (red) and precession (green) parameters⁹¹. (b) Habitat suitability calculated from ref.⁹² for *H. heidelbergensis* (green curves) and *H. neanderthalensis* (red curves) in Europe. (c) Same as b but for *H. heidelbergensis* (green) and *H. sapiens* (orange) in eastern-central Africa. (d) Same as c but for southern Africa.

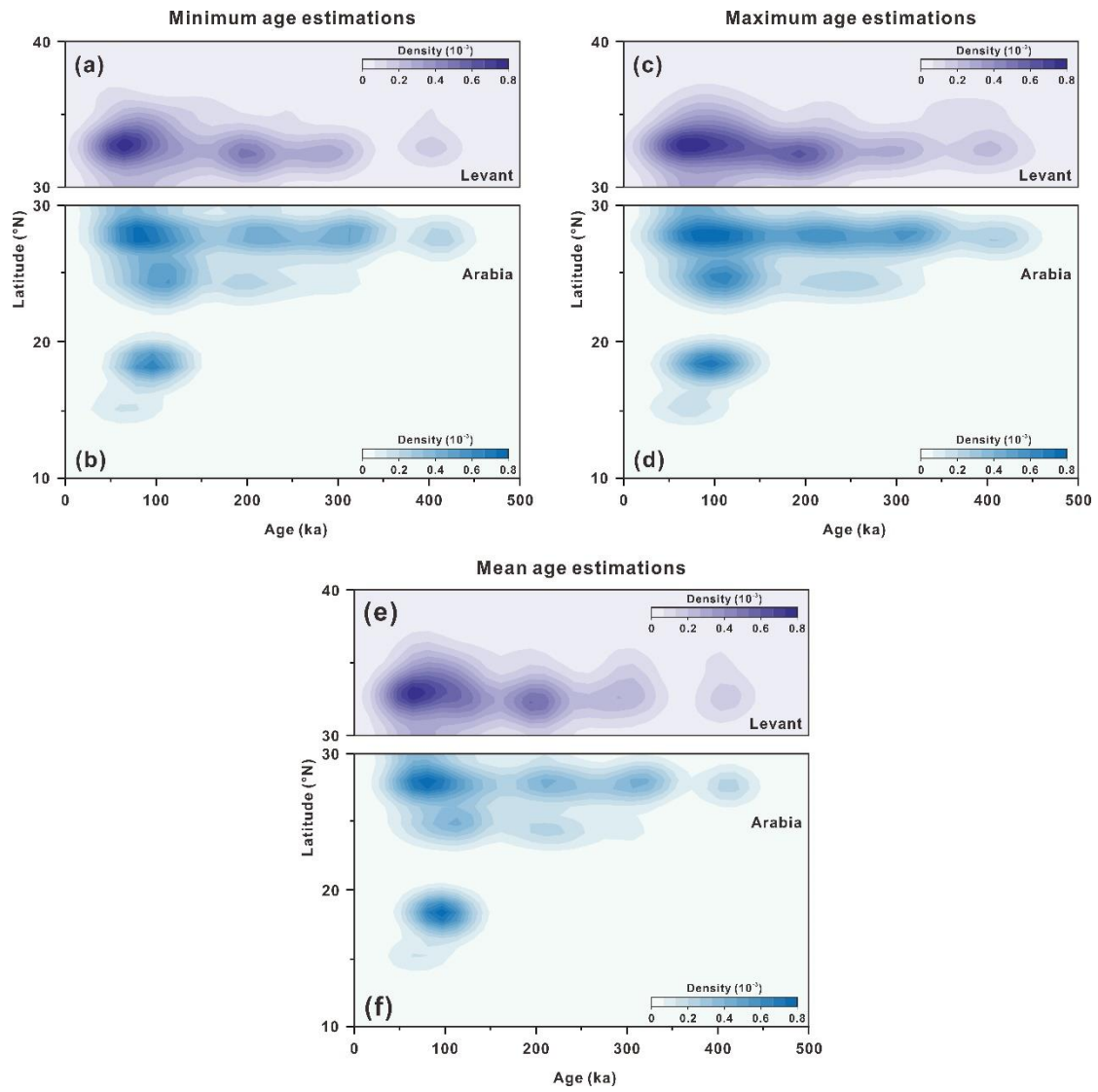


Fig. S14. Potential influences of dating errors on spatial and temporal distributions of Middle-Late Pleistocene archaeological and fossil hominin sites across Arabia and the Levant. (a-b) Minimum age estimations. (c-d) Maximum age estimations. (e-f) Mean age estimations.

Table S1. Sample age of sediment (ka), $^{87}\text{Sr}/^{86}\text{Sr}$, $^{143}\text{Nd}/^{144}\text{Nd}$, and ϵ_{Nd} values.

Age (ka)	$^{87}\text{Sr}/^{86}\text{Sr}$	$\pm\sigma$	$^{143}\text{Nd}/^{144}\text{Nd}$	$\pm\sigma$	ϵ_{Nd}
30.1	0.711275	0.000003	0.512222	0.000003	-8.111099
57.9	0.710994	0.000004	0.512240	0.000003	-7.75986
90.4	0.711640	0.000004	0.512217	0.000003	-8.21827
123.3	0.710485	0.000004	0.512227	0.000002	-8.02321
145.1	0.711294	0.000003	0.512231	0.000003	-7.94128
187.3	0.711373	0.000004	0.512194	0.000004	-8.66693
236.2	0.710617	0.000004	0.512221	0.000003	-8.1383
280.6	0.710796	0.000004	0.512221	0.000003	-8.1383
344.1	0.711564	0.000004	0.512216	0.000003	-8.23973
382.4	0.710773	0.000004	0.512221	0.000003	-8.1344
434.0	0.711146	0.000003	0.512226	0.000002	-8.04076
477.4	0.710274	0.000004	0.512219	0.000003	-8.17146

Note: ϵ_{Nd} was determined using the present CHUR value from Jacobsen and Wasserburg²³,

$$\epsilon_{\text{Nd}} = [(^{143}\text{Nd}/^{144}\text{Nd}_{(\text{measured})}/0.512638 - 1) \times 10^4$$

Table S2. Results of IRM acquisition curve decomposition for selected core CJ02-08 samples.

Age (ka)	Component 1			Component 2			Component 3			Component 4		
	$B_{1/2}$ (mT)	DP	IRM Rel. (%)	$B_{1/2}$ (mT)	DP	IRM Rel. (%)	$B_{1/2}$ (mT)	DP	IRM Rel. (%)	$B_{1/2}$ (mT)	DP	IRM Rel. (%)
61.5	16.7	0.36	26.8	33.3	0.23	42.0	88.0	0.25	25.3	261.7	0.24	5.8
151.4	12.0	0.33	21.7	32.1	0.22	40.1	78.0	0.25	24.2	298.3	0.35	14.1
246.7	11.8	0.30	26.3	35.5	0.22	44.9	78.0	0.25	20.6	318.4	0.29	8.1
397.1	11.8	0.38	27.9	33.6	0.24	55.0	97.4	0.22	12.6	375.6	0.32	4.6

Table S3. Estimated past hominin populations in Arabia during past present-day like desert conditions and the Eemian Green Arabia period (GAP).

State	Ecosystem	Total surface (km ²)	Lake surface (km ²)	Habitable surface (km ²)	Population density (ind./km ²)	Population (individuals)
Eemian GAP	Mediterranean	128361	0	128361	0.1	12,836
	Desert	52157	0	52157	0	0
	Grassland	1085342	1500	1083842	0.01	10,838
	Dry wooded grassland (<700 mm/yr)	2487671	2000	2485671	0.056	139,198
	Total					162,872
Desert	Mediterranean	107870	0	107870	0.1	10,787
	Desert	3342259	0	3340979	0	0
	Grassland	293298	0	293298	0.01	2,933
	Dry wooded grassland (<700 mm/yr)	10104	0	10104	0.056	566
	Total					14,286

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