Impact of climate change on the transition of Neanderthals to modern humans in Europe

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Two speleothem stable isotope records from East-Central Europe demonstrate that Greenland Stadial 12 (GS12) and GS10—at 44.3–43.3 and 40.8–40.2 ka—were prominent intervals of cold and arid conditions, GS12, GS11, and GS10 are coeval with a regional pattern of culturally (near-)sterile layers within Europe’s diachronous archeologic transition from Neanderthals to modern human Aurignacian. Sterile layers coeval with GS12 precede the Aurignacian throughout the middle and upper Danube region. In some records from the northern Iberian Peninsula, such layers are coeval with GS11 and separate the Châtelperronian from the Aurignacian. Sterile layers preceding the Aurignacian in the remaining Châtelper- ronian domain are coeval with GS10 and the previously reported 40.0- to 40.8-ka cal BP (calendar years before present (1950)) time range of Neanderthals’ disappearance from most of Europe. This suggests that ecologic stress during stadial expansion of steppe landscape caused a diachronous pattern of depopulation of Neanderthals, which facilitated repopulation by modern humans who appear to have been better adapted to this environment. Consecutive depopulation—repopulation cycles during severe stadials of the middle pleniglacial may principally explain the repeated replacement of Europe’s population and its genetic composition.

Central Europe | speleothems | millennial-scale climate cycles | stable isotopes | Middle–Upper Paleolithic transition

The replacement of Neanderthals by modern humans is recorded across Europe in a diachronous and culturally complex succession of distinct stone tool assemblages from the Middle–Upper Paleolithic transition (MUPT) roughly between 48 and 36 ka cal BP (calendar years before present (1950)) (1, 2). The succession often includes a regionally distinct “transitional” assemblage of local origin, or an intrusive Initial Upper Paleolithic assemblage between the Neanderthal Mousterian and modern human Aurignacian. The oldest anatomically modern human remains from Europe—found in East-Central Europe and radiocarbon dated to 40.6–38.6 ka cal BP (68% probability) time of the Neanderthals’ disappearance from most of continental Europe—carry genetic evidence for species interbreeding four to six generations earlier (3–6). This individual, however, represents a population that did not contribute to the genome of modern humans present in glacial Europe after the MUPT (7), and the archeologic record provides no site with indication of local coexistence. Within a few millennia after the MUPT, at least two other genetically distinct modern human populations came to subsequently dominate Middle Pleniglacial Europe. During the entire interval, northern hemispheric climate went through several millennial-scale Dansgaard–Oeschger (DO) cold cycles (8, 9). A causality between climate change, the archeologic succession, and modern humans’ genetic makeup has been tentatively suggested but not demonstrated (1, 2, 7). Below, we present the climatic history of continental Europe during the MUPT and derive the impact of climate change on MUPT demography, which may have led to the apparent repetitive genome turnover reported for Europe’s human population during the middle pleniglacial.

Significance

A causality between millennial-scale climate cycles and the replacement of Neanderthals by modern humans in Europe has tentatively been suggested. However, that replacement was diachronous and occurred over several such cycles. A poorly constrained continental paleoclimate framework has hindered identification of any inherent causality. Speleothems from the Carpathians reveal that, between 44,000 and 40,000 years ago, a sequence of stadials with severely cold and arid conditions caused successive regional Neanderthal depopulation intervals across Europe and facilitated staggered repopulation by modern humans. Repetitive depopulation—repopulation cycles may have facilitated multiple genetic turnover in Europe between 44,000 and 34,000 years ago.

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suggest a cold steppe environment with few conifers during interstadials, and a tundra landscape with cryosol formation during stadials (12, 22, 23). During GIs, a temperate forest-steppe prevailed west (marine records) and south of the Alps (Monticchio, Castiglione, and Lagaccione) (12). Dust deposition in the Eifel maar lakes and speleothem carbon isotopes from Villars Cave suggest increasing aridity in Western Europe across the MUPT (24, 25).

The MUPT Speleothem Paleoclimate Record of East-Central Europe

Here, we present two speleothem records from East-Central Europe (Romania). Stalagmite POM1 is from Ascunșa Cave (AC), South Carpathians, 50 km north of the Danube Valley at 1,050-m altitude (SI Appendix, Fig. S1). Stalagmite 1152 is from Tâușoare Cave (TC), East Carpathians, at 950-m altitude, 2.5° north of the AC site. We dated the speleothems by U-Th and measured stable carbon and oxygen isotopes continuously at decadal resolution (SI Appendix, Fig. S2 and Table S1).

The AC speleothem δ13C record reproduces the DO pacing of the Greenland ice core record during the MUPT (Fig. 1). The TC δ13C record reproduces some aspects of the Greenland record. The climatic meaning of the AC δ13C data may be constrained within context of other records from the region, particularly the Black Sea (Fig. 1). The coherency of the AC δ13C record with southern Black Sea IRDc (11) implies that speleothem δ13C responded to changes in the length of the sea ice and winter frost season. Chemically, speleothem δ13C depends on the proportion of low δ13C aqueous CO2 derived from microbial respiration of soil organic matter above the cave (26). Extended frost and shorter plant growth seasons reduces the supply of fresh soil organic matter (SOM) and increases the proportion of CO2 from old SOM in drip water, which has up to 8‰ higher δ13C than fresh SOM under comparable conditions (27). Other potential controls are insufficient to explain the 6‰ amplitude of DO cycles in the speleothem. For example, the effect of a seasonally variable CO2 degassing rate from drip water may add a kinetic fractionation effect on the δ13C of precipitating speleothem calcite (28). However, recent monitoring showed that interannual variability in δ13C of precipitating calcite in AC is less than 1.5‰ (29). Similarity between the AC δ13C and δ18O records is limited (SI Appendix, Fig. S3)—another qualitative argument against major kinetic isotope fractionation. Also, a compositional change in SOM δ13C can be ruled out as variability in Central European paleosol profiles does not exceed 1‰ during the MUPT (30). Finally, variable moisture availability may have influenced soil formation above the cave and speleothem δ13C. The Danube loess record suggests such an interval of enhanced moisture availability after GI8 (20). The AC-speleothem δ18O record does indeed suggest some change to the effect between GS8 and GI7, but the overall pacing across the MUPT resembles that of southeastern Mediterranean records and shows little coherency with DO cycles (see below). In general, soil formation intervals recorded by low values in AC-δ13C are coeval with the range of ages obtained for paleosols in the upper Danube Willendorf II profile (Fig. 2). This suggests that the entire Danube region was inside the same climate zone and responded coherently to temperature change during DO cycles of the MUPT.

Little variance is observed in the East Carpathian TC δ13C record (SI Appendix, Fig. S3). Here, pyrite weathering and enhanced limestone dissolution in the host rock obscures any potential influence on δ13C from SOM (SI Appendix). That record is not considered in this study.

In general, speleothem δ18O reflects a combination of calcite precipitation temperature—the annual average temperature inside the cave—and factors controlling δ18O in drip water. Within SOM, δ18O of water and dissolved CO2 equilibrated in the aquifer above the cave, and drip water composition reflects regional hydroclimatology, that is, moisture source, rain-out history, local rain-out temperature, and the annual distribution of rainfall (26, 31). The AC δ18O signal’s overall amplitude is ∼4‰. Such a large change is beyond reasonable temperature variability (28) and suggests a dominant influence of changes in hydrology on AC δ18O. The dissimilarity of AC δ18O to the coherent δ13C and Greenland temperature records (SI Appendix, Fig. S3) renders a temperature control of the AC δ18O record unlikely. Hydroclimatic dominance on speleothem δ18O is typical for the entire Eastern Mediterranean domain’s Holocene record including at AC, which reflects significant variability in rainfall seasonality and moisture source proportions (31). The similar pacing between AC δ18O and southeastern Mediterranean δ18O record (SI Appendix, Fig. S4) may indicate that the underlying synoptic-scale mechanism was also active during MIS3.

The TC speleothem δ18O record has an amplitude of ∼1‰ and a different pacing compared with AC δ18O (SI Appendix, Fig. S3). This suggests that the East Carpathians were in a different hydroclimatic regime during the MUPT. What defines the East Carpathian regime may be investigated by exploring the partial coherency between TC δ18O and the other paleoclimate record. After ∼60 ka, the δ18O record shows low-amplitude variability with a few pronounced minima of ∼0.5‰ that are coeval with some of the Greenland stadials (Fig. 1). These include GS3-H2, GS5-H3, and GS13-H5. Other minima just after 34 ka and at
41 ka correspond to GS7 and GS10 (Fig. 1). The latter lasted from 40.66 ± 0.47 ka$_{U-Th}$, $P = 95\%$—directly dated—to an interpolated end date of ~39.70 ± 0.43 ka (32). Although there may be a few centuries of overlap with GS9-H4, the measured age interval of 40.7–39.8 ka$_{U-Th}$, ±0.47 to ±0.3 uncertainty, matches GS10 (40.80–40.16 ka$_{GICC05}$) better than GS9-H4 (from 39.90 to 38.22 ka$_{GICC05}$) (9). A poorly resolved relative $\delta^{18}O$ maximum at 39.6 ± 0.4 ka$_{U-Th}$ and a subsequent broad relative $\delta^{18}O$ minimum in the TC record appear to be coeval with GS19 and GS9-H4, which apparently had a lesser impact on Central European climate. The TC $\delta^{18}O$ record thus responded to stadial cooling in the Atlantic but with a regionally controlled amplitude. Similarity with the northern Black Sea paleotemperature archive suggests this response is temperature driven (Fig. 1). For two reasons, TC $\delta^{18}O$ likely records the local summer temperature signal. First, the northern Black Sea record shows that summer sea surface temperature between 40 and 20 ka cal BP dropped by ~2 °C during stadials from a MIS3 average of 4–5 °C (33). Assuming published global empirical temperature relationships for cave calcite—$\Delta \delta^{18}O$/$\Delta T = −0.18$ (28)—and for rainwater in mid-latitudes—where $\Delta \delta^{18}O$/$\Delta T = 0.58$ (34)—the 0.4–0.6‰ $\delta^{18}O$ amplitude in the TC speleothem would correspond to a comparable 1.0–1.5 °C temperature change. Second, palynologic, geomorphologic, and geochemical studies support this interpretation as follows. During the last glacial, the lowlands north of the East Carpathians were situated inside the tundra biome, where field evidence of ice wedges indicates permafrost conditions (35–37). Deep winter frost or even discontinuous permafrost is a likely scenario for the >1,000-m altitude of the ground above TC in the East Carpathians. At 20–40 ka, the average ground temperature of the northern Black Sea drainage basin—comprising the Danube, Dniester, Dniepr, and Don Lowlands—was ~4 °C (38). Common adiabatic gradients of 6–10 °C per km of altitude imply that annual average ground temperatures above TC were lower than ~2 °C. This suggests an alpine near-permafrost environment. Permafrost conditions may have been continuous during extreme stadials, thus preventing groundwater recharge to the cave and interrupting speleothem growth. During GS12—the coldest of all stadials in the Greenland record—the TC speleothem shows a hiatus coeval with the most pronounced $\delta^{13}C$ maximum in the AC speleothem further south. In the Willendorf II loess profile, this interval correlates with tundra grey horizon C9 with evidence of permafrost (22) (Fig. 2). C9 is constrained in time by paleosols C8-3 above and D1 underneath with dating ranges similar to GI11 and GI12, respectively. The Willendorf II profile is located at the same altitude as the TC speleothem, but only at 230-m altitude. Continuous permafrost conditions at 950-m altitude in the East Carpathians were thus likely during GS12. Seasonally permeable frost or discontinuous permafrost at other times would restrict speleothem growth to the summer season. Unlike the hydrologically dominated AC $\delta^{18}O$ record, TC $\delta^{18}O$-only records regional summer temperature change.

**Paleoclimatic Context of the Middle Pleniglacial in Europe**

A reduction of forest and expansion of steppe biomes occurred during all stadials (12–15, 21), but climate records between the Atlantic and the Black Sea show a spatially heterogeneous response to DO cycles (Fig. 2). GS13-H5 is apparent in the TC speleothem $\delta^{18}O$ record but unlike in the Atlantic domain, its amplitude is small compared with subsequent stadials. Southern Black Sea IRDc data (Fig. 1) indicate less sea ice than during subsequent stadials despite apparent colder annual sea surface temperature (Fig. 2). This suggests a different seasonality with less severe winters. Thus, extreme cold was unlikely in East-Central Europe. The extreme aridity apparent in the Aegean region (13) may not have been as severe in the Balkan and Black Sea region (15, 16). GS12, from 43.4–44.3 ka$_{GICC05}$ or 43.3–44.0 ka$_{U-Th}$ (AC), is a very

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**Fig. 2.** (A) Greenland: NGRIP temperature (8). (B) Southern Black Sea: TEX$_{90}$ sea surface temperature core M72-S (10). (C) South Carpathians: AC stalagmite POM1 $\delta^{13}C$. (D) Upper Danube: Willendorf loess/paleosol profile, paleosol ages, age probability density functions (age-pdf), and stratigraphically constrained loess and grey horizons C7-2, C7-3, and C9 (22). (E) East Carpathians: TC stalagmite 1152 $\delta^{13}C$. (F) Eifel maar lakes, ELSA dust stack. (G) Western Massive Centrale: Villars Cave stalagmites $\delta^{13}C$ (25). (H) Iberian Atlantic margin: TEX$_{90}$ sea surface temperature core MD95-2042 (17). Numbers indicate Greenland stadials at the end of the respective DO cycle. A map with locations of records is available in SI Appendix, Fig. S1.

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A number of chronologically well-constrained archeological records allow for testing directly the feasibility of a regional depopulation scenario during severe stadials. Culturally (nearly) sterile layers have been reported—directly $^{14}C$ dated or constrained by $^{13}C$-dated cultural layers above and below—in some regions and mark an archeological hiatus of several centuries duration. While such layers in some places represent short-term sedimentologic events without apparent discontinuity of habitation (42), at many sites their time span is multicentennial. These layers mostly relate chronologically to the time of GS12 and GS10, and suggest regionally widespread depopulation for many centuries between two subsequent cultures. (Fig. 3) (SI Appendix, Table S2). During GS12, this is the case for sites from the upper and middle Danube region. At Geissenklösterle, near-sterile (geologic) layer 17 separates Mousterian and Aurignacian assemblages (43) and is directly dated to 44.4–42.7 and 43.5–42.0 ka cal BP by two samples, with a Bayesian model age of 42.9–42.1 ka cal BP (44). At Sesselfelsgrotte, sterile layer F is directly constrained by Mousterian layer G1, 45.8–44.3 ka cal BP, and the late Mousterian layer E3, 42.2–40.5 ka cal BP, both recalibrated (4, 45). At Willendorf II, sterile layer C9 and subsequent layer C8-3 containing Aurignacian artifacts are constrained between two dated paleosols D1 and C8-2 (Fig. 2) (22). Mousterian Neanderthal presence in the middle Danube region has been redated to until ~45 ka cal BP, just before GS12 (46). Two sites with Aurignacian finds, at Keilberg-Kirche (47), 43.6–41.7 ka cal BP recalibrated (4), and at Pes-kő Cave, 43.8–41.2 ka cal BP (48), immediately postdate GS12. These fit the scenario of a cultural hiatus but lack local constraint by older assemblages. All of the above suggest widespread Mousterian Neanderthal.
depopulation of the upper and middle Danube Valley and—except for the tributary Altmühl Valley (49)—repopulation by Aurignacian after GS12.

Circumstantial evidence points to a similar impact of GS12 in northern Italy, where two sites, Riparo Mocho on the Ligurian Mediterranean coast—layer H—and at Grotta di Fumane in the Venetian Pre-Alps—layer A7—contain sterile layers probably coeval with GS12. However, reliable dating is only available for the onset of the subsequent archeologic layer (50, 51) and age uncertainty is large (Fig. 3). A muted response to GS12 may be inferred for the site of Grotte Mandrin, Rhone Valley, where sterile layers separate strata C, B3, and B2 with post-Neronian-II artifacts (50). Layers B3–B1 of this recent local Mousterian expression were dated in the range between 42.7 and 45.0 ka BP (6). These layers may indicate environmental change but do not appear to reflect longer breaks in the site’s occupation. No such layers associated with GS12 are reported from archeologic sites further west. This suggests that the environmental impact of GS12 may have been less severe in Western Europe.

Sterile or near-sterile layers, or cultural hiatuses in the archeologic succession also occur around the time of GS10 (Fig. 3). This time is coeval with Neanderthals’ disappearance from Western Europe (6). Sites include the Grotte-du-Renne—the low-density artifact layer VIII of intermittent Châtelperronian occupation (52); Les Cottés—sterile unit 5 (53); and Saint Césaire—layer Ejo inf (6). At Willendorf II, dated paleosols with lithic artifacts in C8 and C7 broadly constrain the sterile gylo and loess horizons (C7-3 and C7-2) at that time (22) (Fig. 2). Subsequent archeologic layers are generally Aurignacian and document modern human expansion or reoccupation during GI9 and GI8 but under more arid conditions in Europe (Fig. 2).

A few archeologic profiles from around the Pyrenees—at Labeko Koba, Abric Romani, and possibly at Isturitz, but not at L’Arbreda (54–56)—contain sterile layers coeval with GS11 (Fig. 3). These represent a regional cultural hiatus between the Châtelperronian and the (Proto-)Aurignacian. However, apart from a long-term cooling and drying trend over Western Europe, an unambiguously dated climatic cold event cannot be detected in the present speleothem records for that time (Fig. 2).

A general case for likely adaptation of modern humans in response to climate change during the MUPT has already been made (7). However, the stressing cold event originating around the end of GS12 (Fig. 2) appears to mark transitions in the archeologic sequence attributable to Neanderthals that suggest some adaptation as well. GS12 is coeval with the transition from Moustrie- ranian to archaic Uluzzian, and then to evolved Uluzzian in northern Italy (59), and from Moustierian to Châtelperronian in Western Europe (53). However, late Neanderthals may have had a less diverse diet than modern humans (60). In open grasslands, Neanderthals’ exclusive diet was meat from terrestrial animals, whereas modern human Aurignacian also exploited plant and aquatic foods (60, 61). The frequently observed cultural hiatuses, however, do not suggest a direct competitive displacement of Neanderthals, but rather a higher vulnerability to rapid environmental change and ecologic stress in the open landscape during cold and arid GS12–GS10. While Neanderthals did not survive GS12 in most of the Danube steppe and tundra, modern humans may have been more capable to adapt and habitat track the expanding steppe in Central Europe. GS10 likely caused a repeat of that process in Western Europe.

A depopulation scenario was suggested for GS13-H5 as a trigger for the first intrusion of modern humans into Europe, represented by the Bohunician stone tool assemblage in Moravia (Central Europe) and further east (13, 62). Because a potential cultural hiatus is difficult to prove in the archeologic record of Moravia at given accuracy and precision of applied radiometric dating methods (63), this scenario may not currently be tested. There is also not as strong a climatic evidence in support. Cooling likely was less severe in East-Central Europe compared with subsequent stadials (see above). Severe aridity was apparent between the northern Aegean coastal region and the Levant (13, 64) but not over the Balkan (15), the Adriatic and the Black Sea regions (16, 65), or Central-East Europe.

The Moravian record suggests contemporaneity of modern humans—the Bohunician—and Neanderthals—the Szletian—at close geographic proximity during GI12 and perhaps later but with some chronologic uncertainty (63, 66). During GI11-10, the situation is comparable for the Aurignacian and Mousterian on the upper Danube—at Keilberg-Kirche and Sessefeldgrotte (see above)—and for Aurignacian and Szletian on the middle Danube—at Peš-kô Cave and Szleta Cave (42, 67) (SI Appendix, Fig. S1). Here, however, the interbreeding between species or genetic connections between the Bohunician and the Grotte Mandrin modern human specimen in that time range (5) principally confirms true contemporaneity along the northern fringe of the steppe landscape in the Danube Valley. Nonetheless, in the subsequent ∼6 millennia after GS10, the modern human genome in Europe was replaced twice with a different genetic lineage harboring a much older Neanderthal ancestry (7). One modern European genetic fragment found inside Goyet Cave (Belgium), was attributed to the late Aurignacian and dated to ∼35 ka cal BP—coeval with GI7. This interval of modern human repopulation may have followed the extended cold interval from GS10 to GS9-H4, over 4,000 y long and only interrupted briefly by the 250-y-long GI9 (Fig. 2). A different genetic branch then dominated Europe after ∼34 ka cal BP—coeval to the end of the severe and 1,000-y-long GS7 (Fig. 2). This lineage was attributed to the Gravettian (7), who repopulated Europe during yet another time of low population density (68). The discussion above lays out a general scenario of depopulation–repopulation cycles associated with steppe landscape expansion following extreme or long stadials. The comparable timing of stadials and population changes seen in the archeologic and genetic record suggests that millennial-scale climate cycles may have been the pacemaker for Europe’s demographic history during the Middle Pleniglacial.

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