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The five interglacial periods before the Mid-Brunhes Event (MBE) (c.430 thousand years (ka) ago) are generally considered to be globally cooler than those post-MBE. Inhomogeneities exist regionally, however, which suggest that the Arctic was warmer than present during Marine Isotope Stage (MIS) 15a. Using the first speleothem record for the High Arctic, we investigate the climatic response of northeast Greenland between c.588 and c.549 ka ago. Our results indicate an enhanced warmth of at least +3.5°C relative to the present, leading to permafrost thaw and increased precipitation. We find that δ18O of precipitation was at least 3‰ higher than today and recognize two local cooling events (c.571 and c.594 ka ago) thought to be caused by freshwater forcing. Our results are important for improving understanding of the regional climatic response leading up to the MBE and specifically provide insights into the climatic response of a warmer Arctic.

**INTRODUCTION**

The five interglacial periods before the Mid-Brunhes Event (MBE), which occurred between Marine Isotope Stages (MIS) 13 and 11, are generally considered to be globally cooler than those post-MBE (1–7). In contrast to this global overview, modeling (3) and proxy evidence (8–11) exist at regional scales indicating warmer-than-present climates before the MBE, particularly in the northern mid- and high latitudes, and especially during MIS 15a. MIS 15a, in itself, is of further interest because of its close orbital similarity to the last interglacial MIS 5e (3), which is often used as a baseline in climate studies for a past warmer-than-present climate. Despite the high summer insolation associated with MIS 15a, atmospheric greenhouse gas concentrations (12–15) were relatively low for an interglacial, thus mitigating the effects of the extra warmth (2, 3) provided by insolation, leading to a globally cool climate. Efforts to understand the contribution of CO2 and insolation to interglacial climates showed that despite MIS 15a being globally relatively cool compared to post-MBE interglacials, the regional response was very different, with anomalously high temperatures shown for the Arctic during both summer and winter seasons (3). The regional climatic disparities and their relationship to the global climate leading up to the MBE are poorly understood because of a lack of well-dated paleoclimate records that exist to test these models. Furthermore, given that the Arctic is warming at more than twice the rate of the global average, with temperatures now at their highest since instrumental records began (16), understanding more about how the Arctic responds in a warmer world is of utmost importance. The warm Arctic climate associated with MIS 15a thus potentially offers an important analog for improving our understanding of how the Arctic responds in a warmer world.

Here, we investigate the pre-MBE climate in the High Arctic, specifically during the period MIS 15a-14, using the first speleothem record for Greenland. The record is from the highly sensitive northeast Greenland (17–20), located far from the deep ice cores (Fig. 1) in an ice-free area characterized by extensive deep permafrost. The growth interval of the speleothem provides important constraints on the timing of permafrost thaw related to a locally warmer-than-present climate and enhanced atmospheric hydrological regime. These findings thus confirm modeling results showing that the climate of the Arctic was milder and wetter than today during MIS 15a, despite a globally cool climate, and that warmth continued into the “missing glacial” MIS 14.

**Greenland caves and speleothems**

Solution caves located in Silurian limestones of northeast Greenland (21) (80.38°N, 21.74°E) were first reported in 1960 (Fig. 1) (22). The largely horizontal caves are situated at distinct elevations between c.350 and 670 m above sea level (a.s.l.) in a 1-km-long, north-south trending tributary valley of the larger Grottedal valley (Figs. 1 and 2) (22, 23). The caves reach a maximum of 12 m wide, 100 m long and are all blocked by ice or cave fill. In 1960, the presence of one 10-cm-thick, coarsely crystalline calcite flowstone was reported (520 m a.s.l.) (22).

The caves are located c.35 km from the coast and c.60 km from the Greenland Ice Sheet (GrIS) margin. Today, the region is arid, with an annual precipitation of c.200 mm (24). In the cave region, 50% of the annual precipitation presently occurs on 16 days of the year, with May–June being the driest months and September being the wettest month, although in reality there is limited monthly variability (c.0.35 to 0.81 mm day−1) (24). Winter precipitation sources are centered over the northern (>45°N) North Atlantic and, in particular, the Norwegian Sea. These sources remain in summer but are diminished as a greater contribution is derived from local land and ocean sources (24). The nearest weather stations are situated on the GrIS c.70 km (370 m a.s.l.) and 90 km (870 m a.s.l.) to the southwest of the caves. They recorded a mean annual surface air temperature (MAAT) of −13.2° and −18.4°C, respectively, for the period July 2008 to September 2018 (25). Further afield, MAAT on the coast at Station Nord (180 km north) and Danmarkshavn (400 km south) is c.−18° and −11°C, respectively (26). In contrast, the climate of Grottedal is more continental and warmer during the summer as compared to either the GrIS or coastal stations (27). The Grottedal land area is snow-free during the summer, although the plateau
above the caves (740 m a.s.l.) has previously hosted a small ice cap (22), which is now gone. Patterned ground (fig. S1) containing permafrost and deep ice-wedge depressions exists in the region. Given the contemporary arid climate and frozen ground, as well as the environment largely devoid of soil and vegetation, the modern environmental conditions severely limit karstification and preclude deposition of speleothems, as supported by the absence of drips within the caves and the extensive presence of hoar frost indicating cave air temperatures below freezing. The presence of flowstone thus indicates milder and wetter conditions in the recent geological past.

A 12-cm-thick sample of the flowstone (GD8-1; fig. S2) (22) was collected from a short cave during an exploratory expedition in 2015. For preservation and logistical reasons, the collected sample was not in situ, although a large portion of the sequence was in situ or close to it; thus, it is not expected that GD8-1 traveled far from its original position.

The sampled cave is not officially named by the Greenlandic authorities; hence, it is simply referred to here as GD8, which reflects the cave code it was given during the 2015 Greenland Caves Project (GCP) expedition (28). GD8 cave is 11 m long by 4 m wide by 4 m high (fig. S2). It contains speleogenetic features typical of a phreatic-vadose karst cave including anastomoses on the ceiling, three distinct wall notches indicating former water-table elevations, and scallops that indicate that the direction of flow was away from the present-day entrance and into the cave (23, 28). Similar to all other caves explored in this region, GD8 opens in a steep cliff of the tributary valley, which is formed by fluvioglacial erosion, locally exposing flowstone deposits at the surface. Today’s cave size and morphology as well as the abundance of frost shattering are therefore vastly different from the conditions when the flowstone was deposited. Because of the short length, modern cave air temperature measurements for GD8 offer limited information, as they would simply reflect the weather conditions external to the cave at a given point in time. Some useful information may, however, be obtained from another cave at a similar elevation within the same valley. At 95 m long, U-Tut Ilusilik Qaarusussuaq (U-shaped cave) is the longest cave discovered in this area (23). It is relatively horizontal, “U-like” in shape, and has two entrances that open at an elevation of c.530 m a.s.l. (i.e., at a comparable elevation to GD8 at c.520 m a.s.l.). The middle of the cave, c.48 m from the entrance and found in the

Fig. 1. Greenland cave study area compared to other sites discussed in text and sites of importance. 1: Grottedal caves site (this study); 2 to 10: Sites of ice core drilling (2: Hans Tausen ice cap; 3: Flade Isblink ice cap; 4: East Greenland Ice-core Project (EastGRIP); 5: North Greenland Eemian Ice Drilling (NEEM); 6: North Greenland Ice Core Project (NGRIP); 7: Greenland Ice Core Project (GRIP); 8: Greenland Ice Sheet Project 2 (GISP2); 9: Camp Century; 10: Renland ice cap); 11: Lake El’gygytgyn; 12: Lenskaya Cave; 13: Botovskaya Cave; 14: Ocean Drilling Programme (ODP) 910; 15: ODP 980; 16: Station Nord; 17: Danmarkshavn; 18: Lake Baikal. Figure 1 is constructed from the following datasets: GrIS (69), sea ice extent (70), permafrost (71), and countries (72).
trough of the “U,” is found within the twilight zone and in a location where external air temperatures begin to have a reduced effect on the cave air temperature. Here, a spot temperature measurement of −3.5°C was taken on 6 July 2019. Another horizontal cave within the same area, Inussuk Innartooq Qaarussussuaq (Cairn Climb Cave), located at 540 m a.s.l., but only 19 m in length, also had a comparable air temperature of −3.3°C, c.10 m from the entrance (23) on 7 July 2019. Elsewhere within the wider region, much colder cave air temperatures of −6°C to −17°C have also been recorded, but these are found in caves of higher elevation (up to 825 m a.s.l.) or in gently dipping caves (23) that display cold air-trap characteristics (29, 30). Given that GD8 is truncated and only a relic of the system it was when the flowstone was deposited, it is not possible to say whether it was morphologically horizontal or dipping.

RESULTS

U-Th dating combined with age modeling (figs. S3 and S4 and table S1) indicates that the flowstone was deposited between c.588 and c.537 thousand years (ka) ago. Isotopic measurements are precise despite low...
U concentrations (tens of parts per billion), and corrections for initial Th are negligible. Because the age of the sample is near the limit of the U-Th dating method, uncertainties in age are larger than would be the case for younger samples. Nevertheless, all ages are stratigraphically in order within uncertainty, consistent with closed-system evolution, while two separate slabs taken from the sample yield reproducible results (fig. S4). $\delta^{18}O$ and $\delta^{13}C$ of the speleothem calcite range between c.$-11$ to $-16\%_{\text{VPDB}}$ Vienna Pee Dee Belemnite (VPDB) and $+2$ to $-6\%_{\text{VPDB}}$, respectively, and correlation between $\delta^{18}O$ and $\delta^{13}C$ is minimal (fig. S5), suggesting limited kinetic isotope fractionation (31).

**DISCUSSION**

Results of the U-Th–based age modeling indicate that deposition occurred through MIS 15a and into MIS 14 (Fig. 3A), although it is entirely plausible that growth was not continuous and may have had a stop-start character related, for instance, to the seasonal cycle or extended cold periods on decadal or centennial scales superimposed on a general period of warmth. One might also consider that there may be unidentified hiatuses and that deposition may have occurred during MIS 13; however, the youngest U-Th age from slab 2 indicates that this was not the case (fig. S3). It is also reasonable to question whether or not the speleothem was deposited during a warmer and wetter climate, or possibly beneath a warm-based ice cap. Until recently, a small ice cap was present on the plateau above the caves (22), but despite this, no active drips were reported within the caves, and they remained frozen with air temperatures below 0°C (23). Conditions may have been different in the distant past, but these modern observations indicate that the caves remained in the permafrost when an ice cap was present above. Furthermore, despite
extensive searching, no late Holocene speleothems have been found, which would indicate deposition beneath a warm-based ice cap. Lack of an ice cap, specifically during the MIS 14 period, is also supported by the δ18O profile, which indicates development of a soil at this time (fig. S4). We therefore consider a warmer and wetter climate to be a more probable climate state for the deposition of the speleothem.

The pattern of the δ18O variability yielded by this age model does not exhibit a particularly strong correlation to orbital parameters (Fig. 3B). It is possible, however, to refine the age model within the limits of uncertainties and tune it to 80°N insolation (Fig. 3C) (32). The orbitally refined age model suggests that speleothem deposition occurred between c.622 and c.549 ka ago, i.e., beginning at the start of MIS 15 and ending shortly within MIS 14 (Fig. 3). Naturally, the consequence of this orbital tuning is a better agreement with the δ18O variability in the Asian monsoon composite speleothem record (33) whereby higher δ18O in Greenland corresponds to lower δ18O in Asia and vice versa (Fig. 3G). In contrast, comparison of the U-Th–only age model and the orbitally refined age model with the synthetic Greenland δ18O record (34) shows that both approaches yield δ18O records that are similar in timing and pattern to the Greenland synthetic curve (Fig. 3, D and E). One might even argue that millennial-scale events in the Greenland synthetic record are better correlated with the U-Th–only age model (Fig. 3D). At present, we do not adopt either age model as the best approach because many questions remain open with regard to the relationship between Greenland 80°N speleothem δ18O and 80°N summer insolation. Furthermore, the efficacy of orbital tuning for this particular time period has been shown to be ineffective (35). The remainder of the discussion surrounding the speleothem growth interval will therefore be based around the period c.588 to c.549 ka ago, which is where the two age models overlap.

Regardless of the age model adopted, the speleothem growth interval indicates a warmer and wetter climate as compared to today in northeast Greenland (80°N) during the period of high-amplitude insolation changes leading up to the MBE (Fig. 4). Speleothem deposition was active during MIS 15a to MIS 14, although if the orbitally refined age model is taken into consideration, then deposition may have begun earlier in MIS 15e (Fig. 3C). This period of high-amplitude precessional fluctuations was modulated by the high eccentricity that occurred at the time (35); consequently, the MIS 15a interglacial (1) had the third highest Northern Hemisphere summer insolation maximum of the past million years (fig. S6) (32), an effect that was further compounded at high latitudes by the high obliquity. MIS 15c and the interglacial MIS 15e also experienced relatively high insolation maxima (fig. S6) (32), with MIS 15e generally considered to be of comparable strength to MIS 15a in most records (1).

Both modeling and proxy records indicate that the five interglacial climates before the MBE were generally cooler annually and seasonally on global and hemispheric scales than those post-MBE. The only exception to this is the MIS 15a summer temperature, which models suggest were warmer than present globally and hemispherically, with the Northern Hemisphere summer being the warmest of the past nine interglacials (2, 3). The relatively “cool” annual global climate of MIS 15a is considered to be a consequence of low greenhouse gas concentrations (Fig. 4L) mitigating the effects of high summer insolation (2, 3). The speleothem record presented here, however, provides (the first) radiometrically dated evidence for a warmer and wetter MIS 15a climate in the High Arctic. Further evidence for MIS 15a being a warm interglacial, especially in the high northern latitudes, is further provided by U-series–dated speleothem growth periods in eastern Siberia at 60°N, which indicate permafrost absence (Fig. 4G) (36). At this time, the orbitally tuned Lake El’gygytgyn (67.5°N) sediment cores also show an increase in planktonic concentration and diversity (Fig. 4F) (10), whereas Lake Baikal (53°N) diatom productivity was also very high (Fig. 4I) (9). In the case of Lake El’gygytgyn and Lake Baikal, these increased levels of productivity also partly continue into MIS 14, similar to our speleothem growth period. In contrast, pollen records from south Greenland, which are used as an indicator of reduced ice volume, are not particularly responsive at this time in comparison to other interglacials, although they are approximately 3.5 times higher than for the Holocene (Fig. 4H) (37), indicating a greater ice-free area than is found at present. MIS 9e, which is shown in some records to be one of the warmest interglacials of the past 800,000 years (1), also does not show a particularly strong pollen response in this record (Fig. 4H), although, in contrast, sea surface temperature (SST) records from a nearby core indicate that MIS 9e was the warmest of the previous four interglacials with summer SST 3.3°C above modern values and a reduced size of the GrIS (38). By inference, we therefore would not necessarily expect to see a pollen response in south Greenland for MIS 15a.

Elevated MIS 15a temperatures within the High Arctic, leading to permafrost thawing, an enhanced hydrological cycle, incipient soil development, karstification, and local deposition of speleothems, were likely forced by the high summer insolation, which models show to be the dominant factor controlling interglacial temperature at high Northern Hemisphere latitudes (3). In the Arctic, high MIS 15a obliquity resulted in further enhancement of summer solar radiation in comparison to the rest of the planet. Warm intense summers in the Arctic subsequently led to the “summer remnant effect” (3) whereby summer ocean warming enhanced melting of sea ice and increased warming in the upper ocean, thus delaying formation of sea ice and reducing its thickness and thermal insulation effects the following winter. Consequently, heat was lost from the ocean to the atmosphere in early winter, resulting in an increase in winter atmospheric temperatures, specifically in the Arctic during MIS 15a, despite the low winter insolation (3). The increased temperature from insolation in the summer and increased temperature from the summer remnant effect in the winter thus provide a mechanism by which both summer and winter temperatures were increased, raising MAAT and leading to permafrost thawing in northeast Greenland. The lack of Arctic sea ice during MIS 15a, which is highly important for increasing temperature in the winter months (3), is also supported by proxy data from cores in the Arctic (Fig. 4E) (39), although these point to an extended sea ice–free period before the MBE and not one directly as a result of high insolation. The deposition of speleothem at 80°N in northeast Greenland, which requires a higher MAAT than at present to thaw the permafrost, thus supports model results that show Arctic MIS 15a summer, and winter temperatures were higher than at present (although annually globally they were cooler).

In addition to the increased temperature, an increase in precipitation must have also occurred in this presently arid region. Correlation between atmospheric dust records from Greenland and Antarctica is considered by some to reflect the state of the global hydrological cycle and global atmosphere (35). In comparison to the present, atmospheric dust content in Antarctica was lower during MIS 15a (Fig. 4D) (40), thus suggesting an enhanced global hydrological
Fig. 4. Greenland speleothem $\delta^{18}O$ record compared with other paleoclimate archives for the period 0 to 630 ka ago. (A) Obliquity (dark blue) (73) and precession (light blue) (73). (B) The 65°N summer insolation (73). (C) NGRIP Greenland $\delta^{18}O$ record (dark yellow) (74). U-Th–based (dark blue) and orbitally refined (light blue) age models for 80.2°N Greenland speleothem $\delta^{18}O$ record (this study). (D) Synthetic Greenland $\delta^{18}O$ record (light blue) (34). EPICA Dome C (EDC) atmospheric dust (dark blue) (40). (E) Arctic Ocean Polycope abundance (39). (F) Lake El’gygytgyn plankton diversity index (10). (G) Eastern Siberian speleothem ages (36). (H) South Greenland marine pollen record (37). Solid yellow horizontal bar indicates Holocene baseline. Dashed yellow horizontal bar indicates MIS 15a baseline. (I) Chinese loess grain size >32 μm: Yimaguan (light blue) and Luochuan (dark blue) (43). (K) Asian monsoon composite speleothem record (33). (L) Atmospheric CO$_2$ (dark blue) (12–14) and CH$_4$ (light blue) (12, 13, 15) concentrations. (M) Global benthic $\delta^{18}O$ stack (dark blue) (41). Stacked global SST record (light blue) (44). Vertical gray bar highlights area of overlapping speleothem age models, with hatched vertical bars indicating only single age models. Vertical yellow line marks MIS 15a-14 boundary.
cycle at this time. In agreement, models indicate the largest increase in Northern Hemisphere summer monsoon precipitation during MIS 15a over the past 800 ka controlled by precessional effects at low latitudes (3). The moisture needed for the enhanced hydrological regime in the Arctic, which would have been needed for the speleothem growth, was likely sourced, in part, from an increase in long-range transport but more dominantly from the local ice-free open ocean, as has been predicted for a future warmer Arctic (19). Even under present conditions, moisture sources switch from long-range transport in the winter months to local ice-free waters and continental recycling in the summer months (24). It is thus expected that during MIS 15a, higher temperatures and a greater ice-free local ocean would have resulted in a higher contribution of moisture from local sources.

The continued speleothem deposition into MIS 14 is perhaps more perplexing than the growth during MIS 15a. MIS 14 is, however, considered a “missing glaciation,” suggested to be a consequence of the preceding high-amplitude Northern Hemisphere insolation changes that would have mitigated glacial inception and prevented substantial ice expansion in the Northern Hemisphere (35). This is indicated by a lack of terrestrial evidence for Northern Hemisphere glacial expansion and is also reflected in the marine record, which displays the lowest benthic foraminifera δ18O of the past 10 cold stages (Fig. 4M) (35, 41), indicating lower global ice volume. Sea levels during MIS 14 were also relatively high (~67 m) (42), and perennial Arctic sea ice remained absent (Fig. 4E) (39). For the period that is coincident with the speleothem deposition, benthic foraminifera δ18O were increasing and had not yet reached their maximum (41), pollen assemblages in south Greenland remained as high or higher than the Holocene (Fig. 4H) (37), and Lake El’gygytgyn was marginally productive (Fig. 4I) (43), whereas Lake Baikal became increasingly less productive (Fig. 4H) (42), and perennial Arctic sea ice remained absent (Fig. 4E) (39). For the period that is coincident with the speleothem deposition, benthic foraminifera δ18O were increasing and had not yet reached their maximum (41), pollen assemblages in south Greenland remained as high or higher than the Holocene (Fig. 4H) (37), and Lake El’gygytgyn was marginally productive (Fig. 4I) (43), whereas Lake Baikal became increasingly less productive (Fig. 4H) (42), and perennial Arctic sea ice remained absent (Fig. 4E) (39). 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For the period that is coincident with the speleothem deposition, benthic foraminifera δ18O were increasing and had not yet reached their maximum (41), pollen assemblages in south Greenland remained as high or higher than the Holocene (Fig. 4H) (37), and Lake El’gygytgyn was marginally productive (Fig. 4I) (43), whereas Lake Baikal became increasingly less productive (Fig. 4H) (42), and perennial Arctic sea ice remained absent (Fig. 4E) (39).
Fig. 5. Greenland speleothem δ¹⁸O record compared to marine records for the period 500 to 630 ka ago. (A) U-Th–based (dark blue) and (B) orbitally refined (light blue) age models for 80.2°N Greenland speleothem δ¹⁸O record (this study). (C) ODP 910 80°N Arctic Gateway δ¹⁸O planktic foraminifera record. Arrows indicate freshwater pulses (59). (D) ODP 984 63°N ice-rafted debris (light blue) and *N. pachyderma* (s.) abundance (dark blue) (45). (E) ODP U1314 56°N SST (dark blue) and benthic δ¹³C (light blue) (46). (F) ODP 980 55°N ice-rafted debris (light blue) and *N. pachyderma* (s.) abundance (dark blue) (45). (G) ODP 980 55°N benthic foraminifera δ¹⁸O (dark blue). Summer SST (light blue) (45). (H) Global benthic δ¹⁸O stack (dark blue) (41). Stacked global sea surface temperature record (light blue) (44). Vertical gray bar highlights the area of overlapping speleothem age models, with hatched vertical bars indicating only single age models. Vertical yellow line marks the MIS 15a-14 boundary.
under investigation, closer proximity to the source, and lower elevation. Because the glacial-interglacial isostatic range is only c.30 m (53), it is not expected that glacio-isostatic adjustments had a large impact on $\delta^{18}O$ of the recharge water. As a point of comparison, $\delta^{18}O$ values from the nearby Flade Isblink ice cap (81.3°N, 15.7°W, c.700 m a.s.l.) vary between $−21‰_{VSMOW}$ during the Medieval Climate Anomaly and $−24‰_{VSMOW}$ during the Little Ice Age (54). On the basis of the speleothem $\delta^{18}O$ values of −11 to $−16‰_{VPDB}$, we constrain the $\delta^{18}O$ range of the infiltrating water at the cave site during MIS 15a-14 as c.$−14‰_{VSMOW}$ (for $−11‰_{VPDB}$ calcite) to c.$−19‰_{VSMOW}$ (for $−16‰_{VPDB}$ calcite), assuming a minimum temperature of 1°C (55–57). If temperatures were higher, then the recharge water would need to have been isotopically heavier to produce $−11‰_{VPDB}$ calcite (55–57), thus constraining these as minimum values for the infiltration water. In comparison, congelation ice today within the Grottedal caves varies from $−17$ to $−25‰_{VSMOW}$ (Fig. 8 and table S2), which is in agreement with an annual mean modeled $\delta^{18}O$ of precipitation of c.$−20‰_{VSMOW}$ for the site (58). Infiltration water with $\delta^{18}O_{VSMOW}$ at least c.3‰ heavier than today would therefore be needed to form the $−11‰_{VPDB}$ calcite. We stress, however, that 3‰ $_{VSMOW}$ is the
minimum difference in comparison to today given that temperatures could have been more than 1°C, and that the range in δ18O of modern precipitation extends down to −20 to −25‰. Today, the δ18O temperature gradient for the region is modeled at c.0.9 to 1.3‰ °C−1 (58); thus, while it is not known whether this relationship was the same during MIS 15a-14, it highlights the possibility that a MAAT change of 3.5°C could account for a c.3‰ increase in δ18O of mean annual precipitation. A higher increase in δ18O of precipitation could, of course, be achieved through higher temperatures or by other additional mechanisms. As mentioned, even under present conditions, moisture sources switch from long-range transport in the winter months to local ice-free waters and continental recycling in the summer months (24). Predictions for a future warmer Arctic indicate an increase in precipitation (19, 20) and especially rain in all areas of the Arctic except the interior of the GrIS (20). By the end of this century, the greatest changes in Arctic precipitation are expected over the Arctic Ocean and northeast Greenland and, specifically, in the late autumn-winter months associated with evaporation from ice-free Arctic waters (19). In a warmer MIS 15a-14 Arctic climate in which sea ice was absent (Fig. 4E) (39), a much higher contribution of precipitation was likely sourced locally, thus reducing Rayleigh fractionation, leading to a further increase in δ18O of precipitation. Seasonality changes may have added additional controls, but given that the increase in precipitation is expected to be more dominant in autumn-winter months (when δ18O would be less than for summer months), it is thought that the temperature difference and more local moisture source were likely the main factors affecting the increased δ18O of precipitation during MIS 15a-14.

Given the large uncertainties in the age model, we do not provide a full interpretation of the speleothem δ18O signal but do consider the effects of two freshwater events through the Arctic Gateway (Fig. 5C) (59) on the local climate. The largest of these freshwater events, centered on c.571 ka ago, occurred within age-model uncertainties with a distinctive depletion in δ18O in both models (Fig. 5, A and B), making it likely that this freshwater event had a cooling effect on the local climate. Of interest here is the decoupling between the surface temperatures of the Arctic Gateway and subpolar North Atlantic, which is indicated by the depletion in planktic δ18O in the Arctic Gateway (Fig. 5C) versus a continued low abundance of subpolar North Atlantic polar foraminifera (Fig. 5, D and F) and unchanged SSTs (Fig. 5, E and G). Another smaller freshwater event centered on c.594 ka ago corresponds timewise also to a small depletion in speleothem δ18O in the orbitally refined age model. As for the larger freshwater event, the subpolar North Atlantic records do not respond.

In conclusion, over the past 800,000 years, the five interglacials (MIS 13a-19c) before the MBE are considered to be globally cooler than the six interglacials post-MBE (1). Despite this, modeling results suggest that a period of warmth, higher than at present, existed in the Arctic during MIS 15a as a result of the exceptionally high summer insolation at that time (2, 3). The growth of speleothem during MIS 15a in northeast Greenland, thus requiring the percolation of precipitation into the cave and therefore a warmer and wetter climate in comparison to today, supports the modeling results and the notion that the Arctic can be anomalously warm, even during a period of relatively mild global climate. Such findings do not, at this stage, preclude speleothem deposition during other interglacials, either pre- or post-MBE, and do not limit the driving force for increased warmth to high summer insolation only. Of importance with regard to the results presented here is the novel, first Arctic speleothem record, which has been used as an important test of model results, confirming a warmer and wetter Arctic during MIS 15a and therefore regional heterogeneities with respect to the global climate state. Furthermore, the continued growth of the speleothem into MIS 14 extends the spatial influence of the lagging warmth that existed during the start of this “missing glacial” beyond the eastern North Atlantic.
The growth of speleothem during MIS 15a-14 in northeast Greenland indicating a warmer and wetter climate was associated with high Northern Hemisphere summer insolation and relatively low atmospheric greenhouse gas concentrations. These boundary conditions, which were more comparable to those for MIS 5e, are different to what is expected in the future, i.e., comparatively low Northern Hemisphere summer insolation and high greenhouse gas concentrations. Nevertheless, the results presented here further highlight the sensitivity of the Arctic and, in particular, northeast Greenland to changing boundary conditions and climate states. As is predicted for the future (18–20), warmer climate states in the past have resulted in a wetter climate for which the moisture was likely sourced more locally from an open ice-free ocean.

**MATERIALS AND METHODS**

**U-Th dating**

Sample GD8-1 was collected ex situ from GD8 cave in northeast Greenland. The sample was cut in half, and two slabs were taken as a test of reproducibility. Both slabs were polished to remove cut marks and improve the finish. Subsamples for U-Th dating were drilled using a handheld drill in a thoroughly cleaned laminar flow hood used exclusively for low U concentration carbonate samples. Because of the low uranium concentration and possibility of leaching, this method was preferred to cutting small chips and cleaning in an ultrasonic bath. Typical sample sizes were 300 mg. Extraction and purification of U and Th followed standard chemistry procedures (60) and were performed at the Isotope Geochemistry laboratory at the University of Minnesota. Samples were spiked with a dilute mixed \(^{228}\)Th-\(^{233}\)U-\(^{238}\)U tracer to allow for correction of instrumental fractionation and calculation of U and Th concentrations and ratios. Measurements were conducted on a Thermo-Finnigan Neptune multi-collector inductively coupled plasma mass spectrometer (61).

**Stable isotopes of calcite**

Stable isotopes (\(\delta^{18}\)O\(_{\text{calc}}\) and \(\delta^{13}\)C\(_{\text{calc}}\)) were micro-milled at a spatial resolution of 150 \(\mu\)m (slab 1) and 200 \(\mu\)m (slab 2). Measurements were conducted at the University of Innsbruck on a Thermo Fisher Scientific DeltaPlusXL isotope ratio mass spectrometer linked to GasBench II and measured relative to the NBS19 standard. Analytical precisions of the standard are 0.08 and 0.06‰ for \(\delta^{18}\)O\(_{\text{calc}}\) and \(\delta^{15}\)C\(_{\text{calc}}\), respectively (1σ) (62).

**Reproducibility**

Given that a second speleothem sample was not available for reproducibility testing, we opted to slice two slabs from GD8-1 as a next-best step. Slab 1 was treated as the master slab from which the highest-resolution samples for stable isotopes and dating were taken. Slab 2 was treated as a basic test of slab 1 with lower-resolution spatial analysis. Within uncertainty, slab 2 reliably reproduces the results of slab 1 (fig. S3).

**Secular equilibrium**

Four samples yielding ages at or close to secular equilibrium were obtained for slab 1. Two of these were finite ages within uncertainty of the remainder of the age model. The other two yielded secular equilibrium ages within uncertainty. Higher-precision measurement uncertainties on these two samples that do not intersect secular equilibrium would still result in ages with very large uncertainties that intersect the age model.

**Age model**

Age models were constructed using the Monte Carlo approach in StalAge (63).

**Climate emulator**

The climate emulator (47) was trained using GCM snapshot simulations run using the UK Met Office Hadley Centre Climate Model (HadCM3) (64) with varying orbital parameters, atmospheric CO\(_2\) concentration, and ice sheets, including a set covering the last glacial cycle (51). The emulator was forced at 1-ka resolution by orbital changes (65), a composite record of observed paleo-CO\(_2\) (66), and a record of global sea level reconstructed using a one-dimensional ice sheet–climate model (67). The resulting series of global temperature and precipitation fields were then combined to provide a “continuous” climate simulation, and the climate at the grid box in which the cave is located, and the surrounding region, was examined.

**Stable isotopes of surface precipitation**

During the three expeditions of the GCP (2015 and 2019: Centrumso/Grottedal region, 80.2°N; 2018: Wegener Halvo, 71.7°N) (68), surface waters and cave ice were sampled for analysis of oxygen and hydrogen isotopic composition. At present, the nearest monitoring of isotopes in precipitation takes place at Station Nord, some 180 km to the north and located on the coast. Spot sampling of surface waters and cave ice at the GCP field sites thus provides useful insights into the stable isotopic composition of meteoric precipitation in this region.

In total, 28 samples were collected in 2-ml volume glass vials with plastic screw cap. Where possible, vials were filled completely to the top so that no air bubbles remained. As far as was possible, samples were stored in a cool environment or refrigerated at c.4°C until analysis.

Samples were analyzed for isotopic composition using a Picarro L2140-i CRDS instrument. Analytical precision is better than 0.1 and 0.05‰ for \(\delta^{18}\)O and \(\delta\)D (1σ), respectively. Standardization was accomplished using VSMOW, Greenland Ice Sheet Project 2 (GISP2), and Standard Light Antarctic Precipitation (SLAP) reference solutions. Results are shown in Fig. 8 and table S2.

**SUPPLEMENTARY MATERIALS**

Supplementary material for this article is available at http://advances.sciencemag.org/cgi/content/full/7/13/eabe1260/DC1

**REFERENCES AND NOTES**

1. Past Interglacials Working Group of PAGES, Interglacials of the last 800,000 years. *Rev. Geophys.* **54**, 162–219 (2016).
2. Q. Z. Yin, A. Berger, Insolation and CO\(_2\) contribution to the interglacial climate before and after the Mid-Brunhes Event. *Nat. Geosci.* **3**, 243–246 (2010).
3. Q. Z. Yin, A. Berger, Individual contribution of insolation and CO\(_2\) to the interglacial climates of the past 800,000 years. *Climate Dynam.* **38**, 709–724 (2012).
4. J. H. F. Jansen, A. Kuipjens, S. R. Troelstra, A Mid-Brunhes climatic event: Long term changes in global atmosphere and ocean circulation. *Science* **232**, 619–622 (1986).
5. P. Wang, J. Tian, X. Cheng, C. Liu, J. Xu, Carbon reservoir changes preceded major ice-sheet expansion at the Mid-Brunhes Event. *Geology* **31**, 239–242 (2003).
6. P. C. Tzedakis, D. Raynouard, J. F. McManus, A. Berger, V. Brovkin, T. Kiefer, Interglacial diversity. *Nat. Geosci.* **2**, 751–755 (2009).
7. EPICA Community Members, Eight glacial cycles from an Antarctic ice core. *Nature* **429**, 623–628 (2004).
8. I. Candy, G. R. Coope, J. D. Lee, S. A. Parfitt, R. C. Preece, J. Rose, D. C. Schreve, Pronounced warmth during early Middle Pleistocene interglacials: Investigating the Mid-Brunhes Event in the British terrestrial sequence. *Earth Sci. Rev.* **103**, 183–196 (2010).
10. A. Prokopenko, L. A. Hinnov, D. F. Williams, M. L. Kuzmin, Orbital forcing of continental climate during the Pleistocene: A completely astronomically tuned climatic record from Lake Baikal, SE Siberia. Quat. Sci. Rev. 25, 3431–3457 (2006).
11. J. A. Snyder, M. V. Cherepanova, A. Bryan, Dynamic diatom response to changing climates of 0.1-2 Ma at Lake Efuyogyryg, Far East Russian Arctic. Clim. Past 9, 1309–1319 (2013).

12. E. Willerslev, E. Cappellini, W. Boomstra, R. Nielsen, M. B. Høgsbørg, T. B. Brand, M. Hofreiter, M. Bunce, H. N. Poinar, D. Dahl-Jensen, S. Johnsen, J. P. Steffensen, O. Bennekie, J.-L. Schwenninger, R. Nathan, S. Armitage, C.-J. de Hoog, V. Affmno, M. Christ, J. Beer, R. Muscheler, J. Barker, M. Sharp, K. E. H. Penkman, J. Haile, P. Taberlet, T. M. P. Gilbert, A. Cassoli, E. Campari, M. J. Collins, Ancient biomolecules from deep ice cores reveal a forested southern Greenland. Science 317, 111–114 (2007).

13. D. Veres, L. Bazin, A. Landais, H. T. M. Keel, B. Lemieux-Dudon, F. Parrenin, P. Martinerie, E. Blayo, T. Blunier, E. Capron, J. Chappellaz, S. O. Rasmussen, M. Severi, A. Svensson, B. Vinther, E. Wolff. The Antarctic ice core chronology (AICC2012): An optimized multi-parameter and multi-site dating approach for the last 120 thousand years. Clim. Past 9, 1733–1748 (2013).

14. L. Bazin, A. Landais, B. Lemieux-Dudon, H. T. M. Keel, D. Veres, F. Parrenin, P. Martinerie, C. Ritz, E. Capron, V. Lipenkov, M.-F. Loutre, D. Raynaud, B. Vinther, A. Svensson, S. O. Rasmussen, M. Severi, T. Blunier, M. Leuenberger, H. Fischer, V. Masson-Delmotte, J. Chappellaz, E. Wolff, An optimized multi-proxy, multi-site Antarctic ice and gas orbital chronology (AICC2012): 120–800 ka. Clim. Past 9, 1715–1731 (2013).

15. D. Lüthi, M. L. Floch, B. Bereiter, T. Blunier, J-M. Barnola, U. Siegenthaler, D. Raynaud, J. Jouzel, H. Fischer, K. Kawamura, T. F. Stocker, High-resolution carbon dioxide concentration record 650 000–800 000 years before present. Nature 453, 379–382 (2008).

16. L. Loulergue, A. Schilt, R. Spahn, V. Masson-Delmotte, T. Blunier, B. Lemieux, J-M. Barnola, D. Raynaud, T. F. Stocker, J. Chappellaz, Orbital and millennial-scale features of atmospheric CH4 over the past 800 000 years. Nature 533, 383–386 (2008).

17. Arctic Monitoring and Assessment Programme (AMAP). Snow, Water, Ice and Permafrost. North of 60° (IAEA/WMO, 2013).

18. R. M. Spratt, L. E. Liesiecki, A Late Pleistocene sea level stack. Clim. Past 12, 1079–1092 (2016).

19. Q. Hao, L. Wang, F. Oldfield, G. Zhengtang, Extra-long interglacial in Northern Hemisphere during MISs 15-13 arising from limited extent of Arctic ice sheets in glacial MIS. 14. Sci. Rep. 5, 12103 (2015).

20. J. D. Shukun, D. W. Lea, L. E. Liesiecki, M. E. Rayno, An 800-kyr record of global surface ocean δ18O and implications for ice volume-temperature coupling. Earth Planet. Sci. Lett. 426, 58–68 (2015).

21. M. Alonso-Garcia, F. J. Siemi, M. Kucena, J. A. Flores, I. Casco, N. Andersen, Ocean circulation, ice sheet growth and interhemispheric coupling of millennial climate variability during the last 120 kyr. Quat. Sci. Rev. 33, 3234–3247 (2011).

22. R. W. Petit, L. Carriera, P. Thibaudau, Topoclimatic zones and ice dynamics in the caves of the northern Yukon, Canada. Arctic 41, 215–220 (1988).

23. N. S. Lord, M. Cruccifix, D. J. Lunt, M. C. Thorne, N. Bounceur, H. Dowsett, C. L. O’Brien, A. Ridgwell, Emulation of long-term changes in global climate: Application to the late Pliocene and future. Clim. Past 13, 1539–1571 (2017).

24. B. Lauriol, L. Carrier, P. Thibaudau, Topoclimatic zones and ice dynamics in the caves of the northern Yukon, Canada. Arctic 41, 215–220 (1988).

25. A. Landais, V. Masson-Delmotte, E. Capron, P. M. Langenbrog, P. Bakker, E. J. Stone, N. Merz, C. R. Raible, H. Fischer, A. Orsi, F. Prie, B. Vinther, D. Dahl-Jensen, How warm was Greenland during the last interglacial period? Clim. Past 12, 1933–1946 (2016).

26. J. Peltier, Global glacial isostasy and sea-level observations and glacial-rebound models. Quat. Sci. Rev. 23, 1053–1077 (2004).

27. A. Lemark, thesis, Niels Bohr Institute Copenhagen University (2010).

28. B. Knudsen, Equilibrium and nonequilibrium isotope fractionation factors of geochemical interest, in Data of Geochemistry, M. Fleischer Ed. (U. S. Geol. Survey Rep. Paper 440-K, ed. 6, 1977), pp.1–12.

29. T.-e. Kim, J. R. O’Neil, Equilibrium and non equilibrium oxygen isotope effects in synthetic carbonates. Geochim. Cosmochim. Acta 61, 3641–3647 (1997).

30. F. D. Froelich, Y. Wang, Speleothem calcite formed in situ: Modern calibration of δ18O and δ13C paleoclimatic proxies in a continuously-monitored natural cave system. Geochim. Cosmochim. Acta 75, 4929–4950 (2011).

31. J. Spötl, G. Hoffmann, S. J. Johnsen, B. M. Vinther, V. Masson-Delmotte, C. Sturm, Modelling the water isotopes in Greenland precipitation 1959-2001 with the meso-scale model REMO-ispo. J. Geophys. Res. 116, D18105 (2011).

32. J. Knie, J. Matthiessen, A. Mackensen, R. Stein, C. Vogt, T. Frederichs, S.-I. Nam, Effects of Arctic freshwater forcing on thermohaline circulation during the Pleistocene. Geology 35, 1075–1078 (2007).

33. R. L. Edwards, J. H. Chen, G. J. Wasserburg, 238U–234U–230Th–232Th systemsatics and the precise measurement of time over the past 500,000 years. Earth Planet. Sci. Lett. 81, 175–192 (1987).

34. C.-C. Chen, C.-C. Wu, H. Cheng, R. L. Edwards, Y.-T. Hsieh, S. Gallet, C.-C. Chang, T.-Y. Li, D. D. Lam, A. Kano, M. Hori, C. Spötl, High-precision and high resolution carbonate δ18O dating by MC-ICP-MS with SEM protocols. Geochim. Cosmochim. Acta 99, 71–86 (2012).
C. Spötl, Long-term performance of the Gasbench isotope ratio mass spectrometry system for the stable isotope analysis of carbonate microsamples. *Rapid Commun. Mass Spectrom.* **25**, 1683–1685 (2011).

D. Scholz, D. L. Hoffmann, StalAge—An algorithm designed for construction of speleothem age models. *Quat. Geochron.* **6**, 369–382 (2011).

P. J. Valdes, E. Armstrong, M. P. S. Badger, C. D. Bradshaw, F. Bragg, M. Crucifix, T. Davies-Barnard, J. J. Day, A. Farnsworth, C. Gordon, P. O. Hopcroft, A. T. Kennedy, N. S. Lord, D. J. Lunt, A. Marzocchi, L. M. Parry, V. Pope, W. H. G. Roberts, E. J. Stone, G. J. L. Tourte, J. H. T. Williams, The BRIDGE HadCM3 family of climate models: HadCM3@Bristol v1.0. *Geosci. Model Dev.* **10**, 3715–3743 (2017).

J. Laskar, P. Robutel, F. Joutel, M. Gastineau, A. C. M. Correia, B. Levrard, A long-term numerical solution for the insolation quantities of the Earth. *Astronomy Astrophysics* **428**, 261–285 (2004).

B. Bereiter, S. Egleston, J. Schmitt, C. Nehrbass-Ahles, T. F. Stocker, H. Fischer, S. Kipfstuhl, J. Chappellaz, Revision of the EPICA Dome C CO2 record from 800 to 600 kyr before present. *Geophys. Res. Lett.* **42**, 542–549 (2015).

L. B. Stap, R. S. W. Van De Wal, B. De Boer, R. Bintanja, L. J. Lourens, The influence of ice sheets on temperature during the past 38 million years inferred from a one dimensional ice sheet-climate model. *Clim. Past* **13**, 1243–1257 (2017).

G. E. Moseley, History of exploration in northeast Greenland. *Cave Karst Sci.* **42**, 55–59 (2020).

L. Joughin, B. Smith, I. Howat, T. Moon, T. Scambos, A SAR record of early 21st century change in Greenland. *J. Glaciol.* **62**, 62–71 (2016).

U.S. National Ice Center and National Snow and Ice Data Center. Compiled by F. Fetterer, M. Savoie, S. Helfrich, P. Clemente-Colón. 2010, updated daily. “Multisensor Analyzed Sea Ice Extent—Northern Hemisphere (MASIE-NH, Version 1)” (2002).

J. Obu, S. Westermann, A. Kääb, A. Bartsch, Ground Temperature Map, 2000–2016, Northern Hemisphere—Alfred Wegener Institute, Helmholtz Centre for Polar and Marine Research, 2018.

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A. Berger, M. F. Loutre, Insolation values for the climate of the last 10 million years. *Quat. Sci. Rev.* **10**, 297–317 (1991).

E. W. Wolff, J. Chappellaz, T. Blunier, S. O. Rasmussen, A. Svensson, Millennial-scale variability during the last glacial: The ice core record. *Quat. Sci. Rev.* **29**, 2828–2838 (2010).

H. Cheng, R. L. Edwards, C.-C. Shen, V. J. Polyak, Y. Asmerom, J. Woodhead, J. Hellstrom, Y. Wang, X. Kong, G. Spötl, X. Wang, E. Calvin Alexander Jr., Improvements in U-Th dating, 238U and 234U half-life values, and U-Th isotopic measurements by multi-collector inductively coupled plasma mass spectrometry. *Earth Planet. Sci. Lett.* **371–372**, 82–91 (2013).

A. H. Jaffey, K. F. Flynn, L. E. Glendenin, W. C. Bently, A. M. Essling, Precision measurement of half-lives and specific activities of 234U and 238U. *Phys. Rev. C* **4**, 1889 (1979).

K. H. Wedepohl, The composition of the continental crust. *Geochim. Cosmochim. Acta* **59**, 1217–1232 (1995).

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**Data and materials availability**: All data needed to evaluate the conclusions in the paper are present in the paper and/or the Supplementary Materials. Additional data related to this paper may be requested from the authors. This product includes Intellectual Property from European National Mapping and Cadastral Authorities and is licensed on behalf of these by EuroGeographics. Original product is freely available at eurogeoarchive.sr.org.uk. Terms of the license are available at http://eurogeoarchive.sr.org.uk/form/topographic-data-eurogeographics. Short form—© EuroGeographics. Original product is freely available at eurogeoarchive.sr.org.uk. Terms of the license are available at http://eurogeoarchive.sr.org.uk/form/topographic-data-eurogeographics. This product includes Intellectual Property from European National Mapping and Cadastral Authorities and is licensed on behalf of these by EuroGeographics. Original product is freely available at eurogeoarchive.sr.org.uk. Terms of the license are available at http://eurogeoarchive.sr.org.uk/form/topographic-data-eurogeographics. Submitted 1 August 2020

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