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Drivers of the evolution and amplitude of African Humid Periods

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Abstract

During orbital precession minima, the Sahara was humid and hosted tropical plant species thus providing a corridor for Hominins migration. Uncertainties remain over the climatic processes controlling the initiation, demise and amplitude of these African Humid Periods (AHPs). Here we present transient simulations of the penultimate deglaciation and Last Interglacial period (LIG), and compare them to transient simulations of the last deglaciation and Holocene. We find that the strengthening of the Atlantic Meridional Overturning Circ-
culation (AMOC) at the end of the deglacial millennial-scale events exerts a dominant control on the abrupt initiation of AHPs, as the AMOC modulates the position of the Intertropical Convergence Zone (ITCZ). In addition, residual Northern Hemispheric (NH) ice-sheets can delay the AHP peak. Through its impact on NH ice-sheets disintegration and thus AMOC variations, the larger rate of insolation increase during the penultimate compared to the last deglaciation can explain the earlier and more abrupt LIG AHP onset. Finally, we show that the background climate state modulates precipitation variability with higher variability under wetter background conditions.

Introduction

The climate of tropical North Africa is under the influence of the West African monsoon (WAM), which is characterised by a low level southwesterly flow bringing moist air from the equatorial Atlantic towards the Saharan heat low during boreal summer (1). The WAM is associated with the position of the ITCZ, which lies at the convergence of the southwesterly monsoon flow and the dry northeasterly Harmattan winds. The ITCZ, commonly identified as the latitude of maximum precipitation, is located at the energy flux equator (2), and its position is thus linked to the atmospheric energy transport and the meridional temperature gradient. The African Easterly Jet (AEJ), which is linked to the meridional surface temperature gradient over North Africa (3), has also been shown to modulate precipitation in the Sahel (14°N-18°N), with a weak and northward displaced AEJ leading to wet conditions (1, 4). Variations in Earth’s incoming solar radiation can thus impact the position of the ITCZ, the AEJ and hence the WAM.

As a result of the low precession and associated high boreal summer insolation prevailing during the early Holocene, the ITCZ was likely located further north during summer, thus leading to wetter conditions over the Sahel and Sahara. During this ’African Humid Period’ (AHP, ~11 - 5.5 thousand years before present, thereafter ka) the Sahara also hosted tropical plant
species (5), and an extensive network of drainage channels and lakes (6).

The AHP was not restricted to the Holocene, but occurred during previous periods of high boreal summer insolation, in phase with precession minima (7–11). During the LIG (∼129-116 ka), the warmest interglacial period of the last 800 ka (12), summer insolation at high northern latitudes was more than 70 W/m² higher than during the pre-industrial (PI). As a result, North Atlantic marine records suggest that summer sea surface temperatures (SSTs) were 1.1 to 1.9°C higher than PI (13, 14). The greater summer insolation in NH at the LIG most likely shifted the ITCZ northward, thus leading to wetter conditions over North Africa (15). Paleo-proxy records from North Africa indeed suggest wetter conditions and a likely northward expansion of trees between ∼127 and 122 ka (16–18). Deposition of organic-rich layers in the Mediterranean Sea (‘sapropels’), also suggest increased monsoon run-off from North Africa during the LIG (7, 19, 20).

Time slice numerical simulations of the mid-Holocene (6 ka) and LIG (127 ka) consistently highlight a larger areal extent and stronger WAM during these time periods compared to PI (15), although these simulations could underestimate both the amplitude and extent of the Holocene AHP, because they did not account for vegetation-climate feedbacks (21, 22). While the time evolution of precipitation over North Africa during the last glacial-interglacial cycle was previously simulated (11, 23), millennial-scale variability was not included and the processes leading to the onset and demise of the AHPs during the last two interglacials were not studied in detail. In addition, sapropel records covering the last 150 ka have shown that there is no constant relationship between insolation maxima and sapropel mid-points, suggesting that the timing of sapropel depositions, and thus AHPs, could also be linked to ice-volume/meltwater changes occurring during the preceding deglaciation (20).

Both the penultimate (∼140-129 ka) and last (∼18-10 ka) deglaciations featured millennial-scale climatic events, during which the AMOC weakened significantly: Heinrich stadial 11
(HS11, main phase at \(\sim 133.3 - 128.5\) ka), Heinrich stadial 1 (\(\sim 18 - 14.7\) ka) and the Younger Dryas (\(\sim 12.8 - 11.7\) ka) (24–26). Through its modulation of meridional ocean heat transport, the strength of the AMOC can impact the location of the ITCZ and thus precipitation patterns over North Africa (27–29). The impact of deglacial changes in insolation and follow-on processes in the climate system on the AHP initiation thus needs to be better constrained.

Finally, paleo-proxy records suggest that the AHP termination during the mid-Holocene was locally abrupt in contrast to the slow insolation decrease (22, 30). As such, it has been hypothesised that the abrupt AHP demise could have arisen from vegetation or dust feedbacks. The broad time-transgressive end of the AHP was recently simulated in a transient experiment of the Holocene, due to the different regional controls on precipitation (31). In contrast, the end of the AHP during the LIG has received little attention. Comparing the AHP evolution across the two interglacials can shed light onto the processes controlling its amplitude and timing.

Here, we study the evolution of precipitation and vegetation cover over North Africa across the penultimate deglaciation and the LIG, and analyse the drivers of precipitation changes as simulated by a suite of transient simulations performed with an Earth system model. The simulated precipitation evolution across the period 140 to 120 ka is further compared to a transient simulation covering the last deglaciation and Holocene (18 - 3 ka) performed with the same Earth system model, as well as to selected paleo-records.
Results

Climatic changes across the penultimate deglaciation and the LIG

The transient simulation of the penultimate deglaciation (Full, Methods) is performed with the Earth system model LOVECLIM (32), and is forced by changes in orbital parameters (33), greenhouse gases (34), ice-sheet topography and associated albedo, as well as meltwater input in the North Atlantic following the PMIP4 protocol (35) (Methods). As boreal summer insolation and greenhouse gases increase across the deglaciation (Fig. 1a,b), simulated air temperature increases over Antarctica by 12.4°C between 136.8 ka and 128 ka, in agreement with the EPICA Dome C ice core (36) (Fig. 1d). Across the deglaciation, SSTs off the Iberian margin increase by up to 7.2°C (Fig. 1e), in agreement with the SST estimates from marine sediment core MD01-2444 (37).

These deglacial changes are interrupted by HS11, during which a significantly weakened AMOC (26) induces a ~1° SST decrease in the North Atlantic compared to the penultimate glacial maximum (PGM, here taken at 140 ka). At the end of HS11, the AMOC resumption leads to an abrupt warming in the North Atlantic region as well as an increase in precipitation over southern Europe in agreement with paleo-records (Fig. 1c,e,f) (37–40). The end of HS11 is simulated here ~200 years before the maximum atmospheric CH₄ level is reached at 128.4 ka (41) (Fig. 1b).

At the PGM, the simulated precipitation is low across North Africa, with very dry conditions across both the northwest and northeast (≤20 cm/yr, Fig. 1g,h), but also reduced precipitation over the Sahel and tropical North Africa compared to PI (Fig. 2). As a result, the desert dominates North Africa between ~16°N and 32°N, and reaches 8°N in the east (Fig. 2). An abrupt increase in precipitation is observed at the end of HS11, with a maximum reached at ~128.6 ka (Fig. 1g,h). The AMOC overshoot at 128.6-128.2 ka is followed by a 400-year long AMOC drop to 20 Sv, which leads to a 25% precipitation decrease over North Africa as well as...
a temporary desert advance (Fig. 2).

The simulated precipitation evolution is in good agreement with an Al/Si record (Fig. 1h, green) obtained from marine sediment core GeoB7925-1 from the western African margin, and which provides an estimate of runoff from the western part of the Sahara (18). The initially published age model of this core (18) was revised here using the radiometrically-dated time scale of speleothems from Corchia Cave (37, 42), in order to match the latest chronological improvements proposed for Termination II and the LIG (35). This record suggests very dry conditions during HS11, followed by an increase in precipitation and peak wet conditions at ∼127 ka lasting until ∼124 ka. The Al/Si record then suggests a gradual return to dry conditions over the LIG.

The simulated abrupt shift to wet conditions at the end of HS11 in the eastern Sahara closely follows the timing of African monsoon run-off as derived from a planktic δ18O residual from Mediterranean Sea core LC21 (20) (Fig. 1h, green). A drying event is recorded at ∼127 ka, potentially linked with the C27 event highlighted in the North Atlantic (37), and both proxy and simulation suggest a return to dry conditions at 122 ka.

Both the extent and intensity of the WAM are stronger at the LIG between 128.6 and 122 ka than during PI, with a ∼2.5 mm/day precipitation increase over the Sahara, extending across the Arabian peninsula (Fig. 3a,c). The simulated LIG WAM lasts from June to September over the Sahel, and from mid-June to September over the Sahara, but the coastal Mediterranean region in North Africa also receives increased precipitation in autumn from a more southward westerly flow, and weaker anticyclonic circulation over the Azores (Fig. S1). As a result, annual mean precipitation is higher over North Africa and the Arabian peninsula by ∼35 cm/yr at the LIG, in line with proxy records (43) (Fig. S2).

The increase in precipitation at the end of the penultimate deglaciation induces an expansion of trees to ∼15°N, and an expansion of grass throughout the Sahara, reaching 30°N in the
western and 22°N in the eastern Sahara (Fig. 3a,d). This simulation thus suggests a 'green Sahara', with the total fraction of grass plus trees generally above 50% between ~128.6 and 125.7 ka. These vegetation changes impact surface albedo, which decreases by ~0.1 between 15°N and 30°N over the deglaciation (not shown). The vegetation-albedo feedback significantly contributes to the high LIG precipitation: if vegetation distribution, and thus also albedo, are kept at their PI level, precipitation over the Sahara is 40% lower (Table S1). Changes in albedo alter the radiative balance, and therefore the upward motion of air above the Sahara. The lower LIG albedo thus reduces subsidence (44, 45).

Between 128.6 and 125.6 ka, precipitation and grass cover are higher in the western than eastern part of the Sahara, while they are higher in the eastern than western part of the Sahel (Figs. 1g,h, 2, 3d). An abrupt 19% drop in precipitation and grass cover occurs over northwest Africa at 125.6 ka, while there is no change in the northeast (Fig. 1g,h). This abrupt change is not due to an AMOC weakening, and is not simulated if the vegetation cover is fixed at 127 ka level (Fig. 1g, magenta), highlighting the role of the insolation decrease leading to a vegetation-albedo feedback (46). The precipitation decrease reduces the grass cover, which leads to an albedo increase. This alters the radiative balance, thus decreasing surface air temperature over that region, and impacting the atmospheric circulation (Fig. S3). Precipitation and vegetation cover then further gradually decrease in the northwest, until dropping below PI levels during a centennial-scale AMOC weakening at 122 ka (Figs. 1g, 2). In the eastern side of North Africa precipitation stays at its maximum level until 124 ka, after which it decreases sharply, and drops below PI levels at 122 ka (Fig. 1h). Similar to the abrupt drying at 125.6 ka in the northwest, the 124 ka drying event in the northeast arises from a vegetation-albedo feedback due to the insolation decrease, as it is not simulated when vegetation is fixed at 127 ka (Fig. 1h, magenta line and Fig. S3).

Drivers of precipitation changes over North Africa at the end of the deglaciation
The low precession, and associated high boreal summer insolation, lead to a high tropical Atlantic meridional SST gradient (∼0.7°C, 5°N–30°N compared to 30°S–5°N (47)), and a northward shift of the ITCZ over the Atlantic during the early LIG (Fig. 4a,k, Table S1). The low level westerly flow associated with the WAM reaches 23°N at the LIG, and the Sahara heat low is strong (Fig. 4f). The AEJ is also displaced northward, reaching ∼30°N (Fig. S4).

To constrain the processes that control the initiation, demise and amplitude of the AHP at the LIG, a set of transient sensitivity simulations is performed with forcings each in turn fixed at 140 ka levels (Methods). Figures 4b and 5 clearly highlight that high boreal summer insolation, brought about by low precession, primarily controls the emergence and amplitude of precipitation over North Africa. When orbital parameters are fixed at 140 ka (FixOrb), precipitation over the Sahel and Sahara is reduced by ∼40% and ∼70%, respectively compared to the Full experiment (Fig. 4b). Figure 5a also shows that for positive precession values, which correspond to June insolation at 65°N below ∼500 W/m², precipitation over North Africa is suppressed. Changes in insolation significantly impact the tropical Atlantic SST gradient, which modulates the position of the ITCZ (47). The tropical SST gradient is 60% larger in Full compared to FixOrb (Fig. 4k,l, Table S1), which strengthens the summer ITCZ and shifts its location 4° northward, thus enhancing precipitation over North Africa and in particular over the Sahara (Figs. 4a,b and 5). In addition, the Sahara heat low is much weaker in FixOrb than in Full, and the AEJ is displaced 2° southward, thus leading to a weaker WAM in FixOrb (3, 4) (Figs. 4f,g, S4).

At high precession (≥0), i.e. low NH boreal summer insolation (Fig. 5), precipitation over North Africa is low and displays reduced variability: the possible full range of precipitation level is ≤30 cm/yr. As precession decreases (and thus NH insolation increases), the range of possible precipitation over North Africa increases, with a maximum range of 50 to 60 cm/yr at the peak of insolation (Fig. 5a). Most of this variability is due to changes in AMOC strength:
if the AMOC is weak, precipitation is suppressed over the Sahara and very low over the Sahel irrespective of NH summer insolation level (Figs. 4c, 5). At high insolation levels the AMOC modulates the precipitation over North Africa, with the potential to reduce it by \(\sim 75\%\) over the Sahara and \(\sim 45\%\) over the Sahel, an effect that is comparable to changes in insolation (Fig. 4b,c). The reduced meridional oceanic heat transport to the North Atlantic when the AMOC is weak leads to a cooling of the North Atlantic, while the tropical South Atlantic warms (Fig. 4m). This induces a reversed tropical Atlantic SST gradient and a southward shift of the ITCZ (29, 48) (Table S1). The southward shift of the ITCZ and weaker WAM lead to dry conditions over North Africa (Fig. 4c), with a stronger impact over the western than eastern Sahara. As millennial-scale AMOC weakenings are a common feature of deglaciations, this implies that the AMOC recovery at the end of these events controls the initiation of AHPs during the following interglacials. However, AMOC impacts are not restricted to deglaciations, and centennial-scale periods of AMOC weakening have been evidenced during the LIG (37, 49). The full transient simulation presented here indeed displays some episodes of AMOC weakening, which lead to abrupt drying of North Africa, with an expansion of desert in the middle of the AHP (e.g. from 127.65 ka to 127.35 ka, and at 121.8 ka) (Figs. 1g,h, 2). AMOC variations could thus also trigger the demise of the AHP during the LIG.

Figure 5a shows that changes in precession and AMOC do not explain everything, as even for low precession and a strong AMOC precipitation can be relatively low over North Africa (e.g. 20-30 cm/yr for precession \(\leq -0.03\) and AMOC \(\geq 22\) Sv). While secondary to insolation and AMOC, the presence of glacial ice-sheets over NH land masses impacts tropical hydrology through the position of the ITCZ and the AEJ. The ITCZ is shifted \(\sim 3^\circ\) southward in FixIS compared to the full simulation (Fig. 4d, Table S1). Through the albedo feedback, the extended NH ice-sheets lower NH temperature, maintain cold conditions in the North Atlantic and a weak tropical SST gradient (Fig. 4n, Table S1). The large Laurentide ice-sheet also creates a positive
geopotential height anomaly centred on the North Atlantic, which extends to the northwest of
Africa. This induces \( \sim 1/^\circ \) colder conditions over northwest of Africa, shifting the AEJ \( \sim 3/^\circ \)
southward and maintaining a weaker WAM compared to Full (Figs. 4d,i, S4). As a result, the
transient simulation in which NH ice-sheets are kept at their 140 ka size (FixIS) displays 55%
and 45% lower precipitation over northwestern and northeastern Africa, respectively during the
LIG compared to the Full simulation (Fig. 4d, Table S1).

While the deglacial increase in atmospheric CO\(_2\) plays a significant role in the deglacial
temperature increase (Fig. 4o), it impacts the WAM only marginally (5 to 12%) as it does
not significantly affect the tropical Atlantic SST gradient nor the meridional air temperature
gradient over North Africa (Fig. 4e,j,o, Table S1).

**Comparison with the last deglaciation**

A comparison of the precipitation evolution over North Africa across the penultimate deglaciation
with results from a transient simulation of the last deglaciation previously performed with
LOVECLIM (50) supports our conclusions (Fig. 6). Due to the stronger insolation forcing,
and the AMOC reaching a maximum strength concurrently with the disappearance of the NH
glacial ice-sheets, the simulated precipitation increase over the Sahara is larger and more abrupt
during the penultimate than the last deglaciation, particularly in the west due to the NH ice-sheet
impact. An increase in precipitation and vegetation advance is simulated during the Bølling-
Allerød (\( \sim 14.7-13 \) ka), as this corresponds to a period of relatively high NH summer insolation
(\( \sim 510 \) W/m\(^2\)) and strong AMOC (Fig. 6). However, the precipitation increase over the Sahara
is small and does not reach PI levels because of the large NH ice-sheets still present at the time.

Despite the boreal summer insolation being close to its maximum level during the AMOC re-
covery at the end of the Younger Dryas (here at \( \sim 11.7 \) ka), \( \sim 30\% \) of the NH glacial ice-sheets
are still present (51, 52) (Fig. 6b), thus attenuating the precipitation increase over North Africa.
The disintegration of NH ice-sheets during the early Holocene thus leads to a further increase
in precipitation over North Africa until 8.7 ka.

The stronger summer insolation forcing at high northern latitudes leads to much wetter conditions over the western part of the Sahara during the LIG than during the Holocene (Figs. 3a,b and 6). While the desert still dominates the western Sahara during the Holocene, it features more than 50% of grass and trees during the LIG (Figs. 3d,e and 6). By contrast in northeast Africa, north of 25°N, precipitation is similar during the LIG and Holocene, and the vegetation cover is inferior to 50% during both time periods. The demise of the AHP occurs abruptly at ∼6 ka over northeast Africa, when crossing the same insolation threshold of 0 precession (∼500 W/m²) as during the LIG at ∼122 ka (Fig. S5). In the northwest, precipitation gradually decreases over the Holocene to reach PI levels at ∼5 ka.

**Discussion**

The experiments suggest that during the LIG the WAM was stronger, and impacted a larger area than during both the Holocene and PI (Figs. 3 and 6). While the stronger WAM leads to wetter conditions over North Africa in summer, the northern edge of North Africa on the Mediterranean coast, (∼30-35°N) also benefits from a 50% increase in late autumn rainfall (Fig. S1), in agreement with CCSM3 simulations of the LIG (53) and paleo-data of the Holocene AHP (22,54). The simulated precipitation anomalies at the LIG compared to PI are in agreement with paleo-proxy records (43) (Fig. S2), and are in line with results from PMIP4 equilibrium simulations of the LIG (15), even if larger precipitation anomalies are simulated here north of 20°N, probably due to the vegetation-albedo feedback.

As a result of the wetter conditions during the LIG, vegetation expands over most of North Africa, with more than ∼50% grass cover and ∼10% of trees over the Sahara at 128 ka (Fig. 3). The simulated wetter conditions, and higher vegetation cover over North Africa at the LIG compared to the Holocene are consistent with paleo-environmental reconstructions (8), as well as with a more intensely developed sapropel S5 compared to S1 (20,55). Although the simulated
~10% tree cover over the Sahara at the peak of the LIG is an underestimation compared with previous inference of woodland (8), it is significantly higher than in a previous study, which simulated desert between 18°N and 30°N (53).

Our transient experiments support a rather abrupt initiation of the AHP during the penultimate deglaciation, while the AHP initiation was more gradual during the last deglaciation, particularly in the western Sahara. While their timings broadly follow NH insolation, their initiation is determined by the AMOC strengthening at the end of deglacial millennial-scale events. To a second order, the disintegration of NH ice-sheets also impacts the initiation of the AHP, through warming of the North Atlantic and changes in atmospheric circulation (56).

Based on the AMOC and NH ice-sheet scenarios presented here, the precipitation maximum over North Africa is simulated at 128.4 ka, thus 1.4 kyr before the insolation peak at the LIG, while during the Holocene the maximum is reached at 9.8 ka in the northeast but only at 8.7 ka in the northwest, thus 1.2 to 2.3 kyr after the insolation peak of the Holocene. With respect to insolation, the initiation of the AHP thus occurs earlier in the LIG than during the Holocene, because of the timing of the end of HS11 and the loss of NH ice-sheets. The presence of NH ice-sheets during the early Holocene (51, 52) indeed delays the AHP peak. The higher rate of NH summer insolation increase during the penultimate deglaciation might have induced a faster rate of NH ice-sheets disintegration (57–59) thus potentially leading to low NH ice-sheet volumes and strong AMOC states being reached earlier in the LIG than Holocene.

This set of simulations thus demonstrates that abrupt precipitation transitions in North Africa as a response to slow insolation changes are possible. These abrupt transitions can be triggered by AMOC changes, vegetation-albedo feedbacks, and changes in NH ice-sheets extent. However, these mostly occur for negative precession values, i.e. high northern latitude summer insolation values above ~500 W/m². The background state thus significantly conditions the capacity to generate abrupt changes, with higher insolation values associated with
a higher potential of change. The background state also lead to zonal and meridional timing
differences in abrupt shifts to drier conditions across North Africa.

Some abrupt changes in precipitation and vegetation are simulated during both the LIG and
Holocene, even though the variability is larger during the LIG than Holocene. The larger LIG
variability is due to the wetter background conditions, particularly in northwest Africa, as well
as to the larger AMOC variability. It has been shown that the LIG was punctuated by centennial-
scale AMOC weakening events (37), which would have led to abrupt weakening of the WAM
and abrupt return to arid conditions over the Sahara.

Whether rainfall over North Africa will increase or decrease over the coming century is
highly debated. The experiments performed here suggest a slightly stronger WAM with increasing
atmospheric CO$_2$. However, our simulations highlight the dominant role of the AMOC and
orbital parameters in setting the ITCZ position, and impacting precipitation over North Africa.
As precession is positive, and as the AMOC is projected to slowdown by $\sim$40% over the coming century (60) without taking into account enhanced Greenland runoff (61), an increase in
rainfall over North Africa is unlikely.

**Methods:**

A transient simulation covering the penultimate deglaciation and the last interglacial period
(140 to 120 ka) is performed with the Earth system model LOVECLIM (32). LOVECLIM
comprises an ocean general circulation model ($3^\circ \times 3^\circ$, 20 vertical levels), coupled to a dynamic-
thermodynamic sea-ice model, a quasi-geostrophic T21 atmospheric model, a vegetation model
and a global carbon cycle model. The vegetation model includes two plant functional types:
trees and grasses. The vegetation fraction is the sum of the tree and grass fractions. The vege-
tation fraction is a function of growing degree-days above 0°C and annual mean precipitation.
Vegetation changes impact the surface albedo, but not evapo-transpiration.
The transient simulation (*Full*) follows the PMIP4 protocol described in (35). The model was first integrated to equilibrium under 140 ka boundary conditions, with appropriate orbital parameters, greenhouse gas content (CO$_2$ of 191 ppm, CH$_4$ of 385 ppb and N$_2$O of 201 ppb), and continental ice-sheet geometry and albedo (35, 52, 62, 63). LOVECLIM was then integrated from 140 to 120 ka with varying orbital parameters, greenhouse gas content (34, 64), and continental ice-sheet geometry and albedo (35, 52, 62, 63). To simulate the impact of the deglacial ice-sheet disintegration and associated sea-level rise, meltwater is added into the North Atlantic (50°N-60°N, 60°W-10°W). This meltwater input is a simplified version of scenario fIRD in Menviel et al., (2019) (35), which is based on the time-evolution of IRD input into the North Atlantic. As shown in Figure 1c, here 0.12 Sv is added between 135.8 and 134 ka (sea-level equivalent (s.l.e.) of 19.2 m), and 0.2 Sv is added between 133.2 and 129 ka (s.l.e. of 74.8 m). In addition, a 400-year long negative (-0.2 Sv) freshwater input is added to enhance the AMOC recovery at the end of HS11. The Bering Strait is gradually opened between 132 and 131.8 ka.

To disentangle the impacts of deglacial changes in insolation, northern hemispheric ice-sheet, greenhouse gases and meltwater input, additional transient deglacial simulations are performed. Since there are little changes in most of the forcings, and little climatic variability between 140 and 134 ka (Fig. 1), the sensitivity experiments only start at 134 ka from *Full*, and each forcing is in turn fixed at its 140 ka level: orbital parameters are fixed at 140 ka in *FixOrb*, NH ice-sheets are fixed at 140 ka in *FixIS*, atmospheric CO$_2$ is fixed at 190 ppm in *FixCO$_2$*, and no meltwater is added in *NoFWF*. Additional experiments were also performed to isolate the impact of vegetation changes. A LIG experiment was run with vegetation fixed at PI level (*VegPI*), and another one (*VegLIG*) was run from 127 ka to 120 ka with vegetation fixed at 127 ka.

The seasonal data has been adjusted following Bartlein et al., 2019 (65) and using 20 years of monthly outputs at 128 ka.
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**Data availability** The modelling data has been submitted to Research Data Australia and will be published on 30/06/2021 at

http://handle.unsw.edu.au/1959.4/resource/collection/resdatac148/1

**Author contributions** LM designed the study and performed the numerical experiments. LM and AA analysed the simulations. AG adjusted the chronology, and provided the data of core GeoB7925-1. KMG provided the data of core LC21. LM wrote the manuscript with contributions from all the authors.

**Competing interests** The authors declare no competing interests.
Figure 1: Overview of the penultimate deglaciation and LIG as simulated in the transient *Full* experiment (black) and as inferred from key paleo-records (green). Time series of a) 65°N June insolation (33); b) atmospheric CO₂ (blue) and CH₄ (grey) as measured in the EDC ice core (34, 41, 64) on the AICC2012 chronology (66); c) North Atlantic meltwater forcing (grey), and simulated AMOC (black); d) Annual mean Antarctic air temperature anomalies compared to the temperature estimate from the EDC ice core (36) on the AICC2012 chronology; e) Annual mean Iberian margin SST anomalies (15°W - 8°W, 37°N - 43°N) compared to UK'37 derived SST from marine sediment core MD01-2444 (37); f) Annual mean precipitation on the Iberian Peninsula compared to the temperate tree pollen fraction (%) from joint marine-terrestrial analyses of the MD01-2444 deep-sea core (37); g) Annual mean precipitation over the northwest Africa (20°W - 0°W, 15°N - 30°N) in *Full* (black), VegLIG (magenta) and PI (horizontal thin black), compared to ln(Al/Si) record in GeoB7925-1 (green) indicating runoff from northwest Africa (18); h) Annual mean precipitation over northeast Africa (15°E - 32°E, 15°N - 30°N) in *Full* (black), VegLIG (magenta) and PI (horizontal thin black), compared to LC21 residuals (green) as a proxy for runoff from northeast Africa (20). Blue shading indicates the main part of HS11, purple, pink and grey shading correspond to the AHP, with a simulated drying event in the northwest between the purple and pink, and a drying event in the northeast between pink and grey.
Figure S1: Precipitation and vegetation cover evolution across North Africa. Hovmöller diagrams of annual mean (a, b) precipitation (cm/yr) and (c, d) vegetation cover (%) averaged over (a, c) northwestern (20°W - 0°) and (b, d) northeastern (15°E - 30°E) Africa as a function of latitude and across the penultimate deglaciation and LIG (140 ka to 120 ka).
Figure 3: Precipitation and vegetation cover during the LIG and mid-Holocene. a-c) Boreal summer precipitation (mm/d, shading) and 800 mb winds (m/s, vector) a) during the LIG (128 ka) compared to PI, b) during the mid-Holocene (8.6 ka) compared to PI, and c) during PI. d-f) Annual vegetation cover (%) for d) the LIG (128 ka), e) the mid-Holocene (8.6 ka) and f) PI.
Figure 4: **Processes driving the AHP.** (top) Boreal summer a) precipitation and b-e) precipitation anomalies (mm/d); (middle) f) 800mb geopotential height and winds, and g-j) 800mb geopotential height and wind anomalies (hPa and m/s); (bottom) k) SST and g-j) SST anomalies (°C) for simulations a, f, k) Full at 128 ka, b, g, l) FixOrb, c, h, m) Full at 129 ka with a weak AMOC compared to noFWF at 129 ka, d, i, n) FixIS, and e, j, o) FixCO2 compared to Full at 128 ka.
Figure 5: Drivers of precipitation changes over North Africa. Annual mean precipitation over the western part of North Africa (20°W–0°, 15°N–30°N) as a function of a) precession with symbols shading representing the strength of the AMOC (Sv), and b) the tropical Atlantic meridional SST gradient (°C, 70°W–20°W, 5°N–30°N compared to 40°W–5°E, 30°S–5°N (47)) with symbols shading representing precession, for simulations Full, FixIS, FixCO2, and NoFWF.
Figure 6: Comparison between the penultimate and last deglaciation. Time evolution across (left) the penultimate and (right) the last deglaciations of a) 65°N insolation in June; b) NH ice-sheet extent used as forcing; c) AMOC; d) simulated annual mean precipitation and e) vegetation fraction over the eastern Sahara (15°N - 30°N, 15°E-32°E); f) African monsoon runoff estimate from Mediterranean Sea core LC21 residuals (20); g) simulated annual mean precipitation and h) grass fraction over the western part of the Sahara (15°N - 30°N, 20°W-0°); i) ln Al/Si from marine sediment core GeoB-7925 (cyan) and from core GeoB-7920 (yellow) (67, 68). Blue shading represents North Atlantic cold events such as the main phase of Heinrich stadial 11 (HS11), Heinrich stadial 1 (HS1) and the Younger Dryas (YD). Grey shading represents the Bølling Allerød (BA). Pink shading represents the timing of the simulated AHP, with a particularly wet period over the western Sahara at the LIG compared to the Holocene shaded in purple. Horizontal black lines represent simulated PI values.
Supplementary Files

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