Mid-Holocene West African Monsoon Rainfall enhanced in high-resolution EC-Earth simulation with dynamic vegetation feedback

Ellen Berntell (ellen.berntell@natgeo.su.se)
Stockholm University

Qiong Zhang
Stockholm University

Research Article

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Abstract

Proxy records have shown that the Mid-Holocene was a period of humid conditions across West Africa, with an enhanced West African Monsoon (WAM) and vegetated conditions in areas currently characterized by desert, often referred to as the Green Sahara. However, General Circulation Models regularly struggle with recreating this strengthened Mid-Holocene monsoon in West Africa. The vegetation-albedo feedback has long been viewed as an essential process modulating the monsoon variability in West Africa, and simulations using prescribed vegetation to recreate a Green Sahara have shown a strengthened WAM and increased rainfall. However, these simulations represent an idealized vegetation cover and do not take any environmental heterogeneity into account. Furthermore, this only represents a one-directional forcing by the vegetation on the climate rather than the vegetation-albedo feedback. Using idealized vegetation cover might therefore over-/underestimate the changes of the WAM, as well as over-/understate the importance of the vegetation feedback. To address this, we have simulated the Mid-Holocene (~ 6 ka) climate using the high-resolution Earth System Model EC-Earth3-Veg. The results show that coupled dynamic vegetation reproduces an apparent enhancement of the WAM, with the summer rainfall in the Sahel region increasing by 15% compared to simulations with a prescribed modern vegetation cover. Vegetation feedbacks enhance the warming of the Sahara region, deepens the Sahara Heat Low, results in increased rainfall and strengthens monsoonal flow across West Africa. However, the enhancement is still below what can be viewed in proxy reconstructions, highlighting the importance of investigating other processes, such as the interactive aerosol-albedo feedback.

1. Introduction

Climate reconstructions of early- to mid-Holocene West African climate have shown a significantly different climate compared to present-day conditions, with paleo-lake-level reconstructions indicating more humid conditions and vegetation reconstructions and eolian and leaf wax reconstructions indicating vegetation covering parts of the Sahara region currently characterized by desert. This period is often referred to as the African Humid Period (AHP), and the later stage of the period, the mid-Holocene (6 ka), has received extensive interest from the modelling and proxy reconstruction communities alike (e.g., Bartlein et al., 2011; Brierley et al., 2020; Larrasoña et al., 2013; Otto-Bliesner et al., 2017). However, in spite of the significant focus being put on the Mid-Holocene by the modelling community, General Circulation Models regularly struggle with recreating the strengthened West African Monsoon (WAM), the feature most associated with the more humid climate of the period. The results from the fourth phase of the Paleo Modelling Intercomparison Project (PMIP4) also showed a clear underestimation of the Mid-Holocene rainfall enhancement across the Sahel and Sahara region compared to modern conditions (Brierley et al., 2020).

While the AHP has been shown to be an orbitally forced wet period in West Africa, driven by an increase in boreal summer insolation over the NH (Kutzbach and Liu, 1997), understanding what feedback processes and model features enhance and/or drive the strength and variability of the WAM and its representation in models is an important step in closing this model-proxy mismatch. Several modelling studies have
been dedicated to investigating this over the last few decades, and have increased our understanding of the role ocean (Kutzbach and Liu, 1997), land surface (Chandan and Peltier, 2020; Kutzbach et al., 1996; Lu et al., 2018) and dust (Pausata et al., 2016; Thompson et al., 2019) feedbacks play in enhancing the rainfall over West Africa. Vegetation feedback received special attention after the devastating droughts that plagued the Sahel region in the 1970s (Charney et al., 1975; Charney, 1975). It has been shown that a decrease in vegetation cover over West Africa suppresses rainfall in the region through albedo-vegetation feedback, while a greening strengthens the West African Monsoon (e.g., Messori et al., 2019). Similarly, past greening of the Sahara has been shown to reduce the dust fluxes by 70–80% (deMenocal et al., 2000; Egerer et al., 2016; McGee et al., 2013), which when included in modelling studies has been shown to further strengthen the WAM and shift the rainbelt northward through, e.g., a deepening of the Sahara Heat Low (Pausata et al., 2016), an important feature of the WAM. While most research on the impacts of a “green Sahara” has been focused on the Mid-Holocene, it can be equally important for deepening our understanding of future climatic change in West Africa (Pausata et al., 2020), where changes to the land cover could lead to a greener and less dusty West Africa (Evan et al., 2016; Mahowald and Luo, 2003).

However, while applying approximate Mid-Holocene land surface and dust boundary conditions have been shown to significantly strengthen the WAM and enhance the rainfall close to the levels seen in proxy reconstructions (Chandan and Peltier, 2020; Pausata et al., 2016), these simulations represent coarse, idealized changes to the land surface, vegetation cover and atmospheric dust climatology based on a limited number of proxy records found across, and along the west coast of, West Africa (Chandan and Peltier, 2020; Pausata et al., 2016; Tierney et al., 2017) and do not take any environmental heterogeneity into account. Furthermore, this only represents a one-directional forcing by the land surface and/or dust on the climate, rather than the full vegetation-albedo or dust-albedo feedback. This might therefore over-/underestimate the changes of the WAM, as well as over-/understate the importance of different feedback process and the way in which they interact.

In this paper we address this by using the fully coupled high-resolution Earth System Model EC-Earth3-veg with dynamic vegetation to simulate the Mid-Holocene (6 ka) and investigate the role of more realistic vegetation-albedo feedback in enhancing the West African rainfall.

### 2. Model Description, Experiment Design, Modelling Process And Reconstructions

#### 2.1 Model description

The present study uses EC-Earth 3.3, a fully coupled Earth System Model developed by the European consortium of 27 research institutions across Europe. The atmospheric component is the Integrated Forecasting System (IFS cycle 36r4) developed by the European Centre for Medium-range Weather Forecasts (ECMWF), which includes the revised Tiled ECMWF Scheme for Surface Exchanges over Land incorporating land surface hydrology (H-TESSEL) land model. The ocean and sea-ice components consist of the Nucleus for the European Modelling of the Ocean (NEMO) (Madec, 2008) and the Louvain-
la-Neuve Sea-ice model version 3 (LIM3) (Vancoppenolle et al., 2009). All the components are coupled through the OASIS coupler (Ocean, Atmosphere, Sea Ice, Soil) (Craig et al., 2017). The simulations either use prescribed or dynamic vegetation, with the dynamic vegetation simulated using the Lund-Potsdam-Jena General Ecosystem Simulator (LPJ-GUESS) dynamic vegetation model (Smith et al., 2014). The coupling to EC-Earth 3.3 allows vegetation parameters to be replaced with a global vegetation cover simulated by LPJ-GUESS and forced by shortwave radiation, temperature, precipitation and atmospheric CO\textsubscript{2} concentration. The vegetation is classified into different plant functional types (PFT) (Smith et al., 2014), separated in high and low vegetation types (see Table 1) with each grid box containing a fraction of high vegetation type and low vegetation type.

Table 1: Prescribed and simulated plant functional types (PFTs).

<table>
<thead>
<tr>
<th>Low vegetation types</th>
<th>High vegetation types</th>
</tr>
</thead>
<tbody>
<tr>
<td>Crops, mixed farming</td>
<td>Evergreen needleleaf trees</td>
</tr>
<tr>
<td>Short grass</td>
<td>Deciduous needleleaf trees</td>
</tr>
<tr>
<td>Tall grass</td>
<td>Deciduous broadleaf trees</td>
</tr>
<tr>
<td>Tundra</td>
<td>Evergreen broadleaf trees</td>
</tr>
<tr>
<td>Bog and marshland</td>
<td>Mixed forest/woodland</td>
</tr>
</tbody>
</table>

2.2 Experiment design

The analysis herein is based on three simulations; one pre-Industrial control simulation (PI), and two simulations of the Mid-Holocene with different vegetation set-ups (MH\textsubscript{VEG} and MH\textsubscript{REF}) (Table 2). MH\textsubscript{VEG} uses the fully coupled LPJ-GUESS for a dynamic vegetation, while PI and the mid-Holocene reference simulation MH\textsubscript{REF} use prescribed pre-industrial vegetation. All simulations have an atmospheric resolution of T255 (~80 km) and 91 vertical levels. The Pre-Industrial and Mid-Holocene simulation set-ups follow the PMIP4 experimental design described by Kageyama et al. (2018) (Table 3), with the differences between the simulations being the orbital parameters and concentrations of trace gases in the atmosphere. The difference in orbital forcing between the pre-Industrial and Mid-Holocene simulations leads to seasonal and latitudinal re-distribution of insolation, creating a strengthened seasonal cycle during the Mid-Holocene with the Northern Hemisphere (NH) receiving approximately 5% increase in insolation during the boreal summer and decrease during the boreal winter (Berger, 1978). Aerosols are set as PI climatologies for all simulations, and land-sea mask and topography are set as modern. The prescribed vegetation is the simulated 1850 CE vegetation created by an offline LPJ-GUESS simulation.
Table 2: Experimental set-up for EC-Earth 3.3 simulations. Spatial resolution of the atmosphere model indicated by grid cell extent (in degrees longitude x latitude) and number of vertical layers (L), treatment of vegetation in simulation, trend of global annual mean near-surface air temperature for the years used in the analysis.

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Atmospheric resolution</th>
<th>Vegetation</th>
<th>Description</th>
<th>Trend (K/100 yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>PI</td>
<td>T255 (~0.7 x 0.7), L91</td>
<td>1850</td>
<td>Pre-Industrial with prescribed vegetation</td>
<td>0.03</td>
</tr>
<tr>
<td>MH\textsubscript{VEG}</td>
<td>Same as PI</td>
<td>Dynamic</td>
<td>Mid-Holocene with dynamic vegetation</td>
<td>0.01</td>
</tr>
<tr>
<td>MH\textsubscript{REF}</td>
<td>Same as PI</td>
<td>Same as PI</td>
<td>Mid-Holocene with prescribed vegetation</td>
<td>0.03</td>
</tr>
</tbody>
</table>

Table 3: Orbital parameters and atmospheric trace gas concentrations used in EC-Earth 3.3 simulations.

<table>
<thead>
<tr>
<th></th>
<th>Pre-Industrial</th>
<th>Mid-Holocene</th>
</tr>
</thead>
<tbody>
<tr>
<td>Orbital parameters</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Eccentricity</td>
<td>0.016764</td>
<td>0.039378</td>
</tr>
<tr>
<td>Obliquity</td>
<td>23.459</td>
<td>24.105</td>
</tr>
<tr>
<td>Perihelion</td>
<td>-180</td>
<td>100.33</td>
</tr>
<tr>
<td>Date of vernal equinox</td>
<td>21 March at noon UTC</td>
<td>21 March at noon UTC</td>
</tr>
<tr>
<td>Trace gases</td>
<td></td>
<td></td>
</tr>
<tr>
<td>CO\textsubscript{2}</td>
<td>284.3 ppm</td>
<td>264.4 ppm</td>
</tr>
<tr>
<td>CH\textsubscript{4}</td>
<td>808.2 ppb</td>
<td>597 ppb</td>
</tr>
<tr>
<td>N\textsubscript{2}O</td>
<td>273 ppb</td>
<td>262 ppb</td>
</tr>
<tr>
<td>Other GHGs</td>
<td>CMIP DECK piControl</td>
<td>CMIP DECK piControl</td>
</tr>
</tbody>
</table>

2.3 Initial conditions, spin-up and production

The initial conditions for all simulations are taken from the end of an approx. 1000 years long pre-Industrial spin-up simulation using EC-Earth3-LR (the resolution for atmosphere is T159 and the
resolution for ocean is the same). The PI simulation reaches quasi-equilibrium after 200 years and continues to run for another 100 years for analysis, and 200 years after an initial spin-up of 200 years are used for the $\text{MH}_{\text{REF}}$ and $\text{MH}_{\text{VEG}}$ simulations due to a higher level of centennial variability. The equilibrium state of the simulations is determined by monitoring the trend, with the criteria of $\pm 0.05 \, ^\circ \text{C}$ per century in global mean surface temperature (Kageyama et al., 2018) being fulfilled by all simulations (see Table 2). Following Brierley et al. (2020) a calendar adjustment is not done, as it has been shown to have a limited impact on monsoon rainfall in Mid-Holocene simulations.

2.4 Mid-Holocene reconstruction data

The Mid-Holocene simulations are evaluated against continental mean annual temperature (MAT) and precipitation (MAP) reconstructions from Bartlein et al. (2011), where the quantitative reconstructions are inferred from pollen and plant macrofossil proxy records and represent a mean 5.5-6.5 ka anomaly relative to the present-day (see Bartlein et al. (2011) for details). The simulated vegetation in the $\text{MH}_{\text{DYN}}$ simulation is qualitatively evaluated against reconstructed biomes based on the BIOME6000 dataset (Harrison, 2017), where a $\text{MH}_{\text{VEG}}$ grid box is considered vegetated if the vegetation cover reached 20% at some point during the 200 yr. production period.

3. Results

In this section, we present the changes in monsoon rainfall and dynamics resulting from implementing a dynamic vegetation. We then examine the changes in surface energy properties linked to convection before comparing the vegetation pattern produced in $\text{MH}_{\text{VEG}}$ to vegetation reconstructions produced by proxy records. The significance of the changes are evaluated using a two-tailed student $t$-test at the 95% confidence level.

3.1. Changes in West African Monsoon rainfall and dynamics

The WAM is driven by the seasonally induced Sahara Heat Low, an extensive area of low surface pressure caused by surface warming across the Sahara region during the boreal summer (Lavaysse et al., 2009, 2010; Thorncroft et al., 2011). This drives the low-level westerly monsoonal flow, brings in moisture to the West African region from the Tropical Atlantic region and shifts the rainbelt northward into the Sahel region with a peak in rainfall occurring during the July-September (JAS) period (see Fig. 2 ab for $\text{MH}_{\text{REF}}$ JAS climatology). We therefore start by examining changes to the seasonal cycle of temperature and rainfall in the Sahara and Sahel regions respectively (Fig. 1). The changes to the orbital forcing between the PI and MH time periods result in an expected warming of near-surface air temperature in the Sahara of up to 2.5°C during the Boreal summer (July-October) and a cooling during the year’s remaining months, which can be seen when comparing $\text{MH}_{\text{REF}}$ and $\text{MH}_{\text{VEG}}$ to PI in Fig. 1a. The $\text{MH}_{\text{VEG}}$ simulation
exhibits a slight relative warming throughout the year compared to the MH$_{\text{REF}}$ simulation and a weakening of the cooling during the February to March season compared to PI (Fig. 1a). The MH summer rainfall in the Sahel is enhanced compared to PI in both MH simulations, with the MH$_{\text{VEG}}$ showing higher amounts of rainfall than MH$_{\text{REF}}$ in August with 174.7 mm/day and 159.3 mm/day respectively.

Both MH simulations exhibit an enhancement of the JAS large-scale monsoon pattern compared to PI (MH$_{\text{REF}}$ – PI; Fig. 2cd), with a low-level warming and deepening of the low-pressure area in the Sahara region together with westerly 850 hPa horizontal wind anomalies from the Gulf of Guinea into the Sahel region and positive rainfall anomalies reaching from 8°N to 16°N and a decrease of rainfall over the tropical Atlantic (MH$_{\text{VEG}}$ - PI anomalies not shown). This pattern is further enhanced when comparing MH$_{\text{VEG}}$ to MH$_{\text{REF}}$, where the Sahara region is warmed by up to an additional 0.4°C and the low-pressure area is deepened by 0.2–0.3 hPa (Fig. 2e). The JAS rainfall in the Sahel is enhanced by 30–40 mm/month, constituting an increase relative to MH$_{\text{REF}}$ of approximately 5–50% going from south to north, and the westerlies are strengthened by approx. 1 m/s (Fig. 2f). Using the definition of the northernmost extent of the monsoon as presented in Pausata et al. (2016) we can also see that the WAM has shifted northward from 13.0°N in PI to 15.8°N in MH$_{\text{REF}}$ and 16.5°N in MH$_{\text{VEG}}$ (Table 4). Additionally, the JAS rainfall in central Africa exhibits a checkerboard-like signal. This is a known issue in EC-Earth as well as other Earth System Models such as E3SMv2 (Hannah et al., 2022) with solutions being discussed within the modelling community.

The shallow convective cell induced by the Sahara Heat Low drives the African Easterly Jet (AEJ), which stretches across West Africa at approx. 500–700 hPa (Thorncroft and Blackburn, 1999), and south of the AEJ, and at an altitude of approx. 200 hPa, is the Tropical Easterly Jet (TEJ). The location of the WAM rainbelt is believed to be associated with a latitudinal band of convection, confined between these two jet streams and driven by convergence south of the AEJ and divergence north of the TEJ (Nicholson, 2009, 2013). The WAM reaches its peak as well as northernmost position in August, and in line with previous studies the following analysis will therefore be limited to August mean climatologies (e.g., Nicholson, 2009). While the PI shows an AEJ located at 11.4°N and 700 hPa altitude, the MH simulations show a strengthening and northward shift of the AEJ located at 15.1°N and 15.4°N for the MH$_{\text{REF}}$ and MH$_{\text{VEG}}$ respectively and with its center at 500 hPa altitude (Table 4). The latitude of the AEJ is also consistent with the northernmost extent of the WAM (Table 4). The latitudinal band of negative vertical wind (updraft), located between the AEJ and TEJ and linked to the location of the rainbelt, is markedly stronger in the MH simulations compared to PI, and expands northward following the northward shift of the AEJ (Fig. 3). The updraft is further enhanced in MH$_{\text{VEG}}$ compared to MH$_{\text{REF}}$, with the most significant changes centered around 13–16°N (Fig. 3cd).
Table 4
Northernmost extent of the monsoon and latitudinal location of the African Easterly Jet (AEJ) for the PI, MH\textsubscript{REF} and MH\textsubscript{VEG} simulations respectively, calculated for the peak monsoon month of August. Monsoon extent is calculated over the area 15°W-20°E, and jet stream calculated over a box 20°E-20°E and of an altitude of 500 hPa.

<table>
<thead>
<tr>
<th>Variable</th>
<th>PI</th>
<th>MH\textsubscript{VEG}</th>
<th>MH\textsubscript{REF}</th>
</tr>
</thead>
<tbody>
<tr>
<td>Monsoon extent</td>
<td>13.0°N</td>
<td>16.5°N</td>
<td>15.8°N</td>
</tr>
<tr>
<td>AEJ location (max. velocity)</td>
<td>11.4°N (9.7 m/s)</td>
<td>15.4°N (12.1 m/s)</td>
<td>15.1°N (12.0 m/s)</td>
</tr>
</tbody>
</table>

The majorit of the increase in rainfall over Sahel comes from the convective rather than large-scale, and to further examine the moist convection over West Africa, we compute the Moist Static Energy (MSE) content of the atmosphere. The MSE is defined as the sum of geopotential, enthalpy and latent energy, and is a direct indicator of monsoonal rainfall as the conversion of enthalpy and latent energy into geopotential energy in the upper troposphere is a main signal of convection (Fontaine and Philippon, 2000).

\[
\text{MSE} = gz + C_p T + Lq
\]

where \( g \) is Earth’s gravitational acceleration, \( z \) is the geopotential height, \( C_p \) is the specific heat of dry air at constant pressure, \( T \) is the temperature, \( L \) is the latent heat of evaporation, and \( q \) is the specific humidity. Compared to PI, there is a clear increase of MSE content in the atmosphere in the Mid-Holocene simulations, with positive anomalies from approx. 11°N to 30°N (Fig. 4a, MH\textsubscript{VEG} – PI not shown). These positive anomalies are further strengthened in the MH\textsubscript{VEG} simulation compared to MH\textsubscript{REF}, with significant positive anomalies 13–30°N, reaching up into the atmosphere to an altitude of 200 hPa (Fig. 4b). However, while the MSE content has increased in MH\textsubscript{DYN}, the peak MSE area in the lower troposphere (at 1000 hPa) remains at a similar latitude and magnitude as for MH\textsubscript{REF} (342.1 kJ kg\(^{-1}\) and 341.8 kJ kg\(^{-1}\), respectively at 12–14 °N) with the positive anomalies located to the north of this center (Fig. 4c).

3.2. Changes in vegetation

Figure 5 shows the prescribed vegetation used for the PI and MH\textsubscript{REF} simulations and the dynamically simulated vegetation produced by LPJ-GUESS in MH\textsubscript{VEG}. The main differences between the prescribed and dynamic vegetation are seen in the low vegetation with an approx. 5.5° northward shift from 13.5°N to 19.0°N of the tall grass in the western and eastern Sahel region and a general increase of short grass reaching up to 18.5°N in central Sahel. For the high vegetation, there is a shift from deciduous to evergreen broadleaf vegetation and a northward expansion of the tree line moving from 8°N to 11°N combined with an increase of vegetation reaching up to 15.5°N in central Sahel. However, the vegetation in MH\textsubscript{VEG} is sensitive to climate variability, and with a requirement of a mean JAS vegetation cover above 20% the northward shift remains approximately 1–3° more equatorward. The PFTs in MH\textsubscript{VEG} are also compared to reconstructed Mid-Holocene biomes based on the BIOME6000 dataset (Harrison, 2017). The
reconstructions show a clear dominance of grassland in the Sahara region, some presence of savanna in the Sahel and a combination of tropical and warm-temperate forests south of the Sahel and in central Africa. It indicates a clear underrepresentation of the northward migration of the grassland during the Mid-Holocene in the MH$_{VEG}$ simulation. While some short grass is present in the southern Sahara, the JAS max vegetation cover remains below the threshold of 20%. South of the Sahara, there is a better agreement between the reconstructed warm-temperate forest biome and the shift towards a higher representation of evergreen broadleaf vegetation in central West Africa and Central Africa in MH$_{VEG}$.

4. Discussion

4.1. West African Monsoon enhancement

The PI simulation captures a seasonal cycle over the Sahel with the majority of the annual rainfall (>80%) falling in July-September (Fig. 1b), in agreement with observations (Nicholson, 2009) and in line with previous PI simulations with other climate models (e.g., Chandan and Peltier, 2020). MH$_{REF}$ also exhibits the enhanced summer rainfall expected from the increased NH summer insolation, with rainfall in August increasing by over 120% from 71.5 mm/month to 159.2 mm/month (PI and MH$_{REF}$, respectively), similar to what has been seen in previous studies. However, despite a significant further enhancement of the monsoon rainfall in MH$_{VEG}$ (Fig. 1b and Fig. 2f), with JAS rainfall in Sahel being 15% higher in MH$_{VEG}$ compared to MH$_{REF}$, and regional JAS rainfall increases ranging from 5% in southern Sahel to over 50% at 15–20°N, it is still well below what has been seen in studies using prescribed Mid-Holocene vegetation (Chandan and Peltier, 2020; Pausata et al., 2016). Most notably, the rainfall anomalies are centered around 12°-15° N, and while it coincides with the northern edge of the MH$_{REF}$ rainbelt (Fig. 2b), the monsoon does not exhibit a northward shift in MH$_{VEG}$ compared to MH$_{REF}$ (the difference in monsoon extent is approximately the width of one grid cell, Table 4). Similarly, the AEJ neither shifts in magnitude nor latitude, which can be seen in simulations with prescribed vegetation (Gaetani et al., 2017). However, linked to the statistically increased rainfall over West Africa in MH$_{VEG}$ is a statistically significant warming and deepening of the MSL low-pressure area in the Sahara region and an enhancement of the low-level monsoonal flow which brings moisture to the WAM (Fig. 2ef). It indicates that, while the signal is weaker compared to previous studies with prescribed vegetation, the West African Monsoon was intensified through the inclusion of dynamic vegetation, consistent with our understanding of the rainfall-vegetation-albedo feedback (Charney et al., 1975).

At the same latitudinal band as the positive rainfall anomalies is a strengthened uplift (12°-15°N), located in the northern part of the convective belt between the AEJ and TEJ (Fig. 3cd). Changes to the convective rain, which is responsible for the vast majority of West African Monsoon rainfall, is examined by analyzing the Moist Static Energy content in the atmosphere. The MSE describes the conversion of enthalpy and latent energy near the surface into geopotential energy through convection, and the results show apparent positive anomalies through the atmosphere reaching from 12°N to 30°N, with the most significant anomalies centered at 15°-20°N. These anomalies, which are linked to increased sensible and
latent heat flux at the surface, are located south of the anomalies in previous studies with prescribed vegetation (Gaetani et al., 2017) and markedly weaker (Fig. 4). While the MSE anomalies are significant, they do not shift the low-level (1000 hPa level) MSE peak northward, something that has been suggested to favor the northward migration of the rainbelt (Gaetani et al., 2017).

### 4.2. Vegetation-albedo-rainfall feedback

Previous studies have shown that including prescribed land surface feedbacks, such as prescribed Mid-Holocene vegetation and soil, results in a warming of the Sahara region and a strengthening of the WAM compared to orbital-only simulations (e.g., Chandan and Peltier, 2020). This relative warming is present all year, but the largest warming occurs during late fall to mid spring. However, while our MH$_{\text{REF}}$ simulation exhibits a very similar seasonal temperature cycle anomaly relative to PI as e.g. Chandan and Peltier (2020), the pronounced warming seen in simulations with prescribed vegetation is not present in MH$_{\text{DYN}}$ with a warming of 0.1–0.2°C in February – April (Fig. 1a) compared to 1.0-1.5°C seen in Chandan and Peltier (2020) and a warming over Sahara in the monsoon season of up to 0.3°C (Fig. 2) compared to 1–7°C seen in simulations with prescribed vegetation (Pausata et al., 2016). This seasonal warming of the Sahara region is linked to the onset of the monsoon and the northward shift of the rainbelt during the monsoon season, with the SHL being a driver of the WAM and its strength being strongly correlated to rainfall variability in the Sahel (Biasutti et al., 2009; Lavaysse et al., 2009). Correspondingly, while we see a statistically significant enhancement of the WAM dynamics when including dynamic vegetation (deepening of the Sahara Heat Low and strengthening of the low-level horizontal winds), in line with the expectations from having vegetation-rainfall-albedo feedbacks, it is again markedly weaker compared to simulations with idealized, prescribed vegetation (Pausata et al., 2016). The lack of strong warming seen in MH$_{\text{DYN}}$ therefore explains the relatively weak enhancement of the monsoon rainfall and the lack of the WAM's northward shift, which can be seen when using a prescribed vegetation. This indicates that while vegetation has been shown to sustain an enhancement and northward shift of the WAM, the solely orbitally driven vegetation-rainfall-albedo feedback is not on its own strong enough to enhance neither the monsoon nor the vegetation to mid-Holocene levels. Indeed, when we compare the MH$_{\text{DYN}}$ anomalies (relative to the PI simulation; Fig. 6) to proxy records (Bartlein et al., 2011), we can see that the enhancement is underrepresented in temperature and rainfall. While MH$_{\text{VEG}}$ does capture the overall pattern (warming in Sahara and wetting across West Africa) and the agreement is closer than for MH$_{\text{REF}}$, the signal is markedly weaker than the reconstruction, especially with missing the strong Sahara warming. Although, it is important to note that while our anomalies are calculated as MH-PI, the proxy reconstructions represent the difference between MH and modern climate, using CRU observational data when modern reconstructions are unavailable, which might contribute to differences in magnitude.

Given that the vegetation-rainfall-albedo feedback appears not strong enough to recreate a mid-Holocene WAM and vegetation cover, it is plausible to assume that other feedback processes are equally crucial for enhancing the monsoon and strengthening the vegetation-rainfall-albedo feedback loop, and that running simulations with prescribed vegetation in Sahara indirectly includes the impact of these known and unknown feedbacks in our simulations through their effect on the vegetation. The direct effects of such
processes have been investigated in previous studies, where especially albedo-related feedbacks caused by changes in atmospheric dust concentrations and soil properties in Sahara have been shown to warm the Sahara region, and consequently increase the monsoon rainfall and further shift the rainbelt and vegetation northward (Chandan and Peltier, 2020; Kutzbach et al., 1996; Pausata et al., 2016). As these feedbacks occur as a result of the changes in vegetation, which creates a reduction of atmospheric dust and changes to the soil, coupling an interactive dust-aerosol model with the dynamic vegetation could strengthen the response of the monsoon and vegetation changes. Similarly, improving dynamic soil composition and albedo schemes could further enhance the WAM during the Mid-Holocene. These model improvements have been recommended for future ESM studies of past Green Sahara periods (Lu et al., 2018). However, current state-of-the-art global climate models with dynamic dust have been criticized for failing to reproduce fundamental aspects of dust emissions and transport, and instead introduce more uncertainty in model simulations (Evan et al., 2014; Kok et al., 2017; Zhao et al., 2022), which highlights the importance of choosing a model configuration most relevant to the research question in balance with running more complex, resource-heavy coupled simulations. Chandan and Peltier (2020) also point out that while recent studies have argued for the important role of reduced dust in the Sahara when simulating the mid-Holocene climate (Pausata et al., 2016; Thompson et al., 2019), their study shows that agreement with proxies can be reached solely by including feedbacks from land surface changes, indicating that the pathways for reaching sufficient enhancement of rainfall might be model dependent. Additionally, while the $\text{MH}_{\text{DYN}}$ does not exhibit the pronounced strengthening of the WAM that has been seen in simulations with prescribed vegetation, resulting in a dry bias and lack of vegetation similar to what has been seen in previous studies (Braconnot et al., 2019), other ESMs with dynamic vegetation have successfully been able to recreate a strengthened WAM in line with models using prescribed paleo vegetation (Berntell et al., 2021; Brierley et al., 2020; Stepanek et al., 2020), which again indicates a level of model dependency in the vegetation-rainfall-albedo feedback and highlights the need for additional multi-model studies to draw more robust conclusions about its strength. Several studies have evaluated the relationship between rainfall and vegetation in West Africa, and shown that LPJ-Guess generally produces less vegetation in low-rainfall areas such as Sahara and require over twice as much annual rainfall to reach a 20% grass coverage compared to observations (Hopcroft et al., 2017). This might result in a dampening of the vegetation-albedo feedback and contribute to the lack of expansion of grassland seen in $\text{MH}_{\text{VEG}}$.

5. Conclusion

The Mid-Holocene simulation with dynamic vegetation ($\text{MH}_{\text{VEG}}$) exhibits a significant increase in summer rainfall across West Africa compared to the orbital-only forced simulation ($\text{MH}_{\text{REF}}$), and a weak enhancement of the West African Monsoon dynamics with a warming of Sahara, deepening of the Sahara Heat Low and strengthening of the low-level winds which bring moisture from the Atlantic to the West African continent. However, despite an enhancement of the uplift and an increase in moist static energy in the northern edge of the rainbelt, the increased rainfall and warming of the Sahara region are markedly lower than those in simulations using prescribed vegetation, and there is no apparent northward
shift of the WAM. It indicates that the vegetation feedback, when only driven by orbital forcing, is not strong enough to induce such a shift. Our results suggest that additional feedbacks such as dust-albedo and soil feedback are required to enhance the rainfall-vegetation-albedo feedback and cause the vegetation to reach mid-Holocene levels.

**Declarations**

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**Author contributions.** Both authors contributed to the study conception and design. Ellen Berntell set up and ran the simulations, analyzed the data and wrote the manuscript, QZ commented on previous versions of the manuscript.

**Data availability.** The data is available through communication with the authors, and will be published on the public data repository Zenodo upon acceptance of article.

**References**


Figures

Figure 1

Seasonal cycle of (a) near-surface temperature in Sahara [unit: °C] (b) rainfall in Sahel [unit: mm/month] for MH$_{\text{REF}}$ (blue), MH$_{\text{DYN}}$ (green) and PI (black). Seasonal cycle of temperature is shown relative to PI, which is therefore not shown. The Sahara region is defined as 10°W-20°E, 20-30°N, and the Sahel is defined as 20°W-30°E, 10-20°N.
Figure 2

July-September mean near-surface temperature (TAS), mean sea level pressure (MSL), rainfall (PRE) and low-level horizontal wind (UV) at 850 hPa for all MH simulations; MH\textsubscript{REF} (top panel) climatologies, MH\textsubscript{REF} anomalies (middle panel; MH\textsubscript{REF} - PI) and MH\textsubscript{VEG} anomalies (bottom panel; MH\textsubscript{VEG} - MH\textsubscript{REF}). Significant (95%) MH\textsubscript{VEG} anomalies are indicated by x (temperature and rainfall) and + (sea level pressure) stippling, and only significant wind anomalies are shown.
Figure 3

August vertical wind climatology averaged between 10°W and 10°E [unit: Pa s\(^{-1}\)] for (a) MH\(_{\text{REF}}\), (b) MH\(_{\text{DYN}}\), (c) MH\(_{\text{LR}}\) and (d) Pl. Vertical wind anomalies for (d) MH\(_{\text{LR}}\) and (e) MH\(_{\text{DYN}}\) against MH\(_{\text{REF}}\), with significant anomalies indicated with stippling.
Figure 4

August Moist Static Energy (MSE) in the atmosphere calculated as zonal means between 20°W and 30°E [unit: kJ kg\(^{-1}\)]. (a) MH\(_{\text{REF}}\), (b) MH\(_{\text{LR}}\) and (c) MH\(_{\text{DYN}}\) anomalies against PI and (d) MH\(_{\text{LR}}\) and (e) MH\(_{\text{DYN}}\) against MH\(_{\text{REF}}\). Significant anomalies indicated with stippling.
Figure 5

Prescribed a) low and c) high vegetation for MH$_{\text{REF}}$ and PI, and simulated, mean b) low and d) high vegetation for MH$_{\text{VEG}}$. Bare soil is specified for grid cells with <20% July-September max vegetation cover (indicated in white). Reconstructed biomes, shown in filled circles, based on the BIOME6000 dataset (Harrison, 2017).
Figure 6

Mean annual near-surface temperature (TAS) anomalies [unit: °C] and annual rainfall anomalies [unit: mm/year] for MH$_{\text{REF}}$ and MH$_{\text{VEG}}$. Mid-Holocene anomalies (MH - PI) shown together with proxy-inferred temperature and rainfall anomalies (6-0k) from Bartlein et al. (2011).