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Key Points:

- The Earth system model EC-Earth reproduced the expansion of vegetation in northern high latitudes during the mid-Holocene compared to the PI
- The mid-Holocene Arctic greening contributes 0.33°C increase in annual Arctic warming and 0.35×10^6 km² of sea ice loss
- The increased evapotranspiration and decreased surface albedo due to the Arctic greening amplify the Arctic warming in the growing season

Supporting Information:

Supporting Information may be found in the online version of this article.

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The Contribution of Vegetation-Climate Feedback and Resultant Sea Ice Loss to Amplified Arctic Warming During the Mid-Holocene

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Abstract Understanding influence of vegetation on past temperature changes in the Arctic region would help isolate uncertainty and build understanding of its broader climate system, with implications for paleoclimate reconstructions and future climate change. Using an Earth system model EC-Earth, we conduct a series of simulations to investigate the impact of vegetation-climate feedback on the Arctic climate during the mid-Holocene. Results show Arctic greening induced by the warming resulting from stronger orbital forcing, further amplifies the Arctic warming. The increased vegetation contributes 0.33°C of Arctic warming and 0.35×10^6 km² of Arctic sea ice loss. Increased Arctic vegetation leads to reduced land surface albedo and increased evapotranspiration, both of which cause local warming in spring and summer. The resultant sea ice loss causes warming in the following seasons, with atmospheric circulation anomalies further amplifying the warming. Our results highlight the significant contribution of vegetation-climate feedback to Arctic climate under natural conditions.

Plain Language Summary Vegetation-climate feedback is an essential process between the terrestrial biosphere and the atmosphere. Greening of the Arctic can be invoked by increased solar radiation, for example, during the mid-Holocene (6,000 years ago), or increased greenhouse gas, for example, ongoing global warming. Here we perform simulations of vegetation and climate interactions during the mid-Holocene and demonstrate the impact of vegetation-climate feedback on the Arctic climate. We show that compared to the pre-industrial period (1850 CE), the Arctic region experiences an expansion of vegetation under a warm climate due to increased solar radiation during the mid-Holocene. The greening leads to reduced land albedo and enhanced evapotranspiration. These changes and resultant sea ice loss, in combination with associated atmospheric circulation changes, further amplify the Arctic warming. These changes alone are responsible for 0.33°C increase in surface air temperature and 0.35×10^6 km² increase in sea ice loss. Given the facts of observed ongoing Arctic greening in recent decades, our modeling results suggest that the effect of the vegetation-climate feedback will amplify the future Arctic warming and sea ice loss even more.

1. Introduction

Climate change has led to increased global temperatures over the last century, but this warming has been especially apparent in the Arctic, which has warmed at more than twice of the global rate over the past 50 years and is expected to continue to experience exaggerated warming in the future (Doblas-Reyes et al., 2021; Hind et al., 2016; Serreze & Barry, 2011). This phenomenon, known as Arctic amplification, has been found in reconstructions of past climates (Brigham-Grette et al., 2013; Miller et al., 2010) and model simulations for past (de Nooijer et al., 2020; Q. Zhang et al., 2010; Zheng et al., 2019) and future climates (Barnes & Polvani, 2015). Such regional differences have profound impacts on global climate, ecological systems and socio-economic landscape (Oostdijk et al., 2021; Schuur et al., 2015; Trusel et al., 2018). Many factors have been suggested to explain the Arctic amplification, such as the radiative feedbacks related to surface albedo, cloud cover and water vapor (T. Gong et al., 2017; Taylor et al., 2013; Vavrus, 2004), changes in lapse-rate (Boeke et al., 2021), non-radiative forcing of atmospheric circulation and oceanic heat transport (Cai, 2005; Spielhagen et al., 2011; Yoo et al., 2011). In

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particular, sea ice loss plays a central role in the Arctic amplification either directly or indirectly via these factors (Burt et al., 2016; Dai et al., 2019; Luo et al., 2017; Screen & Simmonds, 2010). As a consequence of the Arctic warming, changes in vegetation have been observed over the Arctic region (Bunn et al., 2007; Myers-Smith et al., 2020; Piao et al., 2020). It could, in turn, affect the local climate system by disturbing surface energy fluxes through changes in albedo and the hydrological cycle (Bonan, 2008; Bonan et al., 1992). It is widely recognized that expansion of vegetation over the Arctic region has a great potential to amplify large-scale warming through decreases in albedo and increases in evapotranspiration (ET) and resultant sea ice loss as observed in recent decades and projected for the future (Bonfils et al., 2012; Jeong et al., 2012, 2014; O'ishi & Abe-Ouchi, 2009; Swann et al., 2010).

Past warm climate periods, such as the Holocene thermal maximum (HTM, about 10,000 to 6,000 years BP; Kaufman, McKay, Routson, Erb, Dätwyler, et al., 2020; Marcott et al., 2013; Renssen et al., 2009), show warm conditions due to the much stronger Northern Hemisphere summer solar insolation than during the pre-industrial period (PI) (Berger, 1978). Proxy-based information reveals that this thermal maximum was most pronounced at high latitudes, including Greenland, Western Arctic, and Northern Europe (Gajewski, 2015; Kaufman et al., 2004; Kaufman, McKay, Routson, Erb, Davis, et al., 2020). Correspondingly, proxy data indicates that Arctic sea ice cover was reduced (de Vernal et al., 2013; Funder et al., 2011). There are also many climate model studies supporting the prominent Arctic warming associated with sea ice loss induced by changes in orbital forcing during the mid-Holocene (Park et al., 2018; Williams et al., 2020; Q. Zhang et al., 2010, 2021), and highlighting the large contribution of the sea ice loss to mid-Holocene Northern Hemisphere warming (Park et al., 2019). During the HTM, the high-latitude tree line shifted poleward (Lindgren et al., 2021; MacDonald et al., 2000), particularly in Eurasia and North America (Binney et al., 2017; Payette & Lavoie, 1994). This poleward expansion of the trees during the HTM provides an excellent opportunity to understand the vegetation-climate feedback in the past warm period under natural conditions. It helps predict and understand future and potential climate. Some modeling studies suggest that past vegetation changes could have amplified the regional warming at northern middle and high latitudes during the HTM (Kaufman et al., 2004; Otto et al., 2009). A similar feature has been found in glacial-interglacial climate transitions during the Late Quaternary (Claussen, 2009; Davies-Barnard et al., 2017). However, only the direct effect of reduced land albedo and the indirect effect of sea ice loss due to increases in vegetation were put forward to explain the mechanisms during the HTM (Foley et al., 1994; O'ishi & Abe-Ouchi, 2011). Many processes of the Arctic vegetation-climate feedback during the HTM remain to be quantified, for instance, the impact of vegetation expansion on sea ice loss and the influence of changes in ET associated with vegetation increases on the Arctic warming. The mid-Holocene is a typical period in the HTM for which plenty of vegetation and climate reconstructions exist, and these could be used to constrain model results. Thus, in this study, we have performed several sensitivity experiments using an Earth System Model EC-Earth. We have systematically analyzed the impact of the vegetation-climate feedback on the Arctic climate and resultant sea ice changes during the mid-Holocene, and then demonstrated the impact pathway through direct and indirect processes and highlighted the importance of changes in ET.

2. Data and Methods

2.1. Model Description

We employ a fully coupled Earth system model: the Coupled Model Intercomparison Project Phase 6 (CMIP6) version of the European Consortium Earth System Model (EC-Earth, version3.3; Döscher et al., 2022). In the current study, components for the atmosphere, ocean, sea ice, land and dynamic vegetation are used. The atmospheric component is the Integrated Forecast System (IFS) developed at the European Centre for Medium-Range Weather Forecast (ECMWF), including the land model H-TESEL (Balsamo et al., 2009). The ocean component is the Nucleus for European Modeling of the Ocean (NEMO3.6; Madec & NEMO Team, 2016), and the sea ice component is the Louvain-la-Neuve sea ice model (LIM3; Rousset et al., 2015). The dynamics vegetation component is the second-generation dynamic vegetation-ecosystem model (LPI-GUESS) based on an individual-based and patch-based representation of land ecosystem structure and dynamics (Smith et al., 2014). The vegetation dynamics in the LPI-GUESS result from growth and competition for light, space and soil resources among woody plant individuals and a herbaceous understory in each grid cell (Smith et al., 2014). The different components and their coupling allow for simulating complex interactions between the atmosphere, ocean and land biosphere. In the Arctic region, the present-day leaf area index (LAI) simulated by the dynamic vegetation module of

EC-Earth is roughly consistent with the satellite observation (Figures S1a–S1c in Supporting Information S1; Song et al., 2021). The simulated surface air temperature (SAT) and sea ice concentration (SIC) during the historical period 1981–2010 are reasonable compared to ERA5 reanalysis (Hersbach et al., 2020), with better results in the model configuration with coupled dynamic vegetation compared to that of without (Figures S1d–S1i in Supporting Information S1).

2.2. Experimental Design

We conduct four simulations using the coupled dynamic vegetation configuration EC-Earth3-Veg-LR and the configuration EC-Earth3-LR without dynamic vegetation. Following the Paleoclimate Model Intercomparison Project Phase 4 (PMIP4) protocol (Kageyama et al., 2018), two simulations were performed with pre-industrial orbital, and greenhouse gas conditions (PIorb) and mid-Holocene orbital and greenhouse gas conditions (MHorb) using EC-Earth3-LR with prescribed vegetation, and another two simulations (PIveg, MHveg) were performed with the same boundary conditions using EC-Earth3-Veg-LR with coupled dynamic vegetation (Table S1 in Supporting Information S1). The dynamic vegetation in PIVEg and MHveg was initialized from the prescribed vegetation (Table S2 in Supporting Information S1). The horizontal resolution of atmosphere, land and vegetation is T159 ($\sim 1.125^\circ$, or ~ 125 km), with 62 vertical levels in the atmosphere. The ocean and sea ice model has a horizontal resolution of $\sim 1^\circ$, and the ocean has 75 vertical levels. We apply the initial conditions for the four simulations from the end of the PI spin-up phase (which already run for about 1,000 years). Another spin-up period of 200 years is used to reach a new equilibrium (global mean surface temperature trend $< \pm 0.05$ K per century; Kageyama et al. (2018)). Subsequently, 200 years of data from the following period in the simulations were used for the analysis. The prescribed vegetation in PIorb and MHorb is taken as the simulated vegetation for 1850 CE from an offline simulation with LPJ-GUESS (Chen et al., 2021). A more detailed description of PIorb and MHorb could be found in Q. Zhang et al. (2021). In addition to the vegetation, there is a substantial difference in the orbital forcing and a slight difference in greenhouse gas concentrations between the PI and the mid-Holocene simulations (Table S1 in Supporting Information S1). Thus, the differences between MHorb and PIorb could represent the climate anomalies due to differences in orbital forcing ($\Delta\text{Orb} = \text{MHorb} - \text{PIorb}$). The differences in climate resulting from both differences in orbital forcing and vegetation–climate feedback could be represented by the differences between MHveg and PIVEg ($\lambda = \text{MHveg} - \text{PIveg}$). Ultimately, vegetation–climate feedback could be further distinguished by comparing the difference between λ and ΔOrb ($\Delta\text{Veg} = \lambda - \Delta\text{Orb}$).

2.3. Proxy Data

The PI and mid-Holocene biome reconstructions from BIOME6000 (Harrison, 2017) were used in this study. Since the vegetation types represented by the biome reconstructions are different from the plant functional type (PFT) output of the model, we compared the two vegetation categories following the method in Prentice et al. (2011) (see Table S3 in Supporting Information S1). We use reconstructed temperatures from the Temperature 12k database (Kaufman, McKay, Routson, Erb, Dätwyler, et al., 2020). It is an extensive new multi-proxy database of paleo-temperature time series. Five different statistical methods were used to generate a multi-method ensemble of global mean surface temperature and latitudinal-zone temperature reconstructions.

3. Results and Discussions

3.1. Changes in the Northern High-Latitude Vegetation

EC-Earth3-Veg-LR simulates a reduced tundra that was replaced by grassland and a northward expansion of deciduous needleleaf trees in northern high-latitude regions during the mid-Holocene compared to the PI (Figure 1). The feature is consistent with previous simulations of vegetation differences between the mid-Holocene and the PI (Kutzbach et al., 1998; Oishi & Abe-Ouchi, 2011; Wohlfahrt et al., 2008). The simulated forest distribution does correspond well with the biome reconstructions in high latitudes (Figures 1d and 1e). However, the simulated vegetation cover is very low in most of these areas (below 10%). This may be due to the underestimation of the vegetation cover as suggested by present-day simulation results (Figure S1c in Supporting Information S1), attributing to a combination of a cold bias and a systematic underestimation of forest biomass as other modeling studies suggested (Braconnot et al., 2019; Guimberteau et al., 2018). Nonetheless, the model shows an evident expansion of vegetation during the mid-Holocene, which is in line with the BIOME6000. Our results further

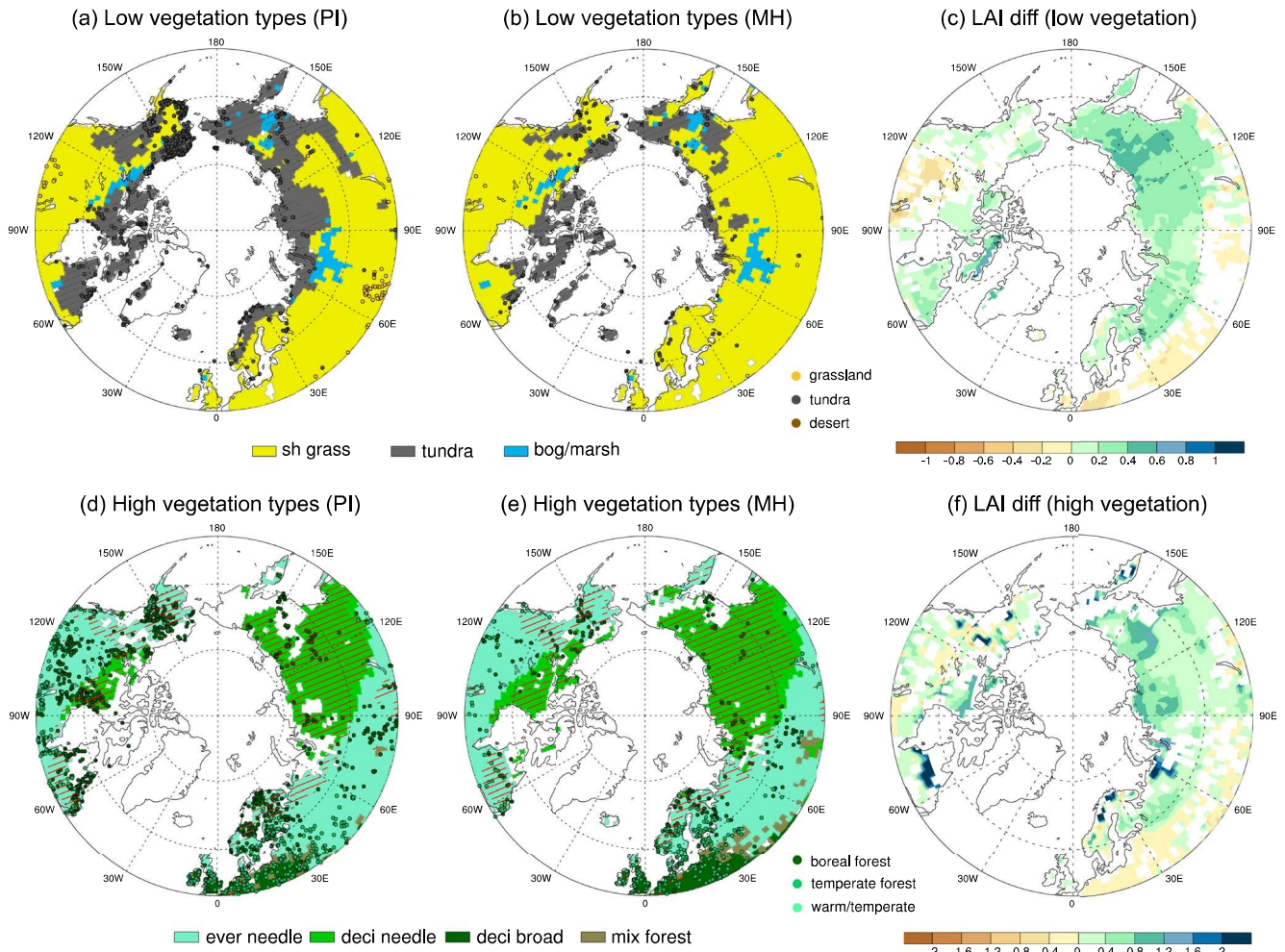


Figure 1. Dominant (a, b) low and (d, e) high vegetation distribution simulated by EC-Earth3-Veg-LR for PIveg and MHveg. (c, f) The summer (JJA) leaf area index (LAI, unit: m^2/m^2) differences between MHveg and PIveg for low vegetation and high vegetation, respectively. The red oblique line is specified where the vegetation cover is $<10\%$. The circles are reconstructed biomes in the mid-Holocene from BIOME6000 data set (Harrison, 2017). The LAI differences exceeding the 95% confidence level based on a Student's t -test are shown in (c) and (f).

show that north of 60°N leaf area index (LAI) of low vegetation (herb) increases by $\sim 14\%$, and that of high vegetation (woody plant) increases by $\sim 30\%$ in summer (Figures 1c and 1f), when comparing the changes of LAI in the mid-Holocene to that in PI.

3.2. Contribution of Vegetation-Climate Feedback to SAT and SIC Changes

Our results indicate a prominent Arctic warming (north of 60°N) throughout the year in the mid-Holocene relative to the PI, with annual mean warming of 1.64°C including the vegetation-climate feedback and 1.31°C excluding them (Figure 2c and Table S4 in Supporting Information S1). Thus, changes in vegetation further amplify Arctic warming in all seasons, with an annual increase ratio of approximately $\sim 25\%$ by comparing the warming caused by vegetation-climate feedback and caused by orbital forcing. The additional warming due to the vegetation-climate feedback is mainly seen in northern Eurasia, northeastern Canada, Greenland and its surrounding oceans (Figure 2a). Our results are consistent with previous studies showing amplified Arctic warming with similar distribution due to vegetation-climate feedback during the mid-Holocene (Foley et al., 1994; O'ishi & Abe-Ouchi, 2011; Otto et al., 2009; Wohlfahrt et al., 2004). The median value of 1.04°C ($0.48 \sim 4.09^\circ\text{C}$, 5th $\sim$ 95th percentiles) obtained from Temperature 12k is relatively lower than our results of 1.64°C . However, compared with other PMIP4 simulations with prescribed vegetation, Temperature 12k gives larger warming in Arctic temperature between the mid-Holocene and the PI (Brierley et al., 2020). Moreover, many reconstructions

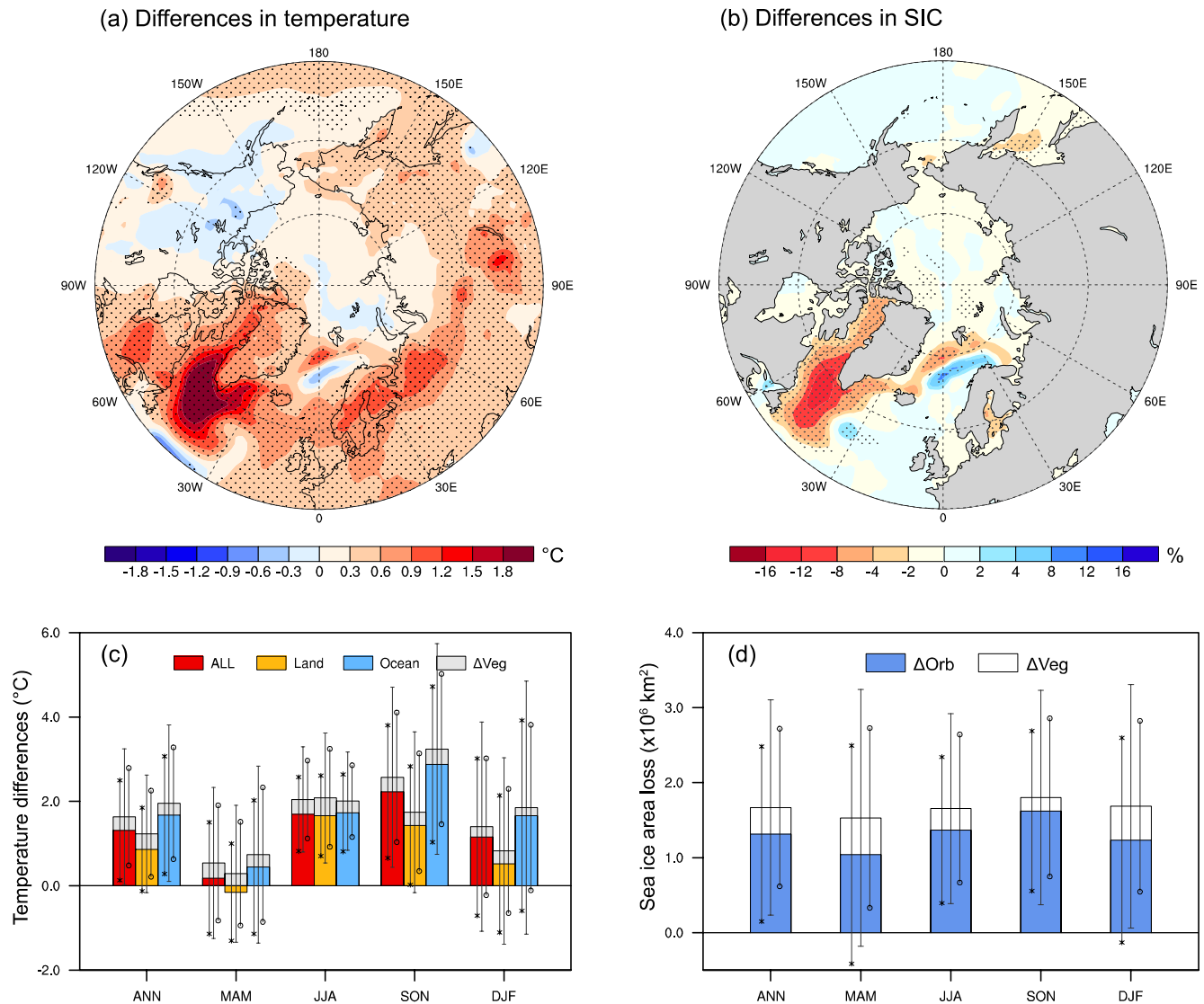


Figure 2. (a) Distribution of annual mean surface air temperature (SAT, unit: °C) and (b) annual sea ice concentration (unit: %) differences due to vegetation-climate feedback (ΔVeg) during the mid-Holocene. (c) SAT differences area-weighted average over north of 60°N and (d) sea ice area loss (unit: $10^6 \text{ km}^2/\text{month}$) due to orbital forcing (ΔOrb) and vegetation-climate feedback (ΔVeg). The dots in (a) indicate that the values exceed the 95% confidence level based on a Student's *t*-test. The error bars with stars, lines and circles denote the ± 1 standard deviations of results due to orbital forcing, vegetation-climate feedback and both of them, respectively.

show a stronger Arctic warming between the mid-Holocene and the PI than model results without consideration of the vegetation-climate feedback (Mann et al., 2009; Park et al., 2019; Sundqvist et al., 2010; Williams et al., 2020). These suggest that the vegetation-climate feedback may be essential in reconciling the data-model discrepancy of Arctic annual temperature during the mid-Holocene.

The simulation shows a significant decrease in the SIC throughout the year due to the orbital forcing, with the most extensive loss in autumn (Figure 2d), which is generally consistent with the other PMIP4 mid-Holocene simulations (Brierley et al., 2020; Williams et al., 2020). The annual mean sea ice loss is $1.32 \times 10^6 \text{ km}^2$ due to the orbital forcing (Table S5 in Supporting Information S1). The sea ice further decreases throughout the year due to the increased Arctic vegetation, especially in spring and winter (Figure 2d). The increased vegetation further contributes $0.35 \times 10^6 \text{ km}^2$ annual mean sea ice loss beyond caused by orbital forcing, corresponding to an increase ratio of $\sim 26\%$ (Figure 2d and Table S5 in Supporting Information S1). The significant annual mean SIC differences between the two mid-Holocene simulations occur in North Atlantic and Arctic Ocean around Greenland, which agree well with the SAT differences here (Figures 2a and 2b).

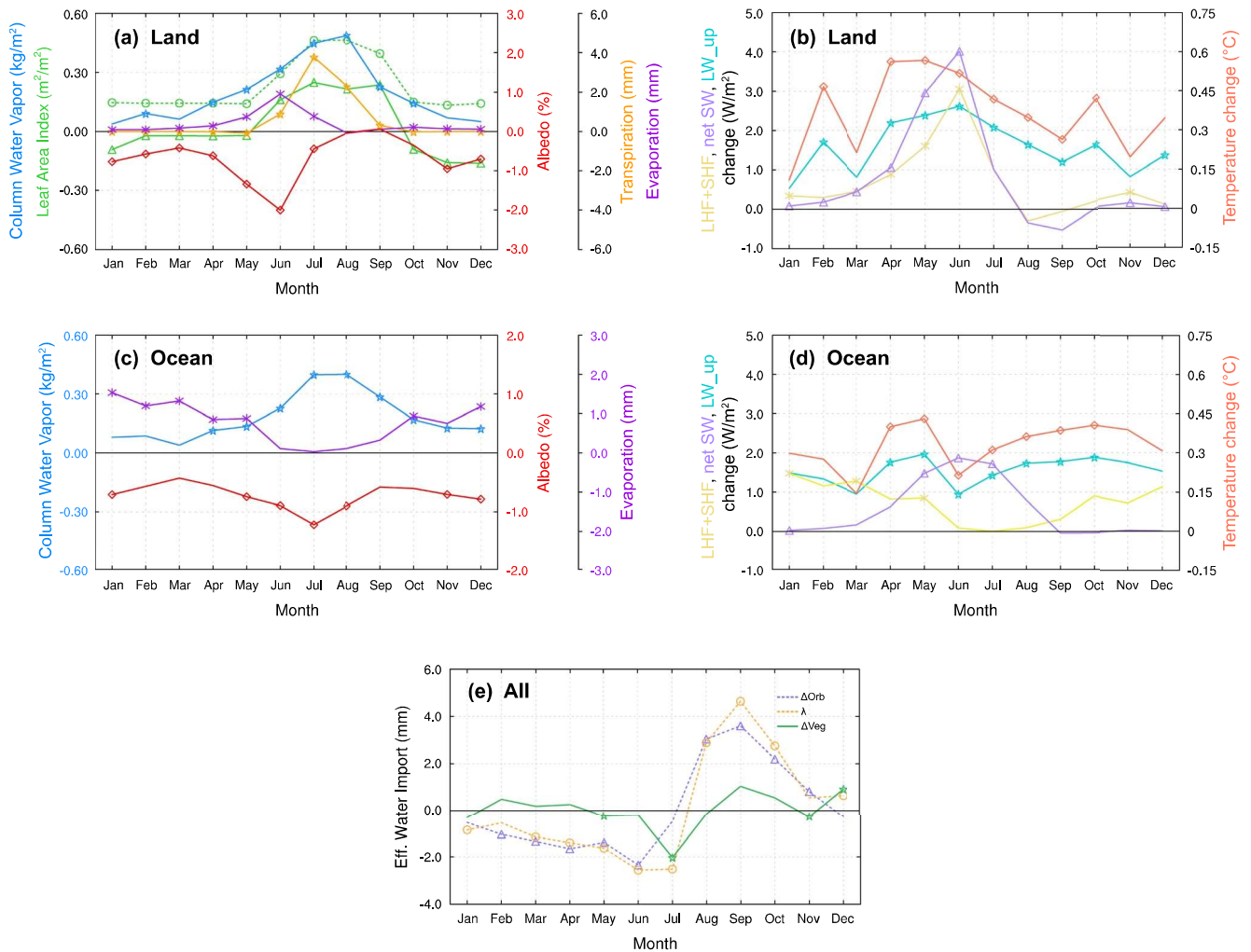


Figure 3. (a) Differences of transpiration (orange solid line, unit: mm/month), evaporation (purple solid line, mm/month), column water vapor (blue solid line, unit: kg/m²), leaf area index of high vegetation (green dashed line) and low vegetation (green solid line, unit: m²/m²), albedo (red solid line, unit: %) area-weighted average over land area north of 60°N due to vegetation-climate feedback (ΔVeg). (c) Column water vapor (blue solid line, unit: mm/month), evaporation (purple solid line, unit: mm/month) and albedo (red solid line, unit: %) anomalies area-weighted average over ocean north of 60°N due to vegetation-climate feedback (ΔVeg). Latent heat flux and sensible heat flux (LHF + SHF, yellow line, positive upward, unit: W/m²), net shortwave radiation (net SW, purple line, positive downward, unit: W/m²), upward longwave radiation (LW_{up}, cyan line, positive upward, unit: W/m²), surface air temperature (orange red line, unit: °C) changes area-weighted average over (b) land and (d) ocean north of 60°N due to vegetation-climate feedback (ΔVeg). (e) The effective water vapor import (unit: mm/month) anomalies over north of 60°N due to orbital forcing (ΔOrb, purple dashed line), orbital forcing with vegetation-climate feedback (λ, orange dashed line), and vegetation-climate feedback (ΔVeg, green solid line). Markers indicate that the results exceed the 95% confidence level based on a Student's *t*-test.

3.3. Direct Influence of Vegetation-Climate Feedback on the Arctic Warming

The low vegetation LAI and high vegetation LAI are larger from June to September in the mid-Holocene than that in the PI, with a peak in July, corresponding to the more greening Arctic (Figures 1 and 3a). Due to the expansion of the deciduous needleleaf trees during the mid-Holocene compared to that in the PI, the high vegetation LAI became larger throughout the year (Figures 1d, 1e, and 3a). The leaf-out across the Arctic in June associated with snow melt in spring when the sun comes out leads to significant surface albedo decreases in May and June. The stem area and a small residual leaf area are maintained in other seasons, causing the reduction of surface albedo throughout the year (Figure 3a). These surface albedo changes result in increased absorbed solar radiation (total of 9.13 W/m²) at the surface in spring and summer. This energy is mainly released in the form of latent heat fluxes (LHF) and sensible heat fluxes (SHF) to the atmosphere (Figure 3b and Figure S3 in Supporting Information S1).

In addition to the albedo feedback, ET may play an essential role in amplifying Arctic warming (Bonfils et al., 2012; Swann et al., 2010). Our results show stronger ET in summer during the mid-Holocene relative to the PI, with the most prominent increase of evaporation in June and transpiration in July (Figure 3a). The resulting increase in atmospheric water vapor leads to an increase in downward longwave (LW) radiation to enhance the greenhouse effect in summer, further heating the atmosphere (Figure 3a and Figures S2b, S3a, and S4a in Supporting Information S1). Meanwhile, the changes of downward LW radiation follow the changes in temperature rather than the atmospheric water vapor. Because surface temperature changes could generate downward LW radiation responses (Vargas Zeppetello et al., 2019). In contrast, the ET also has a cooling effect. The latent cooling offsets some of the warming, but net warming remains in summer (Figure 3b). Besides, some positive feedback amplifying the warming over land exists. For instance, in response to the warming, air can hold more water vapor increasing the greenhouse effect; upward LW radiation release also increases due to surface warming (Figure 3b). The increase in needleleaf trees with a high evaporative resistance reduces low cloud cover and enhances incoming SW radiation due to increased stability from the warming aloft (Figure S3c in Supporting Information S1; Kim et al., 2020).

To further examine whether water vapor from lower latitudes contributes to the increase in atmospheric water vapor in the Arctic, the effective water vapor import (precipitation minus evaporation) anomalies were calculated following Swann et al. (2010). Evaporation includes transpiration, canopy evaporation, land and ocean evaporation here. Results show an increased net water vapor import (2.41 mm/month) from lower latitudes from August to November due to the orbital forcing, and even more net import (2.70 mm/month) when including the vegetation-climate feedback. However, the net import in summer is negative (−0.80 mm/month) due to vegetation-climate feedback, with a maximum negative net import in July (−2.03 mm/month) (Figure 3e). This export confirms that the increase in column water vapor mainly comes from the local supply (transpiration, canopy evaporation, land and ocean evaporation) within the Arctic region rather than large-scale moisture transport. The water vapor anomalies over the Arctic Ocean and the North Atlantic are mainly transported from the Eurasian and North American continents (Figure S4b in Supporting Information S1). We conclude that decreased albedo and increased ET are primarily responsible for the land warming over the Arctic region in spring and summer due to the vegetation-climate feedback.

3.4. Indirect Influence of Vegetation-Climate Feedback on the Arctic Warming via Sea Ice

Anomalous anticyclones over Northern Europe and the North Atlantic and an anomalous cyclone appear over the Arctic Ocean in summer due to the vegetation-climate feedback, leading to the transport of warm air from Northern Europe and North America to the North Atlantic and the Arctic Ocean (Figure 4b). These circulation changes warm the atmosphere and contribute to the sea ice melt in the region (Figures S2b and S5b in Supporting Information S1). The maximum atmospheric water vapor increase over oceans north of 60°N is in summer, while the increased ocean evaporation is scarce (Figure 3c). Meanwhile, there is a similar anomalous water vapor transport with surface air transport (Figure 4b and Figure S4b in Supporting Information S1). Therefore, we could infer that less water vapor transports to northern high-latitude land. The enhanced downward LW related to this increased water vapor contributes to the warming here (Figure 3c and Figure S3b in Supporting Information S1). The resultant decreased low cloud appears due to increased stability from the warming aloft (Figure S4c in Supporting Information S1), leading to an increase in downward solar radiation to favor the sea ice melt. In return, the decreased surface albedo associated with the sea ice loss absorbs more energy to accelerate sea ice loss (Figure S3b in Supporting Information S1).

The associated sea ice loss occurs throughout the year (Figure 2d). The newly opened ocean water resulting from the reduced sea ice could absorb more solar radiation from April to August (Figure 3d). This extra energy is stored in the upper northern high-latitude oceans. Meanwhile, it is released to the atmosphere mainly via LHF and SHF in the cold season when the ocean becomes a heat source for the atmosphere (Figure 3d and Figure S3b in Supporting Information S1; Deser et al., 2010; Serreze & Barry, 2011; Zheng et al., 2019). Additionally, the newly exposed ocean water is much warmer than the cold sea ice surface, allowing the ocean to release more upward LW radiation, heating the lower troposphere (Figure 3d; Dai et al., 2019), thereby contributing to the Arctic warming throughout the year. Notably, the contribution of upward LW radiation is much more substantial than the LHF and SHF (Figure 3d), because of the large temperature differences between the water and ice surface in the Arctic region (Lindsay & Rothrock, 1994). The increased SAT over sea areas around Greenland

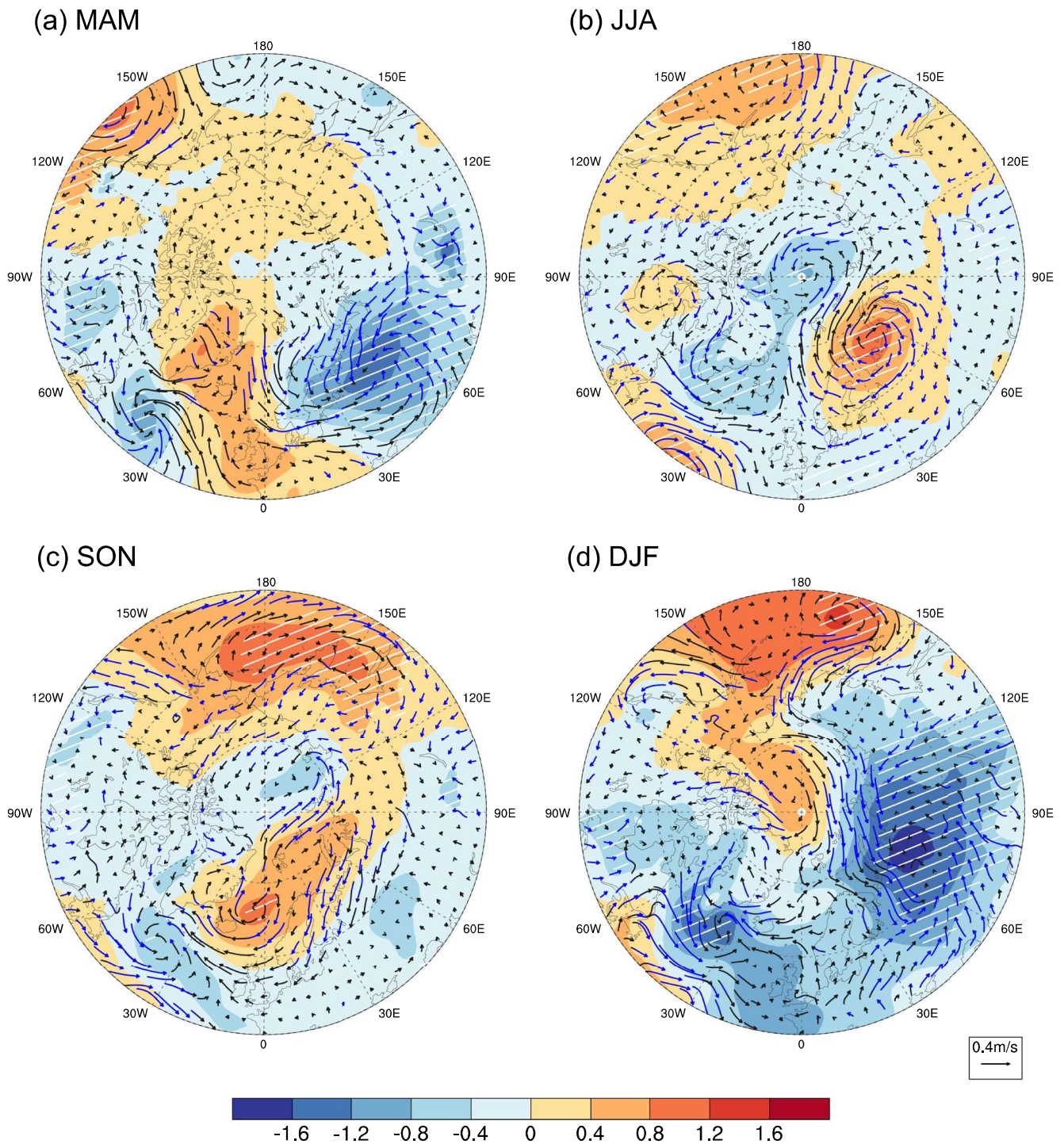


Figure 4. Sea level pressure (color shading, unit: hPa) and surface wind (vector, unit: m/s) anomalies due to vegetation-climate feedback (ΔVeg) during the mid-Holocene in (a) spring, (b) summer, (c) autumn and (d) winter. The white oblique lines and blue vectors indicate that values exceed the 90% confidence level based on a Student's *t*-test.

corresponds well with the portion of sea ice loss (Figures S2 and S5 in Supporting Information S1), confirming the significant contribution of the sea ice loss to the warming over northern high-latitude oceans. Besides, the Arctic oceans release more energy into the atmosphere than they absorb during this process (Figure S3b in Supporting Information S1).

3.5. Indirect Influence of the Vegetation-Climate Feedback on the Arctic Warming via Atmospheric Circulation

The spring experiences the most robust warming over Arctic land and northern high-latitude oceans due to the vegetation-climate feedback (Figures 2 and 3d). In addition to the extra upward LW radiation caused by the sea ice loss and the enhanced LH and SH flux release associated with surface albedo decrease due to the expansion of the trees (Figures 3b and 3d and Figures S2a and S6 in Supporting Information S1), the atmospheric circulation anomalies also contribute to the warming over continental areas in the Arctic. The coastal region of northeastern Canada is heated by anomalous easterlies from the North Atlantic and the Arctic Ocean (Figure 4a). The abnormal low pressure over northern Eurasia associated with the regional warming comes with westerly wind anomalies from the Norwegian Sea and the North Sea and southerly anomalies from lower latitudes with warmer temperature, transporting the anomalous warm air to northern Eurasia, and further promoting regional warming in the area. The anomalous easterlies from the North Pacific to North Asia contribute to transporting warm air to this region (Figure 4a).

The autumn land warming caused by the vegetation-climate feedback mainly occurs in northeastern Canada, Greenland and Fennoscandia. The warming around Greenland due to the sea ice loss could transport relative warmer air to heat Greenland and adjacent areas in northeast Canada (Figure 4c and Figures S2c and S5c in Supporting Information S1). The anomalous anticyclone linked to the sea ice changes over the Barents Sea and the Norwegian Sea may carry relative warmer air to Spitsbergen (Figure 4c). Besides, the contribution of sea ice loss to ocean warming, the increased atmospheric water vapor may also contribute to the warming over the North Atlantic. A weakened westerly water vapor transport from the North Atlantic to northern Eurasia indicates more ocean evaporated water vapor stays in the North Atlantic (Figure 4c and Figure S7a in Supporting Information S1). Together with the increased evaporation, the greenhouse effect amplifies regional warming in the area (Figures S2c and S7b in Supporting Information S1).

An anomalous low-pressure center occurs over the North Atlantic due to local sea ice loss associated with an abnormal high pressure over mid-latitude Atlantic areas, indicating a positive phase of the North Atlantic Oscillation (NAO; Figure 4d). The stronger pressure gradient may drive the storm track further north, causing warm conditions over the North Atlantic and northern Fennoscandia (Figure S2d in Supporting Information S1; Thompson & Wallace, 1998). The positive NAO-like pattern accompanied by an anomalous low pressure over north Eurasia could weaken the Ural high-pressure ridge and the Siberian High, and thus result in the abnormal warm air from lower latitudes to warm the atmosphere and less frequent occurrences of cold surges over Eurasia (Figure 4d and Figure S2d in Supporting Information S1; D. Y. Gong & Ho, 2002; X. Zhang et al., 2012).

According to the results above, the contribution of circulations changes is smaller than the warming caused by albedo/ET changes and sea ice loss. But their specific contributions need to be quantified in the following work. Additionally, different seasons show noticeable high-pressure anomalies in the North Pacific due to the vegetation changes (Figure 4). However, the linkage between the atmospheric anomalies and the vegetation changes and their climate impact still need further study. Further studies using different models are also required for the following work.

4. Conclusions

In this study, we performed four simulations using the global Earth system model EC-Earth to investigate the contribution of vegetation-climate feedback to amplified Arctic (north of 60°N) warming during the mid-Holocene. Our simulations indicate northward expansions of grassland and deciduous needleleaf trees in northern high latitudes during the mid-Holocene compared to the PI, which is roughly consistent with proxy reconstructions. The results show ~14% increase in LAI of low vegetation and ~30% increase in LAI of high vegetation north of 60°N. These changes in vegetation lead to extra Arctic warming and sea ice loss throughout the year. On an annual mean basis, the contribution is 0.33°C (~25% of the increase by comparing the warming due to vegetation-climate feedback and due to orbital forcing) warming and $0.35 \times 10^6 \text{ km}^2$ (~26%) sea ice loss. The direct influence of increased vegetation on this warming owes to decreased albedo and enhanced ET. These two factors are primarily responsible for the Arctic land warming in spring and summer, and the Arctic sea ice loss in summer. The indirect influence of increased vegetation is mainly attributed to sea ice loss and atmospheric circulation anomalies. The SHF and LHF and upward LW radiation associated with the sea ice loss contribute to

the extra warming over the North Atlantic and the Arctic Ocean throughout the year. The atmospheric circulation anomalies related to the increased vegetation and resulting sea ice loss also lead to regional warming in different seasons. This work could provide implications for reducing the uncertainty of simulated temperature and reconciling data-model discrepancy of annual temperature during the mid-Holocene.

Data Availability Statement

The model data to produce the main figures in this paper is accessible at <https://doi.org/10.17043/zhang-2022-holocene-vegetation-1>. The CMIP6 model data can be downloaded at <https://esgf-node.llnl.gov/search/cmip6/>. ERA5 data set can be found at <https://www.ecmwf.int/en/forecasts/datasets/reanalysis-datasets/era5>.

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