

The Influence of Glacial Northern Hemisphere Ice Sheets on Atmospheric Circulation

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Abstract. During the last glacial period, Northern Hemisphere (NH) ice sheets significantly influenced atmospheric circulation, however most studies primarily focused on the North Atlantic. Here, we investigate the influence of ice sheets on global climate during Marine Isotope Stage 3 (MIS 3, 65-25 ka), a period marked by prominent millennial-scale variability. Using the Australian Earth System Model (ACCESS-ESM1.5), we simulate the glacial climate around 49,000 years ago (49 ka). Our findings demonstrate that the NH ice sheets induced a southward shift of the NH westerlies during boreal summer and winter, increasing rainfall over Eurasia during summer but reducing it in winter. The influence of orbital parameters and greenhouse gases alone produces only minor circulation changes. Adding ice sheet albedo intensifies the NH cooling and strengthens the westerlies but does not significantly alter their latitudinal position. The ice sheets' influence also extends across the tropics and the Southern Hemisphere (SH). The most pronounced southward displacements of the Intertropical Convergence Zone (ITCZ) and the NH Hadley cell during austral summer—associated with enhanced Australian rainfall—occur when all ice sheet-related boundary conditions are included. In the SH, the presence of ice sheets leads to an equatorward shift of the SH Hadley cell during JJA, with the largest displacement occurring when all boundary conditions are combined. Additionally, SH westerlies show no latitudinal shift in response to orbital parameters, greenhouse gases, or albedo alone, but exhibit a pronounced equatorward shift when ice sheet topography is included. These findings highlight the non-linear interactions between ice sheets, large-scale atmospheric circulation, and precipitation patterns.

1 Introduction

Earth's climate history has been shaped by the expansion and retreat of Northern Hemisphere (NH) ice sheets, particularly during the last glacial cycle. The expansion of large NH ice sheets has long been associated with global and regional cooling, particularly over the North Atlantic, which is directly influenced by the ice sheet extent and height (Kageyama et al., 2020). Ice sheet growth affects land-ocean geography by lowering sea levels (Lambeck et al., 2014), exposing land bridges, and altering local heating patterns due to the differing heat capacities of land and ocean (Byrne and O'Gorman, 2013). Additionally, changes

in land bridges and oceanic gateways, such as the Bering Strait, influence large-scale ocean circulation, including the Atlantic Meridional Overturning Circulation (AMOC) (Hu et al., 2015). These factors have significant implications for the climate system, particularly through their impact on the atmospheric circulation and ocean currents.

- 25 In today's climate, warm water in the North Atlantic is transported northward via the Gulf Stream, and atmospheric Rossby wave propagation results in a northeastward-tilted Atlantic jet stream. This configuration contributes to the relatively mild winters in Europe compared to regions at similar latitudes in North America (Seager et al., 2002; Brayshaw et al., 2009). However, during past glacial periods, and particularly the Last Glacial Maximum (LGM; 19-21 ka), thick ice sheets covered northern North America and may have reached up to 5 km in height (Peltier, 2004). This so-called Laurentide Ice Sheet (LIS)
- 30 caused a pronounced southward shift of the jet stream, and in some cases, a more zonal orientation compared to today's tilted pattern (Manabe and Broccoli, 1985; Pausata et al., 2011; Löfverström et al., 2014; Ullman et al., 2014). This southward shift of the jet stream, which was primarily driven by ice sheet topography, rather than changes in albedo or greenhouse gas (GHG) forcing (Pausata et al., 2011), had important consequences for European climate, impacting winter storm tracks, precipitation, and seasonal temperatures (Löfverström et al., 2014).
- 35 Previous research on understanding the impact of NH ice sheets on climate (Broccoli and Manabe, 1987; Pausata et al., 2011; Löfverström et al., 2014; Ullman et al., 2014; Wang et al., 2018; Izumi et al., 2023) primarily focused on the LGM, an important period due to its extreme climate conditions, including the coldest global temperatures (Kageyama et al., 2020), lowest atmospheric CO₂ levels (Köhler et al., 2017), and most extensive ice sheets (Peltier, 2004; Clark et al., 2009; Kageyama et al., 2017; Gowan et al., 2021). Most of these studies examined the North Atlantic region, assessing the impact of ice sheets
- 40 on westerlies and the Atlantic jet, with limited investigation into regions beyond this sector. An exception from this list is the work of Löfverström et al. (2014), who investigated the LGM as well as pre-LGM ice sheets of Marine Isotope Stage (MIS) 5b (~88 ka) and MIS 4 (~66 ka). Their study found that the westerlies shifted southward in all ice sheet configurations, but a fully zonal jet orientation emerged only during the LGM. They also suggested that the ice sheet topography dynamically induced warming in Alaska and central Asia, potentially contributing to the westward expansion of the Eurasian ice sheets from MIS 4
- 45 to the LGM. Other studies, such as Ullman et al. (2014) and Zhang et al. (2014a), further explored the impact of varying LIS height on the atmospheric-oceanic system.

- In contrast, the climate of MIS 3 (65–25 ka) has received comparatively less attention in the context of ice sheet–atmosphere interactions. Existing studies of MIS 3 have largely concentrated on large-scale oceanic responses, such as AMOC variability, surface temperature patterns, and sea ice changes (Malmierca-Vallet and Sime, 2023; Brandefelt et al., 2011; Merkel
- 50 et al., 2010; Guo et al., 2019). Some have investigated tropical climate variability, including changes in ENSO behaviour during this period (Brandefelt et al., 2011; Merkel et al., 2010), while others briefly mention reduced global precipitation, a southward-shifted ITCZ, intensified Southern Hemisphere (SH) westerly wind stress, and strengthened NH trade winds alongside weakened SH trade winds (Guo et al., 2019). However, the detailed, mechanistic impacts of NH ice sheets on atmospheric circulation during MIS 3—particularly beyond the North Atlantic—remain underexplored. This gap may be partly
- 55 due to the significant millennial-scale climate variability that characterizes MIS 3—including abrupt temperature fluctuations

in Greenland of 8–10°C (Huber et al., 2006) and multiple episodes of AMOC weakening (Meniel et al. (2020) and references therein)—which make it challenging to isolate equilibrium climate response to boundary conditions.

To address this, we focus on the period around 49 ka within MIS 3, which stands out due to one of the highest obliquity values (24.43°) of the last glacial cycle (Berger, 1978), resulting in greater summer insolation in both hemispheres compared to the preindustrial (PI). In particular, 49 ka exhibits an obliquity value higher than some interglacial periods, such as the Last Interglacial (LIG, 127 ka), yet features a climate state marked by low CO₂ (~199 ppm, similar to LGM (Köhler et al., 2017)) and extensive Laurentide and Scandinavian ice sheets (Gowan et al., 2021), leading to a colder-than-PI climate. Furthermore, global sea levels were approximately 60-65 m lower than PI (Shakun et al., 2015). This combination of factors makes 49 ka an intriguing period for assessing how ice sheets influenced atmospheric circulation and hydroclimate on a planetary scale.

Here, we use the Australian Earth System Model (ACCESS-ESM1.5) to examine the influence of glacial ice sheets on global atmospheric circulation, focusing on their cascading effects across Eurasia, the tropics, and the SH. Our analysis emphasizes the role of the ice sheets in modulating Rossby waves, Hadley circulation, and the Intertropical Convergence Zone (ITCZ), shaping rainfall patterns in both hemispheres, and influencing regional climates, such as over Eurasia and northern Australia. In addition, we examine how the inclusion of the larger MIS 3 ice-sheets compared to PI influence SH westerlies.

2 Methods

2.1 Model description

We employ the Australian Community Climate and Earth System Simulator-Earth System Model version 1.5 (ACCESS-ESM1.5), a comprehensive Earth system model developed for Coupled Model Intercomparison Project Phase 6 (CMIP6) (Ziehn et al., 2020). The atmospheric model is the UK Met Office Unified Model version 7.3 (UM7.3) (Martin et al., 2010; The HadGEM2 Development Team et al., 2011) with a horizontal resolution of 1.875° × 1.25° and 38 vertical levels. The land surface is represented by the Community Atmosphere Biosphere Land Exchange (CABLE) model version 2.4 (Kowalczyk et al., 2013), which includes biogeochemical processes through the CASA-CNP module (Wang et al., 2010). The ocean component consists of the NOAA/GFDL Modular Ocean Model version 5 (MOM5) (Griffies, 2012), featuring a horizontal resolution of 1° × 1°, with the meridional spacing refined to 0.4° in the Southern Ocean and 0.33° near the equator, and with 50 vertical levels. Furthermore, sea ice processes are simulated using the Los Alamos National Laboratory Consortium Model for Sea Ice Development (LANL CICE) version 4.1 (Hunke et al., 2010). The coupling between the atmosphere, ocean and sea ice components is achieved through the Ocean Atmosphere Sea Ice Soil–Model Coupling Toolkit (OASIS-MCT) (Craig et al., 2017). The ocean carbon cycle dynamics are modeled using the nutrient-phytoplankton-zooplankton-detritus (NPZD) scheme in the World Ocean Model of Biogeochemistry and Trophic dynamics (WOMBAT) model for ocean biogeochemistry (Oke et al., 2013).

This model has previously been used to simulate the last interglacial (lig127k) time slice as part of the Tier 1 Paleoclimate Model Intercomparison Project 4 (PMIP4)-CMIP6 experiments (Yeung et al. (2021), MIS9e (336-321 ka; Duboc et al. (2025)), and mid-Holocene (6 ka; Mackallah et al. (2022)).

	PI	49 ka
Orbital parameters		
Eccentricity	0.01674	0.01292
Obliquity (°)	23.459	24.435
Perihelion-180 (°)	100.33	62.451
Greenhouse Gases		
CO ₂	284.3	199
N ₂ O	273	237
CH ₄	284.3	432

Table 1. Orbital parameters and Greenhouse gas concentrations for PI and 49 ka climate.

Experiment	Model years	Years analysed	Orbital parameters and GHGs	Albedo and vegetation	Salinity	Topography and runoff
49ka-co	0 to 618	569 to 618	49ka	PI	PI	PI
49ka-alb	297 to 569	520 to 569	49ka	49ka	PI	PI
49ka-full	824 to 1584	1485 to 1584	49ka	49ka	49ka	49ka

Table 2. Boundary conditions used in the different 49 ka experiments. Model years indicate the total number of years that the corresponding experiment has run for as shown in the time series of Figure A1.

2.1.1 Experimental setup

90 The pre-industrial (PI) simulation is the 1000-year piControl run based on the year 1850 CE as described by Ziehn et al. (2020). We analyse the average of the last 100 years of this simulation (Figure A1, black line). Orbital parameters and GHG concentrations for the PI configuration are provided in Table 1.

To investigate the climate around 49 ka, the boundary conditions are changed in a step-wise manner using the ACCESS-ESM1.5 model. The first experiment, 49ka-co, simulates a baseline climate with orbital parameters (Berger, 1978) and GHG concentrations (Köhler et al., 2017) being set to 49 ka values (Table 1). 49ka-co is run for 618 years (Figure A1, cyan line).

The model uses prescribed Plant Functional Types (PFTs) that remain fixed during each individual simulation. We modified the PFT distributions in selected simulations to reflect the altered boundary conditions. For instance, in a second step, 49ka-alb, we incorporate the ice sheet extent from Gowan et al. (2021) corresponding to 52.5 ka, adopting the maximal scenario with an ice-capped Hudson Bay (Figure 1d). This time period was chosen because the simulated ice sheet at 52.5 ka most closely matches the sea-level estimates for 48–52 ka, based on global seawater $\delta^{18}\text{O}$ stack records (Shakun et al., 2015). The NH ice

sheets at 49 ka include two separate ice sheets over the North American continent—Laurentide in the east and Cordilleran in the west—as well as the Greenland Ice Sheet and Fennoscandian and Eurasian ice sheets (Figure 1d). Their extent and height are smaller than during the LGM (not shown), which featured a single, extensive LIS over North America and continuous ice sheet coverage from Eurasia to Scandinavia, fully covering the Barents Sea. The Antarctic Ice Sheet (AIS) at 49 ka also has a smaller extent compared to that of the LGM, especially over the Ross and Weddell Seas. At 49 ka, the LIS was up to 3 km thick, while the AIS was up to 3.6 km thick. In the present day, permanent ice cover in the NH is restricted to Greenland (Figure 1c).

In the 49ka-alb experiment, additional modifications are introduced at different stages (Figure A1, orange line). The vegetation cover is modified to reflect ice coverage by removing vegetation in ice-covered regions and converting needle leaf evergreen forests in areas between the Cordilleran and LIS to bare soil (Figure 1a,b). These modifications align with the vegetation reconstruction from Allen et al. (2020) for 52 ka. Additionally, C3 crops are removed, and replaced by the next two most dominant vegetation types in the corresponding grid cell. For example, in India, C3 crop values are replaced by broadleaf deciduous and C4 grass. In Europe and eastern North America, they are replaced by broadleaf deciduous and C3 grass, while they are replaced by broadleaf deciduous over China. 49ka-alb (Figure A1, orange line) starts from year 297 of 49ka-co, and was run for a total of 272 years with the above mentioned albedo and vegetation changes (Table 2). As can be seen in Figure A1, 49ka-alb is not in complete equilibrium. To this 49ka-alb experiment, we add a gradual increase in surface salinity from year 547 (Figure A1, magenta line) to account for the 44 m sea level drop relative to PI, based on the prescribed ice sheet, achieving a global average increase of +0.33 psu from PI levels. At year 729, the Bering Strait is closed.

Our final experiment, referred to as 49ka-full (Figure A1, blue line, Table 2), starts at year 824, when the topography is adjusted to reflect conditions at 49 ka, based on Gowan et al. (2021), including an increase in ice sheet height and changes to the land-sea mask, such as the closure of the Bering Strait (done in the previous step), Hudson Bay, the Sahul and Sunda shelves, and the Tasman Gateway (Figure 1). Nearest-neighbour interpolation is applied to determine vegetation types on newly exposed land. The river runoff is also adjusted at year 1258, to reflect the 49 ka topography. In total, 49ka-full is run for 760 years, with 325 years incorporating all the boundary conditions. In the Results section, we compare the average of the last 100 years of 49ka-co, last 50 years of 49ka-alb, and the average of the last 100 years of the 49ka-full experiment (Table 2, Figure A1, darker colors).

The ITCZ position is determined using the precipitation centroid index (Braconnot et al., 2007), which identifies the latitude that evenly divides the tropical precipitation amount between 20°S and 20°N. This approach effectively represents the center of mass of tropical rainfall distribution. The strength of the northern Hadley cell, ψ , is determined by the maximum value of the atmospheric mass streamfunction within the 400–600 hPa pressure range, while the strength of the southern Hadley cell is determined by its minimum value over the same vertical pressure range (Equation 1).

$$\psi = \frac{2\pi R \cos \phi}{g} \int_0^p v dp \quad (1)$$

where, R is the radius of the earth, ϕ is latitude, g is gravitational acceleration, v is the meridional wind and p is pressure.

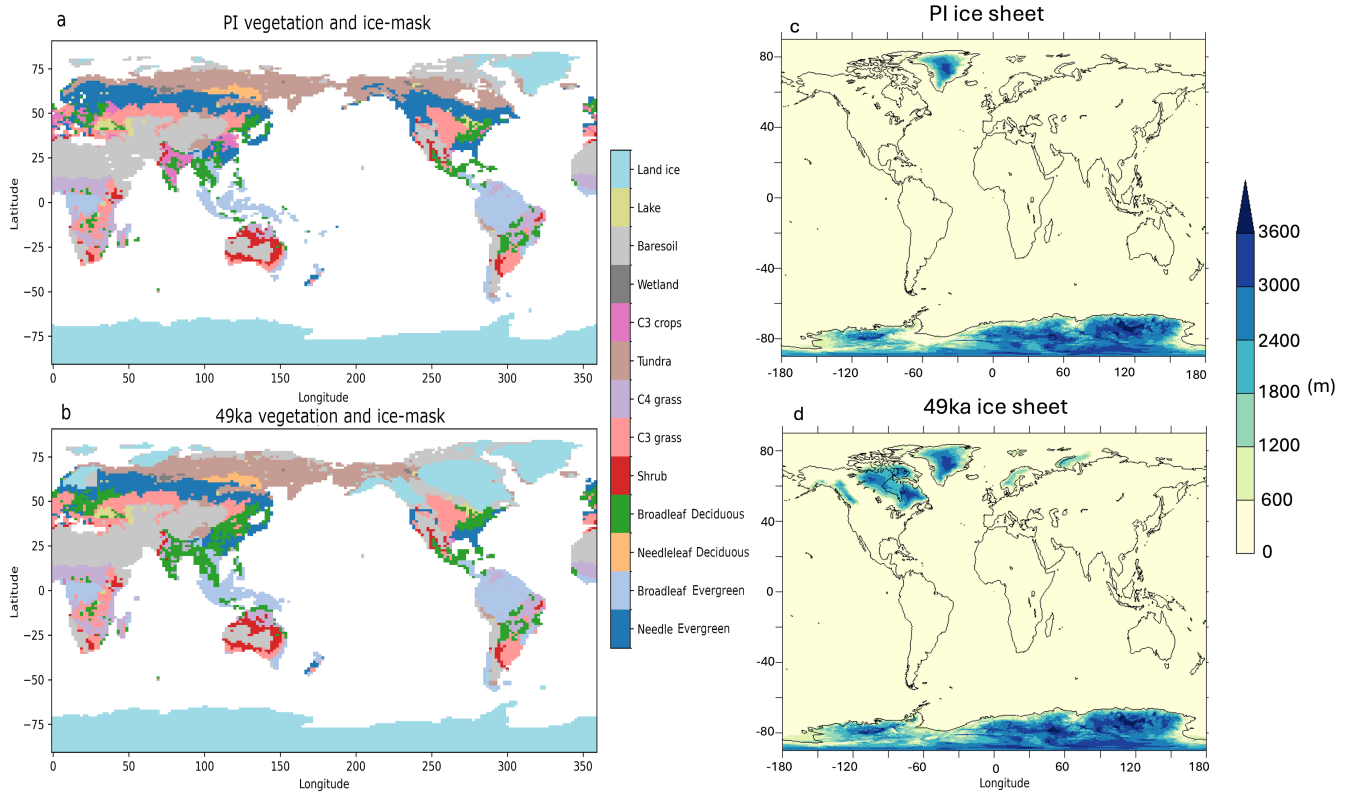


Figure 1. (a,b) Vegetation and ice mask changes at 49 ka compared to PI. The color indicates the dominant Plant Functional Type (PFT) of each grid cell. (c,d) Ice sheet thickness (m) and extent from (Gowan et al., 2021) for PI and 49 ka (represented by 52.5 ka as described in the text)

3 Results

135 3.1 Simulated glacial climate at 49 ka compared to PI

We first describe the 49ka-full experiment in comparison to PI. At 49 ka, the higher obliquity (24.43°) compared to PI (23.45°) results in slightly increased summer insolation in both hemispheres (Figure A2). Despite the higher obliquity, the glacial simulation (49ka-full) exhibits substantial oceanic and atmospheric cooling due to lower GHG concentrations and extensive ice sheet expansion at this time (Figure 2). Global annual mean surface air temperatures (SATs) at 49 ka are 2.7°C lower than PI (Table A1), with the NH cooling by 4.9°C and the SH by 2°C (Figure 2a). The most pronounced cooling is simulated in ice sheet regions and northern high latitudes. Seasonal variations are evident in the NH (Table A1), where temperatures drop by 6.4°C in winter (DJF) and 2.6°C in summer (JJA). Over SH, DJF (austral summer) SATs are 1.7°C lower, while JJA (austral winter) SATs are 2.4°C lower (Table A1).

Global mean ocean temperature is 1°C lower, and global annual mean sea surface temperature (SST) (Figure 2b) is reduced by 1.2°C (Table A1). The simulated SATs and SSTs anomalies show a good agreement with proxy records indicating a qualitative estimate of the changes between Greenland Interstadial (GI 13; 49ka-50ka) and PI climate (Figure 2a,b). Higher SSTs in the Labrador region are linked to a stronger AMOC in 49ka-full, with a transport rate of 31 Sv (about 10 Sv greater than the PI value). This strengthening is attributed to topographical changes and the presence of ice sheets, which enhance westerly winds over Baffin Bay and weaken them over the Labrador Sea (Figure A3). These wind changes modify the North Atlantic gyres, leading to an overall strengthening while shifting the separation between the subpolar and polar gyres southward and orienting the subpolar gyre in a more zonal direction (Figure A3). A strengthening of the cyclonic circulation in the Labrador Sea is also simulated, which increases the westward transport of warm water from south of Iceland to the Labrador Sea and further strengthens the AMOC. The stronger AMOC also influences sea ice extent. During boreal winter, the sea-ice edge at 49 ka expands in the North Pacific and Nordic Sea sectors but retreats in the Labrador Sea due to the influx of warm water (Figure 2b, orange dashed line compared to black dashed line). In boreal summer, the 49 ka sea-ice edge is similar to PI across most regions, except in the Nordic seas, where the sea-ice edge extends further south than in PI by 5 to 6° latitude (Figure 2b, orange solid line compared to black solid line). In the SH, austral summer sea ice retreats in the South Atlantic as stronger North Atlantic Deep Water (NADW) formation leads to a deep-ocean warming of about 0.3°C (not shown), and this warm water gets upwelled in the South Atlantic (Figure 2b, orange dashed line compared to black dashed line). However, the austral winter sea ice edge remains similar to PI in the South Atlantic while expanding in the Indian and Pacific sectors of the Southern Ocean (Figure 2b, orange solid line compared to black solid line).

Drier conditions are simulated at 49 ka compared to PI over NH high-latitude land regions, particularly over the LIS, due to colder conditions and reduced atmospheric moisture availability (Figure 2c). However, precipitation increases in the North Atlantic and Eurasia, between 30°N and 60°N, as well as over India and East Asia. In contrast, most land areas in the SH display decreased precipitation, except over northern Australia, parts of Southeast Africa, and over eastern Brazil.

3.2 Impact of 49 ka boundary conditions on the circulation changes over North Atlantic and Eurasia

At 49 ka (experiment 49ka-full), the presence of large ice sheets over North America significantly alters the atmospheric circulation, particularly in the North Atlantic region. In DJF, the westerlies over the North Atlantic strengthen due to the strong meridional temperature gradient and shift by ~6° southward compared to PI (Figures 3a,c and 4a), with a more zonal flow replacing the northeast-southwest tilt of the jet stream (Figure 5). This pattern is evident both at the surface and in the upper troposphere (Figure A4). Geopotential height (GH) differences (Figure 3e) indicate that the LIS induces a high-pressure system (not shown) over North America, initiating a Rossby wave response. This Rossby wave causes a southward shift of the Icelandic low and a weakening of the Azores High, contributing to the zonal orientation of the westerlies (Figures 3c, 5, and A5). The southward shift and zonal jet alter moisture-bearing storm tracks, which in PI were tilted southwest-northeast, to a more zonal configuration at 49 ka. This reorganization of pressure systems (not shown), with a negative GH anomaly extending from the North Atlantic to eastern Eurasia (Figure 3e), alters precipitation patterns, enhancing precipitation over the

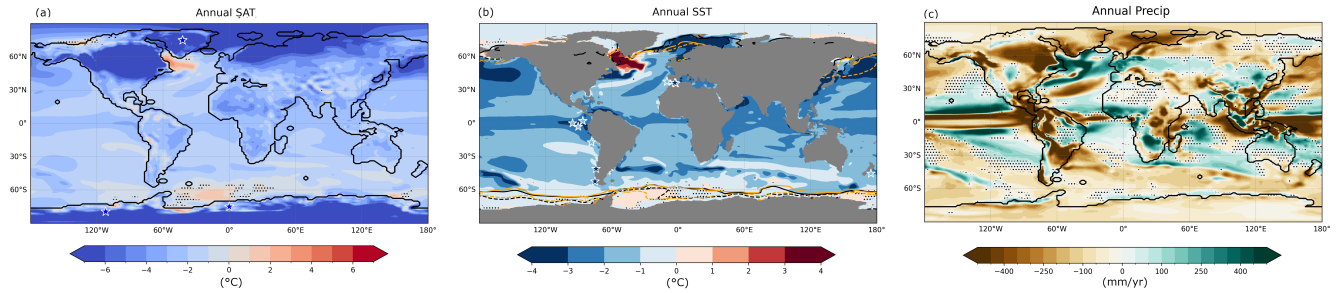


Figure 2. Annual mean anomalies of surface air temperature (SAT), sea surface temperature (SST), and precipitation for 49 ka climate (exp 49ka-full) compared to PI. Contours show seasonal mean 15% sea ice concentration for (orange) 49 ka and (black) PI during DJF (dashed) and JJA (solid). Stippling indicates non-significant changes based on a Student's t-test at 95% confidence level. Stars in subpanels (a) and (b) indicate qualitative estimates of the changes between Greenland Interstadial (GI 13) and PI climate, as inferred from the proxy records (MDO1-2444, ODP-977A, MD95-2043, MD01-2443, MD95-2042 (Martrat et al., 2007); ODP-1233 (Kaiser et al., 2005); TR163-19 (Lea et al., 2000); RC13-110 (Feldberg and Mix, 2003); ODP846B (Martínez et al., 2003); TG-7 (Calvo et al., 2001); MD97-2120 (Pahnke et al., 2003); MD07-3128 (Caniupán et al., 2011); EPICA Dronning Maud Land (EPICA community members, 2006); WAIS (WAIS Divide Project Members, 2015); NGRIP (Huber et al., 2006; Kindler et al., 2014); SO188-17286-1 (Lauterbach et al., 2020). Dark (light) blue colors represent much colder (slightly colder) climate.

North Atlantic and western Europe (Figure 3g). However, colder conditions over Eurasian landmasses lead to drier conditions inland in DJF, despite the persistence of negative GH anomalies (Figure 3a,e,g).

The westerlies are also shifted southward in JJA, with the flow remaining more zonal than during PI (Figures 3d, 5, and 180 4b), coinciding with a positive GH anomaly over the Labrador Sea region while a negative GH anomaly forms over Europe (Figure 3f). These GH anomalies enhance the westerly flow and contribute to moisture convergence (not shown) toward the Eurasian landmass, increasing precipitation along the westerly band in the North Atlantic Ocean and over parts of Eurasia (Figure 3d,f,h). In addition, mild temperature conditions over Eurasia (Figure 3b) further contribute to the increased precipitation. Precipitation also increases north of 60°N in Asia (Figure 3h) due to higher specific humidity (not shown) resulting from 185 warmer conditions and the development of a negative GH anomaly (Figure 3b,f). The warming in this region arises from the topographical changes corresponding to 49 ka, where ocean points have been converted to land. Since land has a lower heat capacity than the ocean, it warms more rapidly in summer, amplifying the temperature-driven moisture increase. Conversely, north of 60°N in Europe precipitation decreases. This is due to the influx of colder north-easterly winds and the presence of a blocking high over the Labrador-Greenland sector, which suppresses rainfall in these areas.

190 During DJF, strong cooling in the 49ka-co experiment reduces moisture availability, leading to widespread drying over NH landmasses (Figure 6a,c). The westerlies are slightly enhanced and shift southward by only $\sim 1^\circ$ (Figure 4a, 9a). This shift is accompanied by a low-pressure anomaly over the North Atlantic and a high-pressure system over Europe, which enhances subsidence and suppresses precipitation over Europe (Figure 6b,c). This pattern is consistent with the overall drying trend in the NH, except for a small increase in precipitation over the North Atlantic Ocean (Figure 6c), likely due to storm activity

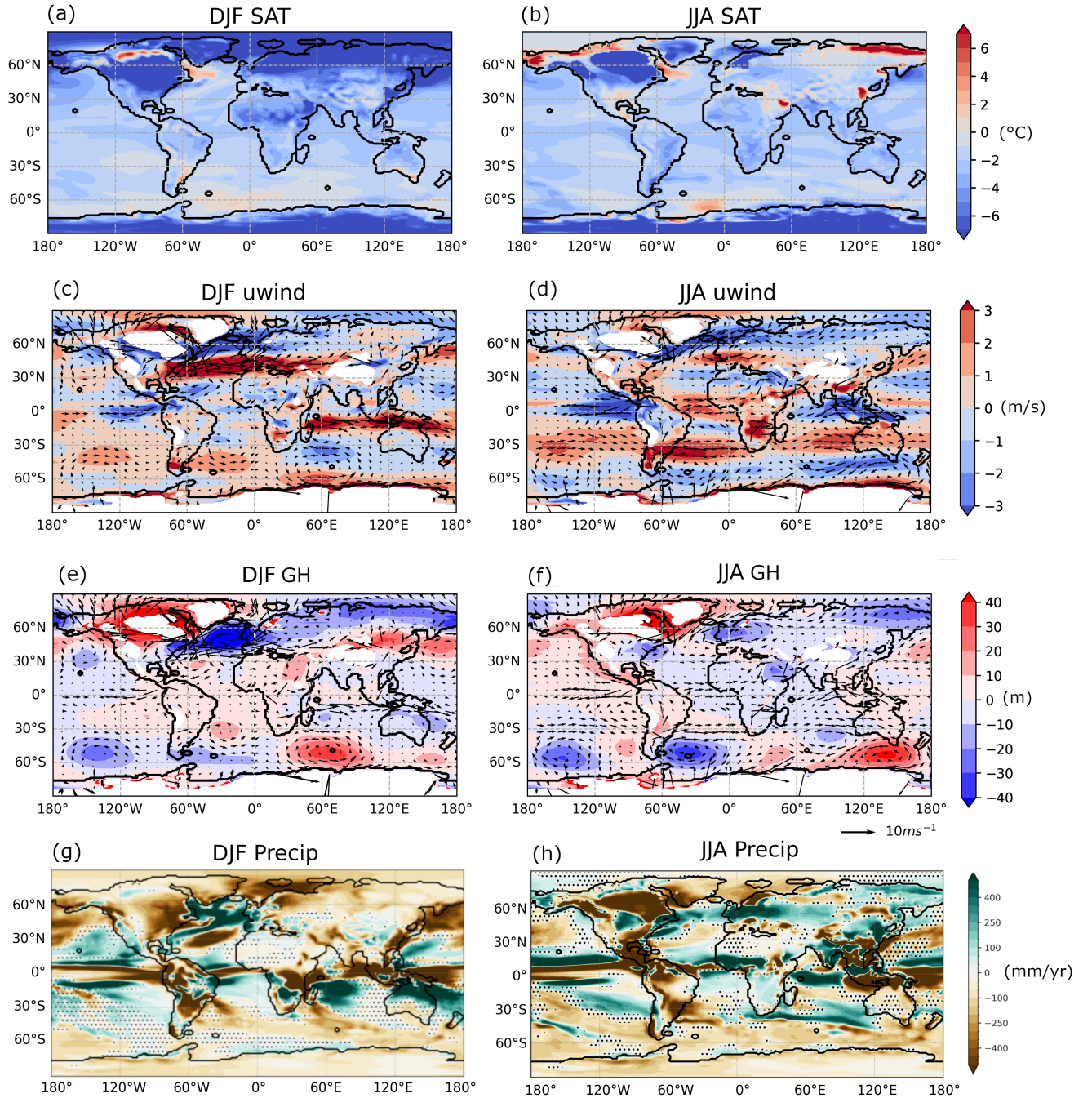


Figure 3. Anomalies between 49ka-full and PI for (a,b) SAT (°C), (c,d) zonal winds at 850 hPa (m/s) overlaid with wind vectors (m/s), (e,f) geopotential height at 850 hPa (m) with zonal mean removed and overlaid with wind vectors (m/s), and (g,h) precipitation (mm/yr) during (left) DJF and (right) JJA. Stippling in subpanels (g) and (h) indicates non-significant changes based on a Student's t-test at 95% confidence level.

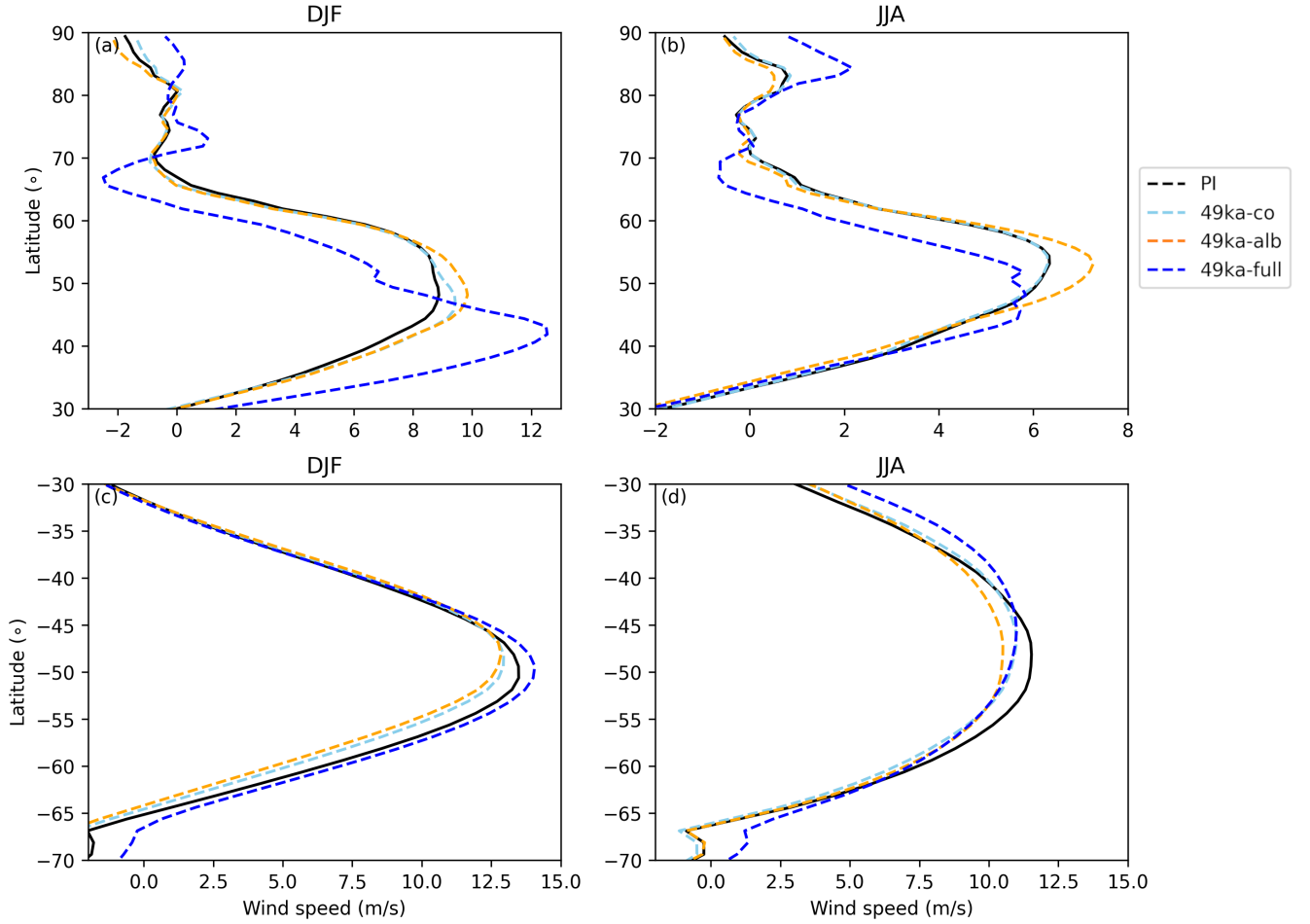


Figure 4. Zonal winds (m/s) averaged (top) over the North Atlantic (80W:0E) and (bottom) averaged over all longitudes representing westerlies over the SH during (a,c) DJF and (b,d) JJA seasons.

195 in this region. In 49ka-alb, changing the vegetation and albedo has little impact compared to 49ka-co, except over northern North America (Figure 6d,f), where cooling is enhanced by 8° and drying increases by 25% due to the albedo change over the LIS. The westerlies intensify further compared to 49ka-co, but their position shifts back to the PI configuration (Figure 4a). The GH response features a low-pressure anomaly over the Hudson Bay and Labrador Sea, which is consistent with increased precipitation in this region (Figure 6e,f).

200 In JJA, the 49ka-co experiment shows no change in westerly strength and position over the North Atlantic (Figure 7b, 4b 9b) and modest temperature and precipitation changes over the NH landmass (Figure 7a,c). In contrast, the 49ka-alb experiment exhibits pronounced drying over the Laurentide region and northern Europe, driven by intense cooling from albedo changes

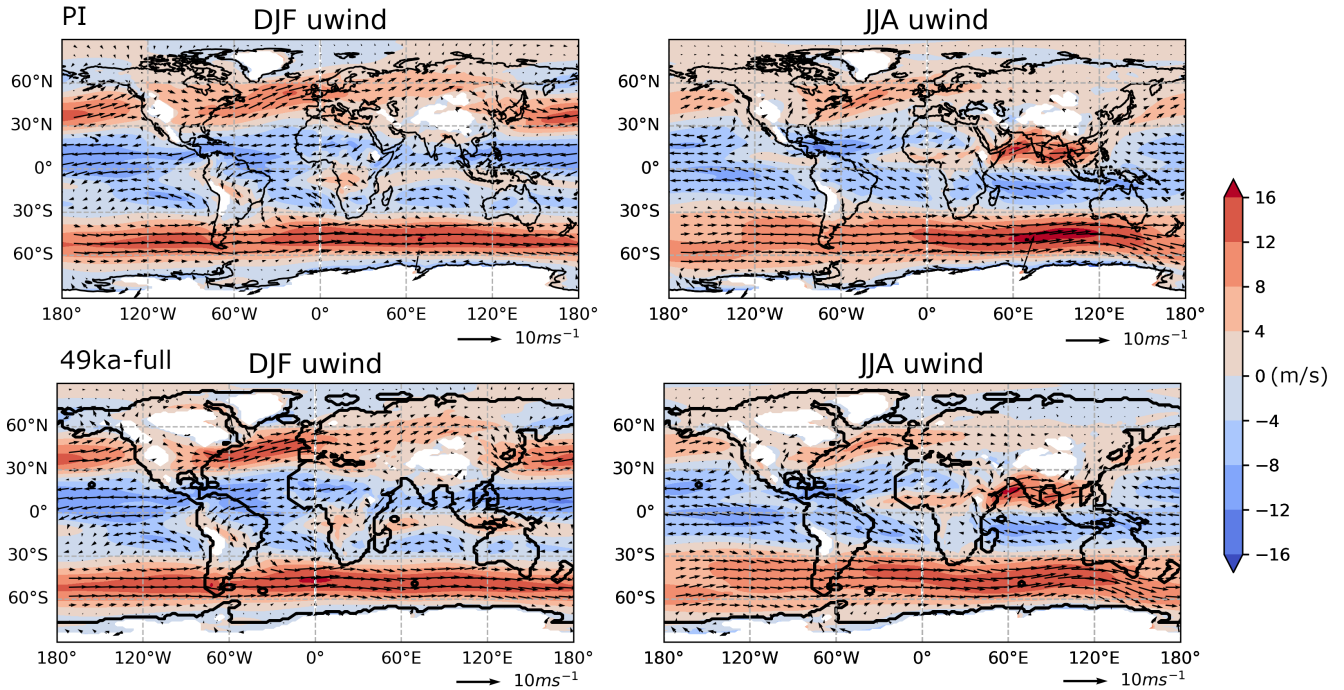


Figure 5. Zonal winds at 850 hPa (m/s) overlaid with wind vectors (m/s) in (left) DJF and (right) JJA for (top) PI and (bottom) 49ka-full.

and persistent high pressure over these regions (Figure 7d-f). While the westerlies strengthen in 49ka-alb, their position remains unchanged (Figure 4b, 9b). This intensification leads to slightly wetter conditions inland over Eurasia in the 49ka-alb scenario.

205 Figures 6, 7, 4a, and 9 show that only the full implementation of the ice sheet, including its topography, can induce a significant planetary wave response in the atmosphere, with southward-shifted North Atlantic westerlies. This response results in substantial drying over the northern high-latitude landmasses, including the Laurentide region during DJF, along with increased milder and wetter conditions over Eurasia during JJA.

3.3 Impact of 49 ka boundary conditions on tropical and SH atmospheric circulation.

210 In DJF, the presence of the LIS forces a southward shift of the Ferrell cell. This displacement further leads to a southward migration of the NH Hadley cell's southern edge, shifting from 7.5°S in PI to 11°S in 49ka-full (Figure 8a, see zero contour), and results in a globally averaged ITCZ shift of approximately 1.5° southward in DJF (Figure 3g, 9a). Additionally, the NH Hadley cell strengthens by 20 Sv relative to PI. The southward shift of the NH Hadley cell and ITCZ during DJF is also captured in 49ka-co (Figure 8c), where the NH Hadley cell strengthens by +11 Sv relative to PI. The shift in the ITCZ is
 215 driven by peak summer insolation in the SH due to higher obliquity at 49 ka. However, the magnitude of change in 49ka-co is smaller than in 49ka-full (Figure 8a, 9a), with an ITCZ shift of 0.6°. Implementing albedo changes in 49ka-alb enhances the Laurentide cooling, further strengthening the NH Hadley cell (by +8 Sv), albeit without additional displacement (9.5°S in 49ka-

49ka-co - PI

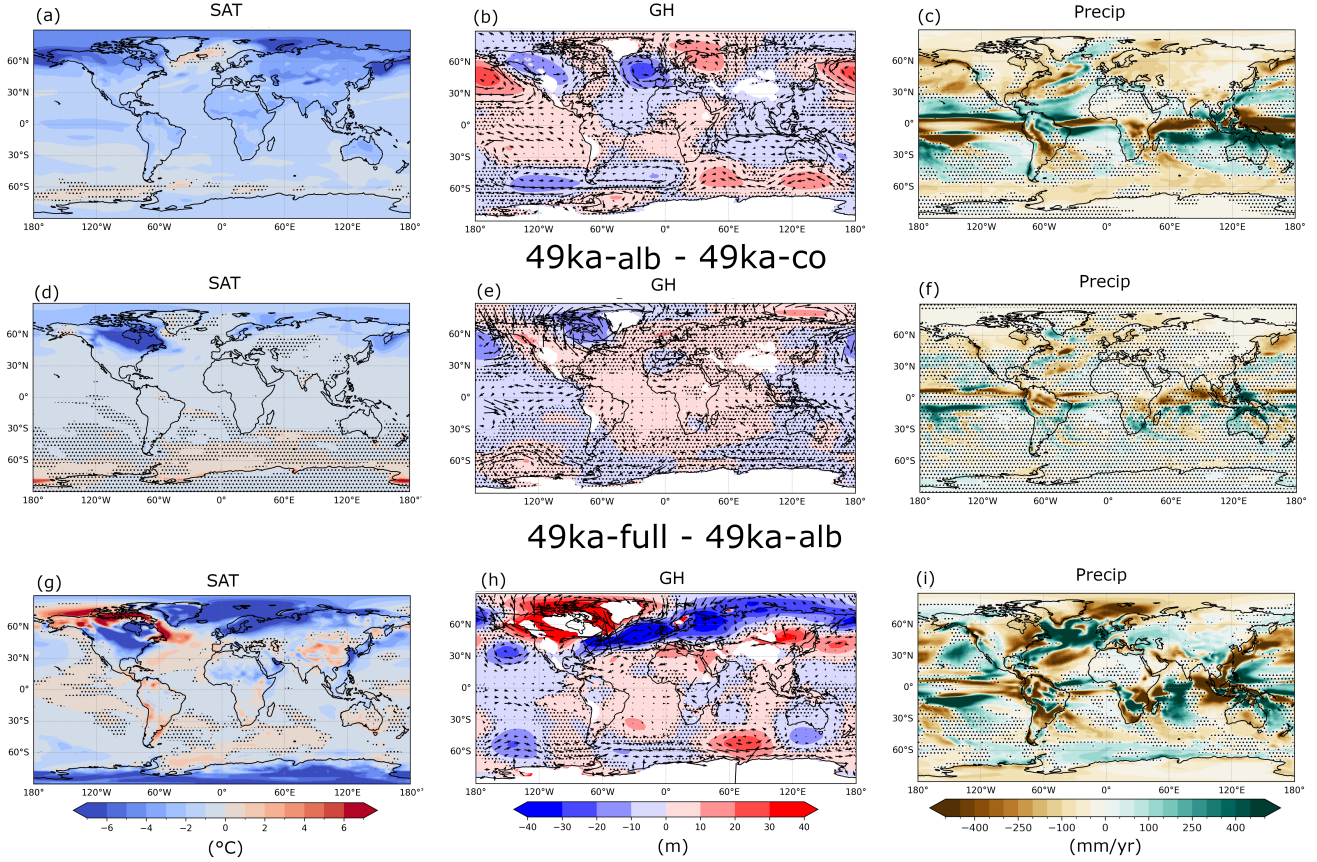


Figure 6. DJF (a,d,g) SAT (°C), (b,e,h) geopotential height (m) overlaid with winds at 850 hPa and precipitation (mm/yr) anomalies for (a-c) 49ka-co minus PI, (d-f) 49ka-alb minus 49ka-co, (g-i) and 49ka-full minus 49ka-alb. Stippling indicates non-significant changes based on a Student's t-test at 95% confidence level.

co vs. 9.6°S in 49ka-alb) (Figure 8c,e). The minor shift in the NH Hadley cell and ITCZ position explains why lower latitude precipitation changes remain insignificant between 49ka-alb and 49ka-co (Figure 6c,f). However, the inclusion of topography in 49ka-full results in a slight additional strengthening of the NH Hadley cell and a more substantial southward ITCZ shift (by 1.5° relative to 49ka-alb), due to the stronger temperature gradient between the two hemispheres (Figure 9a, Table A1).

During JJA, the southern edge of the SH Hadley cell shifts northward by 2° in 49ka-full, moving from 32°S in PI to 30°S, along with a northward shift of the Southern Ferrel cell (Figure 8b, see zero contour). This shift causes a slight northward migration of the globally averaged ITCZ (Figure 3h), while the SH Hadley cell strengthens by 35 Sv. In 49ka-co, a slight warming over the North Atlantic and stronger cooling in the SH shifts the SH Hadley cell's southern edge to 31°S, accompanied by a strengthening of +22 Sv (Figure 8d, 9b). The changes in 49ka-alb do not alter the SH Hadley cell's latitudinal position

49ka-co - PI

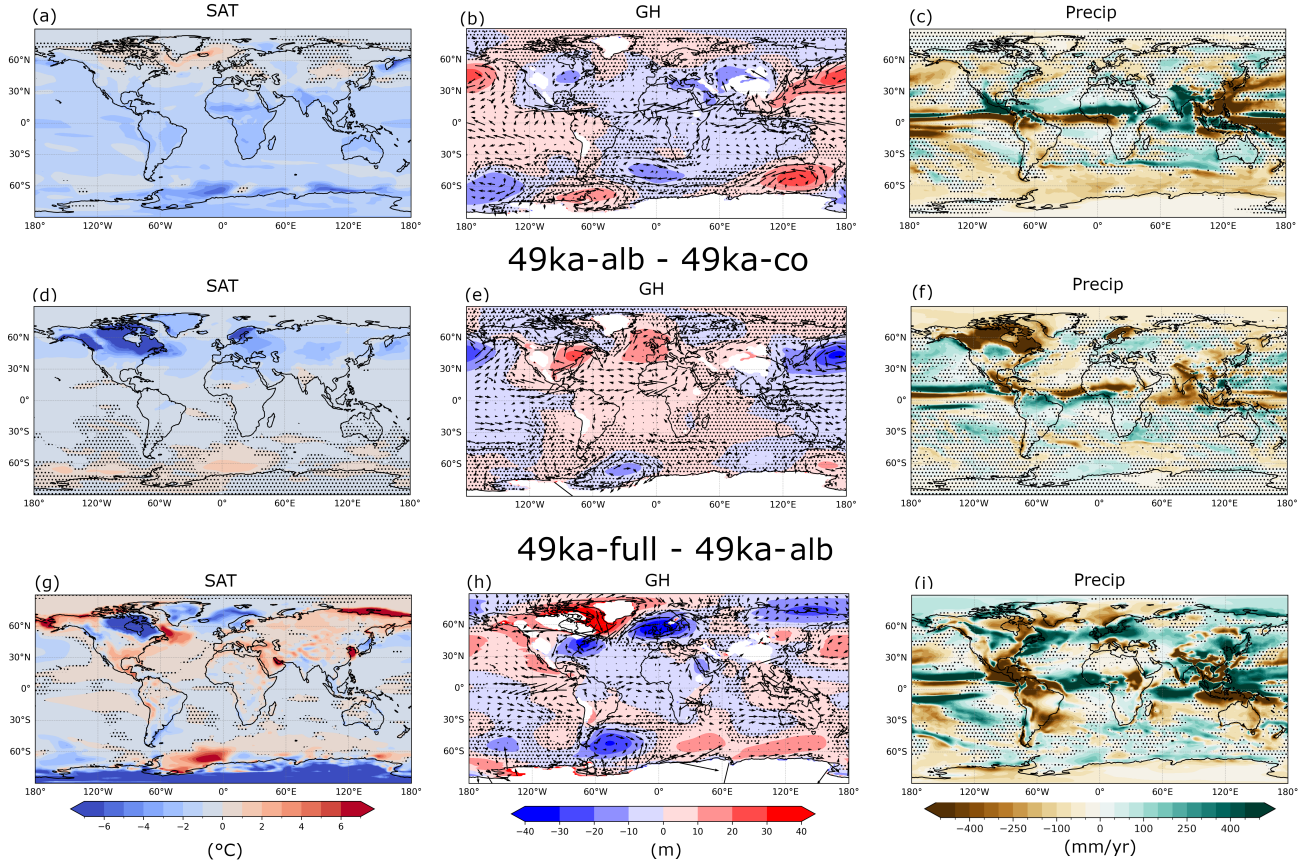


Figure 7. Same as Figure 6 but for JJA.

significantly but weaken its strength to PI levels (Figure 8f, 9b), due to intensified Laurentide cooling. As a result, precipitation changes in the lower latitudes remain small (Figure 7c,f). Temperature contrasts between hemispheres are reduced by ice sheet topography (Table A1), which induces localized warming in Siberia and strong Antarctic cooling, and by changes in the land-sea mask with the replacement of ocean grid cells by land.

This reduced thermal gradient with warming in the NH leads to a slight northward ITCZ shift, further strengthening the SH Hadley cell and shifting it an additional 1° northward, toward the warmer hemisphere, relative to 49ka-alb.

The SH westerlies also exhibit distinct responses across the simulations. In DJF, they weaken slightly in 49ka-co but weaken further in 49ka-alb, accompanied by a small northward shift (Figure 4c). In contrast, in 49ka-full, they remain in the same position as in PI but strengthen due to the increased height and extent of the AIS. During JJA, the SH westerlies weaken across all 49 ka simulations compared to PI, with the strongest weakening occurring in the experiment where albedo is changed

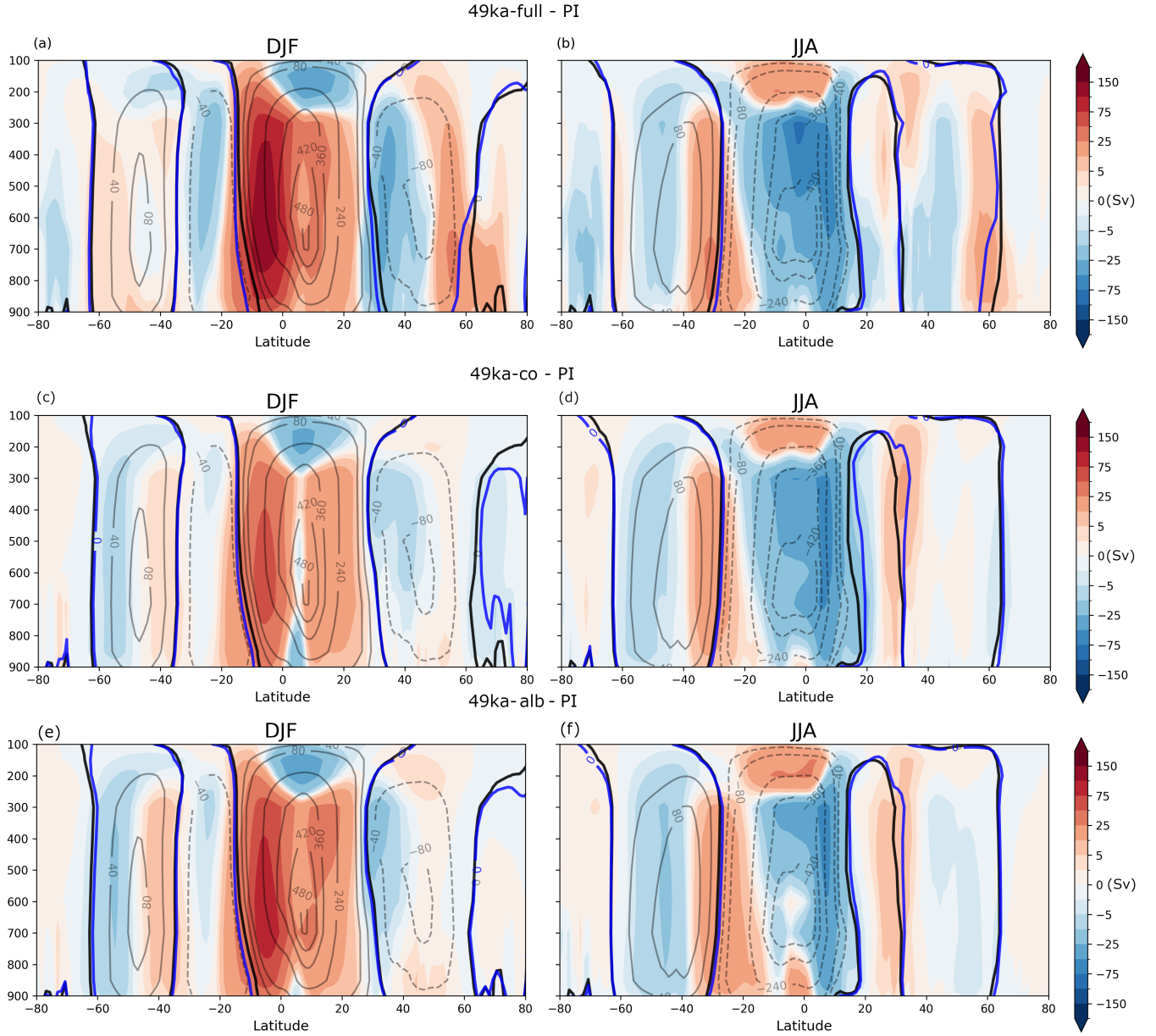


Figure 8. Atmospheric mass streamfunction (Sv) anomalies (shading) between (a,b) 49ka-full and PI, (c,d) 49ka-ic and PI, and (e,f) 49ka-ice and PI, for (left) DJF and (right) JJA. Contours (solid=positive; dashed=negative) are PI (black) absolute streamfunction values. The thick black and blue lines represent the zero contour for PI and 49ka experiments, respectively.

(Figure 4d). However, in 49ka-full, they shift northward by 2.5°, a displacement linked to the northward shift of the SH Hadley cell in 49ka-full (Figure 4d, 9b).

3.4 Impact of 49 ka boundary conditions on Australian rainfall

240 As shown in Figures 3, 9a, the implementation of 49 ka boundary conditions leads to a notable increase in Australian rainfall during the monsoon season. This section focuses on the large-scale processes driving this regional rainfall response. While a detailed analysis of monsoon changes in other regions lies beyond the scope of this study, our results underscore the broader influence of NH ice sheets on hemispheric-scale atmospheric dynamics.

The presence of NH and larger AISs induces substantial atmospheric circulation changes by modifying the Hadley and
245 Ferrel cells and influencing the westerlies in both hemispheres. These shifts propagate through the atmosphere, altering surface pressure systems and triggering a hemispheric-scale response.

During DJF, the NH ice sheet forces a southward shift of the Hadley and Ferrel cells (Figure 8a), enhancing subsidence over East Asia around 40°N and leading to drier conditions (Figure 3e,g). This southward displacement of the NH Hadley cell extends across hemispheres, impacting Australian rainfall by shifting convergence southwards along with the ITCZ (Figure 8a).
250 These changes explain the wetter conditions in northern Australia and drier conditions over the Maritime Continent (Figure 3g). A comparable but weaker response is simulated in the 49ka-co experiment (Figure 6c). In contrast, changes in albedo do not impact Australian precipitation significantly (Figure 6f). However, modifying the topography induces wetter conditions in northern Australia driven by shifts in the Hadley cell and ITCZ, while southern Australia experiences drier conditions (Figure 6h,i).

255 During JJA, in response to the 49 ka boundary conditions, most of Australia experiences drying (Figure 3h), primarily due to the establishment of blocking high-pressure systems south of the continent (Figure 3f). These high-pressure systems reinforce subsidence and restrict moisture transport into the region (not shown). However, the drying response is not significant over central Australia and remains less pronounced along the eastern and western edges of the continent. The southwestern part of Australia becomes wetter, while the southeastern region experiences drying, due to the northward displacement of
260 the westerlies, resembling a negative winter Southern Annular Mode (SAM). Significant drying also occurs across northern Australia and much of the Maritime Continent south of the equator.

When only orbital parameters and GHGs are modified, the SH Hadley cell shifts northward and strengthens. However, in this scenario, the weakening of the SH westerlies and the southward displacement of subtropical high-pressure systems result in increased rainfall over southern Australia (Figure 7b,c). Introducing albedo changes alone has little impact on Australian
265 climate (Figure 7d-f). However, when ice sheet topography is incorporated, the northward shift in SH westerlies linked with the strengthening and equatorward displacement of the SH Hadley cell causes enhanced subsidence and thus drier conditions over Australia, coinciding with the northward ITCZ migration (Figure 7i, 4d). Compared to the PI climate, this drying effect is somewhat mitigated, as the westerlies — despite their northward shift— remain weaker overall.

4 Discussion

Our simulations of the 49 ka glacial climate reveal a complex interplay of ice sheet topography, GHG forcing, and orbital configurations. The results underscore how large-scale climatic features, such as atmospheric circulation patterns, precipitation distribution, and ocean circulation, respond to glacial boundary conditions. Additionally, much of the existing research has focused primarily on the North Atlantic region, driven by the greater availability of paleoclimate data from this area and the presence of extensive North American ice sheets that strongly influenced both regional and global climate. Our results add to this literature by demonstrating that the influence of glacial ice sheets extends well beyond the North Atlantic, with significant climatic impacts also occurring in other regions, including the SH.

Here, the simulated state in 49ka-full experiment represents an interstadial climate. The global cooling in 49ka-full relative to PI (a 2.7°C reduction in global SATs) aligns with the expected glacial forcing associated with lower GHG concentrations and extensive ice sheet expansion. The pronounced winter cooling in the NH, exceeding 6°C, is consistent with PMIP4 simulations of LGM climate, with ice sheet-driven albedo changes amplifying high-latitude cooling during glacial periods (Kageyama et al., 2020). This magnitude of cooling from PI is also broadly consistent with other MIS 3 modelling studies, though direct comparisons are not possible as applied boundary conditions between other MIS 3 studies and ours were different.. For example, Guo et al. (2019) found a 2.9°C cooling for 38 ka time slice using the Norwegian Earth System Model (NorESM) model. Simulations using the Community Climate System Model version 3 (CCSM3) model suggest stronger cooling, with Merkel et al. (2010) reporting a 3.4°C drop at 35 ka, and Zhang et al. (2014b) estimating 3.5°C at 38 ka.

Changes in AMOC

Proxy records suggest significant AMOC variability during the MIS 3 period, with stronger AMOC during the onset of interstadials (Böhm et al., 2015; Henry et al., 2016; Menviel et al., 2020), and weak AMOC during stadials. The processes driving this AMOC variability is still debated (Menviel et al., 2020). Model simulations also show considerable spread in AMOC strength across different MIS 3 timeslices and model setups. For instance, the CCSM3 simulation at 35 ka, initialized from the LGM, showed a ~40% reduction in AMOC strength (Merkel et al., 2010), while the same model produced a ~50% reduction at 44 ka, also initialised from the LGM state (Brandefelt et al., 2011). In contrast, NorESM simulations indicate a ~13% strengthening of the AMOC at 38 ka (Guo et al., 2019). The simulated AMOC in our 49ka simulation is strong (31 Sv) with a ~47% increase relative to PI. This strengthening is driven by changes in the LIS, which enhances westerly winds over Baffin Bay while weakening them over the Labrador Sea, intensifying the cyclonic circulation in the Labrador Sea and facilitating the transport of warm, saline waters to the region. This process is also associated with sea ice retreat in polar waters, further reinforcing the AMOC. These findings align with earlier modelling studies emphasizing the role of topography and ocean-atmosphere interactions in AMOC enhancement (Pausata et al., 2011; Zhang et al., 2014a; Hu et al., 2015; Brown and Galbraith, 2016; Klockmann et al., 2018; Sherriff-Tadano et al., 2018; Guo et al., 2019). For example, Pausata et al. (2011) demonstrated that the presence of an extensive ice sheet during the LGM enhanced AMOC by altering atmospheric circulation patterns. However, their simulations also showed that the cooling effects from lower GHG concentrations and increased albedo

suppressed AMOC strength. In our 49ka-full experiment, these cooling effects appear to be offset by high insolation, allowing topographical impacts to dominate.

Evidence based on foraminifer assemblages and mineral grains in ice-rafted debris also suggests a northward retreat of sea ice and higher temperatures in the Labrador Sea between ~56 and 49 ka, potentially driven by the insolation maxima (Griem et al., 2019). After 49 ka, sea ice variability increased, with near-perennial sea ice cover during most stadials and seasonal ice-free conditions during interstadials. However, it remains uncertain whether interstadials reflect a mean state or are primarily linked to DO variability.

Changes in NH westerlies

The simulated changes in the North Atlantic westerlies and associated precipitation patterns during DJF highlight the critical role of ice sheet topography in driving atmospheric circulation (Figure 9a). At 49ka, the surface westerlies shift southward and adopt a more zonal flow (Figure 3), aligning with previous studies (Löfverström et al., 2014; Ullman et al., 2014; Pausata et al., 2011; Kageyama et al., 2020; Stadelmaier et al., 2024), which emphasize the influence of glacial ice sheet topography on jet stream reorganization. Our results further expand on this, demonstrating that the high-pressure system induced by the LIS initiates a Rossby wave response, shifting the Icelandic low southward and weakening the Azores High. These circulation changes lead to reduced precipitation over NH landmasses during DJF.

During JJA, the changes in westerlies and pressure result in increased rainfall over Eurasia (Figure 9b), with the jet stream extending further inland. Additionally, this rainfall increase is supported by milder land temperatures and enhanced moisture availability (not shown). Our findings on seasonal precipitation changes are consistent with those of Löfverström et al. (2014), as the ice sheet configuration used in our study closely resembles their MIS 4 ice sheet setup. However, while their study reports a full zonal orientation of the jet stream only during the LGM, our results demonstrate that this change can also occur with a MIS 3 ice sheet configuration, suggesting a dynamic response to glacial ice sheets earlier in the glacial period.

Changes in tropical atmospheric circulation and SH westerlies

The presence of ice sheet topography leads to a more pronounced southward shift of the NH Hadley cell during DJF and a stronger northward shift of the SH Hadley cell during JJA, compared to our simulations that include only GHG concentrations, orbital parameters, and an ice mask at 49 ka (Figure 9a,b). This enhanced Hadley cell shift in 49ka-full drives a more substantial southward displacement of the ITCZ during DJF (Figure 9a). Consequently, the northern tropics experience drier conditions, while precipitation increases in the southern tropics, consistent with the mechanisms proposed by Chiang and Bitz (2005).

During JJA, LIS topography induces warming over Siberia, in agreement with previous studies (Liakka and Lofverstrom, 2018; Bakker et al., 2020), which attribute this warming to large-scale atmospheric circulation changes driven by pressure differences linked to LIS height. In contrast, AIS topography enhances cooling over Antarctica during austral winter. This interhemispheric contrast, i.e. relatively warmer conditions in the NH and colder conditions in the SH, pushes the ITCZ northward and strengthens the SH Hadley cell. Additionally, the SH Hadley cell northward shift is associated with the northward displacement of the SH westerlies (Figure 9b), consistent with Ceppi et al. (2013), who demonstrated that interhemispheric teleconnections link ITCZ migration with changes in Hadley cell strength and SH jet positioning. Likewise, NH cooling during

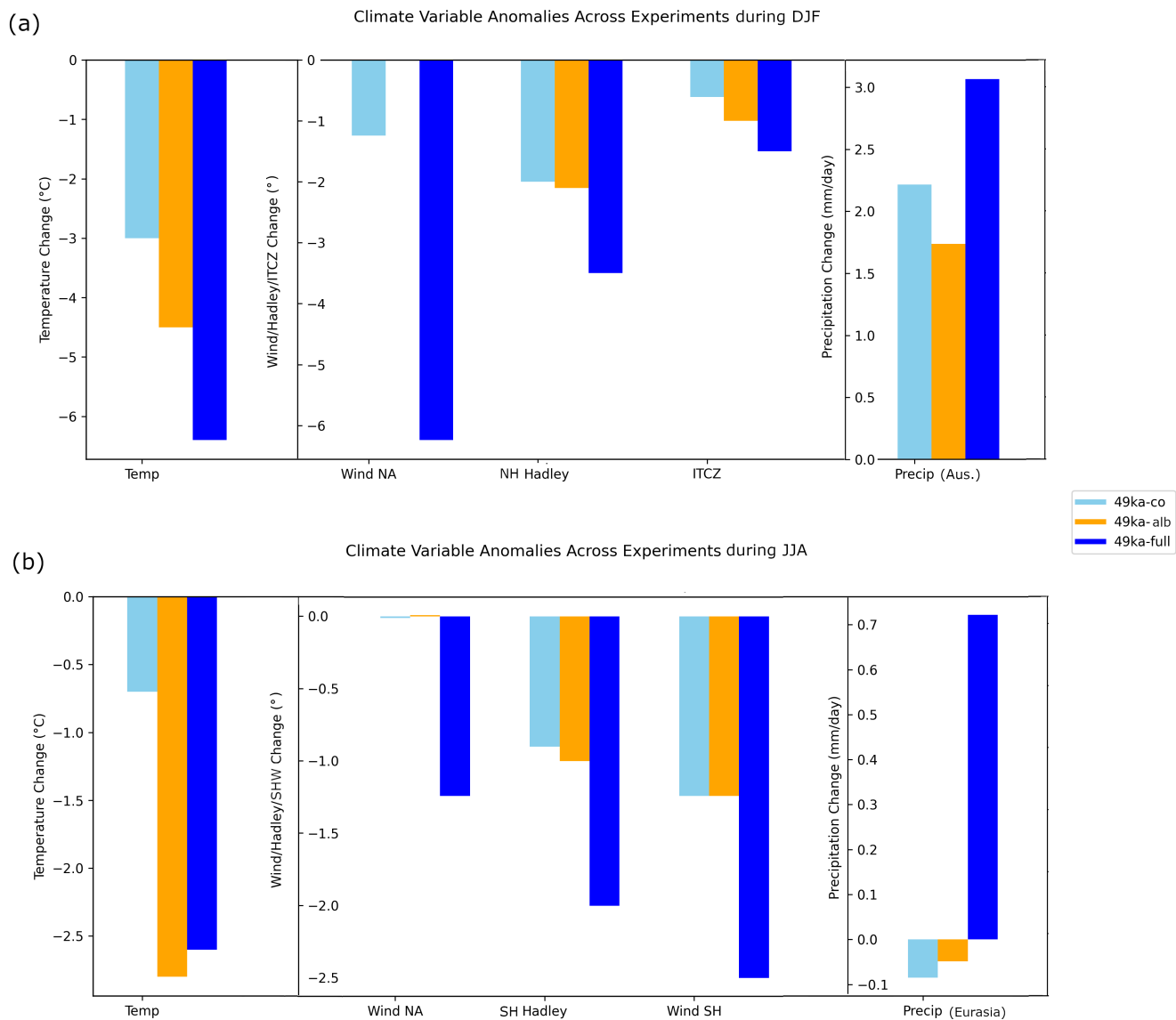


Figure 9. Summary of key atmospheric responses across experiments. (a) DJF anomalies: (left) NH SAT (°C); (middle) latitudinal shifts (°) in westerlies averaged over the North Atlantic (80°W:0°E, 30°N:90°N), NH Hadley cell's southern edge, and globally averaged ITCZ position; and (right) northern Australian precipitation (116°E:152°E, 25°S:12°S, mm/day). (b) JJA anomalies: (left) NH SAT (°C); (middle) latitudinal shifts (°) in westerlies averaged over the North Atlantic, SH Hadley cell's southern edge (°), westerlies averaged over the SH; and (right) precipitation changes over Eurasia (60°W:80°E, 48°N:60°N). Results are shown for the 49ka-co (cyan), 49ka-alb (orange), and 49ka-full (blue) experiments, with all values presented as anomalies relative to PI. Negative values in latitudinal shifts indicate an equatorward shift.

DJF drives a southward shift of the North Atlantic westerlies, influencing the Hadley cell strength and position, and pushing the ITCZ closer to the equator.

These circulation shifts generate regional impacts. Increased subsidence over East Asia creates a high-pressure anomaly, leading to drier conditions, while the southward-shifted ITCZ enhances convergence over Australia, increasing rainfall. The
340 larger shift in the ITCZ and the NH Hadley cell coincides with the maximum increase in Australian rainfall (Figure 9a).

The weakening of the SH westerlies during both DJF and JJA under a high-obliquity climate (49ka-co and 49ka-alb) compared to a low-obliquity climate (PI) aligns with findings from Timmermann et al. (2014). However, the increased height and extent of the AIS counteract this weakening during DJF, leading to SH westerly intensification. Additionally, our results indicate that NH westerlies strengthen with albedo changes relative to orbital parameter and GHG adjustments, whereas SH
345 westerlies weaken under the same conditions (Figure 4). Introducing topographical changes induces an equatorward shift in both NH and SH westerlies during their respective winter seasons (Figure 4a,d).

Overall, our study reveals nonlinear interactions between large-scale atmospheric circulation, surface westerlies, and precipitation patterns under different climate boundary conditions (Figure 9). Orbital forcing, GHGs, albedo, and ice sheet topography collectively shape these interactions in a nonlinear manner.

350 **MIS 3 hydroclimate in Australia**

Proxy records from marine sediments, lake sediments, and river systems, show that Australia experienced spatially variable climates during MIS 3, with greater-than-modern mean water availability peaking between ~ 49 and ~ 40 ka (Kemp et al., 2019). Wet conditions, especially in central and northern Australia between ~ 50 and ~ 45 ka, are linked to an intensified Australian monsoon driven by the summer insolation peak. Our model results align with these observations, showing increased
355 annual precipitation over much of Australia at 49 ka, with particularly intensified monsoonal rainfall in the northern regions (Figure 2c, 3g). This increase in rainfall is consistent with a southward displacement of the ITCZ due to the intensified NH Hadley cell, driven by increased SH insolation during high obliquity. However, our results also reveal that the southward shift of the ITCZ and Hadley cell occur in the 49ka-full simulation (which includes ice sheet topography) as well as in the 49ka-co and 49ka-alb simulations (which lack ice sheet topography). Despite this, the intensified Australian monsoon rainfall is more
360 pronounced in the 49ka-full experiment, as the presence of ice sheets induces larger shifts of the Hadley cell and ITCZ.

Our simulations generally show an insignificant response in southern Australia at 49 ka, with some drying, which contrasts with the increased rainfall seen in southern Australia indicated by paleo-records from a variety of archives, including speleothems, lake sediments, river systems, and offshore sediments (Kemp et al., 2019). In their study, the strengthening or northward movement of the southwesterly winds is linked to increased rainfall in southern Australia, a response that does not
365 align with our simulation. However, our results are somewhat consistent with Weij et al. (2024), in which speleothem growth rates from southwest and southeast Australia indicate drier-than-modern conditions around 49 ka.

5 Conclusions

Our results underscore the far-reaching effects of MIS 3 ice sheets on global atmospheric circulation and regional hydroclimate. We demonstrate that the presence of NH ice sheets during MIS 3 strengthened and shifted the North Atlantic westerlies equatorward by $\sim 6^\circ$ in DJF, leading to increased precipitation over the North Atlantic and Eurasian landmass in boreal summer. These ice sheet-induced changes extend meridionally, with a southward shift of the NH Hadley cell and ITCZ during DJF, resulting in increased Australian rainfall. Although our experiments do not isolate the individual roles of the Laurentide and AISs, simulations with MIS 3 topography suggest that the combined ice sheet configuration also contributes to an equatorward shift of the SH westerlies and SH Hadley cell. Overall, our findings underscore the importance of including realistic MIS 3 boundary conditions when assessing ice sheet–atmosphere interactions on atmospheric circulation.

Data availability. All final data from the modelling simulations will be published on the DRYAD repository under the corresponding author's account (<https://datadryad.org/search?utf8=%E2%9C%93&q=himadri+saini>) and made publicly available upon acceptance of the manuscript.

Appendix A

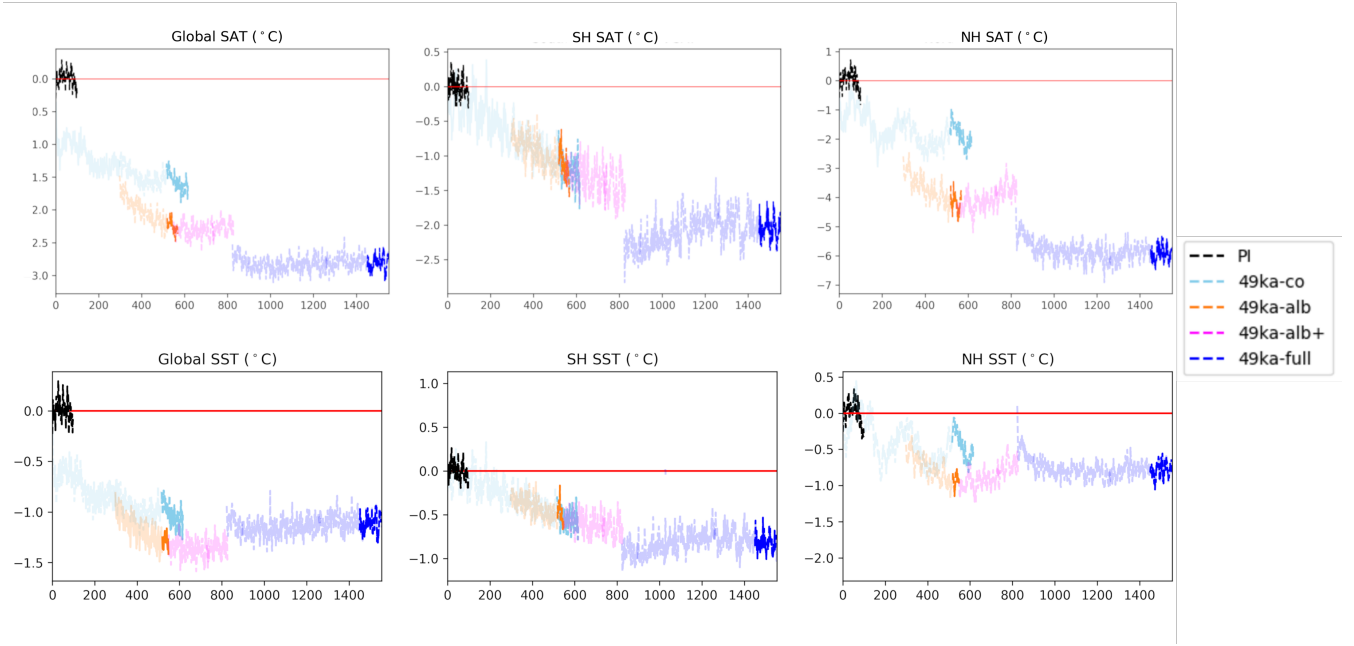


Figure A1. Time series of anomalies of annual mean (left) globally, (middle) southern hemisphere (SH), and (right) northern hemisphere (NH) averaged surface air temperature (SAT) and sea surface temperature (SST) for (black) PI, (cyan) 49ka-co, (orange) 49ka-alb, and (blue) 49ka-full compared to PI. 49ka-co is orbital parameters+GHGs, 49ka-alb is 49ka-co with albedo and vegetation changes. Additional changes to 49ka-alb are shown in magenta colors (49ka-alb+). At year 547, surface salinity is gradually increased to account for the 44 m sea level drop relative to PI, based on the prescribed ice sheet, achieving a global average increase of +0.33 psu from PI levels. At year 729, the Bering Strait is closed. These experiments are not analysed in the manuscript. At year 824, the 49 ka topography is implemented. River runoff is adjusted to reflect the 49 ka topography at year 1258. Darker colors indicate the time-averaged period for each experiment used in the analysis. The horizontal red line is the zero line.

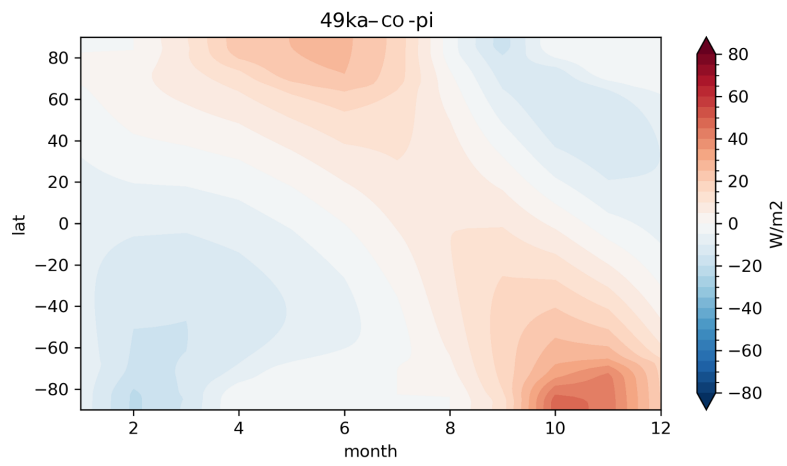


Figure A2. Insolation (Shortwave radiation at the top of the atmosphere) anomalies (W/m^2) between 49 ka and PI as a function of latitude and month.

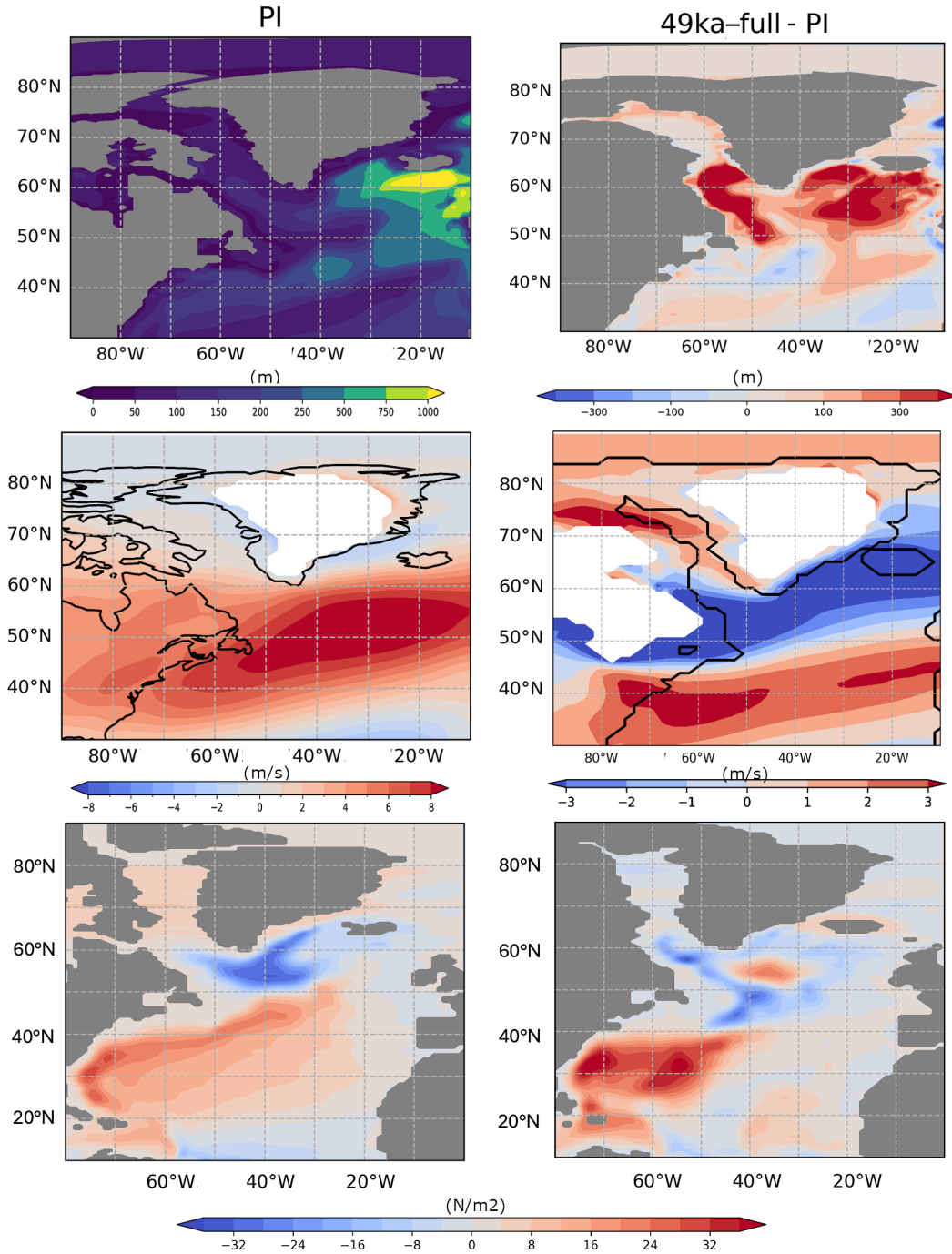


Figure A3. (top) Winter mixed layer depth (m), (middle) Zonal winds (m/s) at 850 hPa, and (bottom) North Atlantic gyre strength (N/m²) for (left) PI and (right) 49ka-full compared to PI. The white areas in the middle panel mark grid cells with continental ice, and for the 49ka-full simulation, continental outlines are shown based on the adjusted land-sea mask.

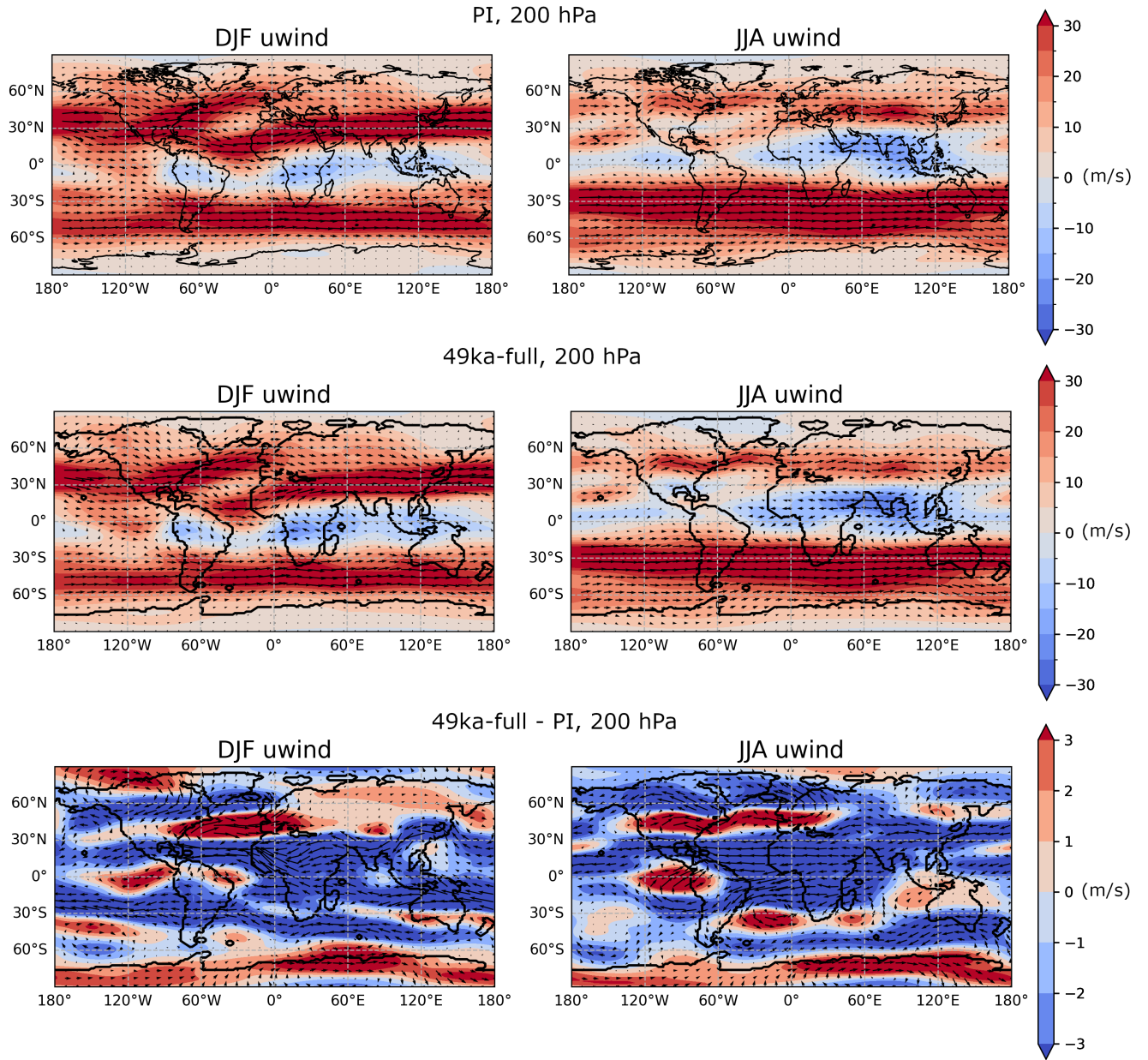


Figure A4. Zonal winds at 200 hPa (m/s) overlaid with wind vectors (m/s) in (left) DJF and (right) JJA for (top) PI and (middle) 49ka-full, and (bottom) 49ka-full minus PI.

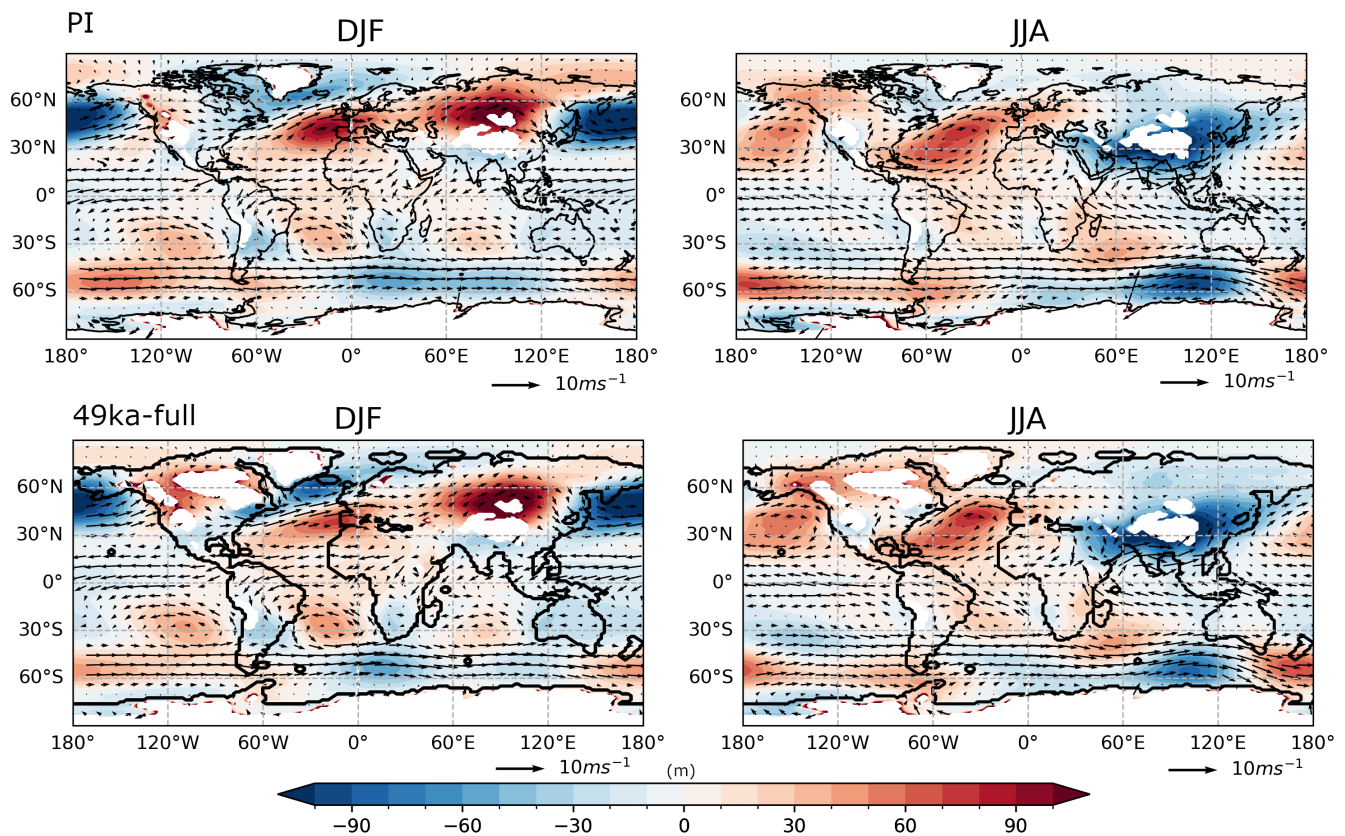


Figure A5. Geopotential height (m) at 850 hPa with zonal mean removed and overlaid with wind vectors in (left) DJF and (right) JJA for (top) PI and (bottom) 49ka-full.

Experiments	PI	49ka-full	49ka-alb	49ka-ic
Annual mean SAT (°C)	15.41	12.71 (2.7)	13.11	13.05
DJF NH SAT (°C)	-4.75	-11.15 (6.4)	-9.25	-7.75
JJA NH SAT (°C)	14.95	12.35 (2.6)	12.15	14.25
DJF SH SAT (°C)	2.85	1.15 (1.7)	2.15	1.95
JJA SH SAT (°C)	-4.85	-7.25 (2.4)	-6.45	-6.45
Annual mean Ocean Temp (°C)	4.32	3.3 (1.02)	3.49	3.62
Annual mean SST (°C)	18.95	17.75 (1.2)	17.65	17.75

Table A1. Global mean values for PI, 49ka-full, 49ka-alb, and 49ka-ic experiments. The values in the brackets show differences from PI.

380 *Author contributions.* HS, LM, JRB, and RND designed the study. HS performed the modelling simulations with assistance from DKH. HS conducted the analysis, interpreted the results and wrote the manuscript with input from LM. JRB, RND, DKH and YD provided valuable comments on the manuscript.

Competing interests. At least one of the (co-)authors is a member of the editorial board of *Climate of the Past*

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