ZEMBA v1.0: An energy and moisture balance climate model to investigate Quaternary climate

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Abstract. A **Z**onally-Averaged **E**nergy and **M**oisture **BA**lance Climate Model (**ZEMBA**) is introduced as a simple and computationally efficient tool for studies of the glacial-interglacial cycles of the Quaternary. The model is based on an energy balance model comprising an atmospheric layer, a land component and a two-dimensional ocean transport model with sea ice. In addition, ZEMBA replaces temperature with moist static energy for calculations of diffusive heat transport in the atmospheric

5   layer and includes a hydrological cycle for simulating precipitation and snowfall. Prior to coupling with an ice sheet model, we present and evaluate equilibrium simulations of the model for the pre-industrial period and the Last Glacial Maximum, using prescribed land ice fractions and elevation. In addition, we test the sensitivity of ZEMBA to a doubling of the atmospheric CO_2 concentration and a 2% increase in solar radiation at the top of the atmosphere. Compared to a global climate model (NorESM2) and reanalysis data (ERA5), ZEMBA reproduces the zonally-averaged climate of the pre-industrial period

10   with reasonable accuracy, capturing features such as surface temperature, precipitation, radiative fluxes, snow cover, sea ice cover and meridional heat transport. The response of ZEMBA to increasing CO_2 concentrations is qualitatively similar to the observational record and climate models of higher complexity, including a polar amplification over the northern hemisphere and during the winter months. The globally-averaged rise in surface air temperature for a doubling in CO_2 is 3.6°C. Finally, ZEMBA shows success in emulating changes in surface temperature and precipitation during the Last Glacial Maximum when

15   compared to reconstructions and global climate models.

1 Introduction

Since the beginning of the Quaternary Period (2.58 Ma BP to present), the Earth's climate has repeatedly switched between cold 'glacial' periods and warmer 'interglacial' periods, which are collectively known as glacial-interglacial cycles. Glacial periods are characterized by the presence of large ice sheets covering North America and Fennoscandia, while interglacials

20   refer to times when ice sheets are restricted to Greenland and Antarctica, such as the current Holocene period (11.7 Kyr BP to present). The glacial-interglacial cycles are widely documented in natural climate archives. For example, the oxygen isotope ratio ($\delta^{18}O$) recorded in the shells of micro-organisms that accumulated on the ocean floor – hereafter referred to as benthic $\delta^{18}O$ – serves as a valuable proxy for global ice volume and deep-ocean temperatures spanning millions of years into the past

(Lisiecki and Raymo, 2005; Elderfield et al., 2012). Additionally, Antarctic ice cores contain valuable information over the
past 800 Kyr BP such as atmospheric CO_2 concentrations (Bereiter et al., 2015) and surface temperature changes (Jouzel et al., 2007; Kawamura et al., 2017).

Despite extensive research, a comprehensive understanding of what caused the glacial-interglacial cycles has remained elusive, although the importance of changes in the Earth's orbital parameters has been widely acknowledged. Changes in these orbital parameters are thought to determine when the Earth switches between glacial and interglacial climates, by redistributing the incoming solar radiation (insolation) the Earth receives across latitudes and seasons. These variations in the orbital parameters encompass changes in the shape of the Earth's orbit (eccentricity) on 100 kyr and 413 kyr cycles, changes in the tilt (obliquity) of the Earth's rotational axis on 41 kyr cycles, and changes in the time of the year when the Earth is closest to the Sun (climatic precession) on 19 and 23 kyr cycles. The most favoured of the orbital hypotheses comes from Milutin Milankovitch, who proposed that glaciation occurs during times of reduced insolation at the high northern latitudes during the summer months; when obliquity is low and the northern hemisphere (NH) summers coincide with the Earth's furthest distance from the Sun (and vice versa for deglaciation). Hays et al. (1976) were the first to discover that benthic $\delta^{18}O$ records contained cycles with periods of 23, 41 and 100 kyr, in correspondence to the Earth's orbital cycles. Subsequent research has extended the benthic $\delta^{18}O$ record further back in time (Pisias and Moore, 1981; Ruddiman et al., 1986) and compiled records from across the globe (Ahn et al., 2017; Lisiecki and Raymo, 2005), which together with other proxy records (Jouzel et al., 2007; Kawamura et al., 2017) provides compelling evidence for an orbital control on climate change. Nevertheless, despite a distinct orbital 'rhythm' to glacial-interglacial cycles, important characteristics of the benthic $\delta^{18}O$ record cannot be easily explained by Milankovitch theory, including a shift in the dominant periodicity observed around 1 Myr BP.

The Mid-Pleistocene Transition (MPT) from 1.25 to 0.7 Myr BP represents a transition in the dominant periodicity of glacial-interglacial cycles from 41 kyr in the Early Pleistocene to ~100 kyr in the Late Pleistocene, occurring without significant variations in the orbital cycles (Clark et al., 2006). Hypotheses for the MPT are abundant in the literature (Berends et al., 2021). Prominent theories include the removal of a subglacial regolith beneath the NH ice sheets (Clark and Pollard, 1998; Clark et al., 2006), a gradual decline in atmospheric CO_2 concentrations (Raymo et al., 1988; Berger et al., 1999), the merging of the Laurentide and Cordilleran ice sheets over North America (Bintanja and Van De Wal, 2008), a phase-locking between Antarctica and the NH ice sheets (Raymo et al., 2006), or some combination of these mechanisms (Chalk et al., 2017; Willeit et al., 2019). Even the ~100 kyr cycles in glacial ice volume following the MPT were not an expected outcome of Milankovitch theory, as they correspond to changes in the Earth's eccentricity which has a negligible direct influence on summer insolation (Imbrie et al., 1993). Instead, these ~100 kyr cycles have been explained as the skipping of one (80 kyr) or two (120 kyr) obliquity cycles (averaging to 100 kyr) leading to longer glacial cycles in the Late Pleistocene (Huybers and Wunsch, 2005), where the precise timing of deglaciation is set by the combined forcing of precession and obliquity (Huybers, 2011; Parrenin and Paillard, 2012). Independent modelling studies have also highlighted the importance of precession, and its modulation by eccentricity, for generating the ~100 kyr cycles of the Late Pleistocene (Abe-Ouchi et al., 2013; Ganopolski and Calov, 2011). Interestingly, models that have simulated the MPT and/or the ~100 kyr cycles of the Late Pleistocene, still struggle to

reproduce the dominant 41 kyr cycles seen in the Early Pleistocene (Berger et al., 1999; Willeit et al., 2019; Watanabe et al., 2023).

60   While modelling experiments of the Early Pleistocene capture 41 kyr cycles in ice volume, in correspondence to the Earth's obliquity, they generate stronger 19 and 23 kyr precession cycles (Berger et al., 1999; Willeit et al., 2019; Watanabe et al., 2023) than observed in the benthic $\delta^{18}O$ record (Lisiecki and Raymo, 2005). This is unsurprising given that climatic precession controls the intensity of summer insolation. Indeed, for various metrics of summer insolation variability, including mid-month insolation (i.e., June 21st), monthly mean insolation or the caloric summer half-year, precession has a strong influence at

65   latitudes where NH ice sheets grow and melt. It's worth noting that precession cycles are detectable prior to the MPT (Lisiecki and Raymo, 2007; Liautaud et al., 2020), but they become more pronounced across the Quaternary, with much stronger signals observed in the Late Pleistocene compared to the Early Pleistocene (Raymo and Nisancioglu, 2003). Theories that account for weaker 19 and 23 kyr precession cycles during the Early Pleistocene include a counterbalancing between summer insolation intensity and summer duration (Huybers, 2006; Huybers and Tziperman, 2008), the cancellation of out-of-phase precession

70   cycles between the NH ice sheets and Antarctica (Raymo et al., 2006; Morée et al., 2021) or the influence obliquity has on the poleward flux of moisture and accumulation rates on ice sheets (Raymo and Nisancioglu, 2003; Nisancioglu, 2004).

Models of varying levels of complexity have been employed to address these questions relating to glacial-interglacial cycles. Global climate models (GCMs) are too computationally expensive for simulations on these time scales, so studies have instead relied on conceptual models (Paillard, 1998; Parrenin and Paillard, 2003; Legrain et al., 2023), energy balance models (Pollard,

75   1978; Huybers and Tziperman, 2008), earth system models of intermediate complexity (Ganopolski and Calov, 2011; Willeit et al., 2019), climate parameterization based on discrete GCM 'snapshots' (Abe-Ouchi et al., 2013) or benthic $\delta^{18}O$ records (Bintanja and Van De Wal, 2008). Among these, zonally-averaged energy balance models (EBMs) provide computationally efficient tools for studying the response of the Earth's climate to changes in the Earth's orbital parameters (Suarez and Held, 1979; Pollard, 1978; Huybers and Tziperman, 2008; Stap et al., 2014). EBMs calculate the distribution of surface temperature

80   with latitude by considering the conservation of energy on a sphere subject to heating by solar insolation, cooling by terrestrial radiation and the diffusive redistribution of heat from the equator to the poles. Since the original works of Budyko (1969) and Sellers (1969), EBMs have long been used to study Earth's climate sensitivity and have shown success at simulating both present-day and glacial climate states (North, 1975; Peng et al., 1987; Harvey, 1988; Jentsch, 1991; Bintanja and Oerlemans, 1996; Bintanja, 1997; Stap et al., 2014). Moreover, the simplicity of EBMs enables the isolation and identification of important

85   processes and feedbacks (Huybers and Tziperman, 2008; Stap et al., 2014). Consequently, EBMs provide valuable tools for exploring the glacial-interglacial cycles of the Quaternary and guiding further investigations with more realistic models.

In this study, we introduce a **Z**onally-Averaged **E**nergy and **M**oisture **BA**lance Climate Model **(ZEMBA)** to study the response of the Earth's climate to changes in the orbital parameters. The model is designed to place a particular emphasis on physical processes that may influence the relative contributions of obliquity and precession to climate variability during periods

90   such as the Early Pleistocene. The model spans both hemispheres and is forced by the full seasonal cycle in insolation. Unlike previous EBMs used for studies of glacial-interglacial cycles (Pollard, 1978; Huybers and Tziperman, 2008; Stap et al., 2014), ZEMBA includes a hydrological cycle to simulate precipitation and snowfall. Before using ZEMBA in experiments relating

to glacial-interglacial cycles, it is important to ensure the model can simulate the present-day (or pre-industrial) climate with reasonable accuracy, and to constrain the sensitivity of the model. Consequently, the following sections document the model and it's climate sensitivity. Firstly, we describe a 'control' simulation of ZEMBA for the pre-industrial period and compare the broad-scale features with a GCM and atmospheric reanalysis (Section 3.1). Then, we test the sensitivity of the model for a doubling of the atmospheric CO_2 concentration and a 2% increase in solar insolation (Section 3.2). Finally, we evaluate its performance for a simulation of the Last Glacial Maximum (LGM) (Section 3.3). Prior to that, we provide a detailed description of the atmospheric (Section 2.1), land (Section 2.2) and ocean (Section 2.3) components of ZEMBA.

2 Model Description

ZEMBA is primarily based on the EBM from Bintanja (1997) which comprises a single atmospheric layer overlying a surface divided into land and ocean. While utilizing the same shortwave and longwave radiation scheme as in Bintanja (1997), in addition to the same ocean transport model, ZEMBA includes a hydrological cycle to estimate precipitation and snowfall rates. Moreover, instead of parameterizing the surface albedo over land as a function of surface temperature, ZEMBA directly estimates snow coverage through the competition between snow accumulation (from the hydrological cycle) and ablation (from the surface energy balance). Following recent studies (Hwang and Frierson, 2010; Rose et al., 2014; Roe et al., 2015; Feldl and Merlis, 2021), atmospheric heat transport is now proportional to meridional gradients in near-surface moist static energy - instead of classic (or 'dry') EBMs that diffuse heat along temperature gradients. Moreover, a Hadley cell parameterization is included to produce an equatorward flux of moisture in the tropics (Siler et al., 2018). The radiative and turbulent heat fluxes are calculated separately over land and ocean, but atmospheric temperatures and humidities are set to the zonal average of land and ocean at the end of each model time-step. The subsequent sections will describe in more detail the individual atmospheric, land and ocean components of ZEMBA.

2.1 Atmosphere

A vertically averaged atmospheric layer simulates the evolution of near-surface air temperature (T_a) through time. It evolves as a function of the radiative fluxes exchanged at the top and bottom of the atmospheric layer, the turbulent exchange of heat with the surface layer, the latent heat released during precipitation and snowfall, and the divergence of atmospheric heat transport. The temporal evolution of T_a is described as follows:

$$C_a H_a \rho_a \frac{\delta T_{a(i)}}{\delta t} = S_{a(i)} + I_{a(i)} + K_{(i)} + L_v P_{r(i)} + L_s P_{s(i)}$$
$$+ \frac{1}{2\pi R_e^2 cos(\theta)} \frac{\delta F_T}{\delta \theta} \tag{1}$$

where C_a, H_a and ρ_a are the specific heat, height and density of the atmospheric layer, respectively, i is the index representing either land or ocean grid cells, S_a is the absorbed shortwave radiation, I_a is the absorbed longwave radiation, K is the exchange of sensible heat with the surface, $L_v P_r$ is the latent heat released during precipitation, where L_v is the latent heat

4

of vaporisation and P_r is precipitation, and $L_f P_s$ is the latent heat released during snowfall, where L_f is the latent heat of snow formation and P_s is snowfall. Finally, the last term on the right side of the equation represents the horizontal diffusion of temperature where R_e is the Earth's radius, θ is the latitude and F_T is the northward flux of dry static energy in the atmospheric layer. All terms in Equation 1 are in $W\,m^{-2}$.

2.1.1 Radiative Fluxes

The model is forced by diurnally averaged solar insolation ($S^\downarrow_{TOA}$) using the orbital parameter solution from Laskar et al. (2004). The amount of solar radiation that is reflected at the top of the atmosphere ($S^\uparrow_{TOA}$), transmitted to the surface ($S^\downarrow_{BOA}$) and reflected at the surface ($S^\uparrow_{BOA}$) is calculated using the parameterization from Bintanja (1996). The shortwave parameterization takes into account several atmospheric properties including surface air temperature, surface albedo, solar zenith angles, cloud optical depth and surface height. The shortwave fluxes are computed for both clear-sky and overcast conditions, with the total radiative flux for a given grid cell determined as the weighted average using prescribed cloud cover fractions. In this study, cloud cover fractions over land and ocean are taken from pre-industrial simulations of the Norwegian Earth System Model version 2 (Seland et al., 2020) (Fig 1a). Daytime-mean solar zenith angles are calculated with equations provided by Balmes and Fu (2020). Of particular importance to the shortwave radiative fluxes is the cloud optical depth parameter (τ). Following Bintanja (1996, 1997) and Stap et al. (2014), τ is kept fixed to a globally and seasonally invariant value. The amount of shortwave radiation that is absorbed by the atmosphere (S_a) and the surface ($S_{(i)}$) - either land and ocean - is described below:

$$S_{(i)} = S^\downarrow_{BOA(i)} - S^\uparrow_{BOA(i)} \tag{2}$$

$$S_{a(i)} = (S^\downarrow_{TOA(i)} - S^\uparrow_{TOA(i)}) - S_{(i)} \tag{3}$$

The outgoing longwave radiative fluxes at the TOA ($I^\uparrow_{TOA}$) and the surface ($I^\uparrow_{BOA}$), together with the incoming longwave flux at the surface ($I^\downarrow_{BOA}$), are also calculated using a radiation parameterization from Bintanja (1996). The longwave parameterization is made a function of surface temperature, surface air temperature, cloud emissivity, surface elevation and atmospheric CO_2 concentrations. As with the shortwave parameterization, the longwave radiative fluxes are calculated separately for clear-sky and overcast conditions. Alterations to the longwave parameterization from Stap et al. (2014), to both increase the climate sensitivity per CO_2 doubling and account for the effects of non-CO_2 greenhouse gases are maintained. The absorbed longwave radiation fluxes at the surface ($I_{(i)}$) and by the atmosphere (I_a) is shown below:

$$I_{(i)} = I^\downarrow_{BOA(i)} - I^\uparrow_{BOA(i)} \tag{4}$$

$$I_{a(i)} = I^\uparrow_{BOA(i)} - (I^\downarrow_{BOA(i)} + I^\uparrow_{TOA(i)}) \tag{5}$$
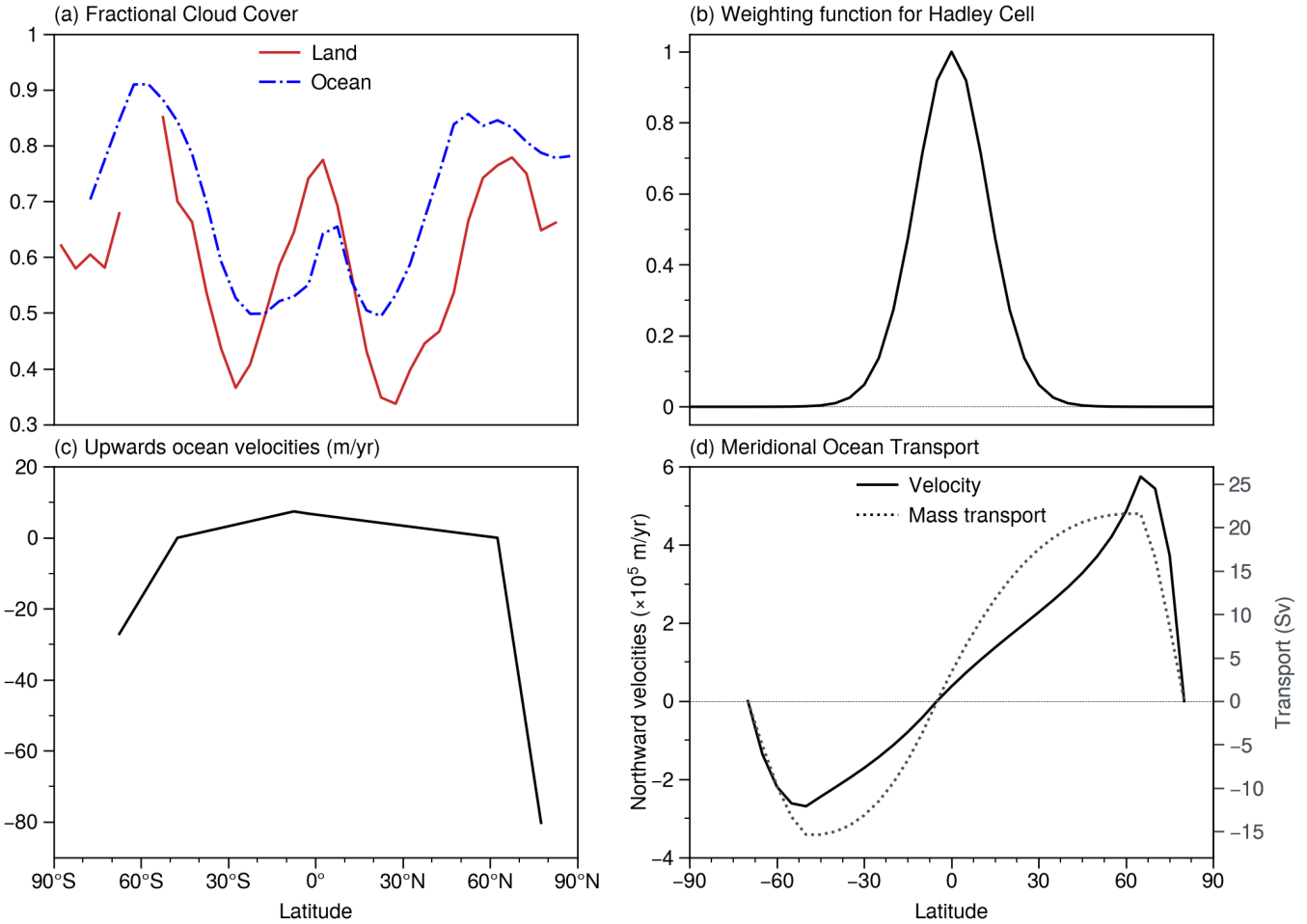
Figure 1. (a) Cloud cover fractions over land and ocean taken from pre-industrial simulations of NorESM2 (Seland et al., 2020) for calculations of the shortwave and longwave radiative fluxes. (b) The weighting function from Siler et al. (2018) which determines the fraction of atmospheric heat transport carried out by the Hadley cell. (c) The prescribed vertical ocean velocities for driving ocean circulation, with upwelling from $50°$S to $60°$N (with an average upwelling rate of 4 m/yr) and downwelling from $70°$S to $50°$S and from $60°$N to $80°$N. (d) The resulting northward horizontal velocities in the uppermost ocean layer (solid black line), together with the mass transport (grey dotted line) in Sverdrups ($1 \times 10^6 m^3/s.$)

2.1.2 Turbulent Heat Fluxes

The aerodynamic bulk relationships are employed to compute the fluxes of sensible heat (K) and evaporation (E) across the atmosphere/ocean and atmosphere/land interface:

$$K_{(i)} = \rho_a \, c_a \, \kappa_{(i)} \left(T_{(i)} - T_{a(i)} \right) \qquad (6)$$

$$155 \quad E_{(i)} = \rho_a \, W_{(i)} \, \kappa_{(i)} \left(Q_{sat}(T_{(i)}) - Q_{a(i)} \right) \tag{7}$$

where κ is the turbulent exchange coefficient, T_i is the temperature of the surface- either land or ocean, W denotes the surface water availability, $Q_{sat}(T_{(i)})$ is the saturation specific humidity of the surface (as determined by the Clausius-Clapeyron relation) and Q_a is the specific humidity of the overlying atmospheric layer. The influence of wind speed and surface roughness on the turbulent heat exchange is not incorporated into κ which remains constant across latitudes and seasons. The latent heat

160   flux associated with evaporation is simply $L_v E$.

2.1.3 Hydrological Cycle

An atmospheric moisture budget is introduced to parameterize the hydrological cycle and simulate precipitation and snowfall (Fanning and Weaver, 1996; Robinson et al., 2010; Ritz et al., 2011):

$$\rho_a H_a \frac{\delta Q_{a(i)}}{\delta t} = (E_{(i)} - P_{r(i)}) + \frac{1}{2\pi R_e^2 cos(\theta)} \frac{\delta F_Q}{L_v \delta \theta} \tag{8}$$

165   where the first term on the right side of the equation represents the sources (evaporation) and sinks (precipitation) and the second term represents the horizontal transport of water vapor, in which F_Q is the latent heat associated with the northward moisture flux. Precipitation occurs once the relative humidity, r, exceeds a maximum threshold, r_{max}:

$$P_{r(i)} = \begin{cases} \frac{\rho_a \, H_a}{3 \cdot \Delta t} \left(Q_{a(i)} - r_{max} \cdot Q_{sat}(T_{a(i)}) \right), & r > r_{max} \\ 0, & \text{otherwise} \end{cases} \tag{9}$$

where $Q_{sat}(T_a)$ is the saturation specific humidity of the atmospheric layer and Δt is the seconds in one day. When the

170   relative humidity exceeds r_{max}, it leads to precipitation. The turnover time for excess humidity in the atmosphere is set to 3 days. Shorter turnover times produces very large and sporadic contribution of latent heat to the atmospheric column. The amount of snowfall is determined using:

$$P_{s(i)} = P_{r(i)} \cdot f_{sf} \tag{10}$$

where f_{sf} is the fraction of precipitation that falls as snow, which is parameterized as a function of surface air temperature

175   (Harvey, 1988):

$$f_{sf(i)} = \begin{cases} 1. & T'_{a(i)} < 260K \\ 0.005(280 - T'_{a(i)}) & 260K <= T'_{a(i)} <= 280K \\ 0. & T'_{a(i)} > 280K \end{cases} \tag{11}$$

where $T_a^{'}$ is the surface air temperature corrected for the zonal mean elevation with a global mean lapse rate of $-6.5\,K\,km^{-1}$. Besides representing the overall quantity of precipitation converted into snow, f_s also determines the fraction of the surface where precipitation takes the form of snowfall. In other words, rather than assuming a uniform distribution of snow across the

180 grid box, only a portion of the land or ocean surface is affected by snowfall.

2.1.4 Atmospheric Heat Transport

The division of the model into land and ocean raises the question of how to parameterize the "zonal mixing" of air belonging to the same latitudes, but overlying each surface type. We adopt an 'infinite wind' mixing scenario (Peng et al., 1987; Bintanja, 1997), in which atmospheric temperatures and humidities over land and ocean are both set equal to the zonal mean ($\overline{T_a}$ and $\overline{Q_a}$,

185 respectively) at the end of each model time-step, according to the land fraction at each grid cell. The assumption of 'infinite wind' mixing between atmosphere over land and ocean is based on previous studies that find a more realistic representation of zonal mixing produces small differences in model sensitivity (Harvey and Schneider, 1985; Thompson and Schneider, 1979). In other words, one atmospheric layer effectively overlies the entire surface. Following the zonal mixing of temperature and humidity, meridional atmospheric heat transport is parameterized as a diffusive process along horizontal gradients in moist

190 static energy ($\overline{m}$), expressed as $\overline{m} = c_a\overline{T_a} + L_v\overline{Q_a}$, where $c_a\overline{T_a}$ and $L_v\overline{Q_a}$ represent the dry static and moist components of atmospheric heat content, respectively. Moist static EBMs behave differently to classic ('dry') EBMs (Hwang and Frierson, 2010; Feldl and Merlis, 2021) and have shown success at emulating the response of more comprehensive GCMs to climate forcings (Hwang and Frierson, 2010; Roe et al., 2015). Consequently, the total northward transport of energy (F_{total}) within the atmospheric layer is described as follows:

$$195 \quad F_{total} = 2\pi R_e cos(\theta)\rho_a H_a D_a \frac{\delta\overline{m}}{R_e\delta\theta} \tag{12}$$

where D_a is the atmospheric diffusion coefficient. The total atmospheric heat transport is divided into the dry static (F_T) and latent component (F_Q). F_T contributes directly to the heating of the atmospheric layer, whereas F_Q transports water vapor within the simplified hydrological cycle (Eq. 8), which is translated into heating of the atmospheric column once precipitation occurs. To capture the equatorward transport of latent heat at the tropics, against meridional gradients in $\overline{m}$, the Hadley cell

200 parameterization introduced by Siler et al. (2018) is included to obtain a more realistic representation of the hydrological cycle. In this parameterization, F_{total} is partitioned into a Hadley cell (F_{HC}) and an eddy component (F_{eddy}) (Siler et al., 2018):

$$F_{total} = F_{HC} + F_{eddy}$$
$$= F_{total}\,\Omega + F_{total}[1 - \Omega] \tag{13}$$

where Ω is a weighting function (shown in Fig. 1b) that ensures the Hadley cell dominates heat transport in the tropics, whereas eddies control the poleward flux of $\overline{m}$ in the mid-latitudes and polar regions. Where $\Omega = 0$, eddies account for all heat

205 transport via the down-gradient diffusion of both dry static (F_{T_eddy}) and latent (F_{Q_eddy}) heat:

$$F_{T\_eddy} = 2\pi R_e cos(\theta) \rho_a H_a c_a D_a \frac{\delta \overline{T_a}}{R_e \delta \theta} [1 - \Omega] \tag{14}$$

$$F_{Q\_eddy} = 2\pi R_e cos(\theta) \rho_a H_a L_v D_a \frac{\delta \overline{Q_a}}{R_e \delta \theta} [1 - \Omega] \tag{15}$$

In the Hadley cell, the poleward flux of $\overline{m}$ in its upper branch slightly exceeds the equatorward flux in its lower branch (Hartmann, 2015). Noting that meridional gradient in $\overline{m}$ are relatively flat within this upper branch, Siler et al. (2018) approximate this difference in $\overline{m}$ between the upper and lower branch as $g = \lambda \cdot \overline{m}_{eq} - \overline{m}$, where $\overline{m}_{eq}$ is the near-surface moist static energy at the equator and λ is the fractional increase in moist static energy in the upper branch of the Hadley cell relative to $\overline{m}_{eq}$. In Siler et al. (2018) λ is set to 1.06 but for this study $\lambda = 1.03$ to improve our simulation of precipitation for the pre-industrial period. Consequently, the net poleward transport of heat within the Hadley cell is given as $F_{HC} = g\,\psi$, where ψ is the mass transport within either the upper or lower branch of the Hadley cell, which can be solved using Eq. 13 as $\psi = (F_{total}\,\Omega)/g$. Assuming latent heat transport is confined to the lower branch, the dry static (F_{T_HC}) and latent (F_{Q_HC}) contributions to heat transport within the Hadley cell are parameterized using (Siler et al., 2018):

$$F_{T\_HC} = \psi[g + L_v\,Q_a] \tag{16}$$

$$F_{Q\_HC} = -\psi\,L_v\,Q_a \tag{17}$$

Finally, the Hadley cell and eddy contributions to the total northward transport of dry static (F_T) and latent (F_Q) heat is simply given by:

$$F_T = F_{T\_HC} + F_{T\_eddy} \tag{18}$$

$$F_Q = F_{Q\_HC} + F_{Q\_eddy} \tag{19}$$

where the divergence of F_T and F_Q (converted into a moisture flux by dividing with L_v) is shown on the right side of Eq. 1 and Eq. 8, respectively.

2.2 Land

The energy balance of the land surface is expressed by:

$$C_l\,\rho_l\,H_l\,\frac{\delta T_l}{\delta t} = S_l + I_l - K_l - L_v E_l \tag{20}$$

where C_l, H_l and ρ_l are the specific heat, depth and density of the ground layer, respectively, T_l is the land surface temperature, S_l is the absorbed shortwave radiation, I_l is the absorbed longwave radiation, K_l is the upwards sensible heat flux and E_l is the upwards evaporation flux over land.

2.2.1 Snow mass budget

The model evaluates both the proportion of the land surface covered by the snow (f_{sc}) and the average thickness of the snowpack (d). A fractional area, f_{sf}, of the precipitation accumulates as snow at the surface according to the hydrological cycle (Section 2.1.3). Similar to the EBM from Harvey (1988), when f_{sf} is less than the existing area of the snowpack (i.e., 235 $f_{sf} < f_{sc}$), this snowfall gain is redistributed over the larger snow-covered area. Conversely, if $f_{sf} > f_{sc}$, the snow-covered area is expanded to match the area of snowfall ($f_{sf} = f_{sc}$) with the average thickness of the snowpack, d, adjusted to conserve the total mass of snow in each grid cell.

If land surface temperatures (T_l) exceeds the melting point of snow (T_K), T_l is reset to T_k and the excess energy is used to melt the snowpack. This rate of melting ($\dot{M}$) is described by:

$$
240 \quad \dot{M} = \begin{cases} 0, & T_l <= T_K \\ (T_l - T_K) \cdot \frac{C_l}{L_f \rho_{ice}}, & T_l > T_K \end{cases} \tag{21}
$$

where ρ_{ice} represents the density of ice. Surface melting of the snowpack is evenly distributed between reducing the average thickness, d, and the fractional area of the snowpack, f_{sc}. Should $\dot{M}$ exceed d, excess melt is reconverted into heating of the land surface.

2.2.2 Land surface albedo

245 The albedo of the land is determined by the fraction of the surface that is covered in snow, together with the albedo of the snow-covered and snow-free surfaces. The fractional snow cover, f_{sc}, is assumed to be evenly distributed between two surface types: bare ground and land ice. The snow albedo (α_s) is expressed as a linear function of surface temperature following Bintanja (1997).

$$
\alpha_s = \begin{cases} \alpha_{cs}, & T_l^{'} <= 263K \\ \alpha_{cs} + [\alpha_{ws} - \alpha_{cs}] \frac{(T_l^{'} - 263)}{10}, & 263K <= T_l^{'} <= 273K \\ \alpha_{ws}, & T_l^{'} > 273K \end{cases} \tag{22}
$$

250 where α_{cs} is the maximum (or 'cold') snow albedo, α_{ws} is the minimum (or 'warm') snow albedo and $T_l^{'}$ is the land surface temperature corrected for zonal mean elevation. The average albedo over bare ground (α_g) and ice (α_i) is then calculated as the weighted average of the snow-covered and snow-free region.

$$
\alpha_g = f_{sc}\alpha_s + (1 - f_{sc})\alpha_{bg} \tag{23}
$$

$$
\alpha_i = f_{sc}\alpha_s + (1 - f_{sc})\alpha_{bi} \tag{24}
$$

255   where α_{bg} and α_{bi} is the albedo of bare ground and ice without snow cover. The average albedo over land (α_l) is then the weighted average of α_g and α_i, depending on the fractional area over which land is covered by ice (f_i):

$$\alpha_l = f_i\alpha_i + (1 - f_i)\alpha_g \tag{25}$$

Additionally, the impact of the solar zenith angle, θ_z, on albedo is accounted for by increasing the land albedo for zenith angles greater than 60° using the following parameterization (Lefebre et al., 2003):

260
$$\alpha_l = \begin{cases} \alpha_l + max\left\{0; 0.32 \cdot \frac{1}{2}\left[\frac{3}{1+4\cos\theta_z} - 1\right]\right\}, & \theta_z >= 60° \\ \alpha_l, & \theta_z < 60° \end{cases} \tag{26}$$

with the constraint that $\theta_z > 80° = 80°$. This adjustment increases land albedo for very high zenith angles in the polar regions.

2.3   Ocean

Meridional heat transport by the oceans is represented by the zonally averaged ocean circulation model from Bintanja (1997).
265   Extending from 70°S to 80°N, the ocean model comprises six layers of increasing thickness with depth (with a total depth of 4000 m). To induce thermohaline circulation, prescribed vertical velocities (shown in Fig. 1c) are used to produce regions of upwelling from 50°S to 60°N and downwelling in the polar regions. These vertical velocities remain constant with depth and drive the poleward flow of water in the uppermost model layer (Fig. 1c), and equatorward flow in the bottom layer, thereby generating a conveyor-belt system of ocean heat transport in each hemisphere. The central point of ocean circulation is placed
270   slightly south of the equator at 5°S to improve temperature estimates in the north polar regions. The model effectively consists of two overturning cells, with an average upwelling rate of 4 m/yr in the upwelling regions (from 50°S to 5°S in the southern cell and from 5°S to 60°N in the northern cell). The transport of ocean heat via eddies and gyres is represented as a diffusive process, along horizontal gradients in surface ocean temperature. Outside of the ocean circulation basin, where ocean fractions are greater than zero (e.g. from 80°S to 90°N), a passive mixed layer of 100 m depth exchanges radiative and turbulent heat
275   with the atmosphere. The temporal evolution of ocean temperature, T_o, is described by (Bintanja, 1997):

$$C_o\,\rho_o\,H_o\frac{\delta T_o}{\delta t} + \frac{1}{2\pi R_e^2 f\cos\theta}\frac{\delta F_{ov}}{\delta\theta} + C_o\,\rho_o\,H_o\frac{\delta}{\delta z}\left(w\,T_o\right) =$$
$$[S_o + I_o - K_o - L_vE_o] + \frac{1}{2\pi R_e^2 f\cos\theta}\frac{F_{eg}}{\delta\theta} +$$
$$\frac{1}{2\pi R_e^2 f\cos\theta}\frac{\delta F_i}{\delta\theta} + C_o\,\rho_o\,H_o\frac{\delta}{\delta z}\left(D_z\frac{\delta T_o}{\delta z}\right) \tag{27}$$

where C_o, ρ_o and D_o are the density, specific heat and depth of each ocean layer, f is the fractional width of the ocean basin,
280   F_{ov} is the northward advective heat flux in the top and bottom ocean layer due to overturning, w is the prescribed vertical

11

velocity, S_o is absorbed shortwave radiation, I_o is absorbed longwave radiative flux, K_o is the exchange of sensible heat with the atmosphere, E_o is evaporation at the ocean surface, F_{eg} is the northward diffusive heat flux at the ocean surface due to eddies and gyres, F_i is the northward diffusive heat flux in the ocean interior and D_z is the coefficient for vertical heat diffusion. The second term on the left side of the equation represents the divergence of horizontal heat advection which pertains to the top

285   and bottom layer of the ocean model, and the third term represents the divergence of vertical heat advection. On the right side, the first to fourth term represents the radiative and turbulent heat fluxes and the fifth term is the divergence of horizontal heat diffusion (via eddy/gyres), all of which apply to the uppermost ocean layer. The sixth term on the right side - the divergence of horizontal heat diffusion in the ocean interior, applies to every layer expect the surface. Finally, the last term on the right side of the equation represents the divergence of vertical heat diffusion. The northward heat flux associated with F_{ov}, F_{eg} and F_i is

290   described by:

$$F_{ov} = 2\pi R_e f cos(\theta) \rho_o C_o H_o u T_o \tag{28}$$

$$F_{eg} = -2\pi R_e f cos(\theta) \rho_o C_o H_o D_o \frac{\delta T_o}{R_e \delta \theta} \tag{29}$$

$$F_i = -2\pi R_e f cos(\theta) \rho_o C_o H_o D_i \frac{\delta T_o}{R_e \delta \theta} \tag{30}$$

where u is the horizontal ocean velocity, D_o is the diffusion coefficient related to eddy/gyre transport at the surface and D_i

295   is the coefficient related to horizontal diffusion in the ocean interior. The distribution of u is calculated using the continuity equation from (Bintanja, 1997, p. 9) and is shown in Fig. 1d.

2.3.1 Sea Ice

The inclusion of sea ice is important for capturing the seasonal range of surface temperatures at the higher latitudes. By reflecting the majority of incoming solar radiation and reducing heat exchange between the atmosphere and ocean surface,

300   the presence of sea ice reduces the effective thermal inertia of the ocean-atmosphere system. Therefore, ZEMBA includes a simple sea ice model to simulate the latitudinal distribution of sea ice and its modification to ocean albedo. The sea ice model does not account for variations in sea ice thickness or sea ice drifting. Sea ice of a prescribed thickness forms or melts when surface ocean temperatures drop below or exceed a critical threshold (T_{fo})- the freezing temperature of sea water. The heat flux available for the formation or melting of sea ice, Q_{si}, is governed by (Gildor and Tziperman, 2001):

305    $$Q_{si} = \frac{C_o \rho_o H_o A_o}{\Delta t} (T_{fo} - T_o) \tag{31}$$

where A_o is the surface area of the ocean. The heat flux is converted into changes in sea ice volume, V_{si}, as follows:

$$\frac{\delta V_{si}}{\delta t} = \frac{Q_{si}}{\rho_{ice} L_f} + \frac{P_s \cdot A_{si}}{\rho_{ice}} \tag{32}$$

where the the last term on the right side of the equation represents the snowfall contribution to sea ice volume, with A_{si} being the the surface area of the sea ice cover. Sea ice volume is then converted into sea ice areal extent by assuming a constant sea ice thickness. When sea ice forms, the temperature of the underlying ocean is reset to T_{fo}, and when sea ice covers the entire ocean surface (i.e. $A_{si} = A_o$), the temperature of the sea ice can drop below the freezing point, while the upper ocean layer remains at T_{fo}. Variations in the temperature of the surface ocean layer (and thereby sea ice) are determined by the surface energy balance, in addition to advective and diffusive ocean heat fluxes within the ocean circulation model.

2.3.2 Ocean Albedo

The albedo of the open ocean, α_{op}, is parameterized as a function of the solar zenith angle, as derived from aircraft observations (Taylor et al., 1996):

$$\alpha_{op} = \frac{0.037}{1.1\theta_z^{1.4} + 0.15} \tag{33}$$

with the average albedo for the ocean surface, α_o, described by:

$$\alpha_o = \alpha_{si} f_{si} + \alpha_{op}(1 - f_{si}) \tag{34}$$

where α_{si} is the albedo of the sea ice and f_{si} is the fraction of the ocean surface covered by sea ice.

2.4 Numerical details

The model has the option to be simulated at a 1°, 2.5° and 5° resolution. All equations are solved using the forward Euler method with the majority of calculations performed using a 1 day time-step. The exception is the atmospheric and ocean heat transport processes, which may be solved using a shorter time-step for numerical stability reasons depending on the choice of model resolution. For all experiments described in the subsequent sections, a model resolution of 5° is chosen. The prescribed thickness at which sea ice grows and melts in both hemispheres is chosen to be 2 m and cloud emissivity is set to 1. The percentage of land and ocean cover for each zonal band is taken from the ICE-6G_C dataset (Argus et al., 2014; Peltier et al., 2014), although a mixed ocean layer is assumed to cover the entire surface from 80°-90°) and land occupies everywhere poleward of 75°. All other important model parameters are summarized in Table 1.

3 Results

We present equilibrium simulations of the model for the pre-industrial period (ZEMBA-PI), a doubling of atmospheric CO_2 conditions relative to the pre-industrial (ZEMBA-2xCO_2), a 2% increase in solar insolation (ZEMBA-S$_o$+2%) and the Last Glacial Maximum (ZEMBA-LGM). The model takes approximately 3000 model years to reach an equilibrium due to the inclusion of the ocean model with a large heat capacity. Key model parameters remain the same between experiments (Table

Table 1. Selection of important parameters used in the atmospheric, land and ocean components of the model.

Parameter	Units	Value	Description
Atmosphere			
τ	-	3.0	Cloud Optical Depth
κ_l	m/s	0.01	Turbulent heat flux coefficient over land
κ_o	m/s	0.006	Turbulent heat flux coefficient over ocean
r_{max}	-	80	Maximum relative humidity
D_a	m/s	0.7×10^6 (SH); 0.84×10^6 (NH)	Diffusion coefficient for atmospheric heat transport
Land			
α_g	-	0.15	Albedo of bare ground
α_{cs}	-	0.8	Albedo of cold ('dry') snow
α_{ws}	-	0.4	Albedo of warm ('wet') snow
α_i	-	0.8	Albedo of land ice
Ocean			
α_{si}	-	0.7	Albedo of sea ice
D_o	m/yr	5×10^{10}	Diffusion coefficient for horizontal heat transport at surface
D_i	m/yr	1.5×10^{10}	Diffusion coefficient for horizontal heat transport in ocean interior
D_z	m/yr	5×10^3	Diffusion coefficient for vertical heat transport

335   1) with the exception of the insolation forcing, atmospheric CO_2 concentration and land ice extent (Table 2). The following sections describe output from ZEMBA prior to coupling with an interactive ice sheet model, with fixed land ice fractions and zonal mean elevations over land taken from ICE-6G_C (Argus et al., 2014; Peltier et al., 2014).

Table 2. Boundary conditions for ZEMBA experiments.

Experiment	Insolation	Elevation/Ice	CO_2
ZEMBA-PI	PD	PD	284 ppm
ZEMBA-LGM	21 Kyr BP	21 Kyr BP	184 ppm
ZEMBA-2xCO$_2$	PD	Present-day	568 ppm
ZEMBA-S$_o$+2%	PD + 2%	PD	284 ppm

3.1 Pre-Industrial Simulation

For the ZEMBA-PI experiment, the model is forced with insolation using the present-day orbital parameters (Laskar et al.,
340   2004) and an atmospheric CO_2 concentration of 284 ppm. To evaluate the accuracy of the ZEMBA-PI model output, we
compare against a selection of zonally-averaged climate variables from a pre-industrial simulation of the Norwegian Earth
System Model version 2 (NorESM2) (Seland et al., 2020). In addition, we compare ZEMBA-PI against the ERA5 atmospheric
reanalysis product averaged between 1940 and 1970 (Hersbach et al., 2023), which belong to a period when CO_2 concentrations
were $\sim$ 26-45 ppm higher than the pre-industrial, but provides one of the earliest observational constraints on the full-field
345   climate. As can be seen in the subsequent figures (Fig. 2-7), zonally averaged climate variables are very similar between
NorESM2 and ERA5 1940-1970.

The zonal mean surface air temperatures are shown in Figure 2. For annual-mean temperatures (Fig. 2a), the latitudinal
structure correspond nicely between ZEMBA-PI, NorESM2 and ERA 1940-1970. Moreover, the DJF-mean (Fig. 2b) and JJA-
mean (Fig. 2c) demonstrates the model successfully captures the seasonal amplitude of temperatures at the higher latitudes
350   in accordance with both NorESM2 and ERA5 1940-1970. The globally averaged surface air temperature from ZEMBA-PI
is 13.87°C, compared with 13.78°C from NorESM2 and 13.72°C from ERA5 1940-1970 (Table 3). When the ZEMBA-PI
experiment is compared only to NorESM2, the difference in annual mean temperatures never exceeds $\sim$4°C (Fig. 2d). The
same is true for ERA5 1940-1970 with the exception of 70-90°N, as ERA5 produces higher temperatures in this region
due to much higher winter temperatures (Fig. 2b), which is presumably in response to higher CO_2 levels. Between 60° and
355   30°S, ZEMBA-PI appears to consistently overestimate temperatures, both annually (Fig. 2d) and seasonally (Fig. 2e-f), by as
much as $\sim$5°C when compared to NorESM2 and ERA5 1940-1970. Between 30° and 60°N, on the other hand, the annual
mean temperatures agree well amongst the models/reanalysis, but ZEMBA-PI seems to underestimate the seasonal range in
temperatures at these latitudes. Overall, despite some biases, ZEMBA captures both the annual mean and the seasonal range of
surface air temperatures with good accuracy for the pre-industrial period.

360   Figure 3 shows climate variables related to the hydrological cycle including precipitation, snowfall and evaporation. Given
the simplicity of ZEMBA, precipitation rates from the ZEMBA-PI experiment are well captured when compared to NorESM2
and ERA5 1940-1970 (Fig. 3a). The inclusion of the Hadley cell parameterization from Siler et al. (2018) generates a con-
vergence of moisture and strong precipitation rates near the equator. The largest difference between ZEMBA-PI and the other
models/reanalysis is located around the equator (Fig. 3c), as the precipitation maxima is located at the equator for ZEMBA-PI,
365   whereas the maxima for NorESM2 and ERA5 1940-1970 is located north of the equator in accordance with the mean position
of the Intertropical Convergence Zone (ITCZ). In the mid-latitudes and polar regions, precipitation rates from ZEMBA-PI are in
close approximation (within 1 mm/day) to those simulated by the complex models. The snowfall rates (Fig. 3) from ZEMBA-
PI over the NH correspond nicely with NorESM2 and ERA5 1940-1970. For the SH, on the other hand, while ZEMBA-PI
captures the location of maximum snowfall at 60°S, it appears to underestimates the snowfall rate by half (Fig. 3c). As for
370   evaporation, the zonally and annually averaged fluxes from ZEMBA-PI is very similar to NorESM2 and ERA5 (Fig. 3b) with
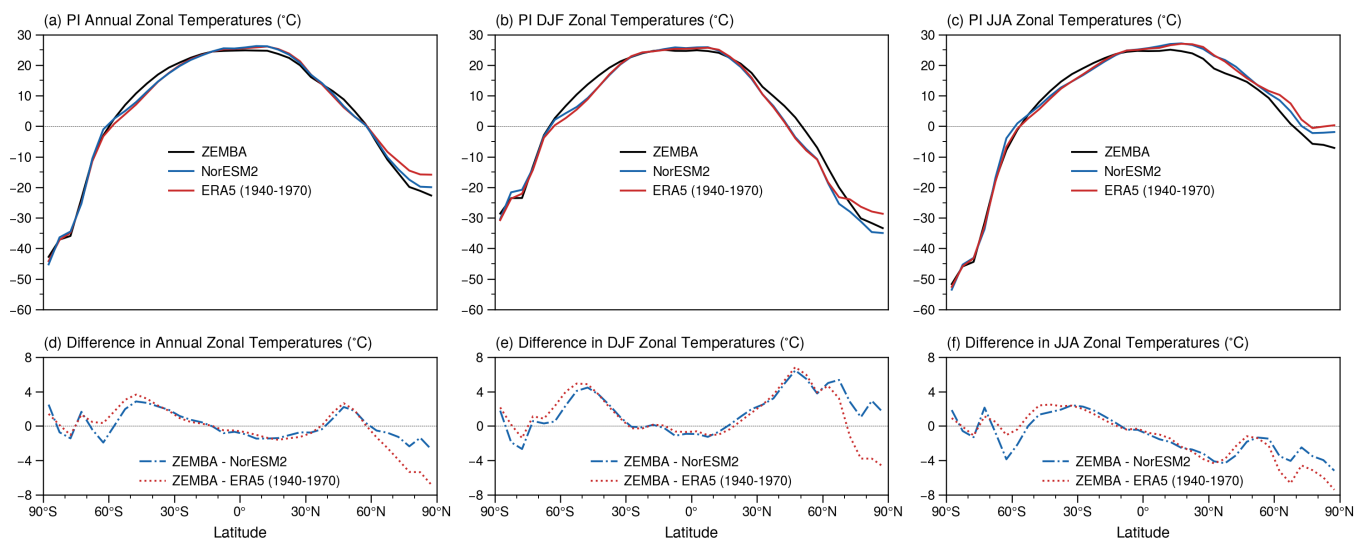the differences between them never exceeding 1 mm/day (Fig. 3d).

Figure 2. (top) The annual (a), December-January-February (b) and June-July-August (b) average of zonal mean surface air temperature for the pre-industrial period (PI), as simulated by ZEMBA (black lines) in comparison to NorESM2 (blue lines) and the ERA5 climatology from 1940 to 1970 (red lines). (bottom) The difference between ZEMBA and the other models/observations.

The surface and planetary albedo compares favourably between ZEMBA-PI, NorESM2 and ERA5 1940-1970 (Fig. 4). Most importantly, ZEMBA-PI reproduces the poleward enhancement of albedo at the higher latitudes due to the presence of snow and sea ice cover. The difference in surface and planetary albedo as simulated by ZEMBA-PI in comparison to the other models/reanalysis never exceeds 0.2 (Fig 4c-d). One limitation of the shortwave parameterization used in ZEMBA is the overestimation of the planetary albedo by ∼0.1 at the polar latitudes (Fig. 4c) as noted by Bintanja (1997). However, as the planetary albedo is slightly underestimated at the mid-latitudes, the globally averaged planetary albedo is very similar between ZEMBA (0.31), NorESM2 (0.32) and ERA5 1940-1970 (0.31) (Table 3). At the surface, the onset of higher surface albedo agrees well with NorESM2 and ERA5 1940-1970 in both hemispheres. Overall, the global mean surface albedo for ZEMBA-PI is 0.15, which is in agreement with both NorESM2 and ERA5 1940-1970 (Table 3).

Figure 5 shows the seasonal cycle in the areal extent of snow cover over land and sea ice coverage. Over land in the NH (Fig. 5a), the ZEMBA-PI experiment underestimates the winter maximum in snow coverage by $> 1 \times 10^{13}$ m^2 and thereby simulates a smaller seasonal amplitude in snow cover when compared to NorESM2 and ERA5 1940-1970 reanalysis. As for the SH (Fig. 5a), all models/reanalysis have similar areal extents in snow cover and produce negligible seasonal variations. The ZEMBA-PI experiment shows most success in simulating sea ice cover in the NH (Fig. 5b), capturing both the amplitude and the phase of sea ice cover changes in reference to NorESM2 and ERA5 1940-1970 (Fig. 5b). For sea ice cover in the SH (Fig. 5b), however, the ZEMBA-PI experiment underestimates the seasonal amplitude.

The annual and zonal mean radiative fluxes exchanged at the TOA and the surface is shown in Figure 6. For the TOA, the amount of absorbed shortwave radiation (ASR) estimated using the shortwave parameterization from ZEMBA (see Section
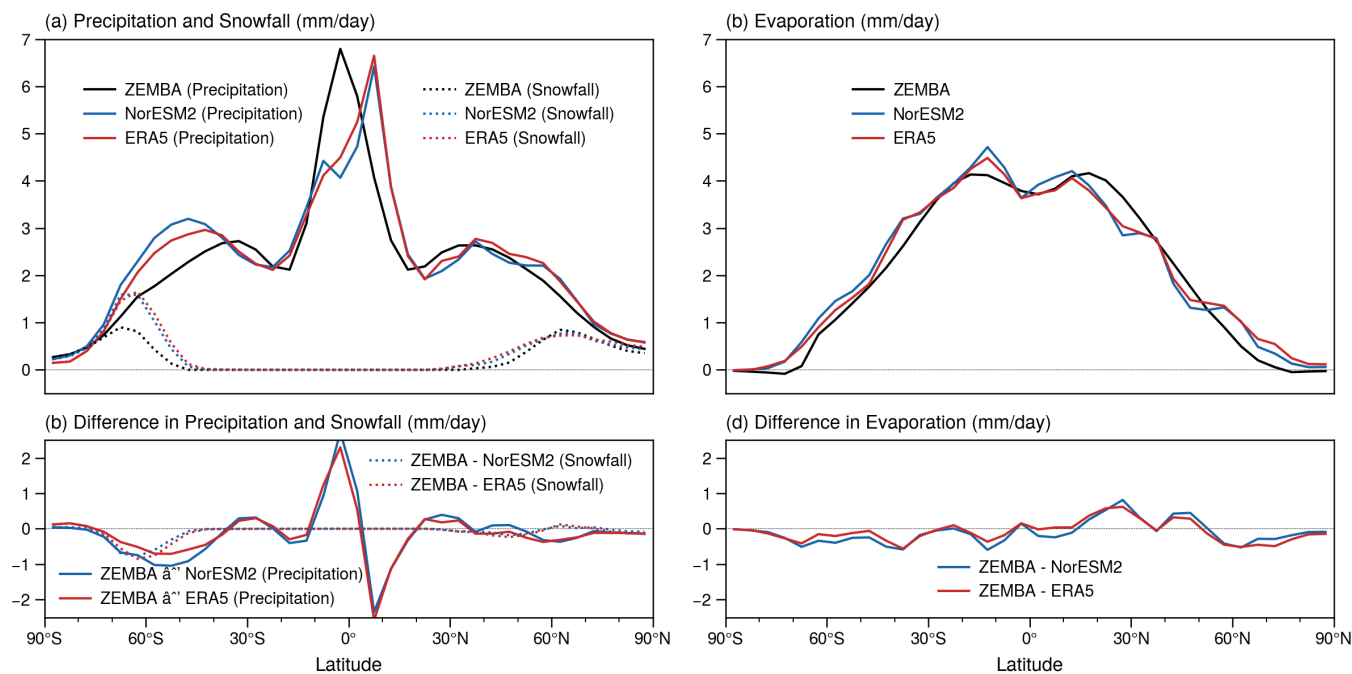
Figure 3. (top) The annual and zonal mean rates of precipitation/snowfall (a) and evaporation (b), as simulated by ZEMBA (black lines) in comparison to NorESM2 (blue lines) and the ERA5 1940-1970 climatology (red lines). Precipitation is shown in solid lines and snowfall is shown in dotted lines (a,c). (bottom) The difference between ZEMBA and the other models/reanalysis.

390   2.1.1) compares favourably with NorESM2 and ERA5 1940-1970, albeit with a slight underestimation of ASR in the polar latitudes by $\sim$ 10-20 W/m^2 (Fig. 6a). The largest differences between ZEMBA-PI and the other models/reanalysis resides in the outgoing longwave flux (OLR) at the TOA (Fig. 6a). In particular, there is a pronounced overestimation of OLR at the tropics (of up to $\sim$ 30 W/m^2) and a pronounced underestimation of OLR in the polar latitudes, especially in the northern hemisphere (of up to 50 W m^{-2}). Consequently, these differences in OLR are reflected in the net radiative flux (NET) received

395   at the TOA, with ZEMBA-PI receiving $\sim$ 25 W/m^2 less NET energy at the equator and up to $\sim$ 35 W/m^2 more NET energy in the high northern latitudes (Fig. 6c). As for the surface radiative fluxes (Fig. 6b), the ASR, OLR and NET radiative fluxes are quite similar but tend to be slightly underestimated in ZEMBA-PI when compared to the other datasets, with the difference in the NET fluxes never exceeding $\sim$ 25 W/m^2 (Fig. 6d). Overall, despite some clear discrepancies in the ASR and OLR, the ZEMBA-PI experiment produces net radiative fluxes at the TOA and the surface which compare favourably with more complex

400   models, generally falling within $\sim$ 25 W/m^2 of those simulated by NorESM2 and ERA5 (Fig. 6 c-d).

Northward heat transport via the atmosphere and ocean is depicted in Figure 7. The total heat transport in ZEMBA-PI is lower than NorESM2 in each hemisphere (Fig. 7a) because of reduced heat transport in the atmospheric layer (Fig. 7b). Ocean heat transport corresponds nicely with that inferred from NorESM2 (Fig. 7b) with a maximum value of $\sim$1.5 and $\sim$1 PW in the NH and SH, respectively. However, the location of maximum ocean heat transport is located at $\sim$ 30°N/S in the ZEMBA-
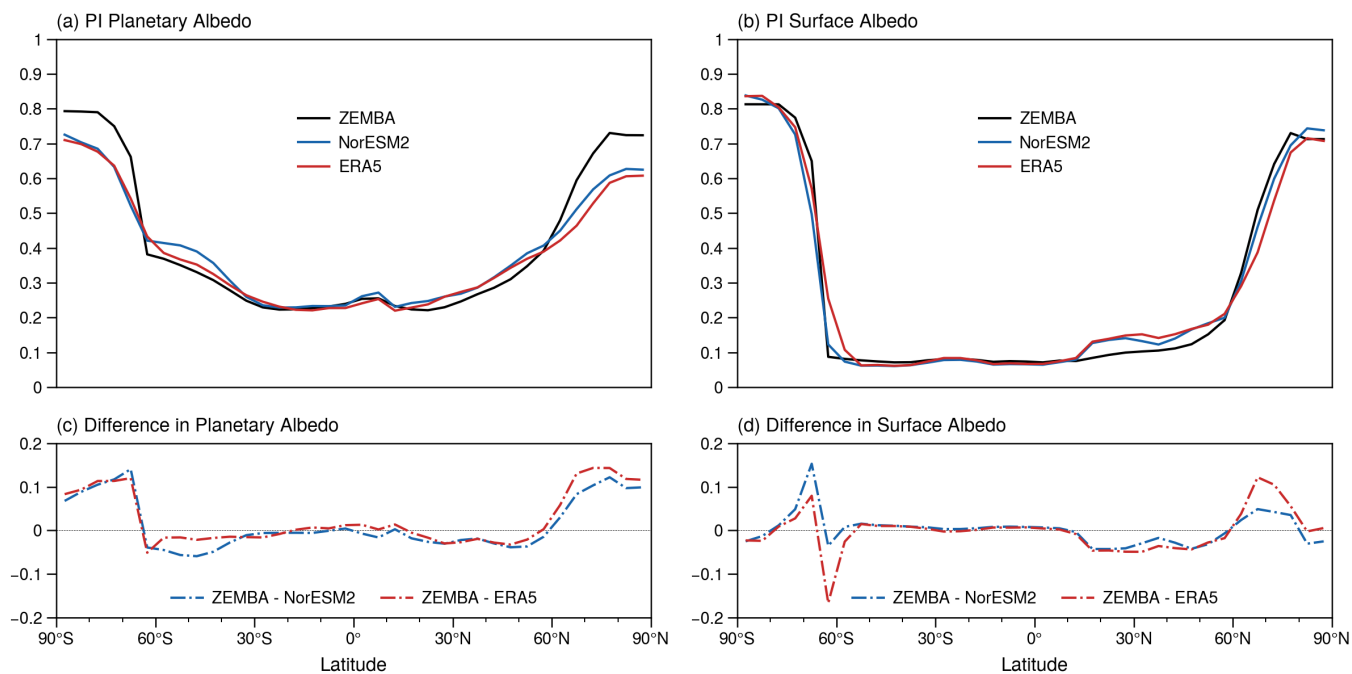
17

Figure 4. (top) The annual and zonal mean planetary (a) and surface (b) albedo for the pre-industrial period (PI), as simulated by ZEMBA (black lines) in comparison to NorESM2 (blue lines) and the ERA5 climatology from 1940 to 1970 (red lines). (bottom) The difference between ZEMBA and the other models/observations.

Table 3. Selection of annual mean variables from a pre-industrial simulation of ZEMBA in comparison to NorESM2 and ERA5 1940-1970 climatology

| Variable	ZEMBA	NorESM2	ERA5 (1940-1970)
Global Mean Surface Air Temperature ($^\circ$C)	13.87	13.78	13.72
Global Mean Planetary Albedo	0.31	0.32	0.31
Global Mean Surface Albedo	0.15	0.15	0.16
Snow cover ($\times 10^{13}$ m^2)	NH: 1.72. SH: 1.41	NH: 2.52. SH: 1.46	NH: 2.37 SH: 1.40
Sea ice cover ($\times 10^{13}$ m^2)	NH: 1.18 SH: 0.86	NH: 1.15 SH: 0.68	NH: 1.14 SH: 1.01
Global Mean Precipitation Rate (mm/day)	2.77	2.85	2.86
Global Mean Snowfall Rate (mm/day)	0.14	0.21	0.22
Peak Atmospheric Heat Transport (PW)	NH: 3.42 SH: 2.84	NH: 4.66 SH: 4.97	-
Peak Ocean Heat Transport (PW)	NH: 1.62 SH: 0.99	NH: 1.63 SH: 0.84	-
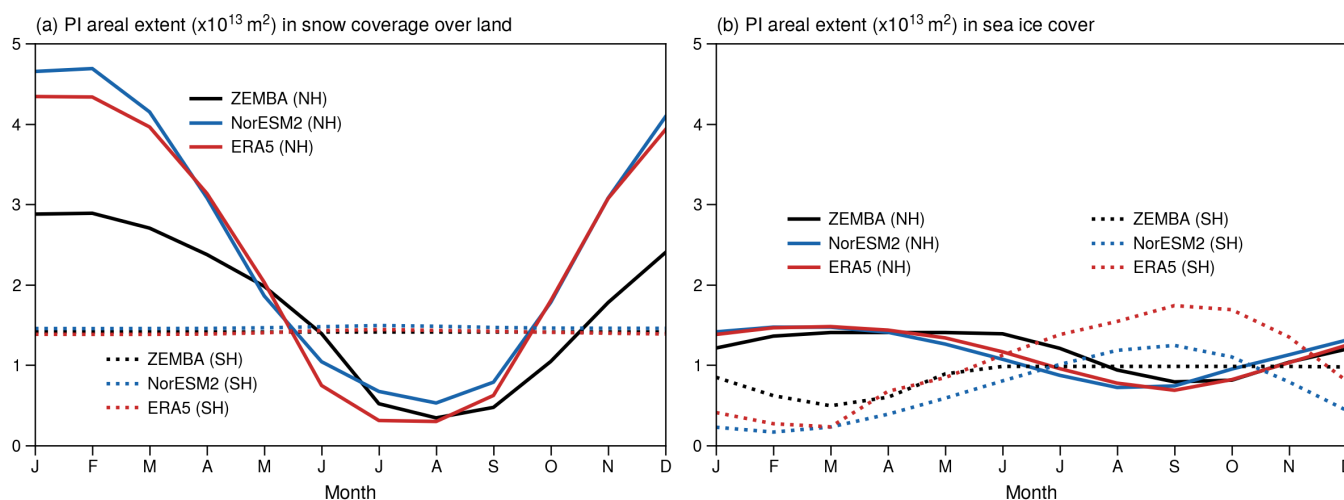
Figure 5. Monthly variations in the areal extent of snow coverage over land (a) and sea ice coverage over the ocean (b) for the pre-industrial period, as simulated by ZEMBA (black lines) in comparison to NorESM2 (blue lines) and the ERA5 1940-1970 climatology (red lines) for the northern hemisphere (solid lines) and the southern hemisphere (dotted lines). For snow cover over land, monthly averaged ERA5 reanalysis is taken over a shorter period from 1950 to 1970 (Muñoz Sabater, 2019) due to data availability.

405   PI experiment compared to $\sim 15°$N/S for NorESM2. While the maximum atmospheric heat transport is underestimated in ZEMBA-PI, the location of maximum heat transport corresponds nicely with NorESM2 at $\sim 45°$N/S. For the ZEMBA-PI experiment, latent heat transport contributes significantly to the poleward flux of atmospheric heat at the mid-latitudes, whereas dry static heat transport dominates the polar regions, in keeping with NorESM2 (Fig. 7c). Moreover, dry static transport peaks at $\sim 15°$N/S and latent heat transport peaks at $\sim 40°$N/S for both ZEMBA-PI and NorESM2. The inclusion of the Hadley cell

410   parameterization (Siler et al., 2018) successfully produces the equatorward flux of latent heat seen in the tropics. Overall, while dry static and latent heat transport in the atmosphere is underestimated in the ZEMBA-PI experiment compared to NorESM2, the overall configuration of atmospheric heat transport is very similar between the models.

3.2   2xCO$_2$ and +2% Insolation

A common method to test the sensitivity of climate models is to impose changes in atmospheric CO$_2$ concentrations or solar

415   insolation (S$_0$). Therefore, the temperature response to a doubling of atmospheric CO$_2$ concentrations (ZEMBA-2xCO$_2$) and a 2% increase in the solar constant (ZEMBA-S$_o$+2%) is shown in Figure 8, keeping all other boundary conditions the same as the ZEMBA-PI experiment. It should be noted that land ice fractions and elevations are kept fixed for these experiments. For a doubling of atmospheric CO$_2$ concentrations, the global mean temperature is 3.6°C higher. The most notable feature is that the NH is significantly more sensitive than the SH, reaching an annual mean rise of $>15°$C in the high northern latitudes, compared

420   to $<8°$C over Antarctica (Fig. 8a). In addition, temperature changes are strongest during the winter months of both hemispheres (Fig. 8b). The response to a 2% increase in the solar constant is very similar to a doubling of atmospheric CO$_2$ concentrations,
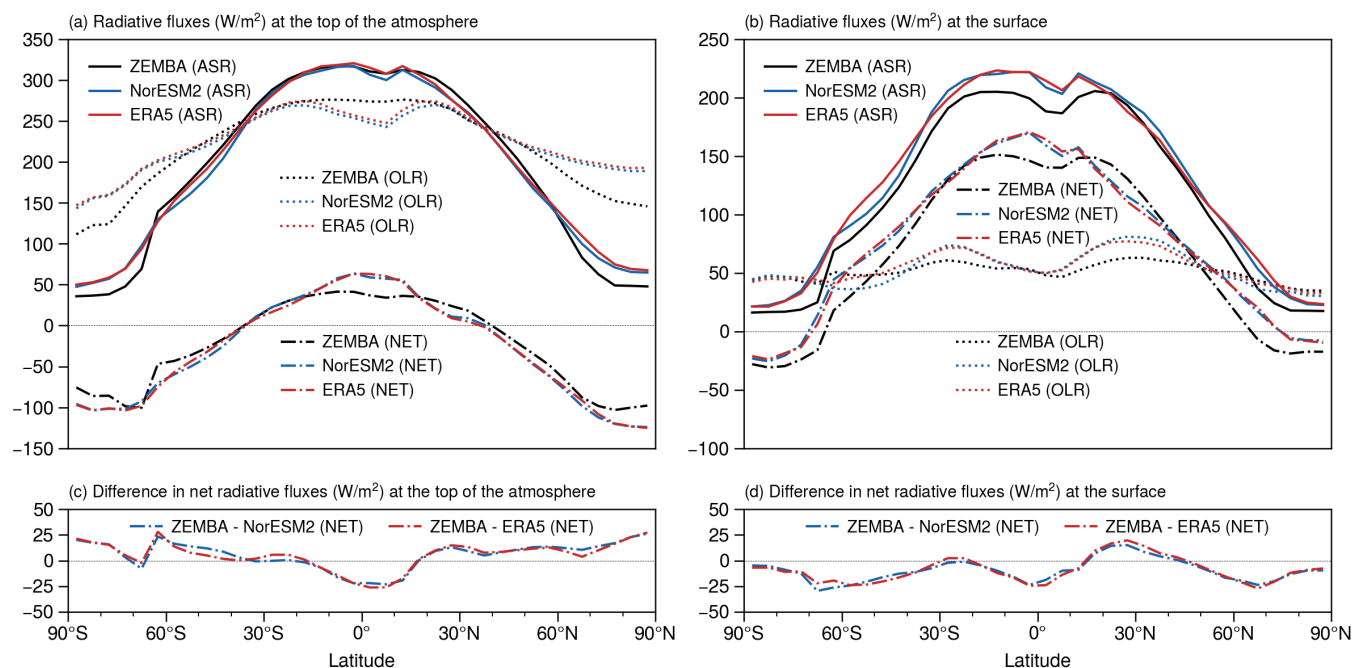
Figure 6. (top) The annual and zonal mean radiative fluxes at the top of the atmosphere (a) and the surface (b), as simulated by ZEMBA (black lines) in comparison to NorESM2 (blue lines) and the ERA5 1940-1970 climatology (red lines). Shown are the absorbed shortwave radiation (ASR: solid lines), outgoing longwave radiation (OLR: dotted lines) and net radiation (NET: dashed lines) radiative fluxes. The outgoing longwave radiation at the surface refers to net upwards longwave radiation, i.e., the upwards longwave radiative flux at the surface minus the downwards longwave radiative flux at the bottom of the atmospheric layer. (bottom) The difference in the net radiative flux between ZEMBA and the other models/reanalysis.

with an equivalent rise in global mean surface air temperature of 3.3°C, but with more muted warming over Antarctica. In reference to other works, comparisons are made complicated by the fact that GCM simulations involving a doubling or quadrupling of atmospheric CO_2 are often not run to their equilibrium climate state due to considerable computing times.
425 However, the global-mean warming from the ZEMBA-2xCO_2 experiment (3.6°C), known as the equilibrium climate sensitivity (ECS), fits comfortably within the 'likely' range of 1.5°-4.5°C estimated in the fifth assessment report of the Intergovernmental Panel on Climate Change (Collins et al., 2013). In addition, a polar amplification in surface warming that is strongest in the NH and during the winter months is consistent with both historical observations (1979-2014) and GCMs responding to an abrupt quadrupling of the atmospheric CO_2 concentration (Hahn et al., 2021).

430 **3.3 Last Glacial Maximum**

To test the ability of ZEMBA to simulate climates other than the pre-industrial period, we perform a simulation of the LGM (hereafter the ZEMBA-LGM experiment). The LGM remains a natural focus for climate models to constrain their climate
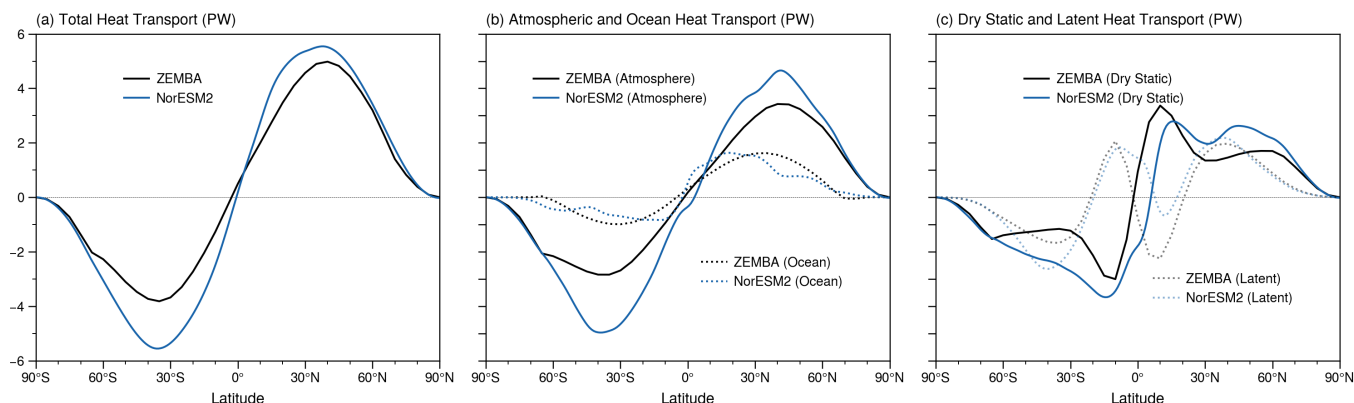
Figure 7. The northward transport of atmospheric and ocean heat in the pre-industrial period, as simulated by ZEMBA (black lines) in comparison to NorESM2 (blue lines). Shown is the total heat transport (a), the atmospheric and ocean components (b) and the partition of atmospheric heat transport into the dry static and latent components (c). For NorESM2, annual mean heat transport is inferred by assuming the system is in equilibrium and the heat transport is equal to energy imbalance at any latitude. More specifically, the energy imbalance at the top of the atmosphere (for total heat transport), in the atmosphere (for atmospheric heat transport) or at the surface (for ocean heat transport) is integrated from the south pole.

sensitivity because it was the most recent cold extreme when atmospheric CO_2 concentrations were ~ 100 ppm lower than the pre-industrial period (Bereiter et al., 2015) and continental ice sheets reached their maximum extent over North America

435 and Fennoscandia. For LGM boundary conditions, the model is forced with insolation using the orbital parameters from 21 kyr BP (Laskar et al., 2004), prescribed changes in land elevation and land ice fractions from ICE-6G_C (Argus et al., 2014; Peltier et al., 2014) and an atmospheric CO_2 concentration of 184 ppm (Bereiter et al., 2015) (see Table 2). The differences between the LGM and PI climates, as simulated by ZEMBA, is compared to an ensemble of state-of-the-art climate models which contributed to the PMIP3 and PMIP4 compiled by Kageyama et al. (2021). In addition, we compare to a recent LGM

440 reconstruction from Annan et al. (2022) which combines an ensemble of climate model simulations with proxy-based estimates of surface temperature using a data assimilation approach. In the ZEMBA-LGM experiment shown in Figure 9, there are no changes to the strength or the configuration of ocean circulation.

The changes in annual and zonal mean surface air temperatures for the ZEMBA-LGM simulation is shown in Figure 9a. The temperature decrease around the tropics (30°S-30°N) of -1.95°C is comparable to the other reconstructions, but falls slightly

445 on the lower end of estimated cooling (Table. 4). In the extratropics, however, ZEMBA appears to generate cooling which is slightly too strong in the NH and too mild in the SH, when compared to the multi-model averages from PMIP3, PMIP4 and the Annan et al. (2022) reconstruction. The latitudinal distribution of cooling always falls within the range simulated across the PMIP3/PMIP4 model ensemble (yellow shaded area: Fig. 9), but this is not always the case at the higher latitudes for the Annan et al. (2022) reconstruction (blue shaded area: Fig. 9) which is constrained by proxies for surface temperatures.

450 Overall, despite some discrepancies in the higher latitudes, the global mean cooling from the ZEMBA-LGM experiment is
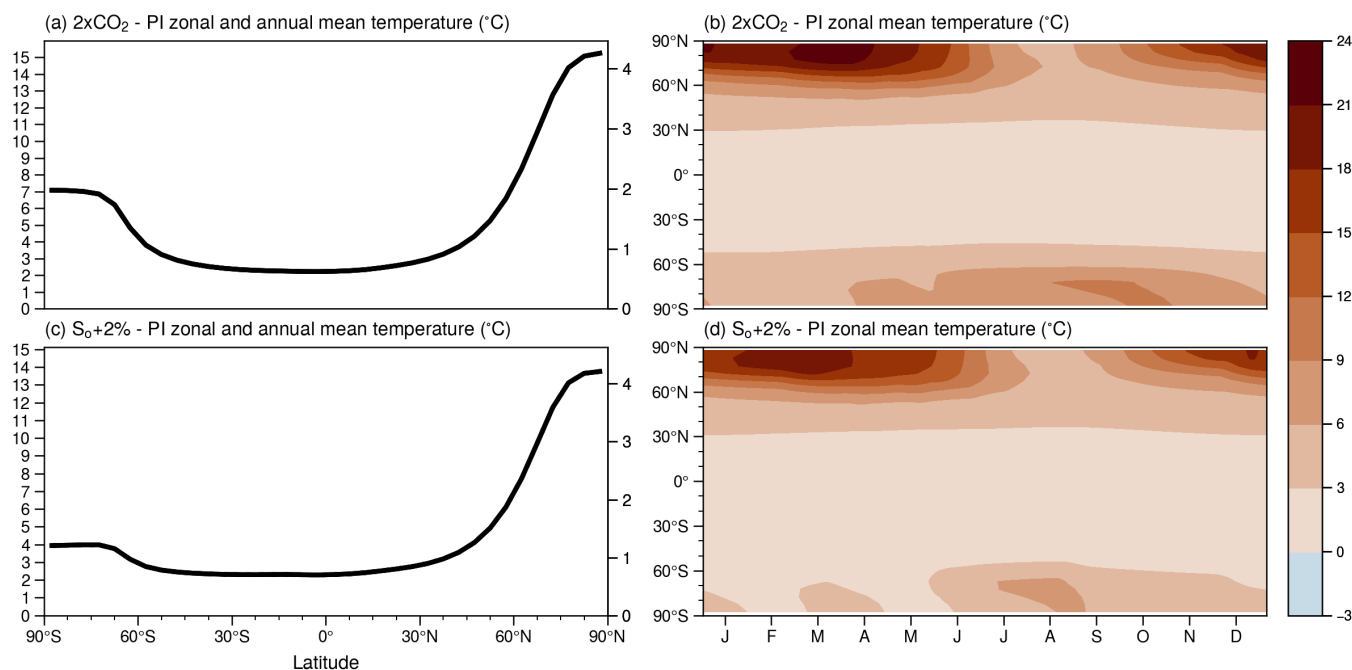
Figure 8. Changes in surface air temperature (relative to the PI) following a doubling of the atmospheric CO_2 concentration from 284 ppm to 568 ppm ($2xCO_2$) and a 2% increase in the solar constant (S_o+2%). (a,c) Changes in the annual mean temperature (left) also shown normalized by the global mean warming (right). (b,d) Changes in the seasonal cycle in temperature.

-4.11°C compared to -4.71°C from PMIP3, -4.77°C from PMIP4 and -4.46°C from Annan et al. (2022) (Table 4). For LGM - PI precipitation rates (Fig. 9b), ZEMBA captures the widespread decrease in precipitation, which again fall within the range estimated by the PMIP3/PMIP4 model ensemble. The reductions are largest in the NH extratropics, similar to surface air temperatures. However, there are a number of discrepancies between ZEMBA LGM - PI precipitation rates and the PMIP3 and

455 PMIP4 model ensembles, such as at 5-10°N where the PMIP ensemble generate a much stronger reduction in precipitation, and at 30-10°S where the PMIP ensemble simulates higher zonal mean precipitation at the LGM due to increased precipitation in the subtropical Pacific Ocean (Kageyama et al., 2021). Overall, the global mean decrease in precipitation is -0.24 mm/day for ZEMBA compared to -0.29 mm/day in PMIP3 and -0.33 mm/day in PMIP4 (Table 5)

The differences in northward heat transport between the ZEMBA-LGM and ZEMBA-PI experiment is shown in Figure 9

460 c-d. Starting with total heat transport (Fig 9c), the most notable feature is a significant peak in northward heat transport (of ∼ 0.5 PW) at 45°N, followed by a significant trough (of ∼ 0.3 PW) at 60°N. The peak is associated with greater atmospheric heat transport in the LGM experiment (Fig 9c-d), which is almost entirely because of larger fluxes of dry static energy in the atmospheric layer (Fig 9d). The reduced northward heat transport at 60°N, on the other hand, is due to a significant 0.5 PW decrease in ocean heat transport (Fig 9c), associated with sea ice expansion in the LGM experiment. As the surface ocean layer

465 underlying sea ice rests at the freezing point of sea water (see Section 2.3.1), there is effectively zero meridional ocean heat
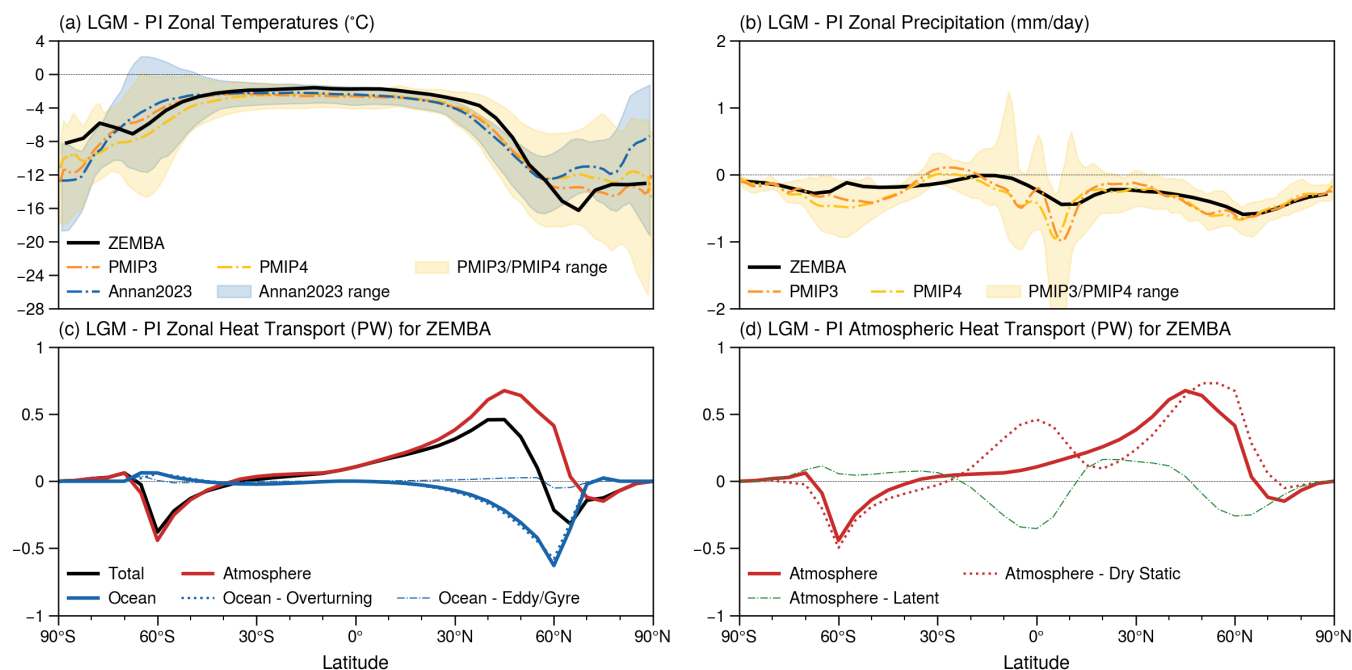
22

Figure 9. (top) The LGM – PI surface air temperatures (a) and precipitation (b) as simulated by ZEMBA (solid black line) in comparison to the multi-model mean from PMIP3 (dashed orange line) and PMIP4 (dashed yellow line) and the data assimilation products from Annan et al. (2022) (dashed blue line). The yellow and blue shaded areas represent the range of LGM-PI surface air temperatures and precipitation as reconstructed by PMIP3/PMIP4 and Annan et al. (2022), respectively. (bottom) The differences in northward heat transport between the LGM and PI experiments for ZEMBA, including: total heat transport (solid black line); atmospheric heat transport (solid red line); ocean heat transport (solid blue line); the decomposition of ocean heat transport into ocean overturning (dotted blue line) and eddy/gyre transport (dashed blue line) (c); and the decomposition of atmospheric heat transport into dry static (red dotted line) and latent (green dashed line) heat transport (d).

fluxes at latitudes covered by sea ice. Consequently, sea ice expansion is accompanied by large drops in ocean heat transport, via both overturning and eddy/gyre transport, at latitudes which were previously free of sea ice in the PI experiment.

In addition to the standard ZEMBA-LGM simulation, we recreate the sensitivity experiment from Bintanja and Oerlemans (1996) by progressively adding each of the LGM boundary conditions (shown in Table 2) followed by changes to the strength of ocean circulation (Fig. 10). Using the EBM from Bintanja (1997), Stap et al. (2014) also found that introducing a mechanism to shift the mid-point of ocean circulation further south during glacial conditions was necessary to produce surface temperatures in closer agreement with observation. Therefore, we additionally investigate the impact of changing the mid-point of ocean circulation from $5°S$ to $15°S$. To summarize, we progressively add each of the following:

- **Ice**: LGM land-ice fractions and zonal mean land elevations.
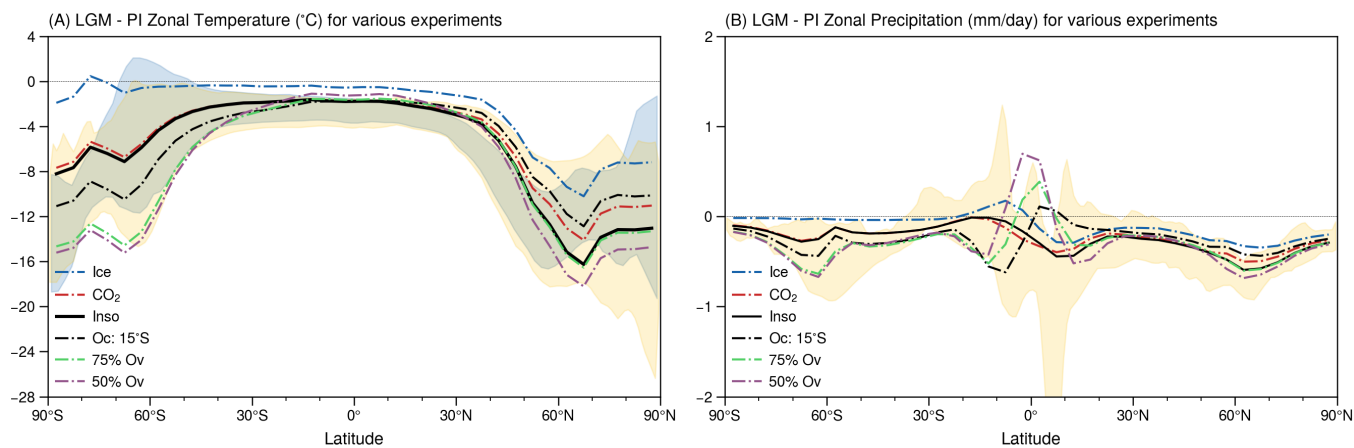- **CO_2**: LGM CO_2 concentrations.

Figure 10. The LGM - PI surface air temperatures (A) and precipitation rates (B) for a series of experiments, which involves progressively adding: LGM land ice fractions and zonal mean land elevations (blue dashed line); LGM CO_2 concentrations (red dashed line); LGM insolation (black line); a shift in the mid-point of ocean circulation from 5°S to 15°S (black dashed line): a 75% decrease in ocean overturning (green dashed line) strength; and 50% decrease in ocean overturning strength (purple dashed line).

- **Inso**: LGM insolation forcing - this is the same as the standard ZEMBA-LGM experiment.
- **Oc: 15°S**: A shift in the mid-point of ocean circulation from 5°S to 15°S.
- **75% Ov**: A 25% reduction in the strength of ocean overturning.
- **50% Ov**: Another 25% (50% total) reduction in the strength of ocean overturning.

480   The changes in temperature and precipitation - averaged over the tropics, extratropics and globally - following the addition of each boundary condition is shown in Table 4 and 5, respectively. Firstly, the addition of LGM land ice fractions and elevation causes a large decrease in temperature (-5.06°C) and precipitation (-0.22 mm/day) in the NH extratropics, whereas the effects in the tropics and the SH extratropics is minimal. A following reduction in atmospheric CO_2 concentrations generates a widespread cooling, drying and a further reduction in global mean temperature of -2.08°C, which is stronger than

485   the initial cooling caused by the LGM ice sheets (-1.69°C). The subsequent addition of the LGM insolation forcing causes a further decrease in global mean temperature of 0.34°C. As in Stap et al. (2014), a shift in the mid-point of ocean circulation to 15°S produces surface temperatures which are in better agreement with LGM reconstructions. In particular, shifting the ocean circulation mid-point 10° further south decreases temperature and precipitation rates in the SH extratropics and increases temperature and precipitation in the NH extratropics, with negligible changes in global mean temperature and precipitation.

490   Finally, reductions in the strength of overturning circulation causes a simultaneous warming in the tropics and cooling in the higher latitudes of each hemisphere, which generates a global mean cooling and a larger disagreement with the PMIP3/PMIP4 ensemble and LGM reconstruction from Annan et al. (2022).

Table 4. Difference in zonal and annual mean surface air temperatures (°C) between the LGM and the PI. *ZEMBA (Inso)* refers to the standard ZEMBA-LGM experiment shown in Fig. 9

Model	90°-30°S	30°S-30°N	30°-90°N	Global
ZEMBA				
Ice	-0.45	-0.62	-5.06	-1.69
CO_2	-3.53	-1.93	-7.69	-3.77
Inso	-3.69	-1.95	-8.83	-4.11
Oc: 15°S	-5.63	-1.99	-6.83	-4.11
75% Ov	-7.74	-1.90	-8.88	-5.11
50% Ov	-8.03	-1.56	-9.91	-5.26
Other models and reconstructions				
PMIP3	-3.81	-2.79	-9.46	-4.71
PMIP4	-4.81	-2.68	-8.91	-4.77
Annan2022	-3.57	-2.63	-9.02	-4.46

Table 5. Difference in zonal and annual mean precipitation rates (mm/day) between the LGM and the PI. *ZEMBA (Inso)* refers to the standard ZEMBA-LGM experiment shown in Fig. 9

Model	90°-30°S	30°S-30°N	30°-90°N	Global
ZEMBA				
Ice	-0.03	-0.07	-0.22	-0.10
CO_2	-0.18	-0.20	-0.34	-0.23
Inso	-0.18	-0.20	-0.38	-0.24
Oc: 15°S	-0.29	-0.20	-0.29	-0.25
75% Ov	-0.35	-0.17	-0.36	-0.27
50% Ov	-0.35	-0.11	-0.39	-0.24
Other models and reconstructions				
PMIP3	-0.25	-0.25	-0.42	-0.29
PMIP4	-0.30	-0.30	-0.43	-0.33

4 Discussion

Our simulation of the pre-industrial demonstrates that ZEMBA can describe the zonally-averaged climate of this period with
495 reasonable accuracy. Surface air temperatures are in strong agreement with NorESM2 and ERA5 1940-1970 both annually

and seasonally (Fig. 2a-c), generally falling within 4°C of those estimated by NorESM2 (Fig. 2d-f). Most notably, given the simplicity of ZEMBA, the model shows success in emulating precipitation and snowfall rates for the pre-industrial climate (Fig. 3a), particularly at the polar latitudes. Furthermore, the model captures the poleward enhancement of the surface and planetary albedo due to the presence of snow and sea ice cover (Fig. 4), with the seasonal range of sea ice cover in the NH corresponding

500   nicely with both NorESM2 and ERA5, varying between ∼0.5 and ∼1.5 x10^{13} m^2 (Fig. 5b). The ocean transport model shows success in emulating zonally averaged ocean heat transport, which makes significant contribution to total heat transport at the lower latitudes, peaking with values of around 1 and 1.5 PW for the SH and NH, respectively (Fig. 7b). Similarly, the model predicts that atmospheric heat transport dominates in the mid-to-high latitudes and reaches its maximum levels around ∼ 45° in each hemisphere. Finally, looking specifically at the partition of atmospheric heat transport into its dry static and latent

505   components, the model captures the relative contributions of each flux to the total atmospheric heat transport, with dry static fluxes peaking at ∼ 15°N/S and latent heat transport peaking at 45°N/S.

    Despite these favourable comparisons with state-of-the-art climate models and reanalysis data, there are limitations of ZEMBA in its current state. For example, while the model does reasonably well at capturing the radiative fluxes at the TOA and the surface (Fig. 6), one of the largest discrepancies between ZEMBA and NorESM2 or ERA5 1940-1970 resides in the

510   outgoing longwave radiation at the TOA. As the outgoing longwave flux is overestimated around the tropics, the net radiation received at these latitudes is up to ∼ 25 W/m^2 lower than NorESM2 and ERA5 1940-1970 (Fig. 6c). Given the comparatively large surface area in the tropics, the underestimation of these net radiative fluxes results in a significant reduction of the surplus radiative energy which needs to be transported poleward. Consequently, both the total and atmospheric heat transport in ZEMBA is much lower (Fig. 7a-b), with atmospheric heat transport peaking at ∼ 3 PW in each hemisphere, which should be

515   closer to 5 PW (Trenberth and Fasullo, 2017). In addition, ZEMBA underestimates the winter maximum in snow cover over land (Fig. 5a), which results in a smaller seasonal amplitude in snow cover. Similar to NH snow cover, the seasonal range of sea ice cover in the SH is reduced when compared to NorESM2 and ERA5 1940-1970 (Fig. 5b) and snowfall rates over Antarctica are underestimated by about 50% (Fig. 3a).

    Given their simplicity, the ability of EBMs like ZEMBA to accurately reproduce the latitudinal pattern of surface temperature

520   is impressive but well-established since the original works by Budyko (1969) and Sellers (1969). In reference to other studies, ZEMBA is most similar to later iterations of EBMs that contain vertical resolution (i.e. an atmospheric and surface layer), the division of the surface into land and ocean, and a seasonal cycle in insolation (Peng et al., 1987; Harvey, 1988; Bintanja, 1997). In comparison to the 'present-day' simulation from the EBM developed by Bintanja (1997), which effectively serves as the basis for ZEMBA, most climate variables exhibit qualitative agreement with the pre-industrial output shown in Section

525   3.1. Although, the global mean surface air temperature reported in Bintanja (1997) is 1.09°C warmer due to a higher 350 ppm CO$_2$ forcing in their model experiment. Other than the hydrological cycle, the primary contrast between the models lies in their representation of land-based snow cover. In the Bintanja (1997) study, the seasonal cycle in snow cover is parameterised as a function of surface air temperature, and appears in better agreement with observations than ZEMBA, which explicitly calculates snow coverage over land. Overall, many such features of the Earth's climate are relatively well produced by various

530   EBMs (Peng et al., 1987; Harvey, 1988; Bintanja, 1997) including ZEMBA, in part because they are 'tuned' to match the

present-day climate. Of greater importance for EBMs than achieving an exact replication of observations and/or GCM output, is the investigation of different climate processes and feedbacks, which necessitates appropriate sensitivities to alterations in external or internal forcings.

The response of ZEMBA to both a doubling of the atmospheric CO_2 concentration and a 2% increase in solar insolation are in qualitative and quantitative agreement to one another (Fig. 8), including a polar amplification of warming, which is stronger in the NH than the SH, and concentrated in the winter months. The climate sensitivity of ZEMBA is larger than the EBM of Bintanja (1997) and similar such studies (Peng et al., 1987). For example, the S_o+2% experiment generates a much stronger annual-mean warming in the high northern latitudes ($>14°$) than Bintanja (1997) ($<5°$). In addition, the ECS of ZEMBA ($3.6°$C) is higher than the original model of 1.9-2.2$°$C (Bintanja, 1997; Stap et al., 2014). However, more recent EBMs have generated a global-mean and polar amplified warming in response to CO_2 forcing which is similar or even greater in magnitude than ZEMBA (Roe et al., 2015; Södergren et al., 2018; Feldl and Merlis, 2021). Moreover, the ECS of ZEMBA is consistent with the range of ECS projected by GCMs that contributed to the Coupled Model Intercomparison Projects, phase 5 (CMIP5) and CMIP6 (Flato et al., 2013; Zelinka et al., 2020), and other works (Collins et al., 2013). The strong seasonal asymmetry of surface warming in the polar regions (Fig. 8b-d) is in accordance with both observations (Screen and Simmonds, 2010) and GCM results (Holland and Bitz, 2003). As in Bintanja (1997), the winter maximum in surface air temperature is related to sea ice loss, causing a greater absorption of shortwave radiation, which is released from the ocean surface into the overlying atmosphere during the winter months. For the 2xCO_2 experiment, the polar amplification in warming - normalized (divided) by the global mean warming - exceeds 4 in the high latitudes of the Arctic and reaches up to 1.5 in the Antarctic (Fig. 8a). While a polar amplification in warming that is strongest in the NH is in agreement with both observations and GCM results, a warming in the high northern latitudes that reaches 4 times the global mean resides in the upper boundary of estimates from GCM simulations (Holland and Bitz, 2003; Hahn et al., 2021). When averaged from $60°$ to $90°$N, however, the normalised Arctic warming is 3.2, which is still less than the 3.5 estimated over the historical period from 1979 to 2014 according to the observational dataset HadCRUT5 (Hahn et al., 2021). Overall, we see ZEMBA exhibits a climate sensitivity - in terms of both global mean and polar amplified warming - that appears broadly consistent with other EBMs, GCMs and observations.

There are comparatively few EBMs which incorporate a hydrological cycle (Jentsch, 1991) and none – to our knowledge – used for studies of glacial-interglacial cycles. Jentsch (1991) developed a EBM, consisting of a vertically averaged atmosphere overlying an ocean layer, to study the influence of the hydrological cycle on climate. Using an optimization procedure, the modelled precipitation and evaporation, amongst other climate variables, compares very well against contemporary observations (Jentsch, 1991). While both ZEMBA and the work of Jentsch (1991) contain a hydrological cycle, ZEMBA includes land cover and a seasonal cycle, which is suitable for investigating the response of climate to variations in the orbital parameters. Previous EBMs used for studies of glacial-interglacial cycles prescribe either a present-day (Pollard, 1978) or spatially uniform (Huybers and Tziperman, 2008) distribution of precipitation, which is not perturbed across climates, or instead parameterize precipitation and snowfall as a function of surface air temperatures and ice sheet size (Stap et al., 2014). The inclusion of a hydrological cycle enables precipitation and snowfall to be calculated internally in ZEMBA, which can account for the influence of changes in both local air temperatures and poleward moisture transport for precipitation and snowfall rates. It has been

suggested that changes in the Earth's obliquity – by altering the meridional gradient in insolation – has a relatively stronger influence on ice sheet volume due to changes in poleward moisture transport (Raymo and Nisancioglu, 2003; Nisancioglu, 2004). The inclusion of a hydrological cycle enables ZEMBA to explore these processes in the context of glacial-interglacial cycles.

570   It is important that ZEMBA, intended for studies of glacial-interglacial cycles, can simulate climates other than the present-day or pre-industrial period. When our simulation of the LGM is compared to other reconstructions, the model compares favourably for changes in both surface air temperature (Fig 9a) and precipitation (Fig 9b). Indeed, ZEMBA captures the polar amplification of cooling in both hemispheres due to positive feedbacks relating to snow cover and sea ice expansion, although this cooling appears somewhat underestimated around Antarctica and slightly overestimated in the northern high

575   latitudes. Moreover, the global mean cooling of -4.11°C is similar to that estimated by the ensemble averages from PMIP3 and PMIP4 (Kageyama et al., 2021), and the data assimilation product from Annan et al. (2022). It should be noted that other data assimilation reconstructions suggest a much stronger global cooling of between 6.1° and 6.8°C (Tierney et al., 2020; Osman et al., 2021), which perhaps signals the importance of including feedbacks relating to clouds, dust and/or vegetation for reproducing the LGM cooling. However, the reconstructions from Tierney et al. (2020) and Osman et al. (2021) are based on a

580   single climate model (CESM1.2) which produces one of the coldest LGM climates in PMIP4, whereas the Annan et al. (2022) reconstruction incorporates the wide range of climates generated across the PMIP ensemble.

Evaluating changes in meridional heat transport during the LGM is made difficult by the large (and often conflicting) range of total, atmospheric and ocean heat transport generated across PMIP3 and PMIP4 at all latitudes (Kageyama et al., 2021). For the ZEMBA-LGM experiment, changes in total heat transport are most intense in the NH (Fig 9c), in response to a large

585   steepening of the equator-to-pole temperature gradient. The upsurge in total heat transport between 15°-45°N is somewhat consistent with PMIP3/PMIP4 (Kageyama et al., 2021), although most of this increase in ZEMBA is generated by stronger atmospheric heat transport (Fig. 9c-d), whereas PMIP3/PMIP4 also shows stronger ocean heat transport at these latitudes. Stronger ocean heat transport in PMIP3/PMIP4 can be attributed to a stronger and sometimes deeper Atlantic Meridional Overturning Circulation (AMOC) (Kageyama et al., 2021), which is inconsistent with proxy-based reconstructions of ocean

590   circulation (Lynch-Stieglitz et al., 2007; Gebbie, 2014; Du et al., 2020), and is associated with stronger surface winds over the northern North Atlantic (Muglia and Schmittner, 2015). For ZEMBA, on the other hand, there is a large decreases in NH ocean heat transport during the LGM, reaching -∼ 0.5 PW at 60°N (Fig. 9c). The prominent reduction in ocean heat transport, not observed in any of the PMIP3/PMIP4 simulations, is associated with the expansion of sea ice and a surface ocean layer that resides near the freezing point of sea-water. Consequently, horizontal heat fluxes due to both advection (representing

595   overturning) and diffusion (representing eddies and gyres) drops to zero at latitudes now covered by sea ice, which perhaps highlights the limitation of the simplified ocean model used in ZEMBA, where meridional heat transport is limited to the surface and bottom ocean layers.

The sensitivity experiments performed for the LGM boundary conditions (Fig. 10) recreates the experiment made by Bintanja and Oerlemans (1996) using the original EBM. For ZEMBA, the global mean cooling caused by the combined ice sheet and

600   CO_2 forcing is 4.11°C, which is stronger than the 3.3°C of cooling noted by Bintanja and Oerlemans (1996), and is in better

agreement with LGM reconstructions. The stronger cooling in ZEMBA can be partially attributed to differences in the LGM boundary conditions used in this study, which involves a stronger CO_2 reduction and larger LGM ice sheet extents than Bintanja and Oerlemans (1996). In our study, the addition of LGM ice sheets generates cooling and drying that is primarily localized to the NH, whereas CO_2 lowering generates widespread cooling and drying with a stronger global-mean cooling than that induced by ice sheet expansion (Table 4 & 5). Additionally, by shifting the mid-point of ocean circulation from 5°S to 15°S, surface air temperatures are in better agreement with the LGM reconstructions, as first noted by Stap et al. (2014). As for the strength of ocean circulation, Bintanja and Oerlemans (1996) found that reducing flow velocities in their ocean model led to temperatures in better agreement with contemporary LGM temperature reconstruction. In this study, however, reductions in ocean overturning instead leads to larger discrepancies with modern LGM constructions. Indeed, it remains uncertain to what degree overturning changed during glacial conditions, with some studies suggesting the mean state of overturning was not 'sluggish' across glacial cycles (Bohm et al., 2015) or the LGM (Lynch-Stieglitz et al., 2007). Therefore, for simulations of glacial-interglacial cycles using ZEMBA, it may be important to consider the impact of variations in both the mid-point and overall strength of ocean circulation.

ZEMBA is intended to be used as a computationally efficient tool for studies of the glacial-interglacial cycles of the Quaternary. The PI and the LGM experiments indicate ZEMBA is able to simulate glacial and interglacial climate states for a given insolation, CO_2 and ice sheet extent. As the model includes: (1) both hemispheres, (2) a seasonal cycle and (3) a hydrological cycle; it is able to explore mechanisms invoked to explain the dearth of precession cycles in ice volume observed during the Early Pleistocene, such as: (1) out-of-phase precession cycles between the hemispheres (Raymo et al., 2006), (2) a counterbalancing between summer insolation intensity and summer duration (Huybers, 2006) or (3) obliquity-induced variations in atmospheric moisture transport (Raymo and Nisancioglu, 2003). In future work, we intend to explore both the equilibrium and transient response of ZEMBA to changes in the Earth's obliquity and precession, prior to simulations of the glacial-interglacial cycles of the Early Pleistocene via coupling to an ice sheet model.

5 Conclusions

In this study, a simple climate model (ZEMBA) is introduced to simulate zonally-averaged climate fields including surface temperatures and precipitation. ZEMBA is largely built on the zonally-averaged energy balance climate model from Bintanja (1997), comprising an atmospheric layer overlying a surface divided into a land component and a six-layered, zonally-averaged ocean transport model. Unlike its predecessor, ZEMBA incorporates a hydrological cycle to estimate snowfall and precipitation with latitude.

Simulations of the pre-industrial period compare favourably with GCMs and reanalysis data, including surface temperatures (and their seasonal cycle), precipitation, surface and TOA radiative fluxes, sea ice, snow cover and meridional heat transport. However, the underestimation of the net TOA radiation received at the tropics leads to a reduction in atmospheric heat transport, and there is also an underestimation of the seasonal amplitude in snow cover over land.

The response of ZEMBA to increases in the atmospheric CO_2 concentration or the solar constant are in qualitative agreement to other EBMs, GCMs and observations, such a polar amplification in surface warming which is strongest over the NH and

635 focused in the winter months. The new additions to ZEMBA appear to increase climate sensitivity in comparison to older EBMs, but it is still broadly consistent with the global-mean and polar amplified warming projected by climate models of higher complexity.

As the purpose of ZEMBA is for studies of the glacial-interglacial cycles of the Quaternary, it is important the model can simulate climates other than present-day or pre-industrial. A simulation of the LGM indicates ZEMBA is able to capture changes

640 in surface temperature and precipitation in qualitative and quantitative agreement with state-of-the-art climate models and data assimilation products, despite neglecting climate feedbacks relating to dust, vegetation and clouds. In particular, ZEMBA reproduces the polar amplification of cooling in both hemispheres and a global mean cooling in accord with reconstructions from more elaborate models.

The overall conclusion from this study is ZEMBA is suitable for studies of climatic change on large spatial and temporal

645 scales, with a particular emphasis on glacial-interglacial cycles and the response of the climate system to changes in the orbital parameters. In future work, we intend to explore both the equilibrium and transient response of ZEMBA to changes in the Earth's obliquity and precession, prior to simulations of the glacial-interglacial cycles of the Early Pleistocene via coupling to an ice sheet model.

Code and data availability. Source code is maintained on GitHub at https://github.com/daniel-francis-james-gunning/zemba (last access: 8

650 May 2024) with the exact version used in this study (including scripts for creating all figures) archived on Zenodo at https://zenodo.org/doi/10.5281/zenodo.11155259

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Author contributions. DFJG designed the model and ran the experiments. DFJG, KHN, EC and RSWvdW analysed the results. DFJG drafted the paper, with input from all co-authors.

655 *Competing interests.* The contact author has declared that none of the authors has any competing interests.

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