Editor Comments:

The authors have thoroughly and satisfactorily responded to the suggestions and comments by the reviewers. Hence I am glad to accept this manuscript for publication in Climate of the Past.

Before final publication I would like the authors to reconsider a minor, but potentially misleading formulation. The authors state in their conclusion (lines 701, 702) that "the simulations provide insights into key dynamics features of the transition, ..." Time slice simulations cannot, in principle, address the dynamics of the transition of the climate system as pointed out by reviewer 1. They can however, as correctly replied by the authors, provide insights into different climate states, i.e. modes of operation, or dynamics, of the atmosphere and the ocean, during the transition. I ask the authors for modifying their formulation.

Thank you for the opportunity to revise this section to avoid confusion. We changed our final paragraph in the conclusion from:

The eight time slice simulations depict the glacial-interglacial transition that is in good agreement with other AOGCM simulations and compares reasonably well with data-based climate reconstructions. The simulations provide insights into key dynamic features of the transition, such as altered NH storm tracks and strengthening of monsoons during the early to mid-Holocene. The data-model and model-model comparisons give us a measure of confidence that our paleo GENMOM simulations are reasonable on broad spatial scales and adds to the growing number of climate models that are capable of simulating key aspects of past climate change when constrained by a relatively small set of global boundary conditions. Future work using the model output produced by this study will address how internal model variability and multidecadal variability influence comparison with proxy data, particularly in North America using dynamical downscaling techniques.

To:

The eight time slice simulations depict the glacial-interglacial transition that is in good agreement with other AOGCM simulations and compares reasonably well with data-based climate reconstructions. The data-model and model-model comparisons give us a measure of confidence that our paleo GENMOM simulations are reasonable on broad spatial scales and adds to the growing number of climate models that are capable of simulating key aspects of past climate change when constrained by a relatively small set of global boundary conditions. While our simulations are not continuous across the deglaciation and do not include events such as freshwater forcing, they do provide insights into between period changes, such as altered NH storm tracks and strengthening of monsoons during the early to mid-Holocene and multi-century time series that are useful, for example, to explore ecosystem responses to changes in mean climate and the related interannual variability in the model. Future work using the model output produced by this study will address

how internal model variability and multidecadal variability influence comparison with proxy data, particularly in North America using dynamical downscaling techniques.

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3 Global climate simulations at 3,000-year intervals for the last 21,000 years

4 with the GENMOM coupled atmosphere-ocean model

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Abstract

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36   We apply GENMOM, a coupled atmosphere-ocean climate model, to simulate

37   eight equilibrium time slices at 3000-yr intervals for the past 21,000 years forced by

38   changes in Earth-Sun geometry, atmospheric greenhouse gases (GHGs), continental ice

39   sheets and sea level. Simulated global cooling during the Last Glacial Maximum (LGM)

40   is 3.8 °C and the rate of post-glacial warming is in overall agreement with recently

41   published temperature reconstructions. The greatest rate of warming occurs between 15

42   and 12 ka (2.4 °C over land, 0.7 °C over oceans and 1.4 °C globally) in response to

43   changes in radiative forcing from the diminished extent of the Northern Hemisphere

44   (NH) ice sheets and increases in GHGs and NH summer insolation. The modeled LGM

45   and 6 ka temperature and precipitation climatologies are generally consistent with proxy

46   reconstructions, the PMIP2 and PMIP3 simulations, and other paleoclimate data-model

47   analyses. The model does not capture the mid-Holocene 'thermal maximum' and gradual

48   cooling to pre-industrial global temperature found in the data. Simulated monsoonal

49   precipitation in North Africa peaks between 12 and 9 ka at values ~50% greater than

50   those of the PI, and Indian monsoonal precipitation peaks at 12 and 9 ka at values ~45%

51   greater than the PI. GENMOM captures the reconstructed LGM extent of NH and

52   Southern Hemisphere (SH) sea ice. The simulated present-day Antarctica Circumpolar

53   Current (ACC) is ~48% weaker than the observed (62 versus 119 Sv). The simulated

54   present-day Atlantic Meridional Overturning Circulation (AMOC) of 19.3 ± 1.4 Sv on

55   the Bermuda Rise (33°N) is comparable with observed value of 18.7 ± 4.8 Sv. AMOC at

56   33°N is reduced by ~15% during the LGM, and the largest post-glacial increase (~11%)

57   occurs during the 15 ka time slice.

1   Introduction

58
59   The history of the climate system over the past 21,000 years reflects the combined

60   changes in earth-sun orbital geometry, atmospheric greenhouse gas concentrations

61   (GHG), the extent of the Northern Hemisphere (NH) ice sheets, and sea level. GHG

62   levels were lowest during the Last Glacial Maximum (LGM, ~21,000 years ago, 21 ka)

63   and increased thereafter to pre-industrial (PI) levels (Brook et al., 2000; Monnin et al.,

64   2001; Sowers et al., 2003). The LGM is further characterized by the large Laurentide

65   (LIS), Cordilleran (CIS) and Fennoscandian (FIS) ice sheets. The height and extent of the

66   ice sheets altered atmospheric circulation patterns, and the extent increased the NH

67   albedo thereby altering the global radiative balance. The effect of the ice sheets on

68   climate progressively diminished from the LGM to the early Holocene as global warming

69   driven by increasing GHGs combined with changes in NH summer insolation to

70   accelerate ice sheet ablation. Abrupt departures from the comparatively smooth transition

71   from the LGM through the Holocene, such as Heinrich and Dansgaard-Oeschger events,

72   the Bølling-Allerød (BA), and the Youger Dryas (YD), are evident in geologic records,

73   and these events likely influenced the overall trajectory of the deglaciation.

74   The climate of the past 21,000 years has been studied extensively, beginning with

75   three international collaborative projects: the Long range Investigation, Mapping, and

76   Prediction (CLIMAP; CLIMAP Project Members, 1981) and the Cooperative Holocene

77   Mapping Project (COHMAP; COHMAP Members, 1988), which evolved into the

78   Testing Earth System Models with Paleoenvironmental Observations (TEMPO) project

79   (Kutzbach et al., 1996a; 1998). CLIMAP focused on reconstructing the LGM climate,

80   COHMAP focused on reconstructing the climate of seven time periods (18, 15, 12, 9, 6, 3

81 ka), and TEMPO focused on reconstructing the climate of 21, 16, 14, 11 and 6 ka. These

82 three projects pioneered data-model comparison through integrating climate model

83 simulations and paleoclimatic data, which motivated the development of new techniques

84 for analyzing geologic data and led to improvements in general circulation models.

85   More recently, the Palaeoclimate Modelling Intercomparison Project (PMIP) is

86 actively working to advance reconstruction of LGM and 6 ka climate through

87 model-to-model evaluations and data-model comparisons. PMIP has now entered the

88 third phase (PMIP3; Braconnot et al., 2012) and is a component of phase 5 of the Climate

89 Model Intercomparison Project (CMIP5). In contrast to CLIMAP, COHMAP, TEMPO

90 and earlier PMIP model experiments that employed fixed sea surface temperatures (SST)

91 and mixed-layer ocean models, some of the PMIP2 experiments and all of the PMIP3

92 experiments include fully coupled ocean and atmospheric models. Braconnot et al. (2012)

93 review some of the highlights of the PMIP2 experiments and the design of the PMIP3

94 experiments and Harrison et al. (2013) evaluate the PMIP3 and PMIP2 simulations of

95 LGM and 6 ka climates with data-model comparisons. In addition, continuous

96 simulations of climate over the last 21 ka have been achieved with earth system models

97 of intermediate complexity (e.g., Timm and Timmermann, 2007), and the TraCE-21ka

98 project at the National Center for Atmospheric Research (NCAR) conducted continuous,

99 transient climate simulations from 22 ka to 6.5 ka with the coupled NCAR Community

100 Climate System Model (Liu et al., 2009). Singarayer and Valdes (2010) simulated the

101 climate of the last 120,000 years using model snapshots at 4 ka and 1 ka intervals.

102   Here we explore past changes in late-Pleistocene climate using the coupled

103 Atmosphere-Ocean General Circulation Model (AOGCM) GENMOM. We simulated

104   multi-century time slices that span the interval from LGM to pre-industrial (PI) every

105   three thousand years (21, 18, 15, 12, 9, 6, 3 ka and PI). The simulations were run with

106   prescribed insolation, GHG concentrations, continental ice sheets, land extent and sea

107   level as boundary conditions. We analyze the within and between climatology of the time

108   slices and compare the 21 ka and 6 ka results with terrestrial and marine climate

109   reconstructions and results from the PMIP2 and PMIP3 simulations. The goal of our

110   simulations is to adopt a methodological framework similar to that of PMIP to simulate

111   time slices between the LGM and mid-Holocene. The simulations also serve as a base

112   line for applying GENMOM to more detailed and focused studies of late-Pleistocene

113   climate such as quantifying the effects of freshwater forcing and dynamic vegetation

114   feedbacks.

115   ## 2   Methods

116   ### 2.1   Model description

117   GENMOM combines version 3 of the GENESIS atmospheric model (Pollard and

118   Thompson, 1997; Thompson and Pollard, 1995; 1997) with version 2 of the Modular

119   Ocean Model (MOM2, Pacanowski, 1996). Version 3 of GENESIS (Alder et al., 2011;

120   Kump and Pollard, 2008; Pollard and Thompson, 1997; Zhou et al., 2008) incorporates

121   the NCAR CCM3 radiation code (Kiehl et al., 1998). GENESIS has been developed with

122   an emphasis on representing terrestrial physical and biophysical processes, and for

123   application to paleoclimate experiments. Earlier versions of GENESIS (Pollard and

124   Thompson, 1994; 1995; 1997; Thompson and Pollard, 1995; 1997) have been applied in

125   a wide range of modern and paleoclimate studies (Beckmann et al., 2005; Bice et al.,

126   2006; DeConto et al., 2007; 2008; Horton et al., 2007; Hostetler et al., 2006; Miller et al.,

127   2005; Poulsen et al., 2007a; 2007b; Ruddiman et al., 2005; Tabor et al., 2014), and

128   GENESIS simulations with fixed and slab ocean SSTs were included in PMIP1

129   (Joussaume et al., 1999; Pinot et al., 1999; Pollard et al., 1998).

130   In our simulations, we employ a coupled model with T31 spectral truncation,

131   which corresponds to a grid of 96 longitudes (3.75°) by 48 Gaussian latitudes (~3.71°).

132   The atmosphere is represented by 18 vertical sigma levels with mid-layers ranging from

133   0.993 at the surface to 0.005 at the tropopause. GENESIS includes the Land Surface

134   eXchange model, LSX, (Pollard and Thompson, 1995) to simulate surface processes and

135   to account for the exchange of energy, mass and momentum between the land surface and

136   the atmospheric boundary layer. MOM2 has 20 fixed-depth vertical levels and is

137   implemented on essentially the same T31 horizontal grid as GENESIS through cosine-

138   weighted distortion (Pacanowski, 1996). Sea ice is simulated by a three-layer model that

139   accounts for local melting, freezing, and fractional cover (Harvey, 1988; Semtner, 1976)

140   and includes the dynamics associated with wind and ocean current using the cavitating-

141   fluid model of Flato and Hibler (1992). The atmospheric and ocean models interact every

142   six hours without flux corrections.

143   GENMOM reproduces observed global circulation patterns, such as the seasonal

144   change in the position and strength of the jetstreams and the major semi-permanent sea

145   level pressure centers (Alder et al., 2011). The simulated present-day (PD) 2-m air

146   temperature climatology (Table 1) is 0.8 ºC colder than observations globally, 0.7 ºC

147   colder over oceans, and 0.9 ºC colder over land. Similar to other AOGCMs (e.g., Lee and

148   Wang, 2014), GENMOM produces a split ITCZ over the equatorial Pacific Ocean.

149   The pre-industrial Atlantic Meridional Overturning Circulation (AMOC)

150   simulated by GENMOM is 19.3 ± 1.4 Sv, which is stronger than, but comparable to, the

151   observed value of 17.4 Sv (Srokosz et al., 2012). Simulated SSTs display a warm bias in

152   some regions of the Southern Ocean, primarily south of 50ºS around Antarctica, and a

153   warm bias exceeding ~2 ºC between 200–1000 m depth in parts of the tropics and mid-

154   latitudes. Alder et al. (2011) note that the warm bias in the Southern Ocean is associated

155   with the relatively weak Antarctic Circumpolar Current (ACC) in GENMOM (62 Sv

156   versus the observed value of 119 Sv) and Deacon Cell upwelling which allows excessive

157   vertical mixing in the present-day GENMOM simulation, and together reduce sea ice

158   around Antarctica, particularly during summer. Both of these features are present to some

159   extent in our suite of simulations. We tested the Gent-McWilliams vertical ocean mixing

160   scheme (Gent and Mcwilliams, 1990) in GENMOM but it did not improve the Southern

161   Ocean warm bias, so we did not implement it in our paleosimulations.

162   The climate sensitivity of GENMOM for a doubling of CO_2 from present day is

163   2.2 ºC, which is in the lower range of other coupled AOGCMs (Meehl et al., 2007) and is

164   consistent with recent estimates of 2.7 ºC based on the PMIP3 LGM simulations

165   (Harrison et al., 2013) and paleodata-model estimates of 2.8 ºC (Annan and Hargreaves,

166   2013) and 2.3 ºC (Schmittner et al., 2011b).

167   Reflecting lower GHG concentrations, the average NH 2-m temperature in our PI

168   simulation is 0.79 °C cooler than our PD simulation, where as the PD simulation is 1.97

169   ºC cooler than observations and reflects the lower GHG concentrations specified in the PI

170   simulation (Tables 1 and 2). The PI-to-PD warming in the NH is similar to the observed

171   warming of ~0.6 – 0.9 °C (Brohan et al., 2006) and is in the range of response of other

172 climate models (e.g., Otto-Bliesner et al., 2006b). The greatest regional warming between

173 the in the PD simulation (not shown) is ~3 °C over the high northern latitudes and

174 northern polar regions during boreal autumn, winter and spring, consistent with the

175 observed polar amplification (Hassol, 2004).

176 **2.2 Experimental design**

177 We applied GENMOM to eight time periods for 21, 18, 15, 12, 9, 6, 3 ka and pre-

178 industrial. We prescribed insolation at the top of atmosphere for each time slice (Fig. 1)

179 by specifying appropriate orbital parameter values for precession, obliquity and

180 eccentricity (Table 1, Berger and Loutre, 1991). The solar constant was set to 1367 W m^{-2}

181 for all time periods. We estimated GHG concentrations from ice-core records by applying

182 a ±300 yr averaging window centered on the time period of interest (Table 1), and we

183 specified the PMIP3 GHG concentrations for our PI simulation (Braconnot et al., 2007a).

184 To derive continental ice sheets for the time slices, we used the ICE-4G

185 reconstructions (Peltier, 2002) for the Fennoscandian (FIS) and Cordilleran (CIS), and

186 the Oregon State University Laurentide Ice Sheet (OSU-LIS) reconstruction (Hostetler et

187 al., 1999; Licciardi et al., 1998) (Fig. 2). The ICE-6G reconstruction was not available for

188 our 8 time slices at time we began our simulations. However, the OSU-LIS

189 reconstruction has similar ice sheet topography to that of ICE-6G (Ullman et al., 2014)

190 and was available for our simulation periods. The combination of OSU-LIS and ICE-4G

191 enables us to use a more realistic LIS topography than that of ICE-5G, particularly over

192 the LIS during the deglacial, and facilitates adjusting sea level throughout our time slices.

193 A similar ice sheet configuration (OSU-LIS and ICE5-G) was used as a boundary

194 condition in the NASA GISS-E2-R LGM simulation submitted to the PMIP3 archive

195 (Ullman et al., 2014). We specified the 10 ka OSU-LIS ice sheet to ensure that Hudson

196 Bay remained covered by the LIS at 9 ka (Dyke and Prest, 1987). The ICE-4G

197 reconstruction includes an Eastern Siberian Ice Sheet, which we removed because it did

198 not exist (Felzer, 2001).

199   Topographic heights of the land masses were altered to reflect relative sea-level

200 change in ICE-4G. We created the topography and land mask for each time slice by

201 applying orographic changes to the present-day Scripps global orography data set (Gates

202 and Nelson, 1975). Orographic changes based on ICE-4G exposed or flooded land grid

203 cells associated with relative sea level (e.g. Indonesia, Papua New Guinea). We set the

204 ocean bathymetry to modern depths for ocean grid cells.

205   We specified the modern distribution of vegetation (Dorman and Sellers, 1989)

206 for all simulations because reconstructions of global vegetation for all time slices either

207 do not exist or are not well constrained. We note that while setting vegetation to modern

208 distribution for all simulations isolates the period-to-period climate response to other

209 boundary conditions, we do not capture dynamic vegetation-climate feedbacks that may

210 be important in some regions such as North Africa (Kutzbach et al., 1996b; Timm et al.,

211 2010) and the high latitudes of the NH (Claussen, 2009; Renssen et al., 2004). The

212 vegetation type on emergent land cells is set to be the same as neighboring existing land

213 cells. The simulations do not include varying dust forcing across the time slices, which

214 may account for up to 20% of the radiative change (Köhler et al., 2010; Rohling et al.,

215 2012). Freshwater flux from land-based precipitation is globally averaged and spread

216 over the world ocean (Alder et al., 2011).

217   In accordance with the PMIP3 protocol, to conserve atmospheric mass we

218   compensated for changes in global topography in each time slice by holding global

219   average surface pressure constant. At T31 resolution the Bering Strait and the Strait of

220   Gibraltar are closed in the default MOM2 bathymetry. We conducted sensitivity tests and

221   adjusted the bathymetry to ensure that key passages (e.g. Drake Passage, Norwegian Sea,

222   and Indonesian Throughflow) were adequately represented. Additional sensitivity testing

223   revealed that the modeled AMOC and salinity of the Arctic are very sensitive to the

224   bathymetry of the Norwegian Sea, particularly to the width of the passage between

225   Scandinavia and Greenland as it narrowed by the growth of the FIS. We removed Iceland

226   from the model to ensure that the passage remained sufficiently wide and deep to prevent

227   unrealistic buildup of salinity in the Arctic.

228   Each time slice simulation was initialized from a cold start (isothermal

229   atmosphere, latitudinally dependent ocean temperature profile, and uniform salinity of 35

230   ppt) and run for 1,100 years. We exclude the first 1,000 years from our analysis here to

231   allow for spin up of ocean temperatures. The temperature drift in the last 300 years of our

232   simulations (SFig. 1) is acceptably small (Braconnot et al., 2007a; Singarayer and Valdes,

233   2010) with values of -0.05 ºC/century for the LGM, 0.01 ºC/century for 6 ka, and -0.02

234   ºC/century for PI. Drift in the LGM and early deglacial simulation is attributed primarily

235   to long-term cooling and the evolution of sea ice in the southern ocean. Simulated

236   AMOC exhibits decadal scale variability, but was free of drift over the last 300 years of

237   the simulations.

238   In what follows, the monthly averages of the model output are based on the

239   modern calendar as opposed to the angular calendar that changes with Earth-sun

240 geometry (Pollard and Reusch, 2002; Timm et al., 2008). The modern calendar is

241 commonly used in data-model comparisons (e.g., Harrison, 2013).

3 Results

3.1 Atmospheric circulation

244   The boreal winter (DFJ) 500 hPa heights in the PI simulation (Fig. 3a) display the

245 observed high- and mid-latitude ridge-trough-ridge-trough standing wave structure (wave

246 number two) that arises from continent-ocean-continent-ocean geography of the NH

247 (Peixoto and Oort, 1992). From 21 ka to 9 ka the LIS, the FIS and Greenland ice sheets

248 alter the NH standing wave structure resulting in persistent, distinct troughs and cyclonic

249 flow tendencies over northeast Asia, the North Pacific, the continental interior of North

250 America, the North Atlantic and Europe (Fig. 3, maps of raw fields SFig. 2).

251   Consistent with previous LGM studies using comparable (Braconnot et al., 2007a)

252 and higher-resolution (Kim et al., 2007; Unterman et al., 2011) climate models, from 21

253 ka through 9 ka the western edge of the Cordilleran Ice Sheet diverts the LGM winter

254 polar jetstream resulting in one branch that is weaker than PI over the Gulf of Alaska and

255 the western and central regions of the ice sheet, and a second branch to the south of the

256 ice sheet that is stronger than the PI (Fig. 3a). The reorganization creates westward wind

257 anomalies over the North American Pacific Northwest. The LIS effectively guides the

258 convergence of the branches, and the meridional gradient of low and high 500 hPa height

259 anomalies in the North Atlantic intensifies flow over North America, the North Atlantic,

260 Europe and Northern Africa (Figs. 3a) thereby altering the path of storm tracks. This flow

261 pattern weakens progressively as the LIS recedes.

262   The influence of the NH ice sheets is also evident in summer (JJA), but to a lesser

263   degree than in winter (Fig. 3b) due to continental heating and the absence of the strong,

264   mid-latitude storm tracks. Between 21 ka and 15 ka, the summer jetstream is constrained

265   and therefore enhanced along and to the south of the southern margin of the LIS

266   extending over the North Atlantic. At 18 ka, a trend toward positive JJA anomalies in 500

267   hPa heights emerges over the regions of the semi-permanent subtropical high pressure of

268   the North Pacific and central Atlantic. The regions of positive height anomalies, and their

269   associated anticyclonic wind anomalies, expand over central North America, peak from

270   12 ka through 9 ka, and diminish by 6 ka (Fig. 3). The DJF pattern of low-to-high height

271   anomalies over the North Atlantic is replaced during JJA by a strengthened subtropical

272   high. Anticyclonic flow around positive height anomalies on the western edge of the FIS

273   alters regional flow patterns over and south of the ice sheet. The GENMOM responses to

274   the NH ice sheets are similar to many previous modeling experiments that have

275   established that changes in tropospheric pressure–surface heights and winds are primarily

276   driven by changes in ice-sheet height, and secondarily by temperature and albedo

277   feedbacks (COHMAP Members, 1988; Felzer et al., 1996; 1998; Otto-Bliesner et al.,

278   2006a; Pausata et al., 2011; Pollard and Thompson, 1997; Rind, 1987).

279   From 21 ka to 12 ka, the largest changes in boreal winter sea-level pressure (SLP)

280   are associated with negative surface temperature anomalies over the continental ice

281   sheets, the landmasses of the NH, and areas of expanded sea ice in the North Atlantic

282   (Fig. 4a) where cooling increases subsidence and thus contributes to cold high surface

283   pressure. From 21 ka to 15 ka, high pressure over the LIS produces anticyclonic flow

284   across the northern Great Plains and over the Puget Lowlands of the US. Similar

285 anticyclonic tendencies are simulated along the margin of the FIS. Between 12 ka and 6

286 ka the winter SLP around the Aleutian low in the North Pacific and the Icelandic low in

287 the North Atlantic is strengthened relative to PI. The Aleutian low is expanded southward

288 whereas the Icelandic low is confined on the northern edge by the FIS and is slightly

289 displaced southeastward.

290   From 21 ka to 9 ka, the JJA SLP anomalies remain strongly positive over the ice

291 sheets and sea ice, whereas from 12 ka to 6 ka the SLP anomalies over Northern

292 Hemisphere landmasses are negative due to enhanced continental warming (Fig. 4b). The

293 patterns of the JJA 500 hPa heights, SLP and the associated circulation over North

294 America and adjacent oceans again illustrate similar responses to time-varying controls:

295 changes from 21 ka to 15 ka are primarily driven by changes in the LIS, whereas from 12

296 ka to 6 ka the circulation changes are related to the changes in the seasonality of

297 Holocene NH insolation (Fig. 2).

298 **3.2   Near-surface air temperature**

299   Our time slice simulations clearly display surface air temperature (SAT) changes

300 attributed to radiative forcing from the presence of the continental ice sheets, GHGs

301   (Clark et al., 2012), and insolation (Fig. 5). The global average mean annual LGM

302 temperature simulated by GENMOM is 3.8 °C colder than the PI (Table 2, Fig. 5a),

303 within the range of cooling in the PMIP2 AOGCM simulations (3.1 °C to 5.6 °C and

304 average 4.4 °C) and the PMIP3 simulations (2.6 °C to 5.0 °C and average of 4.4 °C) that

305 were forced by similar boundary conditions (Harrison et al., 2013; Kageyama et al.,

306 2006). Our LGM cooling is also in agreement with Annan and Hargreaves (2013), who

307 reconciled the PMIP2 ensemble and proxy data to derive an estimated cooling of 4.0 ±

308   0.8 °C, but falls outside the range of Schmittner et al. (2011b) who found a median

309   cooling of 3.0 ºC (66% probability range of 2.1 ºC - 3.3 ºC). GENMOM is also consistent

310   with three transient simulations (Liu et al., 2014) averaged over the periods simulated by

311   GENMOM. Excluding the BA and YD, our simulations reproduce the rate of warming

312   between 21 ka and 15 ka, but are consistently ~1 °C colder than the reconstruction of

313   Shakun et al. (2012) when sampled at the proxy sites (Fig. 5b). During these periods,

314   GENMOM falls at the low end or outside the range of the transient models; however,

315   GENMOM falls within the range of LGM and MH cooling simulated by the PMIP3

316   models, which have similar experimental designs and large scale boundary conditions.

317   Neither GENMOM nor the ensemble mean of the PMIP3 models capture the

318   ~0.5 °C the 6 ka temperature anomaly in the Marcott et al. (2013) reconstruction. The

319   change in the 6 ka mean annual temperature at the proxy sites in the 12 PMIP3 models

320   we analyzed ranged from -0.3 to 0.3 °C with a mean of ~0.0 °C. Three models simulated

321   slight warming, five near zero and four simulated slight cooling. Whether or not some

322   proxies used in the temperature reconstructions have seasonal bias which would

323   exaggerate the mid-Holocene warming remains an open research question (Liu et al.,

324   2014).

325   Seasonal temperature changes across our time slice simulations illustrate the

326   spatial and temporal effect of changing boundary conditions (Fig. 6). From 21 ka through

327   15 ka, both DJF and JJA exhibit cold temperature anomalies exceeding 16 °C over and

328   adjacent to the ice sheets in both hemispheres. With the exception of Europe and the high

329   latitudes of the NH, boreal winters remain generally colder than PI over the continents

330   until 3 ka (Fig. 6), corresponding to reduced insolation. NH atmospheric circulation

331 changes induced by atmospheric blocking from the LIS (Fig. 3) sustain positive winter

332 and summer temperature anomalies over Beringia. Summer warming also occurs south of

333 the FIS across much of Asia. Although the mid-Holocene wintertime deficit in insolation

334 is small at high northern latitudes, changes in short-wave radiation at the surface during

335 boreal summer in the model are large and positive (30 – 40 Wm^{-2}) due to the precessional

336 shift of perihelion and changes in obliquity (SFig. 4). Substantial warming occurs

337 between most pairs of consecutive time slices from the LGM through the Holocene (Fig.

338 7, Table 2); however, over the African and Indian monsoon regions increased cloudiness

339 associated with enhanced summer monsoonal precipitation leads to cooling from 15 to 6

340 ka.

341   The relatively high rate of warming between 18 ka and 15 ka (1.5 °C land and

342 0.5 °C ocean, Fig. 7, Table 2) is commensurate with increased GHGs (Table 1). Periods

343 of peak annual warming from 15 ka to 12 ka (2.4 °C land and 0.7 °C ocean) and from 12

344 ka to 9 ka (1.6 °C land and 0.2 °C ocean) are associated with increasing GHG

345 concentrations, ablation of the NH ice sheets (Figs. 1 and 6a). The simulated rates of

346 annual global warming between the LGM and the early Holocene (Fig. 5) are in

347 agreement with data (Clark et al., 2012; Gasse, 2000), and the analyses by Shakun et al.

348 (2012) and Marcott et al. (2013) who attribute a large component of the warming to rising

349 GHG levels.

350   The DJF and JJA temperature differences in our 21 ka simulation are similar to

351 those of the PMIP3, allowing for differences in between our prescribed NH ice sheets

352 (ICE-4G+OSU-LIS in GENMOM) and the blended ice sheet of the PMIP3 simulations

353 that essentially combines the height of the ICE6G reconstruction with the extent of the

354   Dyke and Prest (1987) reconstructions (SFigs. 5 - 10, Braconnot et al., 2012). In both

355   seasons, GENMOM produces 0.5 - 1 °C less cooling in the tropical oceans and greater

356   warming over Beringia. The positive JJA temperature anomaly south of the FIS in

357   GENMOM persists through 15 ka. Summer warming in the presence of the ice sheet was

358   identified in earlier versions of GENESIS (Pollard and Thompson, 1997) and is

359   associated with subsidence over the ice (Rind, 1987). Similar JJA warming also occurs in

360   some of the PMIP3 models, but is likely a model artifact (Pollard and Thompson, 1997;

361   Ramstein and Joussaume, 1995; Rind, 1987).

362   The DJF and JJA temperature anomalies in our 6 ka simulation are also similar to

363   those of the PMIP3 models (SFigs. 7 and 8). Relative to PI, GENMOM produces slightly

364   greater winter warming over Scandinavia than is evident in the average of the PMIP3

365   simulations, and is generally 0.5 - 1.0 °C cooler over Asia, Africa and South America.

366   During boreal summer, GENMOM simulates warming over the NH landmasses and

367   cooling over the North African and Indian monsoon regions, consistent with the PMIP3

368   models. Continental warming in GENMOM is ~ 0.5 - 1.0 °C weaker than most PMIP3

369   models, particularly in Europe and Asia. A portion of the weaker warming in GENMOM

370   is attributed to the prescribed 6 ka GHG concentrations we derived from the ice-core data

371   that differ slightly from those specified for the PMIP3 experiments (Table 1 caption).

372   **3.3   Precipitation and monsoons**

373   The simulated global precipitation anomalies display a progression from the drier

374   and colder conditions of the LGM to the warmer and wetter conditions of the Holocene

375   (Fig. 8, Table 2). The global mean annual precipitation change of -0.29 mm d^{-1} for the

376   LGM is distributed as greater drying over land and ice sheets (-0.30 mm d^{-1}) than oceans

377 (-0.22 mm d^{-1}). Regionally coherent patterns of precipitation change (Figs. 8 and 9) are

378 indicative of displacement and changes in the strength of storm tracks (Li and Battisti,

379 2008), the ITCZ and the Hadley circulation, and the onset, amplification and subsequent

380 weakening of the global monsoons regions (Broccoli et al., 2006; Chiang, 2009; Chiang

381 and Bitz, 2005).

382 Between the LGM and 15 ka, during DJF areas over and adjacent to the NH ice

383 sheets display predominately reduced precipitation arising from a combination of the

384 desertification-effect of the high and cold ice, lower-than-present atmospheric moisture

385 and cloudiness and the advection of cold, dry air off of the ice sheets (Figs. 3a, 4a, 6a and

386 8a). The topographic and thermal effects of the LIS and the thermal effect of sea ice

387 (Kageyama et al., 1999; Li and Battisti, 2008) alter 500 hPa geopotential heights

388 along the southern margin of the ice sheet (Figs. 3a and SFig. 2a), causing the

389 development of positive precipitation anomalies extending from the eastern Pacific across

390 the Gulf of Mexico, eastern North America and into the Northern Atlantic.

391 Accompanying negative precipitation anomalies over the North Atlantic and positive

392 anomalies over the Nordic Seas are related to changes in the location of storm tracks. The

393 local effect of the ice sheets on precipitation diminishes during the early and mid-

394 Holocene as their influence on circulation weakens and the atmosphere becomes warmer

395 and moister (Fig. 9a).

396 The negative DJF anomalies that persist from 21 ka to 15 ka during austral

397 summer along the equatorial and low-latitude areas of South and Central America, south-

398 central Africa Southeast Asia, Northern Australia, the tropical Atlantic, the Indian Ocean

399 and the western Pacific warm pool are caused by changes in the location of the ITCZ and

400 weakened southern monsoonal circulation. This particularly affects the winter monsoon

401 in central South America (Cheng et al., 2012; Zhao and Harrison, 2012) and in Southeast

402 Asia and Indonesia where additional feedbacks in the energy and water balances over

403 emergent land areas occur during low sea level stands (Figs. 1 and 8a) have been shown

404 to alter the Walker Circulation (DiNezio and Tierney, 2013).

405   Precipitation for JJA also exhibits considerable change over time (Figs. 8b and

406 9b). Similar to DJF, generally drier conditions are simulated over and adjacent to the NH

407 ice sheets where anticyclonic flow tendencies suppress precipitation (Fig. 4b). Along

408 portions of the southern margins of the LIS and FIS, however, orographic lifting

409 enhances precipitation at 21 ka (Pollard and Thompson, 1997). Wetter conditions in the

410 North American Southwest derive from enhanced westerly flow aloft and lower level

411 southwesterly flow off the eastern Pacific that are associated with displacement of the

412 jetstream by the ice sheets and the weakened Pacific subtropical high. Between 21 ka and

413 12 ka the LIS causes an increased pressure gradient from a strengthened Azores-Bermuda

414 high and weakened subtropical high in the eastern Pacific (Figs. 3b and 4b), resulting in

415 amplified and displaced westward winds, drying over Central America, and wetter-than-

416 present conditions over northern South America. At the LGM, North Africa, Europe, and

417 all but the western edge of Asia, are drier than the PI, again reflecting the drier

418 atmosphere of the full glacial.

419   The magnitude, gradients and spatial patterns of GENMOM 21 ka DJF

420 precipitation anomalies are consistent with the PMIP3 experiments. Notable exceptions

421 are greater drying than some models in the North Atlantic and the band of positive

422 anomalies extending across the Gulf of Mexico and the southeast US. GENMOM

423   produces positive precipitation anomalies over Australia, which is present in four of the

424   PMIP3 models. The 21 ka JJA precipitation anomalies are also in agreement with PMIP3,

425   but display weaker drying over eastern NA and slight drying over the North Africa

426   monsoon region.

427   The time evolution from LGM to PI of the African and Indian monsoons reflects

428   the interplay of changes in the location of the ITCZ and Hadley circulation that are linked

429   to the receding NH ice sheets, GHG-driven global warming, enhanced NH JJA insolation

430   and changing land-SST temperature contrast. The North Africa and Indian monsoons are

431   suppressed between 21 ka and 18 ka. After 18 ka, wetter-than-present conditions emerge

432   in the monsoon regions of North Africa and India where increased JJA insolation warms

433   the continents which amplifies the land-sea temperature contrasts that drive monsoonal

434   circulation (Braconnot et al., 2007b; Kutzbach and Otto-Bliesner, 1982; Zhao and

435   Harrison, 2012). The simulated DJF air temperatures in North Africa cool from the LGM

436   until 15 ka, and then warm monotonically through the rest of the deglaciation and

437   Holocene (Fig. 10). Wintertime precipitation over the North African region is minimal. In

438   contrast, JJA temperatures increase throughout the deglaciation, peak at 9 ka, decrease

439   slightly at 6 ka, and increase thereafter. A commensurate increase in JJA precipitation

440   over North Africa between 12 ka and 6 ka is associated with northward migration of the

441   ITCZ (Braconnot et al., 2007a; 2007b; Kutzbach and Liu, 1997), which enhances the

442   transport of moisture into both the North African and Indian monsoon regions.

443   Monsoonal precipitation peaks over both regions between 12 ka and 9 ka (Fig. 10). The

444   change in precipitation between 9 ka and 6 ka over India (0.9 mm d^{-1}) is nearly double

445   the change over North Africa (0.5 mm d^{-1}), consistent with the diagnoses of the mid-

446 Holocene monsoon of Marzin and Braconnot (2009) who attribute the stronger ~9 ka

447 monsoon to insolation related to precession and snow cover on the Tibetan Plateau. The

448 pattern of precipitation in the Indian monsoon region is similar to that of North Africa,

449 but exhibits a greater range between peak Holocene values and the PI.

450   The overall temporal progression and magnitude of precipitation changes in the

451 time slice simulations are in agreement with the PMIP2 (Braconnot et al., 2007a; 2007b)

452 and PMIP3 simulations at 21 and 6 ka, and with other mid-Holocene modeling studies

453   (Hély et al., 2009; Kutzbach and Liu, 1997; Kutzbach and Otto-Bliesner, 1982;

454 Timm et al., 2010). More specifically, the June through September GENMOM

455 precipitation anomaly of -0.6 mm d^{-1} over the North Africa monsoon region during the

456 LGM is within the range (-0.9 to 0.1 mm d^{-1}) of 5 PMIP2 AOGCMs (Braconnot et al.,

457 2007a) and 7 PMIP3 models (range of -0.6 to 0.2 and average of -0.2 mm d^{-1}). The

458 GENMOM LGM anomaly over India (-0.9 mm d^{-1}) is also within the range (-1.7 to -0.1

459 mm d^{-1}) of the PMIP2 simulations (Braconnot et al., 2007a) and the PMIP3 simulations

460 (range of -1.3 to 0.0 and average of -0.7 mm d^{-1}).

461   The northward expansion and spatial pattern of precipitation anomalies of the 6 ka

462 monsoons are in very good agreement with both the PMIP2 and PMIP3 experiments.

463 Summer precipitation in the GENMOM simulation is enhanced by 0.9 mm d^{-1} relative to

464 PI over North Africa, in agreement with the range (0.2 to 1.4 mm d^{-1}) and mean

465 (0.7 mm d^{-1}) of 11 PMIP2 AOGCMs (Zhao and Harrison, 2012) and 12 PMIP3 models

466 (range of 0.1 to 1.0 and average of 0.6 mm d^{-1}). Over India, the 6 ka GENMOM

467 precipitation anomaly of 1.1 mm d^{-1} exceeds the range (0.2 to 0.9 mm d^{-1}) and mean (0.6

468   mm d^{-1}) of the 11 PMIP2 models (Zhao and Harrison, 2012), but is within the range of

469   the PMIP3 models (0.5 to 1.3 and average of 1.0 mm d^{-1}) .

470   **3.4   Sea ice**

471   DJF sea ice is present in the PI simulation over Hudson Bay, the Arctic Ocean,

472   along the coast of eastern Canada, around Greenland, the Nordic Seas and the Baltic and

473   North Sea (Fig. 11), in agreement with observed present-day distributions (Jaccard et al.,

474   2005). Ice fractions of up to 100% are simulated over the Bering Sea and the Sea of

475   Okhotsk. In the SH, sea ice persists through austral summer in the Weddell and Ross

476   Seas and a few scattered locations around Antarctica. While the locations of the ice

477   around Antarctica are in agreement with observations (Gersonde et al., 2005), the model

478   underestimates the ice extent over the Weddell Sea and between the Weddell and Ross

479   Seas. The lack of ice is partly attributable to a warm bias in the Southern Ocean

480   associated with the previously mentioned weak ACC (discussed further below). During

481   August and September, simulated sea ice is greatly reduced in the North Atlantic region

482   (Fig. 11), with remnant ice persisting in the extreme north of Baffin Bay and the east

483   coast of Greenland, also in agreement with observations. In the SH, the corresponding

484   winter sea ice grows substantially and the distribution is in generally good agreement

485   with observations (Gersonde et al., 2005).

486   The simulated annual average ice extents for the NH are 9.8×10^6 km^2 for the

487   LGM, 15.8×10^6 km^2 for 6 ka and 14.1×10^6 km^2 for PI (grid cells with fractional coverage

488   > 15%). Compounded with climate-forcing, changes in both the distribution and areal

489   coverage of the NH ice also reflect the change in ocean area due to the transition of land

490   and ice sheets to ocean as sea level rises (Fig. 11 and SFigs. 13 - 15). For the same time

491   periods, the SH ice area extents, which are minimally affected by land-sea transitions with

492   sea level rise, are 20.9×10^6 km^2, 11.4×10^6 km^2 and 11.1×10^6 km^2, respectively.

493   During the 21 ka boreal winter, the Arctic Ocean and Baffin Bay are fully covered

494   by ice and the ice around Greenland expands. The model displays increased sea ice in the

495   western North Atlantic and decreased ice in the eastern North Atlantic and Nordic Seas

496   where the prescribed FIS margin advances into the water (Fig. 2). The limit of substantial

497   coverage north of 55°N is in agreement with reconstructions (de Vernal et al., 2006) and

498   other LGM simulations (Otto-Bliesner et al., 2006a; Roche et al., 2007); however, slight

499   fractional cover (pack ice) in the model likely extends too far south (to ~45°N) along the

500   coast of North America. Fractional cover of up to 100% is simulated in the far Northwest

501   Pacific and the Sea of Okhotsk with a sharp, southward transition to reduced coverage. In

502   boreal summer of the LGM, simulated sea ice retreats to 65°N in the North Atlantic and

503   persists along eastern Canada, Baffin Bay and south of Greenland and the extreme

504   northern areas of the Nordic Seas.

505   The overall distribution of SH sea ice (Fig. 11) is in good agreement with

506   reconstructions and other model simulations (Gersonde et al., 2005; Roche et al., 2012).

507   The simulated LGM maximum winter sea ice area is 35.5×10^6 km^2 (72% greater than PI)

508   and the LGM summer minimum is 4.8×10^6 km^2 (112% greater than PI); the winter and

509   summer reconstructed areas are $43.5 \pm 4 \times 10^6$ km^2 and $11.1 \pm 4 \times 10^6$ km^2, respectively

510   (Roche et al., 2012). The seasonal amplitude (maximum minus minimum) of LGM ice

511   cover simulated by GENMOM (30.6×10^6 km^2) is comparable with the reconstructed

512   amplitude ($32.4 \pm 4 \times 10^6$ km^2) and the LGM-to-PI change of seasonality is well within the

513   range simulated by the PMIP2 models (Roche et al., 2012 their Figures 2 and 3).

514 **3.5 Antarctic Circumpolar Current and Atlantic Meridional Overturning**
515 **Circulation**

516   The simulated ACC of 62 Sv is ~48% weaker than the observed value of 119 Sv

517 through the Drake Passage (GECCO data; Köhl and Stammer, 2008). Although the T31

518 resolution of GENMOM is a factor in limiting flow through the Drake Passage, we

519 attribute the underestimate of the ACC in part to insufficient wind stress at the latitude of

520 the Drake Passage, which is caused by equatorward displacement of the core of the

521 westerly winds, a shortcoming in common with other low-resolution AOGCMs (Alder et

522 al., 2011; Russell et al., 2006; Schmittner et al., 2011a).

523   Considerable uncertainty exists in the proxies that are used to infer past changes

524 in AMOC strength (Delworth and Zeng, 2008; Lynch-Stieglitz et al., 2007). The

525 ^{231}Pa/^{230}Th record from 33°N on the Bermuda Rise (Lippold et al., 2009; McManus et al.,

526 2004) indicates that after the LGM the strength of the AMOC began to diminish at

527 ~18 ka, was further reduced during Heinrich Event 1 (H1) at ~17 ka, increased abruptly

528 during the BA at 15 ka, and weakened again during the YD cold reversal at ~12 ka. After

529 the YD, the AMOC strengthened again and stabilized. In climate models, a variety of

530 factors including the North Atlantic freshwater budget, model resolution and

531 parameterizations and the characteristics of simulated Antarctic Bottom Water (AABW)

532 give rise to a considerable simulated range of AMOC (Weber et al., 2007).

533   The AMOC in our PI simulation (Fig. 12) is 19.3 ± 1.4 Sv at the core site of

534 33°N, a value similar to the present-day estimate of 18.7 ± 4.8 Sv at 26.5°N (Srokosz et

535 al., 2012). The maximum AMOC simulated by GENMOM in the PI is 21.3 Sv at 41°N, a

536 value outside the range of 13.8 to 20.8 Sv of five models in the PMIP2 experiments

537   (Weber et al., 2007), but within the range of 3.8 to 31.7 Sv of the IPCC AR4

538   models (Meehl et al., 2007; Schmittner et al., 2005). The newer CMIP5 models have a

539   narrower range of AMOC of ~14 to ~30 Sv when sampled at 30°N (Cheng et al., 2013);

540   GENMOM simulates 16.0 ± 1.3 Sv at this location.

541   Our simulated LGM AMOC at the core site is 16.4 Sv, which is a ~14.7%

542   reduction relative to the PI. The maximum LGM AMOC is 22.4 Sv at 40.8°N, an

543   increase of 1.1 Sv (5.1%) relative to the PI maximum and within the considerable range

544   of -6.2 to +7.3 Sv in five PMIP2 simulations (Weber et al., 2007). In the deglacial

545   simulations (21 ka through 15 ka), the northward (positive) AMOC flow extends deeper

546   than that of the PI (Fig. 12) and the southward flow or AABW consequently is somewhat

547   weakened. The maximum AMOC in GENMOM is essentially constant at 40.8°N depth

548   of 1.23 km for all time slices. Although the depth of the maximum is again comparable to

549   the range of the PMIP2 models (1.24 ± 0.20), the invariance of the location and depth in

550   GENMOM is likely a model-specific response.

551   Our time slice simulations display an increase in the strength of AMOC from the

552   LGM to a maximum at 15 ka, decrease to a minimum at 9 ka, and remain more-or-less

553   constant through the PI (Fig. 13), which is in apparent disagreement with the ^{231}Pa/^{230}Th

554   records from which greater variability is inferred (Lippold et al., 2009; McManus et al.,

555   2004). We do not expect to capture rapid and abrupt climate change events such as H1

556   (~17 ka), the BA (~15 ka) and the YD (~12 ka) with only eight time slices, because we

557   did not manipulate freshwater discharge to the North Atlantic in our experimental design.

558   **4   21 ka and 6 ka data-model comparisons**

559   We compare temperature and precipitation from our LGM and mid-Holocene

560   simulations with paleoclimatic reconstructions and the PMIP3 simulations. For the LGM,

561 we use the pollen-based reconstructions of mean annual mean temperature (MAT) and

562 precipitation (MAP) from Bartlein et al. (2011) over land, and the Multiproxy Approach

563 for the Reconstruction of the Glacial Ocean Surface Project (MARGO) reconstructions

564 over oceans (Waelbroeck et al., 2009). The gridded 2° x 2° pollen data include >3,000

565 terrestrial pollen records from Eurasia, Africa and North America, and the global

566 MARGO reconstruction comprises ~700 analyses of planktonic foraminifera, diatom,

567 dinoflagellate cyst and radiolarian abundances, alkenones, and planktonic foraminifera

568 Mg/Ca from marine core sites. For 6 ka, we combine the pollen-based reconstructions of

569 Bartlein et al. (2011) and the GHOST SST reconstructions (Leduc et al., 2010). The 6 ka

570 GHOST data set contains ~100 reconstructed temperature records based on analyses of

571 alkenones and foraminifera Mg/Ca from marine sites located along continental margins

572 and the Mediterranean Sea.

573 **4.1 21 ka**

574   Our simulated 21 ka anomalies of MAT and MAP are comparable with the pollen

575 reconstructions (Fig. 14) and fall within the range of the PMIP3 models. GENMOM

576 captures the mixed pattern of temperature and precipitation anomalies over Beringia that

577 are present in the reconstructions (Fig. 14a,b) and in several of the PMIP3 simulations

578 (SFigs. 8, 9, and 16). The GENMOM SST anomalies indicate broad cooling of the global

579 oceans (mean of -1.7 °C) but not as much cooling as is simulated in the PMIP3 models

580 (mean of -2.9 °C); although, Harrison et al. (2013) found that the PMIP3 models tended

581 to overestimate oceanic cooling. Sampled at the MARGO locations, GENMOM is

582 generally warmer, but within the range of the PMIP3 models (Harrison et al., 2013). The

583 overall agreement of the simulation with the MARGO data is good, but some features in

584   the MARGO data are not reproduced by GENMOM. For example, similar to the PMIP3

585   simulations (SFigs. 5, 6 and 16) the GENMOM simulation lacks the warming over the

586   Greenland and Nordic Seas inferred from the data; although, while the data indicate the

587   Nordic Sea was ice free at the LGM, the magnitude of the warming elsewhere, if it

588   occurred, is somewhat unclear (de Vernal et al., 2006; Moller et al., 2013). The limited

589   cooling along the western coast of North America and Mediterranean in GENMOM is

590   attributed to the inability of the model to resolve the California Current and the

591   Mediterranean circulation (Alder et al., 2011).

592   Over the tropical ocean basins, the 21 ka GENMOM simulation is 1.6 °C colder

593   than the PI, in good agreement with the inferred MARGO cooling of 1.7 ± 1 °C (Otto-

594   Bliesner et al., 2009). Average simulated SST anomalies are also similar to MARGO

595   over the Indian (-1.6 °C versus -1.4 ± 0.7 °C) and Pacific (-1.5 °C versus -1.2 ± 1.1 °C)

596   Oceans, but are warmer than the data in the tropical Atlantic basin (-1.9 °C versus -2.9 ±

597   1.3 °C). In each of these regions, the anomalies simulated by GENMOM fall within the

598   range of six PMIP2 models analyzed by Otto-Bliesner et al. (2009). GENMOM captures

599   the 2 – 4 °C cooling in the eastern coastal Atlantic evident in the MARGO data, and the

600   SST anomalies are ~2 – 4 °C colder over the Western Pacific Warm Pool. Neither

601   GENMOM nor the PMIP3 simulations produce the warming over the central and eastern

602   tropics, or the low latitudes and the North Atlantic that is evident in the MARGO

603   reconstruction.

604   The simulated LGM MAP anomalies are also comparable with the pollen-based

605   reconstructions (Fig. 14c and d). The model simulates general drying of the NH and a

606   mix of increased and decreased precipitation in Beringia, South America, southern

607 Africa, Southeast (SE) Asia and Australia. GENMOM produces strong drying over and

608 around the NH ice sheets, wetter-than-present conditions in the southwestern United

609 States and drying in Central America. The simulation fails to reproduce the drying over

610 eastern North America that is inferred from the pollen-based data. There is considerable

611 variability in the PMIP3 simulations of MAP (SFigs. 9 and 10). In common with the

612 PMIP3 models, GENMOM simulates a general reduction of precipitation over the NH,

613 the North African and Indian monsoon regions, and SE Asia, and increased precipitation

614 south of the LIS, southern Africa and much of Australia (SFig. 16).

615 **4.2   6 ka**

616 Relative to PI, the changes in 6 ka boundary conditions are predominantly in the

617 seasonality of insolation (Table 1) as opposed to the stronger radiative forcing associated

618 with changes in GHGs and continental ice sheets from the LGM through the early

619 Holocene. The resulting changes in 6 ka climatology are thus more subtle than those of

620 the deglaciation. The changes of 6 ka MAT simulated by GENMOM are generally within

621 the range of ±1 °C (Fig. 15b). Enhanced MAP and associated cooling are evident in the

622 NH monsoonal regions (Fig. 15d). Elsewhere, MAP changes are within a range of

623 ±50 mm.

624 Pollen-based data reconstructions indicate highly heterogeneous changes in MAT

625 during 6 ka; however, there are regions with spatially consistent changes in sign, such as

626 warming south of Hudson Bay, areas of warming over Scandinavia and Western Europe,

627 and cooling in the Mediterranean region (Fig. 15a). Larger MAT changes at high-

628 elevation sites and regions with anomalies of mixed sign occur in the data over most

629 continents. The GENMOM 6 ka MAT anomalies also display a mix of warming and

630   cooling in a range of about ±4 °C; however, where pollen-based records exist, the

631   majority of the anomalies are within a narrower range of about ±1.5 °C (Fig. 15b).

632   GENMOM, and many of the PMIP3 models (SFigs. 8, 9 and 16), produce a mixture of

633   warm and cold 6 ka MAT anomalies that are generally in the range of ±1 °C over the

634   North Atlantic, Europe and Scandinavia, which underestimates the proxy-based

635   anomalies by >2 °C at some sites.

636   The Asian pollen-based reconstruction similarly displays a heterogeneous

637   temperature pattern that is reproduced by GENMOM and the PMIP3 models. In all of the

638   models, the sign of the anomalies does not vary abruptly in close proximity to the pollen

639   sites. We note, however, that the smooth topography in GCMs limits the ability of the

640   models to reproduce large and regionally spatially heterogeneous anomalies that are

641   characteristic of the local climate at many high elevation pollen sites in Western North

642   America, the Alps, the central plateau of African and Asia.

643   GENMOM displays cooling in the North African and Indian monsoon regions

644   and warming over the high northern latitudes, consistent with the PMIP3 models (Fig.

645   15). In contrast, GENMOM simulates weak global cooling of 0.39 °C compared to no

646   change in the PMIP3 model average which is partially attributed to our lower prescribed

647   GHG concentrations (Table 1 caption).

648   Precipitation anomalies inferred from the pollen-based data indicate that 6 ka was

649   wetter than the PI in Europe, Africa, Asia and some parts of western North America and

650   drier than PI in much of eastern North America and Scandinavia (Fig. 15c). GENMOM

651   simulates the gradients and coherent patterns of positive and negative MAP anomalies

652   over North America, and North, Central and western Africa, in agreement with the data

653 and the PMIP3 models. The data and GENMOM are also in agreement over the Asian

654 monsoon region and northwest Asia where wetter conditions prevail, but anomalies of

655 opposite sign are simulated over the Great Lowland Plain in north central Eurasia and

656 Southeast Asia. Bartlein et al. (2011) attribute cooling in Southeast Asia to a stronger

657 winter monsoon at 6 ka. Our results (Figs. 6a and 8a), and many of the PMIP3 models,

658 indicate cooler, drier winters (SFigs. 7 and 11) and regionally variable changes in the

659 summer (SFigs. 8 and 12).

660 In Africa, the model captures the increase in precipitation in the northern and

661 continental regions and drying along the southern coastal regions, as evident in the data.

662 Strengthening of the African and Indian summer monsoons during the mid-Holocene

663 corresponds well with the PMIP2 and PMIP3 models (Zheng and Braconnot, 2013). Both

664 GENMOM and the data indicate drying over central Scandinavia, wetter conditions over

665 east central Europe, the Iberian Peninsula and around the Mediterranean but, over

666 Western Europe, the simulated decrease in MAP in GENMOM clearly disagrees with the

667 data and some of the PMIP3 models (Figs. 15, SFigs. 7, 8 and 16); although, the

668 magnitude of the change in the models is very small and the sign of the change varies

669 among models. Wetter conditions also prevail in Indonesia, and a southwest-to-northeast

670 wet-dry gradient is simulated over Australia.

671 **5 Summary**

672 We have presented a suite of multi-century equilibrium climate simulations with

673 GENMOM for the past 21,000 years at 3,000-yr intervals. Each 1,100-yr simulation was

674 forced with fixed, time-appropriate global boundary conditions that included insolation,

675 GHGs, continental ice sheets and adjustment for sea level. The key drivers of climate

676   change from the LGM through the Holocene are retreat of the NH ice sheets, deglacial

677   increased of GHG concentrations, and latitudinal and seasonal variations in insolation.

678   GENMOM reproduces reasonably well the LGM to Holocene temperature trends

679   inferred from the paleoclimate data syntheses of Shakun et al. (2012) and Marcott et al.

680   (2013). The evolution of global temperature change simulated by GENMOM is

681   consistent with three transient simulations, but is generally cooler during the deglacial

682   time slices than the transient simulations when sampled at the proxy locations. The global

683   LGM cooling of 3.8 °C simulated by GENMOM is within the range of 2.6 to 5.0 °C and

684   average of 4.4 °C simulated by the PMIP3 models. Simulated LGM cooling of the

685   tropical oceans is 1.6 °C, which is in good agreement with the MARGO reconstruction of

686   1.7 ± 1 °C. The weaker LGM global cooling is attributed to the sensitivity of GENMOM

687   to CO_2 (2.2 °C for a 2X increase in the present-day value).

688   During the LGM, simulated precipitation is reduced globally by 8.2% and

689   gradually increases through the Holocene to present-day values in response to loss of the

690   NH ice sheets, global warming and related increases in atmospheric humidity. Between

691   15 ka and 6 ka seasonal changes in insolation altered the NH land-sea temperature

692   contrasts, which, combined with shifts in global circulation, strengthened the summer

693   monsoons in Africa and India. Monsoonal precipitation in both regions peaked between

694   12 ka and 9 ka, consistent with pollen-based reconstructions. The spatial patterns of mid-

695   Holocene precipitation change simulated by GENMOM correspond well with the PMIP3

696   models, as do the 6 ka changes in monsoonal precipitation. In contrast to the pollen-based

697   reconstructions, GENMOM simulates slightly drier instead of slightly wetter-than-

698   present in Western Europe.

699   The eight time slice simulations depict the glacial-interglacial transition that is in

700   good agreement with other AOGCM simulations and compares reasonably well with

701   data-based climate reconstructions. The data-model and model-model comparisons give

702   us a measure of confidence that our paleo GENMOM simulations are reasonable on

703   broad spatial scales and adds to the growing number of climate models that are capable of

704   simulating key aspects of past climate change when constrained by a relatively small set

705   of global boundary conditions. While our simulations are not continuous across the

706   deglaciation and do not include events such as freshwater forcing, they do provide

707   insights into between-period changes, such as altered NH storm tracks and strengthening

708   of monsoons during the early to mid-Holocene and multi-century time series that are

709   useful, for example, to explore ecosystem responses to changes in mean climate and the

710   related interannual variability in the model. Future work using the model output produced

711   by this study will address how internal model variability and multidecadal variability

712   influence comparison with proxy data, particularly in North America using dynamical

713   downscaling techniques.

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Deleted: The simulations provide insights into key dynamic features of the transition, such as altered NH storm tracks and strengthening of monsoons during the early to mid-Holocene.

714 **Appendix A: List of abbreviations and acronyms**
715 AABW   Antarctic Bottom Water
716 ACC   Antarctic Circumpolar Current
717 AMOC   Atlantic Meridional Overturning Circulation
718 AOGCM   Atmosphere-Ocean General Circulation Model
719 BA   Bølling-Allerød
720 CIS   Cordilleran Ice Sheet
721 CLIMAP   Climate: Long range Investigation, Mapping, and Prediction
722 COHMAP   Cooperative Holocene Mapping Project
723 DJF   December, January and February
724 FIS   Fennoscandian Ice Sheet
725 GECCO   German partner of Estimating the Circulation and Climate of the Ocean
726 GENESIS   Global Environmental and Ecological Simulation of Interactive Systems
727 GHG   Greenhouse gas
728 ITCZ   Intertropical Convergence Zone
729 H1   Heinrich Event 1
730 JJA   June, July and August
731 LGM   Last Glacial Maximum
732 LIS   Laurentide Ice Sheet
733 LSX   Land Surface eXchange
734 MAM   March, April and May
735 MAP   Mean annual precipitation
736 MARGO   Multiproxy Approach for the Reconstruction of the Glacial Ocean Surface
737   Project
738 MAT   Mean annual temperature
739 MOM2   Modular Ocean Model version 2
740 NCAR   National Center for Atmospheric Research
741 NCEP   National Centers for Environmental Prediction
742 NH   Northern Hemisphere
743 OSU-LIS   Oregon State University Laurentide Ice Sheet
744 PD   Present-day
745 PI   Pre-industrial
746 PMIP   Palaeoclimate Modelling Intercomparison Project
747 SH   Southern Hemisphere
748 SLP   Sea-level pressure
749 SON   September, October and November
750 SST   Sea surface temperature
751 TEMPO   Testing Earth System Models with Paleoenvironmental Observations
752 YD   Youger Dryas
753

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753

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1161 Table 1. Atmospheric greenhouse gas concentrations for each time slice simulation. The

1162 21 ka through 3 ka values for CO_2 (Monnin et al., 2001), CH_4 (Brook et al., 2000) and

1163 N_2O (Sowers et al., 2003) are estimated from ice core records by averaging the gas

1164 concentrations within a ± 300 yr window centered at the time of interest. For comparison,

1165 the PMIP3 concentrations for 6 ka are 280 ppmV, 650 ppbV, and 270 ppbV for CO_2, CH_4

1166 and N_2O respectively, and 185 ppmV, 350 ppbV, and 200 ppbV for 21 ka. In the table, e

1167 is eccentricity, ω-180 is precession and ε is obliquity (Berger and Loutre, 1991).

	CO_2 (ppmV)	CH_4 (ppbV)	N_2O (ppbV)	e	ω-180	ε
PD	355	1714	311	0.0176	101.37	23.446
PI	280	760	270	0.0176	101.37	23.446
3 ka	275	627	264	0.0183	50.30	23.815
6 ka	260	596	227	0.0192	0.01	24.100
9 ka	260	677	244	0.0198	310.32	24.229
12 ka	240	500	246	0.0201	261.07	24.161
15 ka	220	500	216	0.0202	212.04	23.895
18 ka	188	382	219	0.0199	163.04	23.475
21 ka	188	392	199	0.0194	113.98	22.989

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1170   Table 2. Annual average 2-m air temperatures and precipitation rates for the time slice

1171   simulations. NCEP is from the National Center for Environmental Predication

1172   NCEP/NCAR Reanalysis data set (Kalnay et al., 1996), PD2X is the 2xCO$_2$ simulation,

1173   PD is present day and PI is pre-industrial. Parenthetical values are the changes from the

1174   previous time slice, e.g., the global average temperature for the PD is 0.77 °C warmer

1175   than the PI.

	Temperature (K)			Precipitation (mm d^{-1})		
	Global	Land	Ocean	Global	Land	Ocean
NCEP (1980-2000)	287.52 --	281.66 --	289.84 --	3.09 --	2.30 --	3.40 --
PD2X	288.48 (2.2)	282.06 (2.69)	291.29 (1.91)	3.11 (0.11)	2.17 (0.10)	3.53 (0.13)
PD	286.34 (0.77)	279.37 (0.93)	289.38 (0.70)	3.00 (0.04)	2.07 (0.04)	3.40 (0.05)
PI	285.57 (0.07)	278.44 (-0.03)	288.68 (0.15)	2.95 (0.00)	2.03 (-0.02)	3.36 (0.02)
3 ka	285.50 (0.32)	278.47 (0.30)	288.53 (0.33)	2.95 (0.02)	2.05 (0.00)	3.34 (0.04)
6 ka	285.17 (0.23)	278.18 (0.95)	288.20 (-0.27)	2.93 (0.02)	2.05 (0.00)	3.30 (0.02)
9 ka	284.95 (0.74)	277.23 (1.63)	288.47 (0.22)	2.91 (0.05)	2.05 (0.06)	3.30 (0.03)
12 ka	284.21 (1.40)	275.60 (2.44)	288.25 (0.70)	2.86 (0.09)	1.99 (0.12)	3.26 (0.05)
15 ka	282.81 (0.93)	273.16 (1.53)	287.55 (0.53)	2.77 (0.05)	1.87 (0.09)	3.21 (0.02)
18 ka	281.88 (0.16)	271.63 (0.28)	287.02 (0.06)	2.72 (0.01)	1.78 (0.01)	3.19 (0.01)
21 ka	281.72	271.35	286.96	2.71	1.78	3.18

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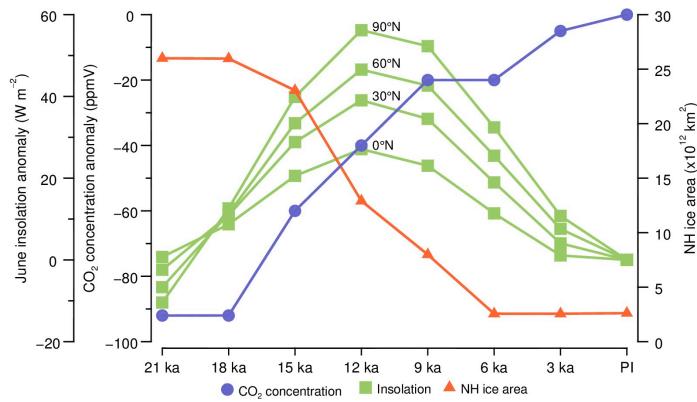
Fig. 1. Boundary conditions for the time slice simulations. CO_2 concentrations are relative to the PI concentration of 280 ppmV. NH ice area is the total area covered by the continental ice sheets. June insolation anomalies are relative to PI at the indicated latitude. Mid-month insolation data from Berger and Loutre (1991).
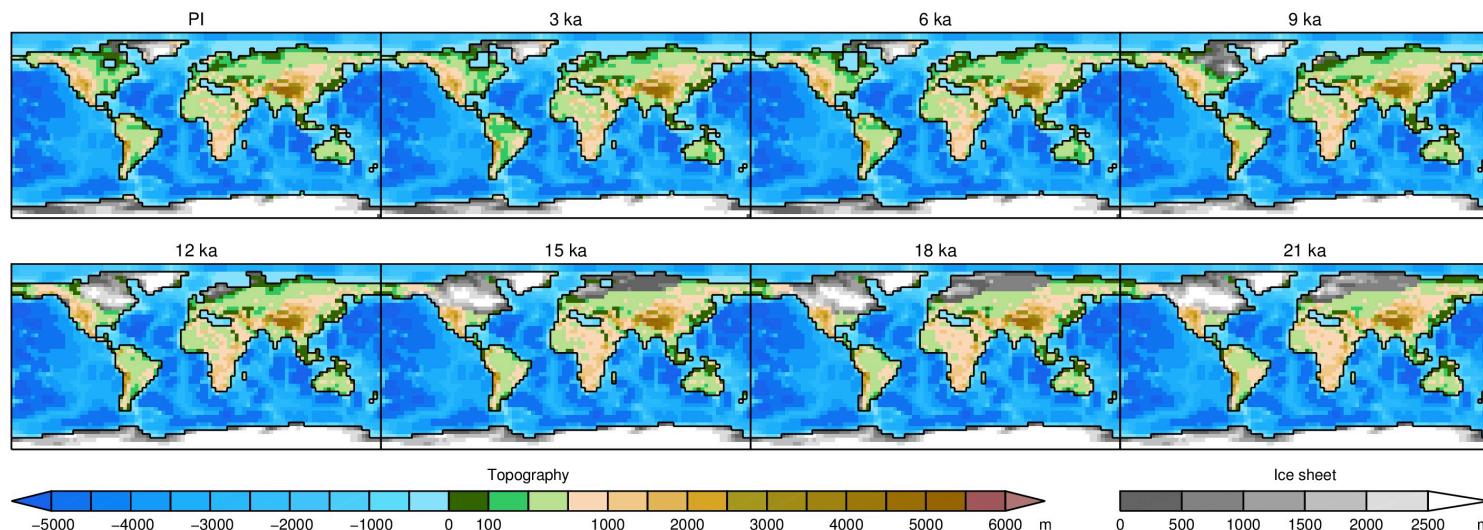
Fig. 2. Orography for the time slice simulations, with ice sheet height and extent derived from ICE-4G (Peltier, 2002) for the

Fennoscandian, Cordilleran and Antarctic, and OSU-LIS (Licciardi et al., 1998) for the Laurentide.
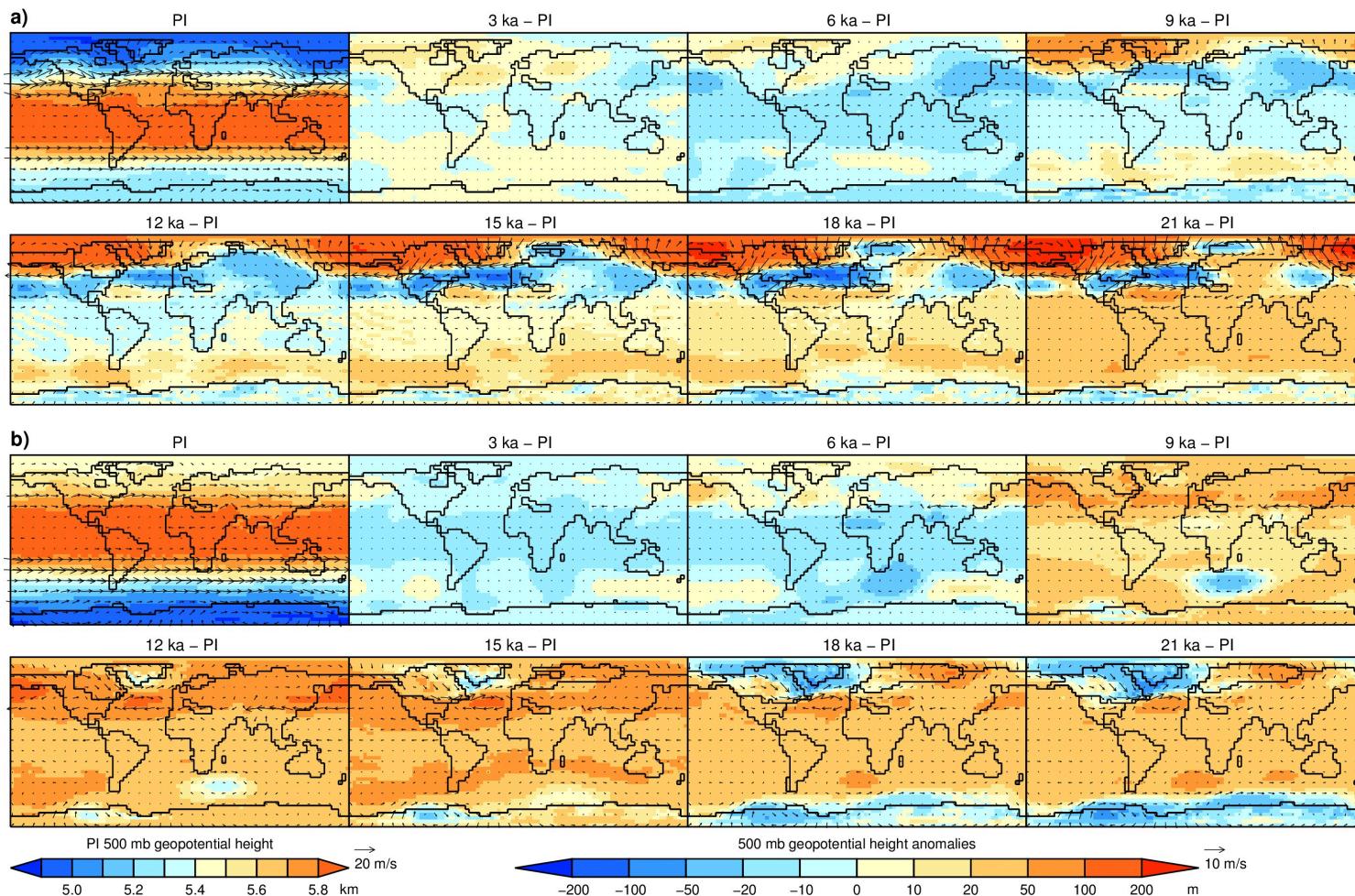
Fig. 3. Simulated seasonal 500 hPa geopotential height and wind anomalies relative to PI. a) December, January, and February and b) June, July and August. Raw 500 hPa geopotential height and wind are shown in SFig. 2.
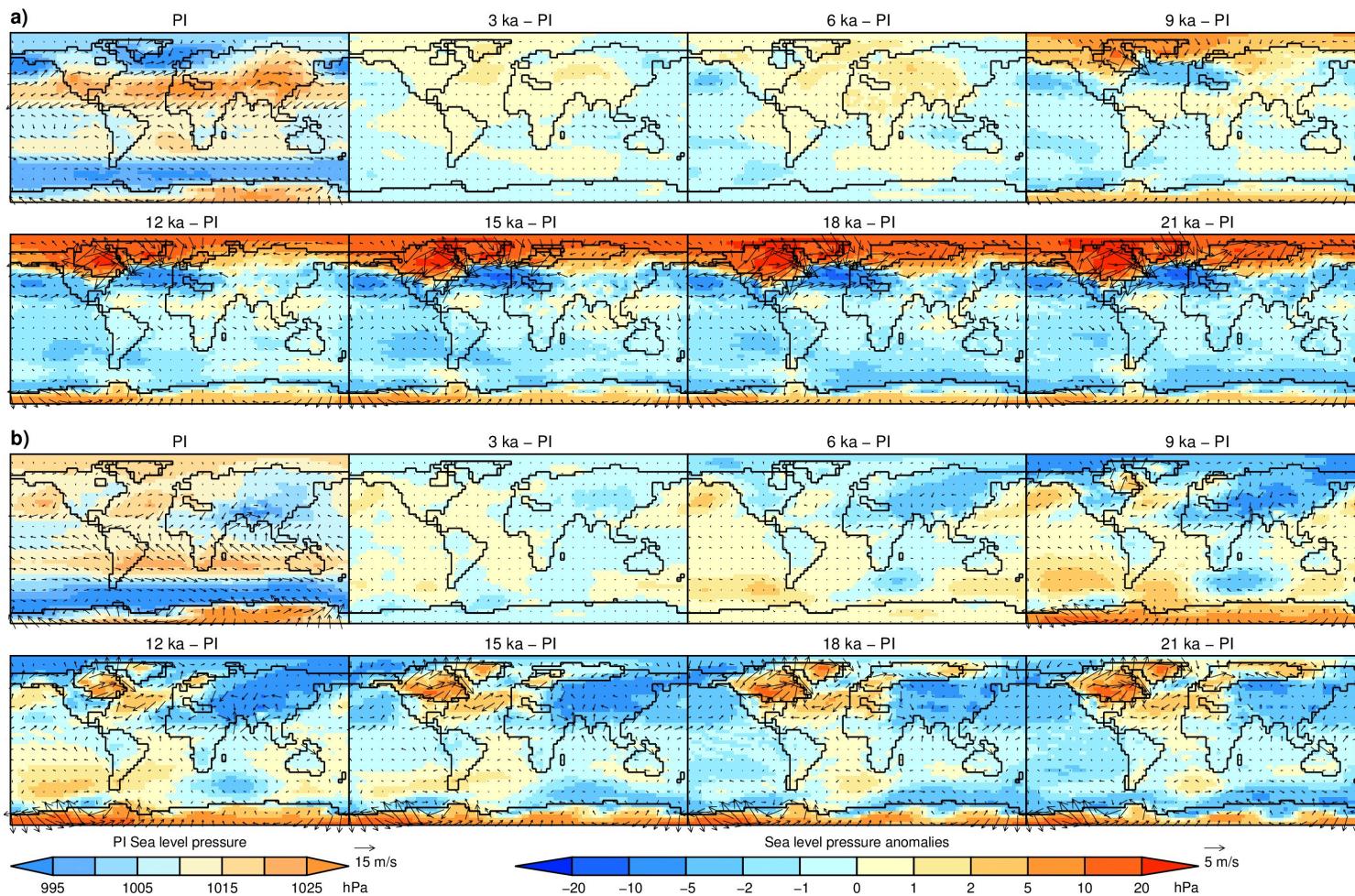
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1192 Fig. 4. Simulated seasonal average sea-level pressure and 2-m wind anomalies relative to PI. a) December, January, and February and

1193 b) June, July, and August. Raw sea level pressure and wind are shown in SFig. 3.

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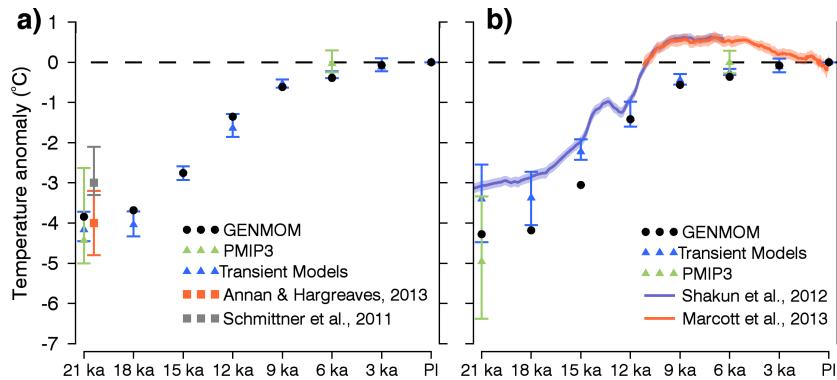
1197   Fig. 5. Simulated and reconstructed changes in temperature from 21 ka to present. a)

1198   Global mean surface air temperature from GENMOM compared to the PMIP3 ensemble

1199   average and three transient models (CCSM (Liu et al., 2009), LOVECLIM (Timm and

1200   Timmermann, 2007) and FAMOUS (Smith and Gregory, 2012)). The transient model

1201   values are averages over a ±50 yr window centered on the eight time slices. The symbols

1202   for the PMIP and transient models are the average of the ensembles and the bars

1203   represent the range of the ensembles. Data-model estimates of mean and range of LGM

1204   cooling by Annan and Hargreaves (2013) and Schmittner et al. (2011b) are offset from

1205   21 ka for legibility. b) Temperature change at the proxy sites used in the reconstructions

1206   by Shakun et al. (2012) and Marcott et al (2013). The models were bilinearly interpolated

1207   and aggregated to the 5° x 5° boxes around the proxy sites as in Marcott et al. (2013).

1208   The 1σ uncertainty in the reconstructions is indicated by the shaded band. Marcott et al.

1209   (2013) is adjusted to a pre-industrial (~1850) base value rather than the original 1961-

1210   1990. Data younger than pre-industrial are removed. The Shakun et al. (2012) and

1211   Marcott et al. (2013) time series are joined at their 11.5 ka – 6.5 ka means.
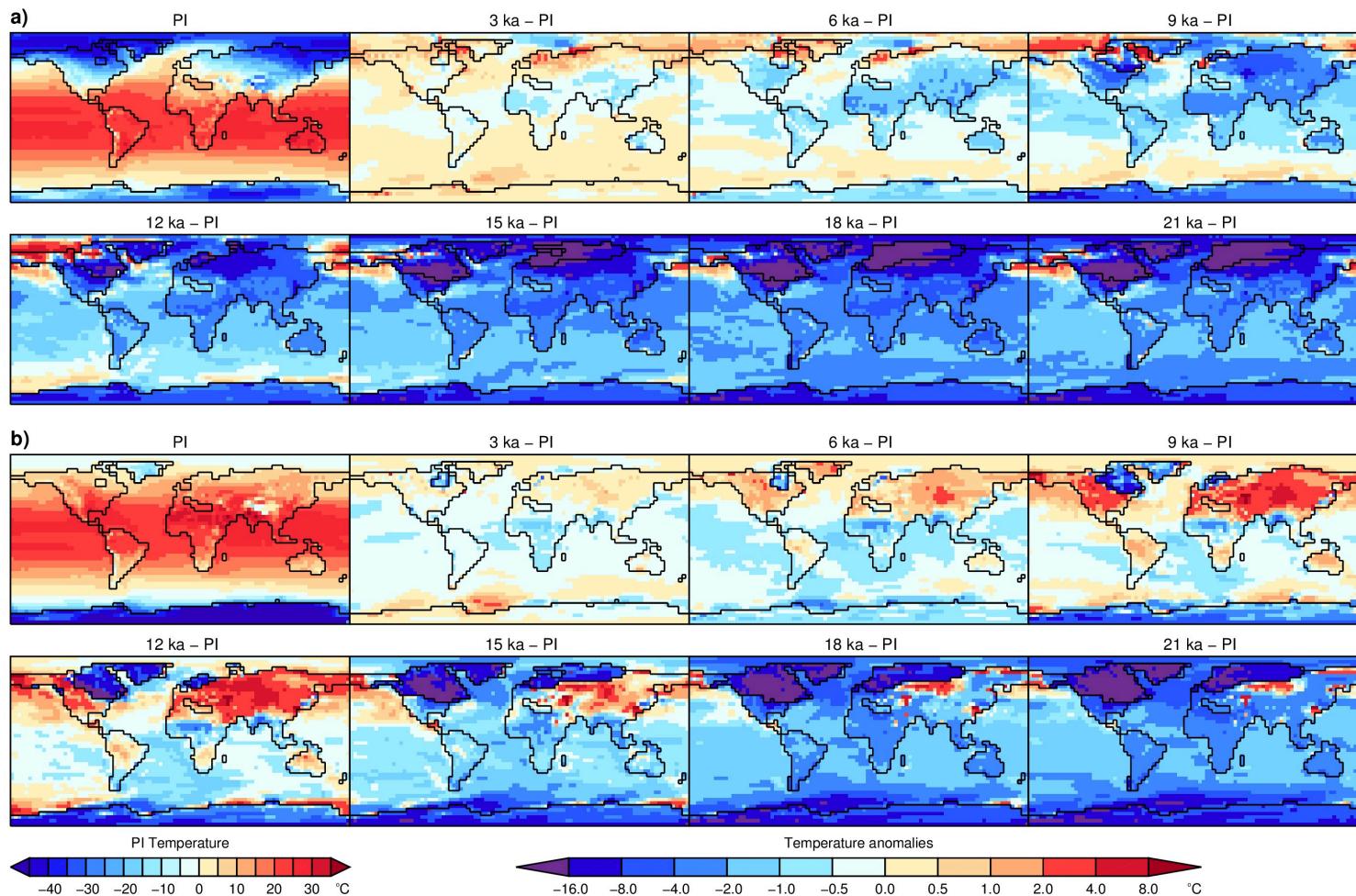
1212
1213 Fig. 6. Simulated seasonal average 2-m air temperature anomalies relative to PI. a) December, January, and February and b) June,

1214 July, and August.
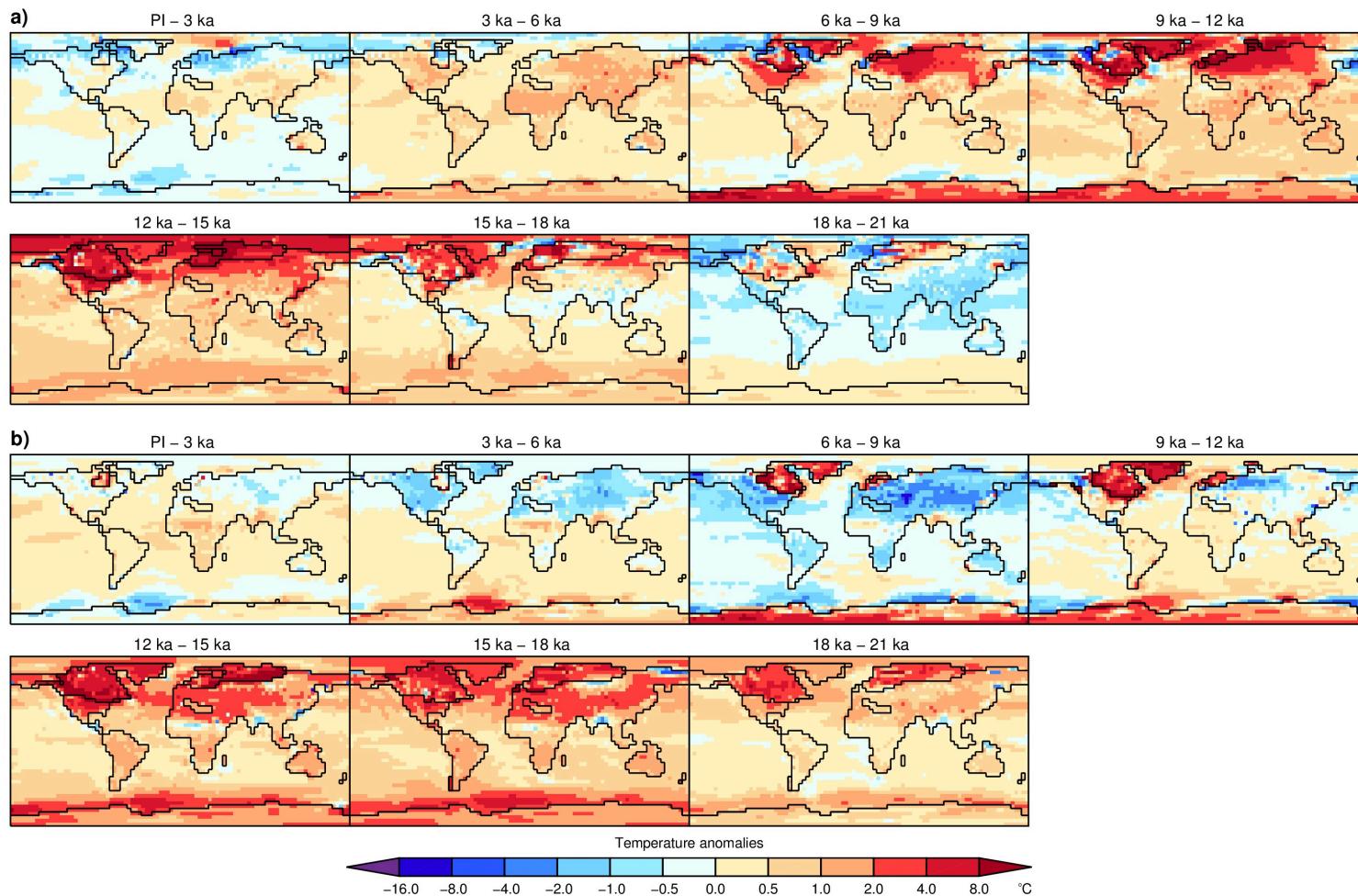
Fig. 7. Simulated seasonal average changes in 2-m air between consecutive time slices. a) December, January, and February and b) June, July, and August.
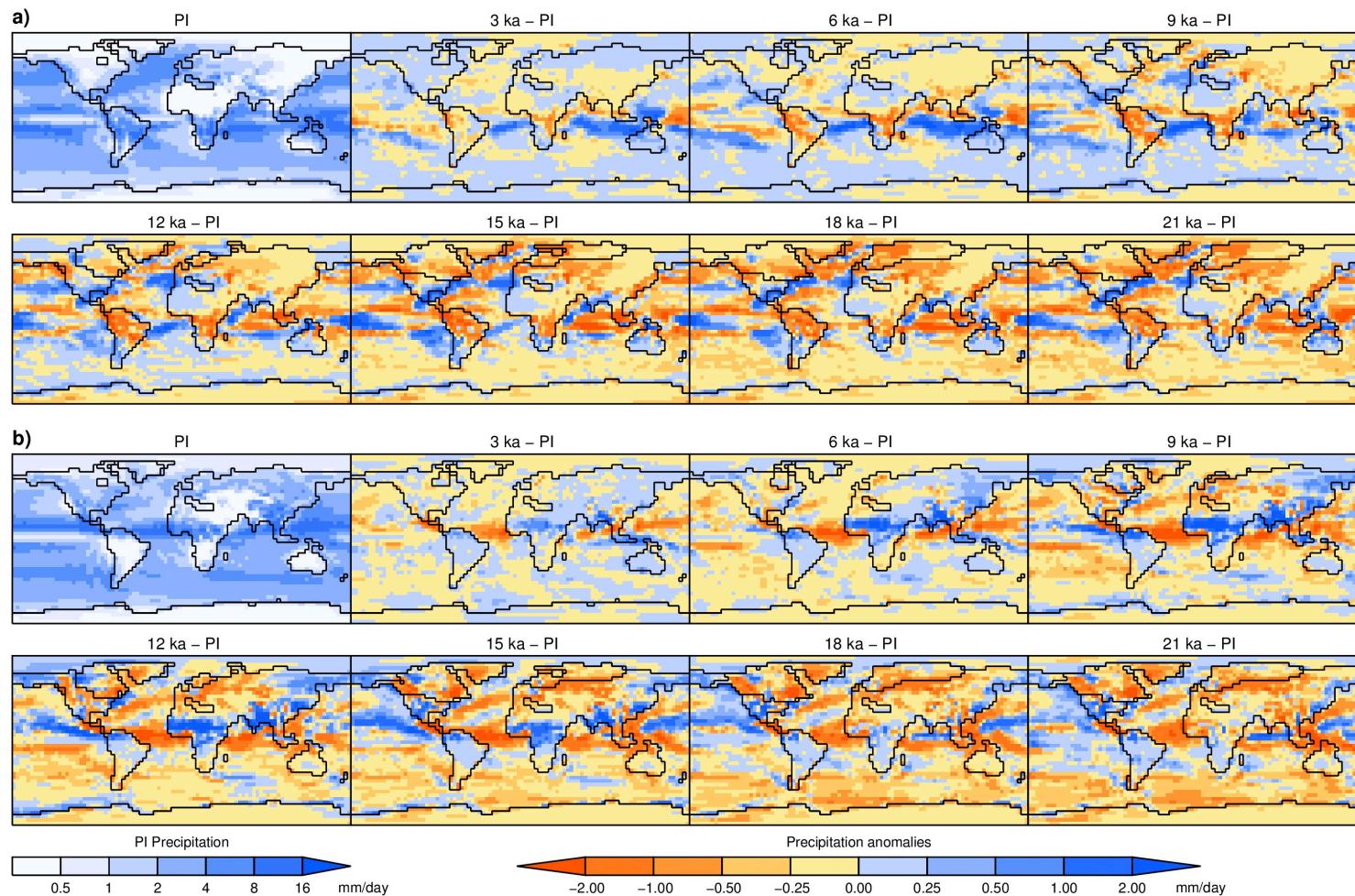
Fig. 8. Simulated seasonal average precipitation anomalies relative to PI. a) December, January, and February and b) June, July, and August.
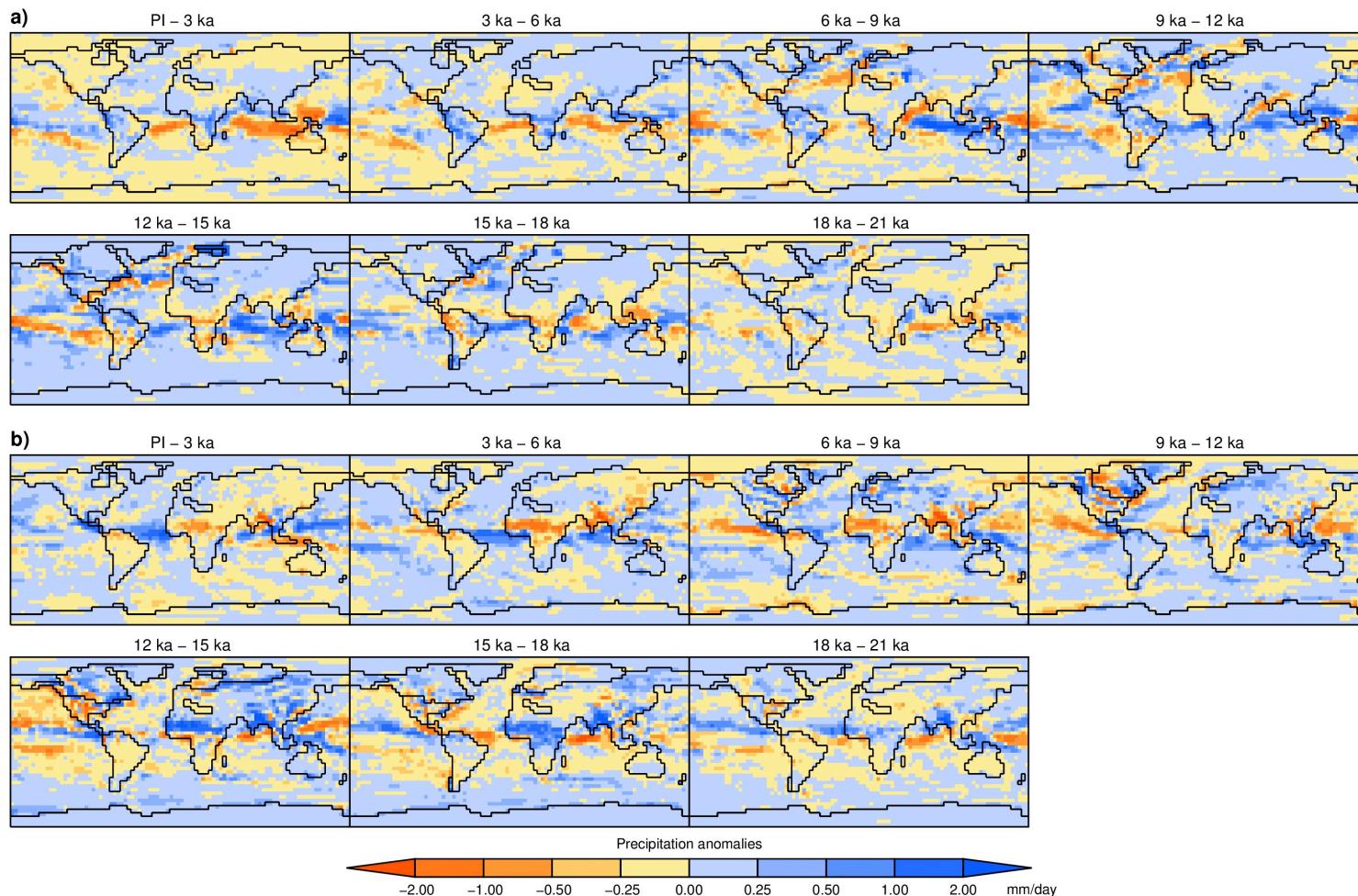
Fig. 9. Simulated seasonal average precipitation changes between consecutive time slices. a) December, January, and February and b) June, July, and August.

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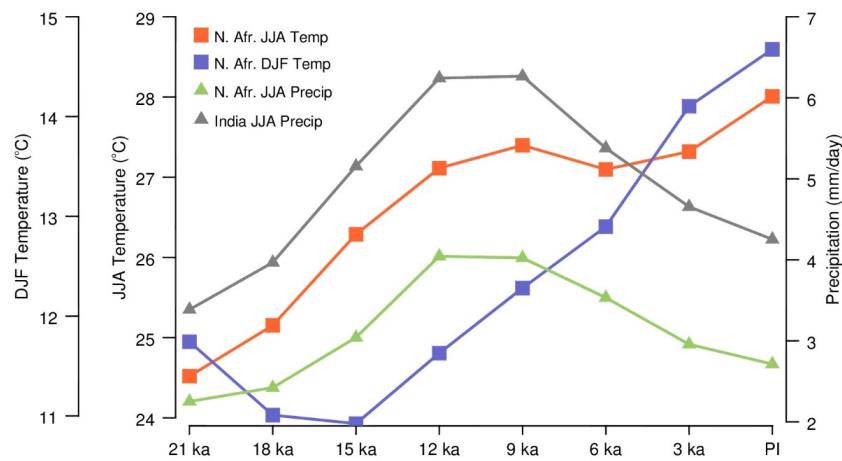
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1227　Fig. 10. Time evolution of North African and Indian summer monsoons. The North

1228　Africa monsoon region is defined as 12°N - 30°N, 20°W - 30°E and India monsoon

1229　region is defined as 20°N - 40°N, 70°E - 100°E (Zhao and Harrison, 2012).

1230

NH Winter

NH Summer

PI

6 ka

21 ka

0.1 0.2 0.3 0.4 0.5 0.6 0.7 0.8 0.9 Frac
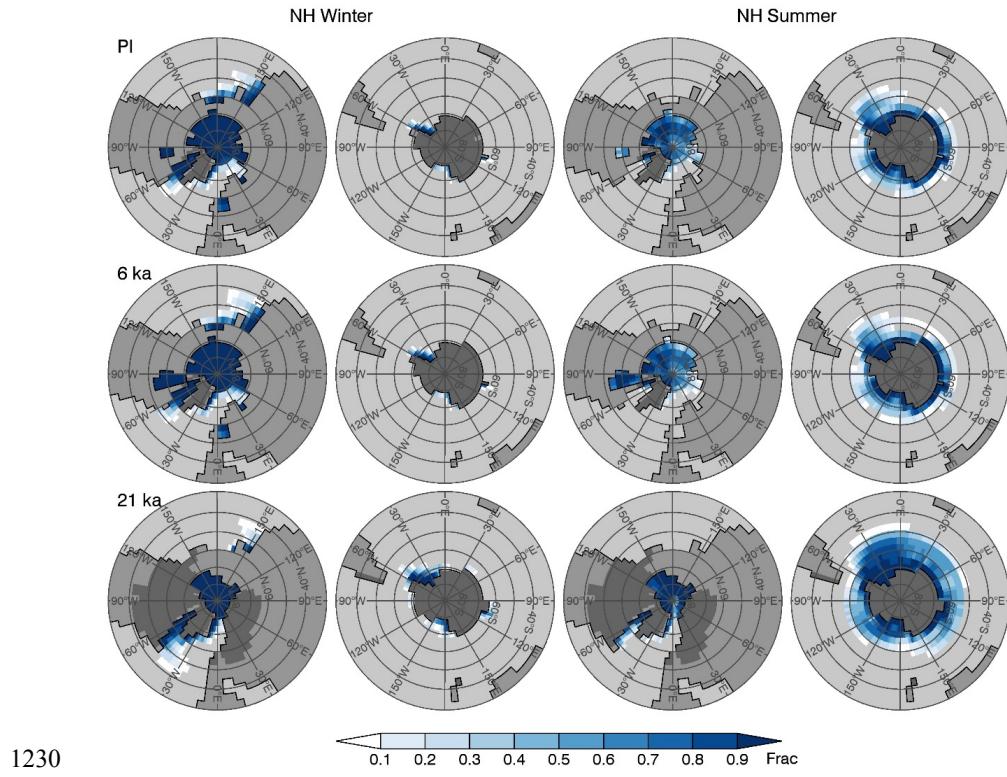
Fig. 11 Simulated sea-ice fraction for PI, 6 ka and 21 ka. Left two columns: February-March and right two columns: August-September. Medium gray is continental land mass and dark gray is continental ice sheet.
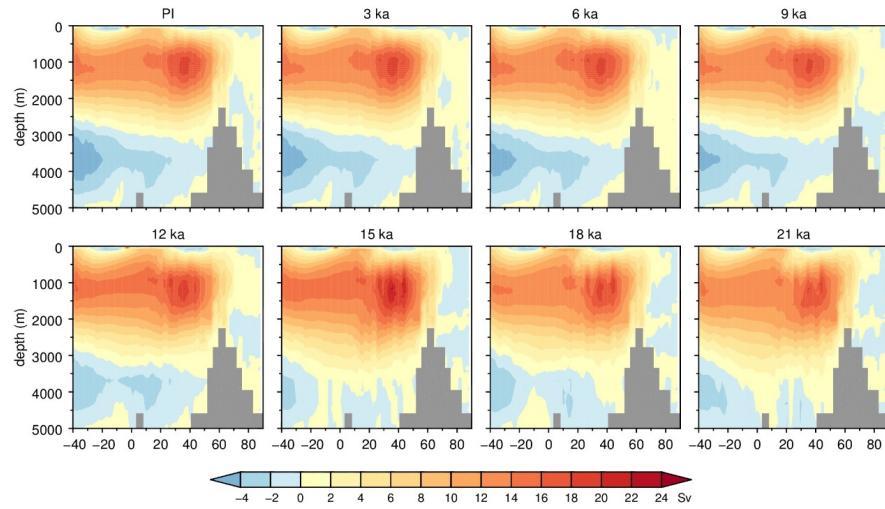
Fig. 12. Simulated annual average Atlantic Meridional Overturning Circulation (AMOC)
for the eight time-slices.
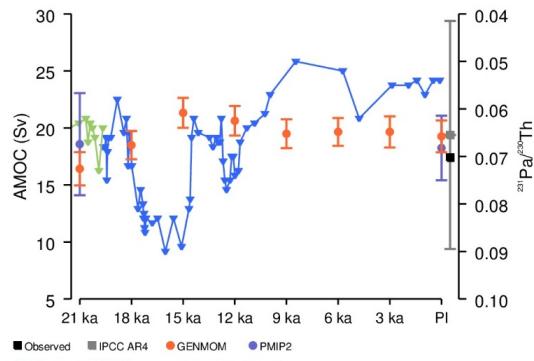
Fig. 13. Simulated Atlantic Meridional Overturning Circulation (AMOC) compared to ^{231}Pa/^{230}Th proxy record at 33°N and other AOGCMs. Observations are from 26.5°N. GENMOM values are 100-yr averages with error bars representing standard deviations. The mean and standard deviation of the maximum AMOC in the five PMIP2 models. The IPCC AR4 point represents the mean and standard deviation from a collection of IPCC AR4 models. ^{231}Pa/^{230}Th data from McManus et al. (2004) and Lippold et al. (2009); observed value from Srokosz et al. (2012), PMIP2 data from Weber et al. (2007), and IPCC data from Schmittner et al. (2005).
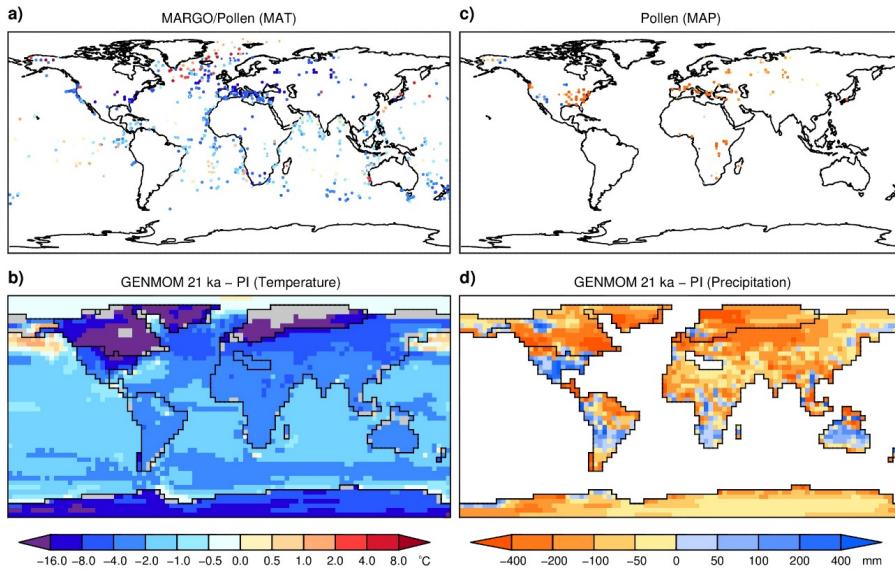
Fig. 14. Changes in 21 ka mean annual temperature (MAT) and precipitation (MAP)

inferred from data and simulated by GENMOM. a) blended sea surface temperature from

MARGO (Waelbroeck et al., 2009) and terrestrial temperature from Bartlein et al. (2011),

b) GENMOM temperature anomalies (blended sea surface temperature and 2-m air

temperature over land), c) precipitation from Bartlein et al. (2011), and d) GENMOM

precipitation anomalies. Grid cells with different land mask types in the 21 ka and PI

simulation are shaded in gray to avoid comparing ocean temperature to land temperature

in emergent cells.
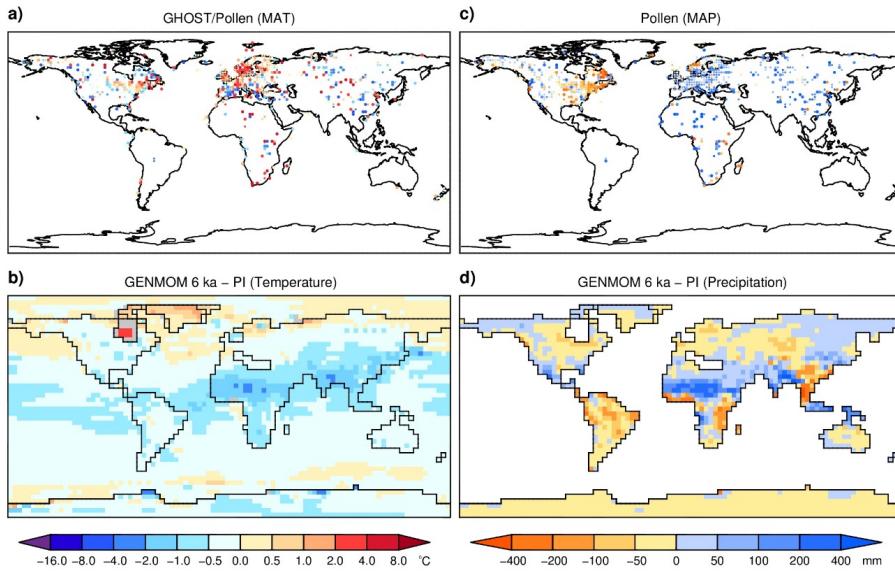
Fig. 15. Changes in 6 ka mean annual temperature (MAT) and precipitation (MAP)
inferred from data and simulated by GENMOM. a) blended sea surface temperature from
Leduc et al. (2010) and terrestrial temperature from Bartlein et al. (2011), b) GENMOM
temperature anomalies (blended sea surface temperature and 2-m air temperature over
land), c) precipitation from Bartlein et al. (2011) and d) GENMOM precipitation
anomalies. Grid cells with different land mask types in the 6 ka and PI simulation are
shaded in gray to avoid comparing ocean temperature to land temperature in emergent
cells.