



1 **The UK contribution to CMIP6/PMIP4: mid-Holocene and Last**
2 **Interglacial experiments with HadGEM3, and comparison to the pre-**
3 **industrial era and proxy data**

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32



33 **ABSTRACT**

34 Palaeoclimate model simulations are an important tool to improve our understanding of the
35 mechanisms of climate change. These simulations also provide tests of the ability of models to
36 simulate climates very different to today. Here we present the results from two simulations using the
37 latest version of the UK's physical climate model, HadGEM3-GC3.1; the mid-Holocene (~6 ka) and
38 Last Interglacial (~127 ka) simulations, both conducted under the auspices of CMIP6/PMIP4. These
39 periods are of particular interest to PMIP4 because they represent the two most recent warm periods
40 in Earth history, where atmospheric concentration of greenhouse gases and continental configuration
41 is similar to the pre-industrial period but where there were significant changes to the Earth's orbital
42 configuration, resulting in a very different seasonal cycle of radiative forcing.

43
44 Results for these simulations are assessed against proxy data, previous versions of the UK model, and
45 models from the previous CMIP5 exercise. When the current version is compared to the previous
46 generation of the UK model, the most recent version suggests limited improvement. In common with
47 these previous model versions, the simulations reproduce global land and ocean temperatures (both
48 surface and at 1.5 m) and a West African monsoon that is consistent with the latitudinal and seasonal
49 distribution of insolation. The Last Interglacial simulation appears to accurately capture Northern
50 Hemisphere temperature changes, but without the addition of Last Interglacial meltwater forcing
51 cannot capture the magnitude of Southern Hemisphere changes. Model-data comparisons indicate
52 that some geographical regions, and some seasons, produce better matches to the palaeodata (relative
53 to pre-industrial) than others. Model-model comparisons, relative to previous generations same
54 model and other models, indicate similarity between generations in terms of both the intensity and
55 northward enhancement of the mid-Holocene West African monsoon, both of which are
56 underestimated. On the 'Saharan greening' which occurred the mid-Holocene African Humid Period,
57 simulation results are likewise consistent with other models. The most recent version of the UK
58 model appears to still be unable to reproduce the amount of rainfall necessary to support grassland
59 across the Sahara.

60



61 **1. INTRODUCTION**

62 Simulating past climates has been instrumental in improving our understanding of the mechanisms of
63 climate change (e.g. Gates 1976, Haywood *et al.* 2016, Jungclaus *et al.* 2017, Kageyama *et al.* 2017,
64 Kageyama *et al.* 2018, Kohfeld *et al.* 2013, Lunt *et al.* 2008, Otto-Bliesner *et al.* 2017, Ramstein *et al.*
65 1997), as well as in identifying and assessing discrepancies in palaeoclimate reconstructions (e.g.
66 Rind & Peteet 1985). Palaeoclimate scenarios can also provide tests of the ability of models to
67 simulate climates that are very different to today, often termed ‘out-of-sample’ tests. This notion
68 underpins the idea that robust simulations of past climates improve our confidence in future climate
69 change projections (Braconnot *et al.* 2011, Harrison *et al.* 2014, Taylor *et al.* 2011). Palaeoclimate
70 scenarios have also been used to provide additional tuning targets for models (e.g. Gregoire *et al.*
71 2011), in combination with historical or pre-industrial conditions.

72

73 The international Climate Model Intercomparison Project (CMIP) and the Palaeoclimate Model
74 Intercomparison Project (PMIP) have spearheaded the coordination of the international palaeoclimate
75 modelling community to run key scenarios with multiple models, perform data syntheses, and
76 undertake model-data comparisons since their initiation twenty-five years ago (Joussaume & Taylor
77 1995). Now in its fourth incarnation, PMIP4 (part of the sixth phase of CMIP, CMIP6), it includes a
78 larger set of models than previously, and more palaeoclimate scenarios and experiments covering the
79 Quaternary (documented in Jungclaus *et al.* 2017, Kageyama *et al.* 2017, Kageyama *et al.* 2018 and
80 Otto-Bliesner *et al.* 2017) and Pliocene (documented in Haywood *et al.* 2016).

81

82 PMIP4 specifies experiment set-ups for two warm interglacial simulations: the mid-Holocene (MH) at
83 ~6 ka and the Last Interglacial (LIG) covering ~129-116 ka. These are the two most recent warm
84 periods in Earth history, and are of particular interest to PMIP4; indeed, the MH experiment is one of
85 the two entry cards into PMIP (Otto-Bliesner *et al.* 2017). This is because whilst the atmospheric
86 concentration of greenhouse gases, the extent of land ice, and the continental configuration is similar
87 in these PMIP4 set-ups compared to the pre-industrial (PI) period, significant changes to the seasonal
88 cycle of radiative forcing, relative to today, do occur during these periods due to long-term variations
89 in the Earth’s orbital configuration. The MH and LIG both have higher boreal summer insolation and
90 lower boreal winter insolation compared to the PI, as shown by Figure 1, leading to an enhanced
91 seasonal cycle in insolation as well as a change in its latitudinal distribution. The change is more
92 significant in the LIG than the MH, due to the larger eccentricity of the Earth’s orbit at that time.

93

94 Palaeodata syntheses indicate globally warmer surface conditions of potentially ~0.7°C than PI in the
95 MH (Marcott *et al.* 2013) and up to ~1.3°C in the LIG (Fischer *et al.* 2018). Recent palaeodata
96 compilations (Capron *et al.* 2014, Hoffman *et al.* 2017) reveal that the maximum temperatures were



97 reached asynchronously in the LIG between the Northern and Southern Hemispheres. Furthermore,
98 model simulations suggest that this may have been caused by meltwater induced shutdown of the
99 Atlantic Meridional Overturning Circulation (AMOC) in the early part of the LIG, due to the melting
100 of the Northern Hemisphere ice sheets during the preceding deglaciation (e.g. Stone *et al.* 2016).
101 During both warm periods there is abundant palaeodata evidence indicating enhancement of Northern
102 Hemisphere summer monsoons (e.g. Wang *et al.* 2008) and in the case of the Sahara, replacement of
103 desert by shrubs and steppe vegetation (e.g. Drake *et al.* 2011, Hoelzmann *et al.* 1998) and inland
104 water bodies (e.g. Drake *et al.* 2011, Lezine *et al.* 2011).

105

106 The driving mechanism producing the climate and environmental changes indicated by the palaeodata
107 for the LIG and MH is different to current and future anthropogenic warming, as the former results
108 from orbital forcing changes whilst the latter results from increases in greenhouse gases. However,
109 these past warm intervals are a unique opportunity to understand the magnitudes of forcings and
110 feedbacks in the climate system that produce warm interglacial conditions, which can help us
111 understand and constrain future climate projections (e.g. Holloway *et al.* 2016, Rachmayani *et al.*
112 2017, Schmidt *et al.* 2014). Running the same model scenarios with ever newer models enables the
113 testing of whether model developments are producing improvements in palaeo model-data
114 comparisons, assuming appropriate boundary conditions are used. Previous iterations of PMIP, with
115 older versions of the PMIP4 models, have uncovered persistent shortcomings (Harrison *et al.* 2015)
116 that have not been eliminated despite developments in resolution, model physics, and addition of
117 further Earth system components. One key example of this is the continued underestimation of the
118 increase in rainfall over the Sahara in the MH PMIP simulations (e.g. Braconnot *et al.* 2012).

119

120 In this study we run and assess the latest version of the UK's physical climate model, HadGEM3-
121 GC3.1. In Global Coupled (GC) version 3 (and therefore the following GC3.1), there have been
122 many updates and improvements, relative to its predecessors, which are discussed extensively in
123 Williams *et al.* (2017) and a number of companion scientific model development papers (see Section
124 2.1). As a brief introduction, however, GC3 includes a new aerosol scheme, multilayer snow scheme,
125 multilayer sea ice and several other parametrization changes, including a set relating to cloud and
126 radiation, as well as a revision to the numerics of convection (Williams *et al.* 2017). In addition, the
127 ocean component of GC3 has other changes including a new ocean and sea ice model, a new cloud
128 scheme, and further revisions to all parametrization schemes (Williams *et al.* 2017). See Section 2.1
129 for further details.

130

131 Following the CMIP6/PMIP4 protocol, here the PMIP4 MH and LIG simulations have been
132 conducted and assessed, comparing the results with available proxy data, previous versions of the
133 UK's same physical climate model, and other models from CMIP5. The focus of this paper is on the



134 fidelity of the temperature anomalies globally and the degree of precipitation enhancement in the
135 Sahara, the latter of which has proved problematic for several generations of models. The results
136 discussed here are split into two sections: after an assessment of the level of equilibrium gained
137 during the spin-up phase, the main focus is on the model-data and model-model comparisons using
138 the production runs. Following this introduction, Section 2 describes the model, the experimental
139 design and the proxy data used for the model-data comparisons. Section 3 then presents the results,
140 divided into two subsections: i) equilibrium during the spin-up phase; and ii) model-data and model-
141 model comparisons from the production runs. Finally, section 4 summarises and concludes.

142

143 **2. MODEL, EXPERIMENT DESIGN AND DATA**

144 **2.1. Model**

145 The MH and LIG simulations conducted here (referred to as *midHolocene* and *lig127k*, respectively,
146 and collectively as the ‘warm climate’ simulations), and indeed the PI simulation (*piControl*,
147 conducted elsewhere as part of the UK’s CMIP6 runs and used here for comparative purposes) were
148 all run using the same fully-coupled GCM: the Global Coupled 3 configuration of the UK’s physical
149 climate model, HadGEM3-GC3.1. Full details on HadGEM3-GC3.1, and a comparison to previous
150 configurations, are given in Williams *et al.* (2017) and Kuhlbrodt *et al.* (2018). Here, the model was
151 run using the Unified Model (UM), version 10.7, and including the following components: i) Global
152 Atmosphere (GA) version 7.1, with an N96 atmospheric spatial resolution (approximately 1.875°
153 longitude by 1.25° latitude) and 85 vertical levels; ii) the NEMO ocean component, version 3.6,
154 including Global Ocean (GO) version 6.0 (ORCA1), with an isotropic Mercator grid which, despite
155 varying in both meridional and zonal directions, has an approximate spatial resolution of 1° by 1° and
156 75 vertical levels; iii) the Global Sea Ice (GIS) component, version 8.0 (GSI8.0); iv) the Global Land
157 (GL) configuration, version 7.0, of the Joint UK Land Environment Simulator (JULES); and v) the
158 OASIS3 MCT coupler. The official title for this configuration of HadGEM3-GC3.1 is HadGEM3-
159 GC31-LL N96ORCA1 UM10.7 NEMO3.6 (for brevity, hereafter HadGEM3).

160

161 All of the above individual components are summarised by Williams *et al.* (2017) and detailed
162 individually by a suite of companion papers (see Walters *et al.* 2017 for GA7 and GL7, Storkey *et al.*
163 2017 for GO6 and Ridley *et al.* 2017 for GIS8). However, a brief description of the major changes
164 relative to its predecessor are given here. Beginning with GA7 and GL7, a once-in-a-decade
165 replacement of the model’s dynamical core, implementing ENDGame, was undertaken for the
166 previous version (GA6) and therefore remains the same in GA7 (Walters *et al.* 2017). In addition, a
167 number of bottom-up and top-down developments were included in GA7. For the former, these
168 include improvements to the radiation scheme to allow better treatment of gases absorption,
169 improvements to how warm rain and ice clouds are treated, and an improvement to the numerics of
170 the convection scheme (Walters *et al.* 2017). For the latter, these include further improvements to the



171 microphysics as well as an incremental development of ENDGame (Walters *et al.* 2017). Together
172 these led to reductions in four model errors that were deemed critical in the previous configuration: i)
173 South Asian monsoon rainfall biases over India; ii) biases in both temperature and humidity in the
174 tropical tropopause; iii) shortcomings in the numerical conservation; and iv) biases in surface
175 radiation fluxes over the Southern Ocean (Walters *et al.* 2017). In addition to these developments,
176 two new parameterisation schemes were introduced in GA7: firstly the UK Chemistry and Aerosol
177 (UKCA) GLOMAP-mode aerosol scheme, to improve the representation of tropospheric aerosols, and
178 secondly a multi-layer snow scheme in JULES, to allow the first time inclusion of stochastic physics
179 in UM climate simulations (Walters *et al.* 2017).

180

181 For the GO and GIS components, a number of improvements to GO6 have been made since the
182 previous version, the first of which was an upgrade of the NEMO base code (to version 3.6) which
183 allowed a formulation for momentum advection (from Hollingsworth *et al.* 1983), a Lagrangian
184 icebergs scheme, and a simulation of circulation below ice shelves (Storkey *et al.* 2018). Other
185 developments included an improvement to the warm SST bias in the Southern Ocean (as detailed by
186 Williams *et al.* 2017), as well as tuning to various parameters e.g. the isopycnal diffusion (Storkey *et al.*
187 2018). For GIS8, along with improvements to the albedo scheme and more realistic semi-implicit
188 coupling, the biggest development since its predecessor is the inclusion of multilayer
189 thermodynamics, giving a heat capacity to the sea ice and allowing vertical variation of conduction
190 (Ridley *et al.* 2018). Testing of these two components produced a better simulation compared to its
191 predecessor, with more realistic mixed layer depths in the Southern Ocean and the aforementioned
192 reduced warm bias, the latter of which was deemed primarily due to the tuning of the different mixing
193 (e.g. vertical and isopycnal) parameters (Storkey *et al.* 2018).

194

195 When all of these components are coupled together to give GC3, there have been several
196 improvements relative to its predecessor (GC2), most noticeably to the large warm bias in the
197 Southern Ocean (which was reduced by 75%), as well as an improved simulation of clouds, sea ice,
198 the frequency of tropical cyclones in the Northern Hemisphere as well as the AMOC, and the Madden
199 Julian Oscillation (MJO) (Williams *et al.* 2017). Relative to the previous fully-coupled version of the
200 model (HadGEM2), which was submitted to the last CMIP5/PMIP3 exercise, many systematic errors
201 have been improved including a reduction in many regions to the temperature bias, a better simulation
202 of mid-latitude synoptic variability, and an improved simulation of tropical cyclones and the El Niño
203 Southern Oscillation (ENSO) (Williams *et al.* 2017).

204

205 Here, the *midHolocene* and *lig127k* simulations were both run on the UK National Supercomputing
206 Service, ARCHER, whereas the *piControl* was run on a different platform based within the UK Met
207 Office's Hadley Centre. While this may mean that anomalies computed against the *piControl* are



208 potentially influenced by different computing environments, and not purely the result of different
209 climate forcings, the reproducibility of GC3.1 simulations across different platforms has been tested
210 (Guarino *et al.* 2019). It was found that, although a simulation length of 200 years is recommended
211 whenever possible to adequately capture climate variability across different platforms, the main
212 climate variables considered here (e.g. surface temperature) are not expected to be significantly
213 different on a 100- or 50-year timescale (see, for example, Fig. 6 in Guarino *et al.* [2019]) as they are
214 not directly affected by medium-frequency climate processes such as ENSO.

215

216 Not including queueing time, both simulations were achieving 3-4 model years per day during the
217 spin-up phase, and 1-2 model years per day during the production run; see below for the differences in
218 output, and therefore speed, between the two phases.

219

220 2.2. Experiment design

221 Full details of the experimental design, and results from the CMIP6 *piControl* simulation, are
222 documented in Menary *et al.* (2018). Both the warm climate simulations followed the experimental
223 design given by Otto-Bliesner *et al.* (2017), and specified at
224 https://pmip4.lsce.ipsl.fr/doku.php/exp_design:index. The primary differences from the *piControl*
225 were to the astronomical parameters and the atmospheric trace greenhouse gas concentrations,
226 summarised in Table 1. For the astronomical parameters, these were prescribed in Otto-Bliesner *et al.*
227 (2017) according to orbital constants from Berger & Loutre (1991). However, in HadGEM3, the
228 individual parameters (e.g. eccentricity, obliquity, etc) use orbital constants based on Berger (1978),
229 according to the specified start date of the simulation. For the atmospheric trace greenhouse gas
230 concentrations, these were based on recent reconstructions from a number of sources (see Table 1 for
231 values, and section 2.2 in Otto-Bliesner *et al.* [2017] for a full list of references/sources).

232

233 All other boundary conditions, including solar activity, ice sheets, topography and coastlines, volcanic
234 activity and aerosol emissions, are identical to the CMIP6 *piControl* simulation. Likewise, vegetation
235 was prescribed to present-day values, to again match the CMIP6 *piControl* simulation. As such, the
236 *piControl* and both the warm climate simulations actually include a prescribed fraction of urban land
237 surface. As a result of this, our orbitally- and greenhouse gas-forced simulations should be considered
238 as anomalies to the *piControl*, rather than absolute representations of the MH or LIG climate.

239

240 Both the warm climate simulations were started from the end of the *piControl* spin-up phase (which
241 ran for approximately 600 years), after which time the *piControl* was considered to be in atmospheric
242 and oceanic equilibrium (Menary *et al.* 2018). To assess this, four metrics were used, namely net
243 radiative balance at the top of the atmosphere (TOA), surface air temperature (SAT), and full-depth
244 ocean temperature (OceTemp) and salinity (OceSal) Menary *et al.* (2018). See Section 3.1 (and in



245 particular Table 2) for an analysis of the equilibrium state of both the *piControl* and the warm climate
246 simulations. Starting at the end of the *piControl*, these were then run for their own spin-up phases,
247 400 and 350 years for the *midHolocene* and *lig127k* respectively. During this phase, ~700 diagnostics
248 were output, containing mostly low temporal frequency (e.g. monthly, seasonal and annual) fields.
249 Once the simulations were considered in an acceptable level of equilibrium (see Section 3.1), a
250 production phase was run for 100 and 200 years for the *midHolocene* and *lig127k* respectively, during
251 which the full CMIP6/PMIP4 diagnostic profile (totalling ~1700 fields) was implemented to output
252 both high and low temporal frequency variables.

253

254 **2.3. Data**

255 Recent data syntheses compiling quantitative surface temperature and rainfall reconstructions were
256 used in order to evaluate the warm climate simulations.

257

258 For the MH, the global-scale continental surface mean annual temperature (MAT) and rainfall (or
259 mean annual precipitation, MAP) reconstructions from Bartlein *et al.* (2011), with quantitative
260 uncertainties accounting for climate parameter reconstruction methods, were used (see Data
261 Availability for access details). They rely on a combination of existing quantitative reconstructions
262 based on pollen and plant macrofossils and are inferred using a variety of methods (see Bartlein *et al.*
263 2011 for further details). At each site, the 6 ka anomaly (corresponding to the 5.5-6.5 ka average
264 value), is given relative to the present day, and in the case where modern values could not be directly
265 inferred from the record, modern climatology values (1961-1990) were extracted from the Climate
266 Research Unit historical climatology data set (New *et al.* 2002).

267

268 For the LIG, two different sets of surface temperature data are available. Firstly, the Capron *et al.*
269 (2017) 127 ka timeslice of SAT and sea surface temperature (SST) anomalies (relative to pre-
270 industrial, 1870-1899), is based on polar ice cores and marine sediment data that are (i) located
271 poleward of 40° latitude and (ii) have been placed on a common temporal framework (see Data
272 Availability for access details). Polar ice core water isotope data are interpreted as annual surface air
273 temperatures, while most marine sediment-based reconstructions are interpreted as summer SST
274 signals. For each site, the 127 ka value was calculated as the average value between 126 and 128 ka
275 using the surface temperature curve resampled every 0.1 ka. Secondly, a global-scale time slice of
276 SST anomalies, relative to pre-industrial (1870-1889), at 127 ka was built, based on the recent
277 compilation from Hoffman *et al.* (2017), which includes both annual and summer SST reconstructions
278 (see Data Availability for access details). The 127 ka values at each site were extracted, following the
279 methodology they proposed for inferring their 129, 125 and 120 ka time slices i.e. the SST value at
280 127 ka was taken on the provided mean 0.1 ka interpolated SST curve for each core location. Data
281 syntheses from both Capron *et al.* (2014, 2017) and Hoffman *et al.* (2017) are associated with



282 quantitative uncertainties accounting for relative dating and surface temperature reconstruction
283 methods. Here, the two datasets are treated as independent data benchmarks, as they use different
284 reference chronologies and methodologies to infer temporal surface temperature changes, and
285 therefore they should not be combined. See Capron *et al.* (2017) for a detailed comparison of the two
286 syntheses. A model-data comparison exercise using existing LIG data compilations focusing on
287 continental surface temperature (e.g. Turney and Jones 2010) was not attempted, as they do no benefit
288 yet from a coherent chronological framework, preventing the definition of a robust time slice
289 representing the 127 ka terrestrial climate conditions (Capron *et al.* 2017).

290

291 3. RESULTS

292 As briefly mentioned above, both the warm climate simulations had a spin-up phase before the main
293 production run was started. The results discussed here are therefore split into two sections: firstly,
294 assessing the level of atmospheric and oceanic equilibrium during (and, in particular, at the end of)
295 the spin-up phase, and secondly assessing the 100-year climatology from the production run.

296

297 3.1. Spin-up

298 Annual global mean 1.5 m air temperature and TOA radiation from both warm climate simulations,
299 compared to the *piControl*, are shown in Figure 2 and summarised in Table 2. Note that the *piControl*
300 spin-up phase was run in three separate parts, to accommodate for minor changes/updates in the
301 model as the simulation progressed. There is a clear increase in temperature during the beginning of
302 this period, as the *piControl* slowly spins up from its original starting point; this levels off towards the
303 end of the period, however, with a final temperature trend of $0.03^{\circ}\text{C century}^{-1}$ (Table 2 and Fig. 2a).
304 For the warm climate simulations, despite considerable interannual variability (particularly halfway
305 through the *lig127k* simulation) both are showing small long-term trends of $-0.06^{\circ}\text{C century}^{-1}$ and -
306 $0.16^{\circ}\text{C century}^{-1}$ for the last 100 years of the *midHolocene* and *lig127k*, respectively (Table 2 and Fig.
307 2a). The same is true for TOA, where the *piControl* has a slow downward trend towards zero until
308 equilibrium was reached, whereas the *midHolocene* and *lig127k* are relatively stable (Fig. 2b).

309

310 For the ocean, annual global mean OceTemp and OceSal are shown in Table 2 and Figure 3. There is
311 again a clear increase in OceTemp during the *piControl* spin-up phase, which again stabilises at
312 $0.035^{\circ}\text{C century}^{-1}$ by the end of the period (Table 2). Whilst OceTemp stabilises in the *midHolocene*
313 and indeed has a smaller trend than the *piControl* (Table 2), it continues to increase in the *lig127k*
314 until it stabilises within the last ~50 years (Fig. 3a). A similar pattern is shown in OceSal, with a
315 steady decrease in the *piControl* spin-up phase which continues during the *midHolocene* and,
316 conversely, starts to increase before stabilising during the *lig127k* (Fig. 3b). Concerning the long-
317 term trends, Menary *et al.* (2018) considered values acceptable for equilibrium to be $< \pm 0.035^{\circ}\text{C}$
318 century^{-1} and $< \pm 0.0001 \text{ psu century}^{-1}$ (for OceTemp and OceSal, respectively); as shown in Table 2,



319 although both warm climate simulations meet the temperature criterion, neither meet the salinity
320 criterion (-0.007 psu and 0.006 psu for the *midHolocene* and *lig127k*, respectively, compared to a
321 criterion of 0.0001 psu). However, running for several thousands of years (and > 5 years of computer
322 time), which would be needed to reach true oceanic equilibrium, was simply unfeasible here given
323 time and resource constraints.

324

325 **3.2. Production runs results**

326 The warm climate production runs were undertaken following the spin-up phase, with a 100-year
327 climatology of each simulation being compared to that from the *piControl*, as well as available proxy
328 data, using either annual means or summer/winter seasonal means. For the latter, depending on the
329 availability of the proxy data, Northern Hemisphere summer is defined as either June-August (JJA) or
330 July-September (JAS), and Northern Hemisphere winter is defined as either December-February
331 (DJF) or January-March (JFM); and vice versa for Southern Hemisphere summer/winter. Using
332 atmospheric diagnostics, the focus is on three separate measures: i) to describe and understand the
333 differences between the current two warm climate simulations and the *piControl* in terms of
334 temperature, rainfall and atmospheric circulation changes; ii) to compare both current simulations,
335 with existing and newly-available proxy data, and iii) to compare both current simulations with those
336 from previous versions of the UK model (where available), such as HadGEM2-ES or HadCM3, in
337 order to assess any improvements due to model advances. In this aim, previous CMIP3 and 5
338 versions of the UK model, alongside other CMIP5 models, will be assessed to address the question of
339 whether simulations produce enough rainfall to allow vegetation growth across the Sahara: the mid-
340 Holocene ‘Saharan greening’ problem.

341

342 **3.2.1. Do the CMIP6 HadGEM3 simulations show temperature, rainfall and circulation 343 differences when compared to the pre-industrial era?**

344 Here we focus on mean differences between the HadGEM3 warm climate simulations and the
345 corresponding *piControl*. Seasonal mean summer and winter 1.5 m air temperature anomalies
346 (relative to the *piControl*) from both warm climate simulations are shown in Figure 4. During JJA,
347 the *midHolocene* is showing a widespread increase in temperatures of up to 2°C across the entire
348 Northern Hemisphere north of 30°N, more in some places e.g. Greenland (Fig. 4a), consistent with the
349 increased latitudinal and seasonal distribution of insolation caused by known differences in the
350 Earth’s axial tilt (Berger & Loutre 1991, Otto-Bliesner *et al.* 2017). The only places showing a
351 reduction in temperature are West and central Africa (around 10°N) and northern India; this, as
352 discussed below, is likely related to increased rainfall in response to a stronger summer monsoon, but
353 could also be due to the resulting increase in cloud cover (reflecting more insolation) or a combination
354 of the two. During DJF, only the Northern Hemisphere high latitudes (north of 60°N) continue this



355 warming trend, with the rest of continental Africa and Asia showing a reduction in temperature (Fig.
356 4b). These patterns are virtually the same during the *lig127k* (Fig. 4c and d), just much more
357 pronounced (with temperature increases during JJA of 5°C or more); again, this is consistent with the
358 differences in the Earth's axial tilt, which were more extreme (and therefore Northern Hemisphere
359 summer experienced larger insolation changes) in the LIG relative to the MH (Berger & Loutre 1991,
360 Otto-Bliesner *et al.* 2017).

361
362 Mean JJA rainfall and 850mb wind anomalies (relative to the *piControl*) from both warm climate
363 simulations are shown in Figure 5, which zooms into Africa. In response to the increased Northern
364 Hemisphere summer insolation, the West African monsoon is enhanced in both simulations, with
365 positive (negative) rainfall anomalies across sub-Saharan Africa (eastern equatorial Atlantic)
366 suggesting a northward displacement of the ITCZ. This is consistent with previous work, with a
367 northward movement of the rainbelt being associated with increased advection of moisture into the
368 continent (Huang *et al.* 2001, Singarayer *et al.* 2017, Wang *et al.* 2014). This increased advection of
369 moisture is shown by the low-level westerlies in Figure 5, drawing in more moisture from the tropical
370 Atlantic, which are consistent with previous work documenting the intensified monsoon circulation
371 associated with a greater land-sea temperature contrast (Huang *et al.* 2001, Singarayer *et al.* 2017,
372 Wang *et al.* 2006). This pattern is enhanced in the *lig127k* relative to the *midHolocene*, again due to
373 the stronger insolation forcing in the LIG relative to the MH, and the northward displacement of the
374 ITCZ is more pronounced in the *lig127k* simulation (Fig. 5c). Interestingly, however, regarding very
375 small anomalies (i.e. < 1 mm day⁻¹), the *midHolocene* is showing wetter conditions further north,
376 throughout the Sahara and up to the Mediterranean, whereas the *lig127k* simulation has small dry
377 anomalies in this region (Fig. 5a and b for the *midHolocene* and *lig127k*, respectively).

378
379 The change to the intensity and the spatial pattern (e.g. latitudinal positioning and extent) of the West
380 African monsoon is further shown in Figure 6, which shows JJA rainfall anomalies by latitude over
381 West Africa from both warm climate simulations. Apart from the clear drying relative to the
382 *piControl* between the Equator and 5°N (which comes almost entirely from the equatorial Atlantic
383 region), both warm climate simulations are showing a large increase in rainfall (of around 2 and 6 mm
384 day⁻¹ for the *midHolocene* and *lig127k*, respectively) during the core monsoon region i.e. between
385 approximately 10-15°N. In terms of the latitudinal extent, an examination of the mean rainfall by
386 latitude suggests that both warm climate simulations are producing a wider monsoon region (i.e. both
387 North and South of the Equator), with rainfall only reducing to near zero at 20°N in these simulations
388 compared to approximately 16°N in the *piControl* (not shown). This is again consistent with previous
389 work, where various theories are compared as to the reasons behind the latitudinal changes in the
390 rainbelt's position, one which is a symmetric expansion during boreal summer (Singarayer &
391 Burrough 2015, Singarayer *et al.* 2017).



392

393 **3.2.2. Model-Data comparison: Do the CMIP6 HadGEM3 simulations reproduce the**
394 **‘reconstructed’ climate based on available proxy data?**

395 Here we focus on comparison with recent proxy data, focusing on surface temperature and rainfall
396 (drawing direct comparisons, as well as using the root mean square error (RMSE), between proxy and
397 simulated data, summarised in Table 4a), to see how well the current warm climate simulations are
398 reproducing the ‘observed’ approximate magnitudes and patterns of change. It is worth noting that
399 both simulated and proxy anomalies contain a high level of uncertainty, and in many locations the
400 uncertainty is often larger than the anomalies themselves (not shown). The following results should
401 therefore be considered with this caveat in mind.

402

403 Before the spatial patterns are compared, it is useful to assess global means (focusing on 1.5 m air
404 temperature, calculated both annually and during Northern and Southern Hemisphere summer, JJA
405 and DJF respectively) for model-model comparisons. Table 3 shows these global means, where it is
406 clear that when annual means are considered, the *midHolocene* simulation is actually cooler than the
407 *piControl*; this discrepancy with the palaeodata, which in general suggests a warmer MH relative to
408 PI, also exists in previous models, and is termed the ‘Holocene temperature conundrum’ by Lui *et al.*
409 (2014). The *lig127k* simulation is, however, warmer than the *piControl* simulation. Given the
410 seasonal distribution of insolation in these two simulations, it is expected that the largest difference to
411 the PI occurs during boreal summer, and indeed it does; during JJA, there is a warmer LIG and a
412 slightly warmer MH (1.69°C and 0.07°C, respectively). Conversely, the opposite is true during DJF.

413

414 Concerning the spatial patterns during the MH, Figure 7 shows simulated surface MAT and MAP
415 anomalies from the *midHolocene* simulation versus MH proxy anomalies from Bartlein *et al.* (2011),
416 both of which have over 600 proxy locations in total (Table 4), although mostly confined to the
417 Northern Hemisphere. For MAT, globally the simulation looks reasonable (RMSE = 2.45°C), and
418 appears to be able to reproduce the sign of temperature change for many locations, with both
419 simulated and proxy anomalies suggesting increases in temperature North of 30°N (Fig. 7a and b).
420 This is not true everywhere, such as across the Mediterranean where the simulation suggests a small
421 warming but the proxy data indicates cooling (Fig. 7a and b). However, regarding the magnitude of
422 change, the *midHolocene* simulation is underestimating the temperature increase across most of the
423 Northern Hemisphere, with for example increases of up to 1°C across Europe from the simulation
424 compared to 3–4°C increases from the proxy data (Fig. 7a and b). In the simulation, temperature
425 anomalies only reach these magnitudes in the Northern Hemisphere polar region (i.e. north of 70°N),
426 not elsewhere. A similar conclusion can be drawn from MAP (RMSE = 280 mm yr⁻¹), where again
427 the *midHolocene* simulation is correctly reproducing the sign of change across most of the Northern
428 Hemisphere, but in some places not the magnitude. Over the eastern US, for example, rainfall



429 decreases of up to 200 mm yr⁻¹ are being shown by the simulation whereas the proxy data suggests a
430 much stronger drying of up to 400 mm yr⁻¹ (Fig. 7c and d). Elsewhere, such as over Europe and
431 Northern Hemisphere Africa, the simulation more accurately reproduces the magnitude of rainfall
432 increases; both simulated and proxy anomalies show increases of 200–400 mm yr⁻¹ (Fig. 7c and d).

433

434 Concerning the spatial patterns during the LIG, Figure 8 shows simulated mean SST anomalies
435 (calculated both annually and during JAS/JFM) from the *lig127k* simulation and LIG proxy anomalies
436 from two sources, Capron *et al.* (2017) and Hoffman *et al.* (2017). When annual anomalies are
437 considered, despite the lack of reconstructions in the Capron *et al.* (2017) data (Table 4), there is
438 relatively good agreement (RMSE = 2.44°C and 2.94°C for the Capron *et al.* (2017) and Hoffman *et al.*
439 *et al.* (2017) data, respectively, and which is within the average uncertainty range), between simulated
440 and observed SST anomalies in the Northern Hemisphere (and in particular in the North Atlantic),
441 with both suggesting increased temperatures during the LIG of up to 3°C (Fig. 8a). There are
442 discrepancies, such as in the Norwegian Sea, where the Hoffman *et al.* (2017) reconstructions suggest
443 a cooler LIG than preindustrial, whereas the *lig127k* simulation shows a consistent warming; this is,
444 however, consistent with previous work, and earlier climate models have also failed to capture this
445 cooling (Capron *et al.* 2014, Stone *et al.* 2016). Note that, over Greenland and Antarctica, the Capron
446 *et al.* (2017) proxy data show SAT, not SST, and are therefore not compared in this figure;
447 comparison with simulated SAT, however, suggests that the model is capturing the sign, if not the
448 magnitude, of annual change over these regions (not shown). During Northern Hemisphere summer,
449 JAS (during which period Capron *et al.* [2017] has the most proxy locations [Table 4]), the simulated
450 anomalies are in agreement with many, but not all, of the proxy locations (RMSE = 3.11°C and
451 2.06°C for the Capron *et al.* (2017) and Hoffman *et al.* (2017) data, respectively); examples of where
452 they differ, not just in magnitude but also sign, again include the Norwegian and Labrador Seas (Fig.
453 8b). In Southern Hemisphere summer, JFM, the model suggests a general (but weak) cooling in the
454 South Atlantic relative to preindustrial and a general (but weak) warming in the Southern Ocean (Fig.
455 8c). However, certain proxy locations (such as off the coast of southern Africa) suggest a much
456 warmer LIG than preindustrial (RMSE = 1.94°C and 4.24°C for the Capron *et al.* (2017) and Hoffman
457 *et al.* (2017) data, respectively), which in stark contrast to the cooling in the same region from the
458 *lig127k* simulation (Fig. 8c). In the Southern Ocean, the majority of simulated anomalies reproduce
459 the observed sign of change, but not the magnitude; the *lig127k* simulation suggests temperature
460 increases of up to 1°C, whereas both proxy datasets suggest SST increases of 2–3°C depending on
461 location (Fig. 8c).

462

463 It would therefore be reasonable to say that, for both warm climate simulations, whilst the model is
464 capturing the sign and magnitude of change (for either temperature or rainfall) in some locations, this
465 is highly geographically dependent and there are locations where the simulation fails to capture even



466 the sign of change. The model also appears to be seasonally dependent, with the *lig127k* simulation
467 (but not the *midHolocene* simulation) correctly reproducing both the sign and magnitude of change
468 during Northern Hemisphere summer in some locations, but not during Southern Hemisphere summer
469 or annually.

470

471 **3.2.3. Model-Model comparison: Do the CMIP6 HadGEM3 simulations show an improvement** 472 **compared to older CMIP versions of the UK model?**

473 Here we focus on model-model intercomparisons, comparing the HadGEM3 warm climate
474 simulations with firstly those from previous versions of the UK model and secondly with those from
475 other models included in CMIP5. It should be noted that although LIG experiments have been
476 conducted previously with both model-model and model-data comparisons being made (Lunt *et al.*
477 2013), all of these experiments were carried out using early versions of the models and were thus not
478 included in CMIP5. Moreover, as part of their assessment Lunt *et al.* (2013) considered a set of four
479 simulations, at 130, 128, 125 and 115 ka, none of which are directly comparable to the current
480 HadGEM3 *lig127k* simulation. Instead, a LIG simulation has recently been undertaken using one of
481 the original versions of the UK's physical climate model, HadCM3, and so this is used here to
482 compare with the *lig127k* simulation. As discussed above, this section is divided into two parts:
483 firstly the mean climate state of the warm climate simulations will be compared to the model's
484 predecessors, focusing again on hydroclimate of the West African monsoon (given the known
485 problem of simulated rainfall underestimation in this region, see e.g. Braconnot *et al.* [2007]). Here,
486 both direct comparisons and RMSE values will again be examined, this time calculating the RMSE
487 between the simulated rainfall anomaly from two older versions of the UK model versus the current
488 HadGEM3 *midHolocene* and *lig127k* simulations (summarised in Table 4b). Secondly, previous
489 generation simulations (from all available models included in CMIP5) will be compared to see
490 whether the most recent HadGEM3 *midHolocene* simulation is now providing enough rainfall to
491 allow vegetation growth across the Sahara; something which previous generations of models from
492 CMIP5 did not (Braconnot *et al.* 2007).

493

494 **3.2.3.1. Mean climate state from predecessors of HadGEM3**

495 Regarding the magnitude and latitudinal extent of the West African monsoon, Figure 9 shows the JJA
496 rainfall differences averaged over West Africa from the current *midHolocene* and *lig127k* simulation
497 versus two of the model's predecessors. During the MH, the two most recent generations of the
498 model (HadGEM3 and HadGEM2-ES) generally agree on drier conditions over the equatorial
499 Atlantic and then wetter conditions over West Africa, however the oldest generation model
500 (HadCM3) does not reproduce the Atlantic drying. Likewise the two most recent generations share a
501 similar latitudinal distribution of rainfall above $\sim 5^{\circ}\text{N}$, with a wetter MH over land, peaking at $\sim 2\text{-}3$
502 mm day^{-1} at $\sim 11\text{-}12^{\circ}\text{N}$. Interestingly, the previous version of the model (HadGEM2-ES) shows the



503 strongest and most northwardly displaced rainfall peak, as discussed in previous work (e.g. Huag *et al.*
504 2001, Otto-Bliesner *et al.* 2017, Singarayer *et al.* 2017, Wang *et al.* 2014); the most recent version,
505 HadGEM3, has lower northward displacement compared to the two older versions of the model. Both
506 recent versions suggest that the monsoon region extends to $\sim 17^\circ\text{N}$, above which the differences
507 between the MH and PI reduce to near zero. In contrast, HadCM3 suggests a generally weaker, but
508 latitudinally more extensive, monsoon region, suggesting a wetter MH (by $\sim 1 \text{ mm day}^{-1}$) as far north
509 as 20°N and beyond. For the LIG, HadGEM3 is showing a much stronger monsoon region relative to
510 the *piControl*, compared to HadCM3. However, in terms of extent, similar results are shown to those
511 for the MH, with HadCM3 showing a generally weaker, but more northwardly displaced, monsoon
512 region. In this older generation model, positive rainfall anomalies of $\sim 2\text{-}3 \text{ mm day}^{-1}$ extend as far
513 north as $17\text{-}18^\circ\text{N}$, whereas in HadGEM3 they fall to $\sim 1 \text{ mm day}^{-1}$ at these latitudes.

514

515 In terms of the spatial patterns of the West African monsoon, Figure 10 and Figure 11 show the JJA
516 daily rainfall climatology differences from the same three model generations for the MH and LIG,
517 respectively. During the MH, consistent with Figure 9, the two most recent simulations generally
518 agree (RMSE = 0.46 mm day^{-1}) and show similar spatial patterns, with a drier equatorial Atlantic
519 during the MH and then increased rainfall around 10°N (Fig. 10a and b for HadGEM3 and
520 HadGEM2-ES, respectively). Both simulations also suggest that the increases in rainfall extend
521 longitudinally across the entire continent, with the largest changes not only occurring across western
522 and central regions but also further east. In contrast, HadCM3 is less consistent than HadGEM3
523 (RMSE = 0.53 mm day^{-1}) and only suggests a wetter MH over West Africa; moreover, again
524 consistent with Figure 9, HadCM3 suggests that although the West African monsoon region is
525 longitudinally narrower, it is latitudinally wider than the other two simulations (Fig. 10c). HadCM3
526 also differs from the other simulations over the equatorial Atlantic, showing a region of drying that is
527 not only stronger in magnitude (with the MH being over 5 mm day^{-1} drier than the PI in HadCM3,
528 compared to $\sim 2\text{-}3 \text{ mm day}^{-1}$ in the two most recent simulations), but also larger in terms of latitude
529 and longitude extent (Fig. 10c).

530

531 During the LIG, only the most recent and oldest version of the model can be compared, as a LIG
532 simulation using HadGEM2-ES is unavailable. In Figure 11 there is a noticeable difference between
533 generations and the level of agreement is the lowest across all simulation combinations (RMSE = 1.57
534 mm day^{-1}), with the most recent HadGEM3 showing greatly increased rainfall across all of northern
535 Africa, centred on 10°N but extending from $\sim 5^\circ\text{N}$ to almost 20°N and beyond (Fig. 11a), again
536 consistent with Figure 9. In contrast, and similar to the MH results, in HadCM3 the largest rainfall
537 increases are confined to Western Africa only, rather than extending longitudinally across the
538 continent (Fig. 11b). However, in terms of latitudinal extent, HadCM3 is showing weak wet
539 anomalies all the way to the Mediterranean, whereas the monsoon region diminishes further south (at



540 ~30°N) in HadCM3 and dry anomalies are suggested North of this. Another noticeable difference is
541 the region of drying, with the most recent generation model placing this over the equatorial Atlantic
542 (consistent with the MH) but HadCM3 shifting this further east, over most of central Africa (Fig.
543 11b). The region of equatorial Atlantic drying shown by the more recent versions of the model is
544 actually wetter during this HadCM3 LIG simulation.

545

546 It would therefore appear that, for the MH, whilst there is less difference between the most recent two
547 configurations of the model (in terms of a more localised West African monsoon region), there
548 nevertheless has been improvement since the oldest version of the UK's physical climate model. For
549 the LIG, where unfortunately there is no intermediate generation, it would be reasonable to say that
550 again considerable change has occurred since the oldest generation model, with the suggestion that,
551 although HadCM3 is identifying an enhanced monsoon which extends to the Mediterranean (albeit
552 with very weak anomalies), at lower latitudes it is not showing the level of northward displacement as
553 the most recent version, apart from in the far western regions.

554

555 3.2.3.2. *Rainfall across the Sahara*

556 Given that the warm climate simulations, and indeed the *piControl*, did not use interactive, but rather
557 prescribed, vegetation, it is not possible to directly test if the model is reproducing the 'Saharan
558 greening' that proxy data suggest. For example, Jolly *et al.* (1998a, 1998b) analysed MH pollen
559 assemblages across northern Africa and suggested that some areas south of 23°N (characterised by
560 desert today) were grassland and xerophytic woodland/scrubland during the MH (Joussaume *et al.*
561 1999). To circumvent this caveat, Joussaume *et al.* (1999) developed a method for indirectly
562 assessing Saharan greening, based on the annual mean rainfall anomaly relative to a given model's
563 modern simulation. Using the water-balance module from the BIOME3 equilibrium vegetation model
564 (Haxeltine & Prentice 1996), Joussaume *et al.* (1999) calculated the increase in mean annual rainfall,
565 zonally averaged over 20°W-30°E, required to support grassland at each latitude from 0 to 30°N,
566 compared to the modern rainfall at that latitude. This was then used to create maximum and
567 minimum estimates, within which bounds the model's annual mean rainfall anomaly must lie to
568 suggest enough of an increase to support grassland (Joussaume *et al.* 1999).

569

570 Therefore, an adapted version of Figure 3a in Joussaume *et al.* (1999) is shown here in Figure 12,
571 which includes the above mean annual rainfall anomalies from not only the current *midHolocene*
572 simulation, but also all previous MH simulations from CMIP5. Firstly of note is that, despite the
573 equatorial Atlantic drying that all the models show (seen, for example, in Figure 5), the HadGEM3
574 *midHolocene* simulation is showing a peak in rainfall further south compared to many other CMIP5
575 models, suggesting less northward displacement of the rainbelt relative to the other models (Fig. 12).
576 Concerning the threshold required to support grassland, it is clear that although the current



577 *midHolocene* simulation is showing an increase in mean annual rainfall further north than some of the
578 models, including its predecessor HadGEM2-ES, and is just within the required bounds at lower
579 latitudes (e.g. up to 17°N), north of this the current *midHolocene* simulation is not meeting the
580 required threshold, neither are any of the other CMIP5 models after ~18°N (Fig. 12). It would
581 therefore appear that, although some improvement has been made since CMIP5 and earlier models,
582 the latest version of the UK's physical climate model it is still unable to reproduce the amount of
583 rainfall necessary to give the 'Saharan greening' suggested by proxy data during the MH.

584

585 4. SUMMARY AND CONCLUSIONS

586 This study has conducted and assessed the mid-Holocene and Last Interglacial simulations using the
587 latest version of the UK's physical climate model, HadGEM3-GC3.1, comparing the results with
588 available proxy data, previous versions of the same model, and other models from CMIP's previous
589 iteration, CMIP5. Both the *midHolocene* and *lig127k* simulations followed the experimental design
590 defined in Otto-Bliesner *et al.* (2017) and under the auspices of CMIP6/PMIP4. Both simulations
591 were run for a 350-400 year spin-up phase, during which time atmospheric and oceanic equilibrium
592 was assessed, and once an acceptable level of equilibrium had been reached, the production runs were
593 started.

594

595 Concerning the results from the spin-up phase, comparison to the metrics used to assess the CMIP6
596 *piControl* suggest that both warm climate simulations reached an acceptable state of equilibrium, in
597 the atmosphere at least, to allow the production runs to be undertaken. From these, both simulations
598 are showing global temperatures consistent with the latitudinal and seasonal distribution of insolation,
599 and with previous work (e.g. Otto-Bliesner *et al.* 2017). Globally, whilst both the simulations are
600 mostly capturing the sign and, in some places, magnitude of change relative to the PI, similar to
601 previous model simulations this is geographically and seasonally dependent. It should be noted that
602 the proxy data (against which the simulations are evaluated) also contain a high level of uncertainty in
603 both space and time, and so it is encouraging that the simulations are generally reproducing the large-
604 scale sign of change, if not at an individual location. Likewise, the behaviour of the West African
605 monsoon in both simulations is consistent with current understanding (e.g. Huag *et al.* 2001,
606 Singarayer *et al.* 2017, Wang *et al.* 2014), which suggests a wetter (and possibly latitudinally wider,
607 and/or northwardly displaced) monsoon during the MH and LIG, relative to the PI. Regarding model
608 development in simulating the West African monsoon, although there has been an improvement since
609 the oldest version of the UK's physical climate model (HadCM3), the two most recent version of the
610 model yield similar results in terms of both intensity and position. Lastly, regarding the well-
611 documented 'Saharan greening' during the MH, results here suggest that the most recent version of
612 the UK's physical climate model is consistent with all other previous models to date.

613



614 In conclusion, the results suggest that the most recent version of the UK's physical climate model is
615 reproducing climate conditions consistent with the known changes to insolation during these two
616 warm periods, and is consistent with previous versions of the same model, and other models. Even
617 though the *lig127k* simulation did not contain any influx of Northern Hemisphere meltwater, shown
618 by previous work to be a critical forcing in LIG warming, it is still nevertheless showing increased
619 temperatures in certain regions. A potential caveat of this conclusion, however, is the matter of spin-
620 up and the fact that neither of the current warm climate simulations were in oceanic equilibrium when
621 the production runs were undertaken. The production runs were undertaken nevertheless because the
622 resources required to run for several thousands of years (needed to reach true oceanic equilibrium)
623 would have been impossible to obtain, but future simulations using this model should endeavour to
624 obtain a better level of oceanic equilibrium. Another limitation of using this particular version of the
625 model is that certain processes, such as vegetation and atmospheric chemistry, were prescribed, rather
626 than allowed to be dynamically evolving. Moreover, for reasons of necessity some of the boundary
627 conditions were left as PI, such as vegetation, surface like, anthropogenic deforestation and aerosols; a
628 better simulation might be achieved if these were prescribed for the MH. Processes and boundary
629 conditions such as these may be of critical importance regarding climate sensitivity during the MH
630 and the LIG, and therefore ongoing work is underway to repeat both of these experiments using the
631 most recent version of the UK's Earth Systems model, UKESM1. Here, although the atmospheric
632 core is HadGEM3, UKESM1 contains many other earth system components (e.g. dynamic
633 vegetation), and therefore in theory should be able to better reproduce these paleoclimate states.

634

635 DATA AVAILABILITY

636 For the MH reconstructions, the data can be found within the Supplementary Online Material of
637 Bartlein *et al.* (2011), at <https://link.springer.com/article/10.1007/s00382-010-0904-1>. For the LIG
638 reconstructions, the data can be found within the Supplementary Online Material of Capron *et al.*
639 (2017), at <https://www.sciencedirect.com/science/article/pii/S0277379117303487?via%3Dihub>, and
640 the Supplementary Online Material of Hoffman *et al.* (2017), at
641 <https://science.sciencemag.org/content/suppl/2017/01/23/355.6322.276.DC1>. The model simulations
642 will be uploaded in early 2020 to the Earth System Grid Federation (ESGF) WCRP Coupled Model
643 Intercomparison Project (Phase 6), but are not yet available. The simulations are currently available
644 by directly contacting the lead author.

645

646 COMPETING INTERESTS

647 The authors declare that they have no conflict of interest.

648

649 AUTHOR CONTRIBUTION



650 CJRW conducted the *midHolocene* simulation, carried out the analysis, produced the figures, wrote
651 the majority of the manuscript, and led the paper. MVG conducted and provided the *lig127k*
652 simulation, and contributed to some of the analysis and writing. EC provided the proxy data, and
653 contributed to some of the writing. IMV provided the HadCM3 LIG simulation. PJV provided the
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656

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871 Note - For temperature, Menary *et al.* (2018) provide SAT. For OceTemp and OceSal, these were
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873 fields were calculated down to a depth of 1045m

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879 a) proxy data from Bartlein *et al.* (2011) for the MH and Capron *et al.* (2017) / Hoffman *et al.* (2017)
880 for the LIG; b) MH and LIG simulations from previous versions of UK model. Regarding the proxy
881 data comparisons in a), for JAS the simulated SST anomalies are compared to Northern Hemisphere
882 summer reconstructions and for JFM the simulated SST anomalies are compared to Southern
883 Hemisphere summer reconstructions. Note that, as shown in Figure 8, proxy locations show SST over
884 ocean and SAT over Greenland/Antarctica; to calculate RMSE values, however, only SST data were
885 used

886

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912 gridded anomalies from HadGEM3 (*midHolocene* production run – *piControl* production run, 100-
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918 anomalies (*lig127k* production run – *piControl* production run) from HadGEM3 (100-year
919 climatology), circles show proxy data (LIG – preindustrial) from Capron *et al.* (2017) and triangles
920 show proxy data (LIG – preindustrial) from Hoffman *et al.* (2017). Proxy data locations are projected
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922 summer (JFM). Note that proxy locations show SST over ocean and SAT over Greenland/Antarctica

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924 Figure 9 – JJA daily rainfall climatology differences (MH and LIG-PI) by latitude, averaged over
925 West Africa (20°W-30°E, including both land and ocean points), for the various generations of the
926 UK's physical climate model, 100-year climatology from each (50-year climatology for HadCM3
927 LIG). Solid lines show MH simulations, dotted lines show LIG simulations. Note that due to the low
928 spatial resolution in HadCM3, values in between latitude points have been interpolated

929

930 Figure 10 – JJA daily rainfall climatology differences (MH-PI) for the various generations of the
931 UK's physical climate model, 100-year climatology from each: a) HadGEM3; b) HadGEM2-ES; c)
932 HadCM3

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934 Figure 11 – JJA daily rainfall climatology differences (LIG-PI) for the various generations of the
935 UK's physical climate model, 100-year climatology from HadGEM3, 50-year climatology from
936 HadCM3: a) HadGEM3; b) HadCM3

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938 Figure 12 – Annual mean rainfall over West Africa, zonally averaged from 20°W-30°E, HadGEM3
939 and CMIP5 *midHolocene* production run minus corresponding *piControl* production runs, 100-year
940 climatology. Solid line shows HadGEM3, dotted lines show CMIP5 simulations. Grey dashes show
941 maximum and minimum bounds of the increase in rainfall required to support grassland at each
942 latitude, within which simulations must lie if producing enough rainfall to support grassland
943



944 TABLES

	<i>piControl</i>	<i>midHolocene</i>	<i>lig127k</i>
Astronomical parameters			
Eccentricity	0.016764	0.018682	0.039378
Obliquity	23.459	24.105°	24.04°
Perihelion-180°	100.33	0.87°	275.41°
Date of vernal equinox	March 21 at noon	March 21 at noon	March 21 at noon
Trace gases			
CO₂	284.3 ppm	264.4 ppm	275 ppm
CH₄	808.2 ppb	597 ppb	685 ppb
N₂O	273 ppb	262 ppb	255 ppb
Other GHG gases	CMIP DECK <i>piControl</i>	CMIP DECK <i>piControl</i>	CMIP DECK <i>piControl</i>

945

946 Table 1 - Astronomical parameters and atmospheric trace gas concentrations used in HadGEM3
 947 simulations

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Variable	<i>piControl</i>	<i>midHolocene</i>	<i>lig127k</i>
TOA (W m²)	-0.002	-0.05	-0.06
1.5 m air temp (°C)	0.03	-0.06	-0.16
OceTemp (°C)	0.035	0.0002	0.02
OceSal (psu)	0.0001	-0.007	0.006

951

952 Table 2 - Trends (per century) in global mean measures of climate equilibrium for the last hundred
 953 years of the simulations, adapted from and including *piControl* results from Menary *et al.* (2018).

954 Note - For temperature, Menary *et al.* (2018) provide SAT. For OceTemp and OceSal, these were

955 calculated using the full-depth ocean for the *piControl*, whereas in the other two simulations these

956 fields were calculated down to a depth of 1045m

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Time period	Means (°C)			Anomalies (°C)	
	<i>piControl</i>	<i>midHolocene</i>	<i>lig127k</i>	<i>midHolocene – piControl</i>	<i>lig127k – piControl</i>
Annual	13.8	13.67	14.29	-0.12	0.49
JJA	15.68	15.75	17.37	0.07	1.69
DJF	11.86	11.55	11.39	-0.31	-0.47

959

960 Table 3 - Global 1.5 m air temperature means and anomalies from HadGEM3 *piControl*,

961 *midHolocene* and *lig127k* production runs (100-year climatology)

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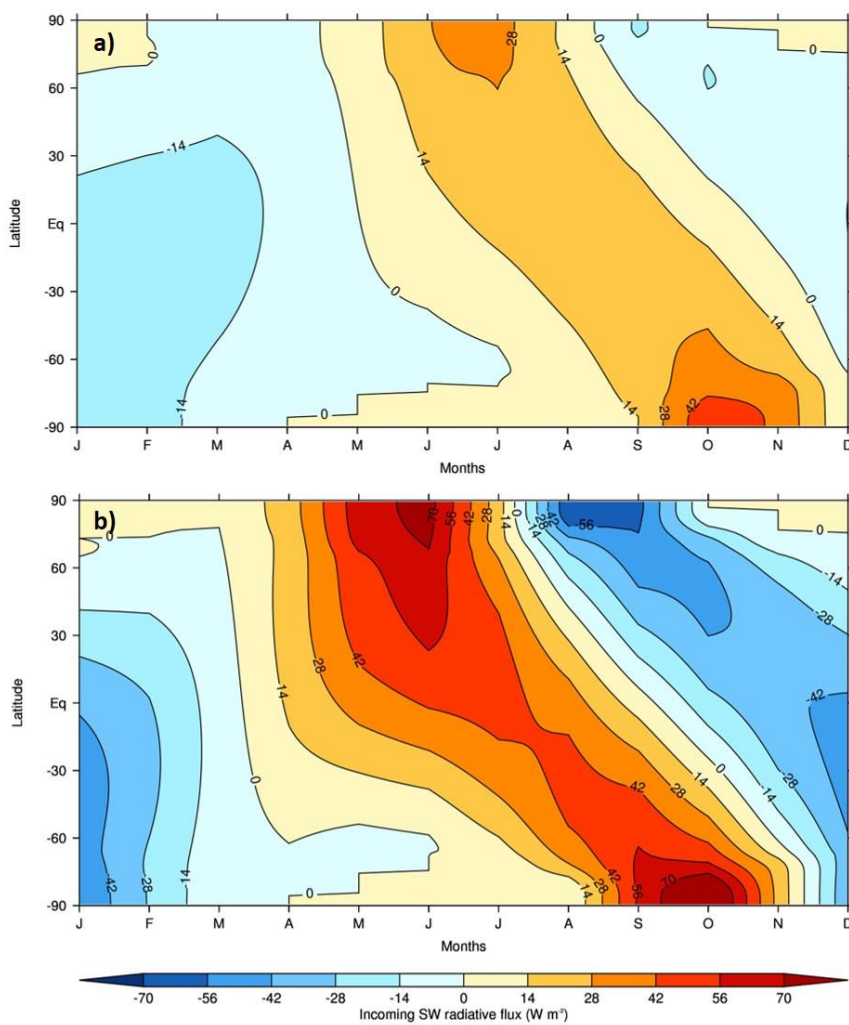
Metric	a) Simulations versus proxy data					
	MH	LIG				
MAT (°C)	2.45	Capron <i>et al.</i> (2017)			Hoffman <i>et al.</i> (2017)	
No. of proxy locations	638					
MAP (mm year ⁻¹)	280					
No. of proxy locations	651					
SST (°C)	Yearly	JAS	JFM	Yearly	JAS	JFM
	2.44	3.11	1.94	2.94	2.06	4.24
No. of proxy locations	7	24	15	86	12	6
JJA rainfall (mm day ⁻¹)	b) Simulations versus simulations					
	MH			LIG		
	HadGEM2-ES v HadGEM3	HadCM3 v HadGEM3		HadCM3 v HadGEM3		
	0.46	0.53		1.57		

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Table 4 - RMSE values for *midHolocene* and *lig127k* production runs (100-year climatology) versus: a) proxy data from Bartlein *et al.* (2011) for the MH and Capron *et al.* (2017) / Hoffman *et al.* (2017) for the LIG; b) MH and LIG simulations from previous versions of UK model. Regarding the proxy data comparisons in a), for JAS the simulated SST anomalies are compared to Northern Hemisphere summer reconstructions and for JFM the simulated SST anomalies are compared to Southern Hemisphere summer reconstructions. Note that, as shown in Figure 8, proxy locations show SST over ocean and SAT over Greenland/Antarctica; to calculate RMSE values, however, only SST data were used



985 FIGURES



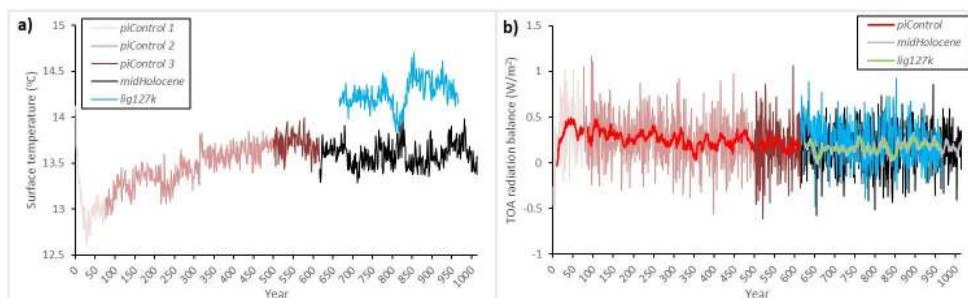
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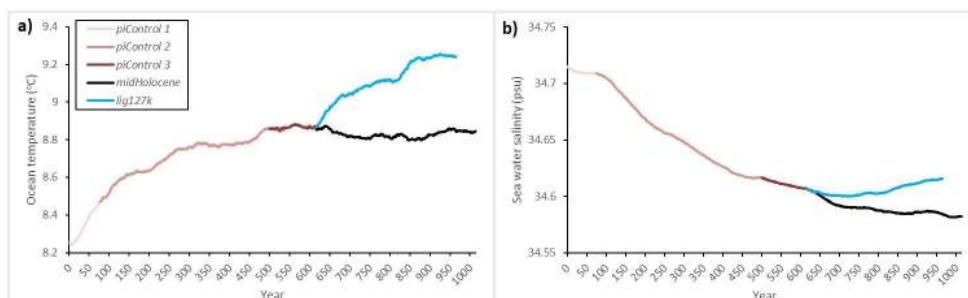


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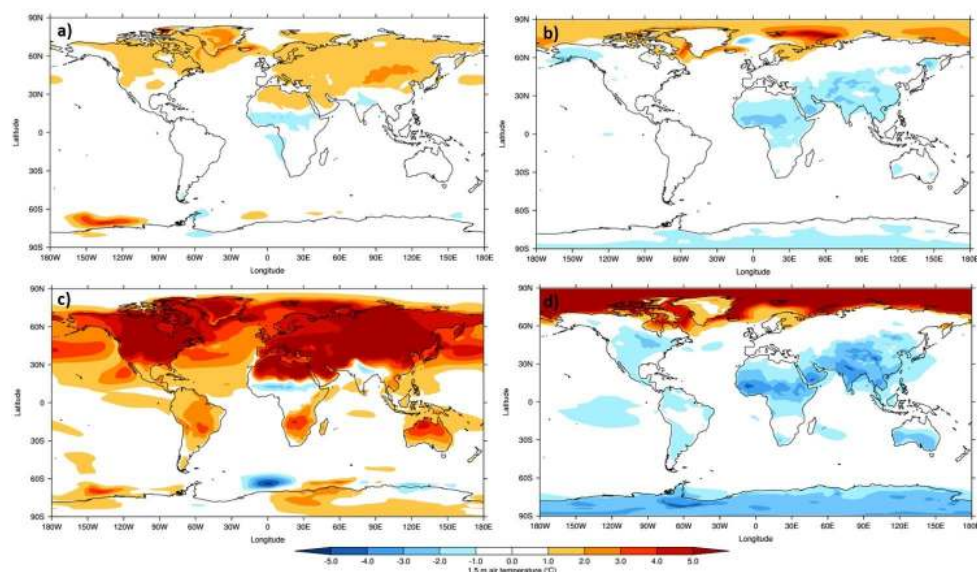


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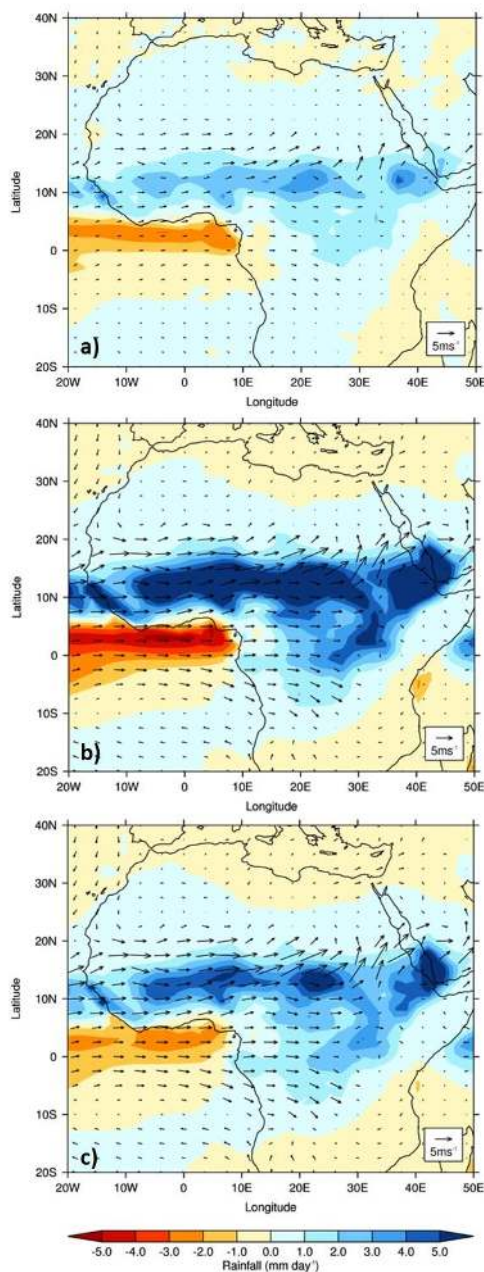
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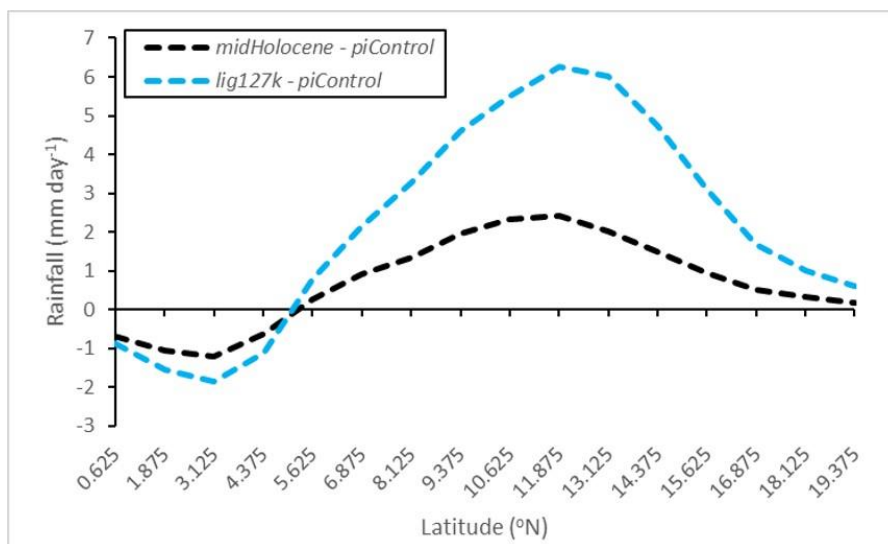
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1004 *lig127k* – *piControl*, DJF

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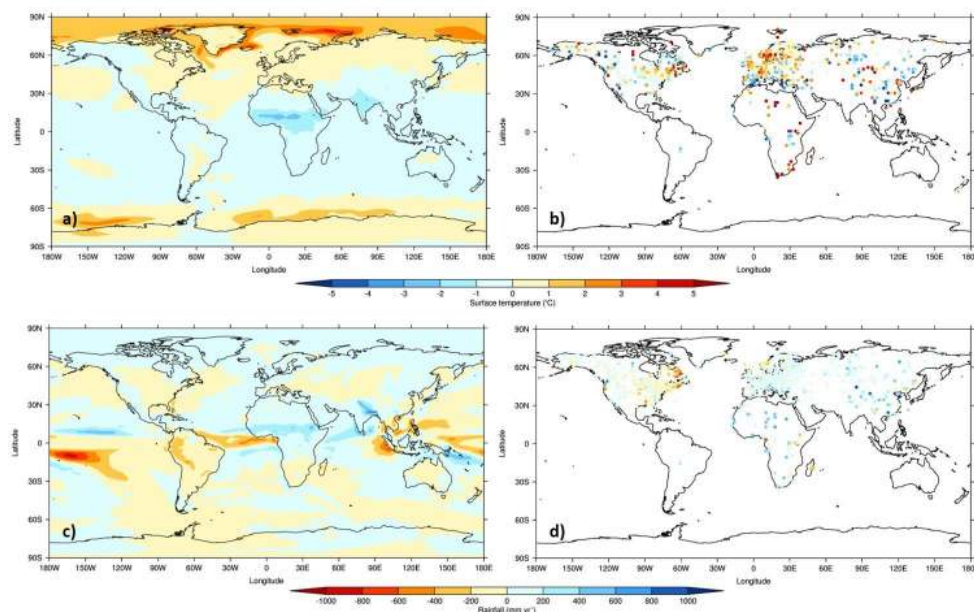
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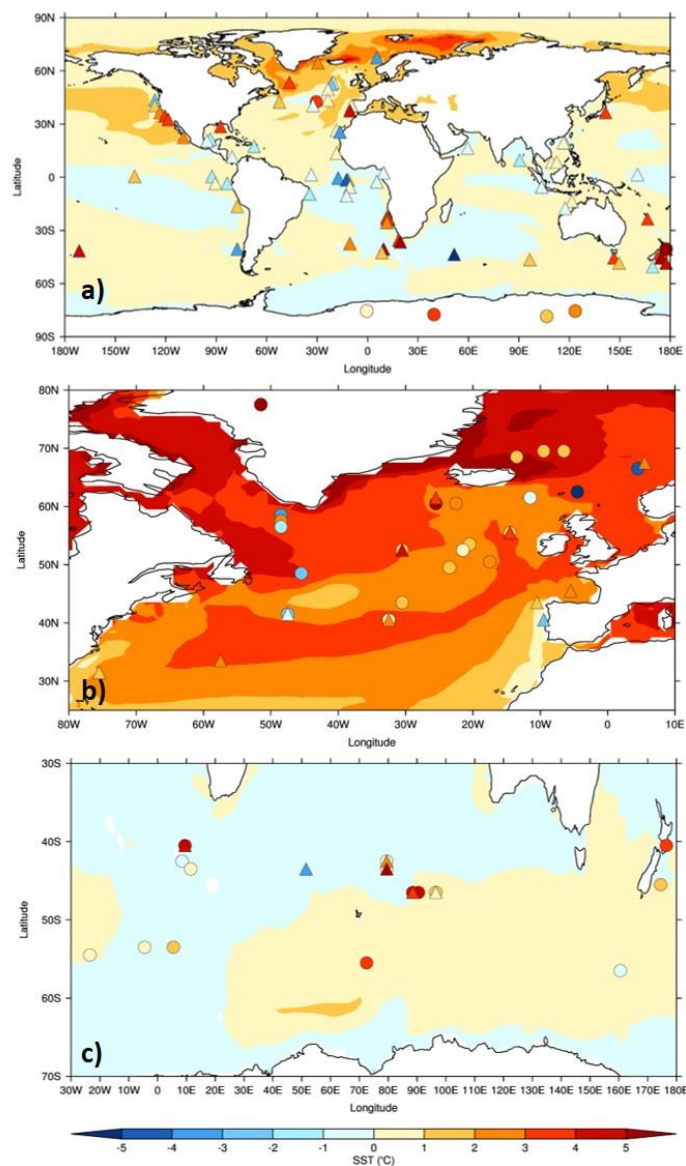
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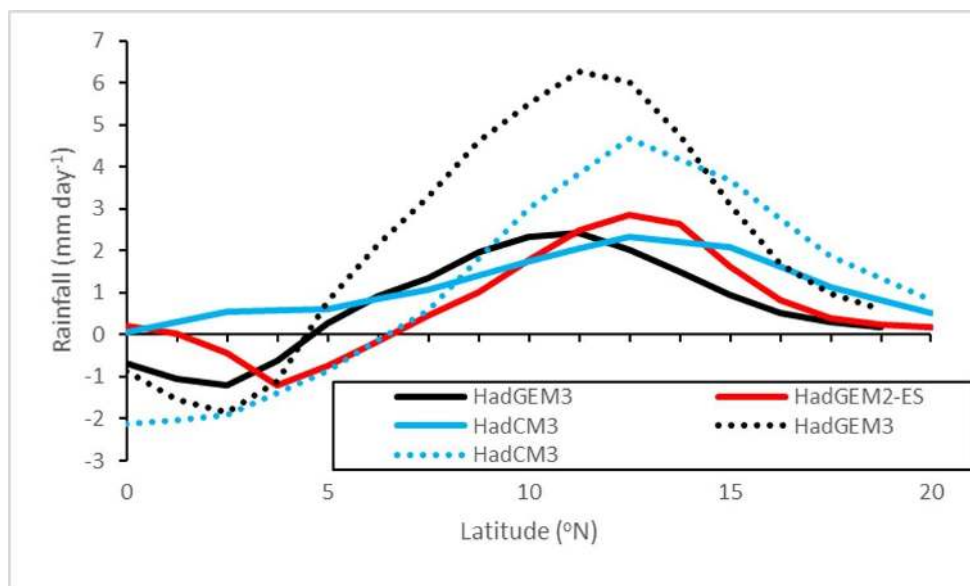
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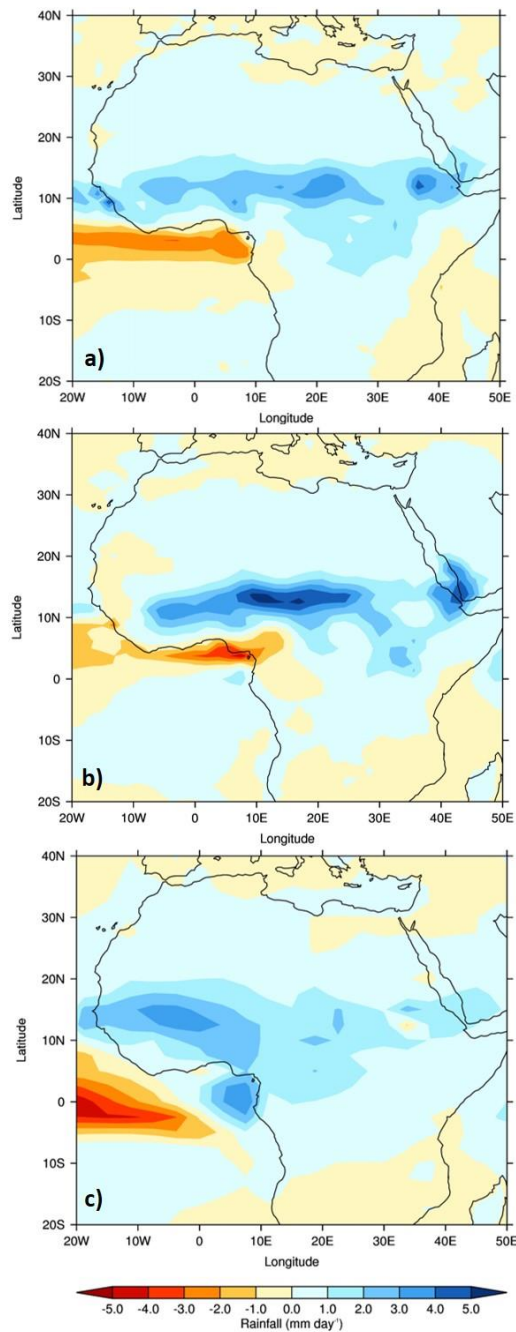
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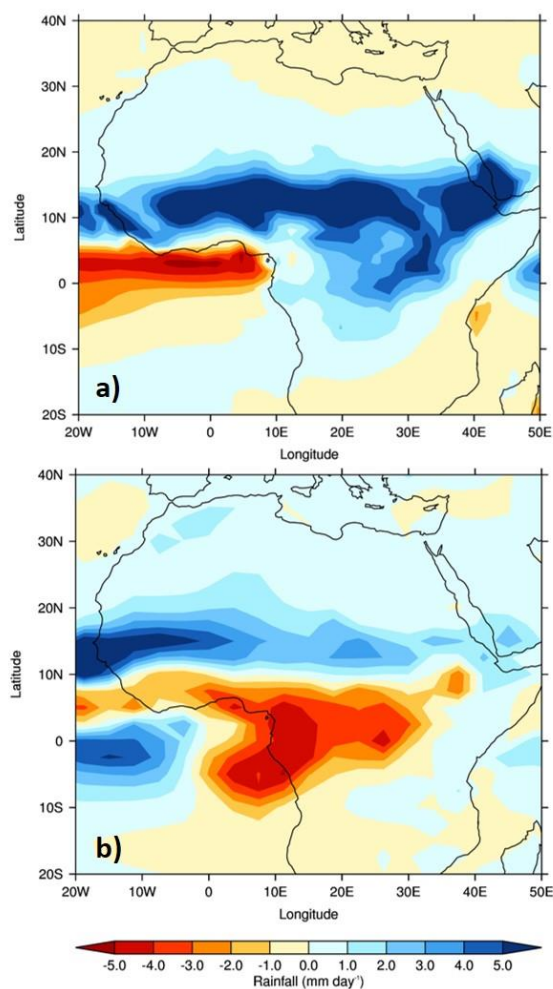


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1040 HadCM3



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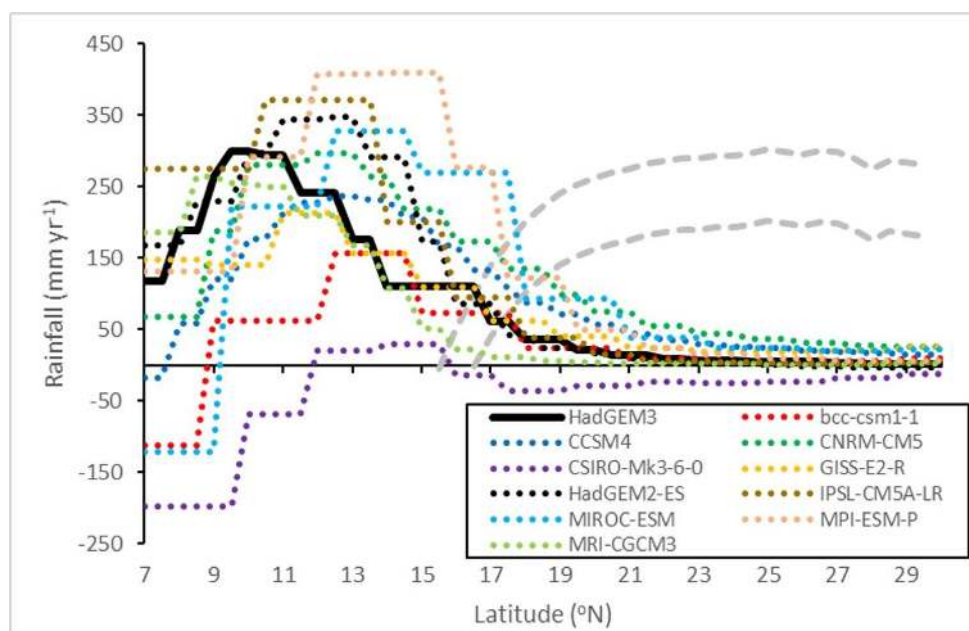
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1043 Figure 11 – JJA daily rainfall climatology differences (LIG-PI) for the various generations of the
1044 UK's physical climate model, 100-year climatology from HadGEM3, 50-year climatology from
1045 HadCM3: a) HadGEM3; b) HadCM3

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1049 Figure 12 – Annual mean rainfall over West Africa, zonally averaged from 20°W-30°E, HadGEM3
1050 and CMIP5 *midHolocene* production run minus corresponding *piControl* production runs, 100-year
1051 climatology. Solid line shows HadGEM3, dotted lines show CMIP5 simulations. Grey dashes show
1052 maximum and minimum bounds of the increase in rainfall required to support grassland at each
1053 latitude, within which simulations must lie if producing enough rainfall to support grassland

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